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(Eds.)

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Continental Evolution: The Geology of Morocco

Structure, Stratigraphy, and Tectonics of the Africa-Atlantic-Mediterranean Triple Junction

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Frontispiece: *Dicranurus monstrusus* (Lower Pragian), J. Issimour (Maider, Eastern Anti-Atlas). Courtesy M. Ihmadi (*Ihmadi Trilobite Center,* Alnif, Morocco)

André Michard · Omar Saddiqi ·
Ahmed Chalouan ·
Dominique Frizon de Lamotte (Eds.)

Continental Evolution: The Geology of Morocco

Structure, Stratigraphy, and Tectonics of the Africa-Atlantic-Mediterranean Triple Junction

With 235 Color Figures

Springer

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Prefaces

Preface from Her Excellency the Minister of Energy, Mines, Water and Environment, Director of ONHYM

Due to its geological diversity, related to the succession of four geological cycles from the Archean to the Alpine Orogeny, and including the major Pan-African and Variscan cycles, Morocco has always been characterized by intense mining activity. The earliest mining works include the remains of exploitations at Imiter (Ag), Bleida (Cu), Zgounder (Ag), Iourirn (Au), Tazalaght (Cu) and have been dated from the 9th to 15th centuries. Today, Morocco is the world's leading exporter for phosphates and produces significant quantities of a wide variety of mineral resources such as silver, zinc, lead, cobalt, manganese, fluorite, barite, bentonite, etc.

The development of the Earth Sciences in Morocco undergone a surge of new activity during the last two decades, based on the diversity of its substratum and on the international achievements in this domain. This new impulse resulted in high level research programs in the frame of national and international partnerships, and in the launching of the National Plan for Geological Mapping aimed at providing a geological infrastructure to the country, and including geological, geophysical and geochemical mapping. This intense activity resulted in a better understanding of the geodynamic framework, based on a more precise stratigraphy and on the systematic use of seismic profiles, geophysical modelling experiments, and modern geotechnologies such as SHRIMP isotopic datings and fission track thermochronology.

It is precisely within this context that the work entitled *Continental Evolution: Geology of Morocco* is presented by Pr. A. Michard and an international, multidisciplinary team. This work follows an earlier book published by the same author in 1976 under the title "Eléments de Géologie Marocaine". As a specialist in tectonics, Pr. Michard is familiar with Moroccan geology. Since the early 70's, he has participated in the education of a number of Moroccan geologists, and today he continues to supervise theses and co-author papers on the geology of the country. In fact, two of his co-editors and several of his co-authors are former students of him. Pr. Michard also supervises the mapping activity of geologists of the Ministry of Energy, Mines, Water and the Environment.

This new book provides an updated overview of the geological knowledge of Morocco based on the achievements of the last few decades. The authors pay homage to their predecessor's work with a large set of references providing a complete review and insight into the geology of Morocco. After an outline of the geological structure and evolution of the country, in the context of the African continent and peri-Atlantic plate tectonics, the book considers successively the Pan-African Belt with a refined geodynamic interpretation; the Variscan events in the Anti-Atlas and southern Morocco as well as in the Meseta Domain; the Atlas System, Rif Belt, Atlantic Basins and Mesetan and Saharan Plateaus, all of which developed during the Alpine cycle; and finally the Quaternary deposits.

We are convinced that this work will strengthen the position of Moroccan geology in the international community of geosciences. This book constitutes essential reading for operators in the sectors of mines, petroleum and the environment. It also provides key insights into the major problems of continental crust evolution from the Archean to the Present.

Prepared by a multi-disciplinary team of specialists in teaching and transfer of knowledge, this work will become the standard work for future generations of geoscientists seeking to understand the geology of Morocco.

We warmly congratulate the entire team of experts who undertook the realization of this work, and hope to see it distributed widely in Morocco and elsewhere.

Amina Benkhadra

Preface from the International Lithosphere Program

Morocco constitutes a unique natural laboratory at the junction between the Atlantic margin, the Mediterranean and the African craton. There, long lasting interactions between mantle and lithosphere dynamics, crustal extension and compression, subsidence and uplift, can be documented in great detail using surface geological records preserved in synrift, postrift and foreland sedimentary series. Outcrops are excellent in the inverted basins of the Atlas, whereas most low lands and the offshore are covered by industry seismic profiles, which have been calibrated by numerous exploration wells.

In October 2007, the geology of Morocco has attracted the interest of more than 700 geologists coming from its conjugate margins in Canada and the US, from Spain, France and other European countries, as well as from other North African countries, all these earth scientists convening for an international meeting hosted by the MAPG, with a strong participation of the ILP Task force VI on Sedimentary Basins.

Morocco has become the focus of discussion for both the academic and industry international communities, which will strongly benefit from this updated compilation of the Moroccan geology. This new Springer volume, co-authored by 36 contributors from various Moroccan (Rabat, Casablanca, Kénitra and Marrakech) and European (Orsay, Cergy-Pontoise, Montpellier, Savoie, Toulouse, ENS-Paris, African Museum of Belgium, Neuchâtel, Granada) universities and the industry (ONHYM and OMV), comprises 10 chapters dedicated to the main geodynamic episodes which controlled the geology of the country, from the Precambrian and Paleozoic orogens to the Mesozoic rifting and subsequent episodes of basin inversion. We are grateful to the editors, André Michard and Dominique Frizon de Lamotte on one hand, and Omar Saddiqi and Ahmed Chalouan on the other hand, who are respectively two leading French and Moroccan experts of the geology of this part of the World, for having stimulated the contributions of such wide range of expertise, and to provide the international community with a very high standard, easy to read volume.

Sierd Cloetingh[*], *François Roure*[**] *and Magdalena Scheck-Wenderoth*[**]
[*]ILP President.- [**]ILP Task Force VI on Sedimentary Basins

Preface from the Société Géologique de France

Morocco is the key point between one passive margin (Atlantic Ocean), one ocean acting to closure (Mediterranean) and an old craton (Africa). Showing Precambrian and Palaeozoic orogenies up to Mesozoic rifting and Cenozoic orogeny, Morocco is a fantastic place where geodynamic interactions between mantle and lithosphere are illustrated.

In this remarkable book which perfectly resumes the last discoveries and state of knowledge of Morocco geology, André Michard, Omar Saddiqi, Ahmed Chalouan and Dominique Frizon de Lamotte, together with a wide panel of specialists, are offering us the "field verity" that is to say, the results of observation and cartography synthesis. These discoveries and state of knowledge have been exposed in the October 2007 International Congress of Marrakech organized by MAPG and ILP, and followed by seven hundreds people coming from either academic or professional world. The authors and contributors of the present book were the leaders of most of the field trips related to the congress.

Since the time of the pioneers of the geological discovery of Morocco (A. Brives, L. Gentil, L. Neltner and others), the Société géologique de France has published hundreds of papers and a number of memoirs concerning the geology, paleontology or mineralogy of Morocco, and its Bulletin remains one of the traditional tribunes for Moroccan publications. Let us also recall the "Réunion Extraordinaire" of the SGF at Rabat, 1973, dedicated to the Gibraltar Arc.

Today, it is a great honour for the Société géologique de France to support the publication and diffusion of such a very high standard volume.

Christian Ravenne and André Schaaf
President and First Vice-President of the Société Géologique de France

Foreword and Acknowledgements

"The scholar is not a foreigner in any foreign country [...].
Culture, wherever we go, accompanies us, guides us, and allows us to arrive safe and sound."
Bartolomeo Platina, *De falso et vero bono (1460)*

The idea of a new overview of Moroccan geology originated from an early project in November 2005 of a guide book related to the educational program of the Ecole Normale Supérieure (ENS, Paris) and the University of Cergy-Pontoise. During 2006, we received substantial moral and logistic support from the Moroccan Association of Petroleum Geologists (MAPG), and by the Office National des Hydrocarbures et des Mines (ONHYM), in relation to the preparation for the 1st MAPG International Convention and Exhibition that was held in Marrakech, Oct. 2007.

While preparing the MAPG field trip program, we decided to undertake a survey of the geological evolution of Morocco to be published as an independent work, in addition to the field trip guide books. The Société géologique de France (SGF) immediately agreed to consider some sort of co-edition of a future "Geology of Morocco", and Michel Faure, Editor-in-Chief of the Bulletin of the SGF organized a peer reviewing process of the successive chapters (Dec. 2006). An early draft of the book, now entitled: *Continental Evolution:The Geology of Morocco*, was ready for submission to Springer Verlag in July 2007. The last months of 2007 were devoted to improve the text and illustrations based on the critical reviews received from the numerous colleagues and specialists that we had solicited.

Producing this book was a team effort involving many people. We are all indebted to our predecessors who investigated Moroccan geology for nearly a century. We are particularly indebted to those geoscientists, from Morocco and elsewhere, who, during the last three decades, brought our understanding of Morocco up to modern, international standards.

We want to thank our colleagues from various universities and institutions who shared with us their knowledge of the subject and/or of the English language by reviewing our chapters: S. Bogdanoff, R. Bortz, J.-P. Bouillin, J. Destombes, M. Durand-Delga, M. Faure, R. Maury, F. Medina, T. Pharaoh, B. Purser, F. Roure, S. Samson, J.-P. Schaer, J. F. Simancas, J. I. Soto, A. Tahiri, Tan Say Biow, J.-C. Trichet.

Warm thanks are also due to the colleagues who allowed us to use the original files of many valuable illustrations: D. Aslanian, A. Baltzer, M. Ben Abbou, late M. Burkhard, I. Contrucci, J. Destombes, H. Echarfaoui, E.H.El Arabi, H. El Hadi, A. El Harfi, J. Fabre, J. Fullea Urchulutegui, R. Guiraud, M.-A. Gutscher, H. Haddoumi, M. R. Houari, M. Ihmadi, H. Jabour, K.B. Knight, P. Krzywiec, E. Laville, N. Loget, A. Maillard, H. Ouanaimi, G. Rosenbaum, D. Rumsey, M. Sahabi, F. Sani, A. Schettino, J. F. Simancas, W. Spakman, G. Stampfli, D. Stich, A. Teixell, E. Teson. We acknowledge the authorizations by the corresponding Publishers to use freely these illustrations.

We acknowledge the moral and material support received from our academic institutions, in particular the ENS (Paris), and the Universities of Casablanca-Aïn Chock, Cergy-Pontoise, and Rabat; from the CNRS-INSU ("Relief de la Terre" Program) and the French-Moroccan "Action Intégrée" Volubilis; from the MAPG and ONHYM in Morocco, particularly to A. Morabet and A. Mouttaqi; and from the SGF in Paris, particularly to F. Rangin, M. Faure, P. de Wever and A. Schaaf. We thank B. El Ilam (ONHYM, Rabat) and A. Paillet (EDYTEM, Chambéry) for their help in the drawing of many figures. Finally, the support of the International Lithosphere Program (ILP) and the Ministry of Energy, Mines, Water Resources and Environment of Morocco (MEM) are also gratefully acknowledged.

Obviously, the present work presents only a small part of what can be seen in the area. We hope that it will provide the reader with an updated insight of the geology of this country, and hopefully stimulate new studies in the very near future.

A. Michard, O. Saddiqi, A. Chalouan, and D. Frizon de Lamotte

Contents

The Editors

André Michard is emeritus professor from the University of Paris-Sud (Orsay) and Ecole Normale Supérieure. Agrégé in Natural Sciences, he received his “Doctorat d’Etat” in Earth Sciences in 1966 at Paris-Sorbonne in 1966. After a two-year stay as a professor at Mohamed V University, Rabat, he joined the Louis-Pasteur University, Strasbourg, and then Paris-Sud University in 1986. He published a first overview of Moroccan Geology in 1976 (3d edition in 2001). His research focussed on the tectonic and metamorphic evolution of mountain belts, in the Western Alps, Oman, Cuba, southeast Turkey, northern Greece, and particularly in the Hercynian and Alpine Belts of Morocco. He shared with Ahmed the responsibility of a number of French-Moroccan *Action Intégrée* programs from 1986 to 1998. André collaborated with the Moroccan Ministry of Energy and Mines in several mapping projects since 1982.

Omar Saddiqi is Professor at the Faculty of Sciences, Casablanca-Aïn Chock University. He obtained his PhD at Louis-Pasteur University, Strasbourg, in 1988. After a two years stay as a lecturer in this University, he joined the Hassan-II University of Casablanca-Aïn Chock, where he received a “Doctorat d’Etat” in 1995. His research has concerned the tectonic evolution of the Oman Mountains and Gibraltar Arc (metamorphic structure, paleomagnetic rotations and fission track analysis). During the last decade, he set up a Laboratory of Thermochronology in his University, focussing on the vertical movements and relief evolution of Morocco through time. This activity has been integrated in international research programs, involving particularly the Ecole Normale Supérieure, Paris, and the Universities of

Cergy-Pontoise, Paris-Sud (Orsay), Grenoble, Potsdam (Germany) and Neuchâtel (Switzerland). He collaborates with the Moroccan Ministry of Energy and Mines in the National Program of Geological Mapping (PNCG).

Ahmed Chalouan is professor at the Faculty of Sciences of the Mohammed V-Agdal University, Rabat. He received his "Doctorat de 3è Cycle" and his "Doctorat d'Etat" at the Louis-Pasteur University, Strasbourg, in 1977 and 1986, respectively. He is in charge of the Laboratory of Structural Geology of his Faculty since 1978. For thirty years Ahmed developed with his colleagues and students a number of national and international programs of tectonic investigations in the Rif Belt, in cooperation with the ONAREP/ONHYM (interpretation of seismic profiles) and SNED (detail structure of the Strait area), the Louis-Pasteur University, Strasbourg, Ecole Normale Supérieure, Paris, and Granada University. He collaborates with the Moroccan Ministry of Energy and Mines in the National Program of Geological Mapping (PNCG). In the last decade he focussed on the neotectonic deformation of the belt. Ahmed co-edited the Geol. Soc. London Spec. Publ. 262 (2006), *Tectonics of the Western Mediterranean and North Africa*.

Dominique Frizon de Lamotte is Professor of Earth Sciences at the University of Cergy-Pontoise (France). He received a "Doctorat d'Etat" from the University Pierre-et-Marie-Curie (Paris) in 1985. After a two years (1978–1980) stay as a lecturer at Mohamed-V University, Rabat, he joined the university of Reims (1980–1987) then the university Paris-Sud (Orsay) (1987–1991). During this period, he was involved in the Rif mapping project of the Geological Survey of Morocco. He has been appointed at Cergy-Pontoise University in 1991 and set up the Department of Earth Sciences of this university. He is currently in charge of a research team (belonging to the Laboratoire de Tectonique, CNRS UMR 7072) which develops integrated approaches combining field geology, physics of rocks and modelling to investigate inverted sedimentary basins. His personal research focuses on two main points: (1) kinematics of fold-thrust structures and reservoir evolution in complex tectonic environments and (2) Geodynamics of West Mediterranean and Middle East.

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Chapter 1
An Outline of the Geology of Morocco

A. Michard, D. Frizon de Lamotte, O. Saddiqi and A. Chalouan

Morocco is one of the most fascinating lands in the world for studying geology. It is a friendly country, provided with a good network of roads. Most of Morocco being situated within the Mediterranean to Sub-Saharan climatic zones, with a mean annual precipitation ranging from 300 to 600 mm, it offers wide landscapes, with subsurface rocks exposed in splendid outcrops. Last but not least, Morocco is located at a triple junction (Fig. 1.1) between a continent (Africa), an ocean (the Atlantic) and an active plate collision zone (the Alpine belt system). This results in a rugged topography with a wide range of outcropping terranes spanning from Archean to Cenozoic in age, as well as diverse tectonic systems from sedimentary basins to metamorphic fold belts. Minerals and fossils from Morocco are curated in museums the world over. Finally, it is worth emphasizing that natural resources extracted from the Moroccan subsoil are important for the national economy (phosphate, Ag, Pb, Zn, barite, fluorspar, etc.). At the moment, there is also active offshore exploration for oil and gas.

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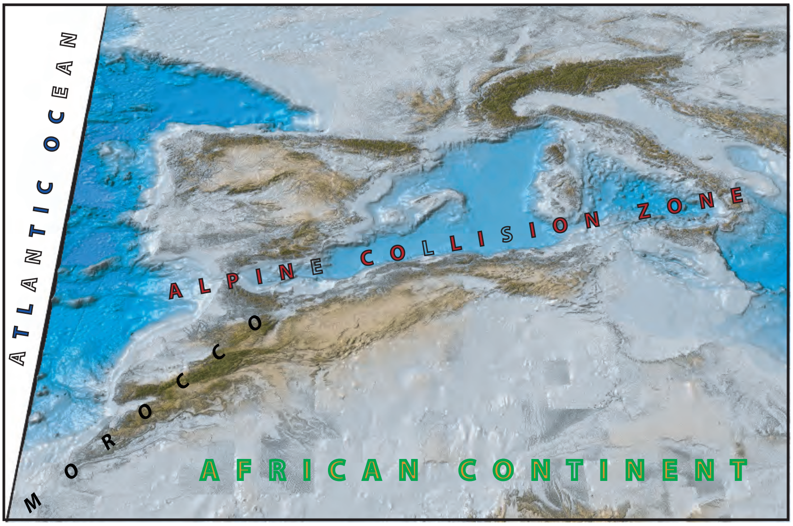

Fig. 1.1 Location of Morocco at the triple junction of the African continent, Atlantic Ocean and Alpine subduction-collision zone. Numerical modelling of elevation and bathymetry data, oblique view of the West Mediterranean area, by courtesy of N. Chamot-Rooke (E.N.S., Paris)

1.1 Topography and Major Geological Domains

At first sight, the topography of Morocco is comparable with that of the central and eastern Maghreb, i.e. Algeria and Tunisia (Fig. 1.2). To the north, the Rif Range extends along the Mediterranean coast (Alboran Sea), maintaining the continuity

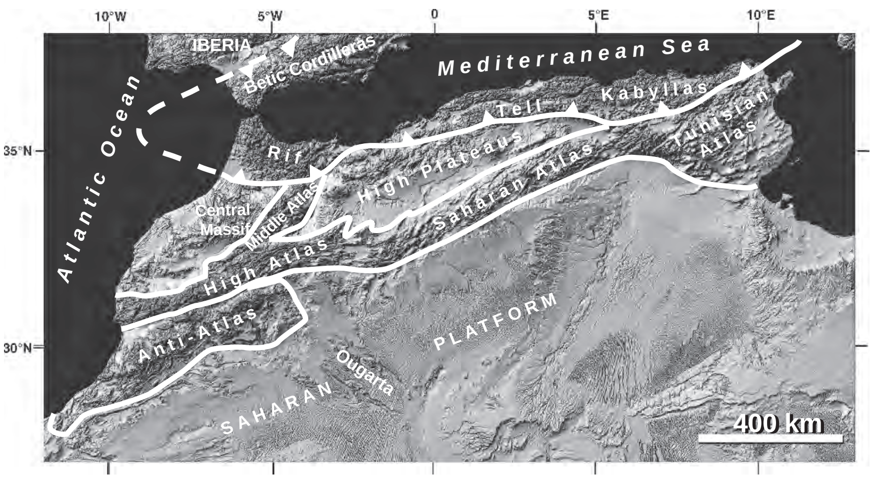

Fig. 1.2 Physiography of northwestern Africa with limits of the main natural regions and young mountain belts. Morocco extends approximately west of the meridian 2°W, and further south of the lower border of the picture (see Fig. 1.2)

of the Kabylian-Tellian belts (Maghrebides) up to the Strait of Gibraltar. South of these coastal ranges, a domain of elevated plateaus or mesetas occur (Algerian High Plateaus and Oran Meseta, Moroccan Meseta), including intramontane basins (Missour and High Moulouya basins). Then the Atlas system rises up, providing a northern boundary to the dominantly low-elevation Saharan domain.

However, Morocco differs from Algeria and Tunisia in several ways. The elevation of most of the country (except the Rif and Atlantic areas) is particularly high (Fig. 1.3). The High Atlas displays several massifs close to 4000 m high, including the highest peak of northern Africa (Jebel Toubkal). A branch of the Atlas system extends obliquely across the mesetan domain, namely the Middle Atlas, which exceeds 3000 m in elevation. The northern, sub-Saharan border of the main Saharan domain also rises and forms a massive mountain belt, the Anti-Atlas, achieving up to 2700 m in Jebel (J.) Saghro and even more in the J. Siroua (Sirwa) recent

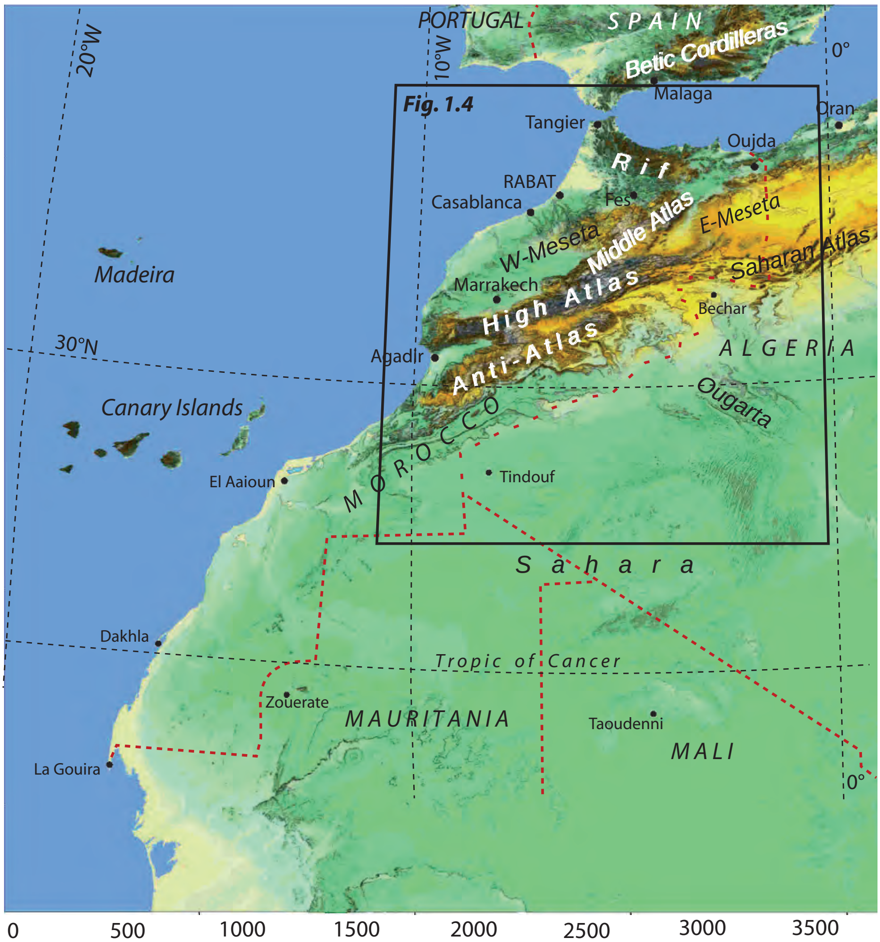

Fig. 1.3 Elevation map of Morocco and neighbouring countries from GTOPO30 database. State borders are indicative

volcano (3300 m). The elevation decreases westward, away from the Middle Atlas mountains to the Central Massif of the Moroccan Meseta, towards the Atlantic coastal basins and finally to the Atlantic abyssal plains. South of the Anti-Atlas and Saghro mountains, in the Saharan "hamadas" (plateaus), elevation decreases both southward, from ca. 1000 m to less than 400 m (Tindouf Basin), and westward to less than 200 m, close to the Atlantic (Tarfaya Basin). Neogene basins are shown along the High Atlas borders (Haouz-Tadla and Bahira Basins to the north, Souss and Ouarzazate Basins to the south) or north and east of the Middle Atlas (Guercif and Missour Basins), whereas a large foredeep basin (Gharb) extends southwest of the Rif belt.

In contrast with Algeria and Tunisia, the continental basement of North Africa is more uplifted in Morocco than in the countries further east, causing Paleozoic and Precambrian rocks to outcrop extensively (Fig. 1.4). Paleozoic rocks form large

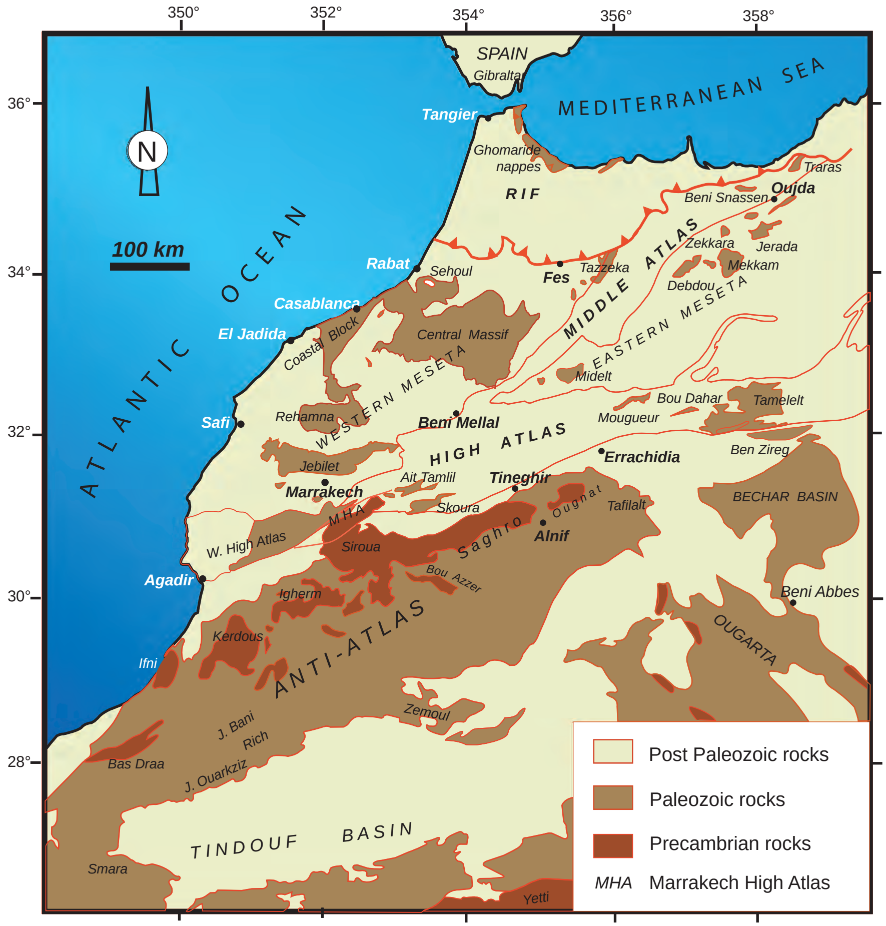

Fig. 1.4 Extension of the Paleozoic and Precambrian oucrops in Morocco (Northern Provinces) and westernmost Algeria (Traras, Ben Zireg, Béchar), modified from Piqué and Michard (1989)

culminations within the Mesozoic Atlas domain, whereas Precambrian rocks form similar culminations ("boutonnières") in the middle of the Anti-Atlas Paleozoic terranes. Paleozoic units also occur in the Maghrebide internal zones, similarly developed in Morocco (Alboran domain) and Algeria (Kabylias), but they belong to a disrupted allochthonous terrane ("AlKaPeCa") and not to the African basement itself.

1.2 Mesozoic-Cenozoic Plate Tectonic Setting

References: The most general works concerning the Alpine Tethys/Central Atlantic plate tectonics are those by Dercourt et al. (1993), Rosenbaum et al. (2002), and Stampfli & Borel (2002). More specific references for Morocco are Piqué & Laville (1995), Frizon de Lamotte et al. (2000), Le Roy & Piqué (2001), Michard et al. (2002), Olsen et al. (2003), Sahabi et al. (2004), Knight et al. (2004), Marzoli et al. (2004), Verati et al. (2007), with references therein.

All the geographic and geological peculiarities of Morocco mentioned above are related to its particular location at the northwestern corner of Africa during the ultimate Wilsonian cycle of plate tectonics. This Mesozoic-Cenozoic cycle begins with the Pangea break-up, goes on with the opening of the Central Atlantic and Alpine Tethys Oceans, and ends with the Tethys closure and Alpine belt formation (Fig. 1.5).

The rifting stage is initially recorded during the Middle-Late Triassic by the development of evaporite basins on both sides of the Central Atlantic rift, as well as in the Atlas and external Maghrebides areas (Tethyan basins). The climax of the rifting stage is marked by a short-lived, but voluminous basaltic magmatism, which defines the Central Atlantic Magmatic Province (CAMP), and probably caused the major climatic and biologic crisis of the Triassic-Jurassic boundary. Magmatism was particularly developed on the African side of the rift where it is recorded by dikes, sills and lava flows, mostly dated at 200 ± 1 Ma (Figs. 1.6A–B, 1.7). Rifting was asymmetric, with a SW-dipping main detachment (Fig. 1.6C) that caused a strong uplift of the Moroccan shoulder of the rift. This setting resulted in the extreme reduction of the Mesozoic cover in west Morocco, except in the Atlantic margin itself and Atlas rift basins, and the extensive exposures of Paleozoic and even Precambrian rocks (Fig. 1.4).

The relative position of Africa with North America by the end of the rifting process – with Morocco opposite Nova Scotia, and the age of the earliest spreading stage have recently been better constrained by fitting the east American and west African margin anomalies (Fig. 1.8), and by dating these oceanic magnetic anomalies to the end of the evaporite deposition (Sinemurian, 195–190 Ma). After this earliest, post break-away stage, spreading developed during the Jurassic in the Central Atlantic and Ligurian (Alpine) Tethys – both connected through a transform fault system north of Morocco (Fig. 1.9A). During the Cretaceous, the onset of spreading in North Atlantic resulted in Iberia's eastward displacement, and eventually brought about its anticlockwise rotation (Fig. 1.9B). At that time

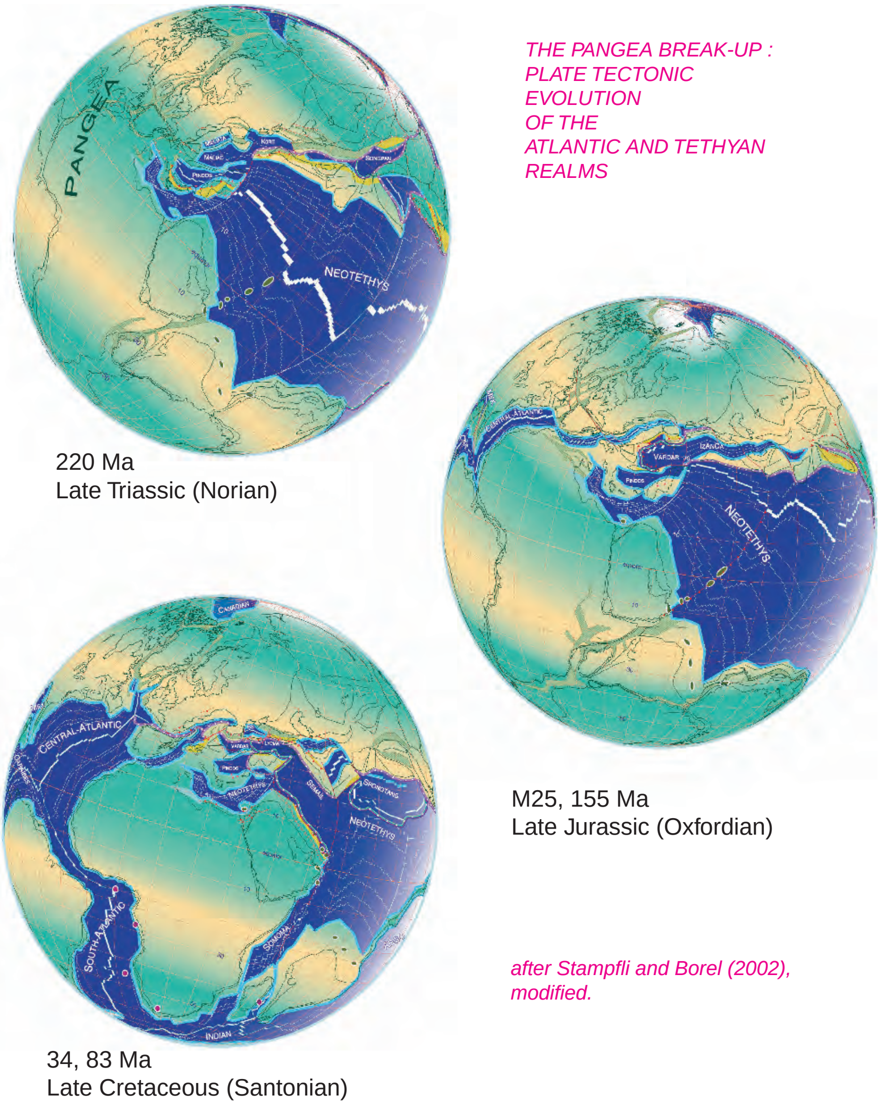

Fig. 1.5 The break-up of the Pangea super-continent and further plate tectonics of the Atlantic and Tethyan domains, after Stampfli & Borel (2002), modified. M25, 34: oceanic anomalies. Rifts are shown as yellow-greenish strips; passive margins are underlined in light blue, active margins in red. Seamounts are shown as large red spots

(Late Cretaceous), due to the opening of the South Atlantic, Africa ceased to translate parallel to southern Eurasia and began to converge with it. This resulted in the Pyrenean-Alpine and Maghrebide-Atlasic shortening events (Figs. 1.9B–D), i.e. in the closure of the Tethyan basins and correlative building of the mountain belts that characterize Morocco. The latter stage (Fig. 1.9D) is also the onset of the

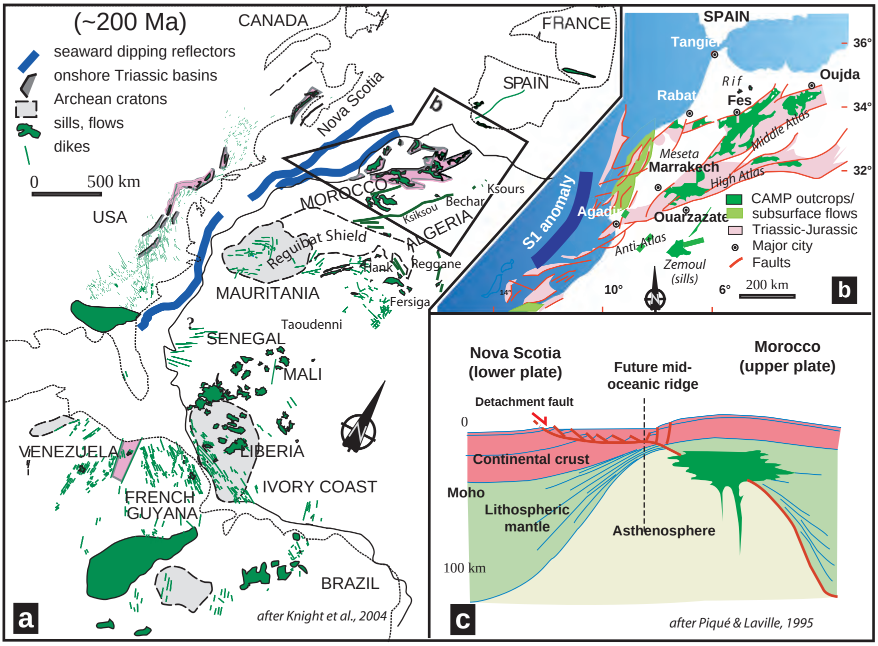

Fig. 1.6 The Atlantic and Neo-Tethyan rifting at the Triassic-Jurassic boundary. **A**: Extension of the Central Atlantic Magmatic Province (CAMP) across four Gondwanan continents (A), and zoom on the Moroccan area (B), after Knight et al. (2004), modified after Chabou et al. (2007) for SW Algeria. **C**: Asymmetric rifting resulted in strong uplift of the rift shoulder on the African side; schematic restoration after Piqué & Laville (1995), modified

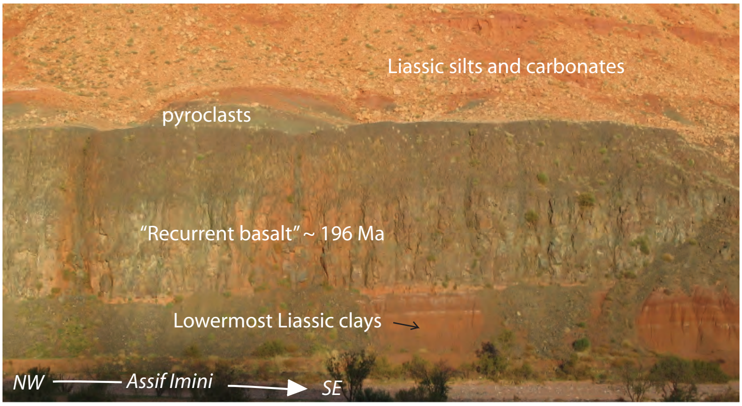

Fig. 1.7 Basaltic trapp on top of (Rhaetian)-Hettangian red clays south of the Marrakech High Atlas. Note the rough prismatic structure within the flow, and the thin pyroclastic level on its top. This is the latest basalt flow ("recurrent basalt") of the High Atlas rift, dated at ~196 Ma (Verati et al., 2007). The pinky deposits above are poorly dated Liassic dolomitic limestones and sandy marls of the post-rift sequence. Tiourjdal, Assif Imini valley, SE of Agouim, 70 km west of Ouarzazate

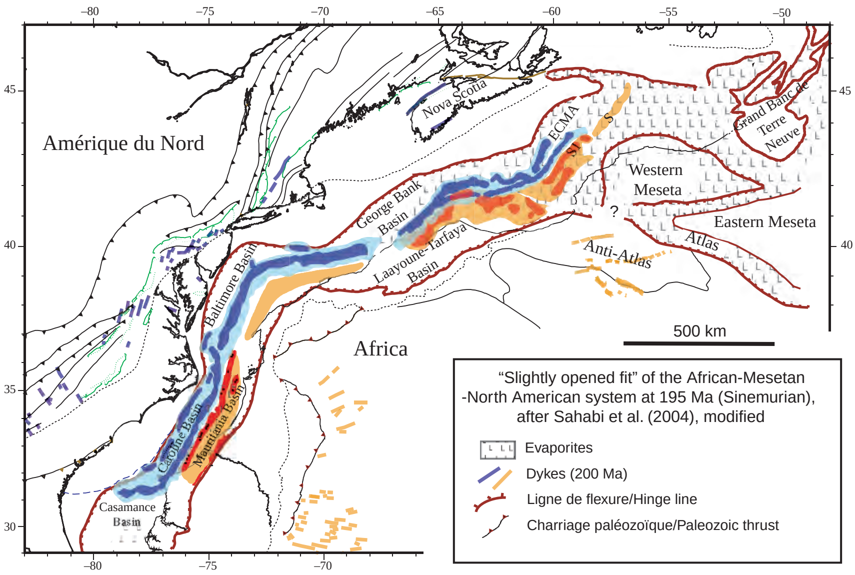

Fig. 1.8 Slightly opened fit of the Africa-Moroccan Meseta-America system at 195 Ma (Sinemurian), after Sahabi et al. (2004), modified. American structures in blue, African ones in red. Large fitted stripes: magnetic anomalies (MA), ECMA (East Coast MA) and S1-S' (West African Coast MA). The evaporite basins on both sides of these anomalies and in the Atlas domain are limited by the hinge lines. Also shown: Triassic-Jurassic dikes (thick bars) and Paleozoic structures (black lines with teeth). The Moroccan Meseta is slightly disconnected from the West African Craton in order to take into account the Atlasic shortening

Ligurian ocean subduction whose roll-back controlled the opening of the Neogene west Mediterranean basins. As Morocco was pinned against Iberia throughout the Late Cretaceous-Cenozoic (whereas wide oceanic areas still occurred to the east), convergence occurred later and at a slower rate in the Moroccan transect than in the Lybian and Egyptian ones (Fig. 1.10). This must be kept in mind to understand the lateral changes in the Alpine belt system from east (Hellenides, Alps) to west (Rif-Betic or Gibraltar Arc), and in particular, the scarcity of oceanic crust remnants in the latter orogen.

Thus, the particular position of Morocco at the NW corner of Africa accounts for the striking differences between its Mediterranean and Atlantic margins. The latter continued up to present being a passive margin. Subsidence allowed sediments to accumulate from the Triassic onward along this margin in the Moroccan Coastal Basins, which were only partially deformed and emergent during the Neogene. The high level seas of the Late Cretaceous-Eocene flooded the majority of Morocco, giving birth to the marly limestones of the Saharan hamadas and eastern Atlas domain, and more interestingly, to the phosphate deposits next to the continental slope ("Plateau des Phosphates" of the western Meseta, Tarfaya Basin). In contrast, the Tethyan margin was associated to a transform system during the Mesozoic, and changed into an oblique subduction-collision margin during the Eocene. The present

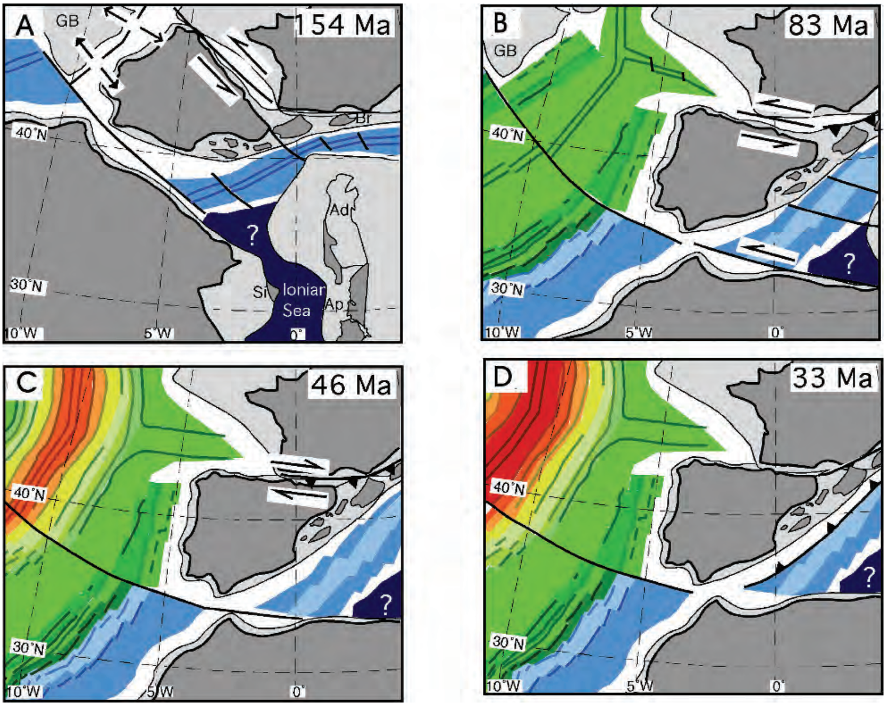

Fig. 1.9 Restoration of the evolution of Central-Northern Atlantic and western Tethys regions, after Rosenbaum et al. (2002), modified. **A**: Late Jurassic; **B**: Late Cretaceous; **C**: Middle Eocene; **D**: Early Oligocene. Ad: Adria; Ap: Apulia; Br: Briançonnais; GB: Grand Banks; Si: Sicily

Mediterranean margin formed during the Neogene, and displays the characteristics of an active margin.

1.3 Continental Growth: The Successive Fold Belts

References: Fundamentals concerning this section are available in Michard (1976), whereas some of the recent references used hereafter are as follows: (i) concerning the Archean, Eburnian and Pan-African belts, Bertrand & de Sá (1990), Black et al. (1994), Villeneuve and Cornée (1994), Dalziel (1997), Ennih and Liégeois (2001, 2003, 2008), Cordani et al. (2003), Fabre (2005), Gasquet et al. (2005), Liégeois et al. (2005), Scholfield et al. (2006), Scholfield & Gillespie (2007); (ii) concerning the Phanerozoic belts, Frizon de Lamotte et al. (2004), and Simancas et al. (2005).

Looking at the overall structure of Morocco in NW Africa (Fig. 1.11) and at the main events of its geological evolution allows us to approach and discuss the concept of continental growth. Morocco displays large parts of the orogenic systems or fold belts that developed around the 2 Ga-old continental nucleus of north-western Africa, i.e. the West African Craton (WAC). These successive

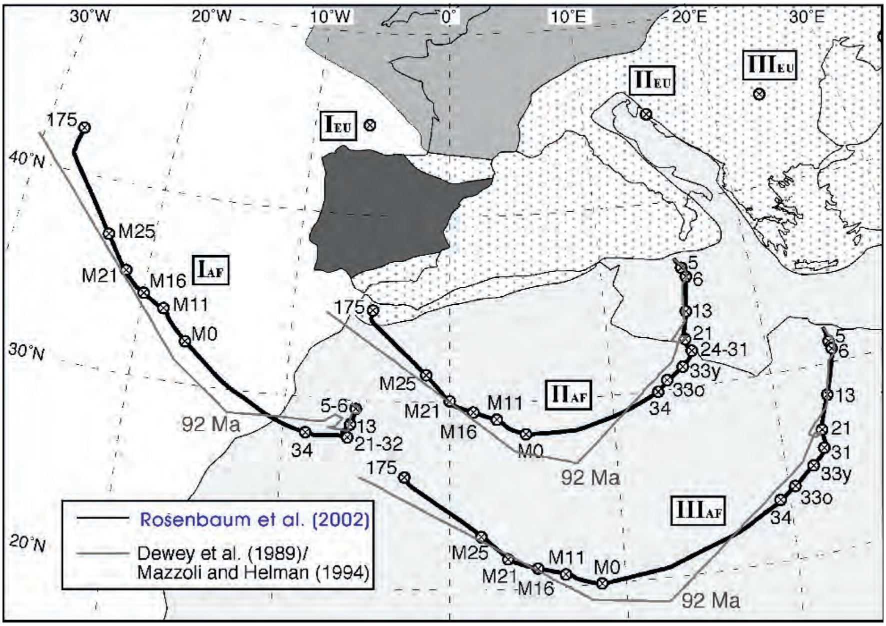

Fig. 1.10 Trajectories of three points of northern Africa relative to fixed points of Europe, plotted as a function of time since 175 Ma, and comparison with previous published trajectories, after Rosenbaum et al. (2002), modified. Each step along the trajectories corresponds to an oceanic magnetic anomaly. The most important dates are M25: 154 Ma; M0: 120.2 Ma; 34: 83 Ma; 31: 67.7 Ma; 21: 46 Ma; 13: 33.1 Ma; 6: 19.2 Ma; 5: 9.9 Ma. Stippled areas are regions of strong Mesozoic-Cenozoic deformation

belts are the Pan-African (Cadomian, Avalonian), Caledonian-Variscan (Hercynian, Alleghanian) and Alpine belts (Fig. 1.11), which are briefly presented hereafter and described in details in Chaps. 2–7. In contrast, the much older belts which constitute the basement of the WAC are presented only (and shortly) in the following section.

1.3.1 The Archean and Eburnian Terranes of the West African Craton (WAC)

1.3.1.1 Definition of the WAC

The WAC extends over millions of square kilometres in the Sahara desert (Figs. 1.11, 1.12). The crystalline basement crops out in the Reguibat Shield or Arch whereas it is hidden beneath thick piles of undeformed sediments in the Tindouf, Reggane and Taoudenni Basins. The age of these sedimentary deposits spans the whole Neoproterozoic-Cenozoic times (about 1 Ga) without any internal unconformity,

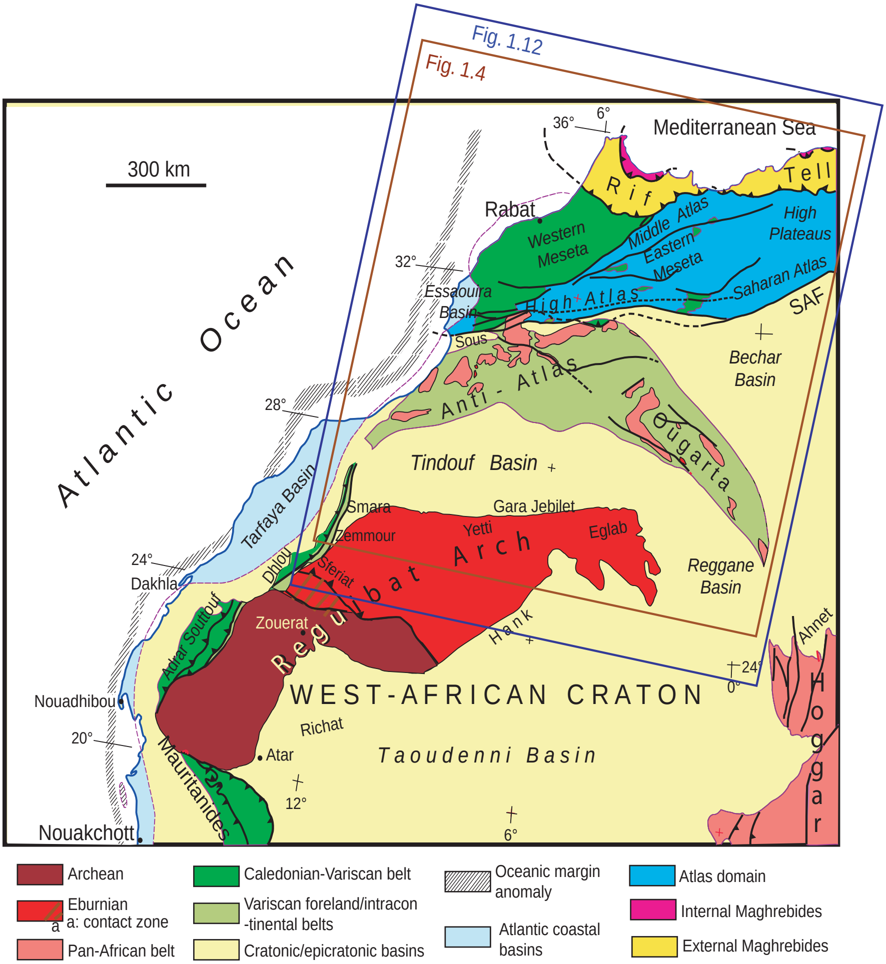

Fig. 1.11 Tectonic map of north-westernmost Africa showing the northern part of the West African Craton (WAC) and the adjoining fold belts, with location of Figs. 1.12 and 1.14 (framed). After Fabre (1976, 2005), Villeneuve & Cornée (1994), Ennih & Liégeois (2001), Villeneuve et al. (2006), Schofield et al. (2006), Schofield & Gillespie (2007). Location of the oceanic margin anomaly after Sahabi et al. (2004). SAF: South Atlas fault

which defines a typical cratonic area. The Neoproterozoic onlap is exposed on the southern border of the Reguibat Shield and in the Zemmour region, consisting of quartzites and stromatolitic limestones 1000–700 Ma in age. By contrast, at the southern border of the Tindouf basin, the base of the sedimentary succession is made up of Upper Ordovician sandstones similar to the Tassilis sandstones north of the Hoggar (or Tuareg) Shield. The thickness of the sedimentary pile in these intracratonic basins reaches 8 km at Taoudenni and up to 10 km at Tindouf.

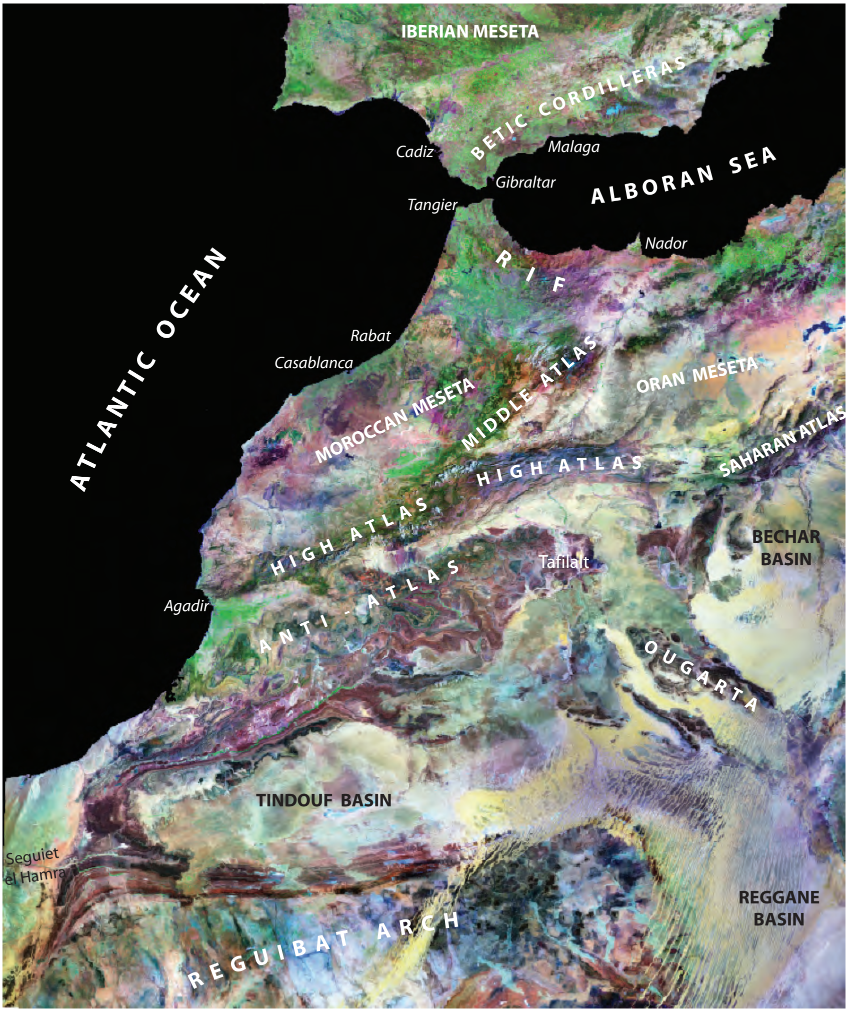

Fig. 1.12 Landsat image of Morocco (see Fig. 1.11 for location). See Fig. 1.4 for the interpretation of the different regions. More detailed Landsat images are also shown in the following chapters

1.3.1.2 The Reguibat Arch

The cratonic basement exposed in the Reguibat Arch (or Reguibat Shield) comprises two contrasting crustal domains, i.e. a western Archean terrane and an eastern Eburnian (or Eburnean) terrane (Fig. 1.11).

The *Archean terrane* is dominated by gneisses and granitic rocks with scattered lenses of metagabbros and serpentinites, and supracrustal rocks including laminated, ferruginous quartzites, felsic gneisses interlayered with pyroxene-rich gneisses, cherts and impure marbles. Younger ages correspond to 3.04–2.83 Ga and were obtained from intrusive granitoids (Lahondère et al., 2003), suggesting a Mesoarchean age of continental crust formation.

In contrast, the *Eburnian terrane* of the eastern Reguibat Shield is largely made up of Paleoproterozoic granitic and metasedimentary rocks such as mylonitic paragneisses with calcsilicate nodules, abundant amphibolites and cherts (bimodal metavolcanic rocks), and ridges of ferruginous quartzites and marbles. The protoliths of the supracrustal rocks indicate a shallow marine basin dominated by clastic sediments, with local development of calcareous nodules and limestones (Schofield et al., 2006). The banded iron-rich formations, widely exploited in Mauritania (Zouerat area), are thought to represent chemical sediments precipitated in the presence of high concentrations of exhalative iron in relation with the activity of cyanobacteriae in an aerobic setting. Together with the associated limestones, they record the increase of the biological productivity during the Paleoproterozoic.

The granitoids plutons which intrude the metamorphic supracrustal rocks have been interpreted as formed during a cycle of subduction and subsequent accretion onto the adjacent Archean continental margin (Lahondère et al., 2006). In the Sfariat region, the Paleoproterozoic continental margin succession has been intruded with synorogenic granitoids and transported SW onto the Archean foreland during sinistral oblique collision (Schofield et al., 2006; Schofield & Gillespie, 2007). U-Pb geochronology reveals that anatexis and sinistral transpression took place between 2.12 and 2.06 Ga. Timing and kinematics of the Eburnian Orogeny in this region are similar to those for the Man Shield in equatorial West Africa.

1.3.1.3 Pre-Gondwana Supercontinents (a reminder)

The Eburnian Orogeny is thought to represent a major tectonic pulse of crustal growth (Bertrand & de Sá, 1990) related to the assembly of a pre-Rodinia, Paleo- to Mesoproterozoic supercontinent termed *Columbia* (Rodgers & Santosh, 2002). However, it must be emphasized that Mesoproterozoic terranes comparable to the 1 Ga-old Grenville terrane of the Laurentian Shield (and also present in the Avalonia basement) are not known in NW Africa.

As for *Rodinia*, it was initially defined as a long-lived supercontinent that assembled all the continental fragments around Laurentia and remained stable from 1000 up to 750 Ma (Dalziel, 1997; Weil et al., 1998, Meert & Lieberman, 2007, with references therein). Nonetheless, recent work has cast doubt on the Rodinia palaeogeography and even on the timing of its assembly and break-up. According to Cordani et al. (2003), a Braziliano Ocean separated most of the South American and African cratons from the Laurentia-Amazonia-West Africa margin. This ocean was closed between 940 and 630 Ma along the Pampean–Paraguay–Araguaia–Pharusian (Pan-African) mobile belts. Moreover, accretion along the South American and African platforms was a diachronous and long-lived process that involved several intra-oceanic and continental magmatic arcs and microcontinents. This evolution started at around 1000 Ma and ended at around 520 Ma with the final assembly of *Gondwana* (Meert et al., 2007, with ref. therein) and, according to some authors (Dalziel, 1997), to *Pannotia*, i.e. the assembly of Gondwana, Laurentia, Baltica and Siberia.

1.3.2 The Pan-African Belt: Gondwana Assemblage

The WAC is surrounded by the Neoproterozoic Pan-African belt that first formed between ~750 and 660 Ma, then between 630 and 560 Ma (Chap. 2). The Pan-African events s.l. are responsible for the building of the supercontinent Gondwana. On the eastern side of the WAC, the Pan-African orogeny is clearly related to the convergence between the East Sahara metacraton and the WAC itself, with building of the Trans-Saharan belt including the Hoggar (Tuareg) Shield. The western part of the Hoggar was essentially built by the Pan-African Orogeny, whereas its central and eastern metacratonic parts were variably reactivated during the Neoproterozoic. The Hoggar massif includes metamorphic terranes and ophiolitic units thrust both eastward and westward. Ultra-high-pressure metamorphism (with coesite relics) has been observed in the Pan-African nappes from Gourma to Mali. In contrast, the Pan-African belt is poorly exposed in the Moroccan Anti-Atlas domain, and displays only low-pressure metamorphic units. It is currently believed that subduction-collision tectonics also occurred along the northern boundary of the WAC, but the Pan-African mobile belts of this region would be now situated elsewhere in the exotic peri-Gondwanan (Avalon) terranes. The unconformities which can be seen beneath the Anti-Atlas Cambrian formations record the end of the Pan-African geodynamic evolution (Fig. 1.13).

1.3.3 The Phanerozoic Belts: from Gondwana to Africa

The *Caledonian-Variscan belt* extends essentially along the northwestern side of Africa, and results from the collision of Gondwana with Laurentia, Baltica and intervening terranes (Avalonia, Armorica etc.) previously detached from the Gondwana continent. The belt developed through repeated oceanic closures and collisional events from about 460–420 Ma (Caledonian, Taconic or Sardic events) to 360–300 Ma (Variscan = Hercynian or Alleghanian events). This is a major orogen associated with several ophiolitic sutures and high- to ultra-high-pressure metamorphism. Only the southern external parts of the belt, devoid of ophiolite and high-pressure metamorphism, are preserved in northern Morocco (Meseta Block, including most of the Atlas basement). East-verging thrusts of "reworked" Precambrian material emplaced upon the western margin of the Reguibat Shield (Adrar Souttouf-Dhlou = Ouled Dhlim) during the Carboniferous, in the northern continuation of the Mauritanide belt. Late Paleozoic deformation remains moderate in the Moroccan Anti-Atlas and Algerian

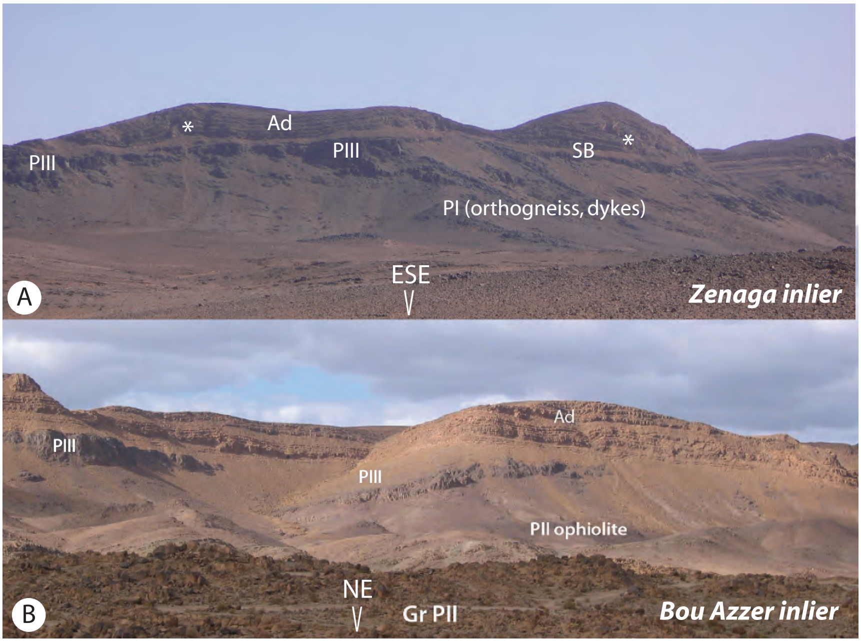

Fig. 1.13 Unconformities on top of the eroded Pan-African belt in the Central Anti-Atlas. **A**: Unconformable Ediacaran volcanites ("PIII") and late Ediacaran conglomerates ("Série de base" SB) on top of the Eburnian granite and pegmatites (Gr PI, c. 2000 Ma) in the eastern Zenaga "boutonnière". The Ediacaran-Cambrian "Adoudounian" dolomites (Ad) are virtually conformable on top of the PIII-SB formations. Asterisks (*): hinges of NE-verging Variscan folds in the detached Adoudounian layers. View from the Tazenakht-Bou Azzer road, 6 km west of Tazenakht (see Chap. 2, Fig. 2.5). – **B**: At the northwest border of the Bou Azzer inlier, the Adoudounian dolomites (Ad) unconformably overlie the tilted Ediacaran volcano-sedimentary formations (PIII) and the underlying Neoproterozoic rocks of the Pan-African belt (PII serpentinite and intrusive granitoid Gr PII). View from the same road as (A), about 20 km further east

Ougarta – both mountains being moulded around the WAC border. In contrast, the Atlas-Meseta domain is strongly deformed, more or less metamorphic, and intruded by varied granite massifs. The Late Permian or, more generally, Triassic unconformity marks the end of the Variscan evolution (Fig. 1.14).

Finally, the youngest orogenic system, i.e. the *Cenozoic Alpine belt*, extends north of the northern boundary of the WAC – being obliquely superimposed onto the Variscan belt and its putative Avalonian-Cadomian basement. The broad organisation of these young mountain belts is directly visible in the topography (Fig. 1.2). The High and Middle Atlas are autochthonous, intracontinental belts, developed by inversion of Triassic-Jurassic aborted rifts. In contrast, the Maghrebides include both parautochthonous, external units originating from the African Mesozoic passive margin, and allochthonous terranes coming from the Maghrebian-Ligurian Tethys and the continental Alboran-Kabylia-Calabria block. In both the Atlas and Maghrebide systems, deformation and mountain building span from the Late Eocene to Present times, i.e. during the last 40 My (e.g. Fig. 1.15).

Fig. 1.14 The Triassic unconformity on top of a Variscan granite from the High Moulouya massif south of Zeida. Transgression of Triassic arkosic sandstones (240–230 Ma?) on top of the 300 Ma-old Boumia granite, emplaced at ca. 10 km depth within the Paleozoic schists of the Eastern Meseta orogen. The post-Triassic oblique fractures are mineralized with barite (cf. Zeida mines)

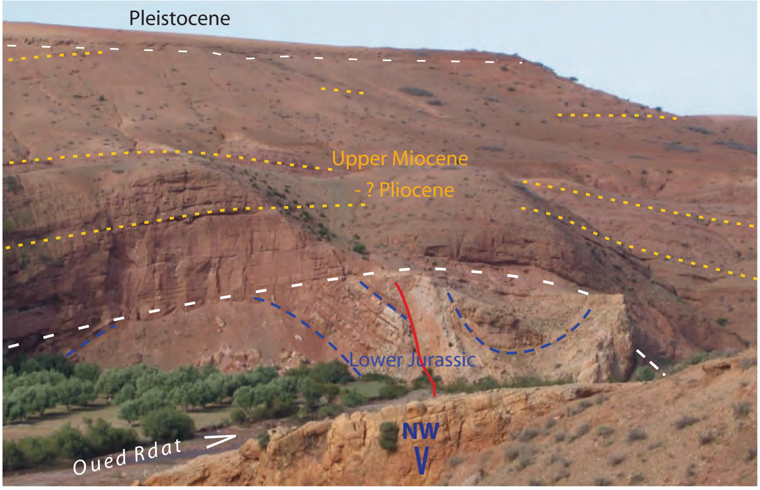

Fig. 1.15 Superimposed unconformities at the northern border of the Marrakech High Atlas, Oued Rdat valley upstream Sidi Rahal (40 km east of Marrakech). Folding of the Lower Jurassic dolomitic limestones initiated during the Late Eocene-Oligocene as the Middle Eocene limestones nearby (Ait Ourir) are also folded. The unconformable Upper Miocene conglomeratic sandstones are deformed into a large anticline which is in turn unconformably overlain by Early Pleistocene torrential conglomerates

1.3.4 *Continental Growth, Continental Reworking, and Structural Inheritance*

Successive orogenic belts, amalgamated against the WAC, constitute a remarkable case study for the concept of continental growth. However, it must be kept in mind that continental breakdown (rifting) also occurs, at least since the Neoproterozoic, prior to each orogenic (convergent) episode. Rifted terranes were drifted away from the Paleo-Gondwana margin, and finally accreted to Laurentia during the Early Paleozoic (Taconic-Caledonian-Acadian events). Part of the continental material was also lost by erosion, and subsequent sedimentation in remote oceanic realms (Variscan and Alpine belts). Moreover, instead of being juxtaposed as crystalline growth zones, the successive orogens are mostly superimposed. As a result, a large part of the rock material of any orogen originated from a previous one. The Pan-African Belt is superimposed onto the border of the WAC and includes rejuvenated schists and granites from the Eburnian Orogen; the Variscan Belt widely extends onto the Pan-African, including Precambrian material in the Anti-Atlas and Mauritanides; and finally, the Atlas and Rif Belts in turn deeply encroach on the Variscan Meseta Domain.

A most important outcome of the addition of superimposed orogens is that old structures (which are mechanical discontinuities) at least partially control the younger ones. A magnificent example of this fact is the *South Atlas Fault*. This Mesozoic-Cenozoic fault system which limits the Atlas Belt to the south (Fig. 1.11) reused a major Variscan fault system, the *Atlas Paleozoic Transform Zone*, which separates the highly-deformed Meseta domain from the mildly deformed Anti-Atlas. Moreover, during the Neoproterozoic this structure apparently corresponded to the northern boundary of the WAC.

1.4 Active Tectonics

References: For both this section and the next one, the sources are essentially Buforn et al. (1995), Morel and Meghraoui (1996), Gutscher et al. (2002), Negredo et al. (2002), Contrucci et al. (2004), Frizon de Lamotte et al. (2004), Spakman & Wortel (2004), Fullea Urchulutegui et al. (2005), Missenard et al. (2006), Sébrier et al. (2006), Stich et al. (2006), and Fernández-Ibáñez et al. (2007) with references therein.

Due to its position at the northwest border of the African plate (Fig. 1.16A), Morocco still exhibits significant tectonic activity, especially in its younger parts. This is clearly shown by the density map of seismicity in the Atlantic-Mediterranean transition zone (Fig. 1.16B).

The most active seismic zone is localized in the Rif and Alboran Sea. This is consistent with the fact that this region corresponds to the contact zone between the two converging plates, Eurasia and Africa. The movement of the African (Nubian)

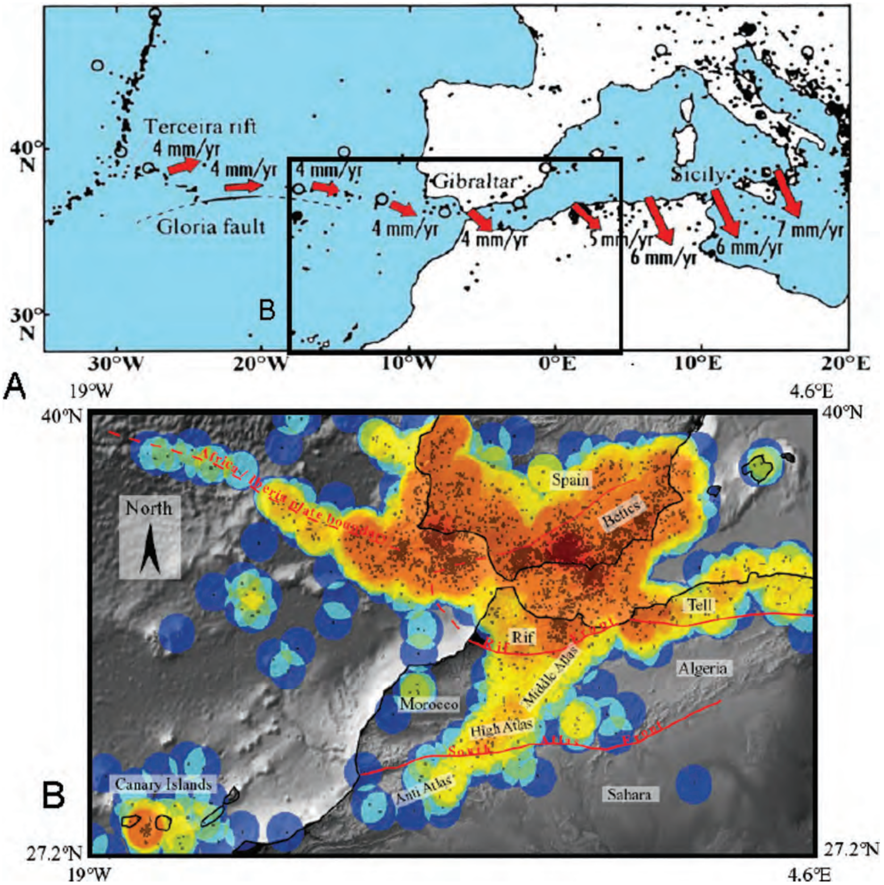

Fig. 1.16 Moroccan seismicity and plate tectonics. **A**: Direction and mean recent rate of the Europe-Africa convergence deduced from global geodetic NUVEL-1 model (DeMets et al., 1994; Morel & Meghraoui, 1996). Black dots: epicentres (1965–1985). The N-S trending seismic lineament on the left is the Mid-Atlantic ridge. The boundary between the Eurasian and African (Nubian) plates trends E-W, being diffuse in the Gibraltar area and Mediterranean Sea. – **B**: Density map of seismicity in the Mediterranean-Atlantic transition zone calculated from ISC database for the 1995–2000 time interval, after Missenard et al. (2006). Grey points indicated epicentres

plate relative to Eurasia which trended N during the Late Cretaceous-Paleogene interval (Fig. 1.10), trends approximately NW, i.e. oblique to the African margin, during the Miocene to Present interval (Fig. 1.16A). The convergence rate decreases from 6 to 4 mm/year from the longitude of Tunisia to that of Morocco due to the actual position of the pole of rotation that imposes an anti-clockwise rotation of Africa with respect to Eurasia.

The vectors of relative displacements for the African and Eurasian plates, deduced from the oceanic geophysical dataset and geodetic observations (e.g., NUVEL-1A model), fit relatively well with more recent GPS observations (Fig. 1.17).

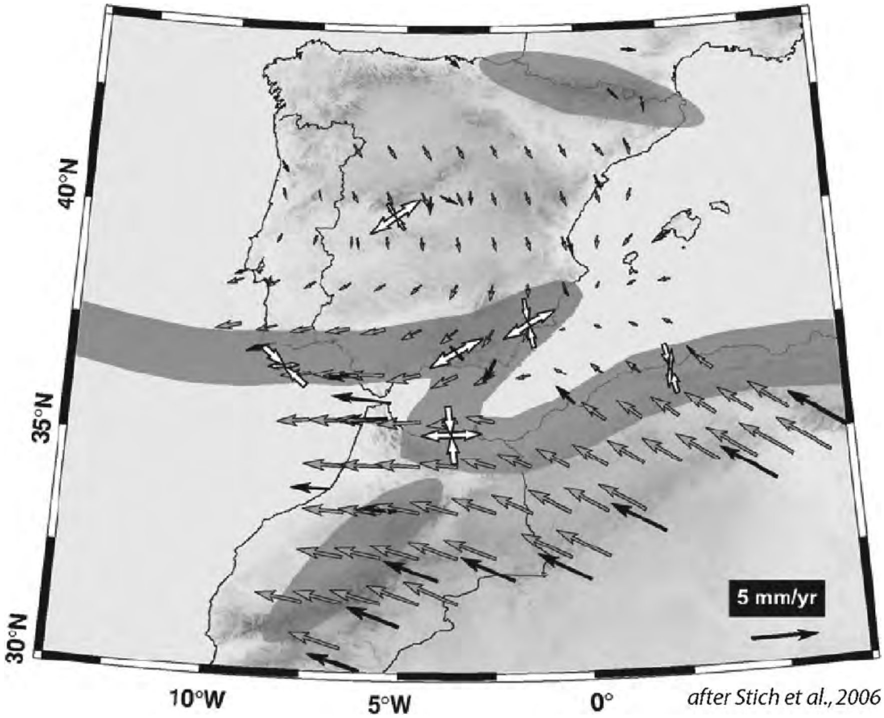

Fig. 1.17 Direction and rate of present-day motion of NW Africa and Spain relative to stable Europe after Global Positioning System (GPS) observations. *Black arrows*: velocities measured in the area or predicted from the dataset. *Grey arrows*: interpolated velocity field. The mean stress tensors deduced from the seismic focal mechanisms (*large white arrows*) are plotted on the Mediterranean seismic belt (*light grey*). After Stich et al. (2006), modified

Discrepancies between the convergence rate and direction predicted among the different global geodetic models occur. Moreover, GPS data also reflect motions by active faulting which induce local motions that deviate from the overall pattern of plate convergence. However, the available data would reflect changes in the last 3 Ma of Africa-Eurasia relative motions, related to the effects of the ongoing collision. Moreover, the overall converging movement is combined with a regional extension in the Alboran area, trending E-W to ENE-WSW, and associated with the building of the Gibraltar Arc since ~20 Ma.

The Maghreb-Alboran-Southern Spain area is characterized by frequent earthquakes, sometimes of great magnitude, resulting in appalling disasters (Al Hoceima, 24 February 2004, M 6.3). Seismic activity may be controlled by the strong spatial gradients in the velocity field. Most epicentres of shallow earthquakes are distributed in a Z-shaped area (Figs. 1.16B, 1.17) that follows the Algerian coast, crosses obliquely the Alboran basin from Eastern Rif to Eastern Betic Cordilleras (Trans-Alboran Seismic Zone, TASZ), and continues further to the west in the Atlantic Ocean. Focal mechanisms indicate a predominance of reverse faulting in the Algerian margin that passes westward to predominant strike-slip mechanisms. Maximum horizontal stresses are consistent with NNW-SSE plate convergence, although it is observed moderate clockwise stress rotation along the TASZ. Within the TASZ relevant seismicity is characterized by predominantly sinistral strike-slip faulting associated with some normal faulting, revealing a left-lateral (transtensive) regime. The compressional deformation in the plate boundary resumes in the west-trending segment of the seismic zone up to the Gorringe Bank west of SW Spain (Fig. 1.3, 348 °E – 37 °N). Intermediate depth earthquakes (60–160 km) also occur beneath southern Spain and in the western Alboran basin, east of the Gibraltar Arc, and some deep events (~600 km) are also reported beneath southern Spain. These features should be related to the east-dipping subduction of a narrow oceanic slab, as discussed in Sect. 1.5 (Sect. 1.5.7).

South of the Mediterranean area, seismicity is scarce and occurs concentrated in a narrow, NE-SW trending zone covering the Middle Atlas, Central High Atlas and part of western Anti-Atlas (Fig. 1.16B). This seismicity mostly corresponds to M < 5 events in the last decades, except the appalling 29 February 1960 Agadir earthquake (M = 5.7), and two events in easternmost Anti-Atlas (M = 5.5). The earthquakes show shallow hypocentres and dominantly strike-slip or reverse mechanisms. It is worth noting that this seismic strip follows the zone where the lithosphere is thin (see below, Fig. 1.18B). Active tectonics along the South Atlas Fault, as evidenced by the deformation of the Quaternary surfaces, suggests the occurrence of an intra-crustal

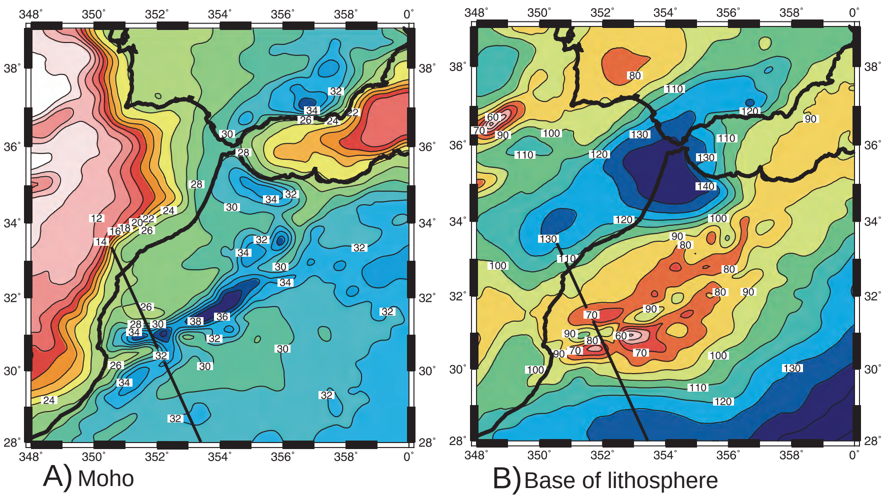

Fig. 1.18 Structure of the lithosphere of the Moroccan-south Iberian region derived from elevation and geoid anomaly modelling, after Fullea Urchulutegui et al. (2006), modified. **A**: Map of the Moho depths. Isolines every 2 km. Note the shallow root beneath High Atlas. **B**: Depth of the lithosphere-asthenosphere boundary. Contour interval is 10 km. Note thin lithosphere (asthenosphere uplift) beneath High and Middle Atlas, thick lithosphere beneath cratonic area to southeast, and decoupling of crustal and mantle lithosphere thicknesses beneath Gibraltar-NW Moroccan margin area. *Bold line*: trace of profile Fig. 1.19

décollement seismically active, which has the potential to create big earthquakes (see Chap 4). Seismicity vanishes further south in the Saharan domain, as well as further east in the Saharan Atlas.

1.5 Lithosphere Structure

References: See section above, particularly Frizon de Lamotte et al. (2004) and Missenard et al. (2006).

The crust beneath inland Morocco is not precisely imaged due to lack of deep seismic profiling. However, modelling of crustal and lithospheric thicknesses was recently performed that integrates elevation, geoid anomalies, surface heat flow, gravity, and seismic data. The resulting map of Moho isobaths (Fig. 1.18A) shows a moderately thick crust underneath the Rif and Betics (~32–34 km), whereas the Alboran Basin in between shows a continental crust (~20–22 km) that thins progressively toward the east, achieving less than 16 km at the transition with the young South-Balearic and Algerian oceanic basins. Inland Morocco, the crust thickness increases to 38 km below the most elevated parts of High Atlas, then decreases southeastward down to the normal thickness of continental crust, 30–32 km.

Remarkably, this High Atlas crustal root is not thick enough to isostatically support such a high topography. Modelling of the lithosphere thickness has led to an explanation of this discrepancy. The map of the base of the lithosphere (Fig. 1.18B) shows a prominent, NE-trending zone of thinned lithosphere, in other words an asthenosphere uplift which reaches ca. 60 km depth underneath the Western High Atlas, central Anti-Atlas and Middle Atlas ranges. Thus, thermal doming accounts for the elevation of the area, in addition to the moderate shortening and crustal thickening. In particular, this explains the high elevation of the Siroua Plateau located in the Anti-Atlas south of the High Atlas between the Sous and Ouarzazate Basins (Fig. 1.19). Moreover, the uplift of hot, asthenospheric mantle accounts for the scatter of Pliocene-Quaternary alkaline volcanoes from J. Siroua to J. Saghro, and to Middle Atlas. This uplift would correspond to the trend of a hot line similar to the Cameroon Hot Line (Deruelle et al., 2007), extending at least from the Canary Islands to southeast Spain and referred to hereafter as the *Morocco Hot Line* (see Chap. 4).

A general description of the lithosphere structure along the southern Iberia-Alboran Sea-Morocco N-S transect has been recently compiled by a large panel of scientists (Frizon de Lamotte et al., 2004). The southern part of this transect crosses the geological domains of Morocco (Fig. 1.20A, B) from the young Alboran Basin and Alpine Rif belt in the north up to the Anti-Atlas in the south. In the latter area, the lithosphere thickness attains ca. 130 km. It thins down to ca. 80 km beneath the Siroua-High Atlas area, in relation with the asthenosphere uplift already discussed (cf. Fig. 1.18B). The lithosphere again thickens beneath the Meseta and Rif domains, before its dramatic thinning beneath the Alboran Basin. These domains of irregular and relatively thin continental lithosphere corresponding to the

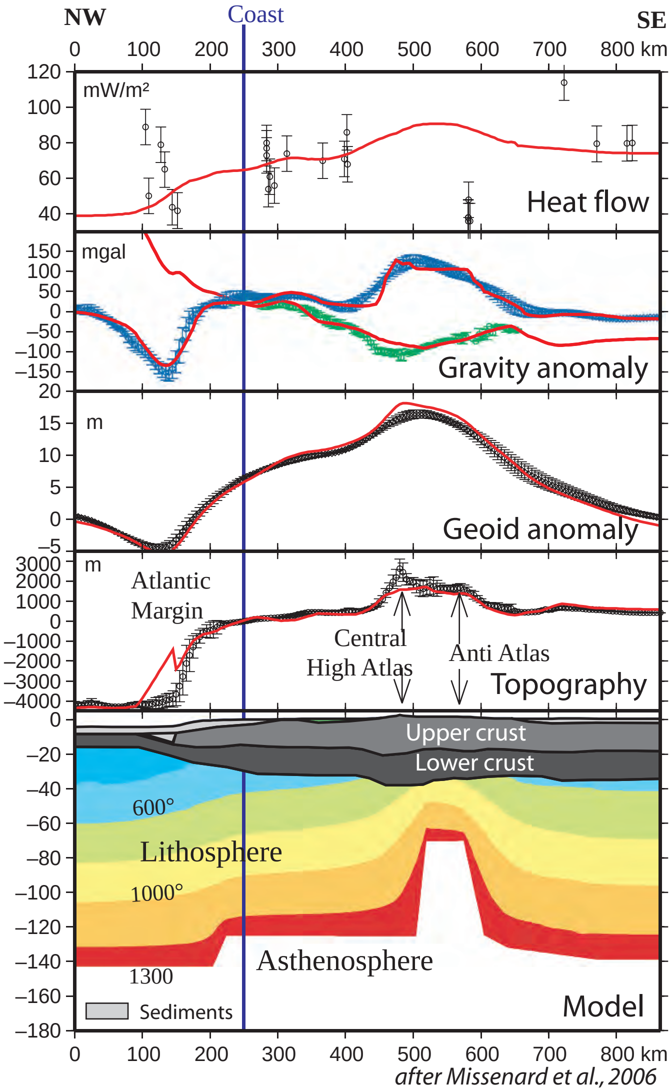

Fig. 1.19 2D lithosphere model across the western High Atlas and Siroua massifs (see Fig. 1.18 for location), with observed (*dots with error bars*) and calculated (*coloured solid lines*) physical properties along the section, after Missenard et al. (2006), modified. The model shows an asthenosphere upwelling up to ca. 60 km depth. Note the strong vertical exaggeration

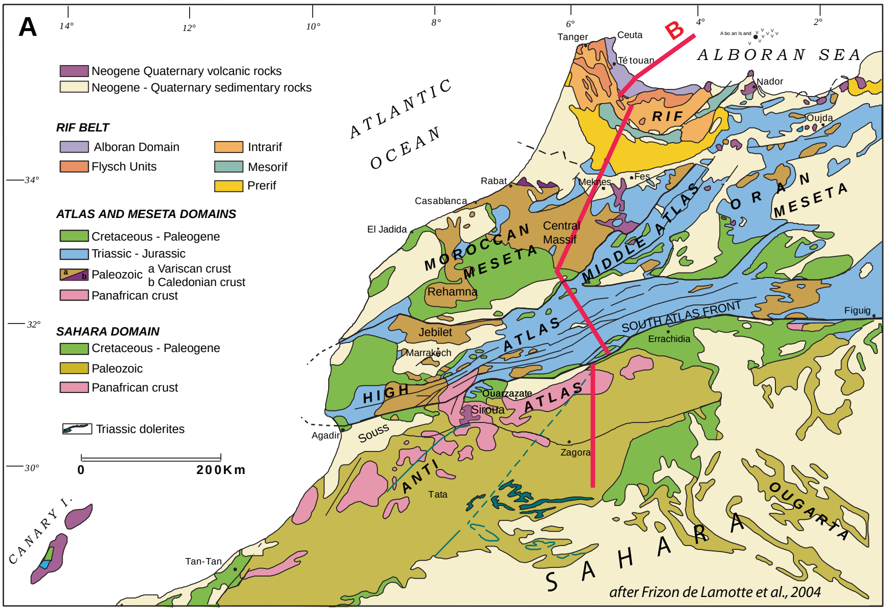

Fig. 1.20 The southern part of TRANSMED I Profile, after Frizon de Lamotte et al. (2004), modified. **A**: Location on the geological map of Morocco (Northern Provinces). – **B**: Lithospheric profile across the different geological domains. For more details see the corresponding figure at greater enlargement on the original CD-rom included in the Transed volume (Cavazza et al., Eds., 2004). The southernmost part of the profile is shown enlarged in Chap. 2

Neoproterozoic-Phanerozoic mountain belts contrast with the WAC domain which displays a remarkably thick (ca. 250 km) lithosphere (Fig. 1.21). The peripheral domain at the border of the craton, deformed during the Pan-African orogeny, then broadly stable and displaying a 130–150 km thick lithosphere are labelled “metacratonic areas”.

Toward the Atlantic Ocean, the Moroccan land is bordered by a 50–100 km-wide marine continental plateau. This shallow marine plateau is limited westward by a large continental slope, about 3000 m-high and 100 km-wide, which leads to the abyssal plains. Depending of the segment considered, this margin shows either active sedimentary progradation or dominant destructional processes such as slumps and slides affecting the seafloor (e.g. Fig. 1.22). The continental slope corresponds to the eastern passive margin of the Central Atlantic Ocean, conjugate to the Nova Scotia margin of North America (Fig. 1.6, 1.8). Its deep structure is better imaged than that of inland Morocco based on offshore seismic reflection profiles combined with wide-angle seismic and gravity data (Fig. 1.23). The shallow structure shows a thick sedimentary cover (up to 6 km at the base of the continental slope) deformed by salt diapirs (see Chap. 6). Basement structures include tilted blocks and a transition zone to the oceanic crust. The crust thins from 35 km underneath the continent to approximately 7 km in the oceanic domain.

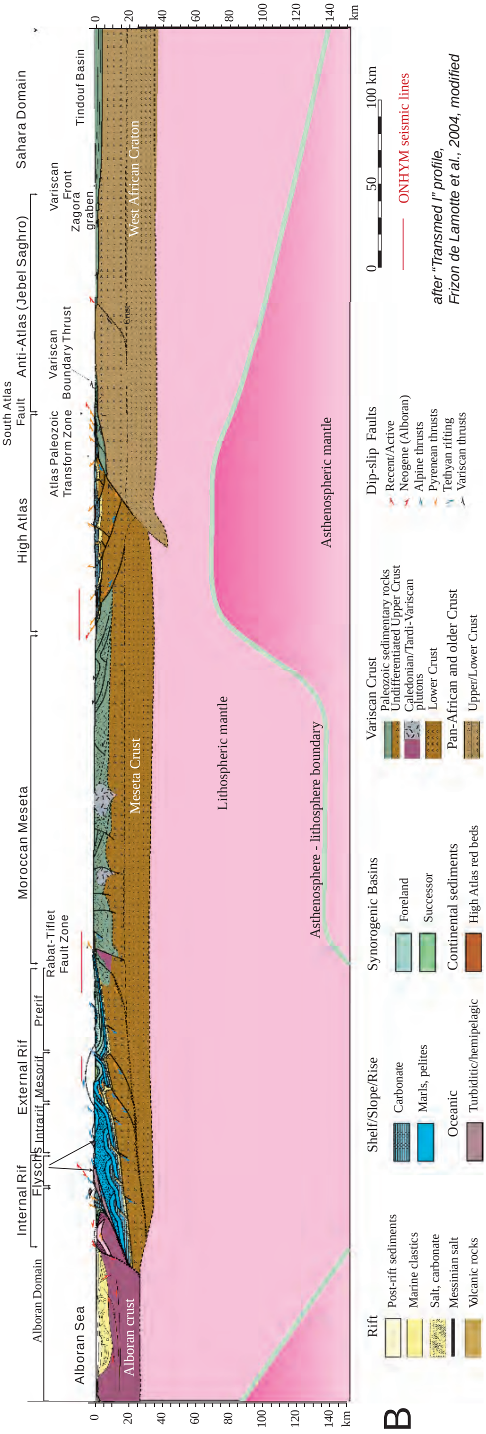

Fig. 1.20 (continued)

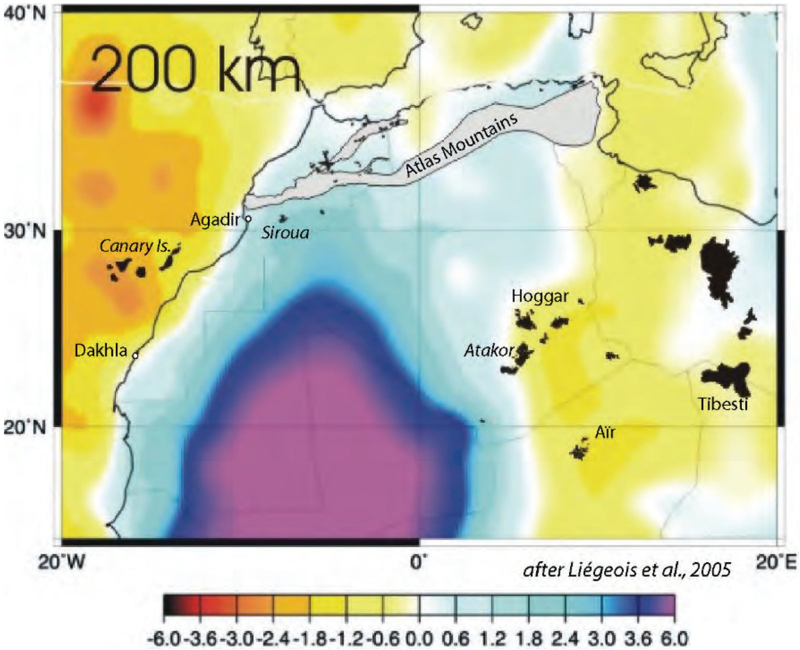

Fig. 1.21 Horizontal cross-section at 200 km depth in a 3D shear-wave velocity tomographic model of North-Western Africa, after Liégeois et al. (2005). Color scale shows the shear wave velocity as % perturbation relative to the reference value that is 4.494 km/s at that depth. *Black*: Cenozoic volcanism; *grey*: Atlas Mountains; fine lines: state borders. The West African Craton is characterized by very high shear-wave velocities which are recognized even at 250 km depth, which indicates a quite thick lithosphere. Cenozoic volcanism is lacking in the cratonic domain, whereas it is abundant in the metacratonic areas further east (Hoggar, etc.)

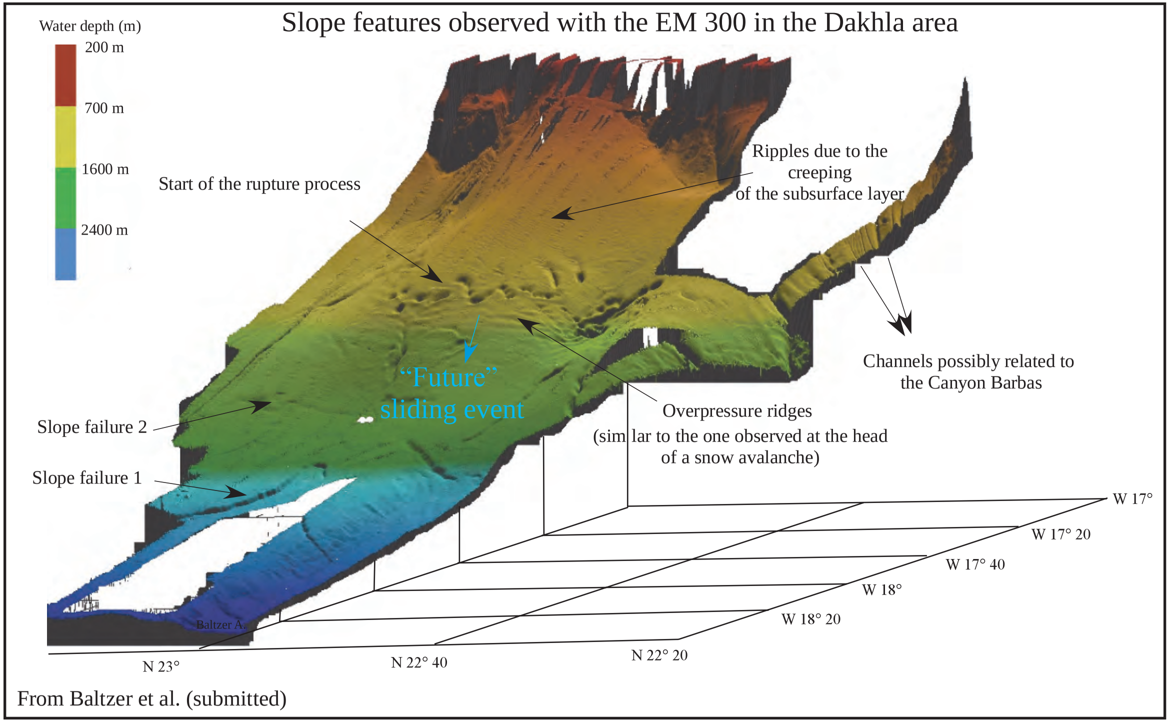

Fig. 1.22 Swath bathymetric data (Simrad EM 300 multibeam echo sounder) on a portion of the West African margin offshore Dakhla, south-western Morocco around 23°N (see Figs. 1.21 or 1.11 for approximate location). The Dakhla cruise (2002) acquired this bathymetric mosaic together with very high resolution seismic data (CHIRP). These complementary data give a 3D view of the shallow sedimentary structures such as creeping processes, slope failures, the initiation of a new failures outlined by overpressure ridges, and channels possibly related to an inactive (?) canyon. Courtesy of A. Baltzer and D. Aslanian (Baltzer et al., submitted)

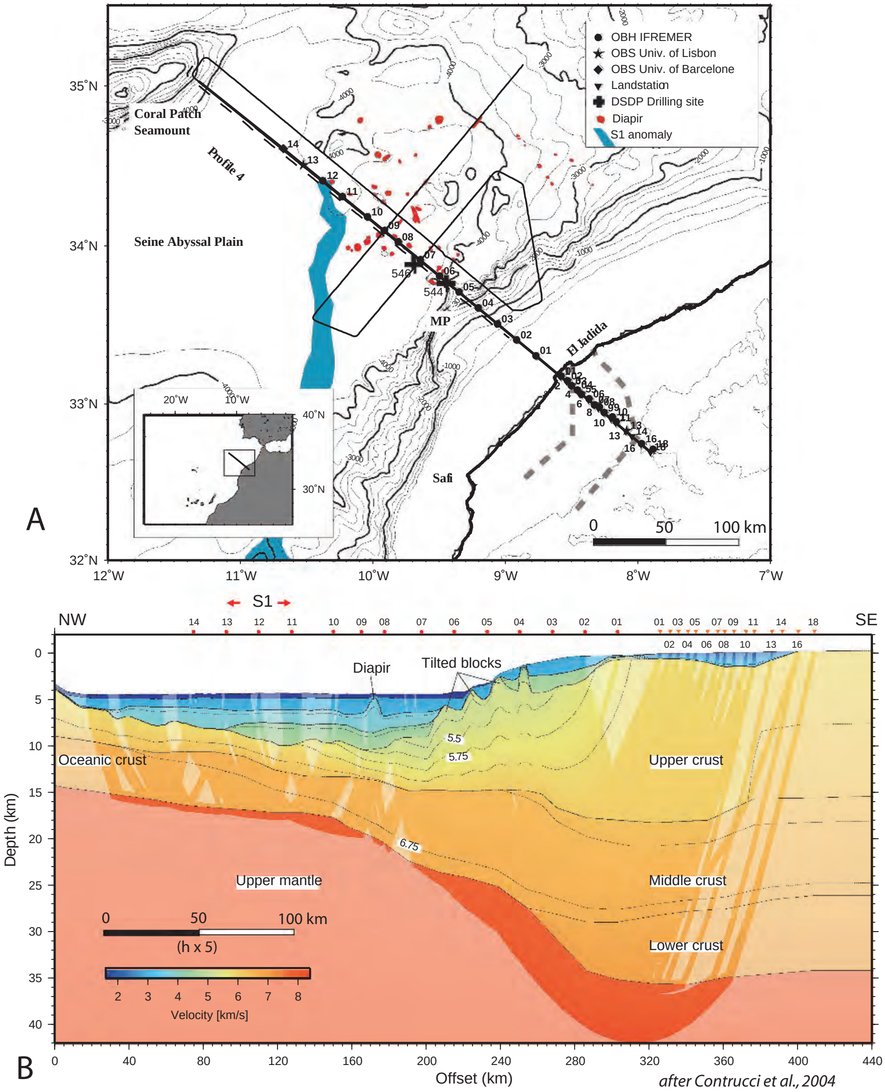

Fig. 1.23 Crustal structure of the NW Moroccan continental margin, after Contrucci et al. (2004), modified. **A**: Location of deep reflection seismic profiles (*dashed line* and *thin lines*) and crustal-scale profile shown in B (*bold line*). The map includes bathymetry. *Thick, broken grey lines* in the vicinity of the land stations indicate the extension of the El Jadida Cretaceous basin. – **B**: Final velocity model deduced from offshore seismic reflection, and both onshore and offshore wide-angle seismic data, integrating the model boundaries used during inversion (*solid lines*) and isovelocity contours every 0.25 km sec^{-1} (*thin lines*). Darker shaded areas show ray paths from the modelling and hence regions of the model that are constrained. Cross-point of the profile with the S1 magnetic anomaly is indicated above the profile, as well as the ocean bottom hydrophones/seismometres (OBH/OBSs) and land stations

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Chapter 2
The Pan-African Belt

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In memory of Georges Choubert, the pioneer of modern Anti-Atlas studies, and Anne Faure-Muret, who joined him for establishing international correlations in the Precambrian of Morocco

2.1 General

References: Since the milestone works of Choubert (1963), and Choubert & Faure-Muret (1970), a number of papers have proposed synthetic, and often contrasting views on the structure and evolution of the Precambrian inliers of the Anti-Atlas, e.g. Benziane & Yazidi (1992), Hefferan et al. (2000), Ennih & Liégeois (2001, 2003, 2008), Thomas et al. (2002, 2004), Inglis et al. (2004), Helg et al. (2004), Gasquet et al. (2005), Deynoux et al. (2006), Liégeois et al. (2006), Bousquet et al. (2008), with references therein. References more specific to the varied tectonic complexes of the area are given in the further sections. The Adrar

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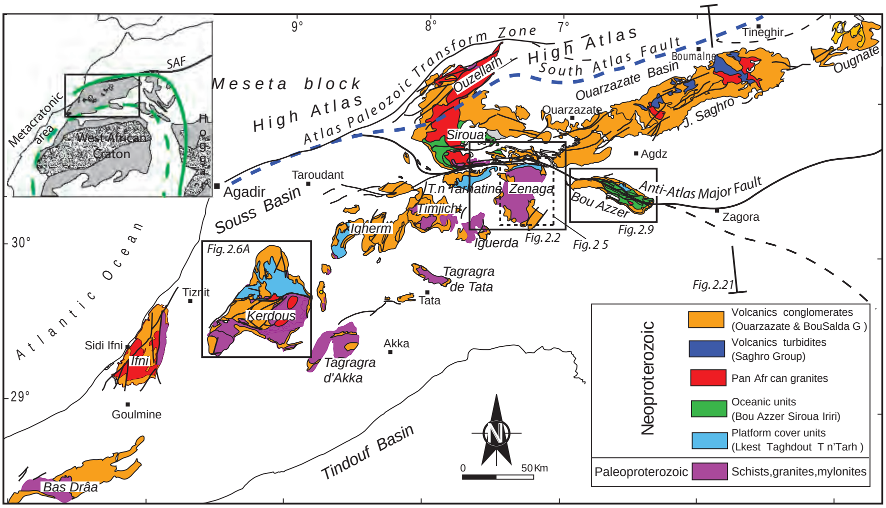

Fig. 2.1 Schematic map of the Anti-Atlas Precambrian inliers (≪boutonnières≫), and location of the maps, satellite views and lithospheric profile presented hereafter

Souttouf Precambrian units are described by Rjimati & Zemmouri (2002) and Villeneuve et al. (2006).

The Pan-African belt exposures in Morocco mainly correspond to large inliers (Fig. 2.1) within the NE-SW oriented Anti-Atlas Paleozoic fold belt (cf. Chap. 1, Fig. 1.4). Each inlier corresponds to a structural culmination of the Variscan fold belt (Chap. 3), but crops out frequently at elevations lower than the overlying, less easily eroded Cambrian sediments, which explains their traditional name "boutonnières" (Fig. 2.2). The exhumation of the Precambrian antiformal boutonnières in the Anti-Atlas axis occurred through erosion of their Paleozoic cover during the Late Carboniferous-Permian, just after the Variscan orogeny, then during the Triassic-Jurassic, when the Anti-Atlas formed the shoulder of the Atlantic and Atlas rifts, and eventually during the Late Eocene-Pleistocene, contemporaneously with the uplift of the Atlas belt itself. Pan-African rocks are also exposed in the eastern block of the High Atlas Paleozoic massif, i.e. the Ouzellarh "promontory", which connects with Anti-Atlas through the Siroua (Sirwa) massif. Additionally, Pan-African rocks make up the allochthonous nappe material of the Mauritanide Variscan belt in the Dhlou and Adrar Souttouf areas (Chap. 1, Fig. 1.11).

The Pan-African belt formed during the Neoproterozoic around the Paleoproterozoic West African Craton (WAC), as reported in Chap. 1. During the last decade, deciphering the successive orogenic events in the Anti-Atlas Precambrian inliers greatly benefited from the Moroccan National Geological Mapping Project, including numerous high sensitivity or conventional U-Pb zircon dates and other geochemical studies. As a result, some of the ideas widely accepted a few years ago concerning, for example, the northern limit of the Paleoproterozoic basement or the

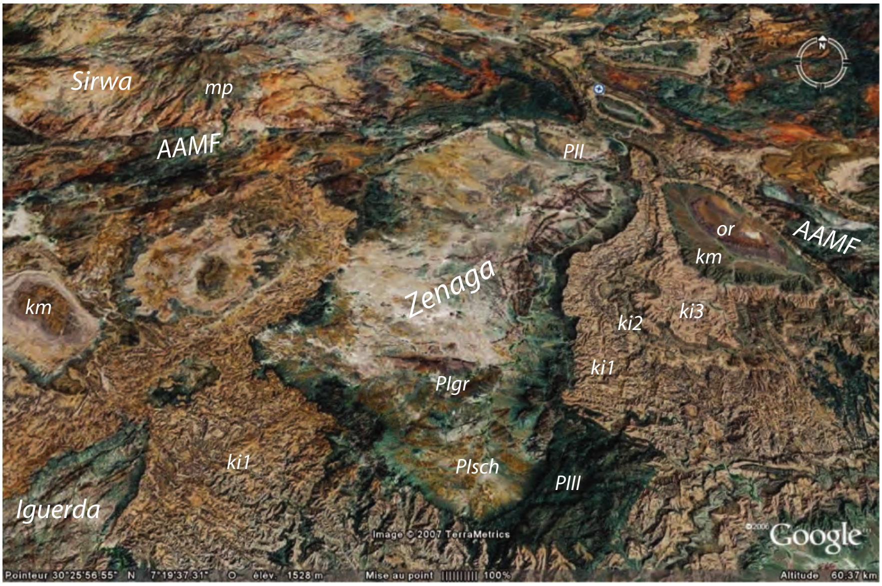

Fig. 2.2 Central Anti-Atlas from space: Google Earth oblique view of the Zenaga-Sirwa area (see Fig. 2.1 for location). The scene is about 70 km wide. Note the shallow elevation of the Zenaga and Iguerda "boutonnières" with respect to their Early Cambrian carbonate blanket (ki1: Adoudounian; ki2: Lie-de-vin Fm; ki3: Calcaires supérieurs). The Zenaga inlier (see map Fig. 2.5) exposes poorly resistant Paleoproterozoic schists and granites (PIsch, PIgr), early Neoproterozoic quartzites and limestones (PII), unconformably overlain by late Neoproterozoic volcanics (PIII). The Sirwa (Siroua) inlier is higher, with the Sirwa Miocene-Pliocene volcano (mp) culminating at 3300 m above sea level. The Anti-Atlas Major Fault (AAMF) marks the southern boundary of the thrust ophiolite and arc units. On the right margin of the scene, note the 5 km left-lateral throw along the AAMF shown by the two white spots that are the two shifted half of the Bou Azzer quartz diorite pluton intrusive in the ophiolite (see Fig. 2.9 for comparison). Variscan deformation is shown by open Middle Cambrian (km)-Ordovician (or) synclines between the Precambrian antiforms

age of the Neoproterozoic Saghro Group, have been challenged. This move is likely to continue for some time, so the present synthesis must be regarded as provisional.

Two groups of Precambrian inliers can be recognized, respectively south and north of the E-trending "Accident Majeur de l'Anti-Atlas" (AAMF, Anti-Atlas Major Fault; Figs. 2.1, 2.2), each group being characterized by partly contrasting lithostratigraphic columns (Fig. 2.3). In the southwestern group (Bas Draa, Ifni, Kerdous, Igherm, Tagragras of Akka and Tata, Iguerda and Zenaga), a Paleoproterozoic crystalline and metasedimentary basement (formerly labelled "PI") occurs beneath Neoproterozoic shallow water metasediments ("PII") affected by the Pan-African orogeny. The basement units are dated at ~ 2 Ga and correlate with the Eburnian basement of the Reguibat Arch (Reguibat Shield or Rise). The Neoproterozoic orogeny generated abundant granites and rhyolites from 630 to 560 Ma (from "PII" to "PIII"). This southwestern part of the Anti-Atlas basement corresponds to the deformed, autochthonous northern rim of the West African Craton (WAC).

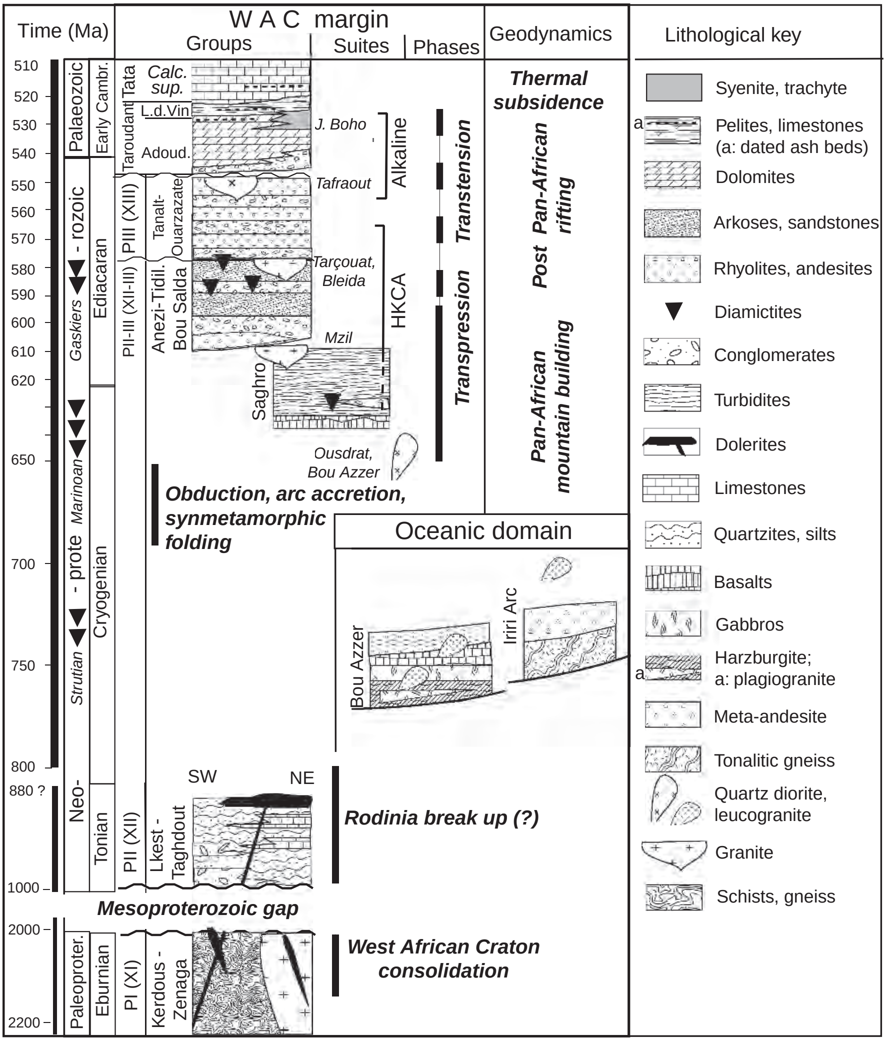

Fig. 2.3 Generalized lithostratigraphic column for the Anti-Atlas Pan-African orogen. “PI” = “XI”, etc. are the classical stratigraphic symbols used on Anti-Atlas geological maps. HKCA: High-K calc-alkaline (granitoids). After Thomas et al. (2004), modified after Gasquet et al. (2005) and Liégeois et al. (2006), and redrawn

In contrast, Eburnian exposures are lacking in the northeastern group of Precambrian inliers, north of the AAMF (J. Siroua, Ouzellarh, J. Saghro, Ougnat). These “boutonnières” no longer display Neoproterozoic platform metasediments, but comprise only oceanic or island arc lithologies below the Pan-African late orogenic formations. This has been considered, and is still considered by some, as the evidence of an ophiolitic suture zone between the WAC and a northern Neoproterozoic continent or island arc, mainly 760–700 Ma in age, represented in the north by the Saghro volcano-sedimentary Group (formerly labelled “PII”) and by the granitoids intrusive into this group. The suture was classically localized in the Bou Azzer-El Graara boutonnière that straddles the AAMF and exposes both ophiolite/arc remnants and supposedly Eburnian orthogneiss units to the north and south of the AAMF, respectively. However, new data have undermined this model: (1) the Saghro granitoids are now precisely dated between 615 and 560 Ma and they are not related to an island arc; their Sr-Nd isotope signatures require an Eburnian basement at depth; likewise, Eburnian xenolithic zircons do occur in the Ougnat rhyolites at Bou Maadine; (2) the detrital zircons within the Saghro Group are coming from the WAC, from the Bou Azzer oceanic complex and from young magmatic bodies between 630 and 612 Ma; (3) the Pan-African magmatic activity on both sides of the AAMF is very similar; (4) the Bou Azzer oceanic complex, precisely dated between 760 and 700 Ma has been overthrust on the WAC at about c. 665 Ma; (5) part if not all of the allegedly Eburnian orthogneisses of the Bou Azzer inlier are Neoproterozoic plutons formed at the expense of a juvenile crust at about 750–700 Ma, and thus are allochthonous as the ophiolite itself (D’Lemos et al., 2006); (6) during the Phanerozoic, the main rheological contrast lies along the SAF and not along the AAMF.

This indicates that the 760–700 Ma rocks are restricted to the allochthonous Bou Azzer/Siroua strip, which underlines the AAMF. The AAMF is bounding to the north the last Eburnian outcrops of the WAC in the Zenaga inlier, but the Eburnian crust continues at depth up to the SAF. Actually, the Pan-African evolution of the Anti-Atlas corresponds to the partial although major destabilization of the northern boundary of the WAC (the “rajeunissement monstrueux” of Choubert & Faure-Muret, 1983), i.e. its metacratonic evolution. The more classical Pan-African belt occurs mostly in the Peri-Gondwanan (or Avalonian) terranes that drifted away during the Phanerozoic.

Similarly, the Pan-African rocks of the Adrar Souttouf include high-grade metabasites and serpentinites (meta-ophiolites) as well as schists and granites in continental units, thrust together upon the Reguibat shield and its Ordovician-Devonian cover (see Chap. 3). An eclogite outcrop south of the Morocco-Mauritania border yielded a U-Pb age of 595 Ma on individual zircon grains, interpreted as the protolith age, whereas garnets from the same rock yielded a Sm-Nd age at 330 Ma (Variscan metamorphism). K-Ar dates from central Adrar Souttouf units are scattered between 1800 and 262 Ma, consistent with a polyphase Paleoproterozoic-Paleozoic evolution. In the following sections, only the Anti-Atlas Precambrian inliers are considered.

2.2 The Paleoproterozoic Basement

References: Structural, geochemical and geochronological data on the Anti-Atlas Paleoproterozoic rocks can be found in Hassenforder (1985, 1987), Aït Malek et al. (1998), Mortaji et al. (2000), Ennih et al. (2001), Ennih & Liégeois (2001, 2008), Benziane et al. (2002), Walsh et al. (2002), Gasquet et al. (2004), Soulaimani & Piqué (2004), Barbey et al. (2004), Benziane (2007). A review of the Eburnian-Transamazonian belt was published by Bertrand & Jardim de Sá (1990). Other references concerning this belt in the Reguibat Shield are given in Chap. 1.

The basement units of the Anti-Atlas Precambrian massifs south of the AAMF include both metamorphic and magmatic rocks, referred to as “PI” or “XI” on geological maps. The metamorphic rocks range from low-grade phyllites and greenschists (“Bas Draa” and “Had n’Tahala Groups”) to amphibolite facies schists and migmatites (“Zenaga Group”). The dominant trend of the Eburnian foliation is ESE, before further deformation, which is consistent with the SE structural trends in the eastern Reguibat Arch (Yetti, Eglab). In the Tagragra of Tata, the schist series contain felsic metatuffs that yielded a zircon age of 2072 ± 8 Ma. However, the Zenaga Group is possibly as old as c. 2170 Ma, based on U-Pb SHRIMP dates from relic zircon cores from intruding granites. The occurrence of Archaean relics (former “P0”) is not yet established by reliable dating.

The metamorphic rocks are intruded by syn-tectonic to post-tectonic dolerites and granitoids. Several granitoids exposed in the Bas Draa, Kerdous (Tahala granite), Tagragras of Tata and Akka (Fig. 2.4), and Igherm inliers yielded dates of c. 2000–2050 Ma (U-Pb zircon), strongly contrasting with the age of the late Pan-African plutons of the same inliers, i.e. 580–560 Ma. A similar, Paleoproterozoic age (c. 2030 Ma, U-Pb zircon) was obtained for the Azguemerzi peraluminous granodiorite and Tazenakht porphyritic monzogranite, which intrude the Zenaga micaschists and migmatized paragneisses (Fig. 2.5).

More precisely, two plutonic events have been recognized in the Paleoproterozoic basement of the western Anti-Atlas: the first consists of a calc-alkaline suite of diorites, monzogabbros-diorites, granodiorites and granites reflecting a lower crustal or mantle origin with variable contamination by crustal material; the second corresponds to peraluminous granodiorites, granites and leucogranites originating from a crustal source. The siliciclastic nature of the host schist protoliths implies the

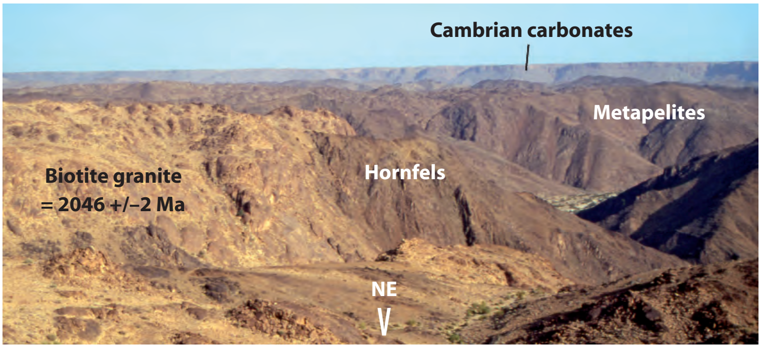

Fig. 2.4 Paleoproterozoic basement of the Tagragra of Akka (southeastern part) surrounded by the Adoudounian (Latest Ediacaran-Early Cambrian) cliffs in the background (NE). Note the importance of recent uplift and correlative deep incision of the Mesozoic peneplain, now elevated at c. 1400 m a.s.l. in the area. Photograph by D.G.

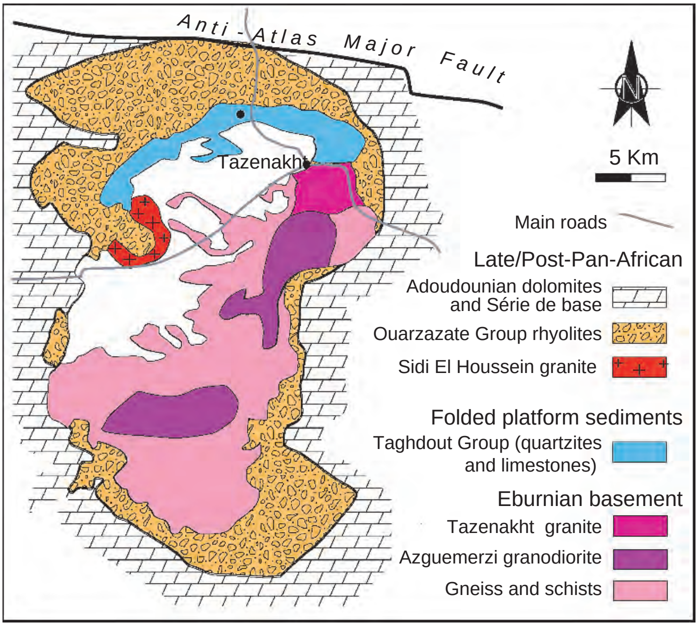

Fig. 2.5 Schematic geological map of the Zenaga boutonnière (compare with Fig. 2.2). The Azguemerzi pluton is a peraluminous granodiorite-monzogranite; the Tazenakht pluton is a porphyritic monzo-syenogranite; both are dated at c. 2030 Ma. The country-rocks are amphibolite facies schists and gneisses. The Pan-African greenschist facies deformation developed under N–S directed stress, which formed tight folds in the Neoproterozoic Quartzites and Limestones (Taghdout Group), and caused mylonitization of the northern part of the Tazenakht pluton with sinistral strike-slip. The Sidi el Hussein ring-dyke granite ("PIII") is dated at 579 ± 7 Ma. After Ennih & Liégeois (2001), redrawn

existence of a neighbouring older domain of probable Archaean age. A few detrital Archaean zircons have been found in the Saghro Group, corroborating this model. Contribution of both juvenile (granites) and recycled (metasediments) Archaean material suggests that, during the Paleoproterozoic, the Anti-Atlas was a zone of accretion close to an Archaean nucleus comparable to that known in the southwestern Reguibat Arch. The geodynamic setting of the Anti-Atlas (WAC) Eburnian belt is reminiscent of the Archean-type granite-greenstone belt associations. By contrast, 2 Ga high-grade gneisses, nappes and syntectonic granites are mostly known as reworked terranes within the Pan-African (-Braziliano) mobile belt, suggesting that the location of the latter belt was controlled by the major structures inherited from the Early Proterozoic (Bertrand & de Sá, 1990).

Numerous mafic (gabbros, dolerites) and felsic (microgranites) dykes crosscut the Paleoproterozoic basement of the western Anti-Atlas. Most can be assigned to the Neoproterozoic rifting of the Paleoproterozoic craton (see next section), but a dolerite dyke from the Tagragra of Tata yielded a SHRIMP age at 2041 ± 6 Ma (Walsh et al., 2002; Benziane et al., 2002). A microgranite from the Kerdous inlier (Fig. 2.6), and pegmatites from the Kerdous and Tagragra of Akka inliers have been

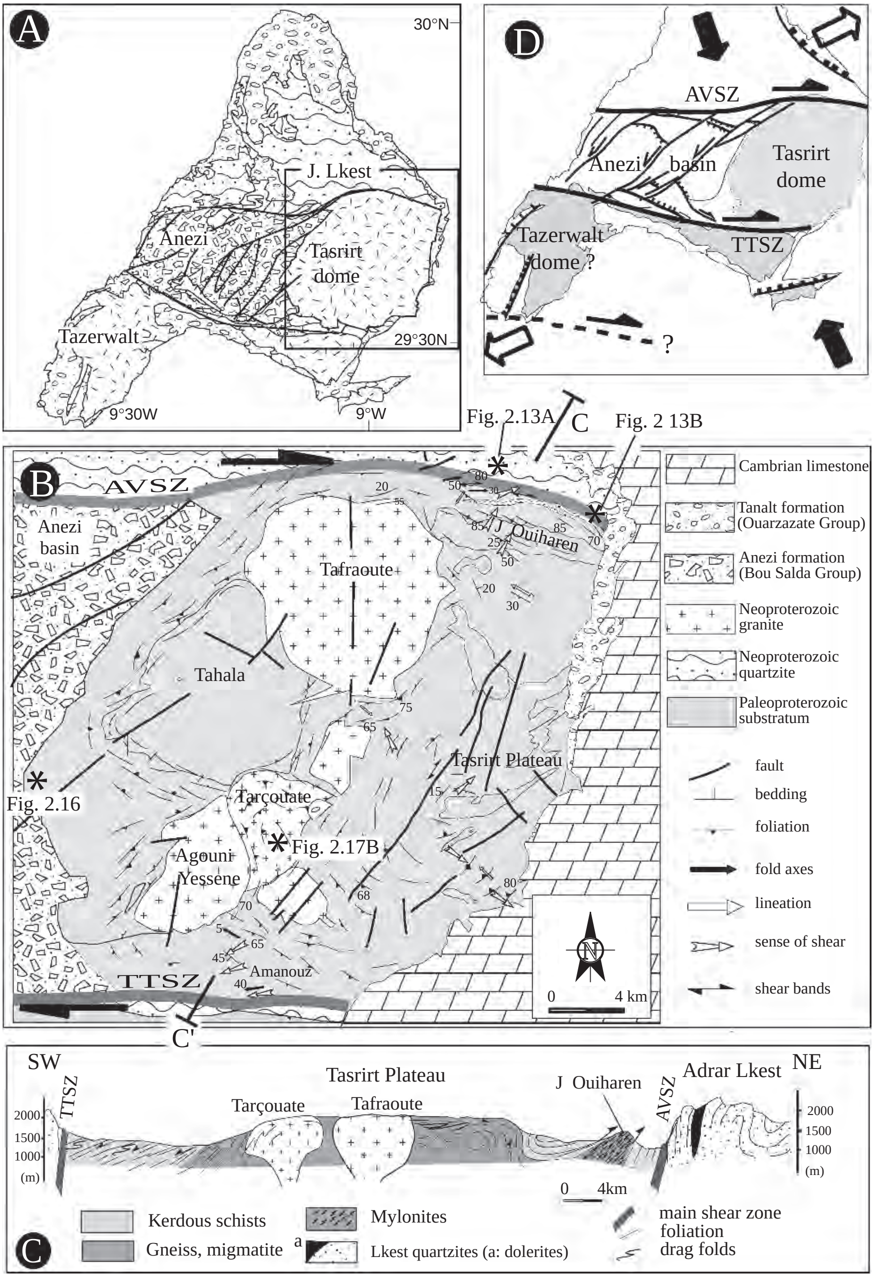

Fig. 2.6 Pan-African structure of eastern Kerdous inlier. (**A**): Schematic map of the inlier (see Fig. 2.1 for location). – (**B**): Structural map of the southeast J. Lkest-Tasrirt Plateau area. – (**C**): Cross-section (trace CC′ in map (**B**)). – (**D**): Interpretation of the regional stress orientation by the end of the synmetamorphic greenschist facies compression. AVSZ, TTSZ: Ameln Valley and Tighmi-Tifermit Shear Zones. Age of granite intrusions as follows: Tahala, 2044 ± 2 Ma (Barbey et al., 2004); Tarçouate, 581 ± 11 Ma Aït Malek et al. (1998); Tafraoute and Agouni Yessen, 549 ± 6 Ma (Pons et al., 2006)

dated at c. 1760 Ma (zircon, monazite and muscovite ages; Gasquet et al., 2005). The latter dykes are linked to a late Paleoproterozoic (Statherian) magmatic event otherwise unidentified in the Anti-Atlas.

Pan-African greenschist-facies retrogression and mylonitization obviously affect the Paleoproterozoic granites and their country-rocks together with their overlying Neoproterozoic cover, although with no significant lithospheric thickening. The Neoproterozoic Taghdout passive margin sediments are still well preserved including sedimentary features such as ripple marks or mud cracks (Fig. 2.7A, B). This Pan-African event has no resolvable imprint on the isotopic system of the zircons but has been able to affect the Sm-Nd isotopic ratios of a part of the Eburnian rocks: this is related to the abundant fluid movements that occurred during the emplacement of the huge Ouarzazate volcanic Supergroup. The exhumation of the Paleoproterozoic units during the Pan-African orogeny can be ascribed either to the dominantly left-lateral transpressive tectonics that occurred between 630 and 580 Ma or to the subsequent transtensional tectonics (580–550 Ma), or else to both these superimposed phenomena. A good example of a Paleoproterozoic dome uplifted at the same level as the folded Taghdout metasediments is found in the Kerdous massif (Tasrirt dome, Fig. 2.6). The dome is bounded north and south by two major dextral shear zones, which operated under greenschist followed by cataclastic conditions. Both shear zones are sealed by the Ouarzazate conglomerates (Tanalt Fm). The Tasrirt dome has been interpreted by one of us (A.S.) as a Late Neoproterozoic diapiric gneiss dome emplaced in a NE-SW pull-apart system.

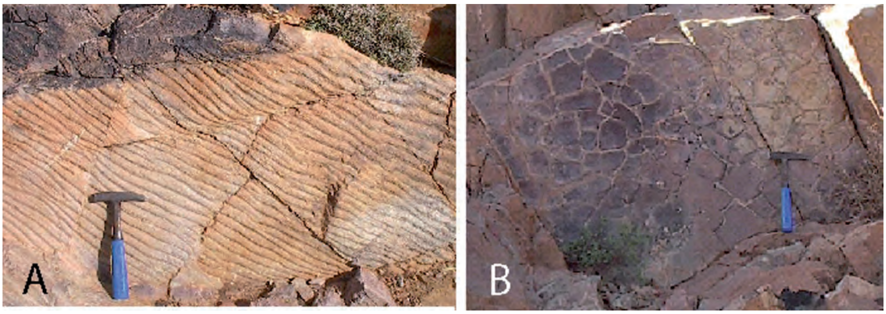

Fig. 2.7 Sedimentological features from the Taghdout Group, Zenaga boutonnière. – **A**: Ripple marks on top of a quartzite layer. – **B**: Mud cracks in argillite interleaved in a sequence of stromatolitic-thrombolitic limestones. Photographs by N.E.

2.3 The Early Neoproterozoic Platform Margin

References: The early Neoproterozoic cover of the WAC margin and the associated basic sills and dykes are described in Choubert (1963), Clauer (1976), Moussine-Pouchkine & Bertrand-Sarfati (1978), Benziane & Yazidi (1982), Hassenforder

(1987), Ikenne et al. (1997), De Beer et al. (2000), El Aouli et al. (2001), Ennih et al. (2001), Benziane et al. (2002), Thomas et al. (2002), Bouougri & Saquaque (2004), Deynoux et al. (2006), with references therein. About the concept of supercontinent Rodinia, see Dalziel (1997), Weil et al. (1998), Cordani et al. (2003). The Proterozoic glaciations have been reviewed by Ramstein et al. (2004).

Considering the recent data that place the Saghro Group in the 630–610 age range (see Sect. 1.3.2), only a single (meta) sedimentary group can be considered now as pre-Pan-African, which is labelled the "Taghdout Group" in the Zenaga inlier and "Lkest Group" in the Kerdous massif (see also "Tizi n'Tarhatine Group"). Both these equivalent groups include thick, layered quartzites with conspicuous sedimentary structures (ripple marks, mud cracks, bioturbations, etc.), stromatolitic carbonates and sandy-pelitic beds (Fig. 2.7A, B). A progressive change of the sedimentary facies is observed from SW to NE: conglomerates and sandstones dominate in the Bas Draa and Ifni inliers, quartzites and pelites in the Kerdous and Igherm inliers, and quartzites, limestones and pelites in the Tata, Zenaga and Bou Azzer inliers, closer to the AAMF. These shallow water formations represent a cover sequence with respect to the Paleoproterozoic basement, although their common contact is generally faulted. The major unconformity on top of the Eburnian schists is locally well-preserved (Taghdout, Tizi n'Tarhatine; Fig. 2.8). The detrital zircons from the quartzites above the Tagragra of Tata schists yielded consistent U-Pb SHRIMP dates between 1990–2080 Ma (Benziane, 2007).

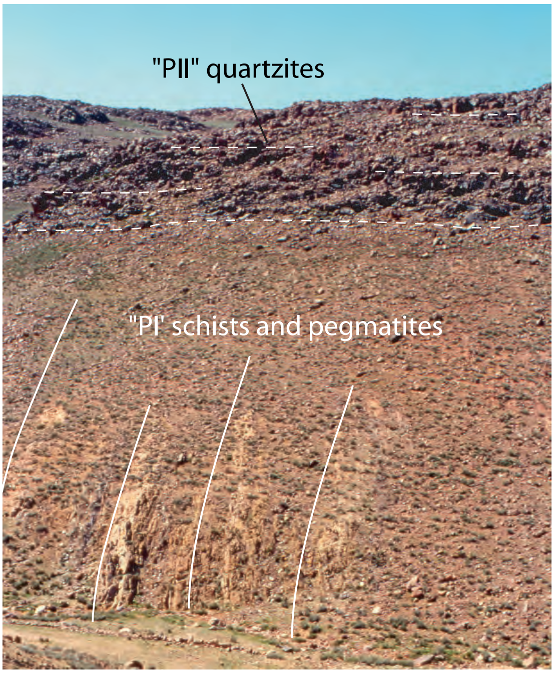

Fig. 2.8 The Tizi n'Tarhatine unconformity between Eburnian schists ("PI") intruded by c. 2 Ga old pegmatites and Neoproterozoic conglomeratic quartzites ("PII", Taghdout-Lkest Group). The first published photograph from the area was by L. Neltner in 1938, as evidence of the "Algonkian" unconformity over allegedly "Archean" terranes. However, G. Choubert observed that the unconformity also corresponds to a décollement level related to tight, dysharmonic Pan-African folds (cited in Michard, 1976, pp. 47 and 56). Photo by D.G.

The Taghdout Group sediments are intruded by abundant dykes and sills of dolerites and gabbros (e.g. J. Lkest, Fig. 2.6C), also found as dyke swarms in the Paleoproterozoic basement (e.g. Igherm, Tagragra of Akka, Kerdous, Zenaga). These rocks (Ifzwane Suite; Thomas et al., 2004) are broadly akin to continental tholeites, although with strong chemical heterogeneity, and could be associated with the increasing rifting of the WAC margin. They are not directly dated.

The depositional age of the Taghdout Group itself is not well constrained. The presence of stromatolites (Choubert, 1963) points to a Neoproterozoic age, i.e. younger than 1000 Ma. Location of the Taghdout Group below the Bou Azzer oceanic complex indicates an age older than 660 Ma, which is the age of obduction. An age of 788 ± 9 Ma (Clauer, 1976) obtained by Rb-Sr on clay fractions from Taghdout metasediments gives a minimum age for the deposition of the Group. However, detrital 880 Ma old zircons from the Saghro Group (Liégeois et al., 2006) suggest a still older minimum age as these zircons can be attributed to the magmatic event associated to the Ifzwane Suite intrusive in the Taghdout Quartzites and Limestones. If correct, the Taghdout shallow water sediments would have accumulated from about 1000 Ma to > 880 Ma, in a proximal passive margin environment upon the northern border of the WAC.

The correlations at the craton scale favourably support the latter proposal. The Taghdout sediments compare with the Char and Atar Groups, which can be traced all along the south border of the Reguibat Arch/north border of the Taoudenni Basin (Atar, Richat, and Hank areas; see Fig. 1.11 for location). These groups form a c. 1000 m thick succession beginning with siliciclastic deposits that range from fluvial to wave- or tide-dominated sediments (Char Group), and passing upward to sandy, stromatolite–bearing carbonate shale sequences (Atar Group; see Deynoux et al., 2006). A clay fraction from the Char Group yielded a Rb-Sr age at 998 ± 32 Ma, whereas Rb-Sr and K-Ar dating of clay fractions from the Atar Group yielded ages between 890 ± 35 and 775 ± 52 (Clauer, 1976). Likewise, correlations can be extended up to the Gourma aulacogen (Moussine-Pouchkine & Bertrand-Sarfati, 1978), on the eastern side of the WAC/southeast side of Taoudenni Basin. Everywhere on the WAC, the dominantly Early Neoproterozoic (Tonian, 1000–850 Ma) platform sedimentation can be related to the break-up of the hypothetical supercontinent Rodinia, as proposed for the Gourma aulacogen. Remarkably, this sedimentation occurred after nearly 1 Ga of quiescence in the West African craton (no event is recorded between 1.7 Ga and the sedimentary onlap).

2.4 The Ophiolite/Arc Complex and its Accretion (Pan-African I Phase)

References: The Pan-African oceanic/transitional units from the Bou Azzer, Siroua and Saghro inliers are described by Leblanc (1981), Saquaque et al. (1989), Admou & Juteau (1998), De Beer et al. (2000), Wafik et al. (2001), Thomas et al. (2002), Hefferan et al. (2002), Samson et al. (2004), Fekkak et al. (2003), Beraaouz et al. (2004), D'Lemos et al. (2006), Soulaimani et al. (2006), Bousquet et al. (2008).

2.4.1 The Oceanic Complex (760–700 Ma)

Well-preserved meta-ophiolites ("Bou Azzer Group") are shown in the Bou Azzer (Ait Ahmane; Fig. 2.9) and Siroua (Khzama, Nqob) inliers. They comprise mantle harzburgites, and a crustal sequence typical for fast oceanic ridges, including layered gabbros and sheeted dykes beneath the pillow basalts section (Fig. 2.10). Only rare metasediments associated with volcanic layers are found on top of the oceanic complex (e.g. Ambed Co-bearing calcareous jaspers at Bou Azzer; ketatophyric tuffites and flows upon the Khzama pillow lavas).

The second component of the lost oceanic domain is best defined as the "Iriri island arc" in the Siroua inlier by a metamagmatic association comprising

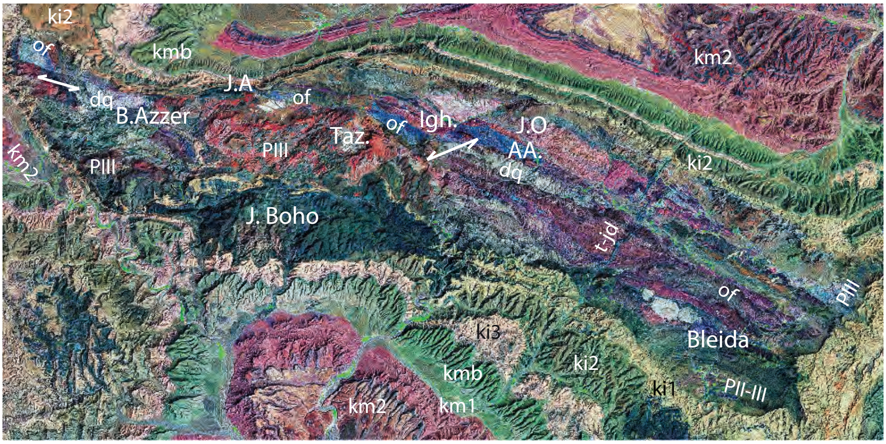

Fig. 2.9 Landsat image of the Bou Azzer-El Graara inlier (see Fig. 2.1 for location). The scene is about 60×45 km large. The WNW-ESE stripes of ophiolitic remnants (of), ~ 750 Ma old, and quartz diorite plutons (dq), c. 650 Ma, mark the trend of the Anti-Atlas Major Fault (AAMF) along the axis of the Precambrian boutonnière. Note the sinistral throw of the quartz diorite pluton on the AAMF axial branch west of the Bou Azzer mining centre. The AAMF and associated (Riedel) faults such as between Igherm (Igh.) and Aït Ahmane (AA) are sealed by the Ouarzazate Group (PIII) volcanics (c. 570–560 Ma) and overlying Lower Cambrian carbonates (ki1). J. Ousdrat (J.O.) is a syn-accretion quartz diorite similar to the Bou Azzer one. In contrast, the 580 Ma old Bleida granodiorite cross-cuts the regional fabric; it would even postdates the Tiddiline series (PII–III) deformation (arcuate north-dipping layers south of Bleida). In the Cambrian blanket, the J. Boho alkaline volcanics are clearly intercalated in the Adoudounian carbonates (ki1). The J. Aghbar (J.A.) is a coeval (530 Ma) syenite sill. Note the mild Variscan deformation of the Paleozoic sequence (ki2: Lie-de-vin series; ki3: Calcaires supérieurs; kmb: Grès terminaux, earliest Middle Cambrian; km1: Schistes à Paradoxides; km2: Grès du Tabanit, late Middle Cambrian). The 200 Ma old Foum Zguid dolerite mega-dyke (t-jd) cross-cuts the Variscan belt and the Precambrian basement with few post-emplacement faulting

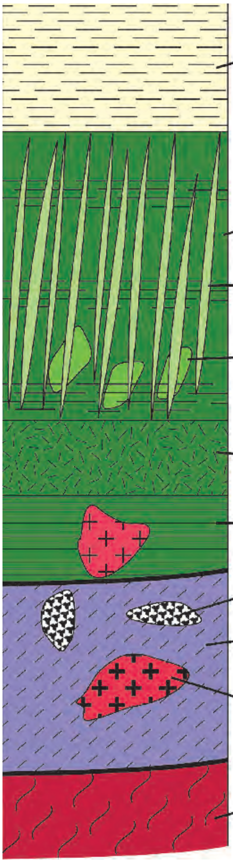

Fig. 2.10 Restored stratigraphic column of the Khzama ophiolite, Siroua massif, after Wafik et al. (2001), redrawn. At regional scale (including Bou Azzer), gabbroic rocks are c. 760–750 Ma old, whereas the juvenile leucogranites and quartz diorites (schematically shown) emplaced at 750–700 Ma and 650–640 Ma, respectively (Thomas et al., 2002, 2004)

medium-grade biotite-rich (andesitic) schists and layered tonalitic orthogneisses (Fig. 2.11). The arc units also include the Ourika complex from the Ouzellarh block.

A main island arc episode occurred in the 760–740 Ma period: an ophiolitic basalt from the Siroua massif was dated by Sm-Nd at c. 740 Ma, two tholeiitic plagiogranites from the ophiolitic sequence of the same massif yielded U-Pb dates of 761 ± 2 Ma and 762 ± 2 Ma; in Bou Azzer the Tazigzaout augen gneiss (previously regarded as Eburnian) yielded an age of 753 ± 2 Ma, and a metagabbro an age of 752 ± 2 Ma; the protolith of the Iriri migmatites has been dated as 743 ± 14 Ma. A second phase occurred at c. 700 Ma with the intrusion of juvenile leucogranites post-dating the deformation of the earlier plutons. A continuous event (arc building) between 760 and 700 Ma is possible.

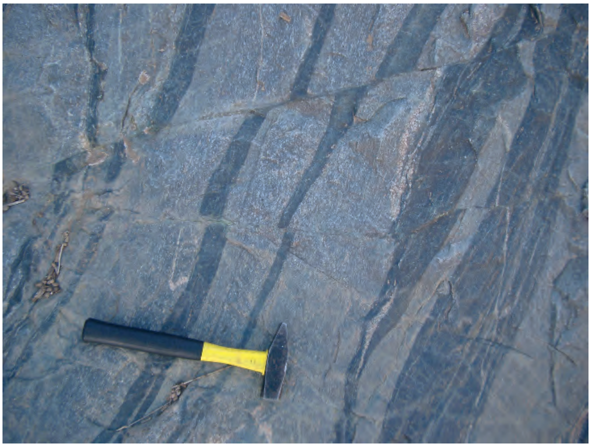

Fig. 2.11 Tonalitic migmatite including foliated metabasic intrusions. Iriri arc, Siroua inlier (*vertical* view). Photo JP.L. The protolith of the Iriri migmatites is dated at 743 ± 14 Ma (Thomas et al., 2004)

The Bou Azzer – Siroua oceanic complex is known only as a discontinuous strip along the AAMF and represents nearly all the Neoproterozoic juvenile magmatic rocks known in the Anti-Atlas. The younger magmatism always comprises an important contribution from the old WAC lithosphere, except in the first basaltic events from the Saghro Group. Nowhere has been found typical high-pressure mineral associations and the alleged blueschist facies mineral associations correspond only to crossite/Mg-riebeckite-bearing HP-greenschist conditions (5–6 kbar, 500–550 °C; Bousquet et al., 2008). The ophiolites and arc units emplaced as south-vergent thrusts (obduction) above a basal mélange onto the WAC passive margin, which means that the root of the oceanic suture has to be located somewhere north of the SAF, not along the AAMF itself. This is supported by the magnetic modelling of the Bou Azzer ophiolitic suture (Fig. 2.12), which suggests the occurrence of a shallow, north-dipping subduction zone up to the High Atlas region.

2.4.2 The Pan-African I Syn-Metamorphic Phase (660–640 Ma)

The accretion of the oceanic complex towards the WAC probably occurred at 663 ± 13 Ma, the age of the zircon rims in the Iriri migmatites. Greenschist facies metamorphism and coeval folding affected the platform beneath the obducted terranes and in front of them. Subsequently, juvenile quartz diorites intruded at 652 ± 2 Ma and 640 ± 2 Ma in the Bou Azzer inlier (e.g. Ousdrat pluton), the significance of which is not well understood (slab breakoff? See last section). These intrusions give an accurate upper limit for the age of the accretion phase.

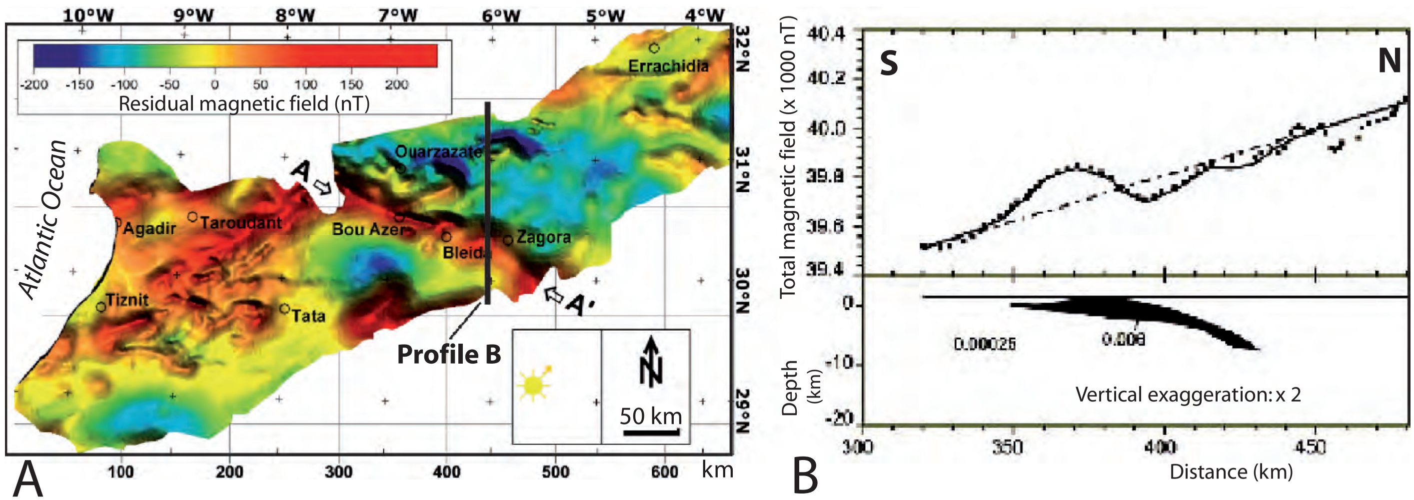

Fig. 2.12 Magnetic modelling of the Bou Azzer-El Graara ophiolite. (**A**): Map of the residual magnetic field of the Anti-Atlas. (1): shading direction; A–A′ arrows: location of the Central Anti-Atlas negative anomaly. – (**B**): Magnetic modelling along profile 40 east of Bleida (for location, see map (**A**)). The major negative anomaly A–A′ coincide with the ophiolite outcrops along the AAMF (compare with Fig. 2.1). Taking into account the latitude and the lack of remnant magnetization, this indicates a north-dipping ophiolitic body, as quantitatively modelled in (**B**). After Soulaimani et al. (2006)

The deformation of the Taghdout Group sediments south of the AAMF is likely coeval with this “Pan-African I phase”. Shortening was associated with widespread décollement from the underlying Eburnian basement. Greenschist facies mineral assemblages, locally with chloritoid and andalusite occurrencies (J. Lkest) are typical for this event. Folding developed in relation with compression or transpression under N–S to NW–SE oriented maximum stress. In the Kerdous massif (Fig. 2.6), north of the Ameln Valley Shear Zone, the J. Lkest range shows WNW-trending fold axes that roughly parallel the mylonitic corridor (Fig. 2.13A). The Tasrirt-Ouiharen basement south of the shear zone tends to overthrust the Lkest cover unit, and displays superimposed Eburnian and Pan-African microstructures. The horizontal shear component is right-lateral in that case, whereas it is mainly left-lateral along the AAMF at Bou Azzer (Fig. 2.9), although dextral ductile shear was also described in part of the latter inlier (Tazigzaout). The Anezi (“PII-III”) and Tanalt-Ouarzazate conglomerates (“PIII”) unconformably overlie the folded platform units after deep erosion (Fig. 2.13B).

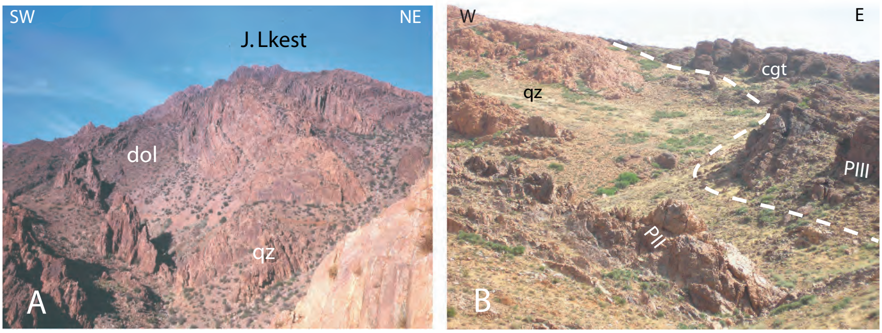

Fig. 2.13 The J. Lkest Pan-African fold range (**A**) and its unconformable late Neoproterozoic cover (**B**), north and east of the Ameln Valley Shear Zone (see Fig. 2.6 for location). (**A**): The elevated Lkest range consists of Neoproterozoic quartzites (qz) and intrusive dolerites (dol) tightly folded together. – (**B**): The slightly tilted Late Ediacaran conglomerates of the Tanalt Group (PIII, cgt) unconformably overlie the nearly vertical Lkest quartzites at the northeast border of the massif

2.5 The Pan-African II Metacratonic “Phase” and Coeval Formations (630–550 Ma)

References: The earliest sedimentary formations (Saghro Group) from this period have been described, and considered as distal equivalent of the platform formations in the following papers: Fekkak et al. (1999, 2001, 2003), De Beer et al. (2000), Thomas et al. (2002, 2004). The relatively young age of these formations and their geodynamic setting are established by Errami et al. (2006), Liégeois et al. (2006).

The more recent, late orogenic to post-orogenic formations, i.e. Bou Salda and equivalents (Anezi, Tiddiline?) and Ouarzazate Groups, and the associated magmatic suites and geodynamics are described in Hassenforder (1987), Lécolle et al. (1991), Benziane & Yazidi (1992), Mokhtari et al. (1995), Aït Malek et al. (1998), De Beer et al. (2000), Errami (2001), Thomas et al. (2002, 2004), Lécolle et al. (2003), Inglis et al. (2004), Levresse et al. (2004), Gasquet et al. (2005), Benziane (2007), with references therein. The importance of extension during the last period (Neoproterozoic-Cambrian boundary) is emphasized by Soulaimani et al. (2003, 2004), Soulaimani & Piqué (2004) and Gasquet et al. (2005).

Late Proterozoic glaciations in the Saharan and Oman areas are discussed in Deynoux et al. (2006) and Le Guerroué et al. (2005), respectively. See also the general review by Ramstein et al. (2004).

Most Anti-Atlas ore deposits are related to the late Neoproterozoic magmatic events and coeval extensional tectonics. Recent metallogenic data can be found in Ouguir et al. (1994), Mouttaqi (1997), Cheilletz et al. (2002), Barakat et al. (2002), Abia et al. (2003), Levresse et al. (2004), Benchekroun & Jettane (2004), Gasquet et al. (2005), Marcoux & Wadjinny (2005).

Three volcano-sedimentary groups occurred during that long lasting period: the Saghro Group (630–610 Ma), the Bou Salda Group (610–580 Ma) and the Ouarzazate Group (580–545 Ma). The earliest group (Saghro Group) still suffered low grade greenschist facies deformation; the Bou Salda Group show significant brittle deformations, whereas the youngest group (Ouarzazate Group) is clearly transitional with the post-orogenic formations.

2.5.1 The Saghro Group and the Greenschist Facies Transpressive Event

The sedimentary-volcanoclastic sequence of the Saghro Group (Sidi Flah, Kelaat Mgouna, Boumalne, Imiter subgroups) corresponds to a great thickness (up to 6000 m) of siliciclastic turbidites, grading upward from shales, siltstones and sandstones at the bottom to coarser deposits on top, with rare limestone layers (Fig. 2.14A). It is worth noting that one of the lowermost turbidite formations contains several diamictite horizons. These deposits alternated with basaltic flows (locally with pillow structures; Fig. 2.14B) having the chemical signatures of rift tholeites and alkaline intraplate basalts, and with volcanoclastic keratophyres and tuffites layers. Small (20–50 m) ultramafic lenses occasionally occur, associated with hydrothermal jaspers and ophicalcites.

Classically, this group was supposed to be a distal equivalent of the "PII" Taghdout Group. However, a comprehensive U-Pb dating (laser ICP-MS) of detrital zircon from the Saghro Group has recently been realized (Liégeois et al., 2006): four samples were dated from the bottom to the top of the Kelaat Mgouna subgroup (Fig. 2.15). The number of Neoproterozoic zircons increases from the bottom (10%) to the top (67%), the rest being Eburnian zircons from the WAC; moreover the

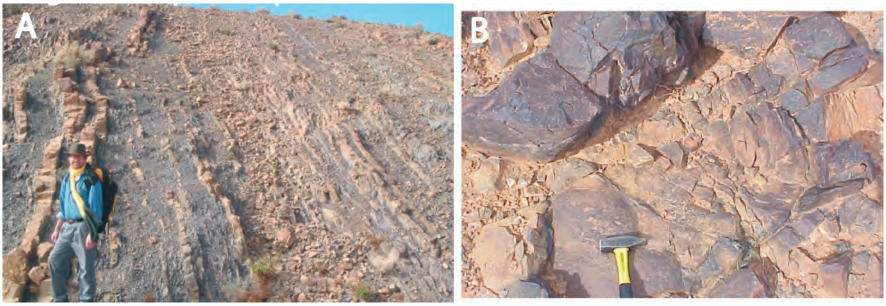

Fig. 2.14 Typical outcrops from the Saghro Group. (**A**): Steeply dipping turbiditic layers, folded under lower greenschist facies condition about 600 My ago. – (**B**): Pillow lavas near the bottom of the turbiditic sequence, recording the early rifting of the metacratonic margin (TDM Nd model ages between 640 and 580 Ma). Photos JP.L.

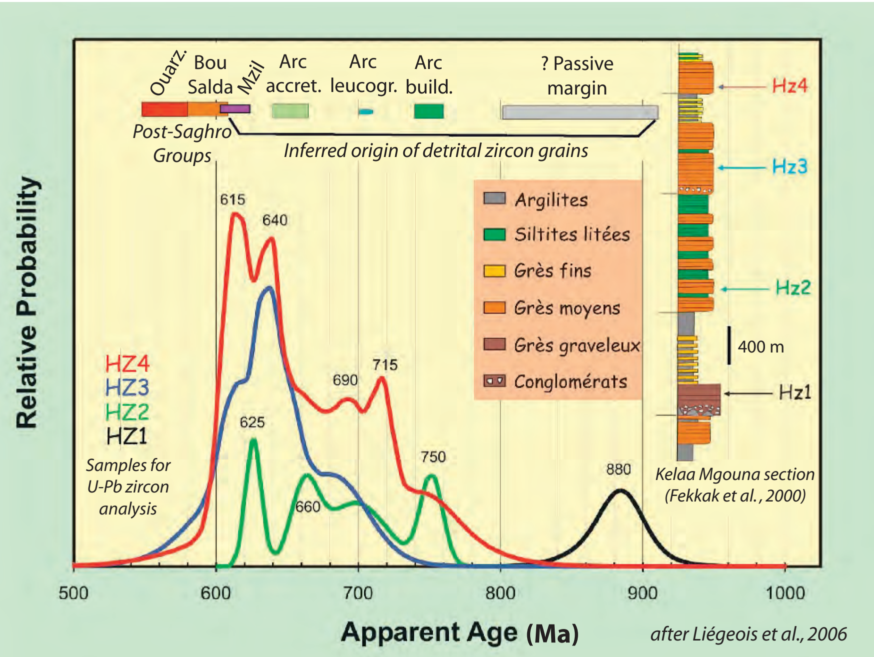

Fig. 2.15 Dating the Saghro turbidites through their detrital zircons (Liégeois et al., 2006). Four samples (HZ1–4) were acquired from bottom to top of the Kelaa Mgouna section (*right*). The dates show that there is a mixing of Paleoproterozoic (2000–2050 Ma) and Neoproterozoic zircons, with a progressive increase of the proportion of Neoproterozoic zircons from the base (10%) to the top (67%). The Neoproterozoic zircon ages correspond to different events known in the Anti-Atlas: (1) the Ifzwane doleritic suite (c. 880 Ma) from the WAC passive margin; (2) the ophiolitic sequence (750, 700 and 660 Ma) and (3) younger intrusions in the ophiolite/arc units and the Saghro Group itself with ages between 630 and 612 Ma (e.g. Mzil granite, c. 614 Ma). The increasing abundance and the younging of the Neoproterozoic zircons indicate that the deposition of the Saghro group occurred during the onset and growing of an Ediacaran volcanism (626–612 Ma) that evolved toward the volcanism of the Bou Salda Group (which overlie the Saghro Group), finally resulting in the huge volume of volcanic rocks and granitoids belonging to the Ouarzazate Group (580–550 Ma)

youngest zircon become younger towards the top of the series (from 626 Ma to 612 Ma). This shows that the Saghro Group was deposited between 630 and 610 Ma during the uplift of the WAC and the beginning of the Pan-African magmatism that culminated during the next Ouarzazate period. The Mzil granite (614 ± 10 Ma) could be a syn-Saghro intrusion, although the age bracket allows this granite also to be contemporaneous with the Bou Salda Group. The pillow basalts present in the Saghro Group have a mean Nd depleted mantle model age (TDM) of 650 ± 30 Ma, indicating a mantle origin without old crust contribution. The diamictite levels from the lower flysch-like formations of the Saghro Group should thus be correlated with the Marinoan ice age (650–630 Ma, late Cryogenian-earliest Ediacaran) widely observed in West Africa and Oman.

The tectonic deformation in the Saghro Group is dominantly shown by SSW- to SE-vergent folds; the metamorphism is low (greenschist facies). Detail tectonic studies are lacking, but in the present state of knowledge, a transpressive system along the northern margin of the WAC can be proposed as it has been demonstrated for the subsequent Bou Salda Group: this can account for the uplift of the WAC, the turbiditic nature of the Saghro group made at the base mainly of detritus from the WAC, and the progressive appearance of the magmatism without any crustal thickening allowing, for instance, the preservation of the Taghdout Group.

2.5.2 The Bou Salda Group and Equivalents

The Bou Salda Group of the Siroua massif rapidly follows the Saghro group: the youngest detrital zircon in Saghro is c. 612 Ma ($206Pb^*/238U$, laser ICP-MS, discordance 0–5%) and two rhyolite sills attributed to Bou Salda have been dated by SHRIMP ages at 606 ± 5 Ma and 606 ± 9 Ma ($206Pb^*/238U$ ages on slightly discordant zircons). We must note however that these two rhyolites are intrusive in the uppermost part of the Saghro Group. They could thus date the end of the Saghro Group deposition. Anyhow, Bou Salda deposition occurred after the folding of the Saghro Group. This group, for many aspects, is intermediate between the Saghro and Ouarzazate Groups. Its contacts with other lithologies are either tectonic or intrusive; it is preserved in narrow fault-bounded troughs. Its thickness is highly variable from a few tens of metres to more than 3000 m in the AAMF region. Its lower part comprises amygdaloidal basalt with subordinate andesite and rhyolite and its upper part is mainly sedimentary with poorly sorted conglomerates (boulders up to 2 m in diameter), arkoses and sandstones with interbedded greywackes, shales, tuffs, basalts and cherts. The Bou Salda Group was deposited during a dextral transpressive period. It is especially deformed along the AAMF (spaced cleavage in the shales, folding in the sandstones).

This Group exists throughout the Anti-Atlas (former “PII-III”). A typical example is that of the “Série d’Anezi”, a 2000 m thick volcanic and alluvial fill preserved in a pull-apart basin within the Kerdous inlier (Fig. 2.16). The “Série d’Anezi”

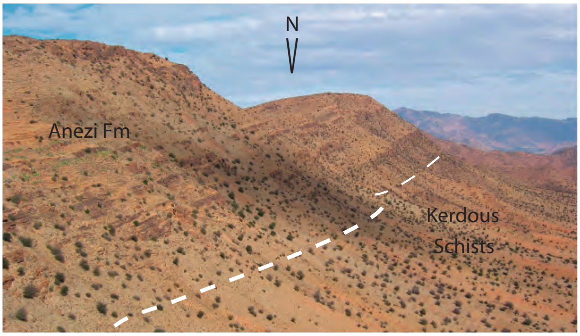

Fig. 2.16 View on the eastern edge of the Anezi basin, central Kerdous massif (see Fig. 2.6 for location). Conglomerates and sandstones of the lowest Anezi member (c. 600 Ma) dip about 30° WNW. They overlie the less resistant, retromorphic Kerdous schists (c. 2 Ga). Acidic tuffites are located at the very base of the Anezi Fm in the foreground (not seen), and rapidly vanish northward

can be subdivided into the basal "Tafraoute Group", dominated by dacitic-rhyolitic ignimbrites, and the "Anezi Group", dominated by sandstones and conglomerates. So far, the age of the Anezi sequence is not precisely known. The periglacial features recorded in the Anezi deposits should correlate with the youngest Gondwana glacial events ("Gaskiers", 590–580 Ma). Most of the basin remained undeformed, except at its margins. Other western Anti-Atlas inliers (Igherm, Aït Abdallah) display proximal diamictite facies with coarse quartzitic breccias and huge quartzite lenses which could originate either from former inselbergs or chaotic collapses.

The Tiddiline series from the Bou Azzer inlier (Fig. 2.9) consists of coarsening upward siltstone-sandstone-conglomerate sequences which contain diamictites interpreted as marine tilloids with dropstones. They could be compared with the Anezi series, but this is controversial because no reliable age is currently available on these series. The fact that the Tiddiline series is more strongly tilted, faulted and folded than Anezi, with local development of axial-plane cleavage can be attributed to its localization along the AAMF, or to an older age, intermediate between that of the Saghro and Bou Salda Groups. The 580 Ma-old (U-Pb zircon) Bleida granodiorite which cross-cuts the regional fabric provides a firm constraint on the latest stage of transpressive brittle movements in the Bou Azzer inlier (Inglis et al., 2004).

The earliest high-K calc-alkaline intrusion is the Mzil granite in the Siroua massif dated at 614 ± 10 Ma. Similar ages have been measured at Ifni from both granite intrusion and trachytic flow. However, most of the high-K calc-alkaline intrusions yield U-Pb ages close to 580 Ma (ex: Amlouggi tonalite, 586 ± 8 Ma; Askaoun granodiorite, 575 ± 8 Ma; Bleida granodiorite, 579 ± 1 Ma), but can be also younger: Imourkhsane granite, 562 ± 5 Ma, Tazoult quartz porphyry, 559 ± 6 Ma, and appear to be coeval with the accumulation of the Ouarzazate Group.

2.5.3 The Ouarzazate Group

The unconformable “Ouarzazate Group” (formely “PIII”) represents a volcano-sedimentary sequence highly variable in thickness (from 0 to 2 km near the Ouarzazate town) consisting of coarse volcanic conglomerates (Fig. 2.17A), ignimbritic rhyolites, trachytes, andesites, basaltic trachyandesites, tuffites, and rare interbedded stromatolitic layers and fault scarp breccias.

Various types of intrusions, such as granitoid massifs, necks, dykes or ring dykes emplaced within the early Ouarzazate Group or underlying units. In the Zenaga inlier, the Sidi El Houssein alkaline granite (579 ± 7 Ma) is a typical example of ring-complex intrusion within the Eburnian basement (Fig. 2.5). All the Ouarzazate plutonic and volcanic rocks belong to a high-K calc-alkaline to alkaline magmatic series (Fig. 2.18). Rhyolites from the Ouarzazate Group have been dated at 577 ± 6 and 571 ± 8 Ma. In the Kerdous massif (Fig. 2.6), the Tarçouate granodiorite laccolith (2.17B) yielded an U-Pb age of 581 ± 11 Ma, whereas the Tafraoute alkaline granite and the equivalent Tazoult pluton yielded U-Pb ages at 549 ± 6 and 548 ± 11 Ma, respectively. The Bou Maadine rhyolitic dome (Ougnat inlier) and the Tachkakacht rhyolitic dyke (Saghro-Imiter) also rank among the youngest, alkaline magmatic events of the Ouarzazate Group, being dated (U-Pb zircon) at 552 ± 5 Ma at 543 ± 9 Ma, respectively. The Ouarzazate Group has not recorded the Pan-African deformation, but was deposited on a highly variable basement topography, which, coupled with the big and rapid variations in thickness of the Ouarzazate Group itself, strongly suggests that this group was deposited during active tectonics, most probably transtensional movements.

This huge late Neoproterozoic (Ediacaran) magmatic event coupled with a strong extensional-transtensional tectonics is linked to an intense hydrothermal activity as exemplified by the main deposits of Imiter and Zgounder (Ag-Hg), Bou Azzer (Co-Ni-As-Ag-Au), Iourirn (Au), Bou Madine (Cu-Pb-Zn-Au-Ag) (Fig. 2.19).

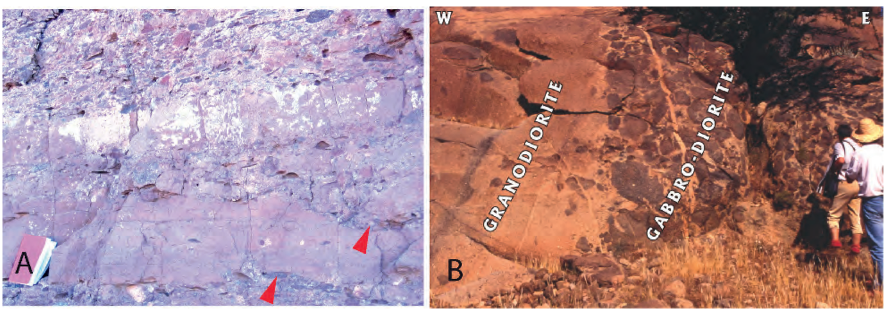

Fig. 2.17 Late Neoproterozoic rocks. (**A**): Ouarzazate volcanoclastic deposits in eastern Siroua inlier; note the uneven surfaces of stratification of the agglomerate layers, whose sub-angular elements are andesites and rhyolites. The notebook is 15 × 20 cm – (**B**): Core of the Tarçouate laccolith (c. 580 Ma) from the central Kerdous inlier (see Fig. 2.6 for location). The steep dip of the modally layered hornblende granodiorite with high amount of monzodioritic enclaves results from tilting above solidus conditions, as suggested by sub-vertical aplite dykes cutting across the igneous layering (Pons et al., 2006). Photo D.G.

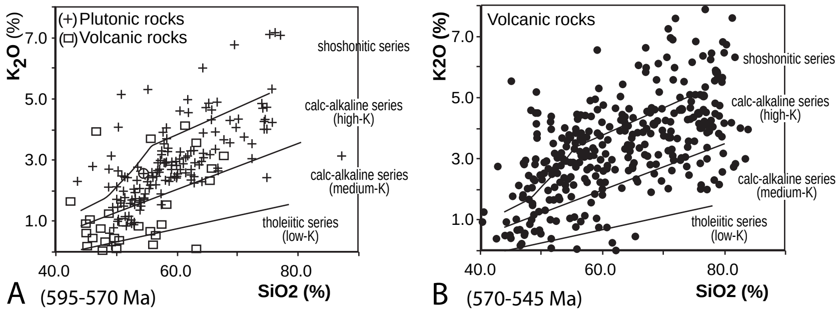

Fig. 2.18 K_2O–SiO_2 plots of Ediacaran calc-alkaline to alkaline plutonic (**A**) and volcanic (**A**, **B**) rocks of the Anti-Atlas belt. The 595–570 Ma magmatism (**A**) mostly show meta-aluminous magmatic rocks (50% < SiO_2 < 75%), with high-K calc-alkaline affinities. The 570–545 Ma volcanism (**B**) is dominantly effusive (mainly andesitic at the *bottom* of the sequence, rhyolitic-ignimbritic on *top*), and belongs to high-K calc-alkaline to shoshonitic series. After Gasquet et al. (2005)

2.5.4 Post-Orogenic Formations

References: For the location of the Precambrian/Cambrian boundary in the Anti-Atlas sedimentary record, see Latham & Riding (1990) and Maloof et al. (2005). Geochronological data for the Early Cambrian volcanites are given by Ducrot & Lancelot (1977), Gasquet et al. (2005), Maloof et al. (2005), and Álvaro et al.

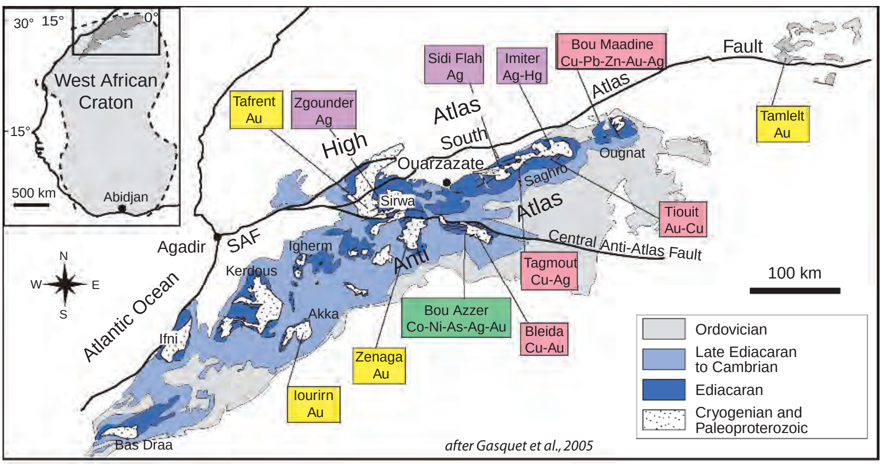

Fig. 2.19 Location of the main ore deposits in the Anti-Atlas belt. The calc-alkaline magmatism of the Ediacaran, late Pan-African metacratonic period was associated with large-scale base metal and gold mineralization. Metallogenic activity was greatest during the final extensional stage, at the Precambrian-Cambrian boundary. It is characterized by world-class precious metal deposits, base-metal porphyry and SEDEX-type occurrences (Gasquet et al., 2005)

(2006). The permanent role of extensional tectonics during that period is emphasized by Soulaimani et al. (2003), Soulaimani & Piqué (2004), Gasquet et al. (2005).

With the quasi-cessation of volcanism and the onset of thermal subsidence, the rifted and eroded Pan-African belt was flooded. This occurred before the end of Neoproterozoic times: top levels of the conglomeratic Adoudounian "Série de base" include carbonate deposits with some Ediacaran-type soft-bodied fauna. Then the main, carbonate part of the Adoudounian Fm straddles the Ediacaran-Cambrian boundary (542 Ma), as Tommotian-type calcified Cyanobacteria occur within the limestone layers of the overlying "Lie-de-vin" Fm. Detailed study of the δ13C record and its comparison with palaeontologically calibrated Siberian sections suggests that the Ediacaran-Cambrian boundary is located close to the base of the Adoudounian carbonates where the δ13C excursion to −6% is observed (Maloof et al., 2005). In the upper Adoudounian carbonates, a +7% high corresponds to the Nemakit-Daldyn/Tommotian boundary (ca. 525 Ma).

The Adoudounian basal conglomerates are virtually concordant on the clastic-ignimbritic Ouarzazate Group although in places a weak disconformity can be observed (cf. see Chap. 3). Weak extension tectonics is recorded in the Adoudounian carbonates by synsedimentary faults and slump structures, and by alkaline flows and sills around the Bou Azzer inlier. The J. Boho volcano at the southern border of the inlier is stratigraphically dated as Early Cambrian as its lowest flows and tuffs are interleaved in the uppermost Adoudounian levels (Fig. 2.9), consistent with an U-Pb age at 529 ± 3 Ma. The Aghbar trachytic sill at the northern border of the same inlier is dated at 531 ± 5 Ma. Ash beds from the upper Adoudou and upper Lie-de-vin Fms have been dated at 525 ± 0.5 and 522 ± 2, respectively.

2.6 Plate Tectonic Interpretation

References: The geodynamical interpretation of the Anti-Atlas Pan-African belt is discussed in Saquaque et al. (1989), Ouguir et al. (1996), Hefferan et al. (2000), Ouazzani et al. (2001), Ennih & Liégeois (2001, 2003, 2008), Thomas et al. (2002, 2004), Inglis et al. (2004), Gasquet et al. (2005), Ennih et al. (2006), Liégeois et al. (2006), Bousquet et al. (2008). The West African framework is described in Fabre (2005), Liégeois et al. (2003, 2005), Deynoux et al. (2006).

Varied plate tectonic interpretations of the Pan-African orogeny in Morocco have been proposed for decades and none was able to reach a consensus. In particular, the origin of the oceanic basin(s) and island arc(s), and the dip of the subduction(s) responsible for the ocean closure have varied considerably. However, the various data obtained in the last few years bring important new constraints limiting the possibilities. Note that the tectonic scenario elaborated for the Anti-Atlas Pan-African belt should correlate with those for the neighbouring segments of the belt, and especially with the most important Hoggar-Iforas belt.

2.6.1 The Three Main Neoproterozoic Epochs

In the Anti-Atlas, the Pan-African orogeny is the result of the convergence of the WAC with other terranes, which is also the case in the Tuareg shield, to the east of the WAC. Some correlations will then be proposed with that area for the convergent period.

First, the new geochronological data indicate three main Neoproterozoic epochs

(1) the deposition of the WAC passive margin sediments, probably beginning at ~ 1000 Ma and followed by the intrusion of doleritic dykes and sills at probably ca. 880 Ma. The Anti-Atlas Taghdout-Lkest Group correlates with the Char and Atar-Hank Groups of NW Taoudeni Basin. Shale horizons from the Atar-Hank Groups yielded Rb-Sr and K-Ar ages between 998 and 775 Ma. This passive margin evolution is a pre-Pan-African event to be linked with continental break-up; there is no information concerning the rest of the continent from which the WAC separated.

(2) the Bou Azzer – Siroua island arc building (760–700 Ma; Cryogenian) with its accretion towards the WAC at c. 665 Ma. This island arc complex is limited in surface, forming a strip along the AAMF. It must be noted that there are no other rock types in the Anti-Atlas within that age range. This event could be called early Pan-African accretion stage. It is accompanied and followed by the earliest greenschist deformation of the foreland between 660 and 640 Ma. This evolution can be correlated with the formation of volcanic arcs around 730–720 Ma in the Hoggar transect, with subsequent deposition of the Green Series, and collision with the WAC from 630 to 580 Ma. The Pan-African I phase is dated at 665–655 Ma in the Bassaride belt (Senegal, Guinea) on the opposite side of the WAC.

(3) the metacratonic Pan-African stage (630–550 Ma; Ediacaran) which began with the Saghro Group (630–610 Ma (up to 600 Ma?), consisting mainly of turbiditic sediments but recording an increasing amount of magmatism, tholeiitic at the start, high-K calc-alkaline at the end. After a minor tectonic phase, the Bou Salda Group (610 or 600–580 Ma) is characterized by badly sorted sediments and abundant high-K calc-alkaline volcanism and plutonism; the final phase is the Ouarzazate Group (580–550 Ma) composed mainly of high-K calc-alkaline rhyolites and granites. These three groups have in common: high variability in thickness, from 0 to several thousands of metres; deposition upon unstable basement with alternation of poorly sorted sediments, high energy sediments and volcanic rocks; finally, there is an increase of the volcanic rock proportion with time. The contemporaneous tectonics regimes were transcurrent, transpressive during the Saghro and Bou Salda Groups, and transtensive during the Ouarzazate Group. Periglacial deposits occur at least in the Saghro, Tiddiline and Anezi (Bou Salda) Groups. The Ouarzazate Group could correlate with the very base of the post-orogenic Série Pourprée of the Iforas-Hoggar transect, which is mostly dated between 560 and 530 Ma.

2.6.2 The WAC Northern Extent and Metacratonic Evolution

The geodynamic behaviours of SW and NE Anti-Atlas are similar. Ediacaran high-K calc-alkaline rhyolites and granites have similar ages and the same Sr-Nd isotopic ratios, indicating the presence at depth of the Eburnian basement even if it is outcropping only to the SW of the AAMF. The AAMF is an important feature, having allowed the Cryogenian oceanic complex to be preserved, but it does not mark the northern boundary of the WAC.

The Anti-Atlas Eburnian basement has not been significantly thickened by the Pan-African tectonics, but only underplated and variably uplifted: the Pan-African effects are limited to greenschist facies metamorphism and mylonitisation accompanied by major amount of fluids, probably especially during the Ouarzazate Group emplacement. The northern WAC border was never transformed to an active margin. In other words, no subduction plane with Franciscan-type metamorphism can be placed below the Anti-Atlas. The northern boundary of the WAC has been dissected by major faults and shear zones which allowed the generation and emplacement of the voluminous Ediacaran magmatism, but has preserved most of its cratonic rigidity: it acted as a rigid or semi-rigid indenter for the peri-Gondawanan terranes during the Pan-African and the Variscan orogenies and currently during the Alpine orogeny. Such a behaviour can be called a "metacratonic evolution", i.e. an evolution after the cratonic stage marked by a partial destabilization of the cratonic area (here the northern margin of the WAC) but preserving a large part of the cratonic rigidity and behaviour. This metacratonic rigidity explains why the early pre-Ediacaran Pan-African structures have been preserved, not only the Eburnian basement but also the c. 900–800 Ma passive margin sediments displaying easily destroyed sedimentary features such as ripple marks or mud cracks, and the thrust Cryogenian oceanic complex.

2.6.3 Location of the Pan-African Mobile Belt

If the Anti-Atlas represents the northern boundary of the WAC metacratonized during the Pan-African orogeny, the true Pan-African mobile belt existed just to the north and west, consisting mostly of peri-Gondwanan terranes which drifted away later, during the Phanerozoic. Complementary information such as the cause of the deformation of the Saghro Group at 610–600 Ma should be searched for within these peri-Gondwanan terranes. Other correlations can be made on the eastern side of the WAC, in the Tuareg shield where the Pan-African terranes that collided with the craton are preserved. It is interesting to note that in the Tuareg shield: (1) several island arc dated at c. 700–750 Ma are known, (2) the collision with the WAC began at c. 630 Ma, which is also the onset of the Saghro Group deposition, (3) the large transcurrent transpressive movements along the mega-shear zones ended mostly at about 580 Ma, passing to transtensional movements accompanied by high-level plutons until c. 560 Ma, corresponding to the Ouarzazate period and (4) that the latest magmatism before the Phanerozoic sedimentation is dated at 535–520 Ma (Taourirt province) contemporaneous to the late alkaline magmatism in the Anti-Atlas. This means that the stress at the origin of the Pan-African orogeny in West Africa can be ascribed to the WAC, as the current stress in Asia can be ascribed to the Indian craton. The collision between the WAC and the Tuareg shield was slightly oblique; in the Anti-Atlas, the Ediacaran Pan-African phase corresponded mainly to a sliding movement, passing progressively from transpression to transtension, metacratonizing the northern boundary of the WAC. This metacratonic reactivation appears to strongly favour magma and fluid generation and movements, mainly during the Pan-African orogeny and subsequently during the early Phanerozoic, including the current uplift and volcanism, eventually leading to the well-known Anti-Atlas mineralizations.

2.6.4 Tectonic Scenario

A relatively simple scenario has been proposed by several authors (including one of us, D.G.) with some variations: it hypothesizes a south-dipping subduction zone operating from ~ 750 Ma to ~ 600 Ma. The protracted subduction would have been

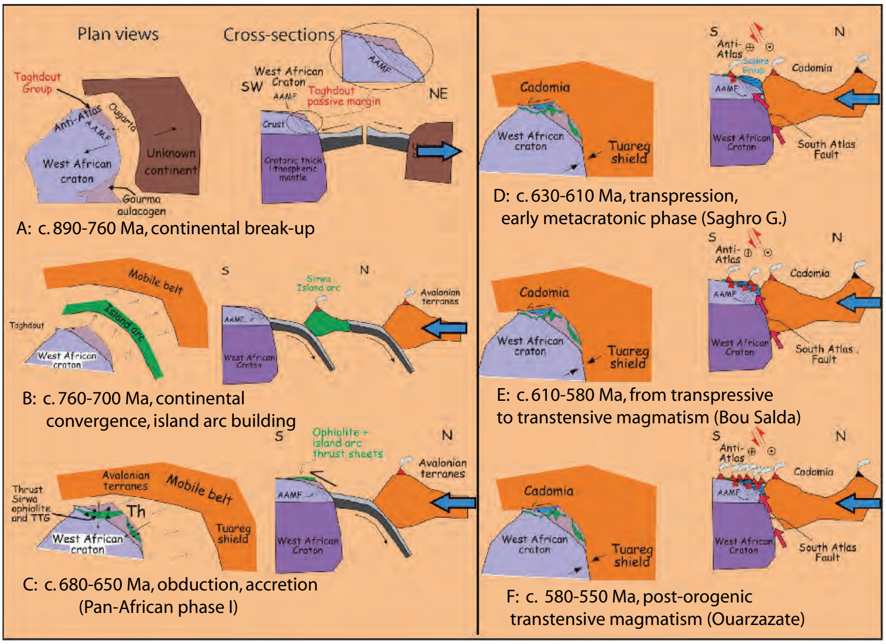

Fig. 2.20 Geodynamic evolution of the Anti-Atlas domain during Neoproterozoic times: a new scenario for the Pan-African mountain building around the West African Craton, after Liégeois et al. (2006)

responsible for the building of an island arc at some distance of the WAC margin, the Bou Azzer ophiolites being formed in a back-arc basin. Thus, this setting would compare with that of the western Pacific margins. Following the obduction and arc collision, the Ediacaran episode of high-K calc-alkaline to alkaline magmatism is interpreted as the result of supra-subduction partial melting, slab break off and post-orogenic extension. However, this model is not supported by the new data reported above (age of the Saghro Group, northern extent of the metacratonic basement, lack of HP-LT metamorphism).

In contrast, our tectonic scenario (Fig. 2.20) assumes a north-dipping subduction from ~ 760 to 660 Ma. The future ophiolite corresponds either to the oceanic part of the WAC plate or to a fore-arc basin, its obduction and the arc accretion (Pan-African phase I) results from the oceanic lithosphere consumption due to Avalonia/Cadomia convergence towards the WAC (a model similar to the Tethyan setting in Oman). Between 660 and 630 Ma, the earliest quartz-diorite magmatism could result from slab break-off. The metacratonic evolution begins with proximal

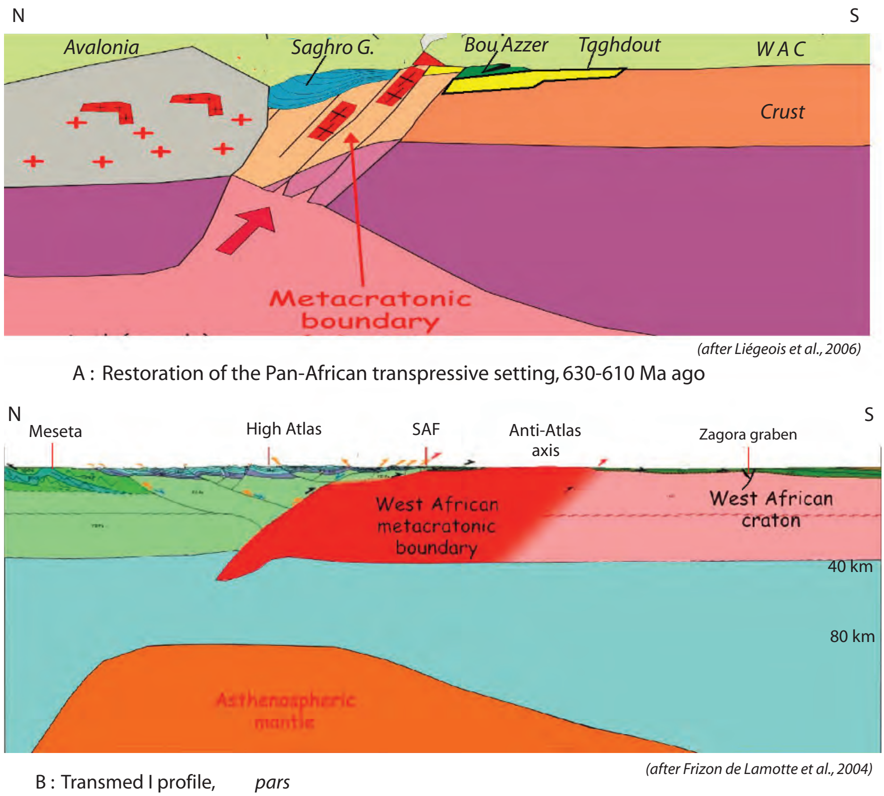

Fig. 2.21 Hypothetic structure of the Anti-Atlas lithosphere by the time of the Saghro Group sedimentation and coeval magmatism (**A**), compared with its present-day structure (**B**). In both case, mantle lithosphere is thinned along the transpressive margin of the metacratonic fringe of the WAC, after Liégeois et al. (2006). See Fig. 2.1 for location

volcanoclastic turbidites (the newly dated Saghro Group) coeval with high-K calc-alkaline magmatism associated with transpressive deformation of the WAC margin. The asthenospheric uplift beneath the transcurrent margin can be compared with that presently observed beneath the San Andreas transform fault, and more simply, in the present-day asthenospheric uplift which follows the Central Atlantic-Atlas transcurrent rift system (Fig. 2.21). The tectonic regime progressively changes from transpressive to transtensive between 610 and 580 Ma, and thus the magmatism becomes more alkaline. This tendency culminates during the Adoudounian-Early Cambrian.

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Chapter 3
The Variscan Belt

A. Michard, C. Hoepffner, A. Soulaimani and L. Baidder

This chapter is a tribute to Jacques Destombes and Solange Willefert, and to the memory of Henri Hollard, whose names are associated to the deciphering of the Paleozoic stratigraphy of Morocco. We want to also evoke the name of Martin Burkhard who renewed our understanding of the Anti-Atlas tectonics.

In the western Maghreb, the Variscan (Hercynian) belt extends into the Meseta and Atlas domains, being widely exposed in the large Paleozoic massifs of the Moroccan Western Meseta and Western High Atlas, whereas it forms smaller massifs in the Eastern Meseta, Middle Atlas and Central-Eastern High Atlas (cf. Chap. 1, Fig. 1.11). Altogether, these massifs define the *Meseta Domain*, whose late Paleozoic evolution was accompanied by significant metamorphism and magmatic intrusions. This domain is also referred to as the Meseta Block (although it was not a single block until the end of the Variscan Orogeny) and corresponds to the south-westernmost segment of the Variscan belt of Europe.

Another strongly deformed Variscan segment is present within the *Adrar Souttouf* and *Dhlou* (Ouled Dehlim) massifs of south-western Morocco. These massifs are entirely distinct from those of the Meseta Domain, both geographically and structurally, and correspond to the northern tip of the Mauritanide belt.

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The *Anti-Atlas belt* extends south of the Meseta Domain and east of the hypothetic northern continuation of the Mauritanides beneath the Tarfaya coastal basin (Fig. 1.11). The Anti-Atlas Paleozoic series were also affected by the late Paleozoic orogeny, although more weakly than those of the neighbouring Variscan segments. Thus, the Anti-Atlas can be regarded as the common foreland fold belt of both the Mesetan Variscides and northernmost Mauritanides. The Anti-Atlas belt connects to the east with the coeval, intracontinental *Ougarta belt* that extends essentially into the Algerian territory.

Finally, Variscan units are also exposed in the Alpine nappes of the *Internal Rif*. These nappes formed at the expense of an allochthonous terrane (Alboran Domain), whose pre-Alpine relationships with the Meseta Domain are debatable. They are described in the chapter of this book dedicated to the Rif Orogen (Chap. 5).

3.1 The Anti-Atlas Fold Belt

3.1.1 An Ancient Fold Belt Exposed in a Young Mountain Range

References: In addition to the classical data presented in Choubert (1952), Michard (1976), Piqué and Michard (1989), and Fabre (2005), this introduction is based mainly on the recent works by Frizon de Lamotte et al. (2000), Missenard et al. (2006), Robert-Charrue (2006), Gutzmer et al. (2006), and Malusà et al. (2007), who emphasize the importance of the Mesozoic-Cenozoic events. Concerning the Triassic intrusions, the general framework is given in Chap. 1 (this volume), whereas the problem of the "pseudo-folded" sills (described by Hollard, 1973) is discussed by Smith et al. (2006) on examples from SW Algeria. See also Chap. 7, this volume.

The Anti-Atlas mountain range parallels the High Atlas at the northern border of the Saharan platform (Fig. 3.1). Both mountain ranges are in direct contact in the Siroua-Ouzellarh area, whereas they are separated by narrow Neogene basins west (Souss Basin) and east (Ouarzazate and Errachidia-Boudenib Basins) of the Siroua plateau. The elevation of the Anti-Atlas axis is generally close to 2000–2500 m (up to 2700 m in the J. Saghro), except in the Siroua area, which is topped by a Neogene volcano culminating at 3300 m. Therefore, the Anti-Atlas can clearly be considered as a young mountain range, uplifted contemporaneously with the High Atlas as the result of both plate convergence and asthenosphere uplift (see Chap. 1, Sect. 1.5). Recent volcanism is widespread in the Siroua and Saghro massifs, as in the Middle Atlas and Missour Basin regions (see Chap. 4).

However, the Anti-Atlas is also a large Paleozoic fold belt, characterized by numerous inliers of Precambrian basement, the so-called "boutonnières" (see Chap. 2), cropping out all along the mountain range axis, which is, in fact, shifted toward the northern border of the fold belt. Each inlier corresponds to a more or less faulted antiform where Precambrian rocks are exposed due to post-Variscan erosion (cf. Chap. 2, Figs. 2.2 and 2.9). In the following account we describe, first, the

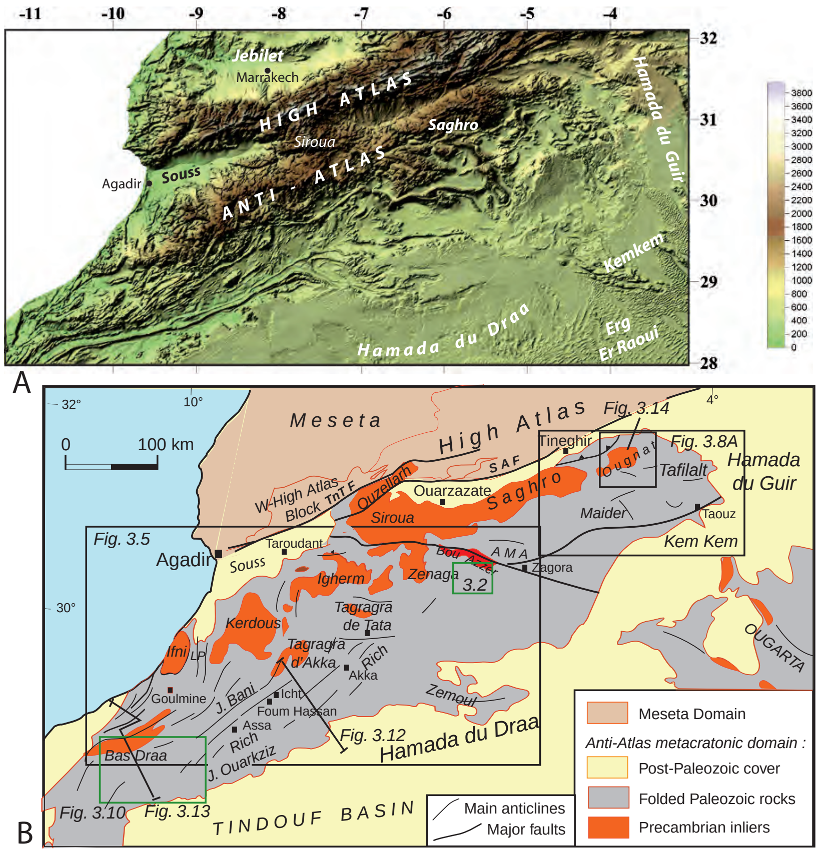

Fig. 3.1 The Anti-Atlas, a young mountain range superimposed on a Paleozoic fold belt. – *Top*, topographic model (courtesy Y. Missenard). The Anti-Atlas parallels the High Atlas, with slightly lower elevation. The two chains converge on the J. Siroua (Sirwa) transect. – *Bottom*, Schematic map of the Anti-Atlas domain showing the Precambrian antiforms ("boutonnières"), the folded Paleozoic cover, with location of some of the following figures (framed). AMA: Anti-Atlas Major Fault ("Accident Majeur de l'Anti-Atlas"); LP: Lakhssas Plateau; SAF: South Atlas Fault; TnTF: Tizi n'Test Fault

pre-orogenic, Cambrian to Early Carboniferous evolution of the Anti-Atlas domain, and second, its orogenic deformation during the Late Carboniferous-Permian.

Due to its location in the vicinity of the Central Atlantic rift (see Chap. 1), the Anti-Atlas belt was intruded by a huge quantity of dykes (Foum Zguid; Fig. 3.2) and sills of Triassic-Early Liassic gabbros and dolerites. Much later, subsequent to a long period of erosion, the former mountain belt was partly overlain by Cretaceous-Tertiary deposits that now form the Hamada Plateaus. The entire Anti-Atlas domain was finally uplifted during the Neogene, contemporaneously with the High Atlas itself.

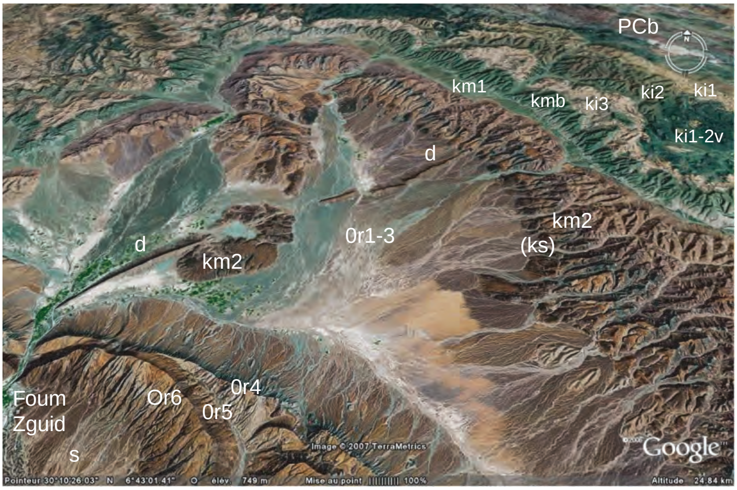

Fig. 3.2 The Foum Zguid gabbroic dyke (d), ca. 200 Ma-old, cutting across the folded Paleozoic sequence SW of the Bou Azzer-El Graara "boutonnière". *Google Earth* oblique view, tilted to the NW. Width of picture: 30 km, depth 40 km. PCb: Precambrian; ki1: Adoudounian = "Calcaires inférieurs" (Latest Neoproterozoic-Lower Cambrian); ki1-2v: Lower Cambrian volcano; ki2: "Lie-de-vin Serie"; ki3: "Calcaires supérieurs"; kmb: "Grès terminaux", now assigned to the base of the Middle Cambrian; km1: "Schistes à *Paradoxides*"; km2: Tabanit sandstones; (ks) discontinuous Upper Cambrian slates. Or1-3: Tremadocian-Llanvirnian (Fezzouata and Tachilla pelites); Or4: Llandeilian ("Premier Bani"); Or5: Caradocian (Ktaoua); Or6: Ashgillian (2nd Bani); Silurian (s). Legend after the 1/500,000 geological map, sheet Ouarzazate, and after Destombes & Feist (1987), Chbani et al. (1999) and, for the age of the "Grès terminaux" ("ki4" on the cited map), Geyer & Landing (2004) and Landing et al. (2006)

3.1.2 The Paleozoic Pre-orogenic Evolution

References: The geological maps of the Anti-Atlas, scale 1/200,000, are fundamental references for the study of this belt. The following sheets are of particular interest: Foum el Hassan-Assa (Choubert et al., 1969), Akka-Tata (Hollard, 1970), Ouarzazate-Alougoum (Choubert et al., 1970), Saghro-Dadès (Hindermeyer et al., 1974–1977), Tafilalt-Taouz (Destombes & Hollard, 1986) and Todgha-Ma'der (Du Dresnay et al., 1988). The Explanatory Notes concerning the Lower Paleozoic series of these maps are being published by the latter Service (Destombes, 2006a,b,c,d)

The post Pan-African sedimentary series of the Anti-Atlas begins with uppermost Neoproterozoic deposits, and then encompasses the whole Paleozoic from the Early Cambrian to the Late Carboniferous (Fig. 3.3). The entire sequence is about 8–9 km-thick in the Western Anti-Atlas, whereas it is only 4–5 km- thick south of the Saghro and Ougnat massifs of Eastern Anti-Atlas. Overall, the Anti-Atlas series were deposited in shallow marine environments. The almost continuous, richly fossiliferous sedimentary succession of the Anti-Atlas domain ranks among the best studied Paleozoic series worldwide.

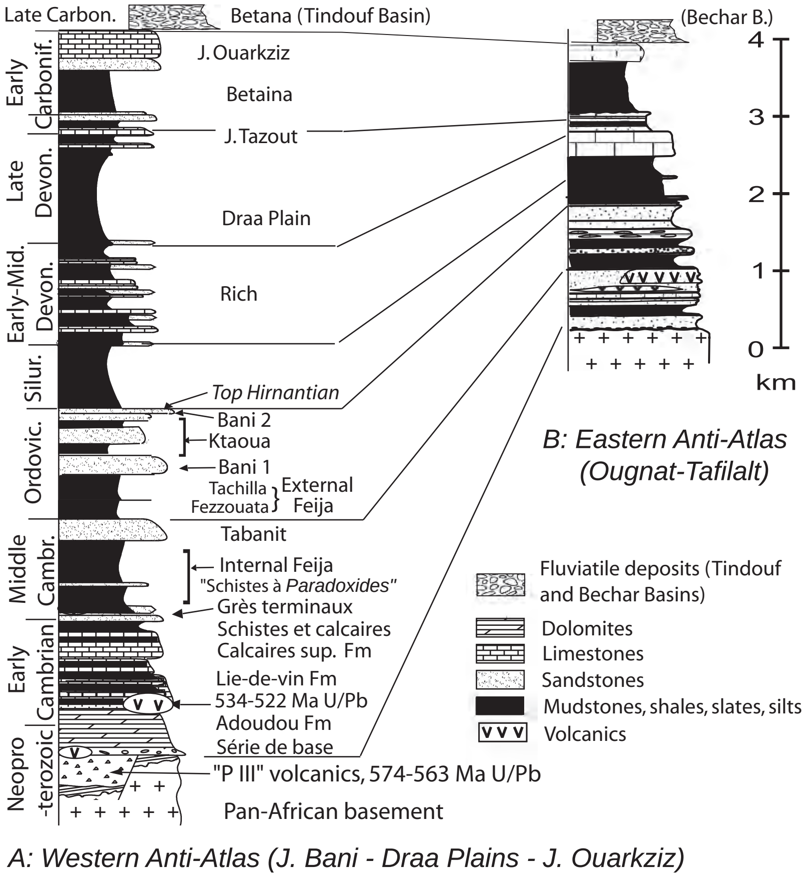

Fig. 3.3 Overall stratigraphy of the Paleozoic formations. (**A**) Western Anti-Atlas, after Helg et al. (2004), modified. Age of the “Grès terminaux” after Landing et al. (2006). (**B**) Eastern Anti-Atlas (Ougnat-Tafilalt), after Baidder et al. (2008), modified. The series are labelled according to their classical, informal names (Choubert, 1952). The formations shown in black correspond to possible décollement horizons during folding

3.1.2.1 Lower Paleozoic

References: The overall Lower Paleozoic stratigraphy of Morocco has been described by Destombes et al. (1985). More recent studies on the Cambrian and Ordovician series/events of the Anti-Atlas have been published by Destombes & Feist (1987), Latham & Riding (1990), Geyer & Landing (1995, 2004), Landing et al. (1998), Chbani et al. (1999), Algouti et al. (2000), Lüning et al. (2000), Gutiérrez-Marco et al. (2003), Benssaou & Hamoumi (2003, 2004), Álvaro et al. (2006a), Landing et al. (2006), El Maazouz & Hamoumi (2007), Álvaro et al. (2007), Pelleter et al. (2007). The Upper Ordovician glacial deposits of the Anti-Atlas and Ouzellarh Block have been described by Álvaro et al. (2004), Villas et al. (2006), Le Heron et al. (2007) and Le Heron (2007), whereas the post-glacial Silurian "hot shales" of North Africa have been reviewed by Lüning et al. (2000).

The Cambrian-Early Ordovician Rifting of the Metacratonic Margin

In the Western Anti-Atlas, the late Neoproterozoic Ouarzazate Group (former "P III"; cf. Chap. 2) is overlain unconformably by a syn-rift sequence labelled "Série de base", the thickness of which is highly variable. This basal series displays, from bottom to top, coarse conglomerates (locally including huge quartzite blocks), stromatolitic dolomites, and siltstones (Fig. 3.4A). Basaltic flows (continental tholeites) are exposed locally at the very bottom of the sequence, on top of the Kerdous inlier. The "Série de base" is followed by the Tata Group trilogy: stromatolitic dolomites ("Calcaires inférieurs" = Adoudounian or Oued Adoudou Formation), reddish calcareous pelitic series ("Série lie-de-vin" = Taliwine Formation), and archaeocyathid-bearing limestones ("Calcaires supérieurs" = Igoudine Formation). The Neoproterozoic-Cambrian limit (542 Ma, according to the ICS geological time scale) is located within the Adoudou dolomites, at an unknown level because of the absence of any characteristic fossil. The uppermost Adoudou Fm, the Lie-de-vin Fm, and the lower part of the Calcaires supérieurs are dated from the Tommotian and characterized by stromatolites (microbialites) typical for restricted, shallow water environment. The upper part of the Calcaires supérieurs and the overlying "Série schisto-calcaire" (Amouslek Fm) are ascribed to the Atdabanian, and characterized by deeper, open-sea conditions promoting archaeocyathid growth and trilobite immigration (Landing et al., 1998; Álvaro et al., 2006a,b). Lava flows on top of the Calcaires inférieurs

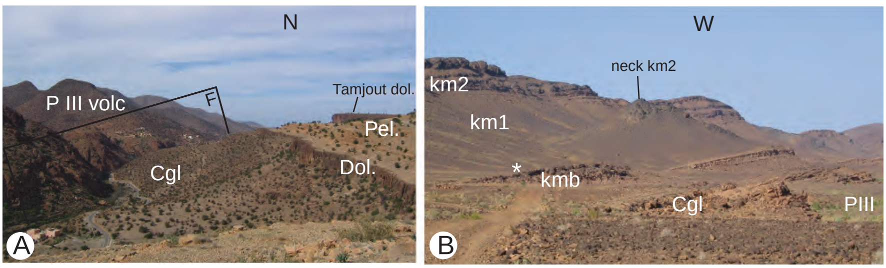

Fig. 3.4 The base of the sedimentary series in the Western (**A**) and Eastern (**B**) Anti-Atlas. – (**A**) North-eastern border of the Kerdous boutonnière, Oued n'Ait Baha valley. "PIII": massive volcanites from the Ouarzazate Group (Late Neoproterozoic). A normal fault limits the syn-rift formations of the Adoudounian "Serie de base" (Latest Neoproterozoic): conglomerates (Congl.), stromatolitic dolomites (Dol), pelites (Pel), beneath the Tamjout dolomite (Adoudounian, Neoproterozoic-Early Cambrian). – (**B**) Southern edge of the Ougnat massif; the Lower Cambrian is incomplete and purely detrital. The onlap on the Neoproterozoic volcanites ("P III") begins with conglomerates and red sandstones (Cgl, assigned to ki2, Serie Lie-de-vin), followed by a thin "Grès terminaux" formation (kmb, base of Middle Cambrian, cf. Landing et al., 2006). Asterisk: "*Micmacca* breccia" (fragmented trilobites); km1: "Schistes à *Paradoxides*"; km2: Tabanit sandstones. The Middle Cambrian neck of andesitic basalt intruding the "km1" pelites feeds a lava flow laterally (out of the picture)

(J. Boho), and tuffites from the Lie-de-vin series have been dated at 534 ± 10 and 522 ± 2 Ma (U/Pb), respectively (see Chap. 2).

The rifting process responsible for the abrupt thickness variations in the “Série de base” was already active during the accumulation of the late Neoproterozoic Ouarzazate Group. Extension continued during the “Adoudounian”-Early Cambrian, allowing the sea to flood the rifted margin of the metacraton as defined in Chap. 1. The marine onlap was favoured by the global high sea level, due to melting of the late Neoproterozoic (Cryogenian) Gondwana polar cap. In the Western Anti-Atlas, the Adoudou Fm. isopach pattern suggests a graben-like depocenter (Fig. 3.5). Likewise, criteria such as slump folds, synsedimentary breccias and normal paleofaults yield evidence for syn-rift sedimentation. The Early Cambrian rifting stage is accompanied by within-plate alkaline volcanism (e.g. J. Boho in the upper part of the Adoudou dolomites). The dominantly fine-grained, calcareous sequence is topped by more homogeneous, post-rift sandy deposits, formerly referred to as the Early Cambrian “Grès terminaux”. They were recently dated from the basal Middle Cambrian (Geyer & Landing, 1995, 2004), and thus they should be now labelled “Grès initiaux”! In the easternmost Anti-Atlas, these sandy facies represent nearly the entire Early Cambrian sequence (Fig. 3.4B), and they disappear north of the Ougnat inlier.

Rifting tectonics resumed during the Middle Cambrian along the northern border of the Anti-Atlas. The Middle Cambrian sedimentation began with the silts and greywackes of the Internal Feija Group (whose name refer to the corresponding circulation corridor = “feija” between the rugged Early Cambrian hills and the cuesta of the upper Middle Cambrian sandstones). This lower group, also referred to as the “Schistes à *Paradoxides*”, overlies unconformably the Precambrian basement

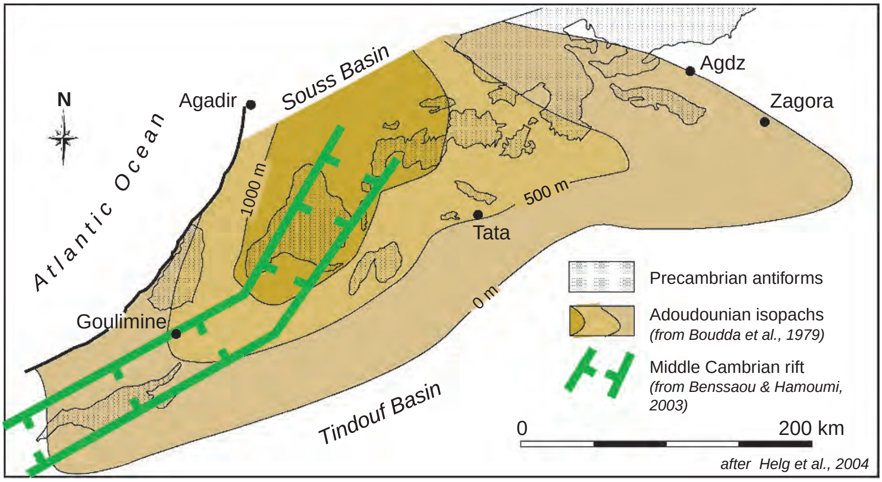

Fig. 3.5 Thickness distribution of the “Adoudounian” carbonates (latest Neoproterozoic-Early Cambrian), and orientation of the Middle Cambrian graben of western Anti-Atlas, after Helg et al. (2004), modified. The original map of the Adoudounian isopachs is from Boudda et al. (1979); the Middle Cambrian rift is after Benssaou & Hamoumi (2003)

along an E-trending paleogeographic high extending from the Marrakech High Atlas (Ouzellarh Block) to the Ougnat Massif (Destombes et al., 1985). This high can be interpreted as the rift shoulder south of the subsiding Meseta domain. The Internal Feija Group is succeeded by the Tabanit Group (= "Grès du Tabanit", or "*Conocoryphe* and Lingula Sandstones"). Coeval extensional tectonics is documented by the emplacement of calc-alkaline to tholeitic basalts or andesitic basalts in the western Marrakech High Atlas (Ounein) and Eastern Anti-Atlas (southeast of the Ougnat massif). Such volcanics are also exposed in the Meseta Domain itself (see below). Thus, the Meseta Domain was rifted away from the metacraton during Cambrian time. The width of the thinned crust/oceanic domain in between, and the geodynamic setting of the rifting stages (either a passive margin or an active margin/backarc extension) are a matter of debate (Sect. 3.3.4.2).

The Anti-Atlas domain was partly emergent during the Late Cambrian-early Tremadoc, as indicated by the frequent absence of Upper Cambrian deposits and by the shallow marine facies of the scarce preserved strata (offshore shales with sandstone shoals and channels, south of the Bou Azzer-El Graara Precambrian inlier, central Anti-Atlas; Fig. 3.2). This can be related to the rifting tectonics and concomitant uplift of the rift shoulder (Álvaro et al., 2007).

Ordovician Platform, Hirnantian Glaciation and Silurian Post-Glacial Onlap

The Tremadoc clays are developed in a sag basin extending south of the Saghro-Ougnat high (Fig. 3.6A,B). Afterward, from the Arenig until the Llandeilo, the Anti-Atlas and Ougarta Domain formed a N-dipping, gently subsiding shallow water platform, covered alternately by sands and silts sourced in the emergent Saharan Domain (either from the Precambrian shield or their Cambrian continental cover). Silt deposits dominated first (External Feija Group), then sandy deposits (First Bani Group), and then both facies alternated within the Ktaoua Group. Fossils are particularly abundant in the shaly facies, with trilobites, brachiopods, molluscs, echinoderms, graptolites, and microfossils.

The south shoreline of this shallow marine platform was located beneath the present-day Tindouf Basin, at some distance from the northern rim of the Reguibat Arch (e.g. Fig. 3.6B). The northeastern parts of the platform show thinner deposits than the central parts, where a W-trending depocenter is indicated by the isopach maps. In contrast, from the Caradoc onward, the isopach maps show a rotation of the depocenter axis in the Eastern Anti-Atlas toward a SE-NW trend (Fig. 3.6C). This points to the occurrence of significant synsedimentary tectonics, rather moderate during the Lower-Middle Ordovician (as exemplified by the intraformational unconformities in the "Premier Bani" sandstones; Fig. 3.7A), but more intense during the Caradoc, at least in the Eastern Anti-Atlas (Fig. 3.7B,C). It is worth noting that Middle-Upper Ordovician magmatic/hydrothermal activity has been described along the northern border of the Anti-Atlas, such as felsic dykes in the Siroua Massif (Huch, 1988) and hydrothermal metamorphism dated at c. 450 Ma (U-Pb individual zircon, ^{40}Ar-^{39}Ar individual phengite) in the Au-bearing albitites of the Tamlelt

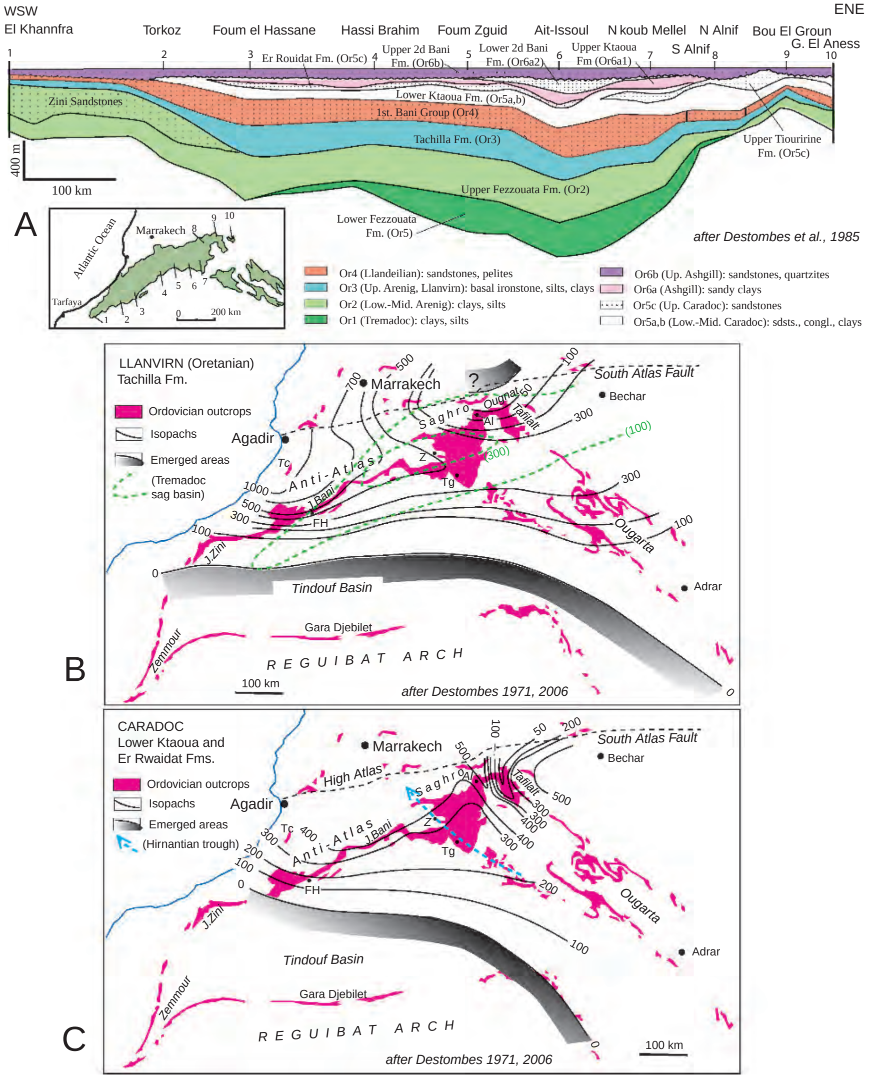

Fig. 3.6 The Ordovician series of the Anti-Atlas and Ougarta domains, after Destombes (1971) and Destombes et al. (1985), modified after Destombes (2006 a,b,c,d) and Destombes (in litt., 2007). – (**A**) Diagrammatic representation of the sedimentary formations and groups recognized in the Ordovician series showing their age and thickness variation along the WSW-ENE trend of the belt. – (**B**) Thickness distribution of the Tachilla Fm. (Llanvirn), with indication of the former Tremadoc sag basin (isopachs 100 and 300 m, *dashed*). – (**C**) Thickness distribution of the Lower Ktaoua and Er Rwaidat Fms. (Caradoc), with indication of the future Hirnantian trough (axis of the Lower Second Bani Fm., *dashed arrow*). Al: Alnif; FH: Foum-el-Hassan; FZ: Foum Zguit; Tc: Tachilla; Tg: Tagounit; Z: Zagora

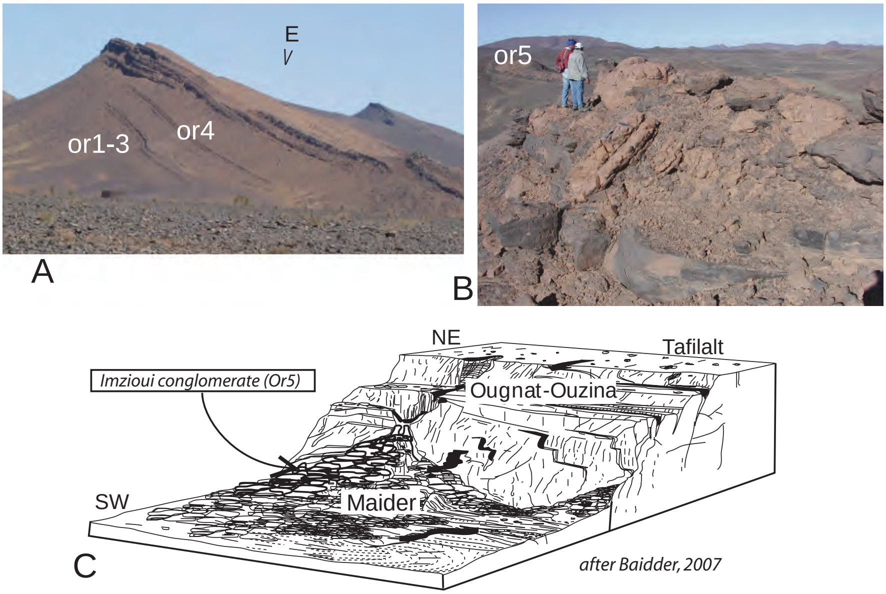

Fig. 3.7 Synsedimentary tectonics in the Ordovician series of the Eastern Anti-Atlas. – (**A**) Intraformational unconformity in the Premier Bani sandstones (Or4, Llandeilo) south of Oukhit (east of the Ougnat massif; for location, see Fig. 3.8). Or1-3: Fezzouata-Tachilla pelites and Fe-oolite (Tremadocian-Llanvirnian) – (**B**) Imzioui chaotic conglomerate (Or5, Caradocian) with blocks of quartzite, ferruginous oolite (Lower Ordovician), *Micmacca* breccia (Middle Cambrian) and limestone (Lower Cambrian). Photo L. Baidder. C: Interpretation of the Imzioui conglomerates as debris flows related to an active fault scarp striking NW-SE along the Ougnat-Ouzina axis, after Baidder (2007). The uplifted Tafilalt block east of the fault displays thin Caradocian sequences with Bryozoan limestones, in contrast to the thick clastic sequences of the Ougnat-Ouzina and Maider basin (Destombes, 2006 a,b,c,d; El Maazouz & Hamoumi, 2007)

(Tamlalt) Massif (Pelleter et al., 2007). This could correspond to a renewed phase of rifting, more clearly visible in the Meseta Domain (see below, Sect. 3.3.2.2). The thickness variations of the varied formations (Fig. 3.6A) depend both on these tectonic movements and on the current orientation and/or length of transport from the source areas. For example, the occurrence of a thick sandstone member in the Fezzouata Fm. (Arenig) of westernmost Anti-Atlas (J. Zini Sandstones, exact equivalent to the Armorican Sandstones from Brittany) is dependant on its proximal location with respect to a source area in the western Reguibat Arch.

The Upper Ordovician formations are characterized, not only in southern Morocco, but in all of North Africa from Libya to Mauritania, by glaciogenic deposits of particular economic importance as oil and gas reservoirs (e.g. Beuf et al., 1971; Ghienne & Deynoux, 1998; Le Heron, 2007, with references therein). In the Anti-Atlas, the Upper Ordovician (Ashgill, Hirnantian) series are labelled Second Bani Group (Or6; Fig. 3.6A). The Lower 2d Bani Fm. (Or6a) consists of preglacial sands and clays, up to 150 m thick in the axis of a NW-trending trough extending from about Tagounit to Zagora (Fig. 3.6C). The pre-glacial deposits are unconformably overlain by the Upper 2d Bani Fm. (Or6b), which consists of typical glaciogenic deposits. The unconformity at their base outlines a system of NW-trending paleovalleys, 0.5–1 km wide and up to 100 m deep, cut probably as tunnel valleys beneath the Late Ordovician ice sheet (Le Heron, 2007), and as paleofjords near Alnif (Destombes, 2006a). The glaciogenic sediments include tabular shallow marine sandstones, massive sandstones and conglomerates with exotic, crystalline boulders (ice contact debrites), meandriform sandstones (ice proximal sandur), stratified diamictites (ice rafted debris) and sigmoidal bedded sandstones (intertidal deposits). Synsedimentary deformations are ubiquitous and include soft-sediment striated pavements with NW- to N-trending striae, metre-scale duplex systems, thrust and folds affecting tens of metres of strata, and pervasive lineations (Le Heron, 2007). Similar glaciogenic deposits and synsedimentary deformations have been recognized by Le Heron et al. (2007) in the Hirnantian series of the Ouzellarh Block (which belongs to the Anti-Atlas domain, as far as its Paleozoic structure is concerned; see Fig. 3.1B and below, Sect. 3.3.1). This strongly suggests that the Hirnantian icecap extended at least up to the northern border of the Anti-Atlas domain. Le Heron et al. (2007) propose, less convincingly, that the icecap reached into the southern Meseta (Jebilet and Rehamna).

The post-glacial deposits of the Silurian Aïn Deliouine Fm. correspond to former anoxic muds deposited during the eustatic transgression caused by the melting of the Late Ordovician icecap. The Rhuddanian (Lower Llandovery) sedimentation was limited to narrow gulfs impinging onto the emergent platform. An example of Rhuddanian gulf extends along the eastern border of the Ougnat-Ouzina Ordovician ridge in western Tafilalt (see Fig. 3.8A for location). The eustatic transgression was generalized from the middle-late Llandovery onward and graptolite-rich black shales deposited all over the former Ordovician platform. The Rhuddanian “hot shales” constitute the main source rock of the Saharan oil and gas fields. During this short period (1–2 m.y.), a favourable combination of factors existed, which led to the development of exceptionally strong oxygen-deficiency in the narrow depositional areas (Lüning et al., 2000). The post-Rhuddanian shales and silts are organically lean and have not contributed to petroleum generation. Carbonate sedimentation resumed by the end of the Silurian (Upper Ludlow) with alternations of *Orthoceras* and *Scyphocrinites* limestone beds and graptolitic shales.

3.1.2.2 Late Paleozoic

References: The fundamentals of Devonian stratigraphy of the Anti-Atlas were established by Hollard, 1967, 1974, 1981, and geological maps 1:200,000 cited above. More recent stratigraphic/paleontological studies have been published by Bensaid et al. (1985), Wendt (1985, 1988), Wendt et al. (1984), Wendt and Aigner (1985), Wendt & Belka (1991), Tahiri & El Hassani, eds. (2000), Fröhlich (2004). The remarkable Devonian mud mounds were thoroughly described by Brachert et al. (1992), Montenat et al. (1996), Mounji et al. (1998), Hilali et al. (1998, 2001),

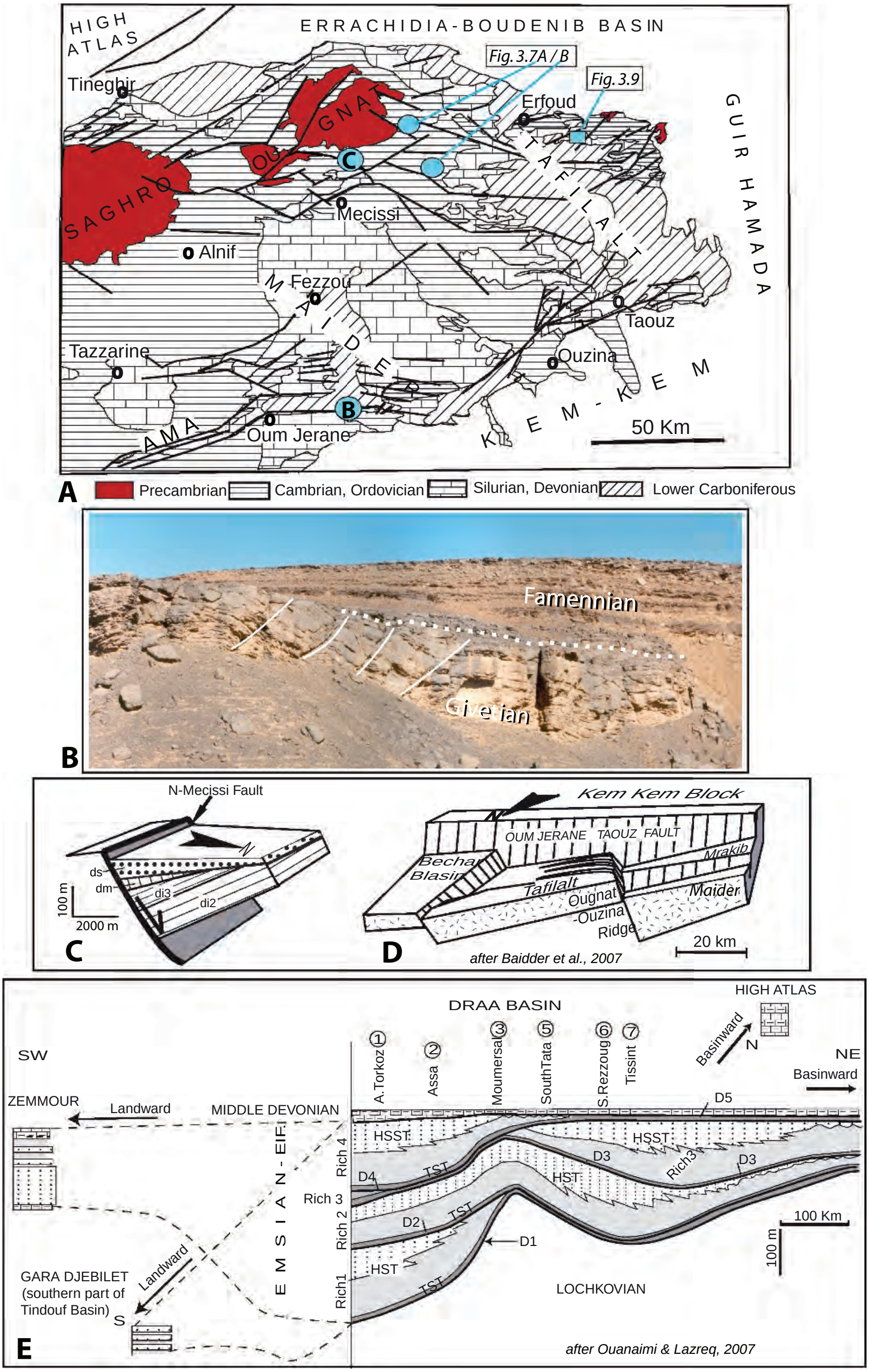

Fig. 3.8 (continued)

Kaufmann (1998), Belka (1998a,b), Wendt and Kaufmann (2006). More recently, Baidder et al. (2008) and Ouanaimi & Lazreq (2008) addressed the synsedimentary tectonics that controlled the Devonian deposits. The Carboniferous series and/or faunas of southern Morocco and southwestern Algeria were described by Cavaroc et al. (1976), Lemosquet et al. (1985), Wendt et al. (2001), Fabre (2005), Guiraud et al. (2005) and Korn et al. (2007, with references therein). The Carboniferous series at the northern border of the J. Saghro were incorporated into the South Meseta Zone during the Late Carboniferous orogeny, but they were deposited on the margin of the Anti-Atlas Domain. They have been described by Soualhine et al. (2003) and Graham & Sevastopulo (2007) with partly diverging conclusions.

The Devonian dislocation of the Pre-Saharan platform

The Late Paleozoic series of southern Morocco are well known for their rich Devonian fossil content and ammonoid stratigraphy. From the structural point of view, this epoch is characterized by the extensional dislocation of the former platform, resulting in the differentiation of basins and highs.

The Lochkovian deposits follow upward the Upper Silurian continuously with similar alternations of mudstones and limestones. However, it is worth noting the occurrence of a peperitic volcano in the northern Tafilalt region (about 20 km east of Erfoud; Fig. 3.8A). The Tafilalt and neighbouring Maider regions of the Eastern Anti-Atlas show striking examples of normal fault activity during the Early-Middle Devonian and particularly during the early Late Devonian (e.g. Fig. 3.8B). These synsedimentary faults account for the contrasting thickness Devonian series of the North Tafilalt Platform (ca. 50 m) and Maider Basin (ca. 900 m). The Early-Middle Devonian platform series comprise nodular limestones and crinoidal limestones locally associated with reef-mounds or, on the upper slope, with mud-mounds such as the Hamar Laghdad (Lakhdad) mounds (Fig. 3.9). Condensed pelagic limestones

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Fig. 3.8 Devonian deposits and synsedimentary extension in the Eastern Anti-Atlas (**A**–**D**) and Western Anti-Atlas (**E**), after Baidder et al. (2008), and Ouanaimi & Lazreq (2008), respectively. – (**A**) Schematic map of the eastern Anti- Atlas showing the main Devonian paleofaults, with location of the following figures. AMA: Anti-Atlas Major Fault, which continues into the Oum Jerane-Taouz Fault. The SE-trending faults of the Ougnat-Ouzina axis are superimposed on Ordovician paleofaults (Fig 3.7B, C). – (**B**) Extensional tilted blocks in the Givetian limestones of J. Mrakib, sealed by Upper Frasnian-Famennian limestones. – (**C**) Diagrammatic representation of the Upper Famennian unconformity onto the tilted Lower-Middle Devonian series of the J. Gherghiz block (di2: Pragian; di3: Emsian; dm: Eifelian, Givetian). The Famennian conglomerates (ds) contain phosphatized black pebbles of Ordovician sandstones shed over the continental shelf during the Upper Frasnian-Famennian transgression (Fröhlich, 2004). – (**D**) Sketch of the Middle-Late Devonian paleofault system of the Tafilalt-Maider region. – (**E**) Lateral variations of the Rich Group sequences of the Draa Plains and correlations with the neighbouring regions. D1–D5: major discontinuities; HST: High stand tract; TST: Transgressive system tract. Each Rich consists of a thin basal calcareous level, followed by sand-mud rythmites then sandstones. Thickness variations and depocenter migrations record extensional deformation of the platform during the Emsian

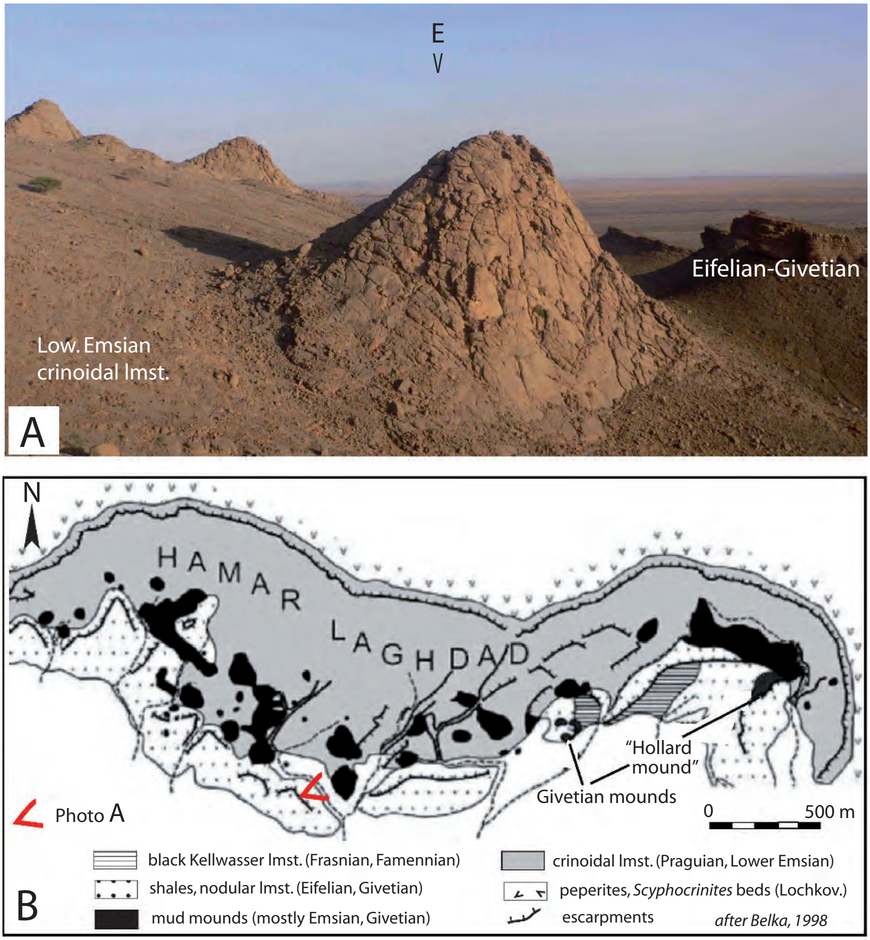

Fig. 3.9 The Hamar Laghdad (Lakhdad) mounds east of Erfoud (Tafilalt). Location: Fig. 3.8A. **A**: Typical view of the Emsian mounds (photograph O. Saddiqi). – **B**: Schematic map of the Hamar Laghdad (Lakhdad) area, after Belka (1998), modified. "The mound facies is composed of skeletal wackestone and mudstone riddled by stromatactis cavities. Small tabulate corals are very abundant. Trilobites and crinoids are ubiquitous, but less abundant than in the intermound carbonates. Nektonic organisms are represented by goniatites, orthoconic nautiloids, conodonts, and rare placoderms. Solitary rugose corals, brachiopods, ostracodes, gastropods and pelecypods are subordinate elements [...] Stromatoporoids and calcareous algae are conspicuously absent." (Belka, 1998). Microstromatoliths, a microbial structure, are common in the fenestrae and over the bioclasts (Hilali et al., 1998).The mud mounds developed in a faulted sub-basin over a south-dipping slope, which can be related to the occurrence of the Lochkovian volcano further north (Montenat et al., 1996; Hilali et al., 1998; Mounji et al., 1998)

rich in nektonic and planktonic faunas formed a thin cover over the platforms during the Frasnian-Famennian, associated with black bituminous shales (Kellwasser facies). In contrast, the thick Devonian series of the subsiding basins (Emsian-Upper Famennian of the Central Maider and South Tafilalt Basins) comprise sandy shales, debris flows and resedimented limestones with pyritic goniatites. Extensional faulting was particularly active during the Frasnian-early Famennian. The Saghro-Ougnat axis and the Tafilalt and Maider platforms emerged at the beginning of the Frasnian, before being flooded again. As a result, the earliest Upper Devonian sediments (Upper Frasnian pelagic limestones, Famennian black phosphatized conglomerates) unconformably overlie the tilted Early-Middle Devonian formations (Figs. 3.8B, C), and even the Ordovician formations locally. The overall system of the Devonian paleofaults (Fig. 3.8D) suggests a multidirectional extension with a dominant N-S component north of the Oum Jerane-Taouz main fault.

The Devonian series extend widely in the Draa Plains of Central-Western Anti-Atlas, between the J. Bani and J. Ouarkziz (Fig. 3.10). In the middle of these plains, the long train of folds of the J. Rich is formed by four sedimentary formations, virtually identical and labelled Rich 1 to Rich 4 (Fig. 3.8E). Each Rich consists

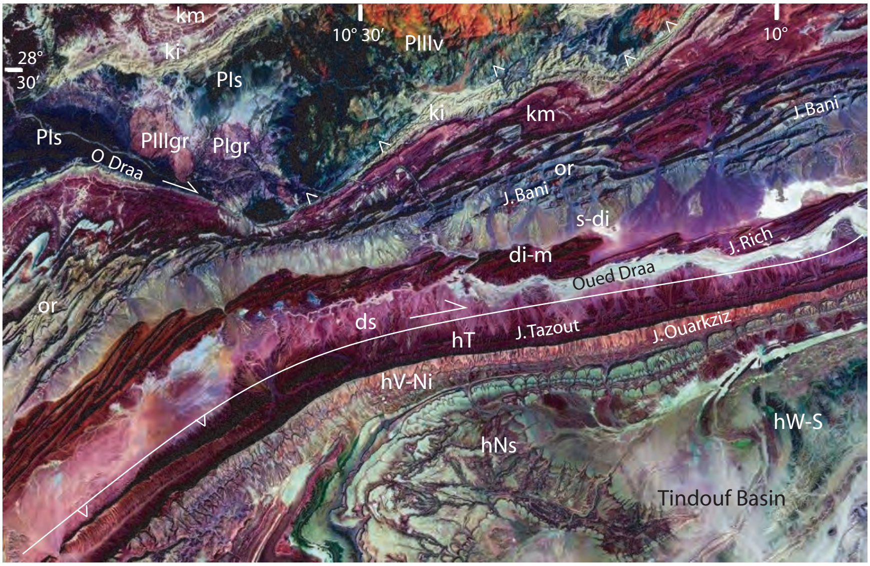

Fig. 3.10 The Anti-Atlas fold belt in the Lower Draa (Bas Draa) area; detail from Landsat N29-25. Location: see Fig 3.1. Structural sketch and cross-section: see Fig. 3.13. Three regions can be distinguished, from NW to SE, (1) the Bas Draa Precambrian inlier, thrust to the SE; (2) the folded Paleozoic cover (Appalachian-style relief), detached from its basement, which is tilted to the SE; (3) the faulted NW border of the Tindouf Basin, underlain by the West-African Craton. The obliquity of the Bani and Rich fold axes relative to this border suggests a right-lateral movement, combined probably with a NW-ward reverse throw (compare with Figs. 3.12 and 3.13). Legend after the geological maps, scale 1/500,000 Marrakech, 1956, and Al Youn Draa, 1/50,000, 2001. PIs: Paleoproterozoic schists and phyllites; PIgr: Mechebbouk Suite (2037–2030 Ma old granites); PIIIgr: Guellaba Suite (579–570 Ma old granites); PIIIv: Late Neoproterozoic volcanics and volcanoclastics; ki Adoudounian/Lower Cambrian; km: middle Cambrian; or: Ordovician; sdi: Silurian and Lochkovian; di-dm: Lower-Middle Devonian; ds: Upper Devonian; hT: Tournaisian; hV-Ni: Visean-Lower Namurian; hNs: continental Upper Namurian; hW-S: continental Westphalian-Stephanian

of a thin, transgressive calcareous member followed by sandy-pelitic rythmites and, finally, by a sandstone bar. The clastic input was sourced south and south-west of the Tindouf Basin in the Reguibat Arch crystalline and its sandy Upper Ordovician cover. The Rich Group ranges in age from the Upper Lochkhovian-Pragian (Rich 1) to the Upper Emsian-Lower Eifelian (Rich 4). The strong lateral thickness variations (Fig. 3.8E) suggest a significant tectonic mobility of the underlying crust. Contrasting with the Eastern Anti-Atlas setting, the Upper Devonian succession displays a monotonous basinal facies until to the transition to the Carboniferous.

The Early Carboniferous Latest Pre-Orogenic Deposits

The Carboniferous faunas and stratigraphy of the Anti-Atlas have been much less documented than the Devonian, due to the more scattered occurrence of fossil localities. The transition from Famennian to Early Carboniferous (Tournaisian) comprises a continuous succession of fine grained clastic strata. However, ferrugineous conglomerates and/or sandstones are observed at the Devonian-Carboniferous boundary in the Tafilalt-Maider area (Fig. 3.8A), for example east of Fezzou (Aguelmous perched syncline). Carbonate sedimentation resumed during the Early Visean (packstones with corals, bryozoan, brachiopods and crinoids intercalated in the clayey and sandy deposits near Erfoud). Reef mounds (J. Begaa) and mud mounds developed in the southern Tafilalt during the beginning of the Late Visean and the youngest ammonoid assemblage of the region indicate the Serpukhovian in the J. Bechar, close to the South-Meseta front.

It is worth noting that the Early Carboniferous strata of the Eastern Anti-Atlas are included in the fold belt. This yields a maximum age for the Variscan orogeny in these eastern regions. In contrast, the Carboniferous strata of the Western Anti-Atlas, south of the Famennian Draa Plains, belong to the northern border of the cratonic Tindouf Basin (Fig. 3.10). There, the Tournaisian pelites and sandstones of the Jebel Tazout are overlain by the Betaina Visean pelites, succeeded by the Upper Visean to Namurian limestones of the J. Ouarkziz. South of the latter ridge, the Betana Plain corresponds to Upper Namurian-Stephanian continental sediments deposited by south-directed fluvial systems (Cavaroc et al., 1976; Fabre, 2005). Thus, they record the orogenic uplift of the Western Anti-Atlas.

Synsedimentary fault tectonics occurred along the Anti-Atlas northern margin during the late Visean-Namurian, resulting in the uplift of ridges overlain by bioclastic limestones, and the emplacement of chaotic breccias and flysch-like deposits within the adjoining basins (Fig. 3.8A: Tineghir, Tisdafine and Ben Zireg olistostromes and turbidites). Similar syn-tectonic deposits are also found in some of the High Atlas Paleozoic massifs close to the Anti-Atlas (Ait Tamelil, Skoura). They immediately predate the Meseta-Anti-Atlas collision.

3.1.3 The Variscan Tectonics

References: After the pioneering structural analysis by Hassenforder (1987), the Variscan structures of the Anti-Atlas have been intensely studied during the last decade, and described in the publications by Soulaimani et al. (1997), Belfoul et al. (2002), Soulaimani et al. (2003), Guiton et al. (2003), Caritg-Monnot (2003), Caritg et al. (2003), Soulaimani & Piqué (2004), Helg et al. (2004), Hoepffner et al. (2005), Burkhard et al. (2006), Álvaro et al. (2006b), Robert-Charrue (2006), Baidder et al. (2008), Raddi et al. (2007), Soulaimani & Burkhard (2008). In contrast, the calibration and dating of metamorphism are still poorly documented: see Bonhomme & Hassenforder (1985), Buggish (1988), Belka (1991), Magroum in Soulaimani and Piqué (2004), Sebti et al. (2008) and Ruiz et al. (2008). The Ougarta belt and Bechar Basin structures are discussed by Donzeau (1974), Kazi-Tani et al. (1991), Haddoum et al. (2001), and Fabre (2005). Further references about the geodynamic interpretation are given in Sect. 3.4.

3.1.3.1 General

In the Western and Central Anti-Atlas, the Paleozoic series exhibit upright folds with large wavelength, clearly visible on satellite images (Fig. 3.10). The "Appalachian-type" relief of the J. Bani and J. Rich areas south of the ridge axis recalls that of the Valley-and-Ridge province of the External Appalachians. However, the Anti-Atlas basement crops out extensively in the ridge axis itself, in contrast with the External Appalachians. This suggests a thick-skinned tectonics in the Anti-Atlas, contrasting with the classical thin-skinned style of the External Appalachians (Burkhard et al., 2006).

The fold axis trend rotates from the NE (Western Anti-Atlas: J. Bani, J. Rich), to the ESE (Central Anti-Atlas), to the SE (Eastern Anti-Atlas and Ougarta). Thus, the Variscan belt seems to reactivate the Pan-African structural trends, and to correspond again to the curved northern boundary of the West African Craton. However, the apparent continuity between the eastern and western segments of the belt is not exclusive of significant differences: The Western Anti-Atlas is much more deformed than Eastern Anti-Atlas, and the age of the main folding episode is probably older in the west (Late Carboniferous) than in the east (Early Permian). The plate tectonic interpretation of these differences will be discussed in Sect. 3.4.

3.1.3.2 Western Anti-Atlas

In most of the Western Anti-Atlas, shortening of the Paleozoic strata occurs through buckling and flexural slip folding under very-low-grade metamorphic conditions. Strongly disharmonic folding results from the mechanical contrasts between competent and incompetent beds (Fig. 3.3). These disharmonic relationships can be observed in map view (Fig. 3.10) and deep natural sections (Fig. 3.11A). Shortening of the thick J. Bani Ordovician sandstones generates large wavelength cylindrical folds often associated with minor folds (Fig. 3.11B). By contrast, conical en-echelon folds with small wavelength develop in the thin multi-layers of the J. Rich Devonian series, bounded at their bottom and top by the thick incompetent levels of the Silurian and Upper Devonian, respectively (Fig. 3.11D). The latter folds display a remarkable, poly-stage development of the fracture systems (Guiton et al., 2003). In the most internal (western) parts of the belt, and especially within narrow shear zones (e.g. Lakhssas plateau, between the Ifni and Kerdous massifs), the folding mechanism changes and an axial plane cleavage develops in response to increasing strain and P-T conditions (Fig. 3.11C).

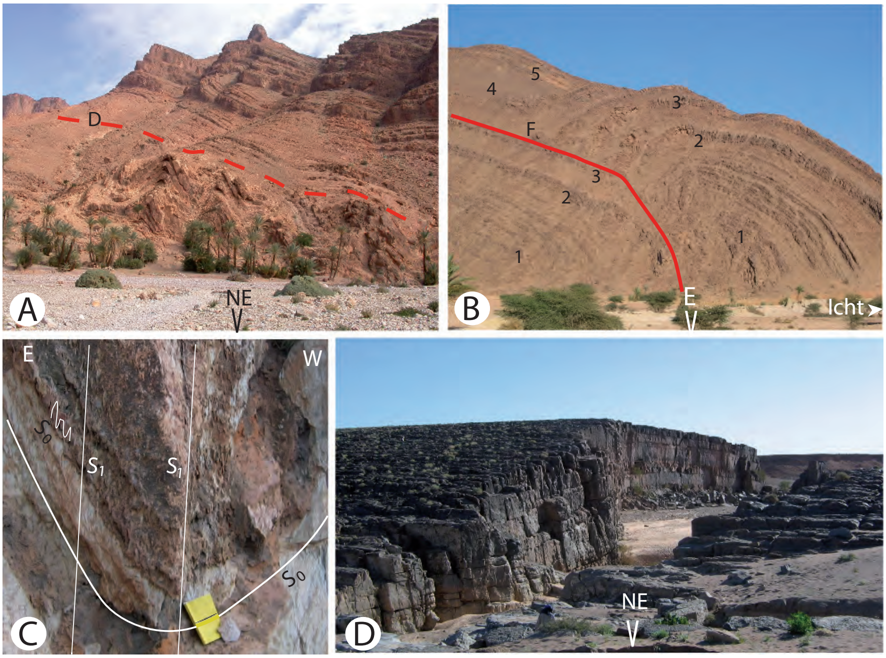

Fig. 3.11 Various types of folds from the Western Anti-Atlas. See Fig. 3.1 for location. – (**A**) Disharmonic folds in the Lower Cambrian limestones between the Igherm and Tagragra de Tata inliers (Photograph M. Burkhard). – (**B**) Fault propagation fold or "rabbit ear" in the southeast limb of a major J. Bani fold at Foum Icht (see Caritg-Monnot, 2003, Fig. 34). Stratigraphic legend after Destombes (2006–2008): 1:First Bani; 2–4: Ktaoua Gr.; 5: Er-Rwaidat Fm. The Second Bani Fm. outcrops at Icht village. – (**C**) Second order fold with axial plane cleavage in the Lower Cambrian calcschists and marbles of J. Inter in the Lakhssas first order synclinal zone between the Ifni and Kerdous Massifs (cf. Soulaimani & Bouabdelli, 2005). Tight, third order minor folds with N-S trending vertical foliation develop in the alternating greywacke-marble layers ("schistes à trous" facies). The notebook is 15 × 20 cm. – (**D**) Open anticline formed in the "Rich 3" sandstones (Middle Emsian), 15 km south of Tata. The polyphase fracturing developed in the fold is described by Guiton et al. (2003)

Folding of the Paleozoic succession implies its detachment from the Precambrian basement, a far more rigid material where faulting is the main deformation mechanism. The main décollement level is in the Lie-de-Vin Series, although the underlying Adoudounian dolomites might also detach on the "Série de Base". As the Anti-Atlas basement is involved in the tectonic shortening (cf. the numerous antiformal Precambrian inliers), we have to consider a balanced cross-section down to crustal depth (Fig. 3.12). The basement shortening would be accommodated by reverse faults connected to a 25 km deep intra-crustal detachment (Burkhard et al., 2006). Some conjugate faults may delimit extrusive blocks or pop-ups. These reverse faults likely result from the inversion of normal paleofaults, particularly active during the Late Neoproterozoic-Early Cambrian rifting. Unfolding of the J. Bani competent beds leads to a minimum shortening of 17–25%, the latter value being obtained by taking into account the continuous deformation by pressure solution. Locally, in the triangular zones associated with the basement faults, shortening may reach 40%.

The contrasting tectonic style in the western and central part of the belt (e.g. Lakhssas *versus* Rich folds) emphasizes the NW-SE polarity of the belt. In the westernmost part of the belt, i.e. in the Ifni and Bas-Draa areas, deformation reaches its maximum, with evidence of east- or southeast-verging thrusts, and pervasive, moderately dipping foliation (Fig. 3.13). The basement itself is deformed through multiple shear zones, resulting in short wavelength folding of the Lower Cambrian (Adoudounian) limestones. The fold virgations around the Ifni and Bas-Draa massifs suggest a significant displacement of these basement massifs toward the foreland. The southeast boundary of the fold belt is marked by the slightly tilted J. Ouarkziz Lower Carboniferous limestones, which belong to the northern border of the cratonic Tindouf Basin. In the Bas-Draa area, the J. Bani and J. Rich fold axes are slightly oblique to the ENE trend of the J. Ouarkziz (Figs. 3.10 and 3.13). This en-echelon setting suggests a moderate dextral displacement of the fold belt basement along the craton, coeval with folding. However, the abrupt decrease of shortening between the Rich and Ouarkziz implies the occurrence of a triangle structure with a blind thrust connected at shallow depth with a NW-verging backthrust (Fig. 3.12). This complex craton boundary likely corresponds to a paleofault inherited from the Late Neoproterozoic to Devonian evolution.

3.1.3.3 Central and Eastern Anti-Atlas

The Central Anti-Atlas displays clear examples of fold interferences between NE-trending early folds, similar to the folds of Western Anti-Atlas, and E-trending younger folds that are typical of Eastern Anti-Atlas. This is illustrated by egg-box structures in the Tata region (Fig. 3.1). Further to the east, folds become very open and show irregular trends controlled by the basement faults. The *Anti-Atlas Major Fault* (AAMF) corresponds to a Pan-African lineament (southern limit of the ophiolite thrusts and late sinistral strike-slip fault; see Chap. 2) reactivated during the

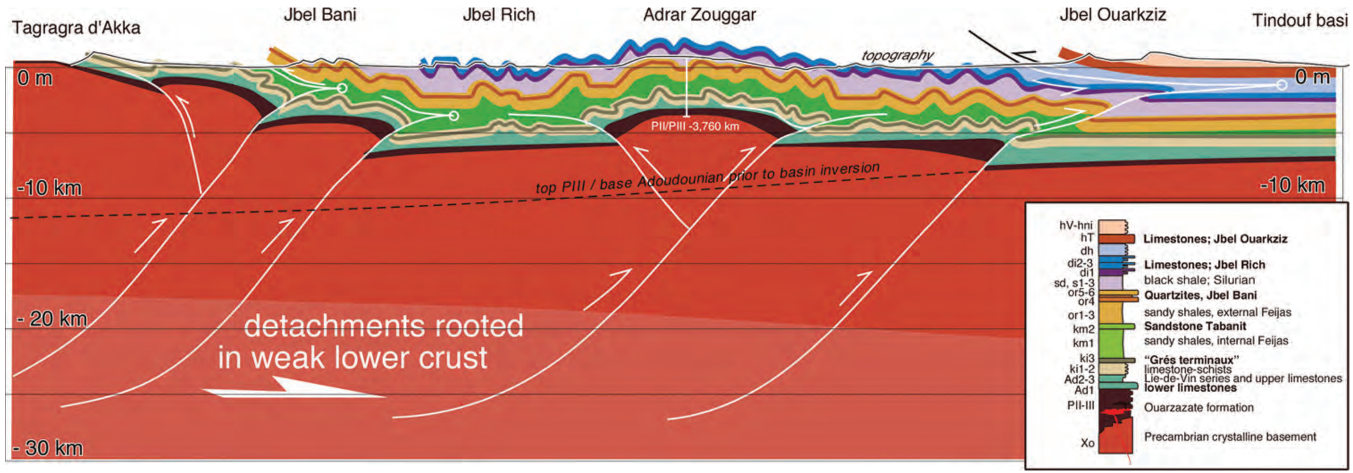

Fig. 3.12 Balanced cross-section of the external part of the Western Anti-Atlas (for location, see Fig. 3.1), after Burkhard et al. (2006), modified. Shortening along the section is close to 20%. Most basement faults have formed by inversion of paleofaults inherited from the rifting stage, but others correspond to new structures such as footwall shortcuts. The decrease of deformation at the craton border is accommodated by a triangle-zone structure within the Paleozoic series below the J. Ouarkziz, probably combined with right-lateral movement in the underlying faulted basement (see Figs. 3.10 and 3.13C)

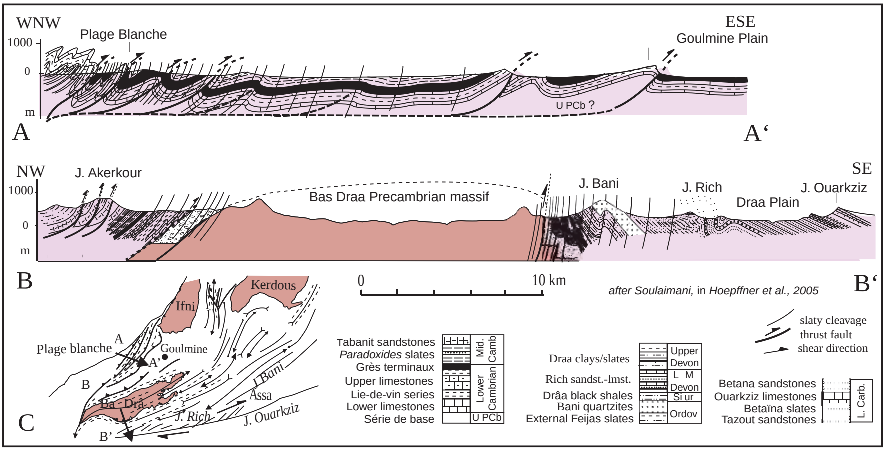

Fig. 3.13 Two cross-sections (AA', BB') and schematic structural map (C) of the Bas Draa region, southwestern Anti-Atlas, after A. Soulaimani, *in* Hoepffner et al. (2005). See Fig. 3.1 for location. The satellite view of the area is shown in Fig. 3.10

Variscan collision. In particular, the strike-slip faults that cut across the entire length of the Bou Azzer inlier, being part of the AAMF, were sealed by the Adoudounian onlap, whose moderate recurrent faulting is visible at the western end of the inlier. Close to Zagora, the AAMF splits into a southern branch directed to the Ougarta range, and an ENE-trending branch directed to Taouz in southern Tafilalt. The latter branch operated as an extensional fault system during the Middle-Late Devonian (Zagora graben; Oum Jerane-Taouz paleofault, Fig. 3.8D).

In the Eastern Anti-Atlas, the Paleozoic series are thinner, and include fewer décollement levels than in the Western Anti-Atlas (Fig. 3.3B). As a consequence, the Paleozoic sediments remain generally attached to the basement, except at the northern boundary of the domain (Sect 3.3). To the south of the Saghro-Ougnat axis, open folds are observed on basement pop-up structures, and en-echelon folds on strike-slip faults. The latter faults correspond to reactivated paleofaults, mostly Devonian in age (cf. Fig. 3.8), but often inherited from Cambrian or Ordovician normal faults. The Variscan collision affected a pre-existing mosaic of tilted blocks, which, due to compression, slid along each other (Fig. 3.14). Fault kinematics (as indicated by the en-echelon folds) and fold axis orientation are consistent with a NE-trending direction of shortening, which is also the shortening direction of the Ougarta chain. Nevertheless, another shortening direction broadly perpendicular to the Saghro axis can be recognized at the northern border of the belt, within the South Meseta units and underlying Anti-Atlas slivers (Tineghir and Tamlelt slivers; see below, Sect. 3.3.3). These two superimposed directions give rise to interference folding south of the Anti-Atlas axis, for example in the Maider Basin near Fezzou (Figs. 3.1 and 3.8A).

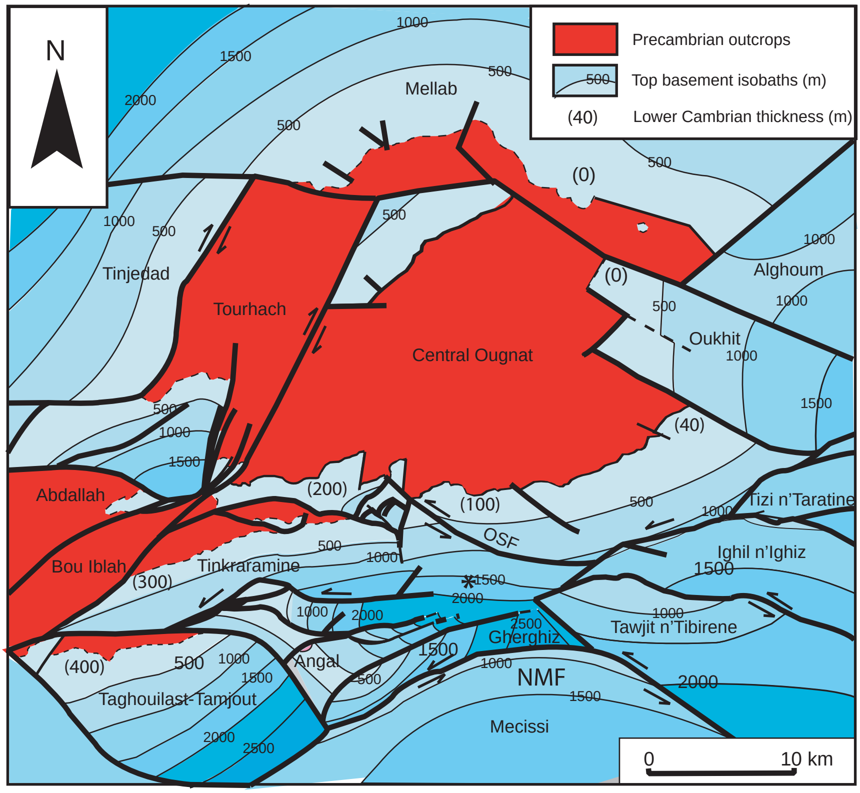

Fig. 3.14 The mosaic of tilted basement blocks inferred from the thickness and mean dip of the Paleozoic cover in the Ougnat area (Eastern Anti-Atlas), after Baidder et al., (2008). See Fig. 3.1 for location. The faults are inherited from the pre-orogenic extensional evolution, particularly active during the Late Devonian. Fault kinematics (arrows) during the Variscan compression has been deduced from the orientation of the en echelon folds (not shown) relative to each fault. NMF: North-Mecissi Fault; OSF: Oued Smile Fault. Asterisk: location of Fig. 3.8C

3.1.3.4 Age of Folding

Folded Lower Carboniferous deposits occur in the Anti-Atlas fold belt, at least within its eastern part, but neither Upper Carboniferous nor Permian sediments are preserved within the belt. Thus, the stratigraphic dating of the folding event(s) relies on indirect arguments inferred from the evolution of the Paleozoic sedimentation in the neighbouring Tindouf and Bechar Basins. Additional information comes from scarce isotopic or fission track datings, and from the inferred correlations with the Meseta Domain, whose tectonic timing is much better known.

The earliest folding episode of the Western Anti-Atlas (i.e. its NW-SE shortening) occurred probably during the Namurian. This assumption is suggested by the fact that, at this time, the J. Ouarkziz marine deposits (Upper Visean to Lower Namurian = Serpukhovian) are replaced by the J. Reouina continental sandstones

(Fig. 3.10), which contain an Upper Namurian flora. Deltaic channels supplied from the NW can be observed there, suggesting that the Anti-Atlas has been uplifted and eroded as early as ~ 320 Ma (Cavaroc et al., 1976; Fabre, 2005). The Tindouf Basin series remained continental during the Westphalian and Stephanian (Merkala Fm.), without any deformation.

In contrast, the earliest folding episode seems younger in the Eastern Anti-Atlas. In the adjoining Bechar Basin, marine sedimentation continues until the end of the Moscovian (Middle Late Carboniferous, cf. the J. Bechar limestones, and J. Antar and Mezarif reefs), although with a few regressive intervals of eustatic origin. Nevertheless, the western part of the basin received abundant continental material by the end of this period ("Kenadzian"), which testifies to the uplift of the neighbouring belts (Ougarta, Anti-Atlas and/or Meseta). A swampy deltaic regime took place during the late Westphalian, followed by the fluviatile Lower Abadla red beds, latest Westphalian to Stephanian-Autunian in age. The latter deposits are unconformably overlapped by the Upper Abadla red beds which are topped by spilitised basalt flows, and thus, assigned to the Upper Triassic. Therefore, in the Eastern Anti-Atlas and Ougarta regions, the earliest deformation would have occurred during the Stephanian-Lower Permian interval (i.e. between 305 Ma and 295 Ma).

A tentative correlation between the shortening episodes recognized in the Anti-Atlas with those of the Meseta Domain (Sect. 3.3.3) allows us to improve chronology of folding in the Anti-Atlas. The NW-SE shortening of the Western Anti-Atlas can be logically correlated with the shortening which affects the Meseta along the same direction during the Namurian-Westphalian. Moreover, the formation of the E-W trending folds of the Central and Eastern Anti-Atlas could be coeval with the N-S shortening of the Meseta Domain, dated from the Upper Westphalian. Finally, the NE-SW shortening phase recognized in the Eastern Anti-Atlas and Ougarta likely correspond to the Stephanian-Lower Permian shortening event of the Meseta Domain (see below, Fig. 3.31).

Isotopic K-Ar analyses of fine micaceous fractions from the Ediacaran and Cambrian formations around the Kerdous Massif yielded ages ranging between 357 ± 9 Ma and 291 ± 7 Ma, with a mean age at 315 ± 9 Ma (Bonhomme & Hassenforder, 1985). K-Ar dates at 360–310 Ma from the Kerdous schists and granites (Margoum, 2001, in Soulaimani & Piqué, 2004) have been assigned to the combined effects of burial heating and Variscan compression. At the scale of the entire SW flank of the Anti-Atlas, Ruiz et al. (2008) observed that illite "crystallinity" improves with stratigraphic age and correlates with estimated paleo-overburden, thus fulfilling the conditions of burial metamorphism. However, the role of tectonic thickening and of hot, pressurized fluids sourced in the most internal parts of the Variscan belt (Mauritanides) must be admitted as burial alone cannot explain temperatures as high as ~ 350 °C in the Cambrian and underlying units (total thickness of the Paleozoic sequence: about 8–9 km). Based on fission track analyses of zircon grains from the Kerdous and Ifni inliers, Sebti et al. (2008) demonstrate that this thermal event was followed by rapid cooling at 330–320 Ma (Late Visean-Serpukhovian). This can be assigned to the uplift of the thickened crust and erosion of the young mountain belt, consistent with the coeval sedimentation of the J. Reouina molasses in the Tindouf Basin.

3.2 The Adrar Souttouf-Dhoul Mauritanide Segment

References: Concerning the Mauritanides as a whole, some of the most recent and significant papers are those by Lécorché et al. (1991), Burg et al. (1993), Villeneuve & Cornée (1994), and Le Goff et al. (2001). The Awsard sheet of the geological map of Morocco, scale 1:50,000 (Rjimati & Zemmouri, 2002) and the paper by Villeneuve et al. (2006) concern particularly the Adrar Souttouf segment.

The Mauritanide fold belt occurs mainly in Senegal, Mali and Mauritania (Fig. 3.15A). However, the belt continues into southernmost Morocco, being exposed in the Adrar Souttouf and Dhlou (Ouled Dhlim or Delim) areas. The large

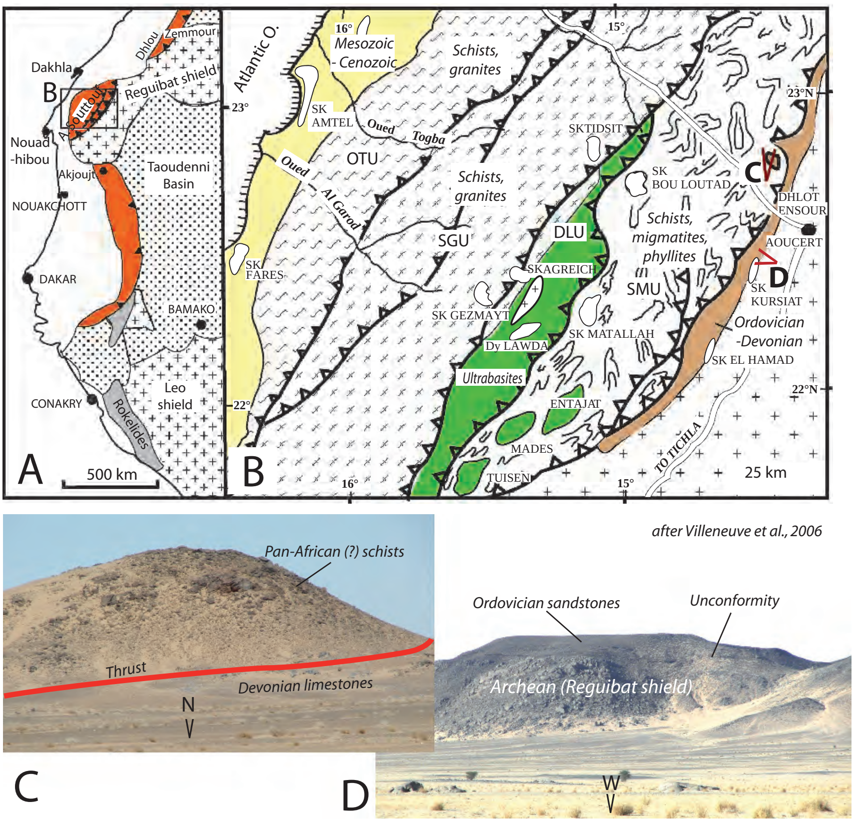

Fig. 3.15 The Adrar Souttouf segment of the Mauritanide belt, after Villeneuve et al. (2006). – (**A**) Location map. – (**B**) Structural map of the metamorphic nappe stack thrust over the Reguibat shield cover. See also the geological map scale 1/50,000, Awsard sheet (Rjimati et al., 2002). SK: Sebkha. Tectonic units, from W to E, OTU/SGU/DLU/SMU: Oued Togba/Sebkha Gezmayt/Dayet Lawda/Sebkha Matallah Unit. – (**C**) The thrust front at Tiznigaten (N 22°44′; W 14°25′), north of Aoucert (Awsard); the allochthonous Pan-African schists overlie autochthonous, bluish Orthoceras limestones. – (**D**) The base of the autochthonous series at Arieles, southwest of Aoucert (N 22°30′; W 14°30′): Ordovician sandstones overlapping Archean granite (Photos M. Villeneuve)

exposures of the Adrar Souttouf hills, in particular, illustrate very clearly the thin-skinned thrust structure which characterizes the entire Mauritanides belt (Fig. 3.15B).

In the central part of the massif, three major tectonic units have been recognized, which include, from west to east and top to bottom (Villeneuve et al., 2006):

(a) the Gezmayet Sebkha schist unit (orthogneisses, garnet granites, amphibolites, quartzites), where a magmatic sample yielded a whole rock (WR) K-Ar age at 665 ± 16 Ma;
(b) the Dawyet Lawda unit, subdivided into four sub-units, (i) amphibole gneisses and amphibolites, dated at 396 ± 9 Ma (WR); (ii) metagabbros, with a K-Ar amphibole age at 1061 Ma; (iii) gabbros, basalts amphibolites and peridotites (probably an ophiolite suite), where an undeformed basalt yielded a 733 ± 17 Ma WR date; (iv) gabbros and gabbros-diorites, from which three samples have been dated between 484–514 Ma, whereas a sample from the top thrust contact yielded a 274 ± 12 Ma WR date;
(c) the very heterogeneous Matallah Sebkha unit, including siliceous or calcareous phyllites, quartzites, granites, migmatites and amphibolites. The amphiboles of a dioritic gabbro yielded a K-Ar result of 1110 Ma.

The varied metamorphic units show a NW-dipping foliation deformed by upright folds or, in the easternmost unit (Gezmayet Sebkha), by eastward verging overturned folds. Foliation is lacking in the gabbros and basalts of the Dawyet Lawda unit, as well as from the granite and basalt dykes which crosscut the varied units. Along the eastern border of the massif, the stacked units overlie the thin Paleozoic cover of the Reguibat Arch, which consists of Ordovician sandstones and quartzites, Silurian-Lower Devonian (?) calcareous pelites, and Middle-Upper Devonian Orthoceras limestones (Fig. 3.15C, D). At the opposite, western side of the massif, the uppermost nappe (Oued Togba unit) is overlain by the tabular Mesozoic-Cenozoic series of the Tarfaya-Dakhla Basin.

Therefore, the Adrar Souttouf is clearly a Variscan segment, even if its allochthonous material seems to include only Precambrian rocks. Garnets from eclogite lenses cropping out in northern Mauritania were recently dated by Sm-Nd at ca. 330 Ma, whereas U-Pb dating of isolated zircon crystals from the same outcrops yielded a late Neoproterozoic age (ca. 595 Ma) for the gabbroic protolith (Le Goff et al., 2001). The occurrence of a high pressure-low temperature event in the Mauritanides strongly distinguishes this belt relative to the Mesetan Variscides. However, the Mauritanide thrust emplacement onto the craton corresponds to a late, low-temperature phenomenon, which was dated at ~ 310 Ma (K-Ar and ^{39}Ar-^{40}Ar datings) on samples from the nappes and from the sedimentary autochthon in the central and southern parts of the belt (e.g. Lécorché et al., 1991). At the northern tip of the belt, i.e. in the Dhlou-Zemmour segment, the Paleozoic series are strongly folded and divided into numerous slivers thrust onto the western termination of the Tindouf Basin. It must be noted that this segment corresponds both to the foreland fold-and-thrust belt of the Mauritanides and to the southern extension of the most internal Anti-Atlas units (cf. Chap. 1, Fig. 1.11).

As for the allochthonous material incorporated within the Mauritanides, it includes ophiolitic remnants of probable Pan-African age (Dawyet Lawda unit) affected by high temperature metamorphism prior to 665 Ma, as well as remnants of a coeval calc-alkaline arc (Sebkha Gezmayet unit). By contrast, ages in the range of 1000–1100 Ma are typical for the Grenvillian Orogen of the Appalachian foreland, being lacking elsewhere in north-western Africa. Remarkably, granulite sample collected at the base of the Mazagan scarp off El Jadida yielded similar dates, close to 900–1000 Ma, suggesting a northward extension of the Mauritanides up to the latitude of El Jadida. This, of course, is important for the reconstruction of the Caledonian-Variscan Orogen before the spreading of the Atlantic Ocean (see below, Sect. 3.4).

3.3 The Variscides of the Meseta Domain

3.3.1 The Meseta Domain: A Complex Collage

References: The classical views on the structure and evolution of the Mesetan Variscides or Hercynides are reported in Piqué & Michard (1989), and El Hassani et al., Eds. (1994). The following description also relies on the synopsis by Hoepffner et al. (2005, 2006) which contains numerous references to recent Moroccan literature. Specific references concerning the pre-orogenic and orogenic periods are proposed in the corresponding sections.

The Variscides (Hercynides) of the Meseta Domain include all the Paleozoic massifs north of the South Atlas Fault (SAF), except the Ouzellarh Block of the Marrakech High Atlas which belongs to the Anti-Atlas Paleozoic domain (Fig. 3.16). The Meseta Domain differs from the Anti-Atlas by the intensity of the tangential deformation, the occurrence of pre- and late-orogenic magmatism (gabbros and granites, respectively), the importance of metamorphism, and the presence of scattered Late Carboniferous-Permian continental basins, moderately deformed before the onset of the Mesozoic-Cenozoic cycle.

The Paleozoic terranes of the Meseta Domain are widely covered by Mesozoic-Cenozoic formations, such as the Triassic-Jurassic of the Atlas domain, or the Cretaceous-Cenozoic plateaus of the Western Meseta (Fig. 3.17). As a result, the Variscan basement is exposed only in isolated massifs and small “boutonnières”, whose names are recalled in Fig. 3.16. This discontinuous setting is a consequence of the complex deformation of the Variscan basement during the Atlas Orogeny, which affected not only the Atlas belt itself, but also the neighbouring regions. A cursory glance at the elevation map of North Africa (Chap. 1, Fig. 1.2) makes it easy to understand that both the Eastern and Western Meseta blocks were more or less involved in the Atlas deformation between the Maghrebide collision belt and the Saharan craton. A major Neogene reverse fault bounds the Jebilet Massif to the north (see Chap. 4, Fig. 4.27), and thus, the Jebilet (“small mountains” in Arabic)

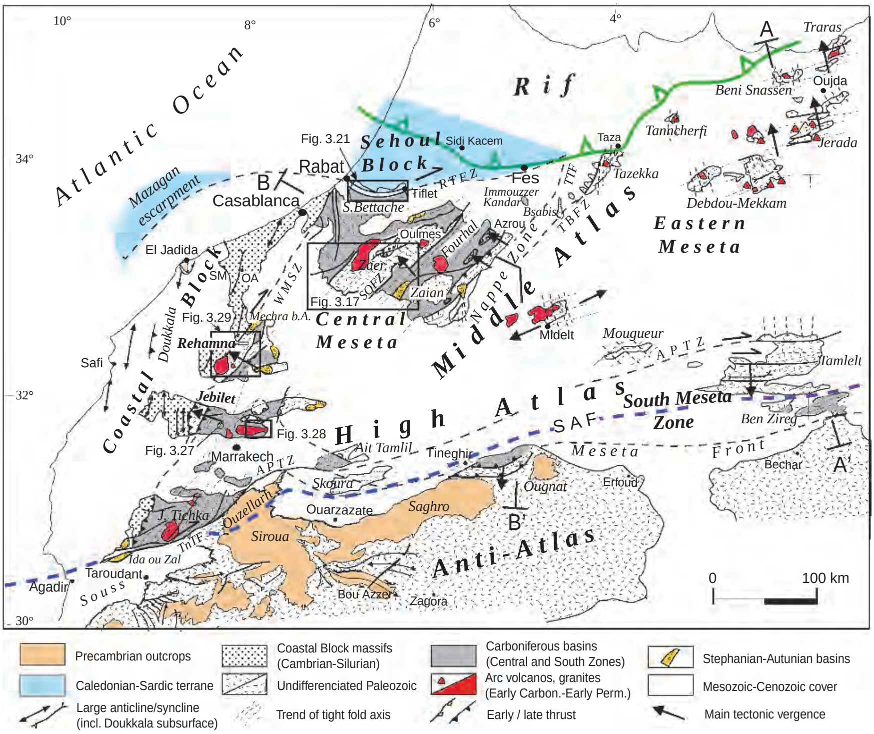

Fig. 3.16 Structural map of the Variscides of the Meseta Domain, after Piqué & Michard (1989) and Hoepffner et al. (2005), modified. APTZ: Atlas Paleozoic Transform Zone; OA: Oulad Abbou; RTFZ: Rabat-Tiflet Fault Zone; SAF: South Atlas Fault; SM: Sidi Said Maachou; SOFZ: Smaala-Oulmes Fault Zone; TBFZ: Tazekka-Bsabis Fault Zone; TnTF: Tizi n'Test Fault; TTF: Tizi n'Tretten Fault; WMSZ: West Meseta Shear Zone

could be considered as a northern branch of the Atlas Chain instead of a southern massif of the Meseta (indeed it is both). Moreover, the Meseta Domain has been affected by the large scale uplift linked to the Neogene asthenosphere upwelling which emplaced abundant alkali basalts over the entire Siroua-Middle Atlas-Eastern Rif magmatic belt (cf. Chap. 4). The vertical movements that affected the Meseta domain are discussed in Chap. 7.

From the point of view of its Paleozoic evolution, the Meseta Domain (Meseta Block) appears, in fact, to be a tectonic complex, including four subsidiary blocks, namely the Sehoul Block, Coastal Block ("Môle côtier"), Central Meseta and Eastern Meseta Blocks, sutured along shear zones and/or thrusts (Figs. 3.16 and 3.18).

The *Sehoul Block* extends from the Rabat-Tiflet Fault Zone (RTFZ) to the basement of the Gharb Basin and El Jadida offshore. This block contrasts with the other Meseta units by the fact that its Cambrian-Ordovician formations have been affected by a "Caledonian" or "Sardic" tectono-metamorphic phase ending with the intrusion of granodiorites at ca. 430 Ma. Thus, the Sehoul Block corresponds to an exotic

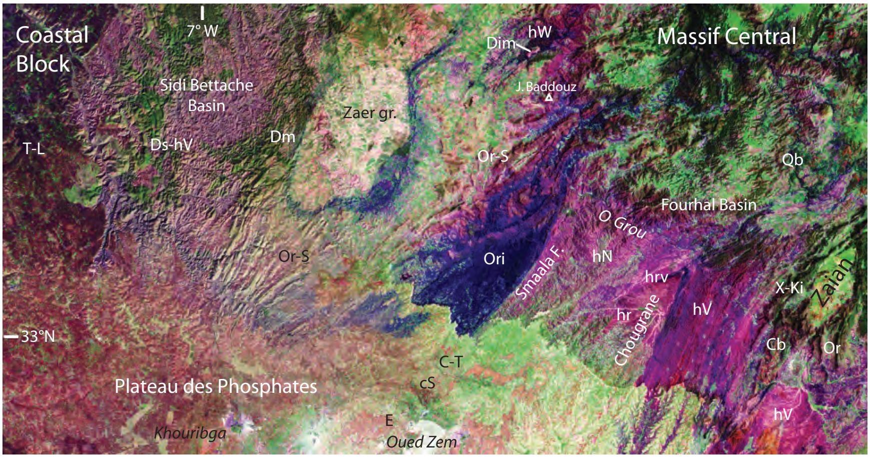

Fig. 3.17 Landsat image of the southwest Central Massif and overlying Meso-Cenozoic cover. See Fig. 3.16 for location. The scene is about 60×100 km large. X-Ki: Late Neoproterozoic-Lower Cambrian (Goaida); Cb: Middle Cambrian (Goaida: see Fig. 3.20); Ori: Lower Ordovician ("Schistes en dalles"); Or-S: Middle-Upper Ordovician quartzites and Silurian black shales deformed into large cylindrical folds; Dm: Middle Devonian limestones; Ds-hV: Famennian-Upper Visean turbidites; hV: Visean; hN: Namurian; hW: Lower Westphalian (Sidi Kassem red beds; the Devonian klippe shown in Fig. 3.30 is visible here as a whitish spot); hr: Autunian red beds; hrv: Autunian rhyolites; Zaer gr: Zaer granite (300–280 Ma); T-L: Upper Trias-Lower Lias; C-T: Cenomanian-Turonian limestones; cS: Senonian; E: Maastrichtian-Middle Eocene (phosphate series); Qb: Quaternary basalts

terrane originating either from the Caledonian Belt or from the South Caledonian (Sardic) Belt (Sect. 3.4).

The *Coastal Block* includes the Casablanca anticlinorium and the western parts of the Rehamna, Jebilet and High Atlas Massifs. This block is also known in wells and by seismic reflection beneath the Doukkala Triassic-Cretaceous basin and again appears in the small El Jadida outcrops. The Coastal Block is characterized by weak Variscan deformation, contrasting with the Central Block further east. A narrow, intensely deformed fault zone extends between both blocks, referred to as the *West Meseta Shear Zone* (WMSZ). The latter zone exposes Lower Devonian "Old Red Sandstones" facies overlying unconformably Cambrian or Ordovician formations. Westward in the Coastal Block (Doukkala), the unconformity is replaced by a sedimentary hiatus between the Upper Ordovician (Caradoc) and the Silurian-Devonian formations.

The *Central Meseta Block* includes the Sidi Bettache Basin, the Central Massif, the central and eastern parts of the Rehamna and Jebilet massifs, and most of the High Atlas Massif. This was a thinned crustal domain during the Late Devonian-Early Carboniferous, judging on the importance of mafic magmatism and turbiditic deposits. The block is now strongly shortened through dominantly NW-verging folds, duplexes and nappes, which developed during the latest Variscan phases (Upper Carboniferous-Permian). Only the eastern parts of the block (Nappe Zone,

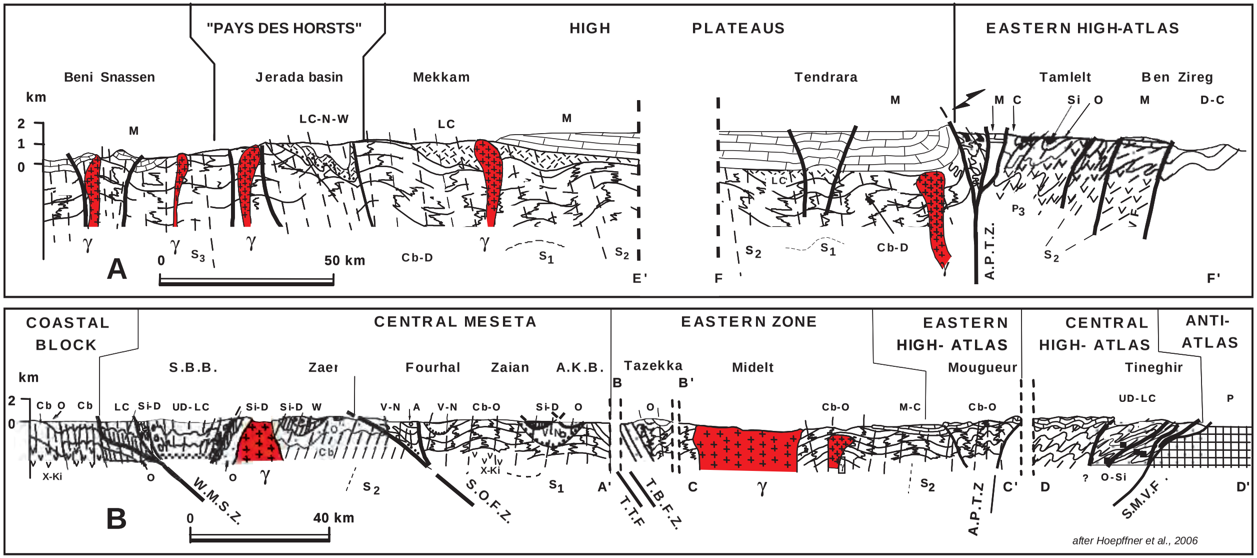

Fig. 3.18 Schematic cross-sections of the eastern (**A**) and central (**B**) regions of the Meseta Domain, after Hoepffner et al. (2006). Location: see Fig. 3.16. X-Ki: Neoproterozoic-Lower Cambrian; Cb: Middle Cambrian; O: Ordovician; Si-D: Silurian-Middle Devonian; UD-LC: Upper Devonian-Lower Carboniferous; C: Carboniferous; V-N: Visean-Namurian; W: Upper Westphalian; A: Autunian; γ: granitoids; S1: slaty cleavage, foliation (Eovariscan phases); S2: *idem* (Variscan phase); A.K.B: Azrou-Khenifra Basin; F.B: Fourhal Basin; M: Mesozoic cover (High Atlas); S.B.B: Sidi Bettache Basin. Other abbreviations as in Fig. 3.16

Azrou-Khenifra and Fourhal Basins) display significant structures related to an early compressional phase (Tournaisian-Early Visean).

The *Eastern Meseta Block* not only comprises the Paleozoic massifs of the Eastern Meseta *s.s.* from Oran (Traras, in Algeria) and Oujda (Beni Snassen, Jerada, Debdou-Mekkam) to Midelt (High Moulouya boutonnières), but also the inliers of eastern Tazekka (Middle Atlas), Mougueur and northern Tamlelt (Eastern High Atlas). The western boundary of the block involves two main faults, mostly hidden beneath the Middle Atlas Mesozoic series, (i) the classical *Tazekka-Bsabis Fault Zone* (TBFZ; cf. Hoepffner et al., 2005), and (ii) the less known *Tizi n'Tretten Fault* (TTF; see Charrière & Régnault, 1989; Willefert & Charrière, 1990). The Eastern Meseta Block is characterized by an early synmetamorphic folding event, poorly dated from the Late Devonian-Tournaisian. This eastern and relatively internal domain formed an almost rigid magmatic arc during the Late Carboniferous shortening phase.

The whole Meseta Domain is bounded to the south by the *Atlas Paleozoic Transform Zone* (APTZ), which widens eastward into a zone of slivers referred to as the *South Meseta Zone* (both being sometimes referred to as "South-Meseta Shear Zone", SMSZ). This is a major boundary zone, notwithstanding the lack of ophiolitic remnants. It is characterized by dextral displacement during the Late Carboniferous. In the High Atlas Paleozoic massif, the APTZ corresponds to the Tizi n'Test-Meltsen fault system which abruptly separates the Meseta metamorphic and granitic units (J. Tichka-Erdouz Massif) from the poorly deformed Ouzellarh block. Further east, the South Meseta slivers are thrust onto the Anti-Atlas along the *Meseta Front* (Tineghir and Tamlelt slivers). The whole Meseta-Anti-Atlas boundary zone was strongly reactivated during the Atlas orogeny (Chap. 4).

3.3.2 Pre-Orogenic Evolution

References: The references used in the present synopsis are indicated in the legend of the varied figures hereafter (and in particular in Fig. 3.23). Useful additional references concerning the pre-Variscan period are as follows: (i) *Sehoul Block and "Caledonian" events*: Hollard et al. (1982), Tahiri (1983), Cornée et al. (1987), Charrière et Régnault (1989), El Hassani (1994a,b); (ii) *Western Meseta, Central Massif*: Willefert & Charrière (1990), Cattaneo et al. (1993), Cailleux (1994), Zahraoui (1994), El Attari et al. (1997), El Kamel et al. (1998), Razin et al. (2001), Attou & Hamoumi (2004), Benfrika & Raji (2003); (iii) *Eastern Meseta*, *South Meseta Front*: Filali et al. (1999); (iv) *Southern Meseta (Rehamna, Jebilet)*: Huvelin (1977), Michard et al. (1982), Aarab & Beauchamp (1987), Baudin et al. (2003), Essaifi et al. (2004); (v) *Atlas massifs*: Ferrandini et al. (1987), Cornée et al. (1987), Ouanaimi (1992, 1998), El Archi et al (2004); (vi) *Cambrian magmatism*, Ouali et al. (2000), El Hadi et al. (2006a). References concerning the Variscan orogenic period and those concerning the geodynamic interpretation are proposed in the corresponding sections.

In all Meseta domains except the Sehoul Block and part of the Coastal Block/ WMSZ, the Paleozoic series correspond to a virtually continuous sequence from Lower Cambrian to Upper Devonian (Fig. 3.19). Rhyolites and granites of late Neoproterozoic age (cf. "P III") crop out locally in the Coastal Block (El Jadida), Central Rehamna (where they are converted into orthogneisses dated at 593 ± 8 Ma, U/Pb zircon; Baudin et al., 2003), and the Zaian Mountains (southeastern Central Massif). This suggests that the Meseta Domain is essentially built on a Pan-African continental crust.

The pre-orogenic evolution of the belt sometimes compares, but frequently contrasts with that of the Anti-Atlas. Two main periods can be distinguished, i.e. the Lower Paleozoic and the Devonian, separated by the "Caledonian" or "Sardic" events of Late Ordovician-Silurian age. The Early Carboniferous basins are considered in the next section, as they are synorogenic at the scale of the entire Meseta Domain.

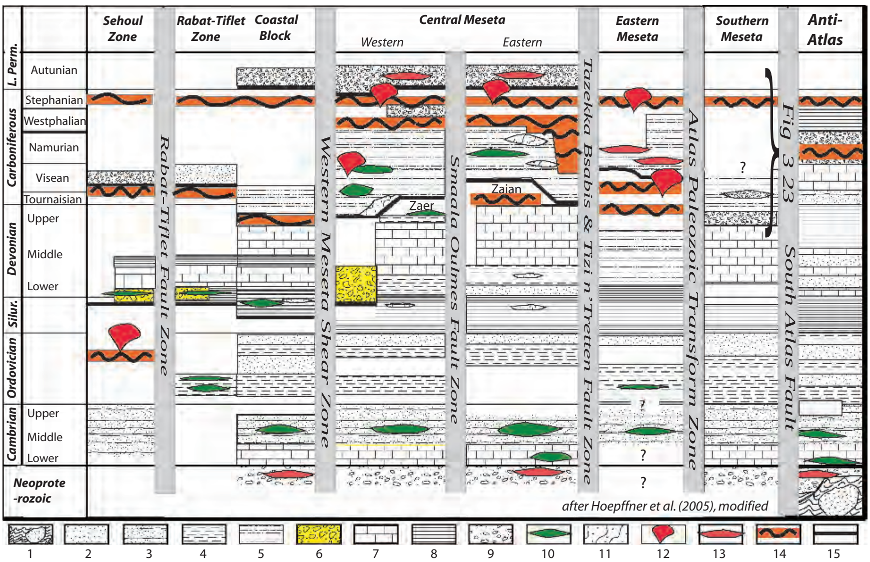

Fig. 3.19 Stratigraphy of the main structural zones of the Meseta Domain, after Hoepffner et al. (2005), modified. The question mark in the Southern Meseta evokes the possibility of Namurian-Westphalian deposits (Soualhine et al., 2003), although it is rejected by Graham & Sevastopulo (2007). The last column on the right shows the Anti-Atlas stratigraphy for comparison. 1: Pan-African basement; 2: sandstones, quartzites; 3: shales, greywackes; 4: argillites, silts; 5: turbiditic deposits; 6: "Old Red Sandstones"; 7: limestones; 8: black shales; 9: calc-alkaline volcanoclastic formations; 10: alkaline to tholeitic volcanism; 11: olistostromes, mélanges; 12: granites; 13: acid to intermediate volcanism; 14: tectonic phase; 15: angular unconformity; White: sedimentary hiatus. See Fig. 3.23 for more detail and references concerning the Late Devonian-Permian orogenic period

3.3.2.1 Lower Paleozoic Extension

The Lower Cambrian carbonates are much reduced relative to those of the Anti-Atlas. They include archaeocyathid limestones in the western Jebilet; and banded limestone-greywacke associations showing the foliated "schistes à trous" facies, similar to those of the Lakhssas plateau (Fig. 3.11C), in the Zaian Mountains, Rehamna and High Atlas. The overlying greywacke formations belong to the Middle Cambrian (cf. the Anti-Atlas "Schistes à *Paradoxides*"). The Lower-Middle Cambrian extensional regime is documented by widespread, dominantly alkaline to tholeitic submarine volcanism, e.g. in the Oued Rhebar Cambrian rift (which extends in the Coastal Block from the O. Rhebar south of Rabat to Sidi Said Maachou east of El Jadida; Fig. 3.19), in the western block of the Marrakech High Atlas (J. Erdouz and Tichka massifs), and in the Zaian Mountains (Fig. 3.20). The calc-alkaline character locally shown by some of these basalts is thought to indicate a contaminated mantle source (El Hadi et al., 2006a). This volcanism matches that of the northern margin of the Anti-Atlas domain, as observed in the Ounein (west of the Ouzellarh Block) and Oukhit area (east of the Ougnat massif).

The Ordovician deposits of the Coastal Block (Casablanca anticlinorium), Rehamna and Central Massif (Zaer, Zaian) are easily correlated with that of the Anti-Atlas, although the former generally tend to display thinner and more distal deposits, richer in micaceous-argillaceous intervals, especially in the northern and eastern massifs. Sedimentological similarities, such as the frequency of clastic biotite, suggest a common southern origin for all these terrigenous deposits, i.e. the erosion of the central Sahara massifs under cold climatic conditions. Le Heron et al. (2007) proposed that the Hirnantian icecap reached into the Jebilet and Rehamna Massifs, but this is a matter of debate.

Remarkably, the Lower Ordovician pelites of the southern Rabat-Tiflet Zone comprise several 5–15 m-thick intercalations of spilitic pillow basalts, as well as a 10 m-thick sill of spilitised gabbro (Tahiri & El Hassani, 1994; right bank of

Fig. 3.20 Middle Cambrian pillow basalts at Bou Acila/Goaida, in the core of the Zaian anticlinorium (eastern Massif Central). Note the hammer on the left. These continental tholeites (Ouali et al., (2003) overlie Lower Cambrian archaeocyathid limestones overlying Neoproterozoic rhyolites ("X-Ki" in Fig. 3.17)

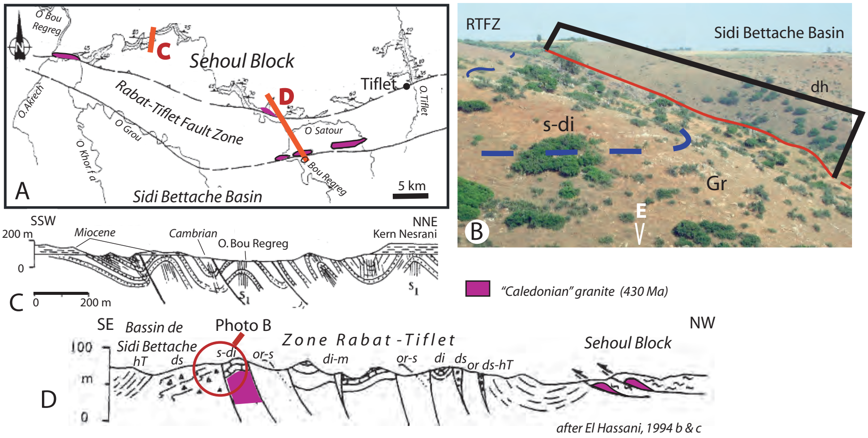

Fig. 3.21 The Sehoul block, an exotic "Caledonian" or "Sardic" terrane accreted to the Central Meseta during the Late Silurian-Early Devonian. – (**A**): Schematic map of the Sehoul Block and adjacent areas, with location of cross-sections (C) and (D), after El Hassani (1994b); see Fig. 3.16 for location. – (**B**) Transgression of the Upper Silurian-Lochkovian (s-di) red beds and limestones upon the Taicha granite (see location in cross-section D). The granite (Gr), dated at c. 430 Ma, and its transgressive cover are included in the Rabat-Tiflet Fault Zone (RTFZ) which overthrusts the Famennian-Tournaisian olistostromes (dh) of the Sidi Bettache Basin. – (**C**) Detail cross-section in the Sehoul Block, after El Hassani (1994a); the section shows the synmetamorphic folding of the Cambrian phyllites, whose recrystallization is dated at c. 450 Ma. – (**D**) General cross-section, after El Hassani (1994c); Caledonian granite slivers (Rb-Sr 430 Ma age) are involved in the Sehoul Block thrust contact on the Rabat-Tiflet Fault Zone

the Oued Bou Regreg and Oued Tiflet, Fig. 3.21). The occurrence of these ca. 470 Ma-old basic intercalations north of the Central Massif suggests the proximity of a continental margin-ocean transition during the Ordovician. The amphibolite lenses intercalated within the Aouli-Mibladen (Midelt) Massif of the Eastern Meseta may represent the metamorphic equivalent of the Rabat-Tiflet spilites and gabbro. It is worth noting that coeval or slightly younger (450 Ma), magmatic/hydrothermal events have been recognized at the northern border of the Anti-Atlas Domain (felsic dykes of the Siroua Massif; Huch, 1988; albitization of the Precambrian volcanics in the Tamlelt gold ore, Pelleter et al., 2007). All these events testify to the particular importance of the Ordovician extension north of the Anti-Atlas Domain.

3.3.2.2 The Caledonian-Sardic Events and Subsequent Silurian Transgression

Tectonic, metamorphic and magmatic events of Late Ordovician-Silurian age (Caledonian-Sardic events; cf. Sect. 3.4) are recorded only in two regions of the Western Meseta Block, i.e. the Sehoul Block and the Coastal Block/West Meseta Shear Zone, respectively.

The Cambrian-Ordovician metagreywackes and metapelites of the Sehoul Block (Fig. 3.21A, C) are affected by south-verging recumbent folds associated with greenschist-facies recrystallization dated at ca. 450 Ma (K-Ar on white mica). These metasediments are intruded by a 430 Ma-old granite (Rb-Sr WR age), which also outcrops within the Rabat-Tiflet fault zone (Taicha granite) beneath unconformable, Upper Silurian-Lower Devonian arkoses and limestones (Fig. 3.21B). Pillow basalts are locally interbedded within the Upper Silurian deposits. Further east, in the Immouzzer-du-Kandar boutonnière of northwest Middle Atlas, granite pebbles occur in a Lower Devonian (Emsian) limestone, associated with pebbles of Ordovician quartzites, Silurian pelites, and ignimbrites (Charrière & Régnault, 1989). Moreover, Ordovician granodiorites have been cored beneath the Gharb Basin and off El Jadida, suggesting that a large Caledonian segment indeed wraps around the NW Meseta (Fig. 3.16).

Accretion of the Sehoul deformed terrane occurred probably during the Late Silurian, as latest Silurian (Pridoli; El Hassani, 1990, 1994c)-Early Devonian strata overlie unconformably some of the granite slivers and the Ordovician series of the Rabat-Tiflet Zone. The fault zone was reactivated during the Variscan Orogeny, as early as the Late Devonian, as suggested by the occurrence of a foliated, chaotic mélange of Famennian age, including blocks of Middle Devonian limestones, along their common boundary of the Rabat-Tiflet Zone and Sidi Bettache Basin. The origin of the Sehoul exotic terrane is controversial, being either the Acadian belt of the Appalachians or the “South-Caledonian” belt of Europe (Sardinia-Corsica-Internal Alps), where tectonic and magmatic events occurred at about 450 Ma (see Sect. 3.4).

Evidence of Late Ordovician-Early Devonian tectonic events is observed in the Coastal Block/Western Meseta Shear Zone (WMSZ, i.e. the eastern border of the Coastal Block; Figs. 3.16, 3.19). In the Oulad Abbou syncline of the Coastal Block, the Silurian sedimentation begins with Telychian (upper Llandovery) platform deposits unconformable onto Caradocian formations. The Upper Silurian series comprise basalt flows typical of an alkaline intraplate magmatism (El Kamel et al., 1998) coeval with those of the Sehoul/Rabat-Tiflet Zones. In the narrow WMSZ, the variably deformed Cambrian-Ordovician series are overlain unconformably by red beds dated from the Lochkovian (Gedinnian) with typical “Old Red Sandstone” facies. These coarse conglomeratic continental deposits occur, from north to south (Fig. 3.22), in the central Rehamna, central-western Jebilet, at Koudiat Mzoudia = J.Ardouz to the east of Marrakech, and up to the western High-Atlas (Ait Lahsen, Talmakent). This suggests that the eastern border of the Coastal Block was faulted, folded and uplifted during the late Silurian-Early Devonian. This “Caledonian” fault zone remarkably prefigures the Namurian-Westphalian WMSZ.

3.3.2.3 The Devonian Platforms and Basins

In spite of the relative displacements related to the Variscan tectonics, mapping of the unrestored distribution of the Devonian sedimentary facies (Fig. 3.22) enable

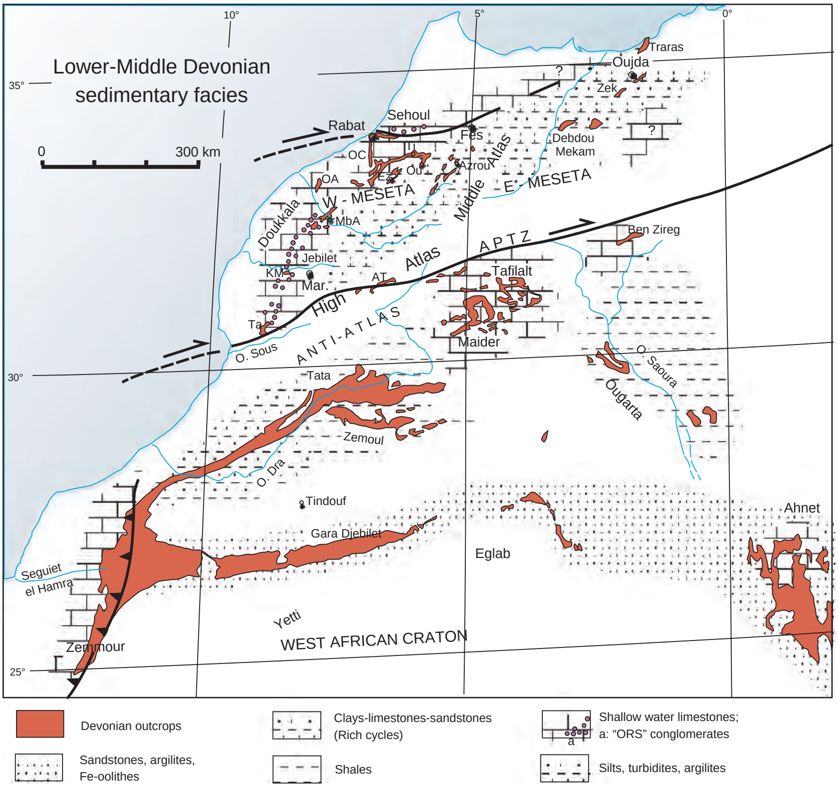

Fig. 3.22 Unrestored distribution of the Lower-Middle Devonian sedimentary facies throughout Morocco and neighbouring areas, after Hollard (1967), Wendt & Aigner (1985), Piqué & Michard (1989), Bouabdelli et al. (1989), El Hassani (1994a), Zahraoui (1994), Fabre (2005), Hoepffner et al. (2005), Wendt & Kaufmann (2006). For clarity, facies extensions are strongly extrapolated. AT: Ait Tamlil; KM: Koudiat Mzoudia; MbA: Mechra ben Abbou; OA: Oulad Abbou; OC: Oued Cherrat; Ou: Oulmes; Ta: Talmakent; Ti: Tilioutine; Zek: Zekkara

us to recognize the occurrence of two major, contrasting paleogeographic domains in the Meseta Domain during that period, i.e. a platform domain in the west and northwest, and a basinal domain elsewhere.

In the northwestern domain, from Tilioutine (northern Central Massif near Oulmes) to Oued Cherrat and Oulad Abbou in the Coastal Block, to Mechra ben Abbou (northern Rehamna), and Talmakent in western High Atlas, the Devonian deposits begins either with Fe-rich argillites interbedded with thin *Scyphocrinites* limestones, or with the conglomeratic red beds described above. These shallow water or continental deposits change progressively upward into massive limestones with minor pelitic intervals, typical for a platform domain. Finally, during the Givetian-early Frasnian, reef mounds develop on the shelf, dominated by tabulata, stromatoporoid and tetracorallia build-ups (Fig. 3.23, columns 1, 2, 8, with references).

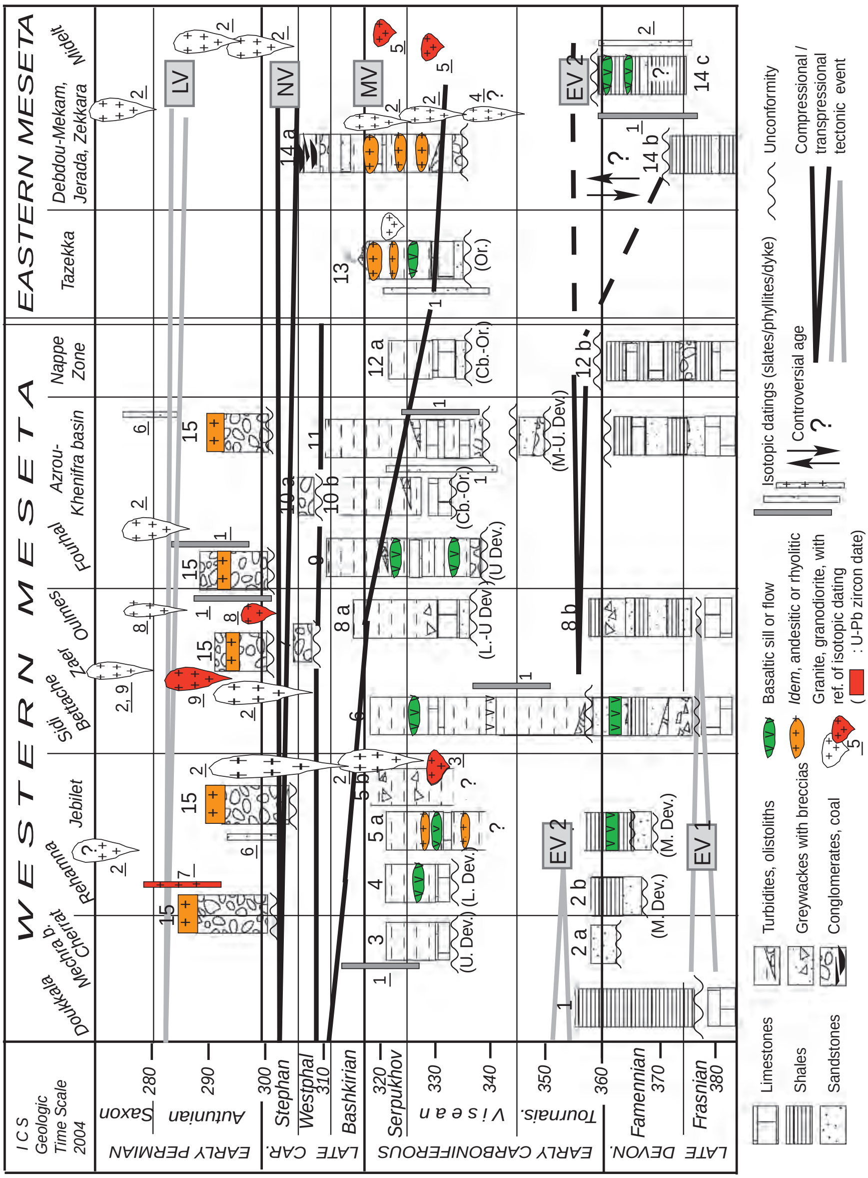
WESTERN MESETA
EASTERN MESETA
ICS Geologic Time Scale 2004
EARLY PERMIAN
LATE CAR.
EARLY CARBONIFEROUS
LATE DEVON.
Saxon
Autunian
Stephan
Westphal
Bashkirian
Serpukhov
Visean
Tournais.
Famennian
Frasnian
Doukkala
Mechra b. Cherrat
Rehamna
Jebilet
Sidi Bettache
Zaer
Oulmes
Fourhal
Azrou-Khenifra basin
Nappe Zone
Tazekka
Debdou-Mekam, Jerada, Zekkara
Midelt
LV
NV
MV
EV 2
EV 1
Limestones
Shales
Sandstones
Turbidites, olistoliths
Greywackes with breccias
Conglomerates, coal
Basaltic sill or flow
Idem, andesitic or rhyolitic
Granite, granodiorite, with ref. of isotopic dating
(: U-Pb zircon date)
Isotopic datings (slates/phyllites/dyke)
Controversial age
Unconformity
Compressional / transpressional tectonic event

Fig. 3.23 (Continued) Stratigraphic record of the Variscan orogenic phases in the Meseta Domain (columns 1–15) and isotopic datings of the related metamorphic and magmatic events (references *1*–*9*). The data are presented along a profile approximately transverse to the belt (see trace A–B, Fig. 3.16 for location). The Sehoul exotic terrane is not represented (see Fig. 3.19). *Tectonic events or phases*: EV 1/2: Eo-Variscan 1/2; MV: Meso-Variscan; NV: Neo-Variscan; LV Late Variscan. *References concerning the stratigraphic columns*: 1: Doukkala: Echarfaoui et al. (2002); 2: Mechra ben Abbou (a), Oued Khibane (b): Destombes et al. (1982); 3: Ben Slimane: Piqué & Michard (1989); 4: Gada Jenabia: Destombes et al. (1982), El Archi et al (2004); 5: Sarhlef (a), Kharrouba (b): Bordonaro et al. (1979), Izart et al. (1997), Essaifi et al. (2004); 6: Sidi Bettache, Mdakra: Piqué & Michard (1989), Vachard & Fadli (1991); 7: Sidi Kassem: Chèvremont et al. (2001); 8: Tougroulmès, Tilioine: Tahiri & Hoepffner (1988), Vachard & Tahiri (1991), Walliser et al. (1995), Izart et al. (2001); 9: Fourhal, Agourai: Berkhli et al. (2000), Ben Abbou et al. (2001); 10a: Migoumess: Berkhli and Vachard (2002); Mouchenkour; Vachard et al; (2006); 10b, Zaian, J. Hadid: Bouabdelli and Piqué (1996); 11: Bou Khadra, Bouechot, Adarouch: Bouabdelli et al. (1989), Huvelin & Mamet (1997), Ben Abbou et al. (2001), Ouarhache et al. (1991); 12a: Mrirt-Khenifra nappes: Piqué & Michard (1989), Piqué (2001); 12b: Bou Nebedou, Ziar nappe: Hollard (1967); 12a: Gara de Mrirt: Walliser et al. (2000); 13: Chalot-Prat & Cabanis (1989); 14a: Berkhli et al. (1993), Torbi (1996); Herbig et al. (2006); 14b: Piqué & Michard (1989), Hoepffner et al. (2005); 14c: Ouali et al. (2001), Raddi et al. (2008); the age of the protoliths of the Midelt schists and metagabbros is controversial (Cambrian or Devonian-Tournaisian ?); 15: Broutin et al. (1987), Youbi et al. (1995), Saber et al. (2001, 2007), Saidi et al. (2002), Hmich et al. (2006). *References concerning the isotopic dating*: *1*: K-Ar white mica, Huon *in* Piqué & Michard (1989); *2*: Rb-Sr, Clauer et al. (1980), Mrini et al. (1992); *3*: U-Pb on zircon, Essaifi et al. (2003); *4*: Ajaji et al. (1998); *5*: U-Pb on zircon, Oukemeni et al. (1995); *6*: ^{39}Ar-^{40}Ar, Watanabe (2002); *7*: U-Pb, Baudin et al. (2003); *8*: U-Pb on zircon (granite) and K-Ar muscovite (leucogranite), Baudin et al. (2001); *9*: U-Pb on zircon, Chèvremont et al. (2001). The vertical extension of the symbols (granitoid plutons, dykes or slates/schists) refers to the error bar of dating. The granites dated by U-Pb on zircon are shown in red. The oldest plutons are granodiorites, whereas the youngest are peraluminous granites/leucogranites

By contrast, a deep water basin extended during the same period southeast of the shelf, the limits of which are unclear owing to the Variscan deformation. Basinal facies have been described recently in the Zaer-Oulmes anticlinorium (Ksiksou Fm.; Razin et al., 2001), consisting of Silurian-Lochkhovian shales passing upward to distal turbidites of Emsian-Famennian age. Likewise, at the bottom of the Azrou Carboniferous basin, the J. Bouechot Devonian anticline displays thick pelitic intervals with minor interbedded limestones (Fig. 3.23, col. 11). Basinal sequences also occur in the Ziar nappes (col. 12b) which overlie the Azrou Basin, and extend southwestward up to eastern Jebilet and Ait Tamlil. They consist either of platy limestones with radiolarian cherts and local effusive mafic intercalations, or of turbidite deposits including slumped beds and sedimentary breccias. The basinal facies extend eastward up to the Beni Snassen, Debdou-Mekam and Traras massifs, which display Early Devonian-Frasnian turbiditic metapelites. The occurrence of a carbonate platform north and/or south of this eastern basin may be postulated in order to account for the Middle Devonian reef limestones reworked within the Upper Visean formations of the Oujda region (Traras, Zekkara; Fig. 3.23, col. 14a). Overall, this paleogeographic setting suggests an extensional or transtensional regime in most of the Meseta Domain during the Devonian interval.

3.3.3 Orogenic Evolution

References: The classical references are recalled in Sect. 3.3.1. The main recent references are indicated in the legends of the figures (especially Fig. 3.23), with additional references as follows: Kharbouch et al. (1985), Lagarde & Michard (1986), Vachard et al. (1991), Aghzer & Arenas (1995), Youbi et al. (1995), Huvelin & Mamet (1997), Ouanaimi & Petit (1992), El Hassani et al. (1994), Bouabdelli & Piqué (1996), Razin et al. (2001), Chèvremont et al. (2001), Baudin et al. (2001, 2003), Ben Abbou et al. (2001), Echarfaoui et al. (2002), Saidi et al. (2002), Roddaz et al. (2002), Soualhine et al. (2003), Houari & Hoepffner (2003), Hmich et al. (2003), Essaifi et al. (2004), Bennouna et al. (2004), Lahfid et al. (2005), Hoepffner et al. (2006), Saber et al. (2007), Graham & Sevastopulo (2007). For more specific information on the structure and/or geochemistry of the Variscan granites, see Lagarde et al. (1990), Bouchez & Diot (1990), Mrini et al. (1992), Oukemeni et al. (1995), Gasquet et al. (1996), Ajaji et al. (1998), Watanabe (2002), Essaifi et al. (2003), El Hadi et al. (2003, 2006b), Boummane & Olivier (2007), with references therein.

3.3.3.1 Late Devonian Events (Eo-Variscan 1)

The transgression of sandy and muddy Famennian strata upon a faulted Middle Devonian shelf has long been known in the Mechra ben Abbou region (Oued Khibane; Fig. 3.23, col. 2; Fig. 3.24A). Recently, well data and seismic lines from

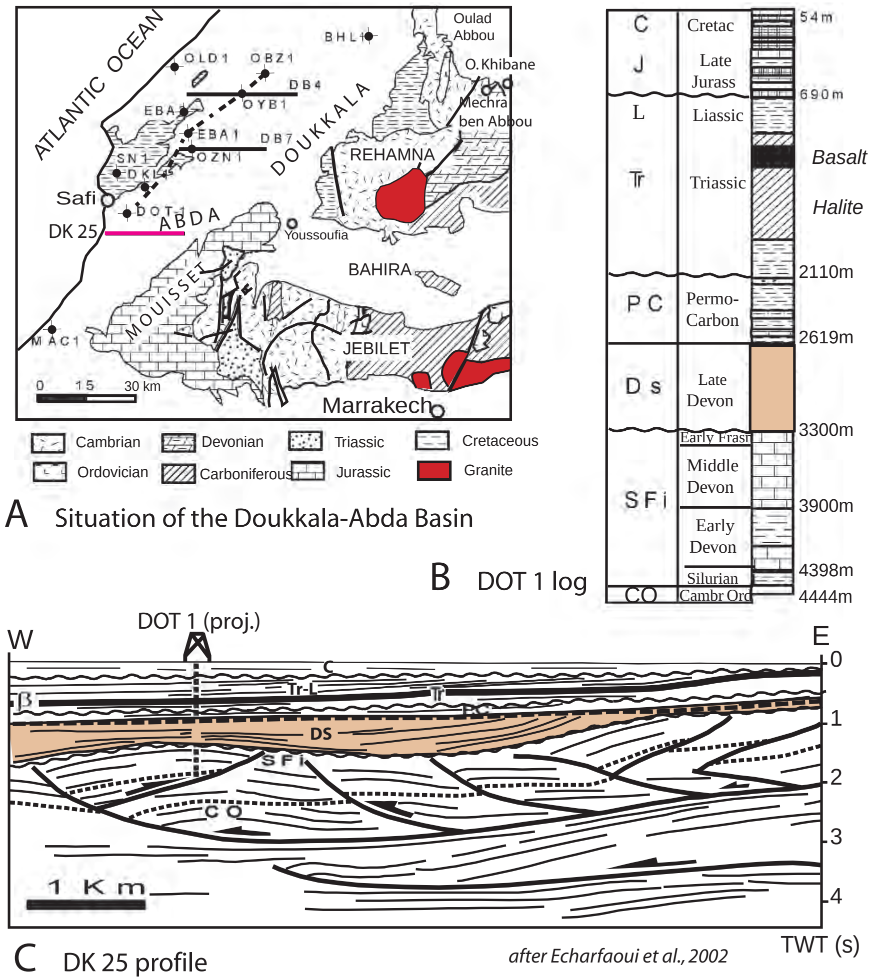

Fig. 3.24 The Silurian-Devonian Doukkala Basin (western Coastal Block), after Echarfaoui et al. (2002), modified. – (**A**) Schematic map with location of the main seismic lines (ONHYM) and wells. – (**B**) Stratigraphy of DOT-1 well. – (**C**) Line drawing of seismic line DK 25. Seismo-stratigraphic correlations are based on eleven industrial wells (e.g. DOT 1, shown in **B**) and a seismic reflection grid totalising about 4000 km. The unconformity between the Upper Frasnian-Famennian beds and the deepest horizons is shown. The geodynamic interpretation of the E-W shortening of the pre-Upper Frasnian formations is debatable (see text)

the Doukkala Basin (Fig. 3.24A–C) have been used to unravel the occurrence of a regional unconformity between the Upper Frasnian-Famennian beds and the underlying Ordovician-Silurian-Lower to Middle Devonian series. The latter series are affected by kilometre-scale open folds bounded by conjugate reverse faults, the formation of which could be correlated with the Eastern Meseta Late Devonian orogenic event (Echarfaoui et al., 2002). The geometry of the Doukkala structures is reminiscent of the toe structures of a detached sedimentary prism which, in fact, could be either a compressional accretionary prism or a mega-slump formed in extensional-transtensional conditions. The latter interpretation seems more reconcilable with the coeval opening of the neighbouring Sidi Bettache Basin.

Indeed, the Sidi Bettache Basin opened south of the uplifted Sehoul Block, and east of the Coastal Block shoulder during the late Frasnian-Tournaisian, as shown by the accumulation of mudflows with olistoliths of Devonian reef limestones (Biar Setla conglomerates) in the Famennian and Tournaisian beds of the basin, particularly in those close to its northwestern borders (Figs. 3.21D and 3.23, col. 6). Alkaline trachy-basalt flows are intercalated in the Famennian greywackes of the same regions, and could have equivalents in the tuffitic layers of the Central Jebilet (col. 5a). Likewise, activity of Famennian paleofaults can be inferred in the Zaer anticlinorium (Tiliouine, Moulay-Hassane, and Ezzhiliga; Tahiri & Hoepffner, 1988).

A pull-apart mechanism has been suggested to account for the opening of the Sidi Bettache Basin, coeval with the Eo-Variscan shortening of the Eastern Meseta (Piqué, 1979, 2001). The concept of an internal Eo-Variscan belt in the Eastern Meseta was introduced in the 1980's (Piqué & Michard, 1989) based on three K-Ar dates (372 ± 8, 368 ± 8, and 366 ± 7 Ma) obtained on fine mica fractions from low- to very low-grade phyllites of Eastern Meseta (Midelt and Debdou-Mekkam; Clauer et al., 1980; Huon et al., 1987). These dates must be taken with caution. Taking into account the abundance of clastic micas in the dated phyllites, and their low-grade of metamorphism, a slightly younger, Tournaisian-Early Visean age of recrystallization as in the eastern part of Central Meseta could be postulated hypothetically. This would be consistent with the lack of Late Devonian compressional structures between the Eastern Meseta and the Sidi Bettache Basin (e.g. pelagic Famennian sequences of the Ziar nappe conformable onto the Frasnian beds at Dchar Ait Abdallah and Bou Nebedou; Fig. 3.23, col. 12b).

3.3.3.2 Tournaisian-Early Visean Events (Eo-Variscan 2)

In the Sidi Bettache Basin, the Tournaisian-Early Visean deposits (goniatite shales, sandy turbidites) overlie the Famennian greywackes almost continuously. In contrast, on each side of this basin, Middle to Upper Visean tidal calcarenites and conglomerates unconformably overlie the pre-Visean series, which often preserve Upper Devonian formations. Therefore, these sections (Mechra-ben-Abbou, Oulmes; Fig. 3.23, col. 3, 8) were affected by Tournaisian-Early Visean folding event(s), labelled hereafter the "*Eo-Variscan 2*" events.

Whenever Upper Devonian sequences are lacking, incomplete or uncertain, one may speculate whether the age of the pre-Upper Visean folding is Eo-Variscan 1 or 2. This is the case in the Eastern Meseta, where the protoliths of the Debdou-Mekkam metapelites and meta-turbidites, determined by palynomorphs and confirmed by macrofloras in the Traras, are Early Devonian-Frasnian in age (Fig. 3.23, col. 14b), and where the Midelt quartz-phyllites and amphibolites and Mougueur shales are undated (either Cambrian-Ordovician or Upper Devonian-lower Tournaisian; col. 14c). By contrast, in the eastern part of the Azrou-Khenifra Basin, Upper Tournaisian conglomeratic sandstones are preserved beneath the Mid-Upper Visean onlap, and the folded sequences beneath the Upper Tournaisian sandstones are dated from the Famennian (col. 11). Thus, the Eo-Variscan 2 folding event must be regarded as Early Tournaisian in this area.

Eo-Variscan folds are dominantly W-verging (Figs. 3.16, 3.18A and 3.25A), and seem to vanish west of the Zaer-Oulmes anticlinorium, in the Sidi Bettache Basin, except for the Doukkala structures discussed above. Eo-Variscan low-grade greenschist facies metamorphism and associated shallow dipping axial-plane foliation are observed in the Debdou-Mekkam, Tazekka, and Khenifra nappes, and in the basement of the Azrou-Khenifra Basin (Cambrian-Ordovician of the Zaian anticlinorium and Bou Guergour window; Fig. 3.23, col. 14–13, 12 and 10, respectively). Metamorphic foliation is lacking in the Devonian series of the Ziar nappes, as well as in the autochthonous Devonian anticlines in front of the nappes (J. Bouechot/J.Bou Khadra, col. 11; El Hammam, col. 8), suggesting that the Upper Paleozoic series were detached from the Lower Paleozoic on the Silurian black shales. The Eo-Variscan metamorphism reaches the muscovite-biotite greenschist and epidote-amphibolite facies only in the Midelt (Aouli-Mibladen) boutonnière of the High-Moulouya valley.

3.3.3.3 Late Visean-Early Westphalian Events (Main Variscan Phase)

The Late Visean-Early Westphalian events can be regarded as the main Variscan phase as they involve, (i) strong shortening of the whole Meseta Domain, including nappe emplacement in the eastern areas; (ii) widespread, intrusive and effusive magmatism, both pre- and syn-tectonic (gabbroic and felsic magmas, respectively), and (iii) high-temperature, low-pressure metamorphism. Moreover, by the end of this period, the Meseta Domain was totally emergent.

Eastern Meseta

From the Early Visean onward, the Eastern Meseta behaves as an internal, relatively rigid magmatic arc. Andesitic basalts, andesites and trachytes emplaced within the Late Visean-Serpukhovian series (Fig. 3.23, col. 13, 14a). Large granodiorite plutons intruded the deeper levels, being surrounded by contact aureoles within which the Eo-Variscan foliation is tilted and sealed by random recrystallizations. The Aouli-Mibladen granodiorite and granite complex yielded zircon U-Pb ages of 333 ± 2 Ma and 319 ± 2 Ma facies, respectively (Fig. 3.23, ref. 5). This magmatism displays calc-alkaline to shoshonitic characters. In the Tazekka Massif (col. 13), Namurian volcanics display evidence of acid-basic magma mixing, consistent with its emplacement on top of a thickened continental crust. Parts of the arc were emergent at this time (Tazekka), whereas other parts remained under shallow marine conditions up to the Middle Westphalian (Jerada coal basin). Late Visean synsedimentary faulting is indicated by olistostromes in the basement

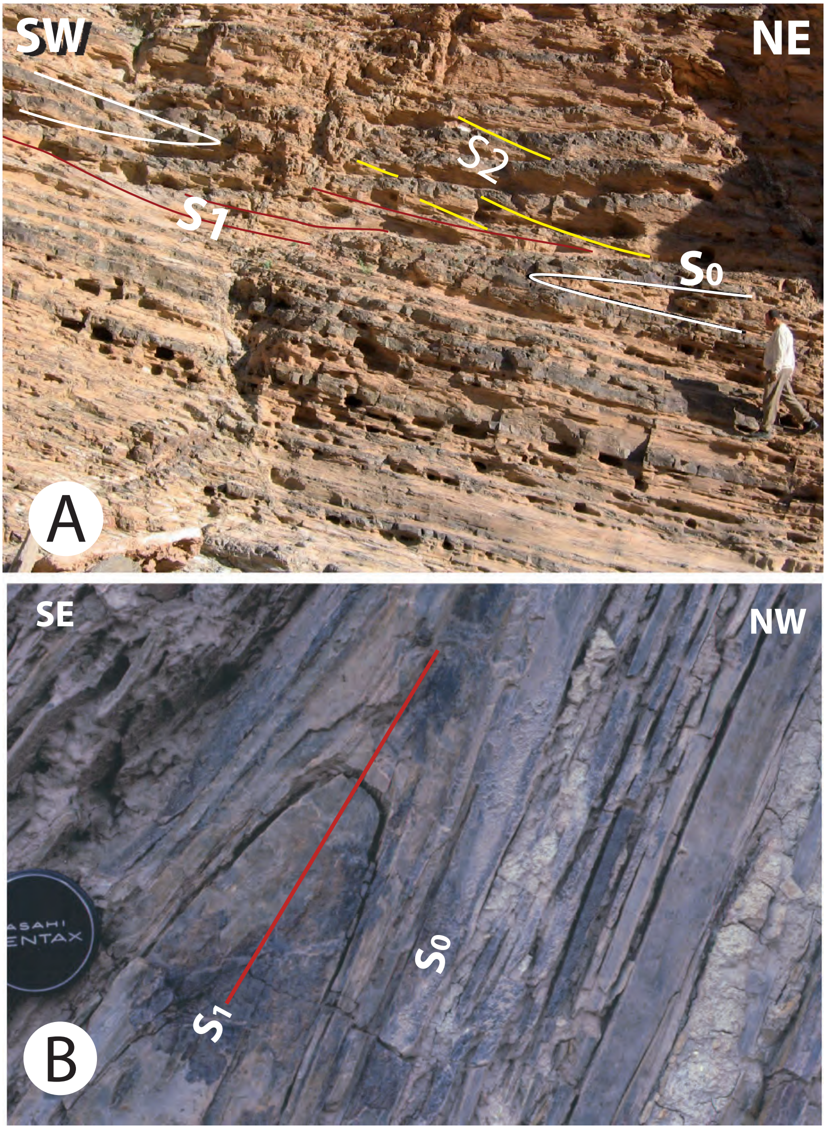

Fig. 3.25 Examples of Variscan folds. – (**A**) Eo-Variscan recumbent folds in epidote-amphibolite schists from the Aouli-Mibladen inlier (Midelt massif) of Eastern Meseta. The age of the protolith is debatable (Cambrian-Ordovician or Upper Devonian?). Note the person on the right for scale. – (**B**) Variscan fold in low-grade metaturbidites of the Fourhal Basin (Upper Visean-Namurian). S_0: bedding; S_1: syn-metamorphic foliation (**A**), or slaty cleavage (**B**); S_2: crenulation cleavage (**A**)

of the Jerada Basin (Zekkara-Jorf Ouazzene). The lithostratigraphic succession in the latter basin is consistent with sedimentation in a back-arc basin (Herbig et al., 2006).

Western Meseta. I: Deformation

In the Nappe Zone, i.e. the west border of the Eastern Meseta arc, two groups of nappes overlie the Lower Carboniferous deposits of the Azrou-Khenifra-Fourhal Basin (Fig. 3.18B). The lowest, Ziar nappes (Bou Khemis, Bou Agri) consist of Devonian material; they were emplaced during the Late Visean in syn-sedimentary conditions, being associated with olistostromes up to the Eastern Jebilet and Ait Tamlil massifs. The uppermost, Mrirt-Khenifra nappes consist of Ordovician metapelites unconformably overlain by Upper Visean beds (J. Aouam), similar to the sequences from the tectonic windows below (J. Bou Guergour or Zaian antiforms), which suggests a proximal origin.

The sedimentary fill of the Azrou-Khenifra Basin and its Lower-Middle Paleozoic basement have been deformed in submarine, synsedimentary conditions. Their overall structure is that of a NW-verging tectonic wedge of duplexes detached on the Middle Ordovician slates and/or Silurian black shales (Fig. 3.26). At the front of the thrust-propagation anticlines, the synclinal sub-basins were filled up with turbidites and debris flows whose age is progressively younger toward the west (Fig. 3.23, col. 11–9). These thin bedded deposits were deformed into tight, W-verging flexural folds with slaty axial plane cleavage (Fig. 3.25B). To the west, the Sidi Bettache Basin played the role of a foredeep with respect to the eastern thrust wedge, and it was not deformed before the end of the Early Namurian, as probably does the Coastal Block.

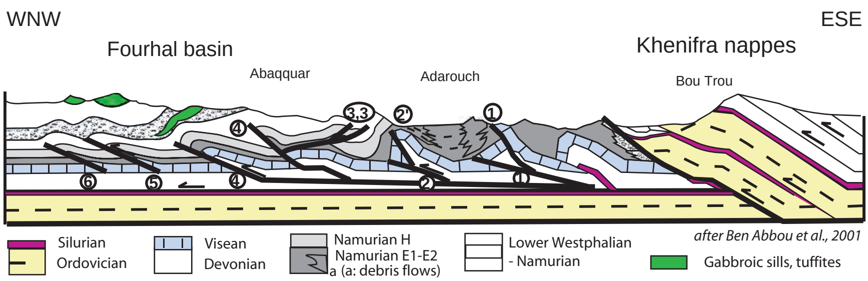

Fig. 3.26 Cross-section of the Azrou-Fourhal synclinorium, after Ben Abbou et al. (2001), modified. Two décollement horizons were active during shortening, i.e. the Lower Ordovician slates and the Silurian black shales. The pre-Visean folds are not shown in this schematic figure, aimed at illustrating the synsedimentary, northwestward propagation of the deformation front from 1 to 6. Fault chronology is based on the age of the associated sedimentary fans and syntectonic facies (debris flows, olistostromes)

Western Meseta. II: Magmatism

Basaltic magmatism occurred in the Western Meseta basins in the form of tuffites, pillow lava flows, and sills of dolerite and gabbro. In the Fourhal Basin, volcanic activity was syntectonic (Namurian-Early Westphalian), and the magma shows a calc-alkaline tendency. In the Sidi Bettache Basin, pillow lavas are intercalated in the Lower Visean turbidites, and display transitional, alkaline-tholeitic geochemical characters. Further south, in the Central Jebilet Basin, tholeitic gabbros and dolerites emplaced as dykes and sills in the Late Visean basinal sediments. The 400 m thick mafic-ultramafic intrusion of Koudiat Kettara (Fig. 3.27), associated with pyrrhotite-chalcopyrite deposits, is particularly evocative of the South Iberian Pyrite Belt (e.g. Onézime et al., 2003).

Felsic magmatism also occurred in the Central Jebilet Basin, in the form of early quartz-keratophyre flows and younger, calc-alkaline granophyric sills and dykes, broadly coeval with or slightly younger than the mafic magmas, as suggested by acid-basic magmatic breccias (magma mixing). A granophyre sill belonging to this early felsic magmatism yielded a U-Pb zircon date at 330 ± 1 Ma (Fig. 3.23, col. 5b, ref. 3). This intrusion occurred shortly before the regional folding and penetra-

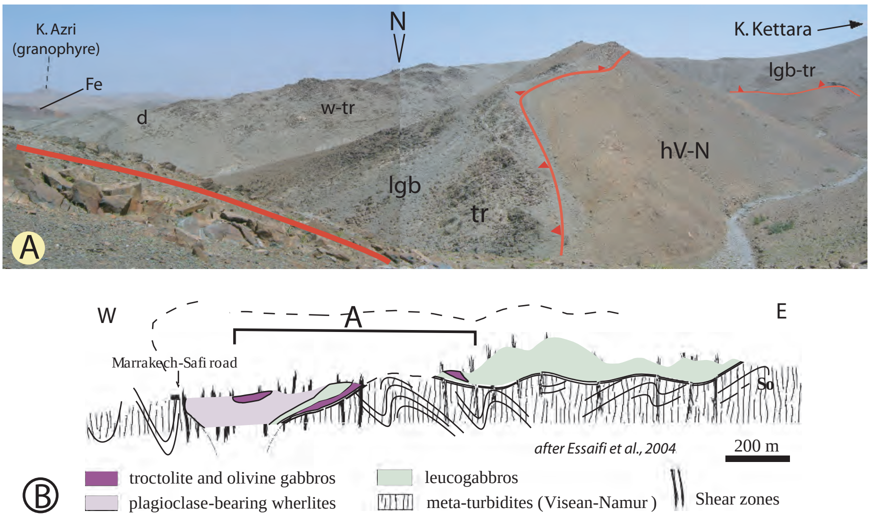

Fig. 3.27 Tholeitic magmatism in the Devonian-Carboniferous basin of the Central Jebilet: example of the Kettara sill, 25 km NW of Marrakech (see Fig. 3.16 for location). – (**A**) View of the basic-ultrabasic sill, intrusive in the Sarhlef turbidites (Upper Visean-Namurian). tr: troctolite, olivine gabbro cumulates; lgb: leucogabbros; w: wherlite; d: folded microgabbro dyke; hV-N: Upper Visean-Namurian metaturbidites cropping out in an anticline window. Fe: gossan associated to the pyrrhotite-chalcopyrite mineralization. – (**B**): Cross-section of the Koudiat Kettara structure, after Essaifi et al. (2004), with location of (**A**). S_0: bedding folded and crosscut by nearly vertical slaty cleavage

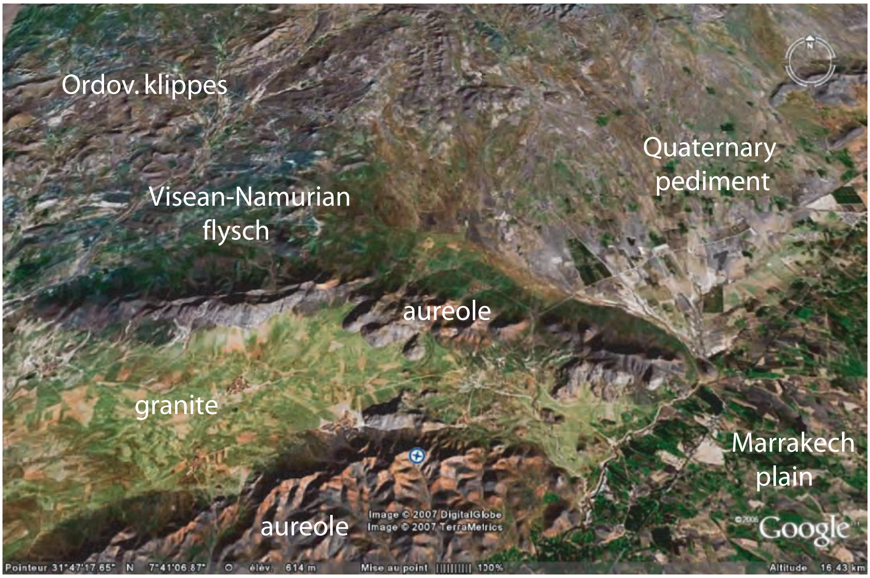

Fig. 3.28 The Ouled Ouaslam laccolith of granite in the eastern Jebilet massif (slightly tilted Google Earth satellite view). The tongue-shaped laccolith grew eastward from a western injection zone located along an early fault (not shown). However, the granite emplacement predates the main, synmetamorphic folding phase (Boummane & Olivier, 2007)

tive deformation. The corresponding rock shows a peraluminous character reflecting crustal anatexis.

The emplacement of the large Ouled Ouaslam laccolith (Fig. 3.28) also predates the shortening deformation as its magnetic fabric is cross-cut by the regional cleavage and shear bands (Boummane & Olivier, 2007). However, a number of the Meseta Domain granites seem to have been emplaced later (cf. Fig. 3.23), i.e. during the Late Stephanian-Autunian (as far as the available datings are reliable). For example, the ascent of the Oulmes granite would have occurred at about 300 Ma (Baudin et al., 2001) by diapirism from ~ 20 km up to ~ 8 km depth after the main regional deformation, and coeval with the latest stages of deformation (Tahiri et al., 2007). Biotite from the Hajar (south of Central Jebilet) and J. Aouam (Central Massif) sulphide districts yielded $^{40}Ar/^{39}Ar$ dates at c. 301 and 280 Ma, respectively (Watanabe, 2002).

Western Meseta. III: Metamorphism

The P-T conditions of metamorphism do not exceed low-grade greenschist-facies conditions during the main Variscan folding (cf. Fig. 3.25B), except in the Western Meseta Shear Zone (WMSZ), from the Central Rehamna to the J. Tichka Massif. In these regions, amphibolites and mica-schists with chlorite-chloritoid

± biotite, or staurolite ± andalusite ± garnet assemblages occur close to the granite intrusions, suggesting recrystallization under a high geothermal gradient related to the syntectonic ascent of the felsic magmas. The axial culmination of the W-verging thrust-fold stack of the WMSZ in the Central Rehamna allows relatively deep units to crop out, and thus garnet-kyanite-staurolite mica-schists are exposed (Fig. 3.29A–D). Their P-T conditions of recrystallization are estimated at about 0.5–0.7 Gpa (15–20 km), 500–550 °C. Peak metamorphism of the WMSZ is not directly dated, but its age lies between the age of the early acidic magmas (330 Ma) and those of the syn- to post-tectonic batholiths (~ 300 Ma).

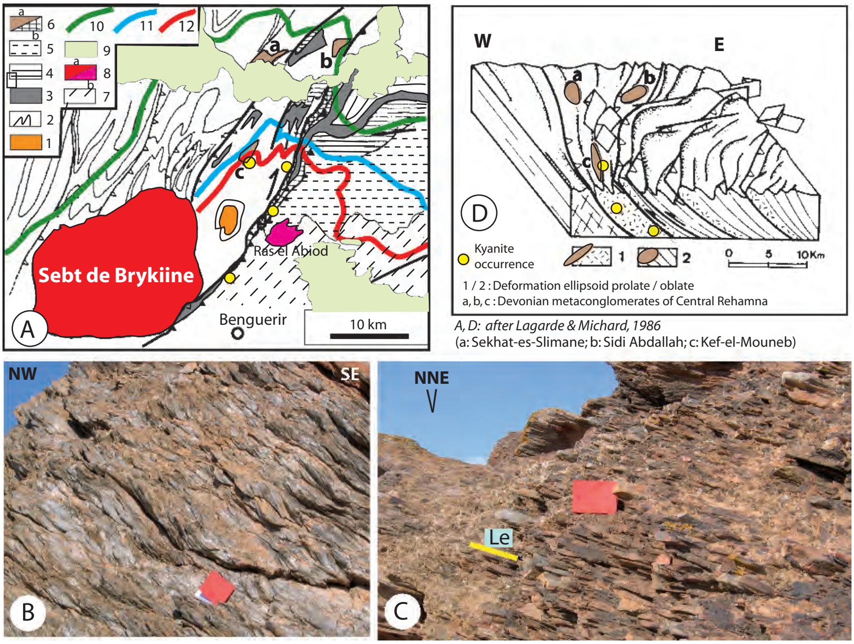

Fig. 3.29 Buchan-type metamorphism in the Rehamna massif (West Meseta Shear Zone; see Fig. 3.16 for location). – (**A**) Schematic map showing the independence of the metamorphic "isograds" with respect to the structural grain, after Michard, ed. (1982), Lagarde & Michard (1986), and Baudin et al. (2001). 1: Orthogneiss; 2: Lower Cambrian marbles, Cambrian-Lower Ordovician metagreywackes and metapelites; 3: Lower-Middle Ordovician quartzites; 4: Upper Ordovician-Silurian metapelites; 5: Devonian metapelites with meta-conglomerate lenses; 6a: Lower Devonian metaconglomerates; 6b: Devonian limestones, quartzites and metapelites; 7: Lower Carboniferous; 8a: granite; 8b: leucogranites; 9: Cenomanian-Turonian marly limestones; 10: chlorite-muscovite *in*; 11: biotite *in*; 12: garnet-staurolite *in*. – (**B**) Sidi Abdallah metaconglomerates (site b in **A** and **D**), where shortening is characterized by an oblate strain ellipsoid with late, top-to-the-west shear planes. The notebook is 15 × 20 cm. – (**C**) Kef-el-Mouneb metaconglomerates (site c), where ductile deformation is characterized by a prolate strain ellipsoid approximately parallel to the regional fold axes; the synmetamorphic lineation (Le) is marked by cigar-like, stretched pebbles of Ordovician quartzite. – (**D**) Idealized kinematic model (Lagarde & Michard, 1986): the intense stretching deformation (type 1 = c) occurs within major shear zones cutting through a thick tectonic pile in an area of high geothermal gradient

Bedding-parallel extensional structures (low angle normal faults) have been observed in the non-metamorphic Ordovician formations of the Zaer anticlinorium, and ascribed to an extensional deformation following the main Variscan contraction and crustal thickening (Razin et al., 2001). Likewise, late orogenic extensional tectonics could have favoured the exhumation of the relatively deep (15–20 km) metamorphic zones of Central Rehamna to the same level as the less metamorphic units (Aghzer & Arenas 1995; Baudin et al., 2003). Even if such a process is likely, its modalities and importance are disputable. In fact, the condensation of the metamorphic isograds results, at least partly, from the high thermal gradients which always characterize the areas with low- to intermediate-pressure, high-temperature metamorphism (Buchan type metamorphism, rather than Barrovian type). In the Rehamna case study, strong gradients are observed not only vertically in the tectonic stack, but also longitudinally in a given unit, for example in the Lower Devonian ferruginous conglomerate unit (Fig. 3.29; compare C and D). Moreover, the observed late metamorphic shear bands (Fig. 3.29B) indicate a late metamorphic overthrust tectonics, and not an extensional regime.

3.3.3.4 Late Westphalian-Early Stephanian Deformation (Neo-Variscan Phase)

The Jerada Basin, which includes Upper Westphalian beds (Fig. 3.23, col. 14a), exposes late Westphalian, E-trending, upright or N-verging folds (Fig. 3.18A). Coeval, but opposite structures occur at the southern border of the Meseta Domain, being overturned toward the Anti-Atlas and Bechar Basin foreland (Tineghir, Tamlelt, and Ben Zireg transects). Likewise, the Upper Westphalian Migoumess formation of the Azrou Basin (col. 10a) is involved in NW-verging thrust tectonics. In contrast, further west in the Zaer anticlinorium, post-Westphalian deformation is SE-verging. This "*Neo-Variscan*" phase is responsible for the emplacement of klippes of Devonian limestones over the Sidi Kassem continental conglomerates (Figs. 3.17, 3.30 and 3.23, col. 7), and of thrusting of the Sidi Bettache sequences onto the Zaer anticlinorium (Tafoudeit nappe). The N-S seismic profile of the Tadla Basin shows also a remarkable thrust of the Late Visean-Namurian flyschs onto the Stephanian-Autunian conglomerates (see Chap. 4, Fig. 4.27).

3.3.3.5 Late Stephanian-Autunian (Late Variscan) Events

Significant tectonic, magmatic and sedimentary events occurred in the Meseta Domain during the Stephanian-Early Permian (Autunian). At that time, this domain formed a large mountainous chain as a consequence of the thickening and heating of the crust which had been intruded by numerous granites. Active shortening resulted in strike-slip tectonics. Conglomeratic molasses accumulated within rhomb-shaped basins and half-grabens scattered throughout the Western Meseta and High Atlas areas (Figs. 3.16 and 3.17). This continental sedimentation began during the Middle-Late Stephanian in the southwestern areas (e.g. Eastern Jebilet, and

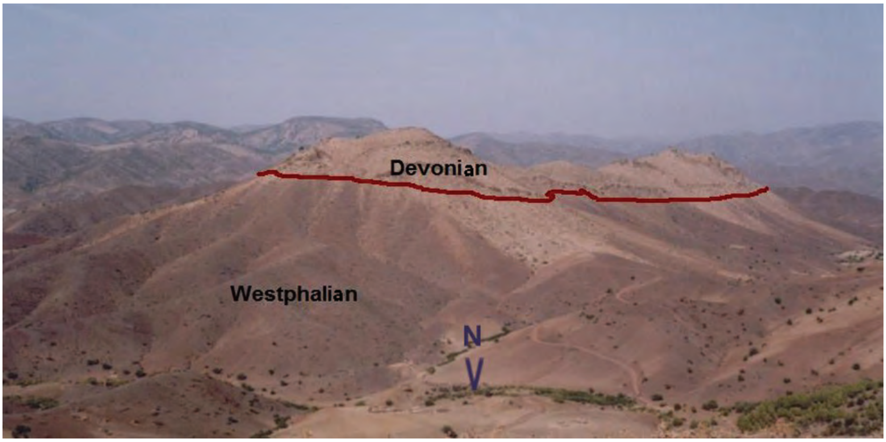

Fig. 3.30 Klippe of Devonian siliciclastic and calci-turbidites (Ksiksou Formation) overlying the Sidi Kassem continental red beds (Westphalian C-D) as a consequence of southward thrusting (Razin et al., 2001). See Figs. 3.16 and 3.17 for location. Photo C. Hoepffner

Ida-ou-Zal/Souss Basin south of the High Atlas), but it generally occurred during the Autunian (Mechra ben Abbou-Daalat, Chougrane, Bou Achouch) or Autunian/Saxonian (Khenifra; Hmich et al., 2006). The orientation of the associated synsedimentary faults suggests a transtensive setting during most of the Stephanian-Autunian, with a dominantly N-S compression, similar to that of the Neo-Variscan phase (Fig. 3.31). A late Autunian rotation in the orientation of the stress field can be inferred from the tilting and mild folding of the Autunian series. This is the last deformation phase of the Variscan cycle, before the Late Permian-Triassic extension (Saber et al., 2007). Late Permian deposits are recognized in the Argana basin of the Western High Atlas, and in the Oued Zat Basin of the Marrakech High Atlas, and in both cases they appear to represent the earliest stage of a rifting process which continued until the Late Triassic (El Arabi, 2007).

Calc-alkaline, mostly granodioritic-leucogranitic magmatism was still active during the late Stephanian-Autunian interval (Fig. 3.23). The persistence of magmatism is proved by the emplacement of widespread trachy-andesite, rhyodacite and ignimbrite flows in the Autunian series (e.g. in the Chougrane Basin, Fig. 3.17). Numerous micro-monzogranite dykes cross-cut the metamorphic zones and granite plutons in the Rehamna and Jebilet massifs. One of these dykes has been dated at 285 ± 6 Ma (U/Pb zircon) in the Rehamna massif (Fig. 3.23, ref. 7), thus marking an upper limit for the age of the granite. In the Western High Atlas Massif, the J. Tichka magmatic complex is fairly well dated from the Early Permian (Sm-Nd isochron at 283 ± 24 Ma, Rb-Sr isochron at 291 ± 5 Ma, best defined for the monzogranite and leucogranite intrusions; Gasquet et al., 1992). This complex includes an exceptional acid-basic association which consists of gabbros, diorites, granodiorites-tonalites, monzogranite and leucogranites, whose isotopic signature implies the existence of contrasting source rocks and the combination of different mechanisms, i.e. the melting of a depleted mantle source, crustal assimilation, fractional crystallization, and

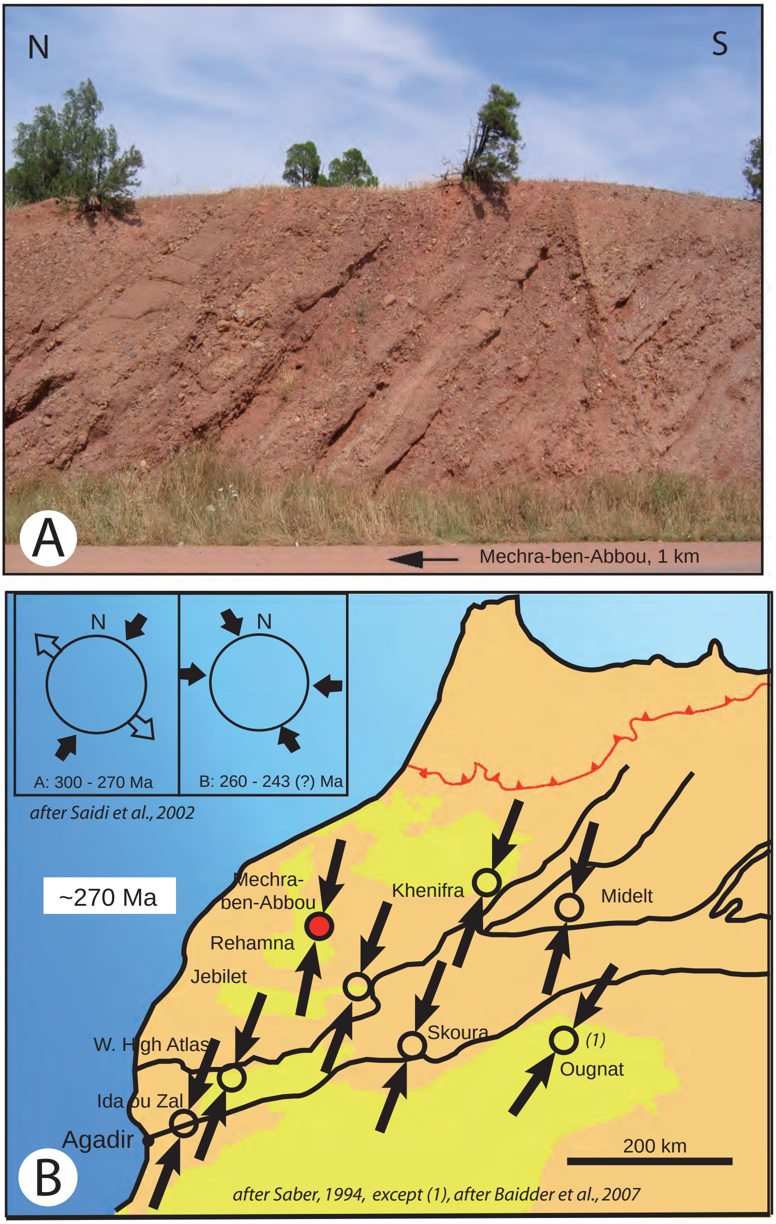

Fig. 3.31 Late Variscan sedimentation and tectonics. – (**A**) Tilted conglomeratic red beds of Stephanian (?)- Autunian age; Mechra-ben-Abbou, Oum-er-Rbia valley north of the Rehamna massif (see **B** for location). – (**B**) Orientation of the maximum compressive stress inferred from the structure of Stephanian-Autunian basins, after Saber (1994) and Saidi et al. (2002), modified, and after Baidder et al. (2008) for the Ougnat area. Initial stage (map and cartouche): transtensive opening of the basins during the Late Carboniferous-Early Permian. Last stage (cartouche): Transpressive deformation during the late Early Permian, prior to the Late Permian-Triassic rifting (Saber et al., 2007; El Arabi EH., 2007)

anatexis of a heterogeneous continental crust (Gasquet et al., 1996). The latter continental basement crops out at Wirgane, some tens of kilometres further east, and appears to be a Pan-African crust including a 625 ± 5 Ma old granodiorite (U-Pb zircon; Eddif et al., 2007).

3.3.3.6 Granite Typology

It is worth noting the presence of a broad geochemical zoning of the Meseta granites (Fig. 3.32). Overall, the Eastern Meseta granites, which postdate an early, Eo-Variscan folding phase, are mainly calc-alkaline, potassic to shoshonitic metaluminous type I granites. According to El Hadi et al. (2006b), they derived from mantle sources enriched by a previous subduction process, either Devonian or Pan-African in age. On the other hand, the Western Meseta granites are generally peraluminous, and correspond to S-type granites where the contribution of an anatectic source is important. This broad distribution evokes that which characterizes supra-subduction settings. Additionally, the Western Meseta is characterized by the occurrence of an early, Late Devonian-Early Carboniferous magmatism with coeval basic and acid magmas, which can probably be assigned to a regional rifting tectonics. The geodynamic interpretation of these data is discussed in the next section.

3.4 Geodynamics of the Moroccan Variscides

References: Geodynamical reconstructions of the Variscide terranes have been proposed in the last decade by Cocks et al. (1997), Crowley et al. (2000), Matte (2001), Stampfli et al. (2001), Stampfli & Borel (2002), Fortey & Cocks (2003), Simancas et al. (2003, 2005). The interpretation of the Moroccan Variscides and their integration in the general scheme of the Caledonian-Variscan evolution has been discussed by Kharbouch et al. (1985), Boulin et al. (1988), Piqué et al. (1990), Khattach et al. (1995), Bouabdelli & Piqué (1996), Piqué (2001), Roddaz et al. (2002), El Hadi et al. (2003), Hoepffner et al. (2005), El Hadi et al. (2006b). The origin and correlations of the "South-Caledonian" fragments included in the Mediterranean Alpine belts have been discussed by Von Raumer et al. (2003), Michard & Goffé (2005), Helbing & Tiepolo (2005), Helbing et al. (2006), Giacomini et al. (2006) (see also Chap. 5).

3.4.1 Displacements of the Meseta Domain with Respect to Gondwana

Certain recent plate tectonic models for Phanerozoic time envisage the "Meseta Block" as a highly mobile terrane with respect to Gondwana, and hence, generally disconnected from the Anti-Atlas (Fig. 3.33). The so-called Meseta Block

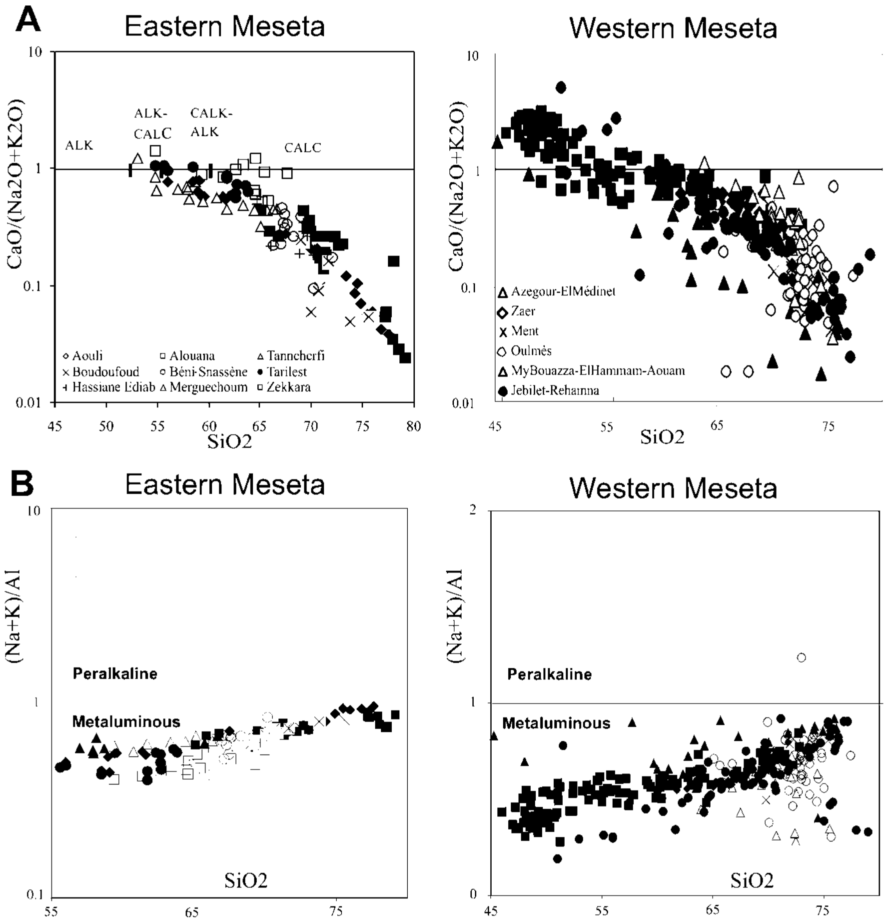

Fig. 3.32 Geochemical typology of the Meseta granites, all ages combined, after El Hadi et al. (2006b). – (**A**) Peacock index ($CaO/Na_2O + K_2O$) versus SiO_2. – (**B**) Agpaitic index (Na + K)/Al versus SiO_2. Symbols in both A and B indicate the varied plutons analysed in the original paper

(or more exactly, the composite Meseta Domain) would have been rifted away from Gondwana at c. 450 Ma, i.e. about 100 Ma after the Avalonian terranes rifting, which would have occurred as early as the latest Neoproterozoic-Cambrian times (Stampfli and Borel, 2002). Then, the "Meseta Block" would have been drifted toward Laurentia, and eventually separated from Africa by a wide Paleo-Tethys Ocean during the Devonian-Early Carboniferous. However, other scenarios of the fragmentation of NW Gondwana have been also proposed with different timing and displacements. Cocks et al. (1997) and Fortey and Cocks (2003) argue that Avalonia split off from Gondwana in early Ordovician time, leaving a widening Rheic Ocean to its south. Crowley et al. (2000) postulate less important

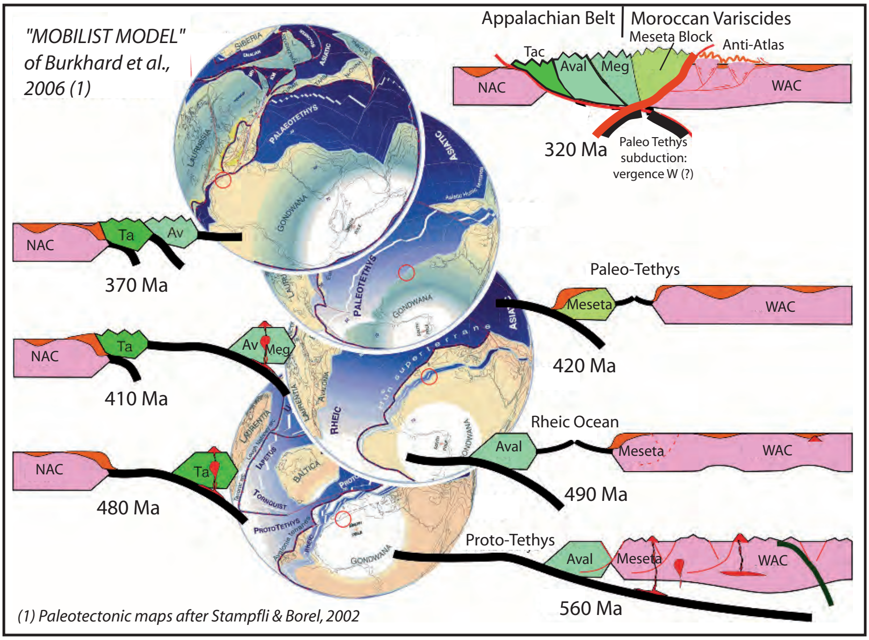

Fig. 3.33 Evolution of the Variscan segments of Morocco during the Paleozoic, after Burkhard et al. (2006), based on the paleogeographic maps of Stampfli & Borel (2002). The figure is modified according to G. Stampfli, personal comm., *in litt*. 2008. The Meseta block is considered as separated from Gondwana (Anti-Atlas) by a Devonian Ocean ("Paleo-Tethys"). Aval, Av: Avalonian terranes; Meg: Meguma; WA/NAC: West African/North American Craton

displacements for the "Armorica Terrane Assemblage" of southwest Europe. They suggest that combined tensional forces and mantle plume convection assisted the early Palaeozoic dispersal of terranes from the N Gondwana margin, resulting in development of an archipelago of related terranes separated by a network of seaways.

Several observations point to a limited mobility of the Meseta Domain (except the Sehoul terrane): (i) the lack of ophiolite in the APTZ – SAF boundary between the Meseta Domain and the Anti-Atlas; (ii) the continuity of the Ordovician lithostratigraphy between the Anti-Atlas and Meseta domains; (iii) the correlations between the Meseta and Anti-Atlas Devonian environments and faunas, which simply suggest a strike-slip movement between both domains (Fig. 3.22). Moreover, the available paleomagnetic data support the hypothesis of restricted displacements of the Meseta with respect to Africa (Khattach et al., 1995). In other words, at least the main part of the Meseta Domain was not involved in great migration during the Lower Paleozoic and could have formed more or less distal fragments of the Gondwana continental margin. On the contrary, the displacements of the Sehoul Block could have been significant before its suturing against the main Western Meseta during the Late Silurian.

During Late Devonian-Late Carboniferous time, i.e. the period of the Variscan Orogeny itself, the displacements of the "proto-Meseta" blocks correspond, first, to the closing up of the eastern and western Meseta blocks, and second, to both the closing up and lateral slip of the whole Meseta complex relative to Gondwana along the Atlas Paleozoic Transform Zone (APTZ)-South Meseta Zone. The obliquity of the fold axes in the Western High Atlas Paleozoic Massif relative to the Tizi n'Test Fault (western APTZ; Fig. 3.16) suggests a dextral displacement of the Meseta Domain along the Anti-Atlas foreland, as M. Mattauer and co-authors noted as early as 1972. Right-lateral movements along the eastern extension of the APTZ have been also observed in the Tamlelt area (Houari and Hoepffner, 2003; Fig. 3.34). The displacement of the Western Meseta relative to the Anti-Atlas likely did not exceed 200 km, assuming a broad initial continuity of the NNE-trending Devonian environmental zones (Fig. 3.22). Thus, the dextral displacement would be less than 100 km in the Tamlelt area, taking into account an E-W internal shortening of the Meseta Domain of about 100 km. The lateral displacement of the Meseta Domain along Africa ended with a nearly N-S collision during the Late Carboniferous-Early Permian. The shortening direction then evolved toward a NE-SW trend during the Early Permian, and finally to a NW orientation (Fig. 3.31).

3.4.2 Geodynamic Interpretation of the Mesetan Variscides

The internal zones of the peri-Atlantic Caledonian-Variscan Belt are characterized by ophiolitic sutures and HP-LT metamorphic recrystallization. Taking into account the above discussion, we may infer that the Meseta Domain, which is characterized by HT-LP metamorphism, corresponds to external parts of the Variscan Orogen, located SE (present coordinates) of the Variscan HP suture zones (Fig. 3.35; Simancas et al., 2005). The latter reconstruction admits close correlations between the Moroccan Meseta Domain and the Iberian Meseta, in line with a number of previous stratigraphic or structural papers (e.g. Matte, 2001; Hoepffner et al., 2005). The Moroccan and Iberian granite typology is in agreement with such reconstructions (El Hadi et al., 2006b). However, the extension of the Ossa-Morena zone into Morocco is not documented. A north-Rheic Avalonian origin is postulated for the Sehoul Block in Fig. 3.35, which is controversial. An alternative origin could be the "South-Caledonian Zone" of the Mediterranean Alpine domain (Von Raumer et al., 2003). These pre-Variscan, Gondwana-derived elements are characterized by a Cambrian-Ordovician rifting evolution (Crowley et al., 2000; Floyd et al., 2000) with widespread Ordovician bimodal magmatism ("Sardic phase"; Helbing et al., 2006). In fact, increasing evidence from the pre-Alpine massifs of the Western Alps, Central Alps, Austro-Alpine and South-Alpine domains suggest that a complete orogenic cycle occurred at least in part of these units (Von Raumer et al., 2003; Michard & Goffé, 2005; Giacomini et al., 2006).

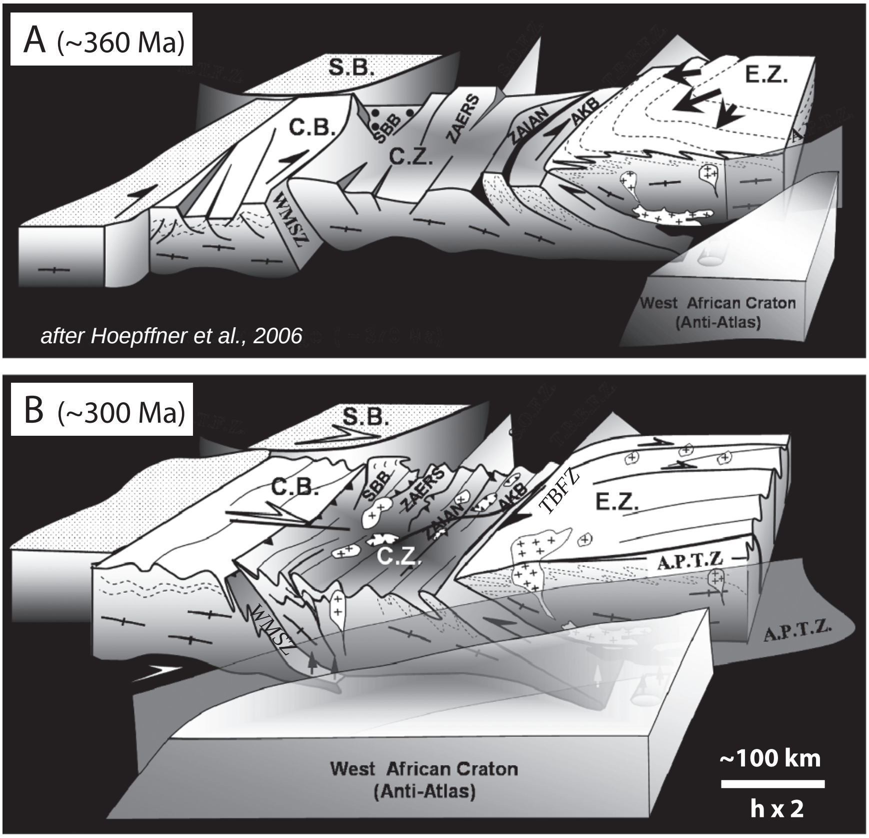

Fig. 3.34 Diagrammatic 3D representation of the Variscan structure and kinematics of the Meseta Domain, after Hoepffner et al. (2006). – (**A**) Upper Devonian-Tournaisian. The Eovariscan phases 1 and 2 affect essentially the eastern Meseta Domain; in the west, strike-slip deformation is accompanied by transtensive opening of pull-apart basins. – (**B**) Late Carboniferous. The main Variscan phase has already affected the entire Meseta Domain (now converted into a Meseta Block) which collides obliquely against the Anti-Atlas. Collisional tectonics continues until the Early Permian. AKB: Azrou-Khenifra Basin; APTZ: Atlas Paleozoic Transform Zone; CB: Coastal Block; CZ: Central Meseta zone; EZ: Eastern Meseta zone; SB: Sehoul Block; SBB: Sidi Bettache Basin; TBFZ: Tazekka-Bsabis and Tizi n'Tretten fault zones; WMSZ: West Meseta Shear Zone

Whatever the major correlations will be, the question arises of how to explain the tectonic, metamorphic and magmatic evolution observed within the Meseta Domain itself. Is it possible to account for this evolution in the frame of a roughly parautochthonous Meseta Domain, as assumed above? Two competing models are currently evoked. The first one considers an entirely intracontinental evolution (Fig. 3.36A). The calc-alkaline magmatism (Eastern Meseta) would relate to deep shear zones within a thickened continental crust/lithosphere, and the western basins opening, coeval with bimodal magmatism, could relate to transtensional pull-apart tectonics. The second model (Fig. 3.36B) invokes a northwest-dipping subduction

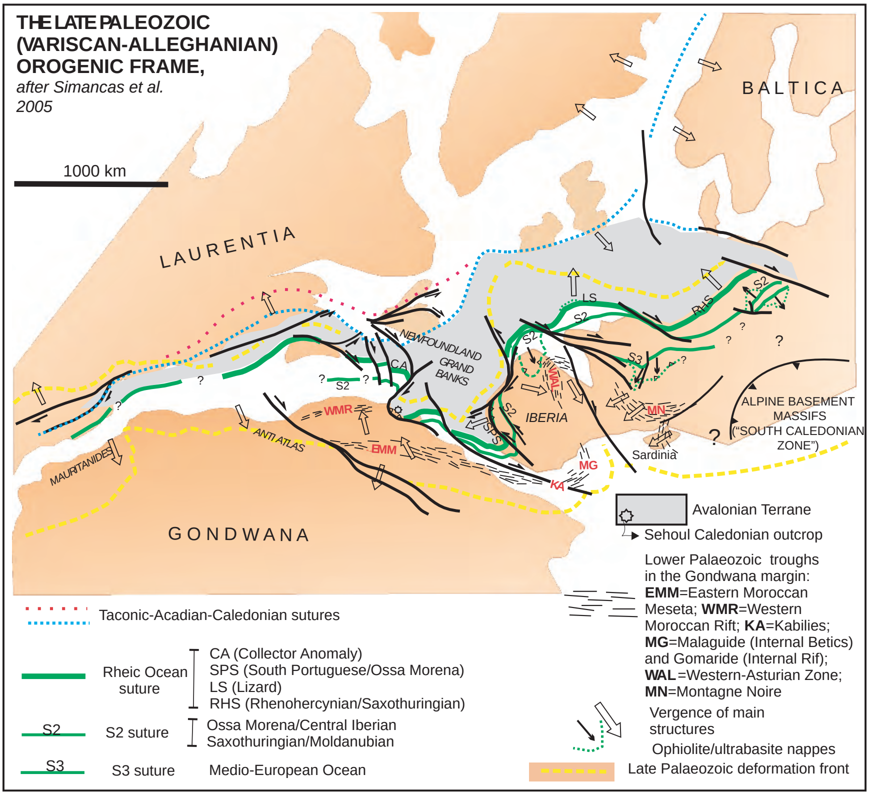

Fig. 3.35 Correlation of the Moroccan Variscides with the peri-Atlantic Caledonian-Variscan belts (Appalachian-Alleghanian belts; Western Europe Hercynian belts), after Simancas et al. (2005), modified. The Moroccan Meseta is opposite to the Meguma Block of Nova Scotia, with the S2 suture (Ossa Morena suture) in between. The S2 suture and the Rheic suture (South Portuguese, Lizard etc.) would pass between the Sehoul Block and the rest of the Meseta. In this reconstruction, the South-Portuguese belt (SPS) has no direct connection with the Devonian-Carboniferous Western Meseta Basin (Sidi Bettache, Jebilet), which is debatable

zone whose hypothetical suture would be located 500 km east of Morocco. The Carboniferous basins would correspond to retro-arc foreland basins, deformation would be retro-arc verging, and magmatism would be triggered by slab break-off.

However, these two models do not take into account the occurrence of at least one southeast-dipping oceanic subduction zone between Africa (Meseta and West African Craton) and Laurentia-Avalonia. This subduction zone is clearly documented in the South-Iberia transect during the Late Devonian-Early Carboniferous (e.g. Onézime et al., 2003; Simancas et al., 2003). The IBERSEIS deep seismic profile gives evidence for such subduction tectonics, and illustrates the detachment of the deformed upper crust above the lower crust. Thence, it is tempting to suggest a new interpretation of the Meseta segment (Fig. 3.36C). Keeping in mind the South-Iberian dataset, we invoke the presence of a shallow, southwest-dipping subduction of the Rheic lithosphere beneath the Meseta Domain during the Late

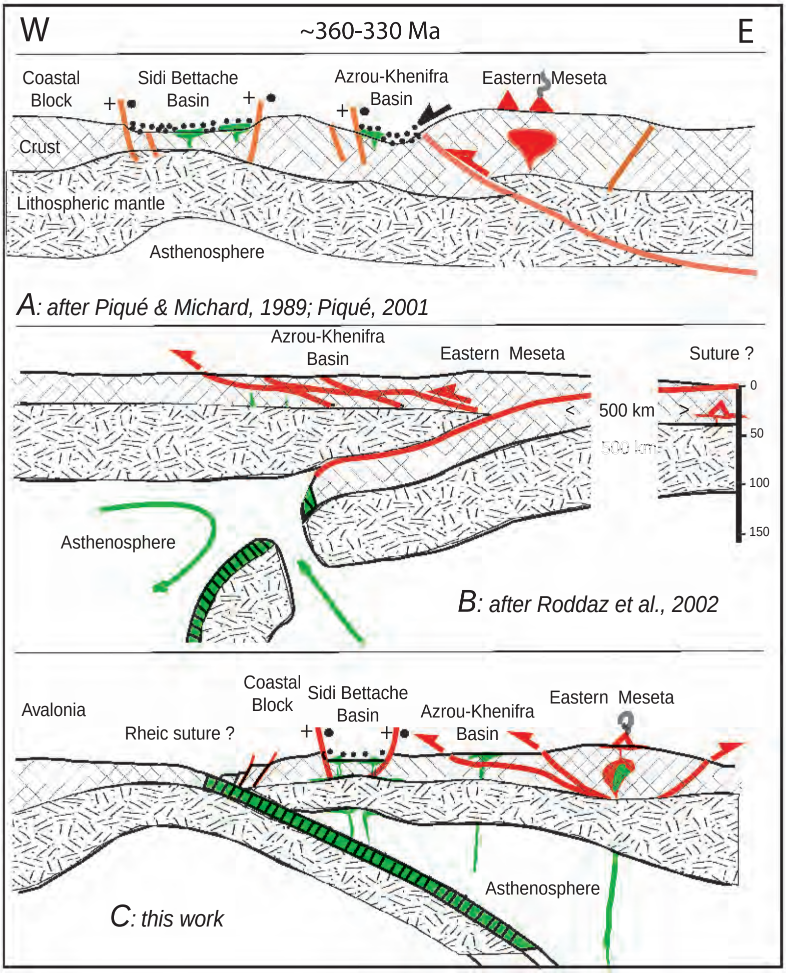

Fig. 3.36 Alternative geodynamic models for the Meseta Orogen during the Late Devonian-Early Carboniferous (all models assume that the Meseta Domain remained always close to Gondwana). – (**A**) Intracontinental orogeny involving pull-apart basins and transcurrent tectonics (e.g. Piqué & Michard, 1989; Bouabdelli & Piqué, 1996; Piqué, 2001; Hoepffner et al., 2005). – (**B**) Hypothetic west-dipping subduction, and retro-arc vergent tectonics in the supra-subduction Meseta block (after Roddaz et al., 2002; see also Kharbouch et al., 1985, and Boulin et al., 1988). – (**C**) This work: east-dipping Rheic subduction; pro-arc and retro-arc tectonics in the Meseta Domain

Devonian-Early Carboniferous. Thus, the Meseta basins might open in a fore-arc setting, due to high-rate subduction roll-back. Magmatism should be bimodal in these basins, whereas calc-alkaline to shoshonitic magmatism is expected in the Eastern Meseta arc. Deformation is fore-arc verging at the front of the postulated eastern Gondwanan buttress. The retro-arc deformation characterizes the Western Anti-Atlas and Mauritanides, as well as some parts of the European Variscan belt (Fig. 3.35). The geochemical zonation of the Variscan granitoids in both Morocco and Iberia is consistent with the latter tectonic model, which also accounts for the similarities between the South-Portuguese Pyrite Zone and the Central Jebilet pyrrhotine-chalcopyrite district.

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Chapter 4
The Atlas System

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This chapter is dedicated to Professor A.W. Bally, who directly or through his PhD students, renewed our understanding of the Atlas System, and to the memory of Mr. R. du Dresnay, the pioneer of modern Atlas studies.

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A. Michard et al., *Continental Evolution The Geology of Morocco*. Lecture Notes in Earth Sciences 116,

4.1 Background

References: Since Michard (1976) synthesis, the Atlas system has been the subject of a number of geological studies synthesized by Stets & Wurster (1982), Laville (1985), Schaer (1987), Jacobshagen et al. (1988), Fedan (1989), Charrière (1990) and Piqué (1994). More recently, industrial subsurface data have been used to document the geometry and subsidence history of associated basins (Le Roy et al., 1997; Beauchamp et al., 1996, 1999; Gomez et al., 1998; Hafid, 2000; Hafid et al., 2000; Frizon de Lamotte et al., 2000; Zizi, 2002; Ellouz et al., 2003; Teixell et al., 2003; Hafid et al., 2006). These studies permitted a reappraisal of the different tectonic events. On the other hand, several geophysical studies during the last decades (Tadili et al., 1986; Makris et al., 1985; Wigger et al., 1992; Mickus & Jallouli, 1999) permitted modelling of the Atlas relief (Frizon de Lamotte et al., 2004; Ayarza et al., 2005; Zeyen et al., 2005; Teixell et al., 2005; Fullea-Urchulutegui et al., 2006; Missenard et al., 2006). The overall plate tectonic setting is presented in the Western Tethys paleogeographic and environmental maps by Dercourt et al. (2000).

4.1.1 Definition

The definition of the Atlas system is given in Chap. 1: fundamentally it is formed by the elevated High Atlas and Middle Atlas mountain belts, which characterise the north-Moroccan provinces and control their physical geography, climatology, hydrology and, consequently, their human geography. The High Atlas trends E to ENE from the Atlantic coast to Algeria where it continues in the so-called Saharan Atlas. The NE-trending Middle Atlas separates obliquely from the main range north of the Central High Atlas (Figs. 1.3 and 4.1). The landscapes of the Atlas Mountains present a great diversity depending on their position on the Mediterranean or Atlantic slopes, which are well watered, or, on the contrary, on the arid continental sides of the belts. The beautiful cedar forests of the Middle Atlas contrast with the esparto moors of the Eastern High Atlas and with the hilly landscapes typical of the Central High Atlas (Fig. 4.2).

From a geological point of view, the Atlas Mountains are fold-belts (Fig. 4.3) developed over a continental basement. The plateaus bordering or included within the belts are also linked to the Atlas system, i.e. the Middle Atlas "Causse" to the north, the "High Plateaus" (Oran Meseta) to the east. These young mountains were uplifted during the Cenozoic and result from the Alpine cycle, as the Rif mountains to the north (Fig. 1.16). However, compared to the Alpine-type, collisional Rif belt, the Atlas system is an intracontinental, autochthonous system, developed over a continental crust which was only slightly thinned during its pre-orogenic evolution. Moreover, the continental basement widely crops out

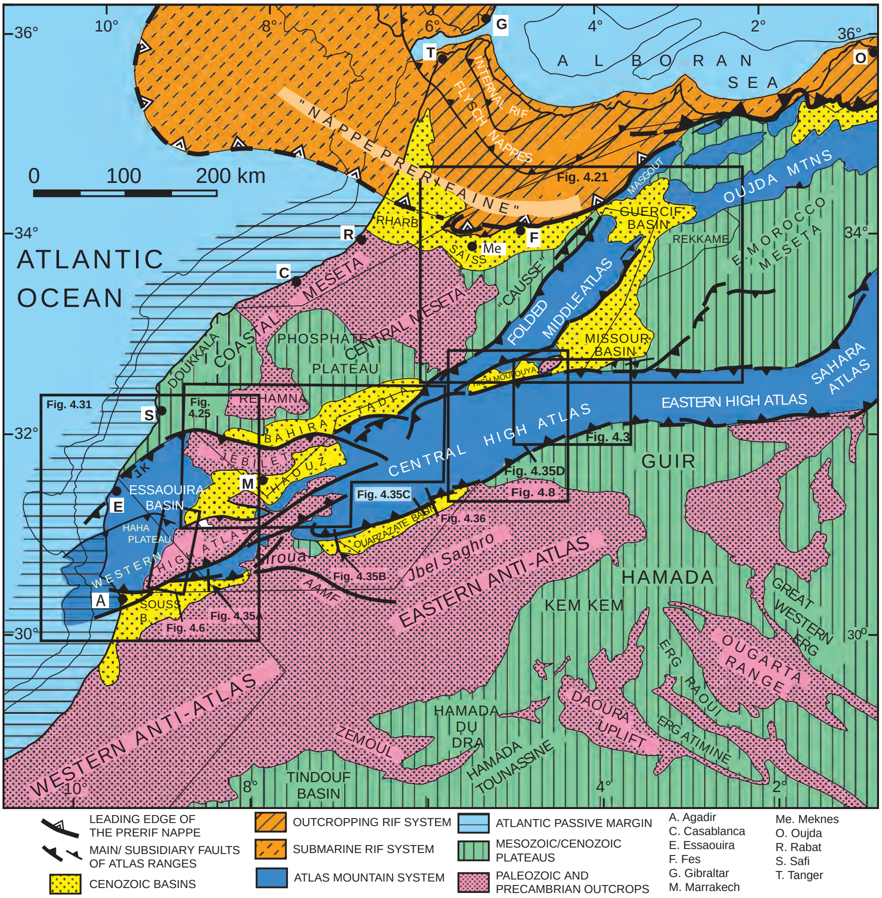

Fig. 4.1 Schematic structural map of Morocco showing the extension of the Atlas system, with location of most of the following figures. After Hafid et al. (2006), modified

in the chain interior and forms its highest peaks south of Marrakech (Fig. 1.4). The chain is almost completely formed over the so-called "Meseta" basement deformed during the Variscan orogeny (see Chap. 3). However, part of the High Atlas incorporates the Anti-Atlas Pan-African basement in the "Ouzellarh Promontory" next to the Siroua Plateau (see Chap. 2, Fig. 2.1). Therefore, the South Atlas Front, which is the present-day southern boundary of the chain, matches only approximately the southern boundary of the Meseta Block (Atlas Paleozoic Transform Zone, APTZ; see Chap. 3). In any case, a Panafrican or Avalonian crust exists everywhere below the Paleozoic units of the Atlas system (cf. Fig. 1.17).

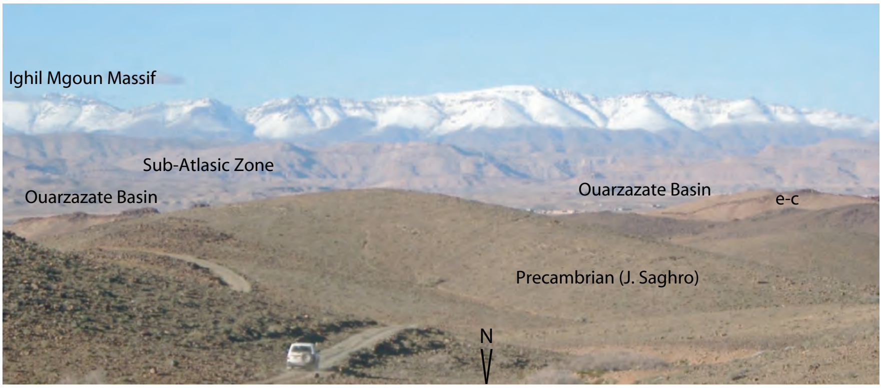

Fig. 4.2 View of the "South Atlas Front": the High Atlas barrier from the south, standing on the north slope of the J. Saghro south of Boumalne-Dadès. Location: see Figs. 4.1 or 4.39. The distance between the snowy Liassic crest (more than 3300 m a.s.l.) and the rocky Precambrian basement in the foreground is about 30 km. The width of the Ouarzazate Basin between the Sub-Atlasic frontal thrust (emergence of the South Atlas Fault, SAF) and the autochthonous Cretaceous-Eocene cover of J. Saghro (c-e, *right*) does not exceed 10 km. Let us imagine this landscape 180 My ago: the sea was on the site of the Atlas. Photograph by J.P. Liégeois

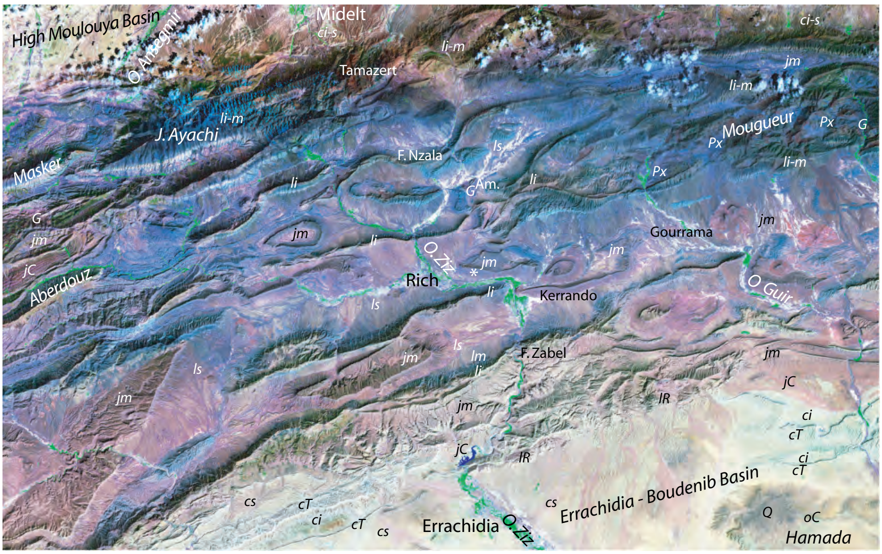

Fig. 4.3 Landsat image of the Central and Eastern High Atlas foldbelt showing the alternation of narrow anticlines and wide synclines. Location: see Figs. 4.1 or 4.39. Px: Paleozoic; li-m, ls: Lower-Middle, and Upper Liassic; jm: Middle Jurassic; jC: Continental Middle-Upper Jurassic; ci: Lower Cretaceous; CT: Cenomanian-Turonian; cs: Upper Cretaceous (Senonian); oC: Continental Oligocene-Miocene; Q: Plio-Quaternary; G: gabbros (Jurassic)

4.1.2 Pre-orogenic and Syn-orogenic Basins; Chapter Contents and Organisation

The geodynamic evolution of the Atlas system comprises two major periods, which will be studied successively. The pre-orogenic period is characterised by the rifting, which affected the Variscan crust, and then by the filling of Mesozoic basins (Sect. 4.2). The orogenic period is characterised by the basin inversion, the shortening of the basement and cover units, and the formation of syn-orogenic basins (Sect. 4.3).

The pre-orogenic period lasts from the Triassic to the Late Cretaceous. An important point is the prefiguration of the future Atlas Mountains in the Liassic paleogeography (Fig. 4.4). At that time, we may distinguish two provinces in the future Atlas system, i.e. an eastern province connected to the Tethys (Central and Eastern High Atlas and Middle Atlas), and a western province opened toward the Central Atlantic (Western High Atlas). The latter corresponds to the Agadir and Essaouira coastal basins, which are parts of the Atlantic passive margin, deformed during the Atlas Orogeny. Both the Tethyan and Atlantic Liassic provinces are basically

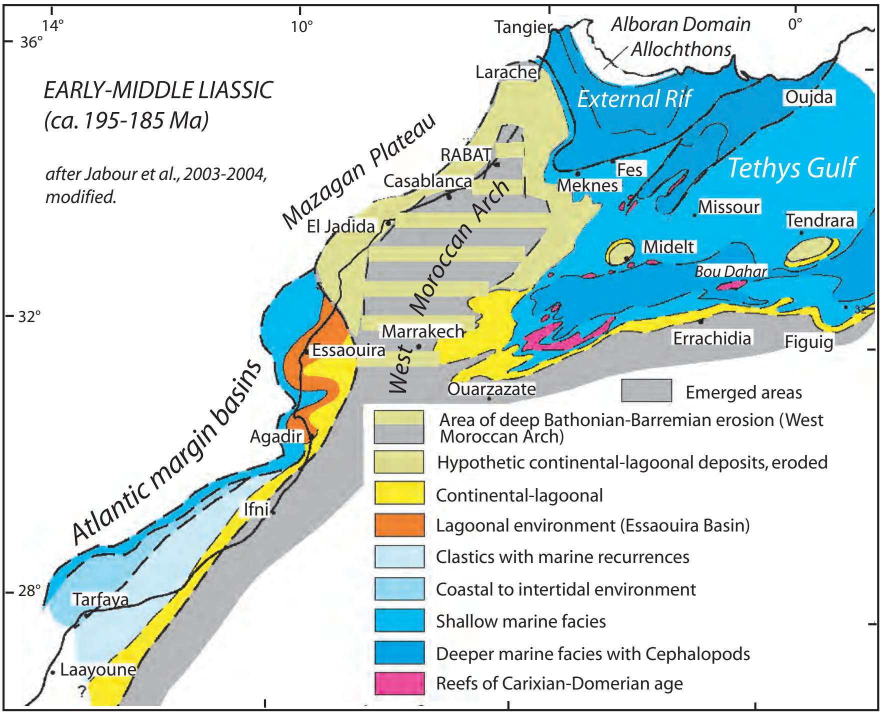

Fig. 4.4 Paleogeographic map for the Liassic epoch, after Jabour et al. (2003–2004), modified. The most important modification concern the West Moroccan Arch, which is no longer regarded as a Liassic emergent land, but as a shallow platform eroded during the late Middle Jurassic-Early Cretaceous interval (Ghorbal et al., 2007; Saddiqi et al., 2008)

inherited from the Late Permian-Triassic rifting. In between the two provinces extended a poorly subsident high, referred to as the West Moroccan Arch (WMA), whose Triassic-Liassic cover was subsequently eroded. Notice that this acception of the WMA differs from the classical concept of a Triassic-Liassic emergent land, defined by Choubert & Faure-Muret (1960–62) under the name of “Terre des Almohades” (Almohades Land) and by du Dresnay (1972, in Michard, 1976) as the “Dorsale du Massif Hercynien Central” (Hercynian Central Massif Rise). This new definition of the WMA is based on the most recent apatite fission-track studies (Saddiqi et al., 2008), which are summarized in the Chap. 7, Sect. 7.4.1 of this volume. Only some parts of the WMA were possibly emergent during the Liassic, in particular the Marrakech High Atlas.

The description of the orogenic period, from the Late Cretaceous to the Present, is based not only on the analysis of structures and unconformities within the chain, but also on that of the syn-orogenic Cenozoic basins fringing the system (Fig. 4.1). Such a presentation is facilitated by the numerous subsurface data issued from petroleum exploration of the South Atlas (Souss, Ouarzazate, Boudenib) and North Atlas basins (Essaouira, Haouz, Bahira, Tadla, Missour, Guercif). Additionally, Sect. 4.4 is dedicated to the quantification of shortening across the chain, and origin of the relief, i.e. the respective roles of crustal shortening and deep thermal processes. This last point has been already addressed in Chap. 1 (cf. Fig. 1.15). The study of the Cenozoic alkaline volcanism (Sect. 4.5) documents the role of the mantle in the recent activity of a zone crossing the Atlas system from the Anti-Atlas to the Rif.

4.2 The Pre-orogenic Evolution of the Atlas System

The Precambrian and Paleozoic evolution of the basement of the Atlas system is presented elsewhere (Chap. 3). Understanding these old structures is important to detect their occasional reactivation during Atlas building. The obvious parallelism of some of the major Paleozoic trends with the overall trends of the High and Middle Atlas attests a reactivation of numerous old structures. For instance, the South Atlas Front is more or less superimposed over the Tizi n’Test Fault (west part of the APTZ; see Chap. 3, Fig. 3.16) and the South Variscan Front (SMF). The Middle Atlas is parallel to the Variscan Tazekka-Bsabis Fault Zone (TBFZ), and the Mesozoic Tizi n’Tretten Fault itself (see below, Fig. 4.21) is also an important Variscan limit (Charrière, 1990). On the other hand, there are also some obvious exceptions such as the E-W trending fault systems of the Jebilet and Western High Atlas, which cut older Paleozoic structures almost at right angles.

After the Variscan orogeny, the break-up of Pangea was expressed in Morocco by successive extensional episodes in an overall rifting context. This initial rifting controls the subsequent evolution of the basins until their inversion during the Cenozoic. For these processes, a unified timing has long been searched for at the scale of the Atlas system or entire Morocco. Recent data, mainly from industrial seismic profiles, help identifying several domains with different evolutions. The major boundary is formed by the West Moroccan Arch defined above, which separates an Atlantic domain to the west from a Tethyan domain to the east. In the eastern domain itself, the Middle Atlas represents a peculiar case.

Seismic and field data suggest two distinct extensional end members. In the west (Argana Corridor, Essaouira Basin), Late Permian to Late Triassic rifting was followed by the formation of a sag basin with widespread evaporites and basaltic flows (Late Triassic-Early Jurassic boundary). In the northeast (Guercif Basin), Liassic to Late Jurassic extension dominates, whereas Triassic extension was more limited. However, the extensional structures are not restricted to these two end-members. An intermediate situation is found in the Central and Eastern High Atlas and the Prerif Ridges, which combine Triassic and Early to Middle Jurassic syn-rift evolution.

A significant role of early Mesozoic transtensional strike-slip faulting was first postulated by Mattauer et al. (1977). This hypothesis was still referred to in many subsequent papers (Laville & Petit, 1984; Laville, 1985, 1987; Favre et al., 1991). More precisely, Laville & Piqué (1991) emphasize the role of pre-existing late Variscan structures and suggest a series of Triassic-Early Liassic pull-apart basins bounded to the north by the Newfoundland and to the south by the Kevin-Tizi n'Test transform zones. By contrast, Jenny (1984), Schaer (1987), El Kochri & Chorowicz (1996) and Qarbous et al. (2003) suggest more limited strike-slip movements, as El Arabi (2007) does for the Triassic rifting itself. Possibly due to a wide spacing and inadequate location of the available seismic data, transfer faults were not identified in the subsurface. Therefore, in the following, we will emphasize the more obvious tensional characteristics of the Middle to Late Jurassic depocenters.

4.2.1 The Atlantic Domain (Western High Atlas)

References: The main references for the Western High Atlas and Atlantic margin are the works by Le Roy et al. (1997, 1998), Hafid (2000, 2006), Le Roy & Piqué (2001), and Hafid et al. (2000, 2006). The most general references on the Triassic rifting and Central Atlantic Magmatic Province (CAMP) are given in Chap. 1, but other data on the Late Permian and Triassic extension in western Morocco can be found in Medina (1991), Jalil (1999), Tourani et al. (2000), Medina et al. (2001), El Arabi E.H. (2007). References on the Jurassic sedimentary formations comprise Bouaouda (1987), Stets (1992), Ourribane et al. (1999), Martin-Garin et al. (2007); and on the Cretaceous sedimentary formations, see Taj-Eddine (1992), Algouti et al. (1999), Labbassi et al. (2000), Nouidar & Chellai (2001, 2002).

Within the Western High Atlas, rifting mainly occurred before the great volcanic flooding which marks the end of the Triassic period and floors a Lower Liassic evaporitic basin (Fig. 4.5). This sag basin linked to thermal relaxation is weakly faulted and acted as a relay between the intensively fractured syn-rift sequence and the post-rift sequence, which began synchronously with the accretion of the Atlantic Ocean at about 195 Ma (cf. Chap. 1, Fig. 1.8). Here and over much of Western Morocco,

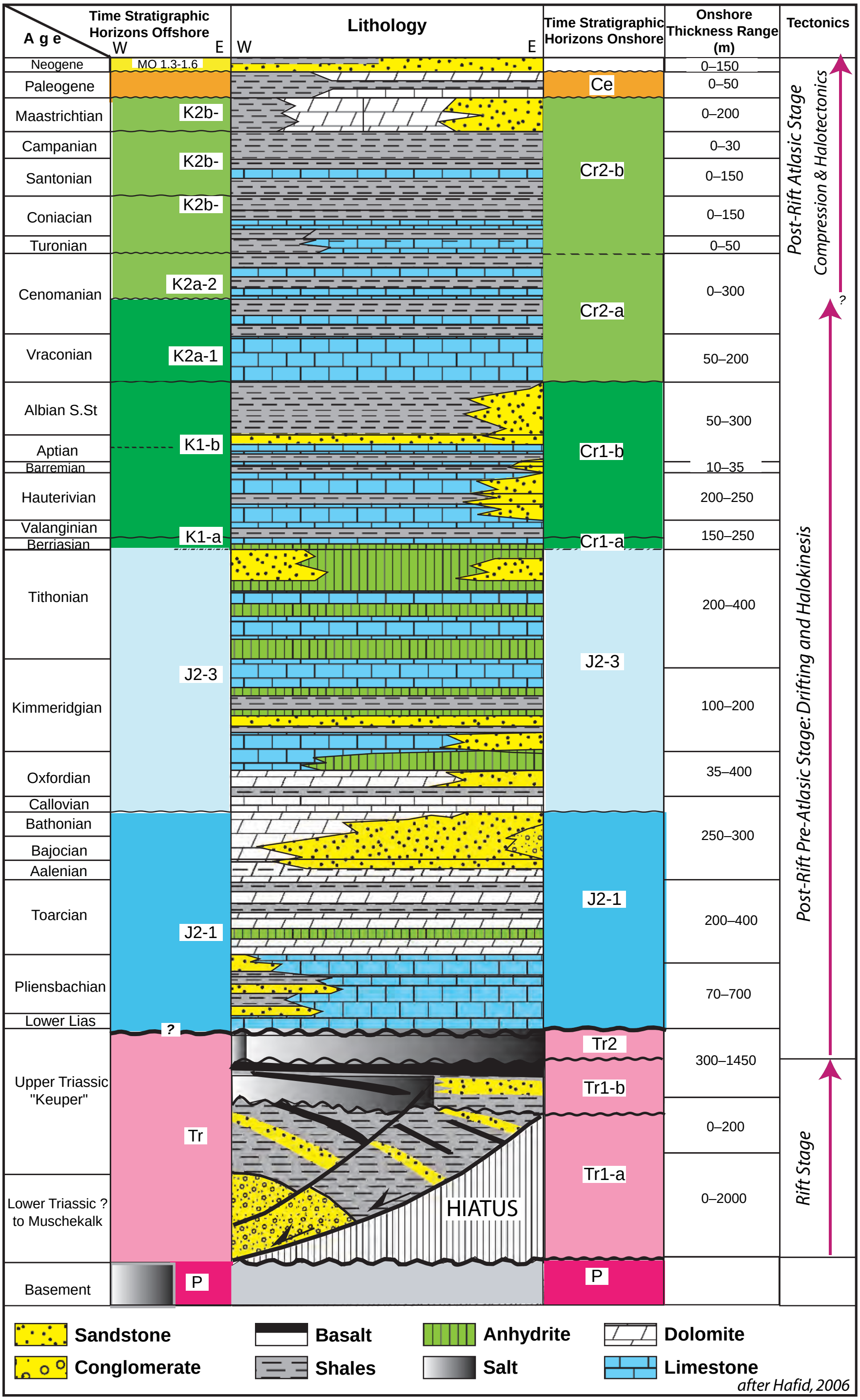

Fig. 4.5 Synthetic stratigraphic column of the onshore Essaouira Basin with age and correlation of seismostratigraphic subdivisions used in different zones of the studied area (after Hafid, 2006, modified)

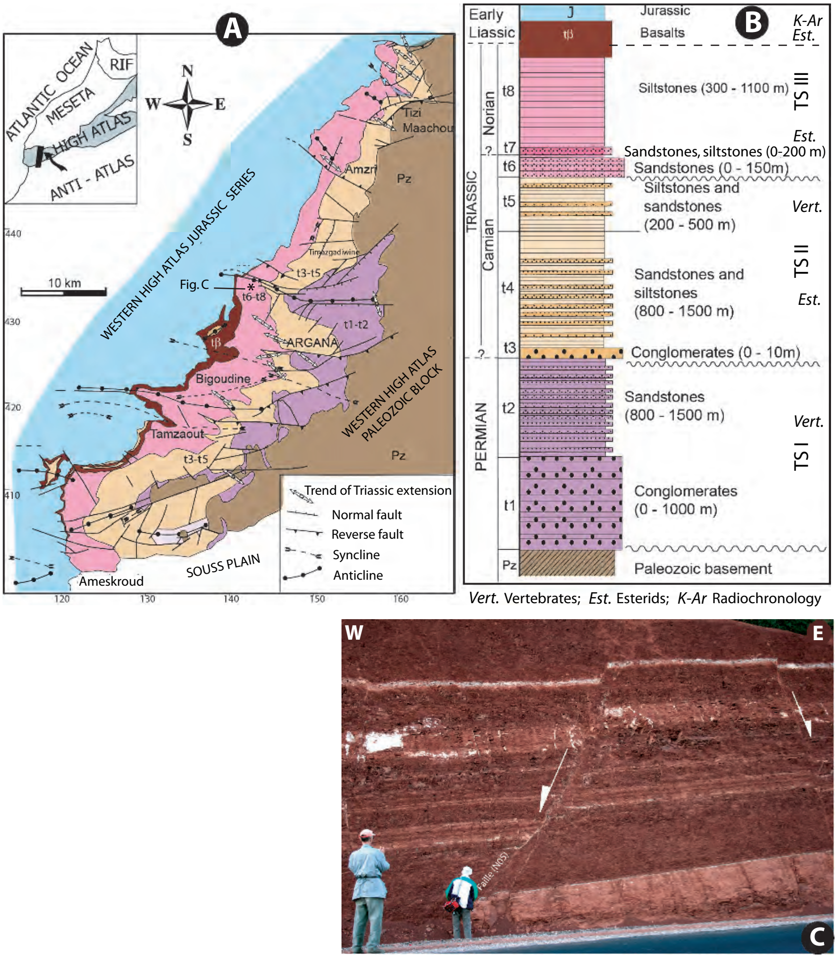

Fig. 4.6 Paleofaults and Triassic syn-rift sediments in the Argana Corridor. (**A**) Structural map from Tixeront (1974) *in* Medina (1991). We observe two different fault systems, E-W and NNE-SSW. They correspond to two distinct extensional phases (Late Permian and Mid-Late Triassic, respectively). – (**B**) Stratigraphic column, from Medina (1991) and Ait Chayeb et al. (1996), modified. TS I-III: Tectono-sedimentary sequences according to Olsen (1997), separated from each other by unconformities (*wavy lines*). Courtesy of H. Ouanaimi (**C**) Conjugate normal faults in Norian siltstones of the Argana Corridor (location: see map Fig. 4.6A). Photo by A. Soulaimani

the general trend of this extensional system is NNE. The syn-rift sequence crops out in the Argana Corridor that has been the subject of detailed studies (Fig. 4.6). It is also known in different wells, and well imaged in seismic data from the Essaouira Basin and Western High Atlas (see below).

To evaluate the regional extent of the Triassic-early Liassic sag basin, it is important to understand the nature of its eastern pinch out against the pre-Jurassic basement of the West Moroccan Arch. Medina (1995) suggests a pre-Carixian age for this unconformity. This author emphasizes that the onlapping strata become progressively younger eastward up to the Upper Bathonian in the western Jebilet. Below this unconformity, the seismic profiles clearly show the truncation of the underlying evaporites and basalts.

During the Jurassic, the sedimentation was dominated by the deposition of carbonates. From the Sinemurian up to the Callovian, the sedimentation pattern varies from one region to the other with development of clastic facies in the eastern area, and of calcareous and marly facies in the western domain. The existence of a flooding surface during the Callovian marks the installation of a carbonate platform, which developed until the early Kimmeridgian. During this epoch, we observe also the stacking of evaporite units particularly in the Doukkala and Essaouira Basins. The Triassic-Lower Liassic salt layers were remobilized almost immediately after deposition and breached the surface during the Middle Jurassic, suggesting that a diffuse extension lasted during the entire Jurassic.

After the Berriasian-Valanginian, the lower Cretaceous sedimentation developed over an almost general unconformity, which signals erosion and marks a complete change of sedimentary environment, now dominated by alluvial plains and siliciclastic facies. The WMA is the source area of the Lower Cretaceous siliciclastic turbidites which invade the Atlantic margin (Price, 1980; Behrens & Siehl, 1982) as well as the Maghrebian margin (e.g. Ketama unit of the External Rif Belt; see Chap. 5). The only observed deformations are linked to continued salt tectonics (Hafid et al., 2006; Hafid, 2006), but at that time the extensional movements had ceased.

4.2.2 The Tethyan Domain (Central and Eastern High Atlas, Middle Atlas)

In this domain two successive rifting episodes can be identified. The first rifting episode is Triassic, as shown by thick red beds capped by basaltic flows as in the Atlantic domain (Fig. 4.7). A second rifting episode occurred during the Jurassic. This second episode is typical of the Tethyan Atlas realm and deserves detail study.

4.2.2.1 The Earliest Rifting, from Permian to Early Sinemurian

References: The earliest rifting phases of the eastern Atlas system have been increasingly studied in recent times. The main references are Petit & Beauchamp (1986), Benaouiss et al. (1996), Jalil (1999), Oujidi & Elmi (2000), Oujidi et al. (2000), Ouarhache et al. (2000), Tourani et al. (2000), El Arabi E.H. et al. (2003, 2006a,b), and El Arabi E.H. (2007). Concerning the associated basalt flows, the most recent and general references are Youbi et al. (2003), Knight et al. (2004), Marzoli et al. (2004), Verati et al. (2007), and Nomade et al. (2007).

During the Late Permian-early Late Triassic, the High Atlas rift was separated from the Atlantic rift by the West Moroccan Arch (WMA), whose southern tip corresponds to the basement of the Marrakech High Atlas (Fig. 4.7A, B). At that time, the WMA represents the common shoulder of the rifted domains. Coarse clastic sediments are mostly found close to the basement rise, whereas silts and evaporites occur in more external areas. Detailed analysis of the tectonosedimentary units (TS) on the Telouet transect, i.e. across the south-westernmost part of the High Atlas rift oulines a southward migration of the depocenters from Late Permian to early Sinemurian in the framework of an asymmetric rift system due to NW-SE extension (Fig. 4.7B). The TS I fluviatile red beds are equivalent to the lowest Argana sequences (t1-t2, Fig. 4.6B), dated from the Late Permian by Vertebrate fossils found in the upper beds (Jalil, 1999). The earliest Triassic sequence (TS II) extends

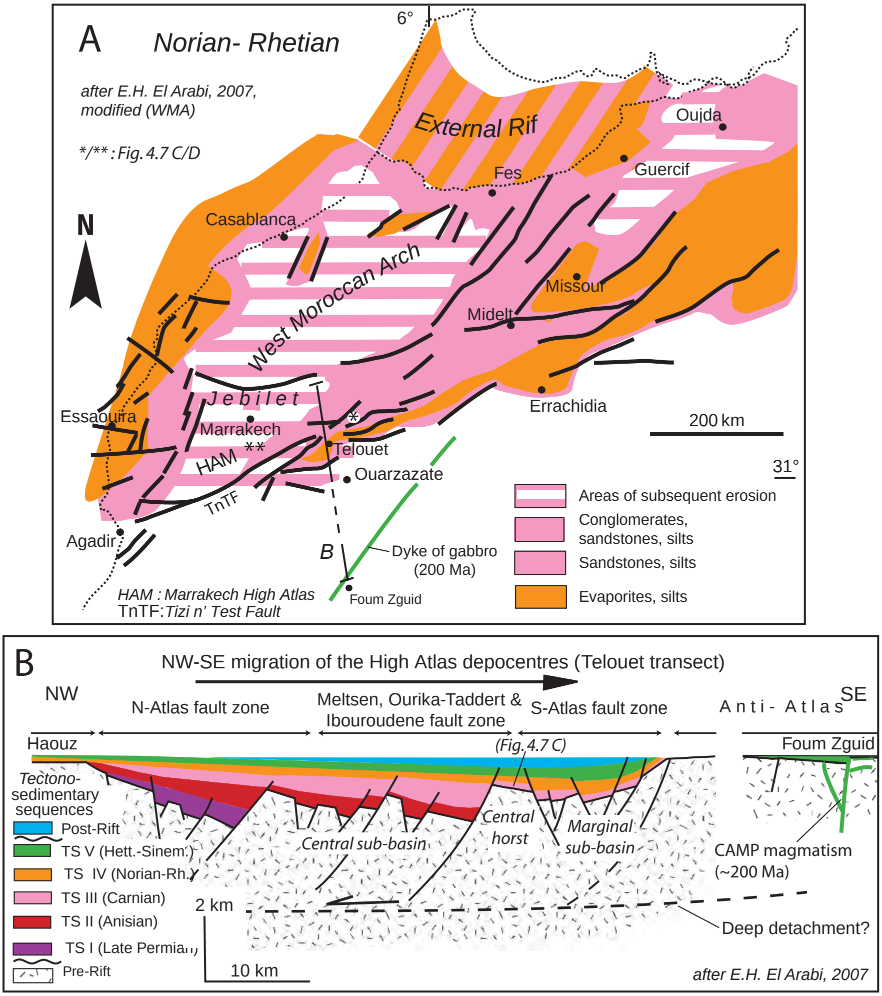

Fig. 4.7 (continued)

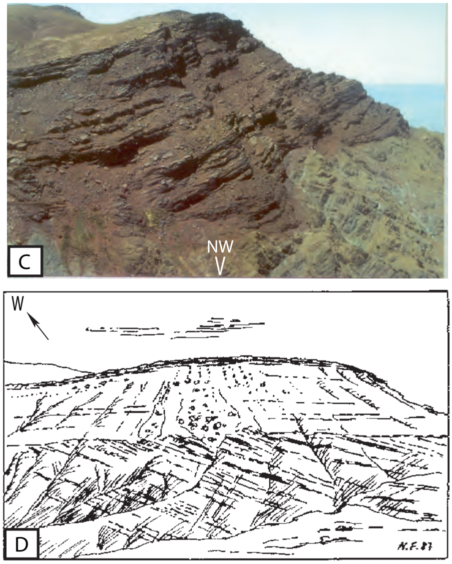

Fig. 4.7 The Permian and Triassic Atlas rift system, after H.A. El Arabi (2007), modified. The most significant modification concerns the West Moroccan Arch, which must have been covered by Triassic and Liassic deposits before its late Middle Jurassic-Early Cretaceous erosion (Saddiqi et al., 2008). – (**A**) Late Triassic paleogeographic setting. – (**B**) Restoration of the depocenters along the Telouet transect (located on **A**) from the Late Permian (TS I) to the beginning of the post-rift sedimentation (Early Jurassic). TS I: Upper Permian fluviatile sandstones, conglomerates and silts. TS II: Middle Triassic fluviatile-lacustrine deposits in central sub-basin, alluvial fans in the marginal sub-basin. TS III: Unconformable Carnian fluviatile-deltaic sandstones, equivalent of the Oukaimeden Sandstone Fm. (Ourika Valley, south of Marrakech). TS IV: Late Carnian-Norian-Rhaetian eolianites, and lacustrine siltites and clays with rare evaporites. TS V: Rhetian-Hettangian-Sinemurian basalts, overlain by silts and marly limestones. (**C**) Transgression of the Oukaimeden Sandstone (TS III) on top of the Precambrian schists of the central horst, High Tessaout Valley (from El Arabi E.H., 2007).– (**D**) Jurassic marls and limestones unconformably overlying the tilted Triassic siltstones west of Tahanaout, Oued Rheraia Valley south of Marrakech (from Froitzheim et al., 1988)

southward onto the basement, hence recording an increased tectonic subsidence of the central sub-basin (Fig. 4.7B). The fluviatile-lacustrine/palustrine beds of this sequence have been dated palynologically from the Middle Anisian in the central sub-basin (El Arabi et al., 2006a). The overlying, unconformable sequence (TS III) corresponds to the famous Oukaimeden Sandstone Fm. (Fig. 4.7C), dated as Carnian based on its palynological associations. This sequence is made up of deltaic, conglomeratic sandstones, which again extend more widely southward with respect to the previous sequences. Their thickness strongly varies from 600 m in the central sub-basin to about 20–50 m on top of the central horst. Eolianites, associated with lacustrine clays and silts, and rare evaporites characterize the overlying sequence TS IV, dated as late Carnian-Norian-Rhetian based on palynologic observations (Marzoli et al., 2004). Consistently, the 120 m-thick lower to upper basalt flows of the overlying sequence (TS V), intercalated with silts and marls, yielded ^{39}Ar-^{40}Ar ages from 199.0 ± 2.0 to 195.6 ± 8.9 Ma, whereas the uppermost "recurrent" flow yielded undistinguishable ages of 196.7 ± 1.9 and 197.6 ± 2.2 Ma (Nomade et al., 2007; Verati et al., 2007). This giant volcanic episode of the CAMP province probably caused the biological crisis, which marks the Triassic-Jurassic boundary (see Chap. 1). The post-rift sequence corresponds to the building of a carbonate platform of Lower-Middle Liassic age. The break-up unconformity is obvious by place (Fig. 4.7D). At least the TS IV and TS V sequences (late Carnian-Hettangian) sequences extended onto most of the WMA, weakly subsiding at that time (Saddiqi et al., 2008).

The sedimentary record and horst/graben geometry of the Triassic rifting have been also described in the Oujda region (Oujidi et al., 2000). The carbonate layer overlying the lowermost basalt flow in the area was formerly attributed to the Ladinian-Carnian. However, the paleontological dates (*Anaplophora sp.* and Ostracods), which have been used in support of this alleged age are in fact ambiguous (H. Bertrand, pers. com., 2007), whereas the underlying lava flow yielded a robust ^{39}Ar-^{40}Ar age at 198.0 ± 0.8 (Marzoli et al., 2004).

4.2.2.2 Early and Middle Jurassic: The Two Branches of the Atlas Rift

References: For the Middle Atlas and Guercif Basin, we mainly used the studies by Fedan (1989), Charrière (1990, 2000) and Zizi (2002). For the Jurassic of Central and Eastern High Atlas, fundamental discoveries have been made by Du Dresnay since 1965. Most of this work remains unpublished but has been widely diffused through oral communications and geological maps by this author, and by some syntheses (Du Dresnay, 1972, 1977, 1979, 1987, 1988). Another recent synthesis on the Atlas rifting by Laville et al. (2004) leads to an interpretation rather different from that presented here. Numerous papers address the question of the successive sedimentary systems in the Tethyan Atlas, among which: i) for *Lower and Middle Jurassic (Middle Atlas)*: Fedan (1989), Benshili (1989), Benshili & Elmi (1994), Charrière et al. (1994a,b), Elmi (1999), El Arabi et al. (1999, 2001), Charrière (2000), Akasbi et al. (2001), EL Hammichi et al., 2002; for *Lower to Middle Jurassic (High Atlas)*: Jossen (1987), Warme (1988), Sadki (1992), El Kochri & Chorowicz (1996), Souhel et al. (1998, 2000), Aït Addi et al. (1998), Elmi et al. (1999), Igmoullan et al. (2001), Kaoukaya et al. (2001), Neuweiler et al. (2001), Milhi et al. (2002), Mehdi et al. (2003), Chafiki et al. (2004), Ettaki & Chellai (2005), Aït Addi (2006).

The Tethyan Atlas system comprises two distinct branches (High and Middle Atlas) which were active during the Early Jurassic. It is worth noting that the trend of the normal faults which controlled Jurassic sedimentation is clearly different from the one observed in the Atlantic domain. These faults are oriented ENE in the High Atlas, slightly oblique on the E-W main trend of the chain, and NE in the Middle Atlas, i.e. parallel to the chain.

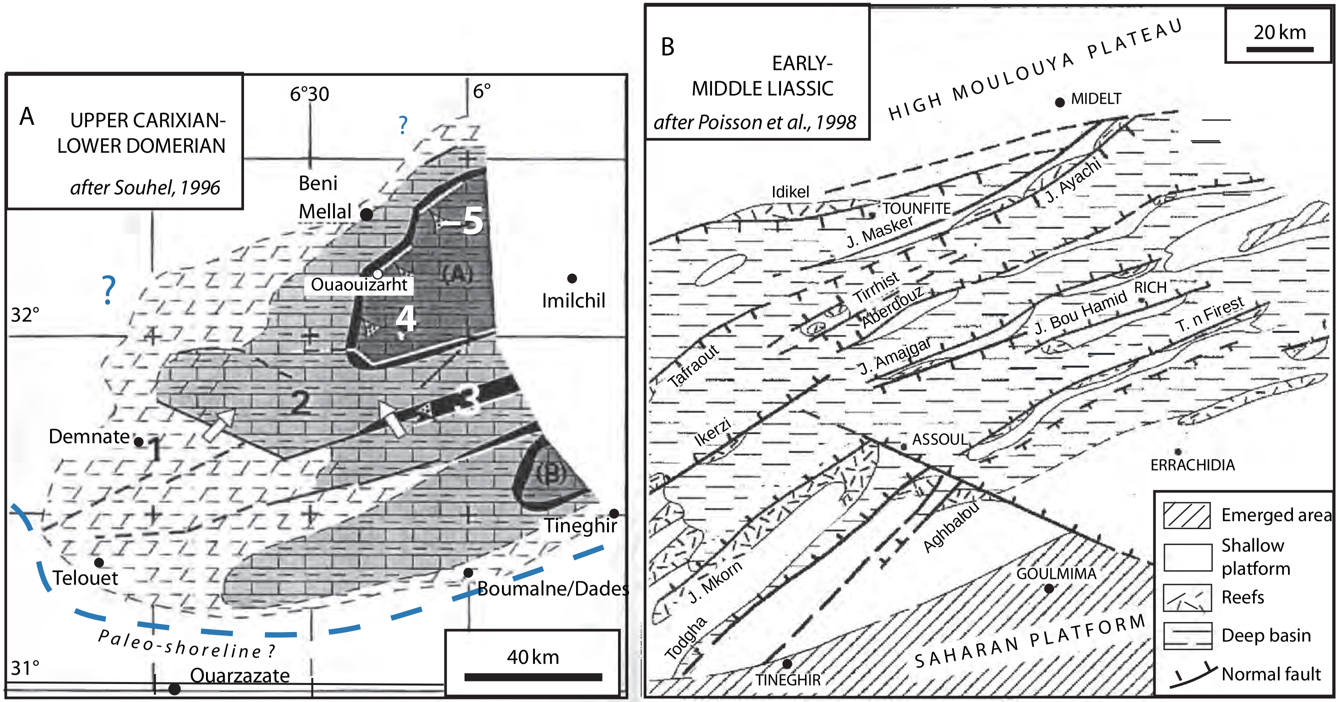

Fig. 4.8 Paleogeography of the High Atlas Basin during the Early-Middle Liassic. (**A**) Southwestern closure of the Tethyan Gulf, after Souhel (1996), modified. The question marks record that the present limits of the Jurassic are erosional limits, related to the Cretaceous and/or Cenozoic uplift episodes. 1: Supra- to intertidal dolomites and marls; 2: Lagoonal limestones, dolomites and marls; 3: Shallow water ridge with oolites, Megalodontids, reefs; 4: Trough facies with marls and limestones alternations; 5: Calci-turbidite fans; arrows: direction of infill thickening. – (**B**) Restored pattern of shallow water ridges and deeper of the Central High Atlas, after Poisson et al. (1998), modified. This pattern is interpreted as the result of extensional block faulting. Notice that both maps (**A**) and (**B**) are not restored with regards to tectonic shortening

The southwestern closure of the Tethyan Atlas realm is clearly illustrated east of the Marrakech High Atlas (Fig. 4.8A). This area is particularly interesting because of the well-documented lithostratigraphic changes which allow us to restore the bathymetric/environmental variations from deep to shallow water, and finally to the interdidal zone. Based on such mapping, the occurrence of a temporarily emergent high in the Marrakech High Atlas (southernmost WMA) is clearly documented.

In Fig. 4.8A, the westernmost part of the Tethyan rift appears to be divided into two sub-basins separated by a shallow water ridge. Farther east in the Central High Atlas, such pattern is repeated several times in the paleogeographic map corresponding to the same Early-Middle Liassic interval (Fig. 4.8B). This is interpreted as the result of extension and block tilting during this time interval. In other words, field evidence shows that a second rifting episode occurred after the Triassic episode and the first post-rift platform sequence (Early Liassic). The second rifting evolution lasted from the Middle Liassic until the Dogger, as demonstrated by the abrupt transition from platform (including reefs) to basinal facies (Figs. 4.8 and 4.9). Due to Cenozoic inversion, these transitions correspond presently to reverse faults carrying the basinal units onto the platform ones. Usually the fault zones display outcrops of

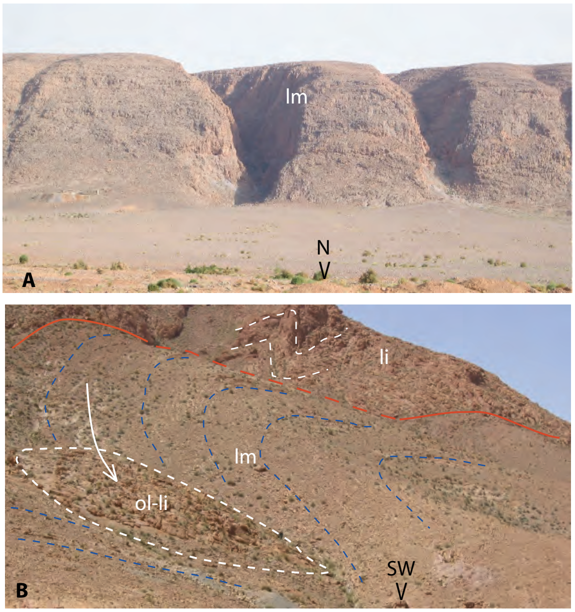

Fig. 4.9 (**A**) Southern border of the Bou Dahar reef (Eastern High Atlas; see Fig. 4..4 for location). This remarkable reef complex has been mapped by du Dresnay [see in Michard (1976), p. 174] and revisited by Elmi et al. (1999). The reef developed during the Middle Liassic over a basement high (block shoulder). It was then flooded during the Late Toarcian-Aalenian. Its southern front, 20 km long, shows canyons with a regular spacing, which were filled by Aalenian mud before being exposed by recent erosion. – (**B**) Olistolite of dolomitic limestone (Idikel Fm, Sinemurian) within bathyal marls (Ouchbis Fm, Domerian) steepened along a paleo-normal fault (northern limb of the J. Bou Gharral anticline, 1/50 000 sheet Talsint-East (Eastern High Atlas)

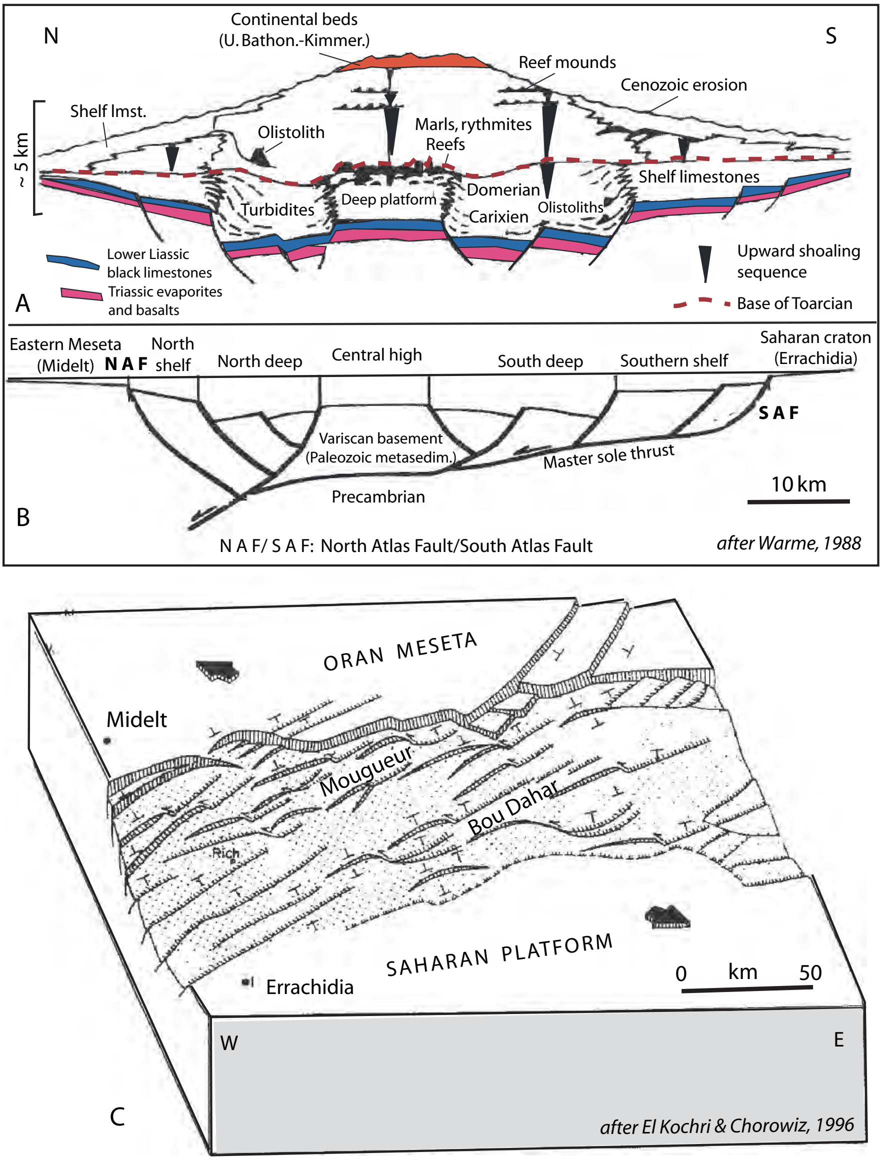

Fig. 4.10 The Jurassic High Atlas rift. **A, B**: Schematic sections showing the general organisation of the sedimentary facies (**A**) and the pre-inversion fault geometry along the Midelt-Errachidia transect (**B**), after Warme (1988), modified. – **C**: Block-diagram suggesting a transtensional regime during the Jurassic (from El Kochri & Chorowicz, 1996)

Triassic argillites and basalts occasionally accompanied with Paleozoic slates as illustrated by the Foum Zabel fault north of Errachidia (see Fig. 4.40 A, B hereafter). Such geometry, repeated in numerous places, is certainly inherited from the Liassic syn-extensional architecture (Fig. 4.10). During the Toarcian, basinal facies are dominant and quite uniform suggesting that their deposition is controlled by thermal subsidence. More than 5000 m of marls and calci-turbidites accumulated until the end of Bajocian.

In the Middle Atlas, after the deposition of Triassic sediments of various thickness, the evolution of the Jurassic basin resulted from three episodes (Fig. 4.11): (i) slow platform subsidence during the Early-Middle Liassic, slightly accelerated during the Domerian; (ii) a period of reduced subsidence during the Toarcian-earliest Bajocian; (iii) a phase of active subsidence and filling from middle Bajocian to late Bathonian, firstly under marine conditions then in a continental environment with frequent marine incursions (El Mers syncline) during the late Bathonian (Charrière et al., 1994a). The same evolution is shown in the seismic profiles analysed by Zizi (2002) in the Guercif Basin, at the northern termination of the Middle Atlas (cf. Fig. 4.23). In contrast, the Late Bajocian corresponds to the latest marine deposits in the High Atlas trough (Fig. 4.12). At that time, the Middle Atlas Basin was separated from the High Atlas one. Subsidence curves representative of Eastern High Atlas and Middle Atlas highlight this difference in the duration of the rifting processes.

4.2.2.3 From the Late Jurassic Emersion to the Late Cretaceous Transgression

References: Regarding the continental red bed stratigraphy, the recent references are Jenny et al. (1981), Charrière et al. (1994a), Haddoumi et al. (1998), Monbaron et al. (1999), Feist et al. (1999), Haddoumi et al. (2002), Charrière et al. (2005), Haddoumi et al. (2008). The magmatic outcrops and associated structures from the High Atlas are described in Schaer & Persoz (1976), Monbaron (1980), Laville & Harmand (1982), Jenny (1984), Studer (1987), Brechbühler et al. (1988), El Kochri & Chorowicz (1996), and Laville & Piqué (1992). For the Cretaceous marine transgressions, see Wurster & Stets (1982), Andreu (1989), Charrière & Vila (1991), Ennslin (1992), Charrière (1996), Charrière et al. (1998), Ciszak et al. (1999), Ettachfini et al. (2005). The paleogeography at the scale of the Maghreb in general is described in Fabre (2005). The petrography of magmatic rocks is discussed in Sect. 4.2.2.4.

During the Bathonian, the marine sedimentation of the Atlas domain changed to fluviatile red beds supplied by the neighbouring highlands, i.e. the Saharan Domain and West Moroccan Arch. The uplift and erosion of the latter domain is documented by the low-temperature thermochronological studies (Ghorbal et al., 2007; Saddiqi et al., 2008). During the Late Jurassic and the beginning of Early Cretaceous, the entire Morocco was emerged except the very borders of the Atlantic and Tethyan passive margins. The former Tethyan Atlas basin was then converted into a transit zone for the clastic flux which fed the “flysch” sedimentation along the

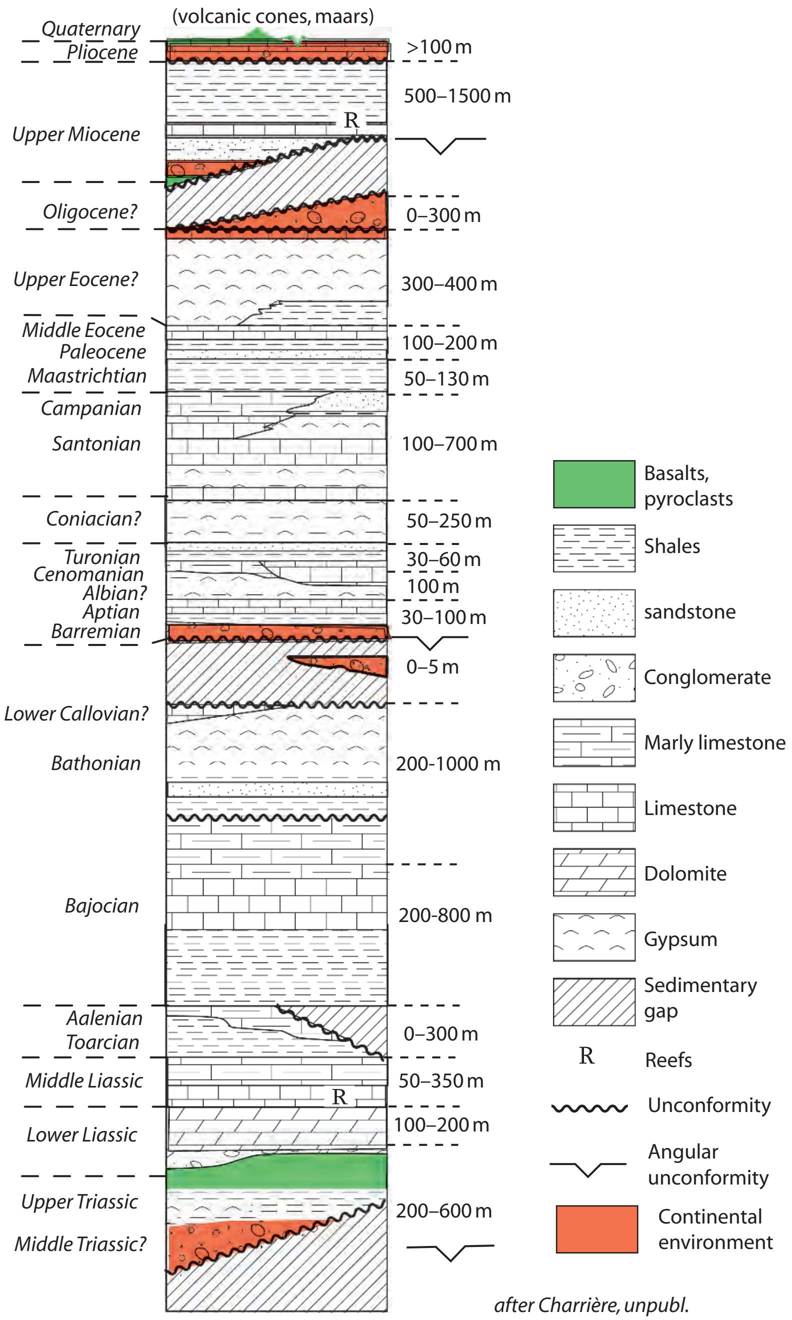

Fig. 4.11 Stratigraphic column of the Middle Atlas after an unpublished document by A. Charrière, *in litt*., 2007 Note the occurrence of three major tectono-sedimentary cycles during the Meso-Cenozoic

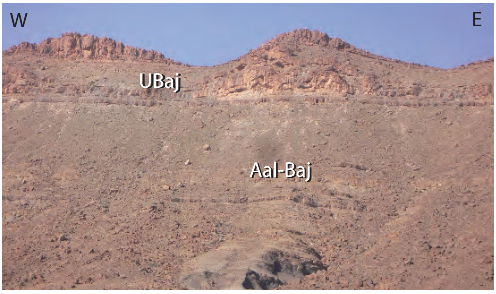

Fig. 4.12 The upward-shallowing sequence of Aalenian-Bajocian marls and *Cancellophycus* limestones (Aal-Baj), culminating in the reef mound clusters of the so-called "Calcaires corniche" (Upper Bajocian, UBaj) of J. Assemour n'Ait Fergane east of Rich, Central-Eastern High Atlas. See Fig. 4.3 (*) for location. These patch reefs are *Scleractinian* coral and algal build-ups (du Dresnay, 1987; Warme, 1988). During the Aalenian-Bajocian interval, progradation of the shelf units extended the reef facies from South (Foum Zabel) to North (J. Assemour; see Aït Addi, 2006). Photo O. Saddiqi

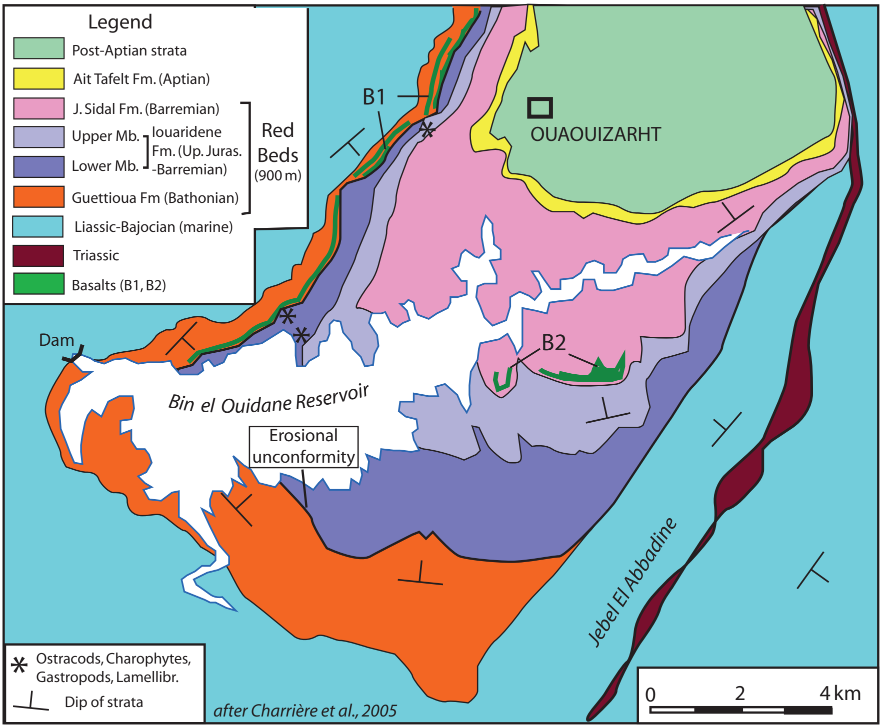

Fig. 4.13 Geological map of the Ouaouizarht syncline south of Beni Mellal (Central High Atlas), after Charrière et al. (2005). Location: see Fig. 4.25. The Middle Jurassic-Barremian red beds extend between the Aalenian-Bajocian shallow marine limestones and the unconformable Aptian-Cenomanian transgression. The B1 basaltic lava flow emplaced on top of the Bathonian red beds (Guettioua Fm) before the sedimentation of the Oxfordian-Kimmeridgian red beds and dolomites (Lower Iouaridene Fm), i.e. between 165–160 Ma. The B2 basalt outpour occurred above the Iouaridene Fm. and at the bottom of the J. Sidal Fm., i.e. around 130 Ma

Tethyan margin of the Rif domain. The stratigraphy of the Atlas red beds has been progressively enlightened, based on the Vertebrate, Ostracod and Charophyte findings (e.g. Fig. 4.13). In particular, the Atlas red beds provided numerous bones and foot prints of dinosaurs (Fig. 4.14) as well as a complete skeleton of giant Sauropod with legs longer than 3.5 metres (Monbaron et al., 1999).

During the Bathonian-early Upper Jurassic, an important plutonic and volcanic activity occurred in the High Atlas, being essentially recorded by gabbro intrusions and basalt lava flows in the Central High Atlas (e.g. Figs. 4.13 and 4.15; cf. also Fig. 4.39). More restricted Early Cretaceous (Barremian) basalt outpours occurred in the northern Central High Atlas (e.g. Fig. 4.13). According to Laville and Piqué (1992), the Jurassic magmatic episode bears the signature of a major phase of folding associated with cleavage development under transpressional regime, and of the subsequent erosion leading to the exhumation of the plutonic rocks. In this interpretation, the plutonic rocks and associated folds should be unconformably covered by latest Jurassic-Cretaceous red beds. This interpretation remains a matter of debate (see Laville, 2002, and Gomez et al., 2002; see also Sect. 4.4 hereafter). Indeed,

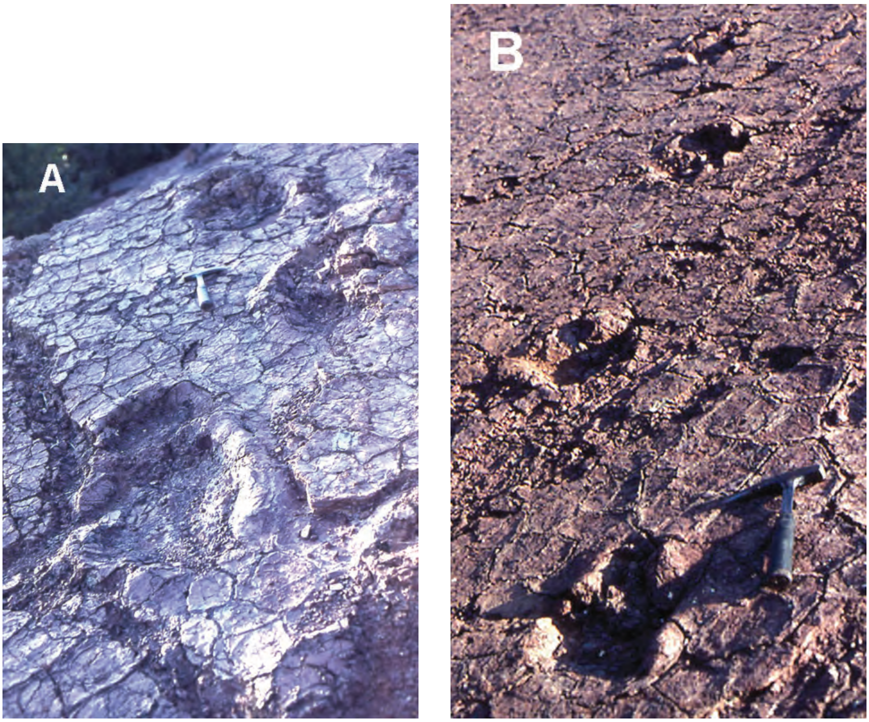

Fig. 4.14 Dinosaur tracks from the Iouaridene syncline southeast of Demnate (see Fig. 4.25. for location). (**A**) Sauropod circular track with mud cracks. – (**B**) Theropod tridactyle track. Both tracks are preserved in the lower part (member "b", Oxfordian?-Kimmeridgian) of the Iouaridene Fm. Photos by A. Charrière

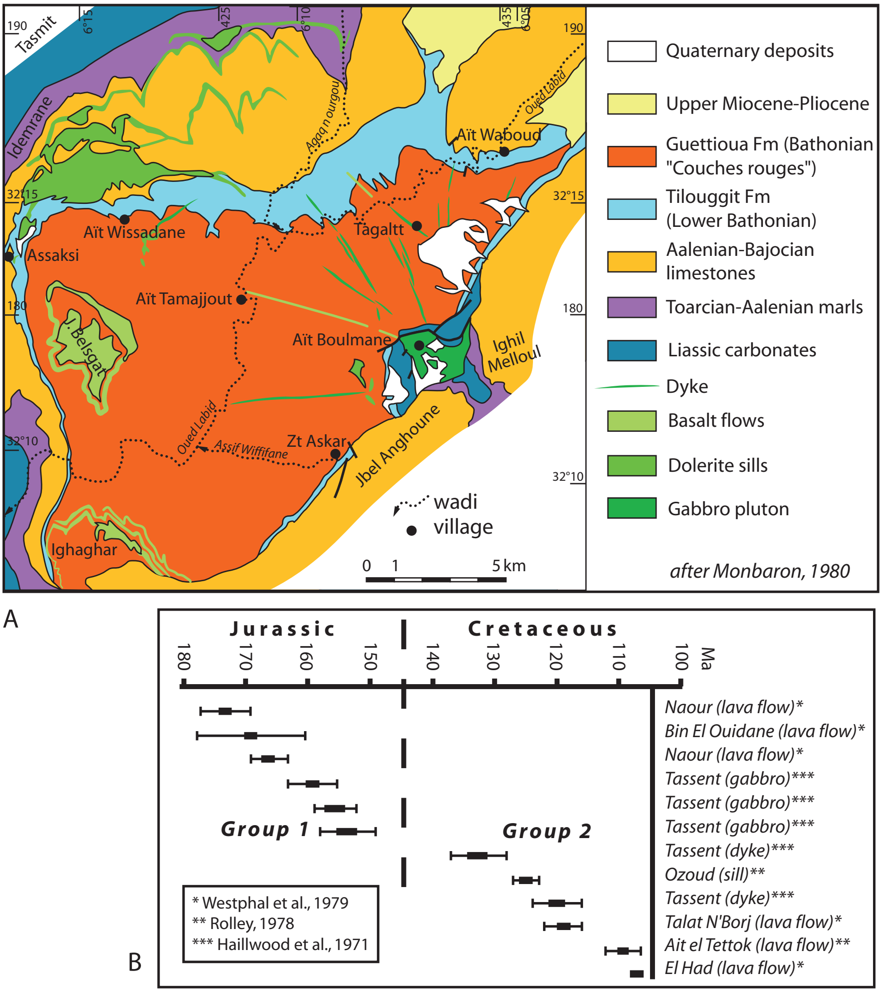

Fig. 4.15 The plutonic and volcanic rocks of the Central High Atlas. – (**A**) Jurassic volcanic and subvolcanic system of the Tagalft Basin SE of Beni Mellal, after Monbaron (1980), modified. See Fig. 4.25 for location and Fig. 4.39 for the general distribution of the plutonic outcrops. – (**B**) Radiometric dating of the High Atlas magmatic rocks, compiled by Souhel (1996), modified. An example of Cretaceous basalt is illustrated in Fig. 4.13

recent thermochronologic data suggest that the plutonic rocks were still situated at depth 90–80 Ma ago (Barbero et al., 2006). It seems more convincing to link the Jurassic magmatism of the High Atlas to the continuation of the previous extensional regime. Moreover, this interpretation appears consistent with the petrologic and geochemical features of the magmatic rocks, as discussed below (Sect. 4.2.2.4).

From the Valanginian to Aptian, the continental lands of Northern Morocco were progressively divided into two distinct emerged lands, due to the formation

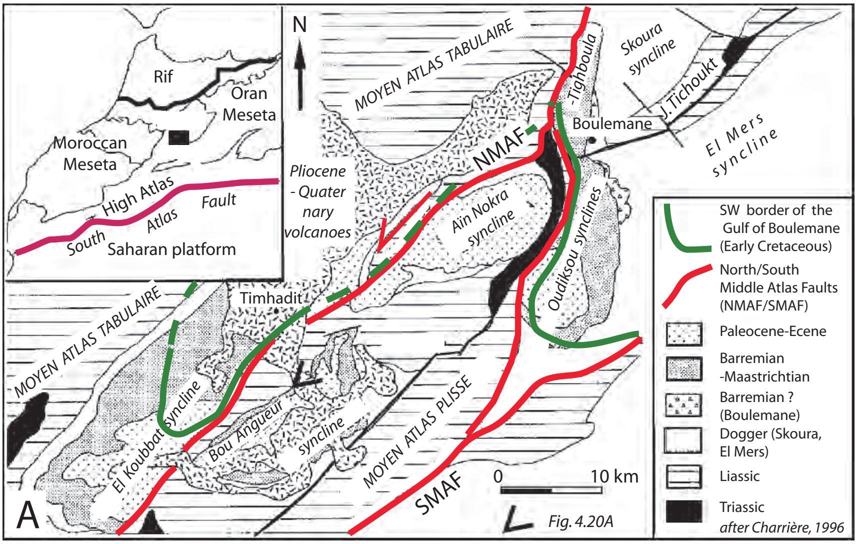

Fig. 4.16 Southwestern limits of the Lower Cretaceous deposits in the Middle Atlas synclines (Gulf of Boulemane=North Middle Atlas Gulf in Fig. 4.17), and distribution of the Upper Cretaceous-Eocene deposits (Atlantic gulf of Timahdite; cf. Fig. 4.19)

of two narrow, elongated gulfs (Fig. 4.17). The North Middle Atlas Gulf (NMAG) developed southeastward up to the Timahdite and Boulemane area (Fig. 4.16), being likely connected with the Peri-Tethyan seas. The North High Atlas Gulf (NHAG) developed northeastward starting from the Atlantic margin (Essaouira basin) up to the Beni Mellal Atlas, thus separating the northern West Moroccan Arch from its "root" (Siroua area) south of Marrakech (Fig. 4.17). These converging gulfs did not connect one to each other, being separated by continental deposits in the south Middle Atlas area. Therefore, the West Moroccan Arch was connected through an isthmus to the emerged lands of the eastern Atlas domain, which we refer to as the "High Atlas Arch". Basically, the latter arch was in continuity with the "Anti-Atlas Arch". In contrast, the West Moroccan Arch was disrupted, not only by the North High Atlas Gulf, but also by the extension of the Doukkala embayment toward the Meseta axis (northwestern Rehamna). The occurrence of a thin marine intercalation (Valanginian?) in the red beds southeast of the Rehamna Massif would suggest that the latter was temporarily an island during the Early Cretaceous. Notice that the "Terre des Idrissides" which was intended by Choubert & Faure-Muret (1960–62) as a Late Jurassic-Early Cretaceous emergent land extending continuously from the Central Massif to the High Plateaus does not appear in our reconstruction.

By the beginning of the Late Cretaceous, a general transgression marked the unification of the Atlas system and the beginning of a common history (even if the diversity of structural inheritance will have direct consequences on the response to the subsequent shortening processes). The high sea level enhanced the previous, Early Cretaceous transgression, and shallow marine conditions prevailed in the entire

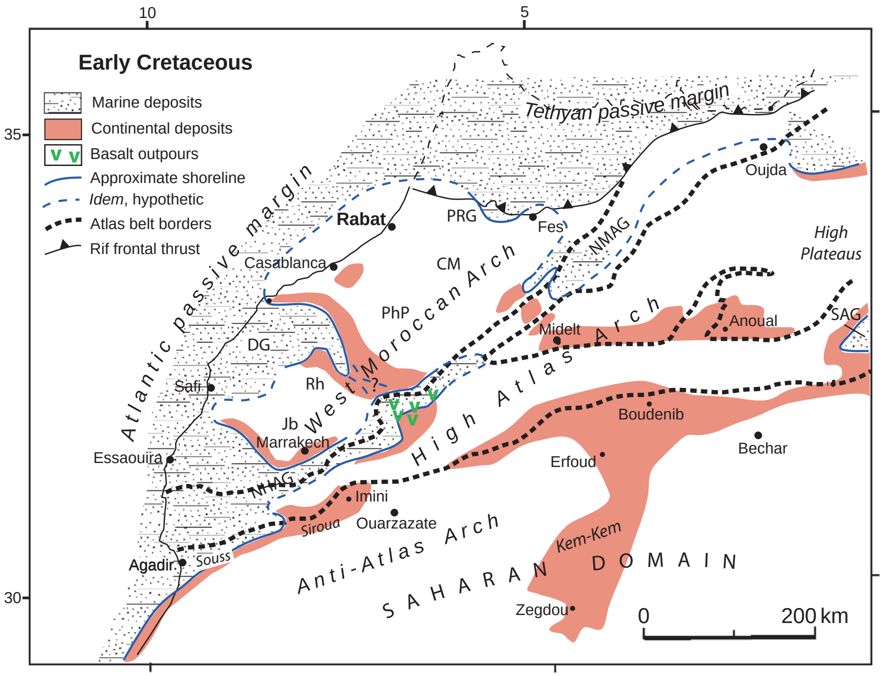

Fig. 4.17 Generalized paleogeography of northern Morocco during the Early Cretaceous, unrestored for tectonic displacements, after Faure-Muret & Choubert (1971), modified. The Atlantic and Tethyan gulfs (western and north-eastern transgressions, respectively) are shown at their maximum extension, i.e. during the Aptian. Notice that Early Cretaceous continental formations occur beneath the marine deposits in the internal parts of the gulfs (cf. the thick Barremian red beds from Ouaouizarht, Fig. 4.13). White areas correspond either to Early Cretaceous highlands devoid of red beds prior to the Cenomanian-Turonian transgression (e.g. Central Massif), or to subsequently eroded areas (e.g. Central High Atlas). CM: Central Massif; DG: Doukkala Gulf; Jb: Jebilet; NHAG: North High Atlas Gulf; NMAG: North Middle Atlas Gulf (Gulf of Boulemane in Fig. 4.16); PhP: Phosphate Plateau; PRG: Prerif Ridge Gulf; Rh: Rehamna; SAG: Saharan Atlas Gulf

Meseta and Atlas domains as well as in the northern Sahara regions (see Chap. 7). Together with the Lower Cretaceous red beds, the Cenomanian-Turonian limestones overlie unconformably the Jurassic rift structures (Fig. 4.18). The Cenomanian-Turonien horizon forms an outstanding benchmark at the scale of the Atlas system, and can be used as a reference for the subsequent vertical movements. Subsidence continued during the Senonian and Paleocene times, highly variable in intensity, and with an alternation of continental and marine sedimentation depending on the global sea level changes. However, by the end of the Cretaceous, a complete change in the tectonic regime occurred. It marked the beginning of the inversion, which subsequently developed during the Cenozoic as a response to the convergence of the Africa and Europe plates. It is notable that an E-trending land (latest stage of the “Terre des Idrissides” of Choubert & Faure-Muret, 1960–62; “Terre Sud-Rifaine”

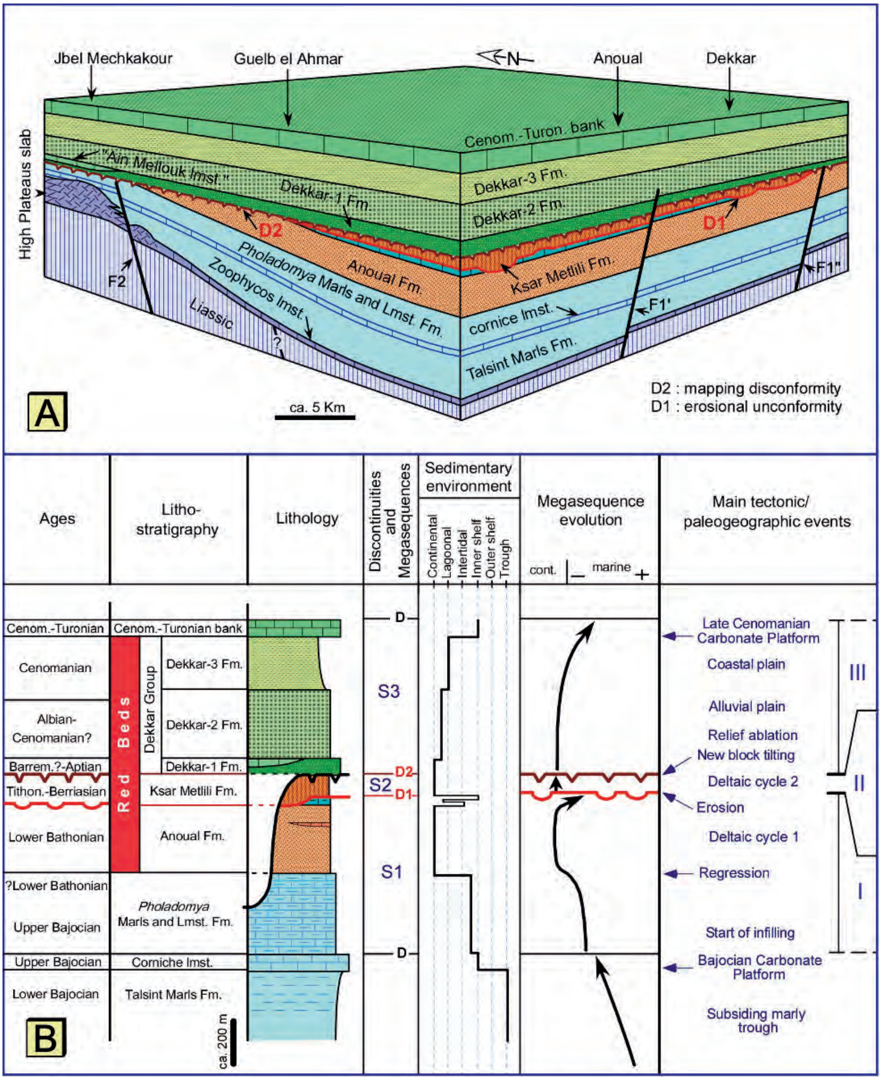

Fig. 4.18 Homogenization of the Atlas rift system beneath the Cretaceous post-rift sequence: example of the Anoual area, after Haddoumi et al. (2008). – (**A**) Schematic organisation of the Middle Jurassic and Cretaceous deposits in the Anoual area. – (**B**) Sequential evolution and paleotectonic-paleogeographic events in the Eastern High Atlas. I: Upfilling of the High Atlas Carbonate Platform; II: Emersion and residual deposits; III: Individualization of the Cretaceous basins, and Cenomanian-Turonian transgression

of Michard, 1976) formed at that time (Senonian) between the Tethyan open marine domain and the Atlas confined seas. This emergent rise, referred hereafter as the "North Moroccan Bulge", could result from a large-scale deformation of the continental lithosphere at the onset of plate convergence.

4.2.2.4 An Outline of Mesozoic Magmatism

References: Numerous studies have dealt with the geochemistry and geochronology of the Triassic-Liassic basalts from High Atlas and Anti-Atlas. The reader will find detailed accounts of these studies in Bertrand et al. (1982), Bertrand (1991), Sebai et al. (1991), Ouarhache et al. (2000), Lachkar et al. (2000), Youbi et al. (2003), Knight et al. (2004), Marzoli et al. (2004), Verati et al. (2007), Mahmoudi & Bertrand (2007) and references therein. Post-Liassic magmatic rocks have been comparatively less studied (Hailwood & Mitchell, 1971; Harmand & Laville, 1983; Beraâouz et al., 1994; Amrhar et al., 1997; Lhachmi et al., 2001; Zayane et al., 2002; Charrière et al., 2005).

Two main magmatic events occurred during the Mesozoic evolution of the Atlas System, i) during the Late Triassic-Early Liassic (200–195 Ma), and ii) during the Middle-Late Jurassic and Early Cretaceous (from ~ 165 to ~ 130 Ma), in fact a much more extended time interval than for the Triassic-Liassic event.

The CAMP Event

The first magmatic event is clearly linked to the Triassic rifting, which led to the opening of the Central Atlantic and Western Tethys oceans. This event characterizes the wide Central Atlantic Magmatic Province (CAMP), and affected the entire Morocco from the Anti-Atlas to the External Rif domain. This widespread, but remarkably short event has been described in the first chapter of the present book (cf. Sect. 1.2 and Figs. 1.6–1.8). The trapp-like pile of fluid basaltic flows emplaced within the Triassic-Liassic basins reaches up to 350 m-thick. Four superimposed units (lower, middle, upper and recurrent units) have been identified within this pile. Their emplacement occurred mainly during a short event straddling the Triassic-Jurassic boundary at 200 ± 1 Ma (Knight et al., 2004), and was associated with a climatic and biotic crisis (Marzoli et al., 2004). The recurrent basalts are dated at ~ 196 Ma in the southern High Atlas (Verati et al., 2007). A volcanic event postdating the main basalt flows is also observed in the southeastern Middle Atlas during the Hettangian-Sinemurian (Ouarhache et al., 2000). Some Triassic-Liassic basalts experienced a strong hydrothermal alteration (evidenced by the occurrence of the well-known amethyst druses). However, their primary geochemical signature, typical of continental tholeites (flood basalts) is generally well preserved, being characterised by weak to moderate enrichments in incompatible elements, together with negative niobium anomalies that indicate significant contamination by continental crust materials during their ascent.

Middle-Late Jurassic to Early Cretaceous

During this second “event”, which is specific for the High Atlas domain, basaltic lava flows and subvolcanic intrusive complexes were emplaced mostly in the Central High Atlas (see Fig. 4.39). However, lava flows were also emplaced during the Dogger in the Western High Atlas (Amrhar et al., 1997), whereas varied intrusions in the Eastern High Atlas are likely to refer to the Late Jurassic event (El Kochri & Chorowicz, 1996). Most plutonic complexes crop out in the cores of anticlinal ridges, but there are clear exceptions to this pattern (Fig. 4.15A). Corresponding K-Ar ages range from Dogger to Barremian, but seem to form two distinct groups, 175–155 ± 5 Ma and 135–110 ± 5 Ma, respectively (Fig. 4.15B). The occurrence of basalt flows on top of the Bathonian red beds and their reworking in the overlying Oxfordian-Kimmeridgian dolomites and conglomerates testifie for the Late Jurassic age of part of this magmatism (Charrière et al., 2005; Fig. 4.13A,B). Although some of the youngest ages could result from hydrothermal alteration (Zayane et al., 2002), the occurrence of Barremian lava flows is evidenced by stratigraphic data in the Aït Attab (Haddoumi et al., 2002) and Ouaouizarht synclines (Fig. 4.13A, C). In some anticlinal ridges, plutonic intrusives are overlain by detrital red beds of questionable age (Upper Jurassic, Lower Cretaceous, or even younger?).

The compositions of the High Atlas plutonic rocks are generally bimodal, i.e. silica-saturated or oversaturated. The mafic rocks range from troctolites to gabbros, and the intermediate to evolved, from diorites to monzodiorites and syenites. The latter group usually derives from the former through low-pressure fractional crystallisation coupled with assimilation of continental crust, and sometimes with magma mixing processes. Their magmatic signatures range from transitional basalt series rather similar to the Afar series (Zayane et al., 2002) to weakly/moderately alkaline series (Beraâouz et al., 1994; Lhachmi et al., 2001). The moderate enrichment in incompatible elements of these magmatic rocks, and the numerous evidences for crustal contamination that they display are typical of continental intraplate magmatism, often described in graben- and rift-related magmatic series.

The emplacement of strongly silica-undersatured alkaline magmas is also commonly observed in such rift-related settings. However, there is presently no evidence allowing to link the Late Jurassic/Cretaceous magmas with the Paleocene-Eocene nephelinites, carbonatites and phonolites of the northeastern Atlas domain, which occur only at discrete locations from Midelt (Tamazert) and Khenifra to Oujda and Taza (e.g. Sidi Maatoug; Chap. 5). We suggest that these Paleocene-Eocene magmatic rocks were emplaced in a specific geodynamic setting linked to some major tectonic change. According to their geochemical features and regional distribution, these lavas might derive from a giant asthenospheric plume which would have ascended below Western Europe and northwestern Africa during the Early Tertiary (Sect. 4.5). The Eocene volcanic events in the High Atlas (Tamazert) and Rekkame might be linked to the activity of this large-scale mantle structure.

4.3 The Orogenic Evolution of the Atlas System

The relative motions of the peri-Atlantic plates changed by the Late Cretaceous, being then characterized by convergence between Africa and Europe. Accordingly, block motions occurred in the Atlas and Meseta realm as early as the Albian-Cenomanian. The geometry of lands and seas progressively evolved, and by the Santonian, the Middle Atlas depocenters were connected to the Atlantic through the "Phosphate Plateau" (Figs. 4.17 and 4.19). Locally, tectonic shortening of Senonian age leads to folding due to syn-sedimentary inversion of inherited faults, with development of breccias along the faults and a clear unconformity of the overlying Eocene strata (Fig. 4.20). However, the deformation remained weak and local, without important relief building. As a matter of fact, the Atlas domain was still submerged until the Middle Eocene (Fig. 4.19), which corresponds to the actual beginning of the Atlas orogeny.

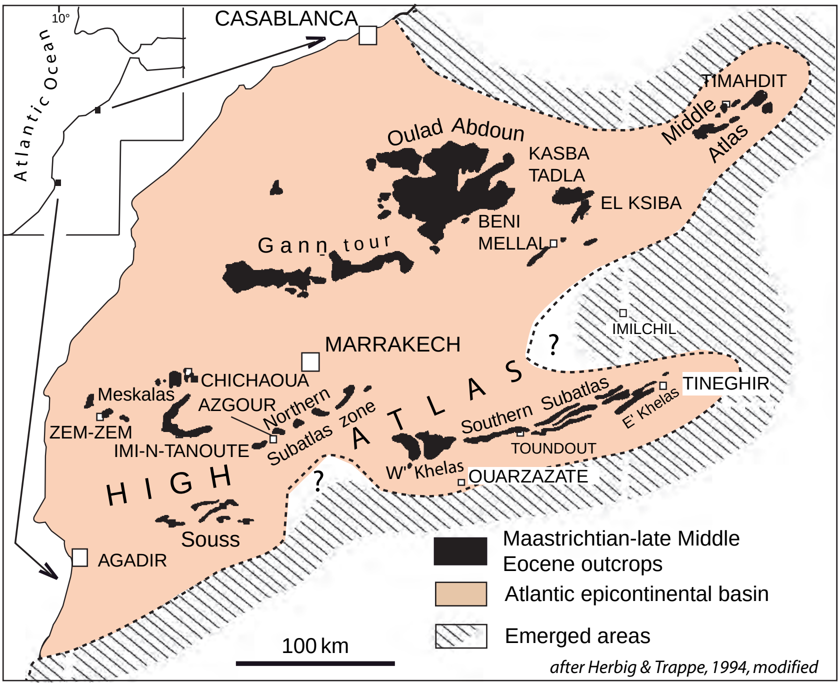

Fig. 4.19 Map of the Maastrichtian-Middle Eocene units in the Atlas system (from Herbig & Trappe, 1994, modified.). These units were deposited in an Atlantic gulf displaying a west-east zonation: siliceous-phosphatic facies (coastal basin: Meskala), phosphatic facies ("Plateau des Phosphates"), carbonate and mixed silicilastic-carbonate facies (Souss, northern and southern Subatlas Zones, Middle Atlas). In the Middle Atlas, Maastrichtian bituminous marls were deposited in small pull-apart basins. Small islands existed in the region of the central High Atlas (H.G. Herbig, *in litt*., 2007). A continental peninsula extended over the Central High Atlas, at least in the Imilchil area (Charrière et al., C.R. Palevol, submitted, Aug. 2008)

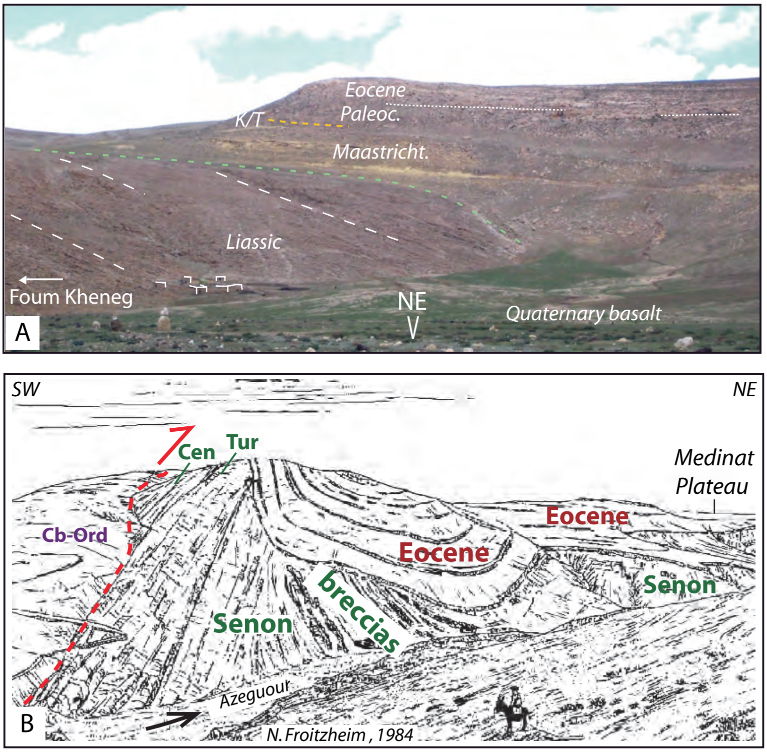

Fig. 4.20 (**A**) The Foum Kheneg unconformity along the north-west border of the Bou Angueur syncline (Martin, 1981; Herbig, 1988). Location: see Fig. 4.16B. The Maastrichtian and Lower to Middle Eocene beds rest unconformably onto Middle Liassic limestones indicating important erosion. – (**B**) Panorama on the folded unconformity between Eocene beds and tilted Senonian strata along the Azegour (Azgour) Fault, after Froitzheim (1984), modified. For location, see Fig. 4.25. Folding developed during the sedimentation as shown by breccias within the upper Senonian. The unconformity has been subsequently folded

The greatest part of information concerning the orogenic evolution of the Atlas system comes from the basins bordering the chain, where the most recent sediments are preserved. This is the reason why we present these data all around the chain, starting from the Middle Atlas to the Eastern High Atlas and then following the North Atlas Front towards west, and finally the South Atlas Front from west to east. The general tectonic interpretation will be discussed in Sect. 4.4 and the late to post-orogenic magmatism in Sect. 4.5.

4.3.1 The Middle Atlas and Adjacent Basins

References: The fundamental references for this domain are the following: Martin (1981), du Dresnay (1988), Herbig (1988), Benshili (1989), Fedan (1989), Charrière

(1990), Ennslin (1992) Herbig (1993).Gomez et al. (1996) and Elazzab & Wartiti (1998) published new data on the active tectonics, using seismicity and paleomagnetism, respectively. Beauchamp et al. (1996) and Gomez et al. (1998) used subsurface data in order to constrain the geometry at depth and the structural style. The studies by Krijgsman et al. (1999) and Zizi (1996, 2002), which concern the stratigraphy and structural geology of the Guercif Basin, led to a better understanding of the timing of deformation in the Middle Atlas. The Plio-Pleistocene evolution of the southeastern Middle Atlas front has been analysed by Laville et al. (2007).

The Middle Atlas region (Fig. 4.21) comprises three structural zones corresponding to distinctive paleogeographic domains during the Mesozoic: the Tabular Middle Atlas (Middle Atlas “Causse”), the Folded Middle Atlas and the Missour-High Moulouya Basin. The Middle Atlas “Causse” extends over the Western Meseta and Tazekka basement, and consists of tabular Triassic-Liassic sequences mostly detached from the Paleozoic basement. The Missour-High Moulouya Basin is

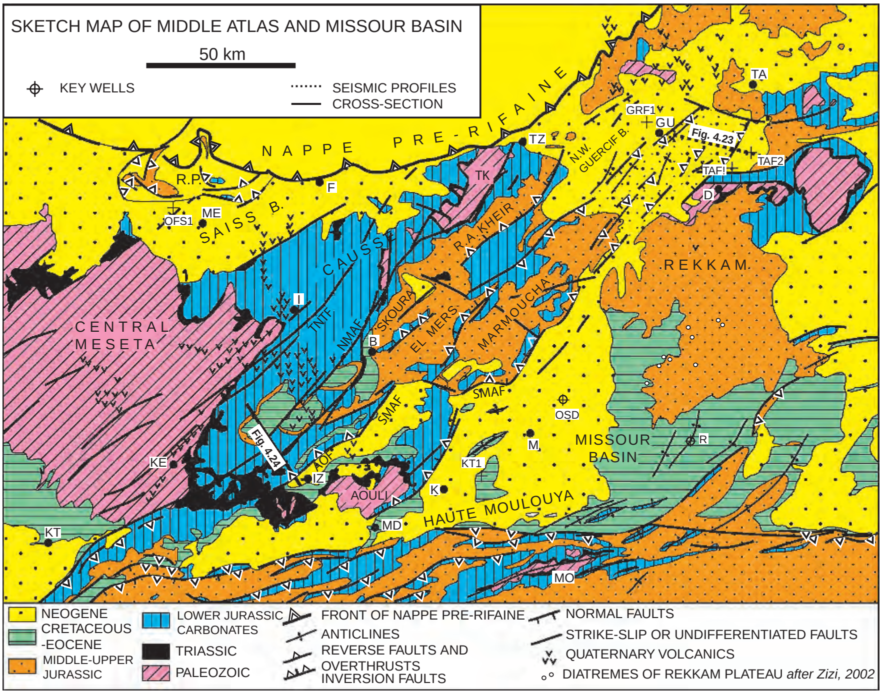

Fig. 4.21 Tectonic map of the Middle Atlas and Missour-High Moulouya and Guercif Basins, after Zizi (2002), modified. AOF: Ait Oufella Fault; B: Boulemane; D: Debdou; F: Fes; GU: Guercif; I: Ifrane; IZ: Itzer; K: Ksabi; KE: Kenifra; KT: Kasba Tadla; M: Missour; MD: Midelt; ME: Meknes; MO: Mougueur; NMAF: North Middle Atlas Fault; RP: Prerif Ridges (Rides Prérifaines); SMAF: South Middle Atlas Fault; TA: Taourirt; TNTF: Tizi n’Tretten Fault; TZ: Taza

the sedimentary cover of the Eastern Meseta that also exhibits mostly tabular Mesozoic-Cenozoic sediments, characterized by Liassic and mostly Middle Jurassic dolomites ("Dalle des Hauts Plateaux"). The Neogene Guercif Basin is located on the northern termination of the Middle Atlas and allows the connection with the South-Rif Corridor to the west and the Oujda Plain to the east. The folded Middle Atlas is bordered by two NE-SW trending faults, i.e. the North Middle Atlas Fault (NMAF) and the South Middle Atlas Fault (SMAF) becoming the Aït Oufella Fault (AOF) to the S-W (Fig. 4.21). The SMAF-AOF fault are complex zones of south-east verging faults and duplexes carrying the chain onto the Missour and High Moulouya Basins, where deformation propagates over several kilometres (Fig. 4.22). In the interior of the chain a net of braided faults and four long anticlinal ridges associated with the major faults delineate wide synclines or sub-basins.

Based on changes in the thickness of the folded strata toward the anticline ridges, several researchers proposed that folding initiated during Jurassic sedimentation (e.g. Fedan; 1989; Piqué et al., 2002; Laville et al., 2004). For these authors, the NE-SW to ENE-WSW trending anticlines correspond to left-lateral wrench faults in an overall strike-slip regime explaining also short E-W compressional ridges.

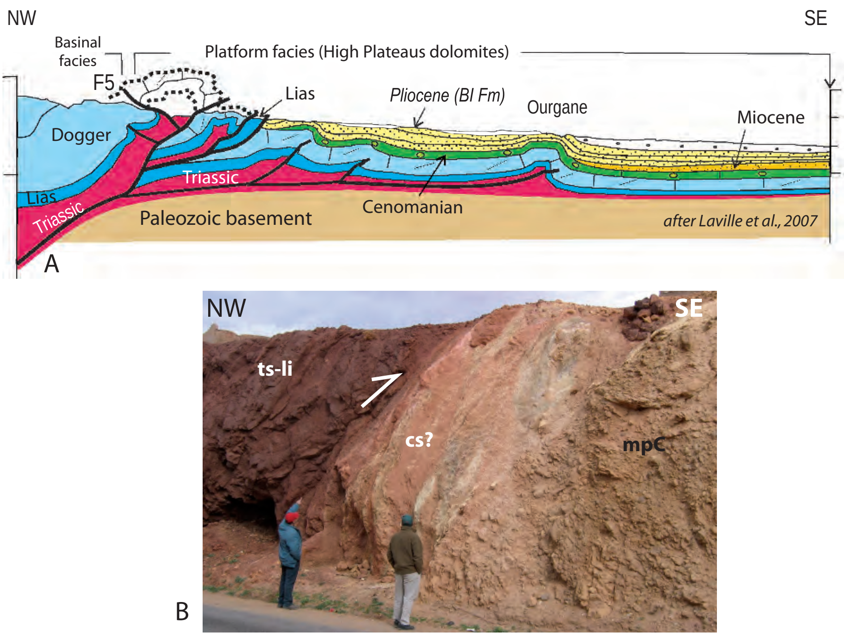

Fig. 4.22 The South Middle Atlas - Aït Oufella Fault Zone. – (**A**) Cross-section of the SMAF in its central segment south of the Middle Atlas highest peak (J. Bou Naceur, 3137 m a.s.l.), after Laville et al. (2007). Note the propagation of the deformation in the west margin of the Missour Basin. – (**B**) The Ait Oufella Fault as seen along the Midelt-Meknes road. The Upper Triassic-Lower Liassic red beds (ts-li) are thrust over Upper Cretaceous (?) gypsum marls (cs?) and overturned Miocene-Pliocene conglomerates (mpC). See Fig. 4.21 for location

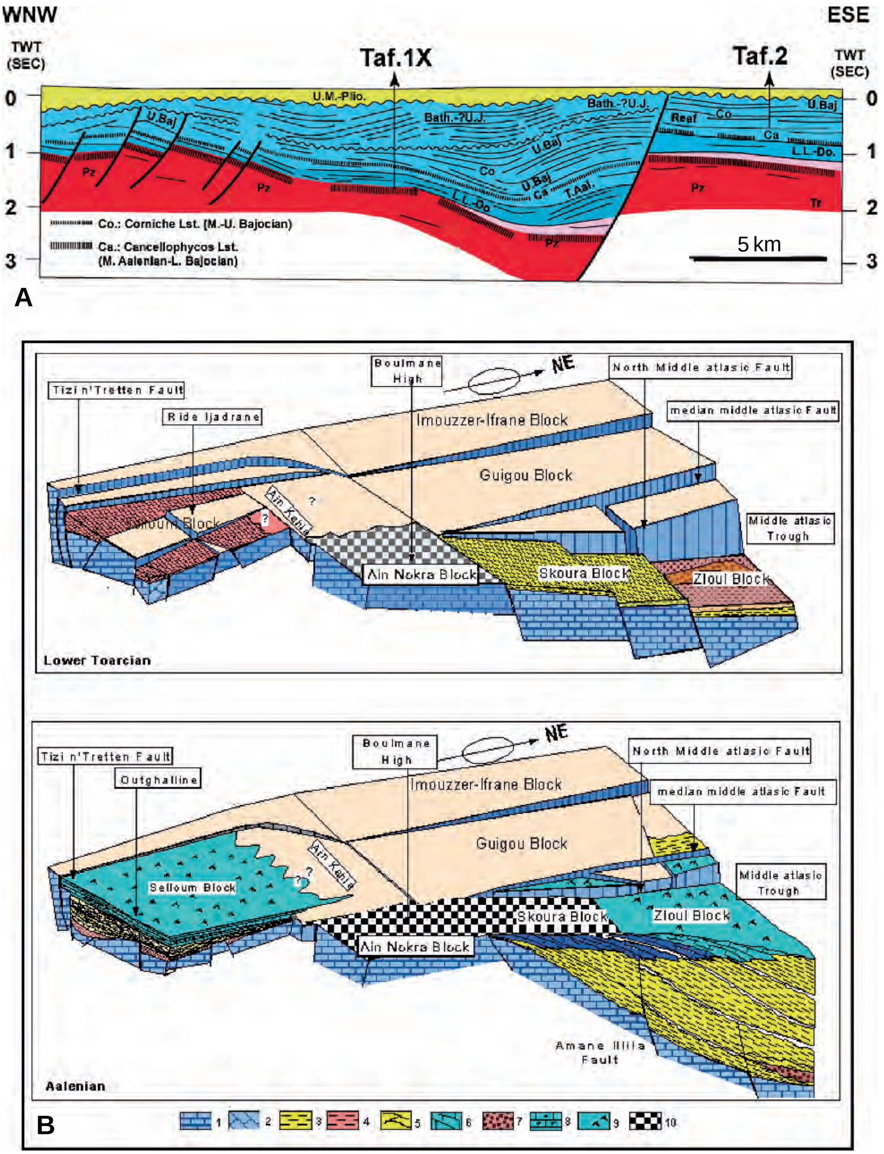

Fig. 4.23 Synsedimentary extension and block faulting in the Middle Atlas domain. – (**A**) Line-drawing of the seismic profile TAF2 crossing the Guercif Basin (from Zizi, 2002). This profile illustrates the half-graben geometry which developed in the Middle Atlas itself during the Early-Middle Jurassic. Location: see Fig. 4.21. – (**B**) Block faulting and sedimentary infill from Toarcian to Aalenian in southwestern Middle Atlas (from El Arabi et al., 2001). See Fig. 4.21 for location. The Middle Liassic rifting activated both longitudinal and transverse faults, resulting in a mosaic of tilted blocks with varied bathymetry. Faulting and block tilting continued during the Toarcian-Aalenian sedimentation

In this scenario, the compressional and extensional structures are interpreted as basically coeval and developed progressively during the Early-Middle Jurassic, whereas compression would have dominated during the Late Jurassic. Subsurface data show that these models, mostly developed during the eighties, do not sufficiently distinguish the Jurassic structures from those related to the Cenozoic inversion. The analysis of seismic profiles, in particular those crossing the Guercif Basin (Fig. 4.23A), as well as detail stratigraphic mapping (Fig. 4.23B) allow us to propose another interpretation, according to which the Jurassic period is dominated by extensional tectonics and block faulting with development of half-grabens bounded by ridges in the hanging wall of major normal faults. The thickness changes in the Jurassic beds are linked to the extensional movement along the faults. Unconformities are shown at the bottom of Toarcian, Bajocian and Bathonian beds. The extensional regime lasted until the Upper Jurassic. Therefore, the Skoura, El Mers and Marmoucha depocenters (Fig. 4.21) are interpreted as a half-graben system segmented by E-trending transfer faults. The fan-like geometry of the Jurassic units is related to block tilting during the rifting and progressive sedimentary infill, whereas the anticline-syncline geometry results from the Cenozoic shortening, with the anticline ridges being located along the crests of the tilted blocks.

The compressional regime leading to the inversion of the Middle Atlas Basin began during the Late Cretaceous, but culminated during the Late Eocene-Pleistocene. During the Late Cretaceous-Middle Eocene, the Middle Atlas was slightly submerged as shown by the occurrence of shallow water sediments. Deformation was limited to fault reactivation (cf. Fig. 4.20A). During the Cenozoic folding, a left-lateral component on the faults parallel to the chain is required in view of the obliquity of the Middle Atlas with respect to the more or less N-S shortening direction. This strike-slip component was mainly accommodated along the border faults, whereas in the folded Middle Atlas the shortening direction remained perpendicular to the main structures (concept of partition of deformation). Within the fold belt, the resulting shortening is low, slightly above 5 km following the available balanced cross-sections (Fig. 4.24), but additional shortening has to be considered in relation with the strike-slip movements along the main faults, and with the Tabular Middle Atlas detachment.

The uplift of the Middle Atlas domain is basically not due to shortening and isostatic rebound, but to thermal processes (lithospheric thinning) as discussed in Chap. 1 (Sect. 1.5) and below (Sect. 4.6). In this context, it is worth noting the occurrence of Middle Miocene to Quaternary basaltic rocks with some precursors as old as Paleocene in the whole Middle Atlas province (see Sect. 4.5). The detailed stratigraphy of the Neogene Guercif Basin shows that it emerged at 6 Ma. This result can reasonably be extended to the Middle Atlas mountain range where the unconformable Miocene deposits correspond to a major sedimentary cycle whose peak marine conditions are dated from the Messinian (Wernli, 1987). At Skoura, Pliocene conglomerates rest unconformably over the folded Mesozoic and Miocene sequences, and seal the J. Tichoukht Thrust Fault. However, they have been themselves subsequently deformed (Martin, 1981). Vertical axis rotations of

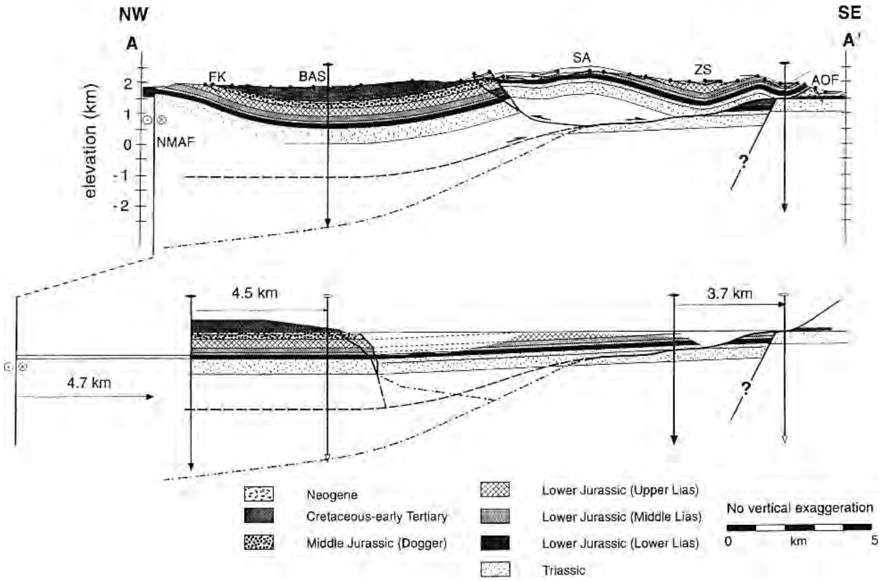

Fig. 4.24 Balanced cross-section across the central Middle Atlas fold belt and its restored section, after Gomez et al. (1998). Location: see Fig. 4.21. AOF: Ait Oufella Fault; BAS: Bou Angueur syncline; NMAF: North Middle Atlas Fault; SA: Lake Sidi Ali anticlinorium; ZS: Zad syncline. The section is balanced using the equal-area and line length methods. Dashed lines indicate different possible fault geometries at depth. A total of 4.7 ± 0.7 km of horizontal shortening is demonstrated within this part of the foldbelt regardless the deeper fault geometry, and most of the shortening is accomodated in the southern half of the cross-section. Shortening northwest of the Bou Angueur syncline may be greater because of possible horizontal shear along the NMAF. The nonreactivated normal fault near the AOF is interpretative, but may explain abrupt stratigraphic changes

Plio-Quaternary basalts, shown by paleomagnetism, and the diffuse regional seismicity confirm the persistence of tectonic activity in the Middle Atlas.

4.3.2 The Northern Flank of the Central High Atlas and the Tadla-Bahira and Haouz Basins

References: The main recent references comprise the studies by Rolley (1978), Petit et al. (1985), Dutour & Ferrandini (1985), Morel et al. (1993), Chellaï & Perriaux (1996), Salomon et al. (1996), Beauchamp et al. (1999), Benammi et al. (2001), Hafid et al. (2006), Najine et al. (2006), Missenard et al. (2007). The text below is also based on an unpublished synthesis by ONHYM, made under the supervision of Pr. A.W. Bally (Rice University, Houston, USA).

The junction between the Middle Atlas and the northern front of the High Atlas corresponds to a Jurassic ridge, which, east of Beni Mellal, separates the High Moulouya Valley from the Tadla Basin (Fig. 4.25). Surface geology shows a reverse

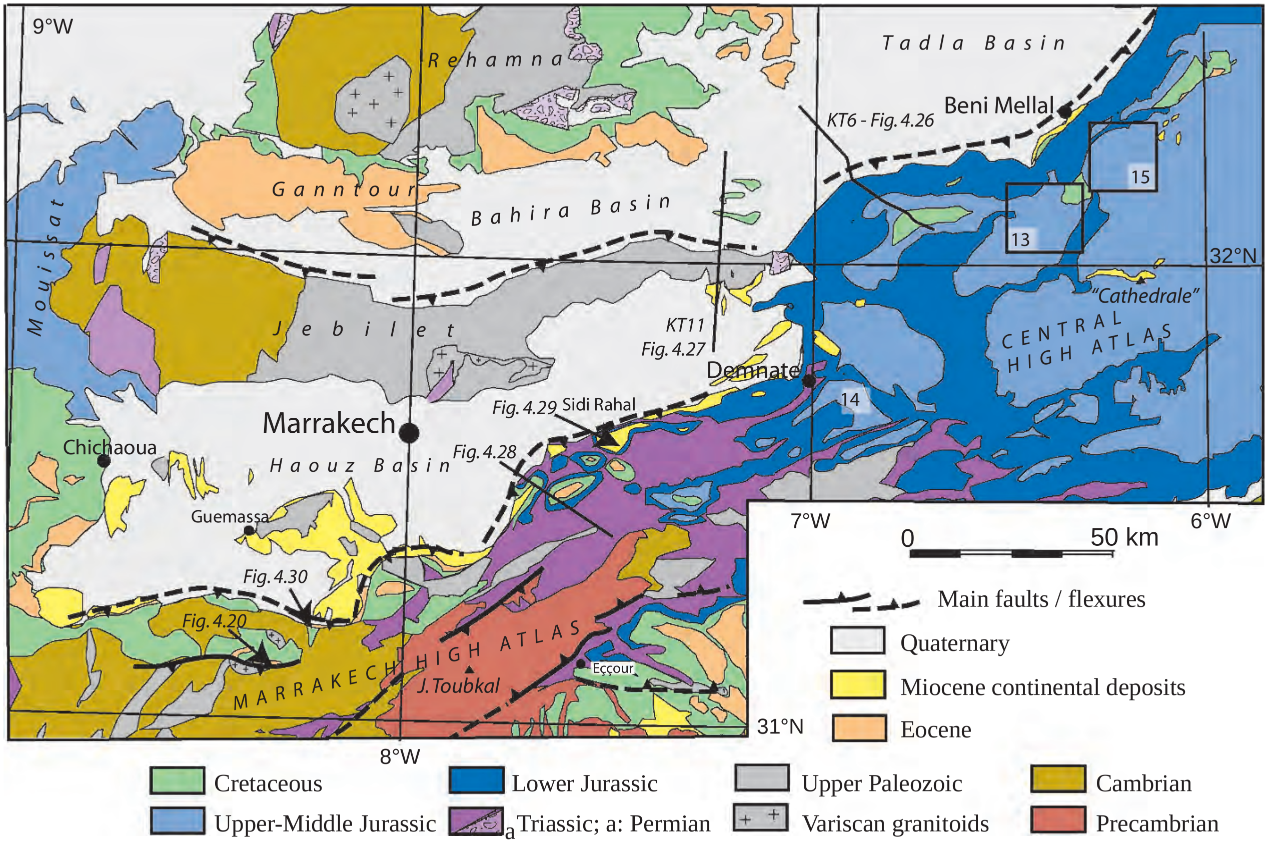

Fig. 4.25 Simplified geological map of the northern part of the Central High Atlas and neighbouring areas. Framed: location of Figs. 4.13 (Ouaouizarth), 4.14 (Iouaridene) and 4.15A (Tagalft = Taguelft)

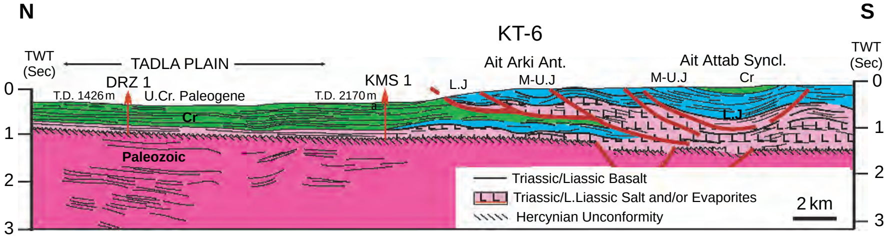

Fig. 4.26 Seismic profile KT6 across the northern front of the Beni Mellal High Atlas, interpretated by A.W. Bally and M. Zizi (unpublished). Location on Fig. 4.25. Note the thin-skinned tectonic style linked to the presence of the Triassic décollement level. The deformation propagated northward with large fault bend faults (Missenard et al., 2007)

fault along the latter basin. This structure is likely inherited from a former normal fault, as the seismic profiles show thinning or even lack of Jurassic beds in the Tadla Basin in contrast with the at least 2000 m thick Jurassic sequence observed in the hanging wall of the fault. In particular, the KT6 seismic profile cuts out the North Atlas Front as well as the Aït Attab syncline (Fig. 4.26). This profile, tied by different borehole data, shows very thin Mesozoic units beneath the Tadla Plain with thinning and then vanishing of Jurassic beds northward beneath the Cretaceous unconformity. The Cretaceous is well-developed (about 200 m of Lower Cretaceous, mostly continental deposits, and 400 m-thick Cenomanian-Turonian and Senonian marine to lagoonal sediments) and supports about 400 m of Tertiary layers. From the frontal slices to the Aït Attab syncline, the structures are developed over an important décollement level situated in the evaporites just above the Upper Triassic-Lower Liassic basalts. A basement fault may occur south of the Aït Attab syncline and should connect to the N-Jebilet Fault (Fig. 4.25). Therefore, the whole Mesozoic cover is probably duplicated above the Triassic décollement in the frontal part of the Beni Mellal Atlas.

Between the Bahira Plain and the Jebilet (Fig. 4.27), the seismic profiles show the existence of a major post-Middle Eocene E-trending, south-dipping reverse fault,

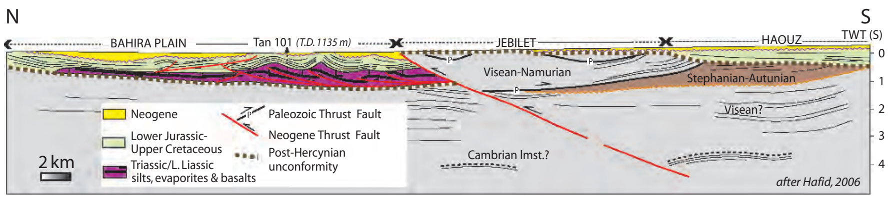

Fig. 4.27 Line drawing of seismic profile KT11 after Hafid (2006). Location on Fig. 4.25. Note the importance of the Neogene N-Jebilet reverse fault, the propagation of the deformation in the Bahira Basin (décollement over the Triassic beds), and finally the combination of thin-skinned and thick-skinned deformation

which accommodates about 4 km of shortening. If we consider the Jebilet range as part of the Atlas system, it is worth noting that it has not developed on the site of a former Triassic-Liassic graben. Moreover, the north-Jebilet fault is at right-angle to the Variscan synmetamorphic structures, and must be interpreted either as inherited from Permian structures (see Chap. 3), or as newly formed during the Cenozoic.

The Haouz Basin, where Marrakech City is situated, is located between the Jebilet and the High Atlas. Usually interpreted as the flexural basin at the front of the High Atlas, it rather belongs to the orogenic system and is better interpreted as an intra-mountain basin. The basin shows thin Mesozoic and Cenozoic units, ending with poorly dated Miocene-Pliocene molasse deposits, which onlap the Jebilet range to the north. In the southeastern part of the Haouz, the "Mio-Pliocene" molasses are coarser and thicker than in the north, and they are obviously folded. They are preserved in wide synclines located behind the frontal anticline (Fig. 4.25), on top of the Mesozoic-Eocene series of the so-called Northern Sub-Atlas Zone.

In the Demnate-Sidi Rahal segment of the latter zone, the folds (e.g. Ait Ourir bowls; Fig. 4.28) are cored by the Triassic-Liassic basalts and involve the thin Liassic series of the west border of the Tethyan gulf (cf. Fig. 4.4). A conspicuous unconformity separates the Mio-Pliocene beds from the folded Jurassic beds (Fig. 4.29). We observe here the effects of two superimposed folding phases in the building of the Atlas Mountain. The first tectonic step, probably Late Eocene in age (see Sect. 4.3.4), predates the sedimentation of the Miocene-Pliocene molasses. Some outcrops of such unconformable molasses are found in the core of the chain (Zawyat Ahansal or "Cathedral" conglomerates near Imilchil), which suggests that after the first folding step, the chain was covered extensively by

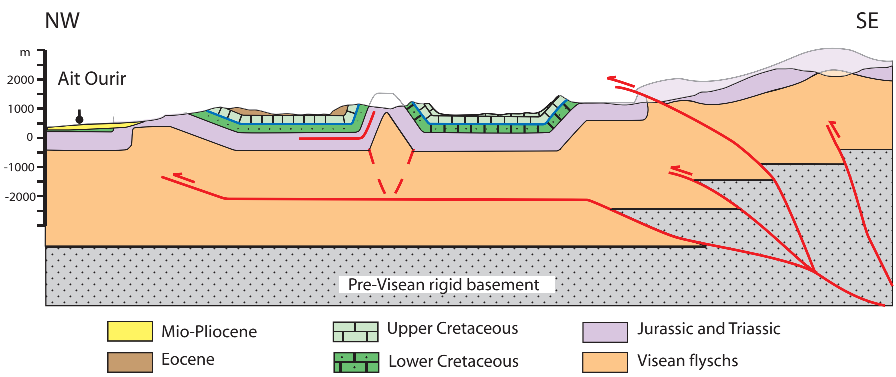

Fig. 4.28 Balanced cross-section of the North Atlas Front in the Aït Ourir area (from Missenard et al., 2007). Location on Fig. 4.25. The presence of thick Visean incompetent formations (dominantly flysch facies) allowed a deep décollement to propagate and trigger the development of large detachment folds (Ait Ourir bowls). The overall northward tilting of the Sub-Atlas Zone and the subsequent gravity-driven sliding of the innermost syncline on the Triassic evaporites described by Ferrandini & Le Marrec (1982) are not shown here

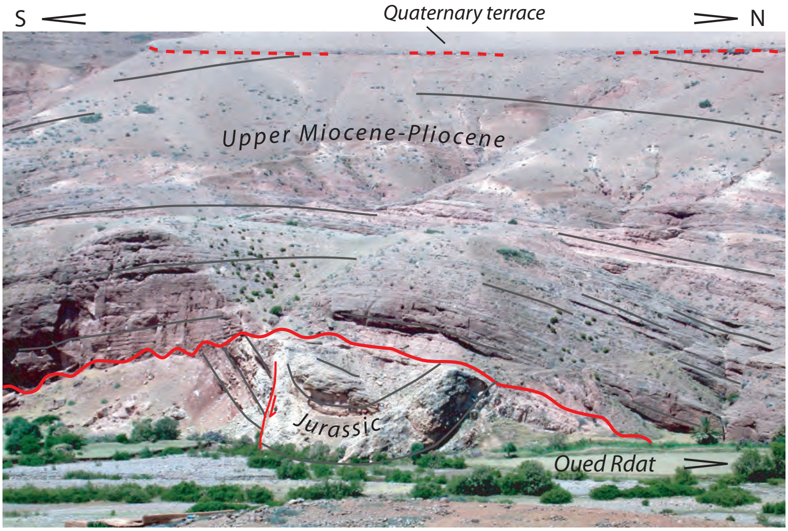

Fig. 4.29 Folded unconformity of the Miocene-Pliocene molasses over the previously folded Jurassic beds from the northern Sub-Atlas Zone, Sidi Rahal region southwest of Marrakech (see Fig. 4.25 for location). Notice the presence of growth strata in the molasses, showing the persistence of a compressional regime during the molasse deposition. The unconformable, almost horizontal terrace is assigned to the late Pliocene?-Villafranchian. After Missenard et al. (2007), modified

fluviatile-lacustrine deposits (similar to the Skoura conglomerates in the Middle Atlas). The second folding phase mostly postdates the molasses sedimentation. However, the molasses exhibit growth strata, which show that deformation persisted during their sedimentation. At the top of the pile the Pliocene-Quaternary horizontal beds are uplifted by about 300 m with respect to the adjacent plain, testifying a recent to active vertical motion. By place, Quaternary terraces are overlain by Triassic, Paleozoic or Precambrian formations along steeply dipping reverse faults (Dutour & Ferrandini, 1985).

Further to the west, the High Atlas is devoid of Jurassic deposits and shows only restricted Triassic series preserved in the Tizi n'Test (Oued N'Fis) corridor. There, the mountain range was built on the site of the West Moroccan Arch (cf. Figs. 4.4 and 4.7). In the Marrakech High Atlas, the substratum consists of Precambrian basement and poorly deformed Paleozoic series. In contrast, in the Western High Atlas Paleozoic (WHAP or Erdouz) Massif, the substratum displays strongly folded and recrystallized Paleozoic material, intruded by granite plutons such as the J. Tichka Massif. In both transects the front of the chain is underlain by a rapid basement step. It shows steep units of Late Cretaceous to Paleocene age (drape folds) over which the Mio-Pliocene molasses rest unconformably on a regional scale (Fig. 4.30). In the internal, highest part of the Sub-Atlas Zone, Late Cretaceous tectonic movements are recorded by the unconformity of the Eocene series on top of Senonian fault scarp breccias (cf. Fig. 4.20B).

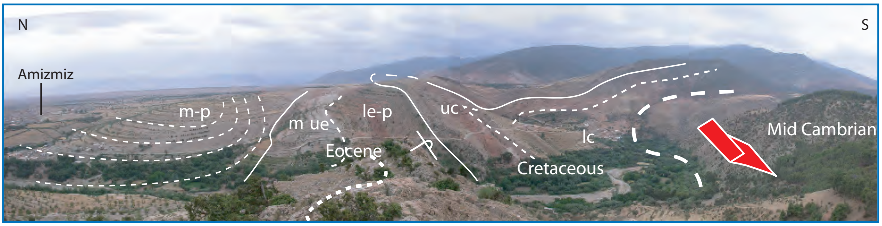

Fig. 4.30 The North Atlas Front in the Amizmiz area SE of Marrakech (from Missenard et al., 2007, modified). Location on Fig. 4.25. Notice the development of a "rabbit ear" fold in front of the main anticline cored by competent Cambrian rocks. The second order fold developed over the Senonian silts, which form an efficient décollement level. The Mio-Pliocene molasses rest unconformably on the older formations and contain growth strata as at Sidi Rahal (Fig. 4.29)

4.3.3 The Western High Atlas and the Essaouira-Haha and Souss Basins Onshore and Offshore

References: The evolution of the Atlantic margin as a whole is addressed in the Chap. 6, with references therein. The present section concerns the integration of the Atlantic margin in the Atlas system; the main references are Hinz et al. (1982a,b), Hafid et al. (2000, 2006), Bouatmani et al. (2003, 2004), Hafid (2006). For the Souss Basin, the useful references comprise Outtani (1996), Mustaphi et al. (1997), Frizon de Lamotte et al. (2000) and Sébrier et al. (2006).

The region studied now (Figs. 4.31 and 4.32) corresponds to the westernmost part of the High Atlas. It extends offshore until the western limit of the continental margin. An outline of the lithostratigraphy of this domain is given in Fig. 4.5. From a paleogeographic point of view, this area belongs to the Atlantic domain (Essaouira-Haha Basin) and is separated from the Tethyan domain by the West Moroccan Arch, which played a major role during the Liassic (cf. Fig. 4.4). The virtually undeformed, eastern part of the Basin is situated between the Jebilet and the Western High Atlas. It widens westward in the offshore Essaouira Basin. The deformed part of the basin (Haha Basin) corresponds to the Mesozoic Western High Atlas onshore (west of the Paleozoic Massif), whereas it intercepts the undeformed passive margin in the Cape Tafelney fold-belt.

At the end of the seventies and the beginning of the eighties, the "Deep Sea Drilling Project" (DSDP) involved drilling off the Moroccan coast. The main objective of this programme was to understand the stratigraphic and structural evolution of the Moroccan passive margin, but the specific question of the western termination of the Atlas was not addressed. The latter aspect has been addressed recently thanks to a set of new data from industrial seismic profiles.

The post-Variscan geological evolution of this domain began during the Late Triassic-Early Liassic (see Sect. 4.2.1) with the formation of NNE-SSW half-grabens, segmented by E-trending transform faults. These half-grabens were subsequently buried under a thick evaporitic sequence intercalated with scarce Middle

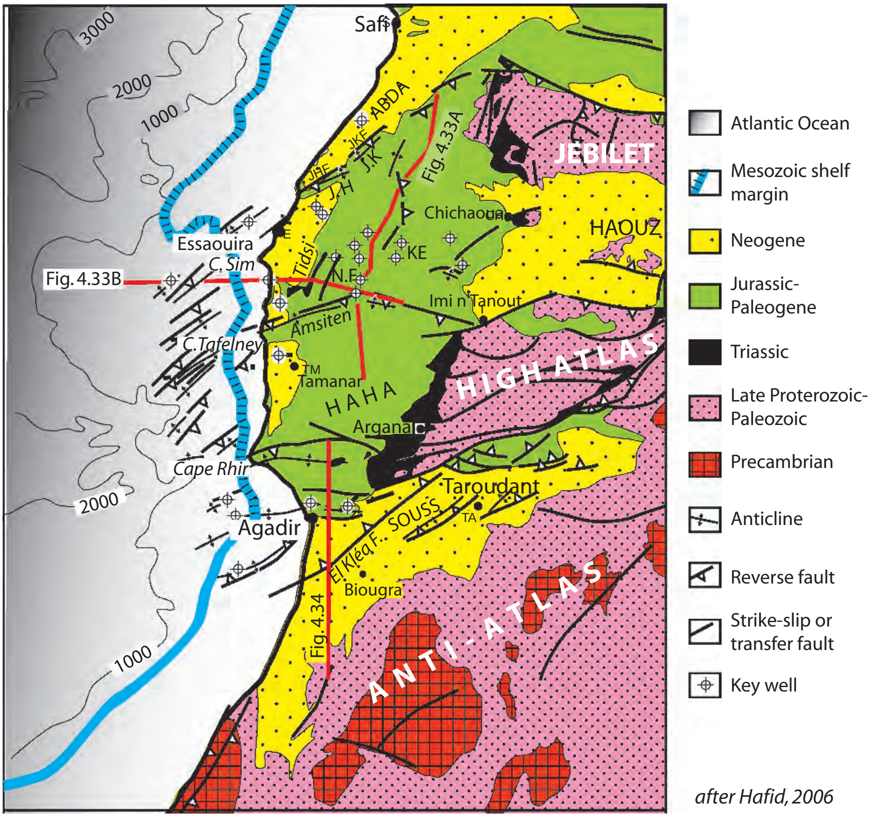

Fig. 4.31 Structural map of the Western High Atlas onshore and offshore, after Hafid et al. (2000) and Hafid (2006). Location: see Fig. 4.1. – J.H: J.Hadid; J.K: J. Kourati; K.E: Kechoula boreholes; NF: East-Necnafa Fault

Jurassic basalt (Amrhar et al., 1997). From the Late Jurassic to the Early Cretaceous, the Essaouira Basin and Western High Atlas formed a shallow platform which connected to the Atlantic margin. From the Late Cretaceous onward, the region suffered a general NNW-SSE compression, leading to the development of folds and reverse faults.

From a structural point of view, the onshore part of the Essaouira Basin appears on a N-S profile (Fig. 4.33A) as a wide synclinorium between the Jebilet and the Western High Atlas. This profile images the post-Cretaceous uplift of the Jebilet Massif as well as the existence of large evaporite-cored compressional structures such as the Kechoula one (KE1 and KE2 boreholes). Westwards, following a bending toward the SW, the Jebilet front crops out (Fig. 4.31) at the J. Kourati and J. Hadid structures which correspond to inverted Triassic normal faults. Westward again, the Cape Tafelney fold-belt marks the northern offshore front of the Atlas system. The structural axis linking the Jebel Kourati with the Cape Tafelney fold-belt

Fig. 4.32 Oblique satellite view of the NW corner of the Atlas system showing the transition between the Western High Atlas and Essaouira-Haha Basin (GoogleEarth website). See Fig. 4.31 for location

trends NE with a clear obliquity to the Atlas direction. It must be consequently interpreted as a lateral ramp.

A long E-W profile (Fig. 4.33B) documents the transition between a basinal domain to the west and a platform domain to the east. This transition occurs in the vicinity of the East-Necnafa Fault, along which an offset of the basement is observed. The importance of salt movements is well expressed. Timing of halokinesis along the East-Necnafa Fault is documented by the thinning of the Mesozoic series towards the TAB1 borehole, which shows that salt displacement began as early as the Early Jurassic and lasted until the Late Cretaceous. We also observe that only the westernmost faults, corresponding to the Atlas front, were inverted during the Cenozoic.

South of the Western High Atlas, the Souss Basin (Fig. 4.31) represents the southern foreland of the Atlas system. This basin exhibits a puzzling triangle shape due to a strong structural inheritance. The El Kléa Triassic normal Fault obliquely crosses the basin and connects to the well-known Tizi n'Test Fault, which is superimposed on the major limit between the Variscan Meseta Block and the Anti-Atlas domain (APTZ, see Chap. 3). The El Kléa Fault is buried below the younger sediments. However, industrial seismic data show that it was partly inverted during the Cenozoic (Fig. 4.34). Farther south, the Biougra Fault has not been inverted. Northward, the Tagragra anticline is interpreted as a fault-bend fold developed above the Upper Triassic-Lower Liassic salt layers. To the north, this décollement level connects with a basement fault continuing westward toward the Ameskroud Fault.

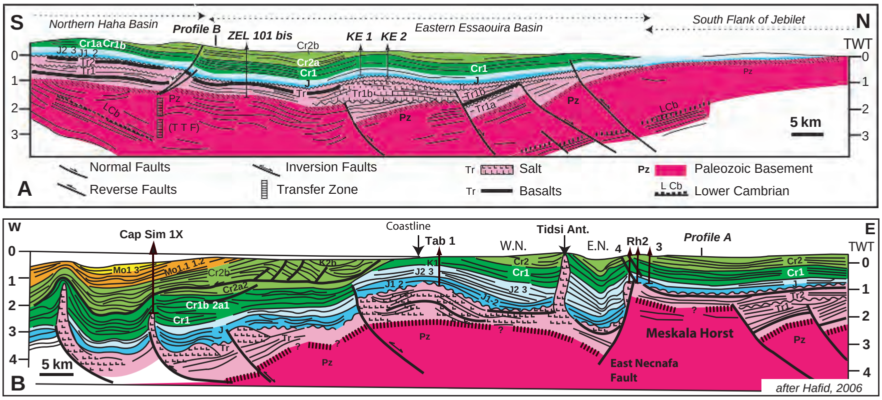

Fig. 4.33 Interpreted seismic profiles across the Western High Atlas and Essaouira-Haha Basin, after Hafid et al. (2000) and Hafid (2006), modified. Location: Fig. 4.31. Seismo-stratigraphic abbreviations: Fig. 4.5. – (**A**) Line drawing of the seismic profile 42C across the northern flank of the Western High Atlas and the Essaouira Basin. – (**B**) Line drawing of the seismic profile 44C across the Essaouira-Haha Basin onshore and offshore. The latter profile is roughly perpendicular to the main shortening direction and then mainly illustrates the role of salt tectonics. KE: Kechoula boreholes; W/EN: Western/Eastern Necfana

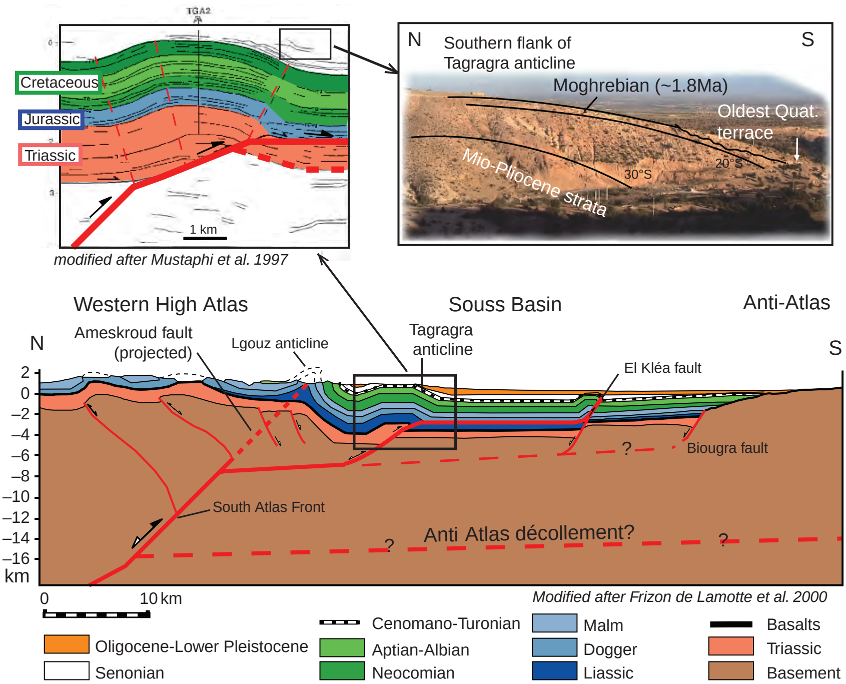

Fig. 4.34 Cross-section of the South Atlas Front and Souss Basin [from Sébrier et al. (2006) modified from Frizon de Lamotte et al. (2000)]. Location: Fig. 4.31. Note the tectonic inversion of the El Kléa Triassic Fault contrasting with the absence of inversion along the Biougra Fault further south

These structures are active, as evidenced by the earthquake which almost completely destroyed Agadir city in 1960. At a larger scale, the Souss Basin constitutes a dissymmetric synclinorium. Along its southern boundary, the regular, gentle dip of the Mesozoic reflectors outlines the recent Anti-Atlas uplift, the thermal origin of which being presently demonstrated (Missenard et al., 2006; see Chap. 1, this volume).

4.3.4 The South Atlas Front from the Eastern Souss Basin to the Boudenib Basin

References: A recent synthesis on the South Atlas Front is given by Frizon de Lamotte et al. (2000). For the Siroua shelf (between the Souss and Ouarzazate Basins), a recent reference is Missenard et al. (2007). Due to spectacular tectonic structures and the presence of a well-developed Cretaceous-Eocene series, the Ouarzazate Basin is somewhat over-studied (Laville et al., 1977; Fraissinet et al., 1988; Görler et al. (1988), Zylka (1988), Jossen & Filali-Moutei, 1992; Gheerbrant et al., 1993; Herbig & Trappe (1994); Errarhaoui, 1998; El Harfi et al., 2001; Benammi et al., 2001; Tabuce et al., 2005; Teson & Teixell, 2006). Eastward the data are from Saint Bézar et al. (1998) and the ONHYM synthesis under the supervision of A.W. Bally (Rice University, Houston). For the recent volcanism (Siroua Plateau, Saghro Massif and Ouarzazate Basin), see Sect. 4.5.

The structure of the South Atlas Front is extremely varied (Figs. 4.1 and 4.35). The classical geometry expected for a mountain front, involving a major thrust fault carrying allochthonous units onto a foreland basin, is observed only in the western and central parts of the system (Souss and Ouarzazate Basins, respectively). Elsewhere, the contact is either direct with the unflexured Sahara platform (Boudenib area), or uplifted to an unusual elevation (Siroua).

In the eastern part of the Souss Basin, the High Atlas Paleozoic and Triassic series are limited by a high-angle reverse fault connected to the Tizi n'Test Fault system. As in the western part of the basin (Fig. 4.34), a foothill domain of Paleozoic cored anticlines involving Late Cretaceous-Paleogene beds occurs between the Atlas range and the basin (Fig. 4.35A).

Further east, the Siroua Plateau forms a major topographic threshold between the Souss and the Ouarzazate Basins. This step is accentuated by a Miocene volcanic complex, namely the Siroua volcano (see Sect. 4.5). We suggest that the Siroua Plateau constitutes a wide compressional relay-zone between the South Atlas Front (SAF) to the north and Anti-Atlas Major Fault (AAMF) to the south (Figs. 4.1 and 4.25). However, the high elevation of the Siroua Plateau results also from the occurrence of the regional mantle anomaly discussed above (Sects. 1.5 and 4.1).

Immediately east of the Siroua area, i.e. in the "Khelas" (plateaus) area south of Telouet, the South Atlas Fault propagated as a blind structure within the Lower Paleozoic strata, leading to the development of a major back-thrust, which explains the

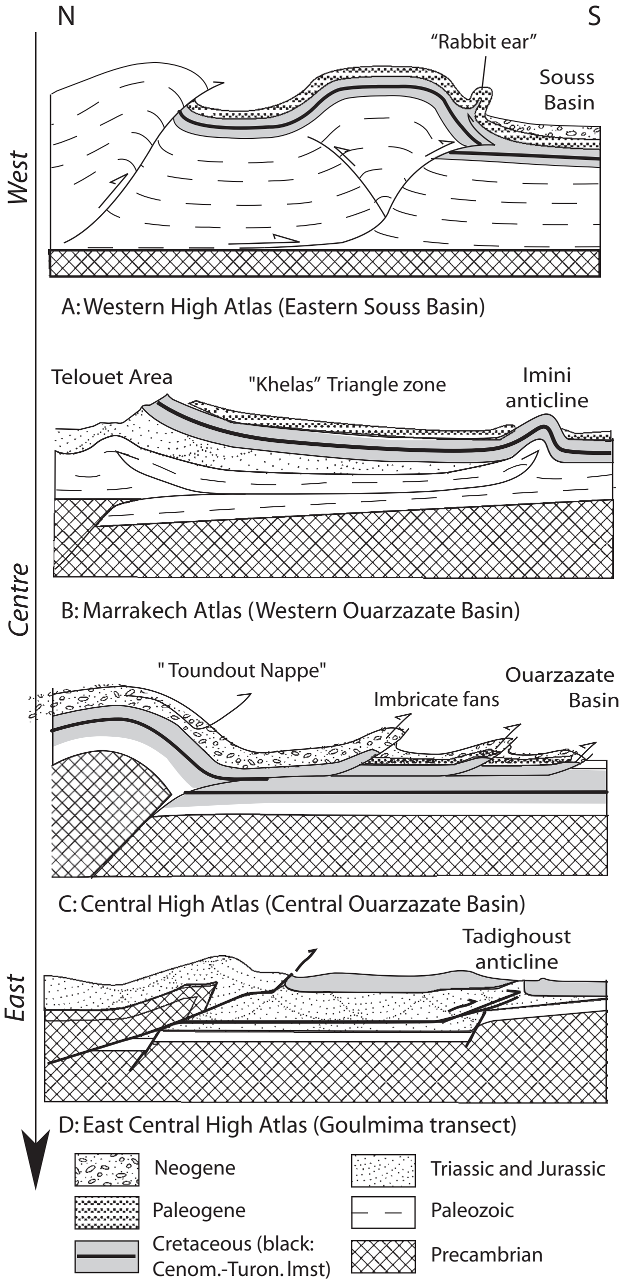

Fig. 4.35 Schematic cross-sections showing the changes of tectonic style along the South Atlas Front, in relation with the importance of the Paleozoic series on top of the Precambrian basement. **A, B, C** after Missenard et al. (2007); D after Saint Bézar et al. (1998). Location of the sections: see Fig. 4.1

south-dipping monocline formed by the overlying units (concept of triangle zone; Fig. 4.35 B).

Further east, the Ouarzazate Basin forms a narrow, 150 km long topographic low along the High Atlas (Figs. 4.1 and 4.2). The southern border of the basin lies over the northern flank of the J. Saghro (eastern Anti-Atlas domain), whereas the northern border is included in the South Sub-Atlas Zone, more or less deformed along the South Atlas Front (Fig. 4.35C). The activation of a décollement in the Senonian gypsum-bearing red beds determined the development of a small, but typical imbricate fan, clearly documented in the natural cross-sections (Fig. 4.36) as well as by seismic-reflection (e.g. Toundoute area; Benammi et al., 2001). The stratigraphy of these imbricate units yields critical dates for the mountain building chronology. The earliest record of the Atlas uplift corresponds to the onset of continental sedimentation, sourced in the uprising belt, during the Late Eocene (Hadida and Ait Arbi red beds; Figs. 4.36 and 4.37). The onlap of Oligocene?-Lower Miocene deposits (Ait Ouglif Fm) above the folded and eroded Mesozoic-Eocene beds allows the first significant folding event of the Sub-Atlas Zone to be dated as Late Eocene-Early Miocene. The lacustrine facies of the Middle-Late Miocene deposits (Ait Kandoula Fm) suggest a relative tectonic relaxation at that time. Subsequently, shortening resumed during the Pliocene-Quaternary, being recorded by thrusting and refolding structures. Therefore, the South Sub-Atlas Zone yields evidence of two main shortening periods, Late Eocene-Oligocene and Pliocene-Pleistocene, respectively (Fig. 4.38), similar to the North Sub-Atlas Zone (cf. Sect. 4.3.2).

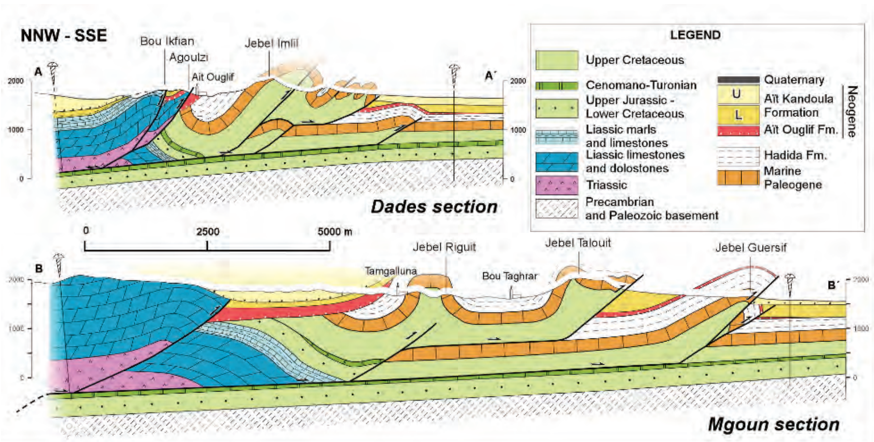

Fig. 4.36 Balanced cross-sections of the Southern Sub-Atlas fold-thrust belt, after Teson & Teixell (2006). Location on Fig. 4.39. – AA': Dades Valley north of Boumalne. – BB': Mgoun Valley, 20 km further west. Notice the décollement level situated within the Senonian gypsum silts, and the superimposed unconformities of the Oligocene?Miocene Aït Ouglif Fm. and Miocene-Pliocene Ait Kandoula Fm. These cross-sections suggest shortening values of about 7–8 km accomodated at a low rate of about 0.3 mm/y during the last 20–25 Ma

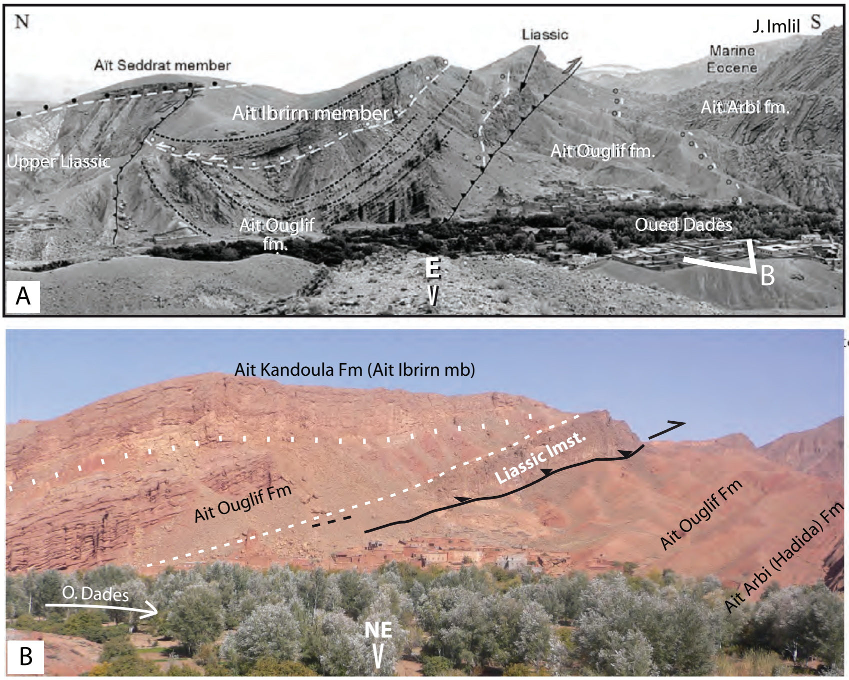

Fig. 4.37 The Southern Sub-Atlas fold-thrust belt in the Dades Valley north of Boumalne, Ait Ouglif area. Location: see cross-section AA′, Fig. 4.36, and map Fig. 4.39. – (**A**) Wide-angle photograph and geological interpretation by Teson and Teixell (2006). Three tectonic units are visible, from south to north, (i) the J. Imlil unit, mainly Cretaceous-Eocene, ending with the Ait Arbi and Ait Ouglif continental deposits, Upper Eocene and Oligocene?-Miocene, respectively; (ii) a thin, Liassic cored unit (Agoutzi sliver on cross-section AA’, Fig. 4.36) showing two superimposed, folded unconformities, i.e. the Ait Ouglif Fm. onlap onto the Liassic limestones and the Ait Ibrirn member (Ait Kandoula Fm, Miocene-Pliocene) unconformity over the Ait Ouglif folded conglomerates; and (iii) a massive Liassic unit (cf. Bou Ikfian nappe on cross-section AA′) overlain by the unconformable Ait Seddrat member (Pliocene) of the Kandoula Fm. – (**B**) Frontal view of the unconformity between the Ait Ouglif and Ait Ibrirn red beds, as seen from a more southern point (located on **A**). This unconformity records a Late Oligocene?-Early Miocene folding event

Seismic data show the lack of Triassic beds below the Ouarzazate Basin. This is a major difference with the Souss Basin which could explain, at least partly, the different subsidence rates observed between the two basins: the Triassic rifting could be responsible for the low rigidity of the basement of the Souss Basin.

East of the Ouarzazate Basin, the transition between the Atlas and its foreland is not outlined by any flexural basin, indicating a low tectonic load and, probably, a rigid behaviour of the Sahara Platform, which had not been weakened by the Triassic rifting. The South Atlas Front is localized near the inherited normal faults bounding the Atlas Basin. In front of it, smaller structures developed using the décollement level located within Upper Triassic-Lower Liassic evaporites (Fig. 4.35D).

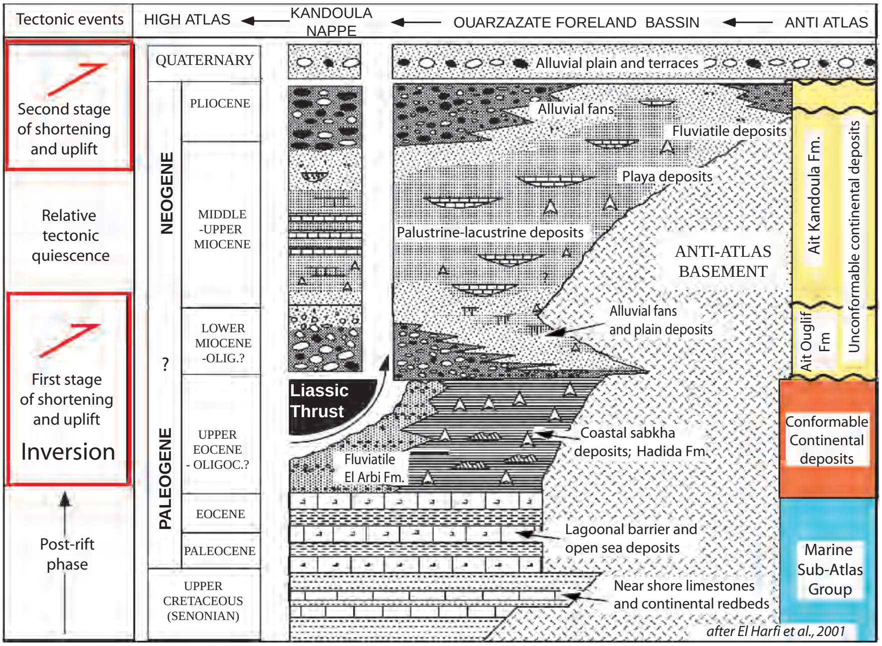

Fig. 4.38 Lithostratigraphic table for the Ouarzazate Basin after El Harfi et al. (2001) modified according to Teson & Teixell (2006). Notice the occurrence of coarse conglomerates in Oligocene?-Early Miocene and Pliocene-Pleistocene series, following the two main tectonic episodes. These conglomerates are sourced from the raising High Atlas. The Hadida-El Arbi red sands and pebbles, first considered as sourced in the Anti-Atlas (Görler et al., 1988), are more likely recycled Lower Cretaceous material from the uplifted High Atlas (Teson & Teixell, 2006)

The propagation of Cenozoic deformation south of the South Atlas Front, i.e. in the Anti-Atlas domain, has been recognized (see Chap. 7, Sect. 7.3), even if its mechanism (thrusting or strike-slipping) remains unknown. The dissymmetry of the Anti-Atlas topography suggests that a tectonic component acted during its formation. The moderate seismicity recorded in the area, for instance the Md=5.3 Rissani earthquake in 1992, confirms the persistence of deformation.

4.4 Structure of the Atlas Chain: Orogenic Processes and Overall Shortening

References: The more comprehensive cross-sections of the Atlas system have been presented by Poisson et al. (1998), Beauchamp et al. (1999), Morel et al. (1999), Frizon de Lamotte et al. (2000), Benammi (2002), Teixell et al. (2003), Arboleya et al. (2004) and Missenard et al. (2007). The ductile deformation (axial planar cleavage) which occurs along the anticlines in the axial zone of the Central

High Atlas was described by Laville & Harmand (1982), Jacobshagen et al. (1988), Brechbühler et al. (1988), Laville and Piqué (1992), with divergent interpretations.

In the above sections, we have focused on the geometry of the High Atlas borders, because the age of the tectonic events is better defined there by the occurrence of recent deposits and availability of industry seismic data. A particular Sect. (4.3.1) has been dedicated to the structure of the Middle Atlas. We now focus on the core of the High Atlas, and address the questions of its deep structure and of the amount of shortening accommodated in the orogenic system. For this purpose, we examine the recently published, synthetic cross-sections of the Central High Atlas (CHA) and Marrakech High Atlas (MHA) located along the accessible transects of the belt (Fig. 4.39).

The two easternmost sections follow the classical road from Midelt to Errachidia at the transition between the Central/Eastern High Atlas (Figs. 4.40A and 4.41A). The slight differences in the surface geology of the two sections are due to their slightly different positions. More significant differences are observed in the structural geometry at depth. Following Benammi (2002), the Paleozoic substratum is divided into blocks separated by former normal faults of the Mesozoic rifting. No detachment level is proposed in the post-Paleozoic cover. The inherited faults are only partially inverted, i.e. their reverse movement does not completely balance their previous normal movement, except for the north and south boundaries. In contrast, for Teixell et al. (2003) as well as for Arboleya et al. (2004), the Paleozoic substratum is clearly folded together with the cover despite the existence of a decoupling zone at the bottom of the Mesozoic unit. Moreover, almost all the inherited faults are completely inverted.

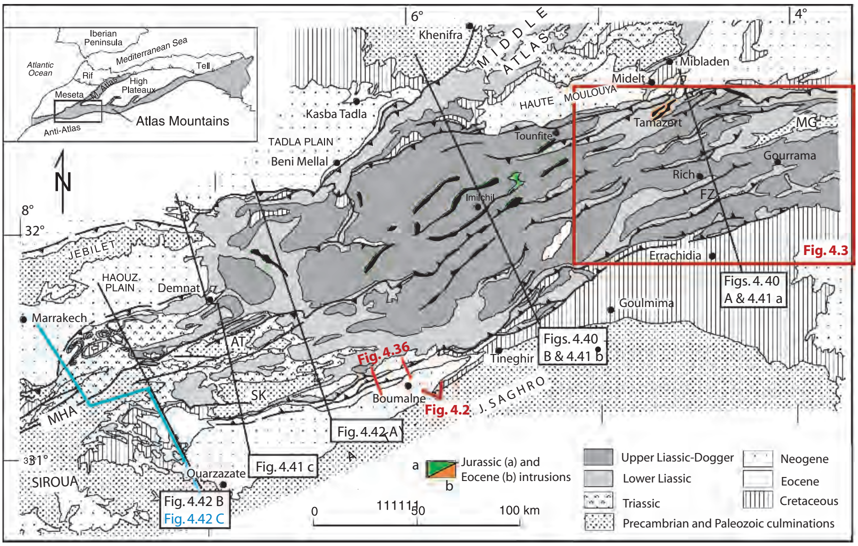

Fig. 4.39 Geological map of the Central High Atlas (from Teixell et al., 2003, modified) with location of the cross-sections Figs. 4.36, 4.40–4.42 and of the Figs. 4.2 and 4.3

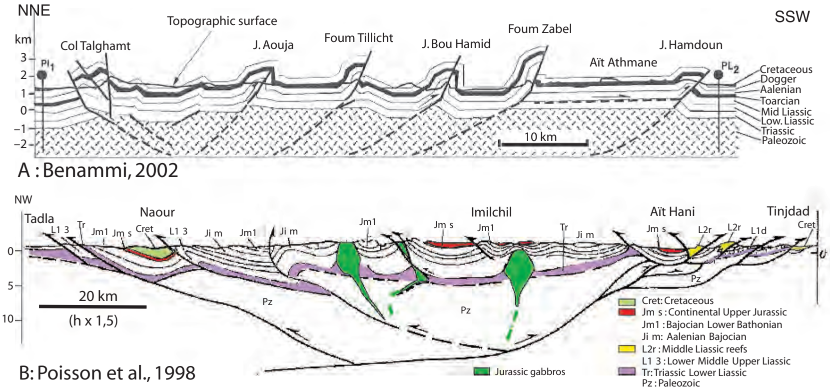

Fig. 4.40 Generalized cross-sections of the Central and Eastern High Atlas (see Fig. 4.39 for location). – (**A**) Balanced cross-section of the classical Errachidia-Midelt (Oued Ziz) transect, after Benammi (2002). The author proposes a shortening value of about 17 km between pile lines PL1 and PL2 (now distant by 75 km). Compare with Fig. 4.41a. – (**B**) Crustal-scale cross-section proposed by Poisson et al. (1998) through the widest segment of the High Atlas. This was formerly the more stretched part of the Atlas rift, characterized by large gabbro intrusions during the Middle-Late Jurassic

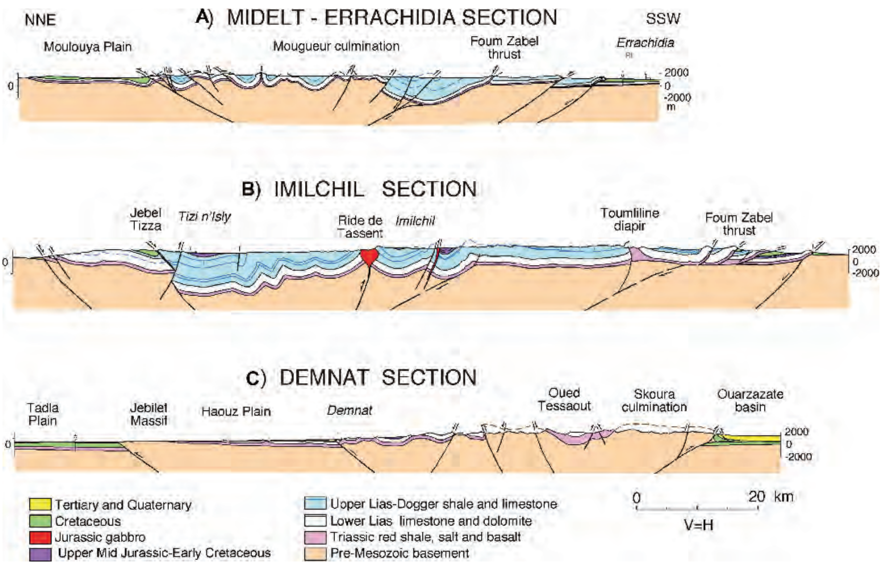

Fig. 4.41 Serial geological cross-sections through the High Atlas (see Fig. 4.39 for location) slightly modified from Teixell et al. (2003): (**A**) Midelt-Errachidia section; shortening estimate: 26 km, contrasting with the estimate by Benammi (17 km; see Fig. 4.40A). (**B**) Imilchil section (compare with Fig. 4.40B). (**C**) Demnat-Skoura section; note the propagation of the crustal shortening up to the southern Meseta domain (North Jebilet reverse fault)

Another critical cross-section more to the west has been represented by two different groups (Figs. 4.40B and 4.41B). The transect corresponds to the wider segment of the CHA, between the High Atlas-Middle Atlas junction to the north and the Ouarzazate Basin to the south, passing through the Lake Plateau (Imilchil) and the main anticlinal ridges intruded by Jurassic gabbros. In both interpretations of this profile, the complete decoupling along the Triassic beds and the particular intensity of the deformation along the borders of the chain are evidenced. However, for Poisson et al. (1998), the northern front is controlled by a Triassic décollement whereas, to the south, the ramps directly emerge from the Paleozoic. For Teixell et al. (2003), detachment within the Triassic level is also responsible for the imbricate structures shown along the southern front. Moreover, the basement deformation is interpreted differently. For Poisson et al. (1998), the decoupling between Triassic and Paleozoic units remains weak. Folding in the cover occurred as a response to thrusting in the substratum. At depth, the thrust faults display a low-angle attitude and merge on a crustal detachment, equivalent to that already proposed by Giese & Jacobshagen (1992). For Teixell et al. (2003), like in the eastern section of the same authors, the pre-Mesozoic series are folded together with the cover and form wide detachment folds. They do not propose an interpretation for the geometry at depth. None of these authors distinguishes Paleozoic terranes from the Precambrian basement, which obviously acts as the mechanical basement.

In both the preceding transects, ductile deformation (axial planar cleavage) occurs along the anticlines of the axial zone, particularly close to the plutonic rocks. Syn-tectonic, low- to very low-grade metamorphic conditions have been estimated in the same area (Tounfit, Imilchil, Rich) based on the illite cristallinity method and fluid inclusions by Brechbühler et al. (1988). This can be assigned to burial metamorphism, as the Triassic-Jurassic series alone are about 7–8 km thick (Studer, 1987), below potential Cretaceous-Cenozoic successions. In line with Jacobshagen et al. (1988), we consider that the synmetamorphic cleavage can be assigned to the Cenozoic deformation (accentuated against the rigid plutonic bodies), contrary to the interpretation of Laville & Harmand (1982) and Laville & Piqué (1992).

Two sections cross the western termination of the former Jurassic gulf along the Demnat transect (Fig. 4.41C) or slightly east (Fig. 4.42A), from the Tadla Plain to the Skoura culmination and the Ouarzazate Basin. We note on both sections the existence of a shallow detachment level in the Senonian beds that permits the propagation of the deformation in the southern border of the chain. To the north, the Aït Attab syncline is detached over the Triassic beds. Teixell et al. (2003) assume the same basement behavior as in the more eastern sections, i.e. without any separation between the Paleozoic strata and the Precambrian basement, the cover being harmonically folded with the substratum. Beauchamp et al. (Fig 4.42A) differentiate the basement from the Paleozoic cover and propose a duplex geometry for the Paleozoic. However, such geometry is not constrained and their restored section (with flat Paleozoic layers) is not consistent with what is known from the Variscan structures in the region.

The Marrakech High Atlas (MHA) is cored by Precambrian rocks. This region of the High Atlas has been the subject of a limited number of geological studies. Only

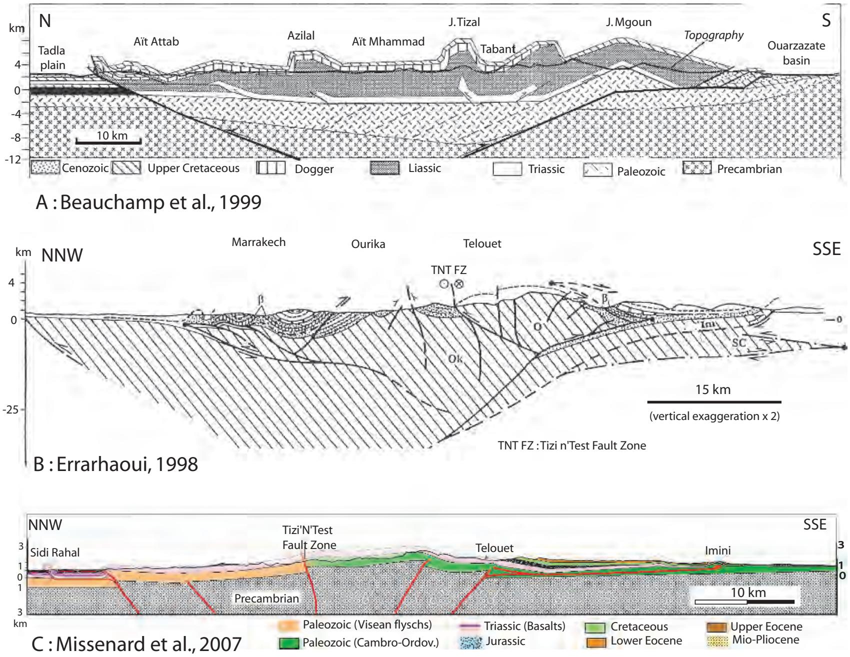

Fig. 4.42 Cross-sections of the Central High Atlas-Marrakech High Atlas regions (see Fig. 4.39 for location). – (**A**) Balanced profile from the Toundoute-Mgoun area to the Tadla Plain, after Beauchamp et al. (1999). The shortening estimate is ca. 36 km, about three times more than the estimate by Teixell et al. (2003) for the neighbouring traverses (Fig. 4.41B, C). – (**B**) Cross-section of the Demnate-Telouet traverse (Haouz Plain-western Khelas) after Errarhaoui (1998). – (**C**) Cross-section of the same traverse without vertical exaggeration, after Missenard et al. (2007). Shortening seems to be much reduced across the Marrakech High Atlas, about 10 km according to Morel et al. (1999) or even less according to Missenard (2006)

three complete sections exist at the moment. Note that Léon Moret, who mapped the region in 1931 (see Chap. 9), already distinguished different mechanical units allowing a decoupling between the substratum and the Mesozoic-Cenozoic cover (the accuracy of his data along this section is exceptionally good). In her thesis (1998), K. Errahraoui proposed a complete section of the MHA (Fig. 4.42B). Her interpretation is close to that proposed by Poisson et al. (1998) further east (Fig. 4.40B). We note the existence of a mid-crustal detachment and a deep duplex to explain the uplift of the western termination of the Ouarzazate Basin. The Paleozoic is not differenciated from the Precambrian. Missenard et al. (2007) separated the Precambrian basement (with a rigid behavior) from the Paleozoic series within which the tectonic style depends on the structural heritage. In the southeastern regions, where the Variscan deformation is mild, the Paleozoic series are folded together with the overlying Mesozoic beds (Fig. 4.42C). In the western region (not presented here) where the Variscan deformation is accompanied by metamorphism and granitic intrusions, the Paleozoic substratum behaves as a rigid material and the Upper Triassic décollement is activated. The southern flank of the Atlas and the contact with the Ouarzazate Basin are interpreted as a large triangle zone without duplex at depth.

At the western termination of the High Atlas, the combination of the cross-sections by Frizon de Lamotte et al. (2000) (Fig. 4.34) and Hafid et al. (2006) (Fig. 4.33A) gives an idea of the structural style in this area. It is worth noting the importance of the decoupling along the lower Liassic salt and evaporites. The metamorphic Paleozoic substratum, intruded by granitic plutons, behaves as a rigid body together with the underlying Precambrian basement.

The appraisal of shortening values considerably differs from one section to the other, and even for a given section, from one author to another. The estimates are comprised between 10 and 35 km of shortening. According to Teixell et al. (2003), the shortening varies from 20 km in the Marrakech-Ouarzazate transect up to 30 km in the Midelt-Errachidia transect. The occurrence of strike-slip deformation along oblique faults, and that of ductile shortening could increase these estimates. In fact, axial planar dissolution cleavage occurs in the anticlines of the axial domain of the CHA, being developed at T ~ 300 °C under 7–9 km of sedimentary load (Jacobshagen et al., 1988; Brechbühler et al., 1988). In the MHA, the important strike-slip movements along the dense fault array oblique to the axis of the belt makes almost impossible to reasonably estimate the amount of shortening. However, the present-day tendancy is to accept shortening ratio ranging from 10 to 20%, because we know that the Atlas relief is not due solely to crustal shortening, but also to thermal phenomena in the upper mantle (see Chap. 1, and Sects. 4.5 and 4.6 below). In any case, deep seismic profiles would be required for a better understanding of the deep geometry of the Atlas system. We express the wish that they will be done in the near future.

4.5 The Cenozoic Volcanic History of the Atlas System

References: The volcanic geomorphology of the Middle Atlas has been described by Martin (1981), and the corresponding units mapped by Faure-Muret and Meslouh (2005) for the Azrou area and by Baudin et al. (2001,b) for the Oulmès region. K-Ar datings of lavas have been published by Harmand & Cantagrel (1984), Berrahma (1995), Rachdi (1995) and El Azzouzi et al. (1999). The petrogenesis and the tectonic setting of the alkali basalts and related lavas have been discussed by a number of authors, including Bernard-Griffiths et al. (1991), Berrahma (1995), Rachdi (1995), Mourtada et al. (1997), El Azzouzi et al. (1999, 2003), Maury et al. (2000), Coulon et al. (2002), Savelli (2002), Duggen et al. (2005), Teixell et al. (2005) and Missenard et al. (2006).

The Cenozoic volcanism of the Atlas system is exclusively of intraplate alkaline type (alkali basalts, basanites, nephelinites, and associated intermediate and evolved lavas), whereas in the Rif it evolved through time from calc-alkaline to shoshonitic and finally alkaline (see Chap. 5, Fig. 5.46). The Atlas volcanism is located within a SW-NE trending strip, underlain by thinned lithosphere (Frizon de Lamotte et al., 2004; Teixell et al., 2005; Zeyen et al., 2005; Missenard et al., 2006), that we propose to call the *Morocco Hot Line*. This trend extends towards the Mediterranean coast near Oujda where it is dated from 6.2 to 1.5 Ma (El Azzouzi et al., 1999), and in the Oran area, Algeria (4 to 0.8 Ma, Coulon et al., 2002). It could be connected with the linear trend defined by the Pliocene-Quaternary alkaline lavas of southeastern Spain and southern France (see below, Sect. 6).

According to the available K-Ar datings, the Atlas volcanism is mostly Middle Miocene to Quaternary in age (14.6–0.3 Ma). However, some older (Eocene) ages have been published. Dykes and sills of lamprophyres, phonolites, nepheline syenites, nephelinites and carbonatites crosscutting the alkaline intrusion of Tamazert (High Atlas) and its limestone country rocks have been dated between 45 and 35 Ma (Bernard-Griffiths et al., 1991). Other Eocene ages obtained on the Zebzat nephelinite in the Middle Atlas (35 ± 3 Ma, Harmand & Cantagrel, 1984) and on basanites and nephelinites from Rekkam (55.2 to 33.8 Ma, Rachdi, 1995), need to be confirmed.

The main volcanic edifice of the Anti-Atlas domain is the Siroua, a dissymetric massif ca. 30 km large, which ranges from Late Miocene to Pliocene in age (10.8 to 2.7 Ma, Berrahma, 1995). Its southern part consists of intermediate (trachybasaltic, trachyandesitic) and evolved (trachytic) lava flows, whereas necks and dykes of peralkaline trachytes and rhyolites associated to pyroclastics and phonolitic domes occur in its northern part (Berrahma, 1995). In the Jebel Saghro (or Sarhro) area, located east of Siroua, nephelinite, tephrite and phonolite formations are mostly dated from 9.6 to 6.7 Ma, but younger (4.8 to 2.9 Ma) nephelinite flows containing carbonatite xenoliths also occur. In Central Morocco (Oulmès region), young volcanics dated from 2.8 to 0.3 Ma (Rachdi, 1995) have been mapped (Baudin et al., 2001a). They include ca. 20 strombolian cones and short lava flows of melilite nephelinites, nephelinites, basanites and tephrites, and several phonolite domes.

The Middle Atlas basaltic province comprises the largest and youngest volcanic fields in Morocco. A hundred well-preserved strombolian cones and maars occur along a N-S trend ca. 70 km long between El Hajeb and Itzer (Figs. 4.43 and 4.44). Some maar deposits (Tafraoute, Bou-Ibalrhatene) contain large and abundant lithospheric mantle xenoliths (spinel lherzolites, pyroxenites) and lower crustal (granulitic) ones. Numerous fluid basaltic flows emitted from the strombolian cones, some of them 30–50 km long, overlie the dolomitic limestones of the Middle Atlas Causse. The K-Ar datings indicate two periods of emplacement, during the Miocene (14.6 to 5.5 Ma) and, mainly, during the Quaternary (1.8 to 0.5 Ma). However, the occurrence of younger eruptions cannot be discarded given the excellent preservation of volcanic landforms. The total surface covered by the volcanic rocks is rather large (960 km^2), but the corresponding volume remains low (~ 20 km^3) because of the limited thickness (20–30 m) of the lava accumulation.

Four types of mafic lavas are distinguished in the petrologic map (Fig. 4.43), based on a hundred new major and trace element analyses. Intermediate and evolved compositions are lacking, a feature which contrasts with other Moroccan volcanic fields (Oulmes, Siroua and Saghro). Nephelinites (SiO_2 = 36–41%) usually form small strombolian cones and associated lava flows located along the borders of the

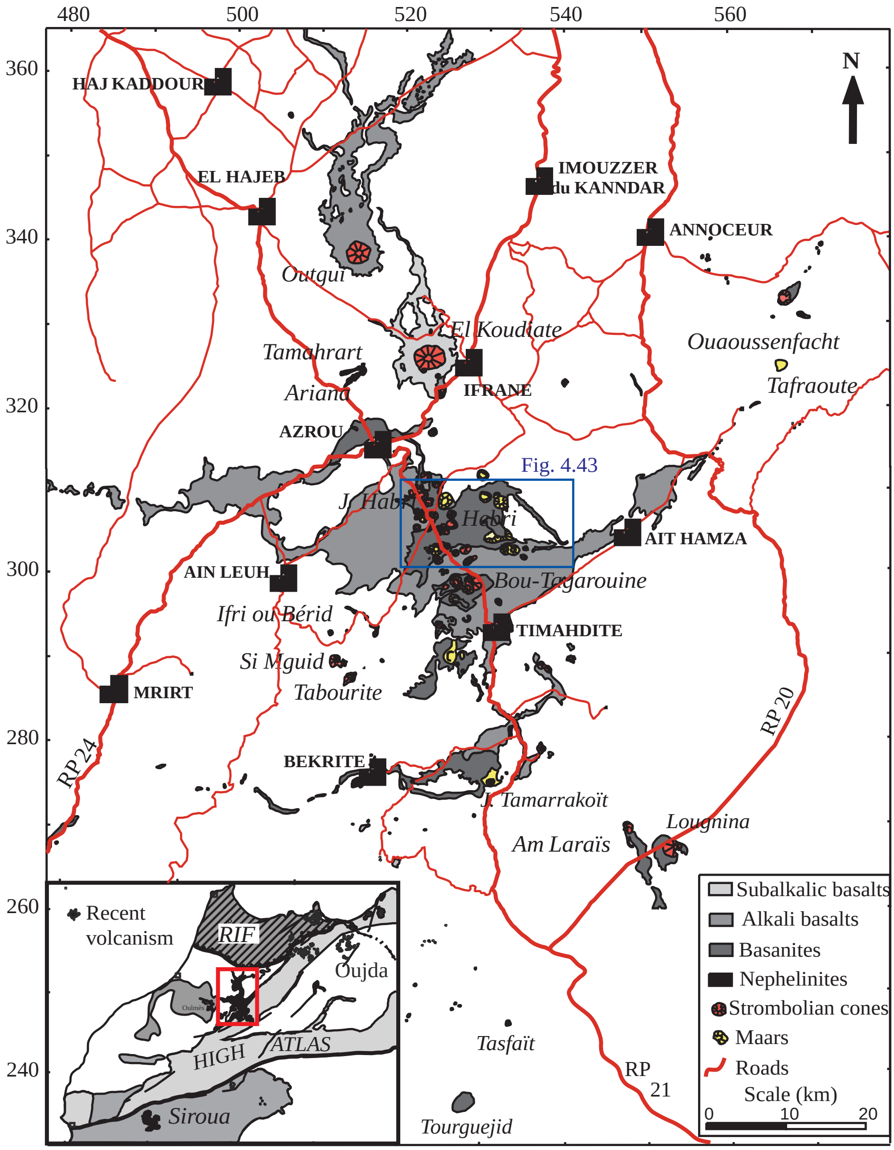

Fig. 4.43 Petrologic map of the Middle Atlas basaltic province, after El Azzouzi, *in prep*. The percentages of the total surface covered by the various lava types are 68.5% for the alkali basalts, 22.5% for the basanites, 7.8% for the subalkaline basalts and only 1.2% for the nephelinites. The latter include the oldest (Miocene) lavas exposed in the chain

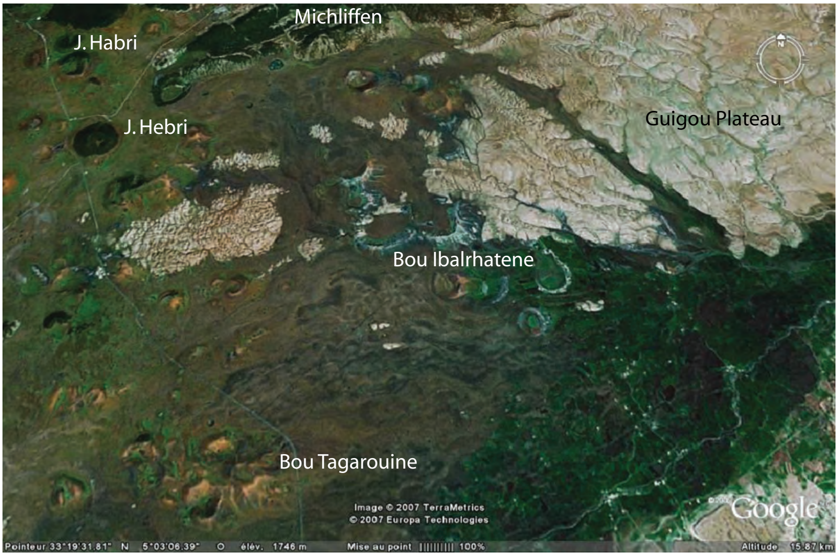

Fig. 4.44 The Middle Atlas basaltic province. View from 15 km above Timahdite, GoogleEarth image with small obliquity. Location: see Fig. 4.43. Note the young J. Hebri and J. Habri strombolian cones and the maars further east (J. Bou Ibalrhatene)

volcanic plateau, and most of them were emplaced prior to the other petrologic types. The basanites (SiO_2 = 41–45%) are the youngest lava type, and make up most of the well preserved cones located between Azrou and Itzer. The corresponding lava flows generally overlie the alkali basalt flows. Alkali basalts (SiO_2 = 46–51%) represent the dominant petrographic type, and their fissural lava flows cover most of the plateau surface (Fig. 4.44), especially to the east (Oued Guigou Valley) and the west (Oued Tigrigra Valley) of the main volcanic axis. They also form the large northern cone of J. Outgui, whose flows covered the Quaternary formations of the Saiss plain. Finally, subalkaline basalts, richer in silica than the former types (SiO_2 = 52%), make up the El Koudiate cone and associated 20 km long lava flows.

According to available datings (Harmand et Cantagrel, 1984; El Azzouzi et al., 1999), the Middle Atlas Miocene volcanic events emplaced only nephelinites, from 14.6 Ma (Bekrit) to 5.9 Ma (Talzast). However, nephelinites also erupted during the Quaternary, around 1.6 Ma (J. Tourguejid) and 0.75 Ma (J. Tahabrit). Alkaline et subalkaline basalts as well as basanites seem to be exclusively Quaternary in age, and the youngest published ages have been measured on basanites (0.8 Ma at J. Tahabrit, 0.6 Ma at J. Am Larais, 0.5 Ma at J. Aït el Haj). The alkali basalts, basanites and nephelinites display strongly enriched incompatible element patterns (Fig. 4.45). Their geochemical signatures are typically intraplate alkaline, and hardly distinguishable from those of ocean island alkali basalts (OIB) and related rocks. The progressive enrichment in the most incompatible elements observed from alkali basalts to nephelinites (Fig. 4.45), is consistent with decreasing degrees of partial melting of an enriched mantle source.

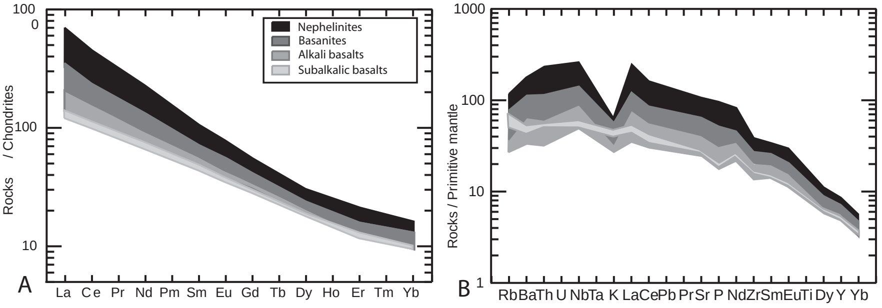

Fig. 4.45 Chondrite-normalised rare earth element patterns and primitive mantle-normalised incompatible element patterns showing the geochemical diversity of the Middle Atlas lavas. Alkali basalts, basanites and nephelinites represent relatively primitive magmas (9% < MgO < 13%, 45 < Co < 60 ppm). The El Koudiate subalkaline basalts display some petrographic (quartz xenocrysts) and geochemical (selective enrichments in Rb, Th, K, depletion in Nb) evidences for crustal contamination. They probably derive from alkali basalt magmas, contaminated by continental crust during their ascent

Sr, Nd et Pb isotopic ratios measured on alkaline lavas from Tamazert (Bernard-Griffiths et al., 1991), Middle Atlas and Oulmès (El Azzouzi et al., 1999) and Oujda area (Duggen et al., 2005) indicate an enriched mantle source, with almost no radiogenic Sr, showing rather variable Nd isotopic ratios, and consistently rich in radiogenic lead. This isotopic signature is close to the HIMU end-member recognized in oceanic islands such as St. Helens and Tubuai. Such a signature is frequently found in Cenozoic alkali basalts and basanites from Europe, the western Mediterranean, northern Africa, and eastern Atlantic islands (Madeira, Canary archipelago). These lavas are thought to derive from a 2500 to 4000 km large giant asthenospheric plume which would have ascended below these areas during the Early Tertiary (Hoernle et al., 1995). The Eocene volcanic events in the High Atlas (Tamazert) and Rekkame might be linked to the activity of similar large-scale plume.

However, the small volume of Miocene to Quaternary lavas erupted within the Atlas domain seems hardly consistent with the activity of such a giant plume. In addition, their silica-undersaturated character implies small degrees of melting of their mantle source (less than 5% for the nephelinites). Morevoer, the strong negative K spikes observed in incompatible multi-element patterns (Fig. 4.45) indicate that their source contained residual hydroxyl-bearing minerals (pargasite and phlogopite), a feature which implies that this source was in a lithospheric situation during partial melting. These minerals, which commonly appear during magma-mantle interactions, could have formed when alkaline magmas derived from the large-scale Early Tertiary plume percolated through the sub-Atlas mantle (Duggen et al., 2005). The thermal anomaly responsible for the lithospheric thinning below the Atlas volcanic fields could also have generated the Miocene to Quaternary alkaline volcanism. The corresponding partial melting would be linked to the thermal erosion of the base of the lithospheric mantle, previously enriched through metasomatic interactions with the Early Tertiary giant plume-derived magmas. This melting process might have started at around 15 Ma below the southern Middle Atlas, generating the Miocene nephelinites, and then propagated toward SW (Siroua, Saghro) and NE (Guilliz, Oujda) along the Morocco Hot Line, crosscutting the earlier tectonic boundaries. The lack of correlation between the age and the geographic position of the Moroccan alkaline volcanoes is indeed typical of hot lines (Easter Island: Bonatti and Harrison, 1976; Cameroon: Déruelle et al., 2007) and contrasts with the regular trends observed for Hawaiian-type hot spots.

4.6 Conclusion: Geodynamics of the Atlas System

The Moroccan Atlas illustrates numerous aspects of the evolution of an intra-continental orogenic domain and in particular the relationships between tectonics, magmatism and geodynamics. We outline below some points that we consider as essential.

The Atlas system is developed on the site of a rift domain presenting a complex geometry and chronology. We emphasize the existence of numerous fault directions ranging from NNE to ENE and the different timing from one side of the West Moroccan Arch to the other. The latter block formed a common shoulder at the beginning of the rift process (Late Permian-early Late Triassic), but changed into a submarine high during the Late Triassic-Liassic. Extensional tectonics resumed during the Liassic or even the Dogger only in the Tethyan domain. The normal faults defining the half-grabens are partly inherited from Variscan faults. They developed under a general extensional and transtensional regime during the opening of Central Atlantic and the related left lateral movement between European and African plates. The net result was the splitting of the Atlas Tethyan domain into numerous sub-basins forming a complex network.

As shown by subsurface data as well as by a reappraisal of magmatic data, there is no convincing evidence for a major compressive deformation during neither the Late Jurassic nor the Early Cretaceous as proposed by some authors from local and disputable data. The shortening in the whole Atlas system began during the Late Cretaceous with the onset of convergence between Europe and Africa. It began likely by large scale buckling of the whole lithosphere then developed essentially during the Cenozoic.

The Atlas relief is not related solely to crustal shortening, which remained low (less than 20%). A thermal factor is involved in the Atlas uplift, being linked to an oblique NE-SW strip of thinned lithosphere (the Morocco Hot Line), which extends from the Western Anti-Atlas and Siroua region, to the Central High Atlas, to the Middle Atlas and, finally, to the Eastern Rif at least, considering inland Morocco. To illustrate the importance of this thermal component on the relief, we have extrapolated the data obtained along four lithospheric cross-sections to draw a virtual topographic map, i.e a map of what the Moroccan Atllas system would be without lithospheric thinning (Fig. 4.46). On this map, the geography is quite different from the actual one: the Souss and Tadla basins are below sea level, the Anti-Atlas

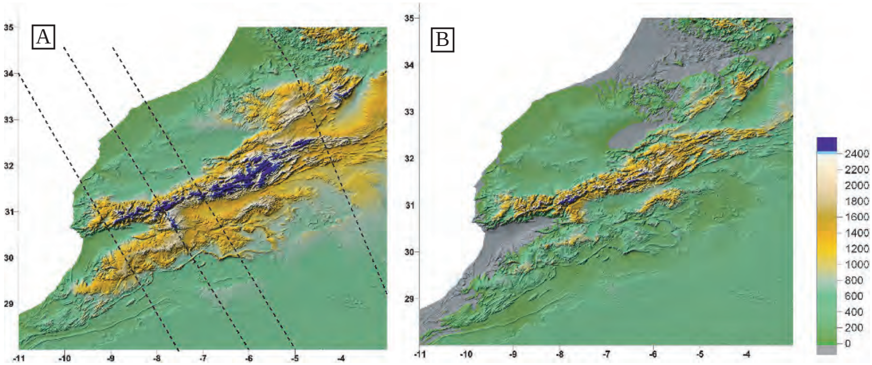

Fig. 4.46 Comparison of the present topography of the Atlas system of Morocco (**A**) with its "virtual" topography resulting from the subtraction of the thermal component (**B**). The thermal component of topography has been calculated on four profiles and then extrapolated using minimum curvature method (modified from Missenard, 2006; see explanations in the text, and Missenard et al., 2006)

disappears and the High Atlas looses about 1000 m of elevation. Last but not least, the South-Rif corridor, which was a marine connection between the Atlantic Ocean and the Mediterranean Sea during the Late Miocene, is restored as a sea-way in this hypothetic geography. Therefore, we may suggest, as a working hypothesis, that the Messinian salinity crisis of the Mediterranean area could result partly from thermal doming of the NW African corner, and not exclusively from the Rif nappe emplacement and subsequent compression, as currently proposed.

The Morocco Hot Line is outlined by a diffuse seismicity and by an intraplate-type alkaline volcanism, Miocene to Quaternary in age. The emplacement of the lithospheric anomaly, being responsible for all these phenomena, took place during the Middle-Late Miocene. Consistently, the recent apatite fission track analysis by Missenard et al. (2008) in the Marrakech High Atlas yielded ages between 9 ± 1 Ma and 27 ± 3 Ma, suggesting that this area underwent significant denudation during the Miocene.

The timing of the Cenozoic inversion events in the Atlas remains a matter of study and debate. From this point of view, it would be useful to gain reliable datings of the so-called Mio-Pliocene molasses which sealed a first tectonic event. Overall, the compressional tectonic processes were not progressive from the Late Cretaceous up to the Present, but they occurred by more or less discrete steps. By reference to what is known elsewhere in the Maghreb, we propose that the two main steps occurred during the Middle-Late Eocene and the Pliocene-Quaternary, respectively, disregarding subsidiary movements during the Oligocene(?)-Miocene. The origin of such discontinuity in the tectonic shortening could be related to the spatial and temporal pattern of the Mediterranean subduction(s) and to the variable efficiency of the coupling between the African and European plates (Fig. 4.47). It seems that the periods of strong coupling would correspond respectively to the initiation (Fig. 4.47B) and cessation (Fig. 4.47G) of the subduction roll-back processes

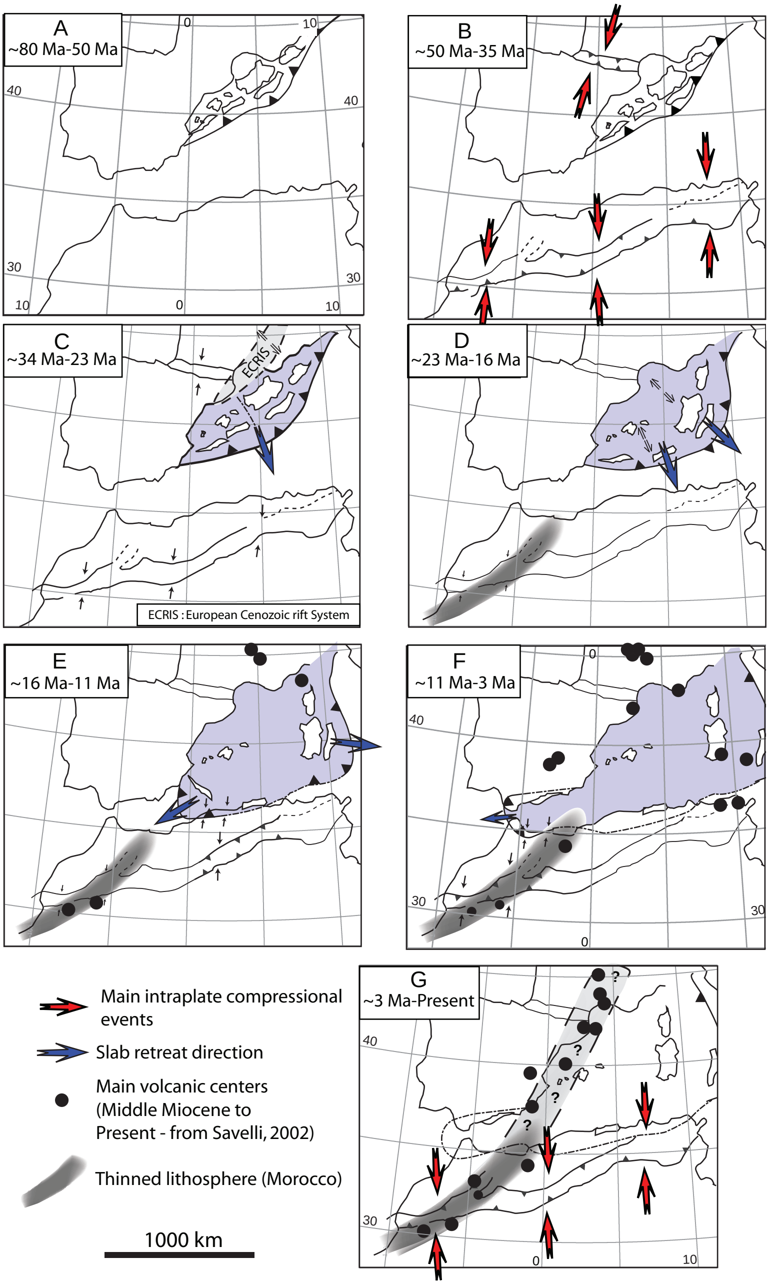

Fig. 4.47 Chronology of the Atlas tectonic/magmatic and lithospheric events in the frame of the Western Mediterranean.geodynamics (modified from Missenard, 2006). The most important shortening phases in the Atlas system (red arrows) occurred before and after the time when the Appenine-Maghrebide subduction and slab roll-back processes operated (blue arrows). Notice that the Late Cretaceous-Eocene tectonic setting is highly simplified in this figure (see Chap. 5 for details)

in Western Mediterranean (Fig. 4.47C-F). The latter, Oligocene-Miocene period is precisely the one during which displacements are rather accommodated in the coastal mountain belts such as the Betic-Rif Arc (Chap. 5). This period of relative tectonic quiescence in the Atlas is also the one during which the lithospheric thinning took place.

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Chapter 5
The Rif Belt

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This chapter is dedicated to the memory of our friend Pr. Ahmed Ben Yaïch, whose promising scientific activity has been interrupted too early, and of Gabriel Suter, the father of the Rifian geological mapping, just disappeared after a long Moroccan career.

5.1 General

References: Considering only the most recent and general references, we may cite Frizon de Lamotte et al. (1991), Martínez-Martínez & Azañón (1997), Chalouan et al. (2001), Platt et al. (2003a), Chalouan & Michard (2004), Frizon de Lamotte

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5.1.1 The Rif Belt, a Segment of the Mediterranean Alpine Belts

The Rif Belt belongs to a much larger orogen, i.e. the Betic-Rif-Tell orogen, which itself is part of the still larger Mediterranean Alpine belts (Fig. 5.1; see also Chap. 1.1, Fig. 1.2). The Rif Belt occupies a key position in this orogenic system. On the one hand, it forms the westernmost part of the *Maghrebide belt*, which extends along the North African coast, and continues eastward to Sicily and Calabria in southern Italy. On the other hand, it forms the southern limb of the Gibraltar Arc, the northern limb of which corresponds to the Betic Cordilleras (Fig. 5.2).

The Gibraltar Arc is one of the tighter orogenic arcs (oroclines) worldwide. It also corresponds to the western tip of the Alpine belts. The Gibraltar Arc closes almost completely the Mediterranean to the west (Fig. 5.3), and exemplifies the

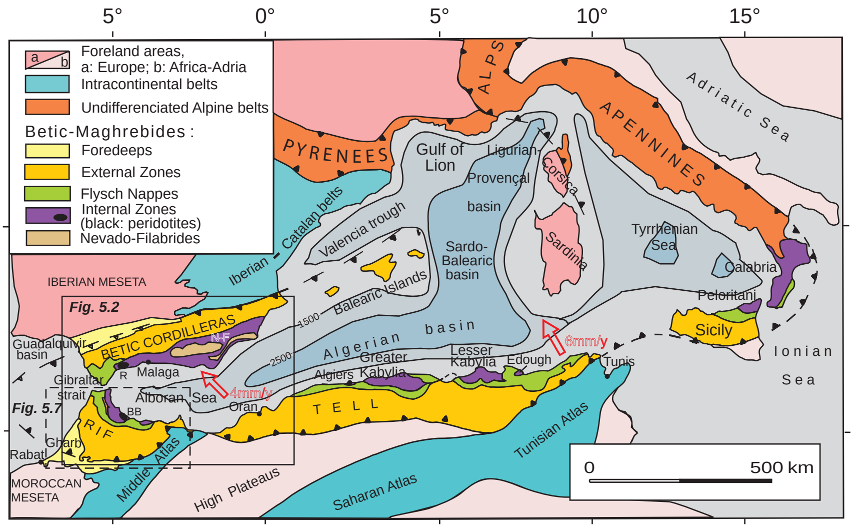


Fig. 5.1 The Rif Belt in the frame of the West Mediterranean Alpine belts. *Empty arrows*: direction and rate of recent Africa-Europe convergence (NUVEL 1A model, DeMets et al., 1994; Morel & Meghraoui, 1996). BB: Beni Bousera; N-F: Nevado-Filabrides; R: Ronda

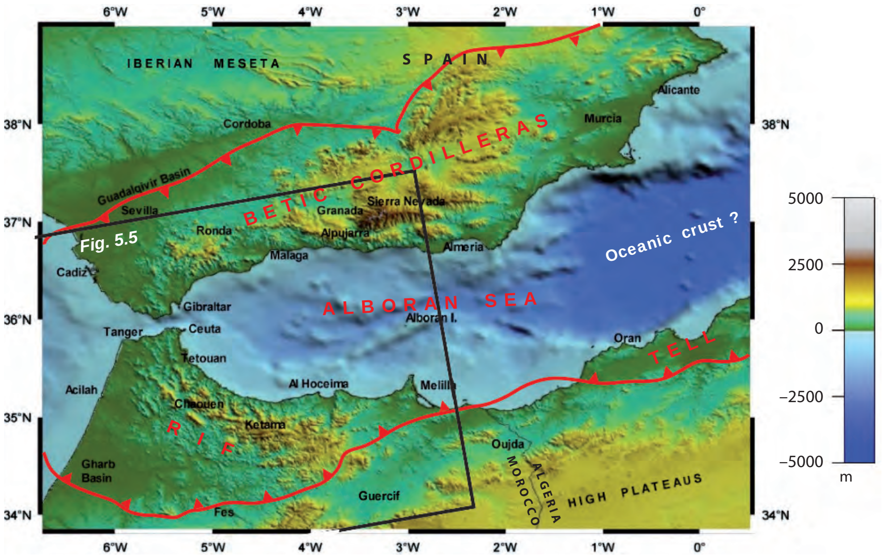


Fig. 5.2 Elevation map of the Gibraltar Arc and Alboran Sea area from ETOPO2 Global Data Base (Courtesy of F. Negro). For location, see Fig. 5.1. *Red line* with teeth: External front of the Betic and Maghrebide orogens

intimate association of two opposite processes, i.e. mountain building and subsequent collapse. The latter process formed the Alboran Sea in the core of the system while thrusts and folds propagated towards the external zones. The Gibraltar Arc has a counterpart at the eastern end of the Maghrebides, i.e. the Calabrese Arc. Both arcs originate from the same geodynamic process, which resulted in the closure of the Ligurian-Maghrebian Tethys and opening of the West Mediterranean

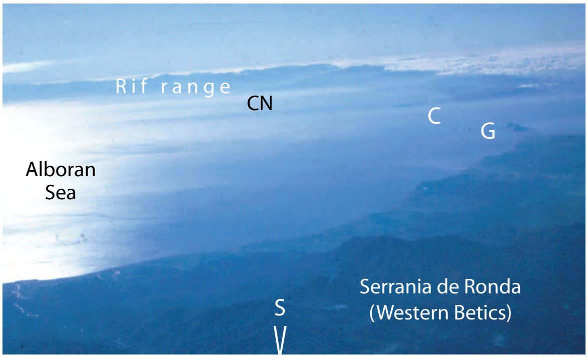


Fig. 5.3 Aerial view looking southeast-ward on the Alboran Sea surrounded by the Gibraltar Arc. The plane is crossing the Serrania de Ronda. G: Gibraltar; C: Ceuta; CN: Cabo Negro. The Strait of Gibraltar passes between G and C. The clouds from the Atlantic Ocean are stopped by the Dorsale range of the Haouz, south of Ceuta

Sea. Paradoxically, this evolution started from the Late Eocene and lasted for about 30 million years while the Africa-Europe convergence was going on (see Chap. 1, Fig. 1.12). This is explained by the roll-back of the Tethyan subduction beneath the European lithosphere (Sect. 5.7).

Due to its position between the Atlantic Ocean and the Mediterranean Sea, the Gibraltar Arc played a major role in the evolution of the famous Messinian salinity crisis (5.9–5.3 Ma), which deeply marked the morphology of the Mediterranean areas and the coeval sedimentation. Folding in the External Zones related to plate convergence during the Messinian (~6 Ma) contributed to the closure of the *South-Rif Corridor*, i.e. the last gate for the Atlantic waters entering in the Mediterranean. Closure of the seaway was ensured by the regional uplift related to the development of the Trans-Moroccan Hot Line (Chap. 4), also responsible for the Trans-Alboran magmatism. By the early Pliocene, the opening of the Strait of Gibraltar allowed the cool, low-salinity Atlantic waters to enter again the Mediterranean Basin (Fig. 5.4). This event can be ascribed either to normal fault activity within the Strait, or to a general downthrow of the whole Arc due to westward roll-back of the underlying subduction (see Sect. 5.7), or to both causes. Moreover, during the late Messinian when the Mediterranean Sea level was at its lower stand, the Gibraltar threshold has been incised by a deep canyon where Atlantic waters finally rushed.

5.1.2 Structural Domains

Three main structural domains form the Gibraltar Arc, from inside to outside and bottom to top, (i) the *Internal Zones*, or *Alboran Domain*; (ii) the *Maghrebian Flyschs*, and (iii) the *External Zones* (Fig. 5.5). Each domain consists of tectonic complexes of stacked units or nappes with similar lithologies within a given complex, but contrasting from one complex to the other. This results in different surface morphologies and spectral signatures (Fig. 5.6). The structural map (Fig. 5.7) and associated cross-section (Fig. 5.8) highlight the principal structural lines of the Rif Belt based on extensive, integrated studies and mapping at scale 1:50000.

5.1.2.1 Internal Zones (Alboran Domain)

The Internal Zones consist of continental units displaced westward over several hundreds of kilometres, thus representing a genuine exotic terrane. Considering the grade of Alpine metamorphic recrystallization in these units, we can recognize two complexes, which respectively form the upper and lower plates of a metamorphic core complex. In the Rif and western Betics, the lower plate corresponds to the *Sebtide* and *Alpujarride* units, respectively, both dominantly consisting of relatively deep crustal rocks such as mica-schists, migmatites and granulites associated with mantle peridotites (Beni Bousera, Ronda). Another deep complex occurs in Central and Eastern Betics beneath the Alpujarrides, namely the *Nevado-Filabride* complex,

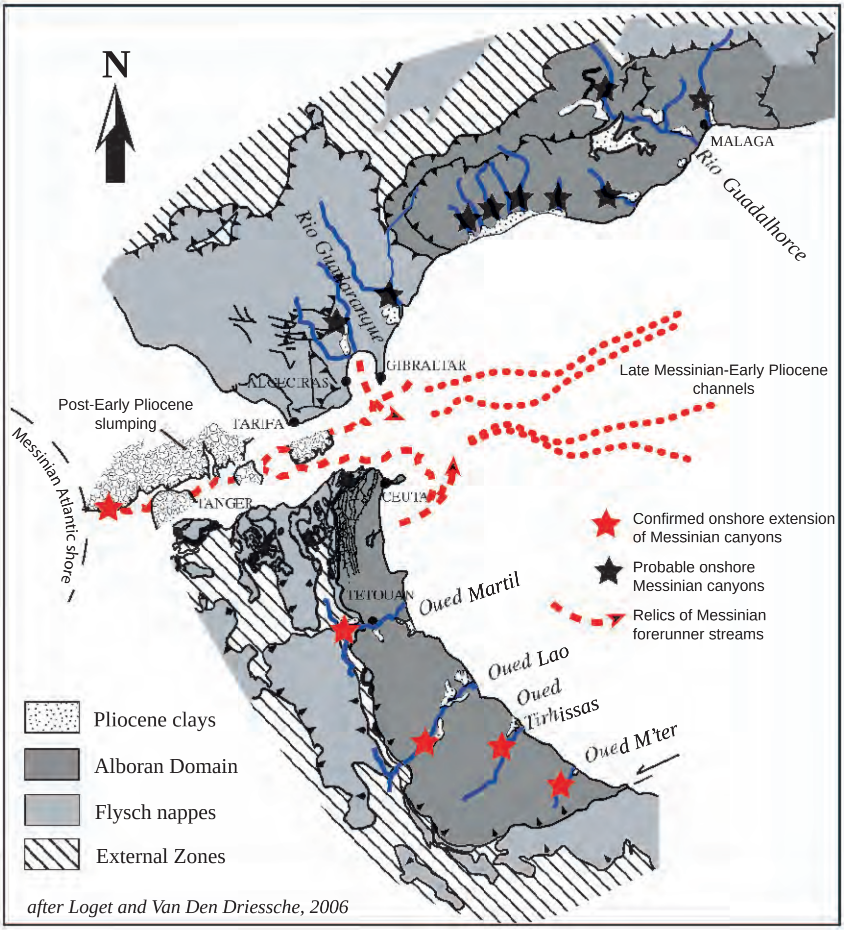


Fig. 5.4 Distribution of the Pliocene deposits in the Messinian canyons of the internal part of the Gibraltar Arc, after Loget & Van Den Driessche (2006), modified. The canyon of the Strait of Gibraltar, now filled with breccias, was used by the Atlantic waters 5.3 Ma ago to rush into the Mediterranean Basin

which displays meta-ophiolites at its top. The upper plate consists of the *Ghomaride* (Rif) and *Malaguide* (Betics) complexes, which overlie the Sebtide-Alpujarride through a regional detachment. They include Paleozoic rocks affected by a Variscan metamorphism partly superimposed by weak Alpine recrystallization, and relicts of their Mesozoic-Cenozoic cover.

Sebtide-Alpujarride rocks were cored in the Alboran Basin (Fig. 5.5), which fully justifies the name of "Alboran Domain" given to the Internal Zones. A similar metamorphic structure can be recognized in the Algerian Kabylides, Peloritan Mountains of Sicily, and Calabria. Thus, these varied exotic terranes were admittedly parts of the same continental domain *AlKaPeCa* (Bouillin, 1984), including the "Alboran

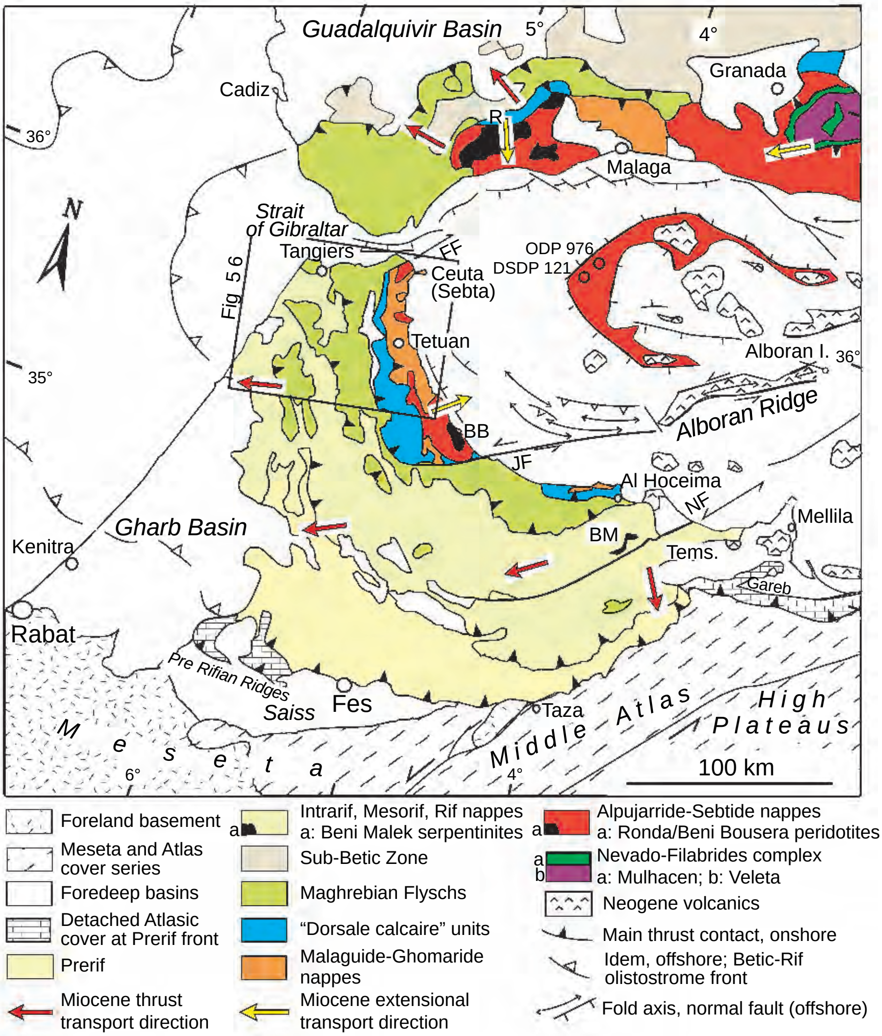


Fig. 5.5 Structural map of the Gibraltar Arc, modified from Chalouan & Michard (2004). Alboran Basin after Comas et al. (1999); Gulf of Cadiz after Maldonado et al. (1999); Miocene kinematics after Frizon de Lamotte et al. (1991) and Martínez-Martínez & Azañón (1997). BB: Beni Bousera; BM: Beni Malek; FF: Fahies Fault; JF: Jebha Fault; NF: Nekor Fault; R: Ronda; Tems.: Temsamane

microplate” of Andrieux et al. (1971), before being deformed and dispersed. The initial location of AlKaPeCa is controversial as this domain may correspond either to a distal part of the Iberian margin or to an isolated microcontinent within western Tethys (Sect. 5.7). Note that the name of “Mesomediterranean Block”, sometimes used instead of AlKaPeCa, is misleading since the Mediterranean Sea developed after the AlKaPeCa split.

The Alboran Domain and its eastern equivalents include also a complex of Mesozoic-Cenozoic thrust sheets dominated by Triassic-Liassic carbonates. This is the *Dorsale Calcaire*, which appears as discontinuous ranges at the front of, and generally below the more internal crustal units. The Dorsale Calcaire units represent remnants of the former southern passive margin of AlKaPeCa. These units can be traced northward around the Gibraltar Arc up to the Granada meridian.

5.1.2.2 Maghrebian Flyschs

The Maghrebian Flyschs nappe complex originates from the Ligurian-Maghrebian Ocean, which connected the Central Atlantic and Alpine Oceans from Jurassic to Paleogene (see Chap. 1, Figs. 1.5, 1.8). The final suturing of the Maghrebian Ocean between AlKaPeCa and Africa by the Late Oligocene-Early Miocene resulted in the formation of a nappe stack consisting of the dominantly turbiditic sediments ("flyschs") accumulated within the oceanic basin. These nappes root beneath the Internal Zones and overlie the External Zones (Figs. 5.7, 5.8). Part of the Flysch nappes has been back-thrust over the northern Ghomarides (J. Zemzem and Riffiene massifs south of Ceuta).

The Flyschs nappes are exposed widely in western Betics (Campo de Gibraltar; Fig. 5.5), up to the Granada transect. According to Algerian nomenclature

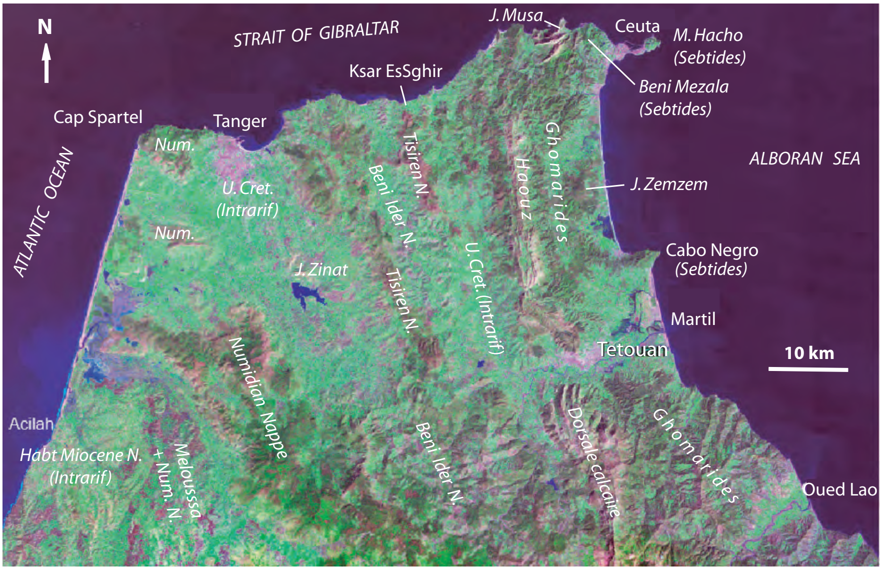


Fig. 5.6 Landsat image of northern Rif. Location: see Figs. 5.5 and 5.7. The different structural zones shown here are (from E to W) the Internal Zones (Sebtides, Ghomarides, Dorsale calcaire), the Flysch Nappes (Tisiren, Beni Ider, Meloussa, Numidian), and the underlying External Zones (restricted to their internal part, i.e. Intrarif from Tanger and Habt). Num. N: Numidian Nappe. Basically, the image shows a collapsed accretionnary prism in front of a partly collapsed and immerged buttress

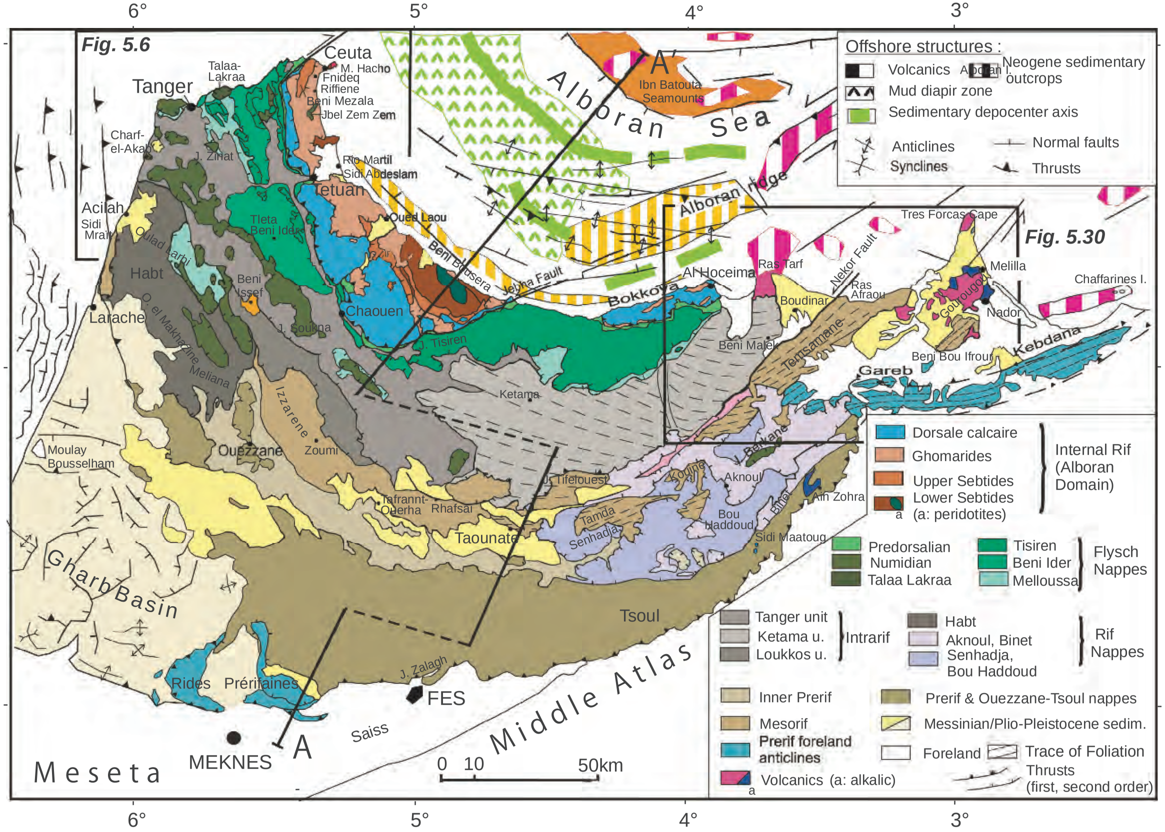


Fig. 5.7 Structural map of the Rif Belt, after Suter (1980a, 1980b) and Chalouan et al. (2001), modified. Alboran Basin after Chalouan et al. (1997) and Comas et al. (1999). Gharb and Atlantic margin after Flinch (1996). AA': trace of cross-section shown in Fig. 5.8

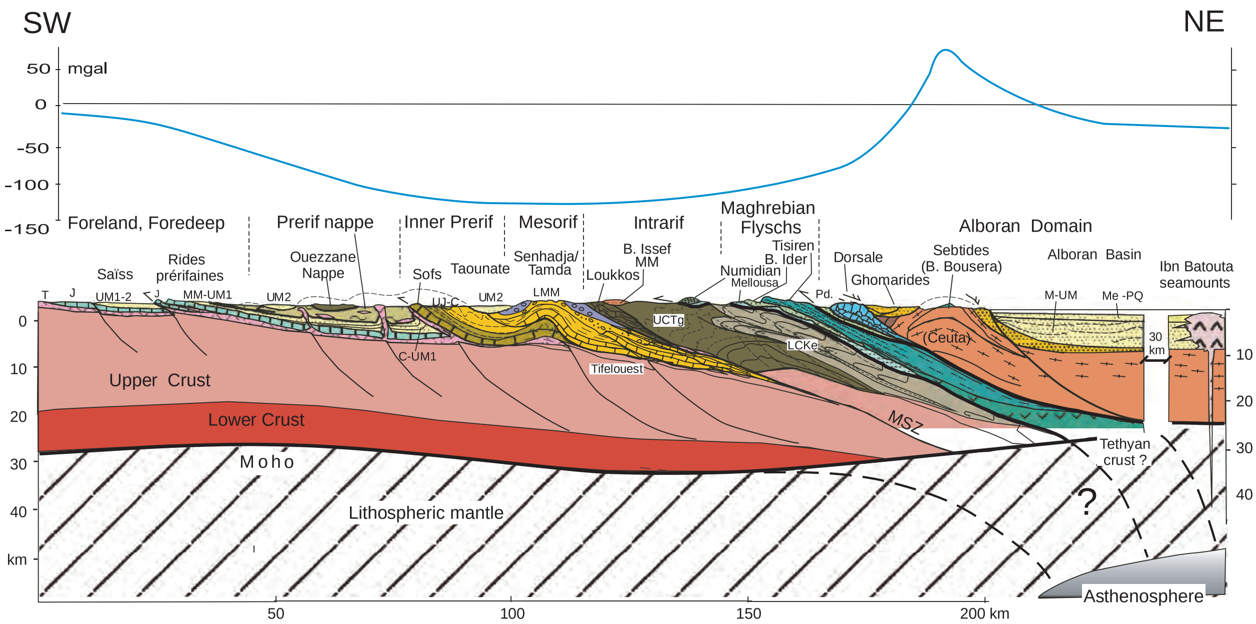


Fig. 5.8 Crustal cross-section of the Rif Belt, modified from Chalouan et al. (2001), and corresponding Bouguer anomaly after Favre (1995). Moho and base of lithosphere depth after Favre (1995), Torné et al. (2000), Frizon de Lamotte et al. (2004), Fullea Urchulutegui et al. (2006). Location in Fig. 5.7. Abbreviations: B.: Beni; C: Cretaceous; J: Lower-Middle Jurassic; LCKe: Lower Cretaceous of Ketama; LMM: Lower-Middle Miocene; MM: Middle Miocene; MSZ: Mesorif Suture Zone; Pd: Predorsalian; T: Triassic; Tg: Tanger Unit; UM: Upper Miocene (1: Tortonian “pre-nappe”; 2: Upper Tortonian-Pliocene “post-nappe”); UJ-C: Upper Jurassic-Cretaceous. Conventional lithologic signatures (schematic)

(cf. Sect. 5.3), the more internal and higher units are referred to as the *Mauretanian nappes* (in the Rif, J. Tisiren and Beni Ider nappes), whereas the more external and lower units are named the *Massylian nappes* (Chouamat-Melloussa and Numidian nappes). All these nappes of oceanic origin mark a suture zone beneath the Alboran terrane (Fig. 5.8), although ophiolitic remnants are lacking along the suture, except in Algeria and, more significantly, in Calabria.

5.1.2.3 External Zones

The External Zones of each limb of the Gibraltar Arc originate from two distinct paleomargins of Africa and Iberia, respectively. Therefore, contrary to the Internal Zones and Flyschs nappes, the Rif and Betic External Zones do not display any stratigraphic/structural continuity across the Strait of Gibraltar, except in the uppermost and youngest units of the accretionary prism located offshore in the Gulf of Cadiz. The *Subbetic Zone* proceeds from a starved paleomargin and displays a thin-skinned tectonics, whereas the more complex Rif-Tell Externides proceed from an abundantly sedimented margin and display both thick-skinned and thin-skinned structural styles.

In the External Rif (Figs. 5.7, 5.8), we may distinguish three structural zones, from north to south and top to bottom, the *Intrarif*, *Mesorif* and *Prerif*, which derive from more and more proximal parts of the African paleomargin, respectively. Within each of these zones, we may again distinguish deep rooted, parautochthonous units (more or less detached from their original basement), and diverticulated, superficial gravity-driven nappes. Metamorphic recrystallizations reaching the chloritoid-bearing greenschist-facies conditions occur in the deep Intrarif (Ketama) and eastern Mesorif (North Temsamane) units, on both sides of the Nekor fault. Serpentinite and metabasite slivers (Beni Malek) crop out along the latter fault, which likely corresponds to a segment of an intracontinental (intra-margin) minor suture recognisable eastward up to Algeria (Sect. 5.4).

Two major, ENE to NE-trending left-lateral faults, namely the Jebha Fault, south of the Northern Rif Internal Zones, and the Nekor Fault in the Eastern Rif, give evidence, at the map scale, of the obliquity of the movement of the Alboran Domain relative to Africa. The tectonic structures observed in the External units and overlying Flysch outliers all show an externalward displacement, i.e. toward the Neogene *Prerif foredeep*. The foredeep is widely developed and poorly deformed in the west, i.e. in the Gharb (Rharb) Basin, equivalent to the Guadalqivir Basin (Betic foredeep). In contrast, the foredeep changes to a deformed, narrow corridor south of Eastern Rif, and finally vanishes east of Taza, where the Mesorif outliers directly overlie the Middle Atlas foreland.

5.1.2.4 Alboran Basin and Trans-Alboran Magmatism

The *Alboran Basin* opened at the rear/east of the belt during the latest Oligocene-Early Miocene. This is a synorogenic basin with a thinned continental crust, which incorporates stretched elements of the orogen. Up to 8 km of turbidites and muds accumulated in the western sub-basin where numerous mud diapirs occur. In the central and eastern parts of the basin, as well as on its southern and northern borders, an important calc-alkaline magmatism developed during the Middle-Late Miocene. This is the *Trans-Alboran magmatic province*, i.e. the western part of the Maghrebian magmatic province, which includes numerous granitic massifs in Algeria (Sect. 5.6).

The Alboran sediments are affected by open folds and strike-slip faults (Alboran Ridge) dated from the Late Miocene (Messinian) and resulting from the ongoing Africa-Europe convergence.

5.1.3 Lithospheric Structure

References: The crustal structure of the African margin was described by Favre (1995). Overall descriptions of the Gibraltar Arc lithosphere, based mainly on seismological data and geophysical modeling, have been published by Seber et al. (1996), Calvert et al. (2000), Gurría & Mezcua (2000), Torné et al. (2000), Gutscher et al. (2002), Spakman & Wortel (2004), Frizon de Lamotte et al. (2004), Fullea Urchulutegui et al. (2006), Bokelman & Maufroy (2007). The thermal structure is described by Polyak et al. (1996) and Rimi et al. (1998). Concerning the Eastern Alboran Basin and next Algerian-Balearic Basin, see Mauffret et al. (2004), Domzig et al. (2006), and Schettino & Turco (2006).

The continental crust north and especially south of the Alboran Basin is rather poorly known due to lack of deep seismic survey. However, the varied geophysical methods already used indicate great thickness variations in the crust and heterogeneities in the mantle.

In the Rif foreland, seismic refraction studies image a 30 km thick continental crust, with a 9–10 km thick lower crust (Chap. 1, Fig. 1.20; cf. Fig. 5.8 above). Modelling of crustal and lithospheric thickness or density variations that integrates both elevation and geoid anomalies yields evidence (assuming local isostasy) of a poorly marked orogenic root beneath Central Rif (Moho at 34 km depth), followed by a quick Moho rise toward the Alboran Basin. In the western sub-basin where the water depth does not exceed 1 km (Figs. 5.2, 5.8), the Moho is located at about 18–20 km, associated with a thinned, 15–18 km continental crust. East of the Alboran Ridge, water depth increases up to 2.5 km, and the crust thickness decreases to ~12 km: the eastern Alboran sub-basin is transitional toward the Algerian-Balearic Basin the crust of which is likely oceanic.

The mantle lithosphere thickness is less constrained. Taking into account the high thermal gradients observed in the basin axis, Torné et al. (2000) inferred a thickness of about 20–30 km. In contrast, seismic studies led Calvert et al. (2000) to suggest that asthenosphere comes in contact with the Alboran thinned continental crust, and even penetrates as a wedge between the Gibraltar Arc crust and underlying lithospheric mantle (delamination). The latter hypothesis is neither confirmed nor contradicted by the more recent geophysical studies (Sect. 5.7). Fullea Urchulutegui et al. (2006) modelling leads to rather greater thicknesses beneath the western sub-basin (Fig. 1.18). In the cross-section (Fig. 5.8), we schematically delineate the depth to the asthenosphere at shallow level beneath the Alboran Sea and assume that it crosscuts the subducting African lithosphere, consistent with the slab break-off hypothesis. The latter hypothesis is supported by the geochemistry of the Trans-Alboran magmatism (Sect. 5.6) and by the 3D seismic tomography imaging (Sect. 5.7). Waveforms of body waves that traverse the Alboran Sea confirm the presence of an anomalous mantle underlying the basin (Bokelman & Maufroy, 2007).

North of the Alboran Sea, the continental crust again thickens beneath the Betic Cordilleras, following a roughly symmetrical pattern as described above in the Rif. Below the Strait of Gibraltar, the crust is up to 30–32 km thick (Fig. 1.18). This is consistent with the current concept of an orogenic arc above an active east-dipping subduction zone (Sect. 5.7).

5.2 Internal Zones (Alboran Domain)

The following sections describe the stratigraphy and structure of the Rif Internal Zones, i.e. of the Moroccan part of the Alboran Domain. However, we will also refer frequently to Betic data, as they complement usefully the geological data on this exotic terrane.

5.2.1 Sebtides

References: Most of the recent literature concerning the Sebtide complex deals with its metamorphic structure and petrology: see Bouybaouene (1993), Saddiqi et al. (1995), Bouybaouene et al. (1998), Montel et al. (2000), El Maz & Guiraud (2001), Haissen et al. (2004). Negro et al. (2006) consider both the Sebtides and their Betic counterparts (Alpujarrides). The Beni Bousera peridotites were repeatedly studied: see Reuber et al. (1982), Saddiqi et al. (1988), Pearson et al. (1989), Kornprobst et al. (1990), Kumar et al. (1996), Tabit et al. (1997), Pearson et al. (2004), and Downes (2007). Their Betic equivalents (Ronda), even more frequently investigated, are recently considered by Sánchez-Gómez et al. (2002), Platt et al. (2003b), Tubía et al. (2004), Cuevas et al. (2006), and Downes (2007), with references therein. The isotopic ages reported below are from Platt & Whitehouse (1999), Argles et al. (1999), Montel et al. (2000) and Sánchez-Rodriguez & Gebauer (2000). Kabylian data are summarized by Michard et al. (2006), with references therein.

The Sebtide complex crops out in four tectonic windows beneath the Ghomaride nappes (Figs. 5.7, 5.9), i.e. from N to S, the Beni Mezala anticline, Ceuta-Monte Hacho massif, Cabo Negro promontory, and eventually the much larger Beni

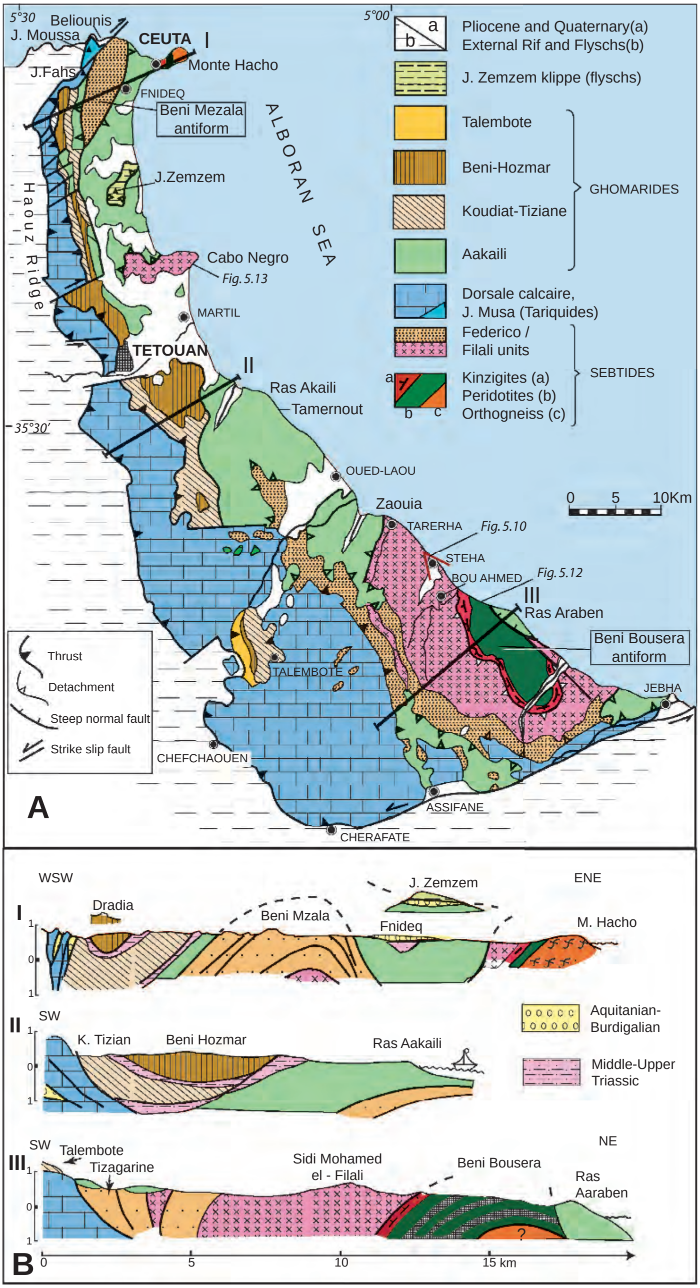


Fig. 5.9 Structural map (**A**) and cross-sections (**B**) in the Northern Rif Internal Zones, after Chalouan & Michard (1990), modified. The Ghomaride-Sebtide nappe stack was folded before and after the unconformable onlap of the “post-nappe” formations (e.g. at Fnideq in Sect. B-I). The J. Zemzem and Haouz backthrusts occurred during the second shortening event

Bousera antiform. Each of these windows, except Cabo Negro, consists of a stack of nappes with different pre-Alpine protoliths and contrasted metamorphic grades (Sect. 5.4.2).

5.2.1.1 Basement Units

The oldest rocks, likely Paleozoic in age or even Precambrian, crop out in the Beni Bousera antiform (Fig. 5.10). This large antiformal structure deforms a stack of two thick basement units, from bottom to top, the *Beni Bousera* and *Filali* units, and their more or less detached metamorphic cover units, referred to as the *Federico* imbrications (Fig. 5.11A).

The Beni Bousera unit consists of a huge peridotite body, at least 2 km thick, topped by discontinuous slivers of granulites (kinzigites). Together with their Alpujarride counterpart (Ronda peridotites of the Los Reales nappe), the Beni Bousera peridotites are among the largest infracontinental mantle massif worldwide (Figs. 5.10, 5.12). The dominant lithology is a spinel lherzolite including pyroxenite layers which yielded graphitized diamond pseudomorphs. This indicates an early equilibration at more than 140 km depth followed by a retromorphic evolution at about 50 km depth. The metagabbroic pyroxenite and garnet pyroxenite layers interbedded within the ultrabasites are deformed by isoclinal folds associated with a high temperature foliation. Only some of these pyroxenites display geochemical evidence of being related to recycled oceanic crust, whereas the others are best explained by crustal accumulation in mantle-derived magmas (Downes, 2007). Intensely foliated harzburgites occur close to the top of the ultrabasite massif, some of them including garnet crystals whose origin is controversial. Serpentinization increases toward the sheared envelope of the peridotite body.

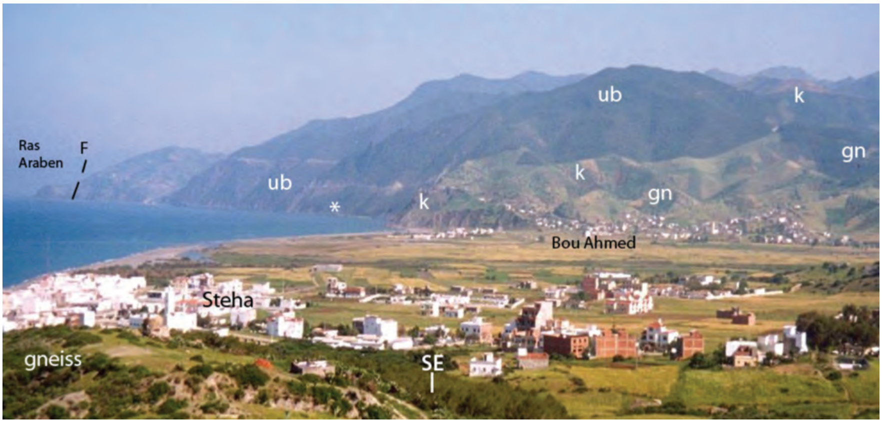


Fig. 5.10 View of the Beni Bousera massif looking SE-ward. See Fig. 5.9A for location. ub: ultrabasites (*: location of Fig. 5.12); k: kinzigites (top of Beni Bousera unit); gn: gneisses (base of Filali unit). The Ras (Cape) Araben is made of a downthrown element of the Aakaili nappe (Ghomaride). Photo by O. Saddiqi

The crustal rocks immediately above the peridotites consist mainly of acidic granulites (kinzigites) with garnet-sillimanite±kyanite-graphite assemblages (Sect. 5.5). Some basic granulites are interbedded within the acidic ones, such as the Ichendirene HP-granulites, which exhibit mineral assemblage including pyrope-rich garnet, jadeite-rich clinopyroxene and rutile, indicative of an equilibration at $P > 16\,\text{kbar}$, 760–820 °C. The latter P-T conditions are closely similar to those of the garnet-bearing peridotite equilibration.

It is worth noting that the Beni Bousera and Ronda peridotites have become disconnected with the mantle as they were thrust over a deeper Sebtide-Alpujarride crustal unit, which corresponds to the migmatitic orthogneiss of Monte Hacho

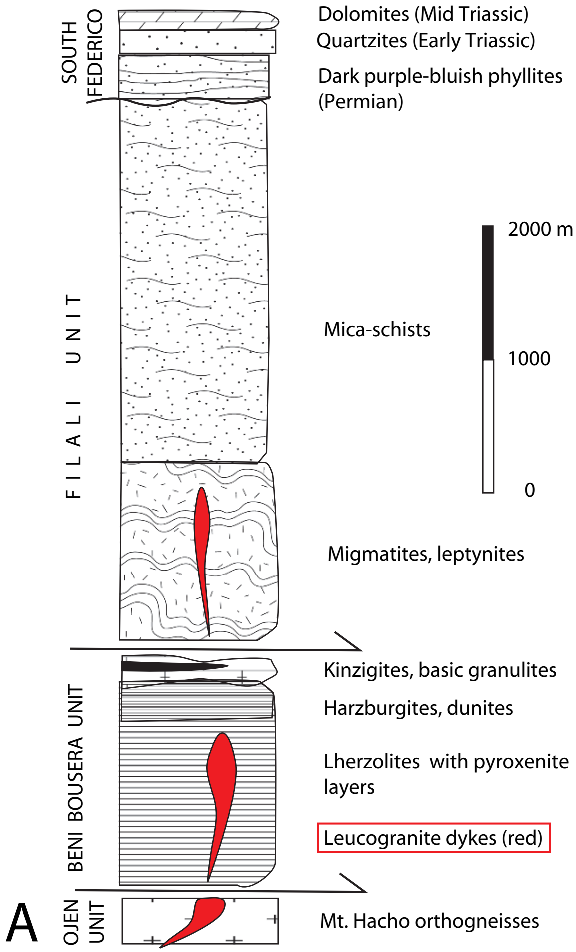


Fig. 5.11 Synthetic litho-stratigraphy of the pre-Triassic series of the Alboran Domain in the Rif Belt. (**A**): Sebtides, after Bouybaouene (1993), Saddiqi (1995), Negro (2005); (**B**–**D**): Ghomaride nappes, after Chalouan & Michard (1990), modified

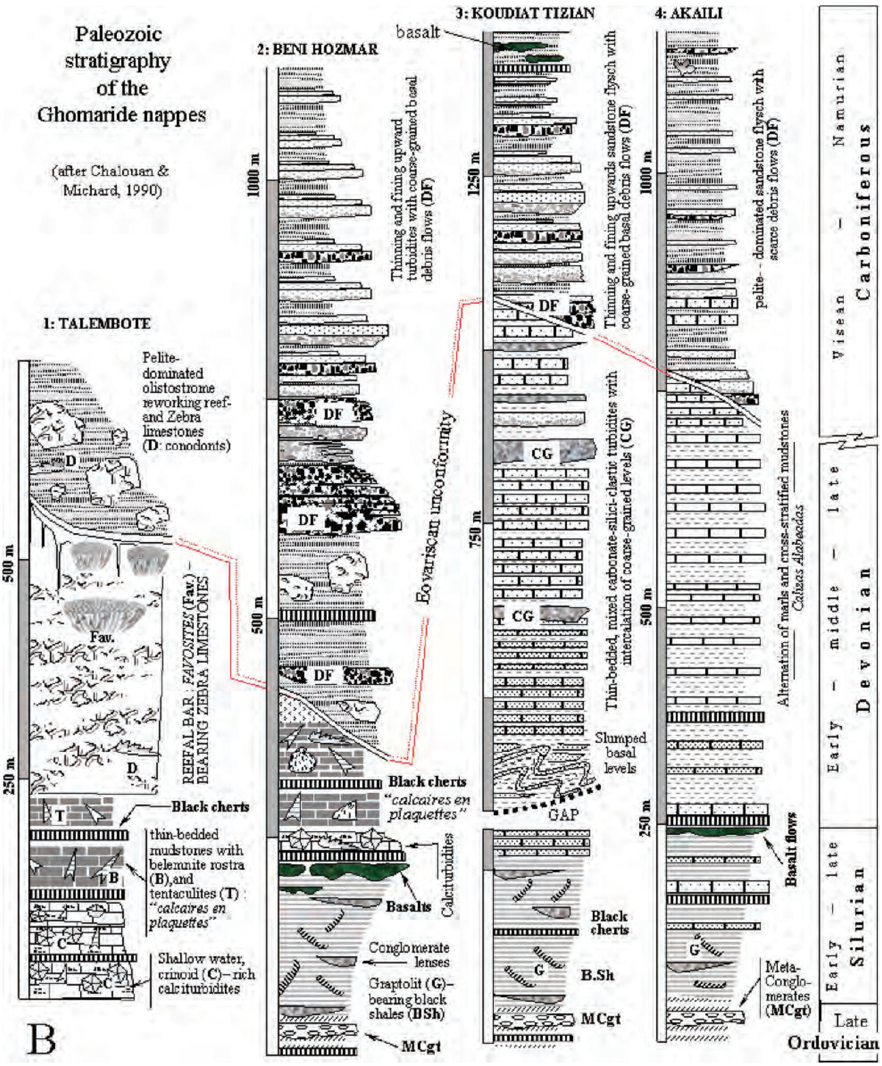


Fig. 5.11 (continued)

(Ceuta) and Ojen (Betics), associated with thick marbles in Andalucia (Sierra Blanca). Partial melting within this deep crustal material resulted in the intrusion of cordierite-andalusite-bearing granite dykes across both the northern and southern peridotite massifs.

The age and emplacement mechanisms of the Ronda and Beni Bousera peridotites have been repeatedly questioned. Some authors assume that they originate from a Miocene hot asthenospheric diapir coeval with the opening of the Alboran Sea, whereas others argue that they formed during the Paleozoic, being uplifted first during the opening of the Tethys Ocean, then during the Alpine orogeny

Fig. 5.12 Lherzolite outcrops along the Beni Bousera shoreline east of Bou Ahmed (see Figs. 5.9 and 5.10 for location). *Arrow*: pyroxenite layers. The steep, NE-dipping fractures are likely related to the Ras Araben normal fault (*background left*)

(Sect. 5.7.4.4). The latter hypothesis is supported by a great variety of isotopic datings such as: the ages close to 300 Ma obtained on relict minerals from the kinzigites (U-Pb on zircon cores, and garnet-armoured monazite grains); the garnet-whole rock Sm-Nd isochron age of 235.1 ± 1.7 Ma obtained from Ronda pyroxenites; the 286 ± 5 Ma U-Pb age from oscillatory-zoned domains from the euhedral zircon fraction in Ronda garnet pyroxenites; and the U-Pb zircon core ages from the same rocks, 178 ± 6 Ma, 143 ± 16 Ma, 131 ± 3 Ma recording successive steps of cooling during the Triassic-Early Cretaceous Tethyan rifting.

The Filali unit overlies the Beni Bousera unit through a subtractive ductile shear zone, as the migmatites at the bottom of the Filali (Fig. 5.13) are equilibrated under 8 kbar, 780 °C, i.e. at much lower pressure than the underlying granulites. Above the gneisses, the Filali unit is made of mica-schists the mineral assemblages of which change more progressively upward, from garnet-biotite-sillimanite to garnet-biotite-staurolite-kyanite, and finally chlorite-chloritoid-muscovite±biotite±kyanite. Andalusite is ubiquitous in these metapelites, and the coexistence of the three Al silicates suggests a polycyclic metamorphic history with an Alpine evolution (Sect. 5.5) superimposed on a Variscan one. Indeed, Variscan to Jurassic ages were obtained from some high-grade Alpujarride rocks, such as the Torrox gneiss (equivalent to the Monte Hacho gneiss), dated at 285 ± 5 Ma (U-Pb zircon), and the eclogite included in the Ojen gneisses (183 ± 3 Ma, U-Pb on magmatic zircon core).

5.2.1.2 Metamorphic Cover Units

The *Upper Sebtides* or *Federico units* consist of metasediments, affected only by Alpine recrystallizations. They form thrust imbrications where the same lithostratigraphic sequence is repeated several times with downward increasing metamorphic grade. In the Beni Mezala antiform (Figs. 5.9, 5.14), four superimposed, relatively thin units (500–1000 m) are folded together forming an antiformal stack beneath the Ghomarides units. Each unit displays the three main formations ascribed from bottom to top, to Permian-Triassic, Lower Triassic and Middle-Upper Triassic Alpine-type sequences (Fig. 5.15A1). The Permian-Triassic deposits consist of red pelites in the uppermost unit (Tizgarine), purple phyllites in the intermediate unit (Boquete

Fig. 5.13 Migmatitic gneiss from the Filali unit, with a pervasive boudinage fabric affecting a dyke of garnet-bearing meta-aplite (below the hammer). Note the complex, polyphase structure of these rocks (superimposed Variscan and Alpine deformations). Cabo Negro south cliffs (see Fig. 5.9A for location). Photo by O. Saddiqi

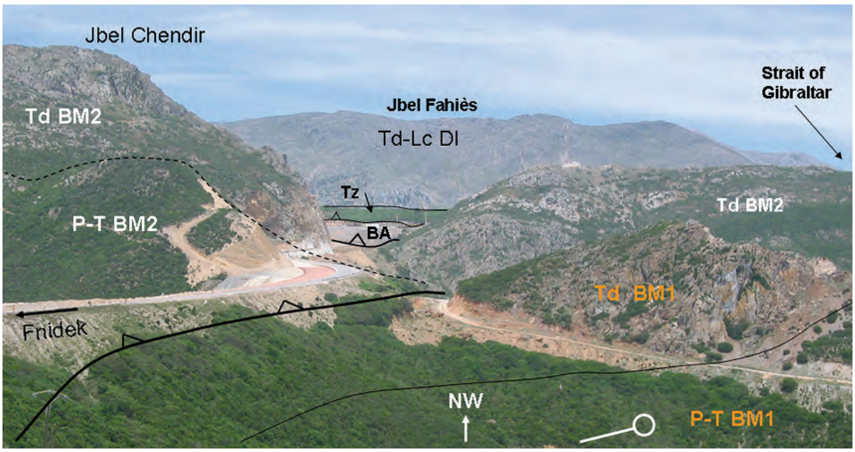


Fig. 5.14 The Beni Mezala (BM) massif along the Fnideq-Ksar Es-Sghir road, 3 km NW from Fnideq (see Fig. 5.9 for location). Note the duplication of the stratigraphic levels of the lowest BM1 and BM2 Federico units in the foreground: P-T, Permian-Triassic metapelites and quartzites; Td: Middle Triassic dolomites. Background: Upper Federico units (BA: Boquete Anjera; Tz: Tizgarine), dominated by the Jebel Fahs Dorsale unit (DI), which includes Upper Triassic dolomites and Hettangian limestones (Td-Lc). The post-nappe fold axis plunges SSW

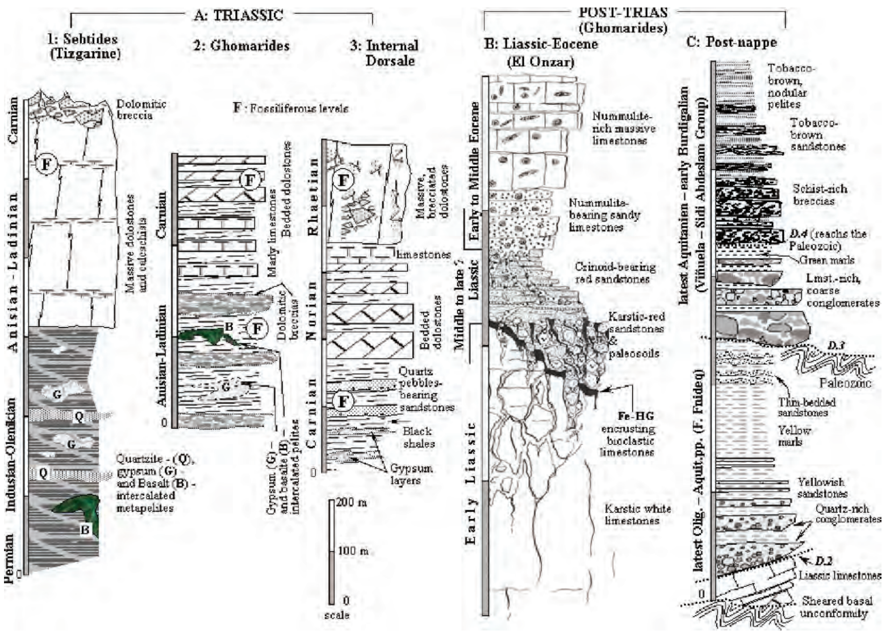


Fig. 5.15 Detail stratigraphy of the cover sequences from representative units of the Alboran Domain in the Rif Belt. (**A**): Triassic series, after Chalouan (1996). (**B** and **C**): Liassic-Eocene and "post-nappe" séries of the Ghomarides, after El Kadiri et al. (1992) and Serrano et al. (2006), respectively

Anjera), and of dark ("color de humo", smokg) quartz-phyllites in the lowest units (Beni Mezala 2 and 1). This color evolution can be correlated with change in metamorphic grade (Sect. 5.5). Light-colored quartzites are placed in the Lower Triassic, and compare with the Briançonnais quartzites of Western Alps. They are overlain by Middle-Upper Triassic dolomitic marbles, which locally yielded *Gyroporella* (Dasycladaceae) of Middle Triassic age. Remarkably, no younger stratigraphic level was ever recognized in the Sebtides or the Alpujarrides, which is possibly related to a tectonic detachement of the upper units (Sect. 5.7.4.3). By contrast, the stratigraphic sequence of the upper Federico unit is completed downward by probable Upper Carboniferous greywacke formations. The base of the Beni Mezala unit is made of the Benzu mica-schists, closely similar to the Filali ones.

On top of the Beni Bousera antiform, similar metamorphic imbrications occur. However, in this southern region the lowest Federico unit, here labelled the *Souk-el-Had unit*, seemingly remained in stratigraphic contact over the Filali mica-schists, as similar metamorphic assemblages are found on both sides of their common limit (Sect. 5.5). Note that similar Permian and Triassic imbrications also occur on top of the Los Reales nappe of western Betics (Casares units, Benarraba imbrications).

5.2.2 Ghomarides

References: The Paleozoic stratigraphy of the Ghomaride–Malaguide Complex has been described by Herbig & Mamet (1985) and Herbig (1989) in Spain, and by Chalouan & Michard (1990) in Morocco. The pre-Alpine relationships of this complex, and of the Alboran microplate in general with the other Hercynian segments of Western Mediterranean have been discussed by von Raumer et al. (2003), Trombetta et al. (2004), Helbing et al. (2006), Sanz de Galdeano et al. (2006), Micheletti et al. (2006).

The Triassic unconformable deposits have been described by Baudelot et al. (1984), Chalouan (1996), and Diez (2000). On the other hand, Feinberg et al. (1990), El Kadiri et al. (1992), Durand-Delga et al. (1993), Maaté (1996), Martín-Martín et al. (1997, 2006), Martín-Algarra et al. (2000), El Kadiri et al. (2001), Serrano et al. (2006) described the post-Triassic deposits. Concerning the late orogenic basins of eastern and central Betics (i.e. the youngest deposits which overlie the Ghomarides as well as the deeper nappes complexes), see Weijermars et al. (1985), Montenat et al. (1987), Weijermars (1991), García-Dueñas et al. (1992), Orozco et al. (1999).

The Ghomaride complex includes four nappes (Fig. 5.9) with different Paleozoic stratigraphy (Fig. 5.11B1–4). The nappes are separated from each other by relics of their Mesozoic-Cenozoic cover, mostly Triassic in age. The larger nappes, i.e. from bottom to top, Aakaili, Koudiat Tizian, and Beni Hozmar nappes crop out in northern Rif and partly in the Bokkoya. The highest, Talembote nappe forms a large tectonic klippe over the Dorsale in the Oued Lao area and some small outcrops in the Bokkoya (Fig. 5.7).

5.2.2.1 Paleozoic Formations

The Ghomaride Paleozoic formations are folded and recrystallized (Fig. 5.16), contrary to their Mesozoic-Cenozoic cover deposits. Therefore, they must be regarded as Variscan chips in the Rif Alpine belt.

The Lower Paleozoic sequences are rather homogeneous from one nappe to another, contrary to the Devonian (Fig. 5.11B2–4). The terrigeneous Ordovician deposits consist of phyllites with interleaved quartzite and meta-conglomerates. Carbonate layers first appear in the Silurian sequence, which contains black Graptolith-bearing cherts (Llandovery). This sequence is followed upward by a trilogy pillow basalts-lydites-micrites with *Orthoceras*, Tentaculites and Conodonts, all indicating the Lochkovian. These pelagic limestones are substituted by much thicker *Orthoceras* limestones in the Talembote nappe, with incipient reef buildings: this is the beginning of a strong paleogeographic differenciation. In the Aakaili nappe, the Devonian sedimentation continues with distal calci-turbidites (“calizas alabeadas” = “tortuous limestones”: cf. Fig. 5.16) where Famennian levels are dated near the top of the sequence. In the Koudiat Tizian nappe, a proximal, terrigeneous flysch (greywackes and pelites) is found. In the Beni Hozmar nappe, the latter deposits are substituted

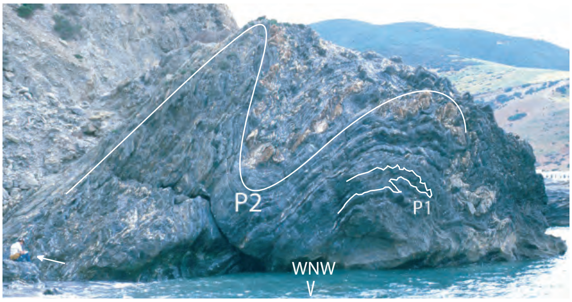


Fig. 5.16 Superimposed Eovariscan folds in the Upper Devonian calciturbidites of the Aakaili nappe (Lower Ghomarides) at Tamernout beach (see Fig. 5.9 for location). Scale is given by the person sitting on the left (*arrow*). A tight syn-metamorphic P1 fold is deformed by the P2 folds, which are associated with a SE-dipping crenulation cleavage

by thinner Tentaculite-bearing marls and limestones. Eventually, massive limestones ("calcaires zébrés") with *Favosites* reef-mounds occur in the Talembote nappe.

The Lower and Middle Paleozoic formations are affected by an Eovariscan folding phase associated with greenschist-facies metamorphism prior to the transgression of the unconformable Carboniferous sediments. The earliest Carboniferous levels are dated from the late Tournaisian in the Malaguides, which indicates a late Famennian-early Tournaisian age of the Eovariscan phase. The Carboniferous deposits begin with a lower sequence of proximal, turbiditic greywackes ending (at least in the Beni Hozmar and Talembote nappes) with coarse olistotromal conglomerates and boulders of Late Visean shallow water limestones. This lower, proximal sequence is followed upward by a more distal sequence interleaved by limestones with Late Visean-lower Namurian microfossils. Conglomeratic mass flows toward the bottom of this upper sequence contain pebbles reworked from the earlier Carboniferous sediments, the Silurian – Devonian deformed series, and even from a granitic and metamorphic basement, which is unknown in the Ghomaride-Malaguide complex, but occurs in the Kabylides.

The Aakaili Carboniferous sedimentation ends at Ceuta with coarse conglomeratic deposits comparable to the early Late Carboniferous Marbella Conglomerates in the Malaguides. Both conglomerates include peraluminous granite pebbles and resemble coeval conglomerates from Menorca (Balearic Islands), likely sourced from Central Iberia (Sanz de Galdeano et al., 2006). Likewise, the Early Carboniferous sediments of both the Ghomarides and Malaguides strongly recall those from Menorca. However, they also resemble the coeval deposits from the Moroccan Eastern Meseta, and contrast with those of Western Meseta where the Eovariscan phase is virtually lacking (Chalouan & Michard, 1990). Both cases support an eastern origin for the Ghomaride-Malaguide Complex in the Alpine, pre-collisional setting (cf. Sect. 5.7.4.3).

The restoration of the initial position of these Variscan chips, and more generally of the AlKaPeCa basement (Alboran microplate), at the beginning of the Paleozoic evolution is hindered by the superimposed effects of both the Variscan and Alpine orogenic cycles. Recent attempts of restoration have been based on

U-Pb zircon datings of the orthogneisses from the Alboran microplate elements. The Lesser Kabylia orthogneisses and the Peloritan porphyroids yielded Mid-Ordovician magmatic ages, which compare with the age obtained from the Sardinia gneisses (Trombetta et al., 2004; Helbing et al., 2006), suggesting relationships with the Variscan segments of south-western Europe. Contrastingly, Micheletti et al. (2006) obtained Late Neoproterozoic-Early Cambrian magmatic ages from the Calabrian augen gneisses, suggesting relationships with the high-K granitoid suite of the Anti-Atlas. The latter authors assume that the “Alboran microplate” – or, at least, its Calabrian part, detached from Gondwana after the early rifting and drifting of the Avalonian-Cadomian terranes (Raumer et al., 2003). Accordingly, the AlKaPeCa basement appears to be composed of different terranes whose collage resulted from the Variscan orogeny. The position of this complex after the western Tethys opening is discussed in Sect. 5.7.4.3.

5.2.2.2 Mesozoic-Cenozoic Cover

The post-Variscan sedimentation begins with Middle-Upper Triassic unconformable red beds (Fig. 5.15A2). These are thick, mainly fresh water deposits such as channelled quartzose conglomerates, arkosic sandstones, and gypsum intercalated ferrugineous clays. Paleogeographically, the entire succession evokes a mostly emergent continental shelf. Palynological data indicate the late Anisian-Ladinian (Baudelot et al., 1984) or Ladinian-Carnian (Diez, 2000). Locally, the clastic sequence is followed upward by dolomitic carbonates dated from the Carnian. Extensional tectonics related to the Tethyan opening is recorded by frequent synsedimentary faults and some alkaline basalt flows.

The remainder of the cover sequence occurs only in discrete strips. North of Tetuan, the El Onzar strip (Fig. 5.15B) overlies the Koudiat Tizian nappe, whereas two other strips (Belouazene, Kellalyine) occur beneath the latter nappe and overlie the Aakaili one through a tectonic contact which partly or totally cuts out the Triassic sandstones. Further to the north and west of Fnideq, the truncated Dradia strip lies on top of Beni Hozmar Paleozoic sediments (Fig. 5.9B).

In the most representative outcrops, the Triassic red beds are followed upward by Upper Triassic dolomites, passing up into massive limestones of Early Liassic age. This is a typical shelf sequence of the pre-rift stage. A possible emersion is suggested by karst-like structures formed during the Early-Middle Liassic: at that time, the area would correspond to the Tethyan rift shoulder or to the head of some large tilted block. However, the area is certainly submerged during the Toarcian as ammonites from that stage are reworked within breccias topping the deeply fractured Liassic limestones. The breccias are in turn topped by *Microcodium* limestones (Dradia) followed upward by whitish, bioclastic sandy limestones with Nummulites, Alveolines, and Discocyclines of Ypresian, Lutetian and Bartonian age. These Lower-Middle Eocene layers are, in turn, locally topped (Jebha, Fig. 5.9) by Upper Eocene conglomerates, which contain Hercynian granite pebbles (Iberian or Kabylian basement-sourced?)

The lack of any post-Toarcian sedimentary record up to the Late Cretaceous can be interpreted as being the result of a long-lasting emersion, or alternatively as the consequence of the Late Cretaceous-Eocene erosion of a thin Jurassic-Cretaceous sedimentary veil (Sect. 5.7). Indeed, some Malaguide sections show preserved Mesozoic deposits. In such sections, Martín-Martín et al. (2006) describe the following events: (i) emersion of the Triassic-Liassic platform before the Domerian (karst); (ii) rifting and submersion of the platform during the Middle-Upper Liassic (Toarcian – Dogger cherts and micrites); (iii) break-up unconformity at the bottom of Upper Jurassic pelagic limestones, either nodular or not; (iv) Cretaceous marls/calcschists deposited during thermal subsidence of the rifted area. These data allow us to correlate the Malaguide-Ghomaride domain with the Internal Dorsale one (Sect. 5.3.2).

5.2.2.3 Oligocene-Miocene "Post-Nappe" Cover

The youngest levels of the late orogenic, deeply unconformable Ghomaride-Malaguide cover, include two groups of formations (Fig. 5.15C), defined at the Betic-Rif scale, i.e. from bottom to top, (i) the late Oligocene-Aquitanian *Ciudad Granada Group*, and (ii) the Burdigalian *Viñuela Group* (Serrano et al., 2006). The *Fnideq Formation* and its Betic counterpart the *Alozaina Fm*. both belong to the Ciudad Granada Group. They begin with quartzose conglomerates passing upward to alternating sandstones and marls with benthic and pelagic fossils of Late Oligocene and Aquitanian age. These deposits are diachronous, being dated from the latest Rupelian (~29 Ma) in Eastern Betics (Sierra Espuña), and from the Aquitanian near Malaga (Western Betics).

The *Sidi-Abdeslam Fm* (Viñuela Group) outcrops are usually separated from those of the Fnideq Fm. For example, the Beni Maaden coarse conglomerates belonging to the Sidi Abdeslam Fm directly overlie the Paleozoic sediments. However, at Talembote (Fig. 5.15C), the Fnideq Fm. is unconformably overlain by conglomerates and green siliceous marls with Radiolaria and Foraminifera from the latest Aquitanian-earliest Budigalian. Thus, these levels can be correlated with the *Viñuela* (*Las Millianas*) *Fm*. defined around the Malaga basin. Their pebbles and boulders originate from local sources (Paleozoic and Triassic-Eocene units of the Ghomaride-Malaguide nappes). Cordierite migmatites, orthogneisses and phyllites comparable to the Kabylides upper plate basement or to the Sebtides-Alpujarrides crustal rocks can also be observed. However, in Andalucia, the Viñuela Fm. clearly contains Alpujarride rocks such as peridotites, kinzigites and garnet gneisses. Therefore, parts of the previously buried Alpujarride units have been exposed as early as 20–19 Ma ago.

Again at Talembote, the upper levels of the Viñuela Group correspond to brown-tobacco pelites, which yield nannoplankton assemblages of Early-Middle Burdigalian age. These levels contain sandstone intercalations comparable to the Numidian layers ("Neonumidian" of Andalucia), as well as olistostromes carrying rocks from the whole Internal complexes and locally from the Flysch nappes. By that time (~19–18 Ma), the Sebtide-Alpujarride units are totally exhumed, and the slopes inside the Gibraltar Arc are directed toward the Alboran Basin.

The latter formations are referred to as “post-nappe” formations as they unconformably overlie the varied Ghomaride-Malaguide nappes, and are never pinched in between. The Late Eocene uplift of the Paleozoic basement (granite pebbles at Jebha) and the deep erosion prior to the Oligocene-Aquitanian onlap suggest strong compressional deformations during the Late Eocene-Oligocene. In the eastern Malaguides (Sierra Espuña), some of the cross-sections constrain the age of nappe emplacement in the “middle” Oligocene (i.e. ~28 Ma ± 1 Ma). The Ciudad Granada deposits postdate this phase, being coeval with the beginning of the mountain belt collapse. The Viñuela deposits accumulate under the same extensional regime, and are contemporaneous with the opening of the Alboran Sea. Since the late Burdigalian (~18–17 Ma), these internal “post-nappe” formations are affected by major tectonic events, which cause the Dorsale and Flyschs to be thrust onto the more external domains, with backthrusting of parts of the Dorsale (Haouz range) and Flysch units (J. Zemzem) over the Ghomarides. Therefore, the so-called “post-nappe” formations of the Alboran Domain are indeed synorogenic deposits.

In Central and Eastern Betic Cordilleras, transgressive deposits of Serravallian, Tortonian and Messinian age unconformably overlie the “post-nappe” formations and underlying terranes, including the Nevado-Filabrides. They occupy large synclines, which can be regarded as emerged parts of the Alboran Basin itself, and formed through two main stages, before and after the Messinian. They represent late- to post-orogenic deposits: the exhumation of the Nevado-Filabride domes from 14 to 9 Ma controlled the depocentre evolution between 12 and 8 Ma. In the internal Rif, the post-Burdigalian deposits are restricted to Pliocene marls and conglomerates accumulated within the canyons formed during the Messinian salinity crisis (Fig. 5.4).

5.2.3 Dorsale Calcaire and Predorsalian

References: The most important and more or less recent works concerning the Dorsale are those by Wildi et al. (1977), Wildi (1979), Nold et al. (1981), Baudelot et al. (1984), Ben Yaïch et al. (1986, 1988), El Hatimi et al. (1991), El Kadiri et al. (1992), Maaté et al. (1993), El Kadiri (2002a, 2002b), El Kadiri et al. (2005). The neptunian dykes from the Algerian Dorsale (Djurdjura), Sicily and Calabria are described by Bouillin & Bellomo (1990) and Bouillin et al. (1999). Concerning the Predorsalian units, see Mourier et al. (1982), De Wever et al. (1985), Durand-Delga & Olivier (1988), El Hatimi et al. (1988), Olivier (1990), El Kadiri et al. (1990), Hlila et al. (1994), Durand-Delga & Maaté (2003), Durand-Delga et al. (2005, 2007). The Jebha Fault is described by Olivier (1981–1982) and Leblanc & Olivier (1984).

5.2.3.1 General

The Dorsale calcaire (Calcareous Range) is a complex tectonic domain, but its Mesozoic-Paleogene paleogeographic evolution is well constrained by a wealth of paleontological and sedimentological data. There, the deformed relics of the south paleomargin of the Alboran Domain may be observed, as exposed below (see also Sect. 5.7.4.3). However, the initial relative location of the ca. 30 elementary units or "nappes", which compose the Dorsale from the Strait of Gibraltar to Jebha and the Bokkoya (Fig. 5.17) is controversial. As for the Predorsalian, it consists of highly

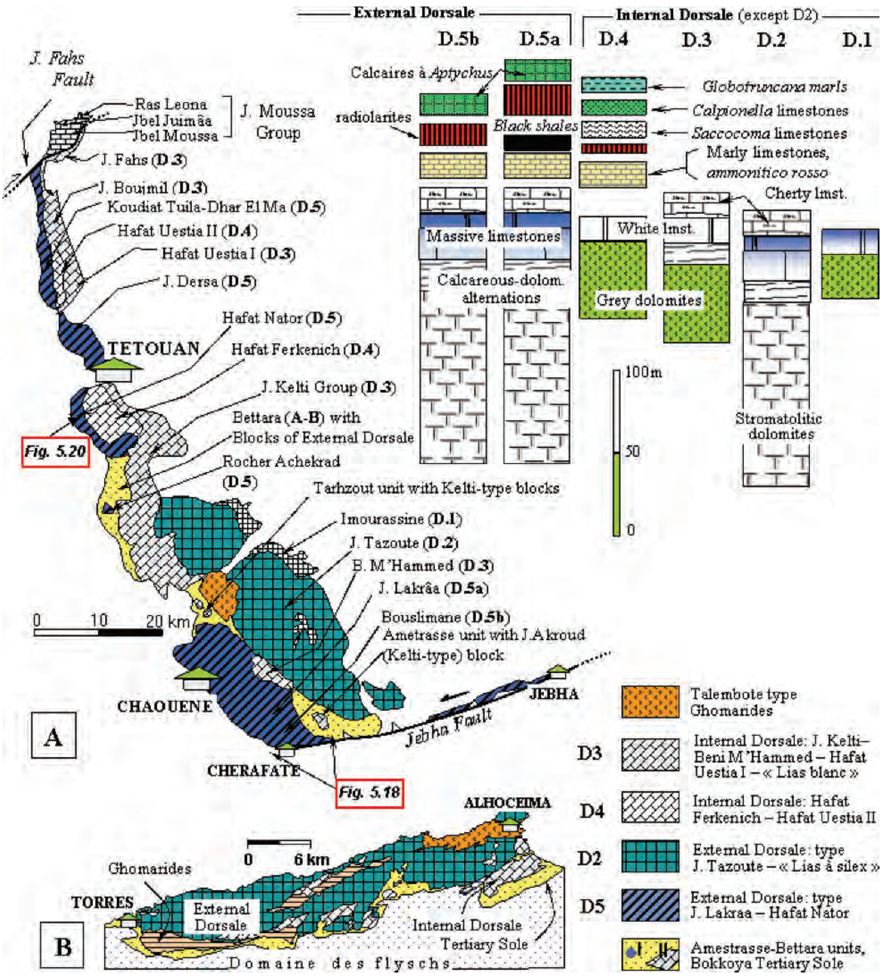


Fig. 5.17 Structural map of the "Dorsale calcaire" from Northern Rif (**A**) and Bokkoya area (**B**), and synthetic stratigraphic columns of the units, classified as Internal and External Dorsale units according to their stratigraphy. The Liassic strata are white massive limestones ("Lias blanc") in the Internal Dorsale, whereas they are dark cherty limestones ("Lias à silex") in the External Dorsale. The Internal-type unit D2 is intercalated within the External domain, either due to paleogeographic variations or out-of-sequence thrust. The maps also show the syntectonic formations of Late Oligocene-Aquitanian age (Ametrasse-Bettara units, Bokkoya Tertiary Sole) at the front of the nappe stack. Note that the town of Chaouen is frequently referred to as Chefchaouen

disrupted stratigraphic sections pinched within the narrow suture zone beneath the Alboran continental block, and admittedly corresponds to a former transition zone toward the Maghrebian Flyschs oceanic basin.

The two regions where the Dorsale units crop out in the Rif range, i.e. the Northern Rif and Bokkoya, are separated one from each other by the Jebha Fault, which extends offshore beneath the Alboran Ridge (Fig. 5.7). The geometry of the Jebha Fault is that of a sinistral lateral ramp with respect to the Internal Zone main thrust. The fault likely developed as a tear fault in the Internal Zone allochthon in front of a transverse structure of the underlying African margin. Due to an accessory vertical throw, the fault exposes a beautiful section across the Dorsale nappes and underlying units (Fig. 5.18). From there to the Tetouan valley, the Dorsale corresponds to a stack of moderately dipping imbrications, thrust over the Predorsalian, the Flyschs and the External Zones, and overlain either by the Ghomarides (Talembote) or the Sebtides (lower Oued Lao valley). This suggests a complex evolution of the main Alboran thrust (out-of-sequence thrust, and late extensional inversion; see Sect. 5.7). By contrast, in the Haouz range (north of Tetouan, Fig. 5.6) the varied Dorsale imbrications are nearly vertical and show a fan-like structure.

The Dorsale and Predorsalian stratigraphy corresponds to three contrasting sequences, which encompass the Triassic – Miocene time interval. The oldest, Late Triassic – Liassic sequence forms the competent part of the nappes, and on the other hand, allowed the authors to classify the varied nappes into “Internal Dorsale” and “External Dorsale” although, in some cases, this traditional classification does not correspond to the actual position of the nappes. The intermediate, Liassic–Paleocene sequence records the evolution of a Tethyan paleomargin, and finally the youngest, Eocene-Miocene sequence records that of the Alpine orogeny.

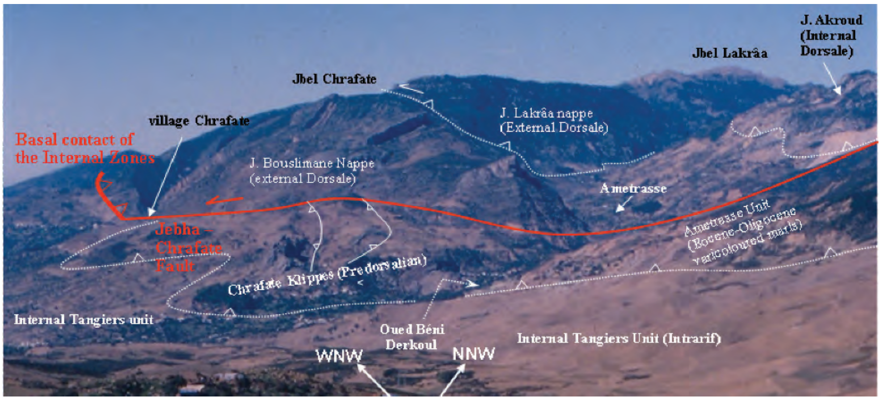


Fig. 5.18 Southwest corner of the Northern Rif Dorsale, looking NW-ward from the Ketama-Bab Taza road. Location: see Fig. 5.17. Note the lack of any Flysch Nappe outcrop between the Predorsalian and Intrarif (Tanger) units along the Jebha-Cherafat wrench fault. A huge rock fall slid recently from J. Akroud down to the Ametrasse houses

5.2.3.2 Triassic-Liassic Massive Carbonates

Triassic-Liassic massive carbonates characterize the Dorsale range. In some cases these calcareous slabs overlie upper Carnian evaporites (Fig. 5.19A), which clearly compare with the corresponding levels of the Ghomarides (Fig. 5.15A2, A3), and

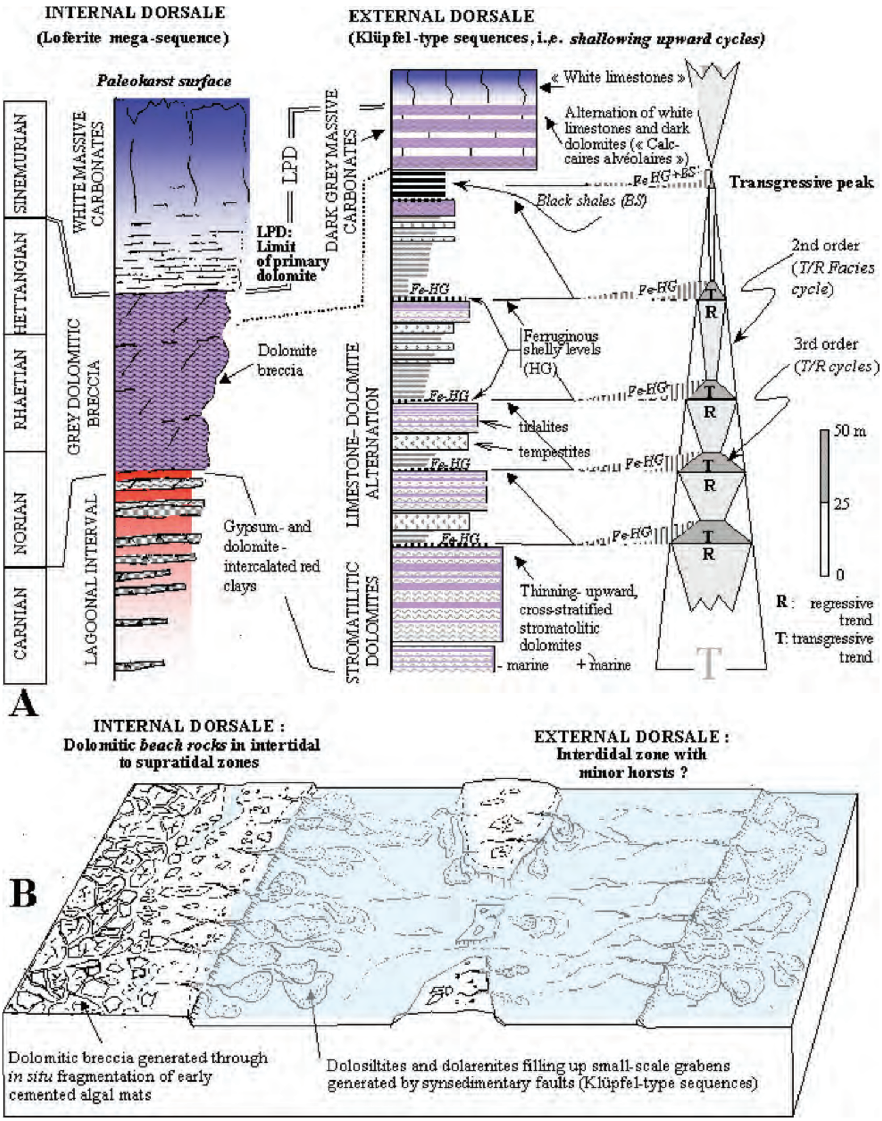


Fig. 5.19 (**A**): Comparison of the Late Triassic-Early Liassic stratigraphy of the Internal *versus* External Dorsale domains. LPD: Upper limit of primary dolomite. – (**B**): Hypothetic paleogeography, after El Kadiri & Faouzi (1996), modified. The horst within the external platform would correspond to unit D2, Fig. 5.17. A simpler paleogeographic pattern could be alternatively imagined, assuming a complex thrust tectonics (out-of-sequence D2 thrust)

played a major role in the detachment of the Dorsale units from their original (and controversial) basement.

The carbonate sequence itself begins with greyish dolomites. These are dolomitic breccias in the Internal Dorsale units (ID), and laminated, stromatolithic dolomites followed by alternating limestones-dolostones layers in the External Dorsale units (ED). They represent the Norian, Rhetian and Hettangian *pro parte* intervals. Their facies are typically Alpine, comparable with the coeval deposits of internal Briançonnais, Eastern Alps or Tuscany, and strongly contrast with the German facies of the Maghrebide and Betic External Zones.

The differences between ID and ED units increase in the Hettangian-Sinemurian massive limestone levels. ID units comprise white limestones, devoid of laminated dolomicrite facies, but still rich in stromatolitic structures and Dasycladaceae. In contrast, ED units display dark limestones, partly dolomitic (cellular limestones), but where Ammonoidae now occur. These lateral changes suggest that two neighbouring paleogeographic domains occurred at that time, i.e. a confined, shallow water, although subsiding basin (ID), and a deeper basin opened toward the oceanic domain (ED). In other words, ID units correspond to the internal carbonate shelf of the former passive margin, and ED to the external shelf, which is supported by their Jurassic-Cretaceous evolution (Sect. 5.2.3.3). This initial setting has been so deeply altered by the Oligocene-Miocene tectonics that it seems risky to infer from the present-day location of the varied units any more specific arrangement of the broad paleogeographic reconstruction such, for example, the occurrence of an internal-type horst in the middle of the external shelf (a proposal of one of us, K.E.K, which is shown in Fig. 5.19B).

5.2.3.3 Jurassic-Paleocene Pelagic Formations

During the Jurassic, sedimentation became condensed and pelagic throughout the Dorsale domain, with some significant gaps in the stratigraphic record. This suggests a submarine plateau setting. However, some differences still occur between the ID and ED.

In the *Internal Dorsale* (Fig. 5.17, columns D3, D4), following sedimentation of the Sinemurian white limestones, we observe a gap, which may attain a few millions years (D4: late Sinemurian – early Carixian), and involves an emersion with paleokarst. Then the karst is submerged and is filled and progressively covered by thin middle Carixian sediments, followed by Domerian and Toarcian *ammonitico rosso* nodular limestones. A second gap involving some erosion in part of the sections corresponds to the Middle Jurassic and the basis of Late Jurassic. It ends with the sedimentation of late Kimmeridgian-Tithonian radiolarites, followed upward by *Saccocoma* micritic limestones, then *Calpionella* limestones from the Tithonian-Berriasian. A third gap corresponds to the Early Cretaceous and part of the Late Cretaceous. It is accompanied by some erosion of the previous pelagic sediments and locally (Hafa Ferkenich) by fault scarp breccias. Cenomanian and/or Turonian pelagic sediments are locally preserved, but in most cases sedimentation resumes later with the "Couches rouges" and "Couches blanches" *Globotruncana* marls sedimentation, dated from the Campanian and Maastrichtian, respectively. The pre-orogenic sedimentation ends with Paleocene *Globigerina* black shales.

The origin of the successive gaps, and particularly of the younger ones (Toarcian-Kimmeridgian and Berriasian-Campanian) is a matter of debate. The intervention of aerial erosion and karstification has been suggested by one of us (K.E.K.) even for the two younger gaps, based on the observation of dissolution breccias and speleothems (palissadic calcite) on both sides of the pelagic infillings. However, the fact that these gaps occur between deep, pelagic episodes would suggest (A.M.) submarine gaps related to the existence of sea floor currents. It is worth noting that neptunian dykes with calcite prisms disposed radially to any interface between pelagic infilling and country rock was described in Paleozoic phyllites and granites from varied eastern Dorsale segments (Djurdjura, Peloritan Mountains, Calabria). Additional studies are needed to check these alternative interpretations.

In the *External Dorsale* (ED), there is no aerial karst during the Early Liassic, and *ammonitico rosso* deposits are observed as early as the late Sinemurian (Fig. 5.20). During this time interval, the ID belongs to the rift shoulder by contrast with the subsiding ED. Cherty limestones (which correspond to slope facies) are deposited there during Middle Liassic, then marly limestones during the Toarcian, and radiolarites during the Middle Jurassic and most of the Late Jurassic. The middle-late Tithonian and Berriasian correspond to *Aptychus* micrites as in the ID, but in the ED these micrites form the matrix of chaotic breccias (Cherafat, El Queddane). This suggests an increasing activity of normal faults related to the Tethyan rifting (Sect. 5.7). Subsequently, during the Cretaceous-Paleocene times, the DE evolution compares with that of the DI.

5.2.3.4 Eocene-Lower Miocene Unconformable Formations

A dramatic paleogeographic change took place during the Eocene, like in the Ghomaride-Malaguide domain. As early as the Early Eocene, unconformable sandy bioclastic limestones (J. Gorgues) and Nummulite rich, chaotic breccias accumulate over the varied Mesozoic formations. In places the latter breccias are overlain by Middle-Late Eocene bioclastic limestones (J. Lakraa). Other sections display conglomerates, calcareous sandstones and bioclastic limestones from the Late Eocene (J. Akroud). Hence, the Dorsale domain was strongly uplifted during the Eocene, compared to its Cretaceous position.

The Eocene sediments sometimes covered by ferrugineous crusts are overlain by Lower-Middle Oligocene coloured marls, followed upward by reddish-brown micaceous sandstones dated as Late Oligocene-Aquitanian. Conglomeratic and even chaotic formations first appear in the coloured marls, and then become abundant (Bettara, Taghzoute, Ametrasse, Tertiary sole of the Bokkoya). These deposits again evoke those from the Ghomaride-Malaguide (see Fig. 5.15, and also Fig. 5.42). They are synorogenic formations deposited just before the emplacement of the Dorsale slivers and overturned folds onto the Predorsalian domain. Likewise, the Haouz

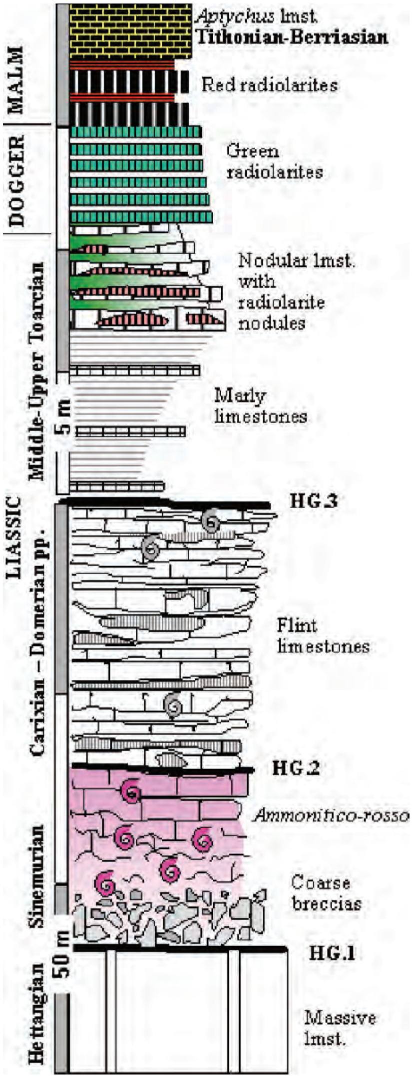


Fig. 5.20 An example of Jurassic "condensed series" from the External Dorsale, east flank of Hafat Nator, 4 km SW of Tetouan (unit D5 in Fig. 5.17), after El Kadiri (1992), modified. Note that the pelagic facies begin later in the Internal Dorsale, i.e. in the Domerian (*ammonitico-rosso*) and Tithonian (radiolarites)

backthrusting over the Ghomarides arose after the Early Burdigalian. Note that the Oligocene-Miocene deposits labelled "post-nappe" in the Ghomaride-Malaguide realm are "pre-nappe" in the Dorsale domain indicating a progressive outward migration of the deformation.

5.2.3.5 Predorsalian Domain

The narrow Predorsalian domain corresponds to relics of the transition zone between the continental margin of the Internal Zones (Dorsale domain) and the oceanic domain of the Maghrebian Flyschs. In fact, it is strongly disrupted along the suture, which separates these major domains. Two typical examples of Predorsalian sections from Northern Rif deserve illustration, namely the J. Moussa and Cherafate sections.

The *J. Moussa* (Fig. 5.21) is homologous to the Gibraltar Rock north of the Strait: they are the Ancients Pillars of Hercules. Their specific stratigraphic succession defines the "Tariquide Ridge" sub-domain (after the Arabic name of the Gibraltar Rock: Jebel Tariq). The J. Moussa succession (Fig. 5.20, right) begins with Triassic-Lower Liassic carbonates comparable with those of the Internal Dorsale, then continues from Toarcian onward in a way similar to that of the External Dorsale, except that the Aalenian-Bajocian radiolarites are substituted by nodular, ammonite-rich limestones. During the Late Jurassic, the ridge status of the sub-domain is only justified relative to the distal DE domain. Interestingly, the sequence continues upward with terms comparable with both the Dorsale and Beni Ider Flysch successions: "*Aptychus* complex" of Early Cretaceous age, Maastrichtian-Paleocene "Couches rouges", Upper Eocene-Oligocene coloured marls including upward increasing clastic-nummulitic mud flows. The sequence ends with Aquitanian holoquartzose sandstones with quartz pebbles (Numidian facies) and Lower Burdigalian brown pelites, which compare with the Ghomaride "post-nappe" formations as well as with the "pre-nappe" Dorsale equivalents.

The *Cherafate (Chrafat) klippes* occupy a smaller surface than the J. Moussa unit, and correspond to sedimentary klippes slid within the Tertiary clays of the Ametrasse unit (Fig. 5.18). The main, Beni Derkoul klippe shows an overturned

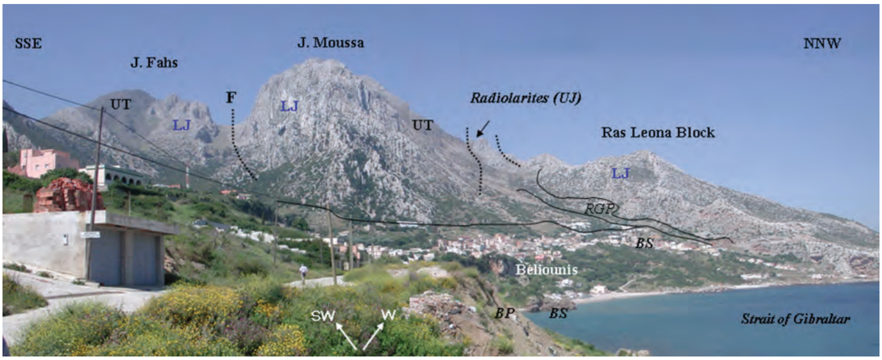


Fig. 5.21 The J. Moussa massif as seen from Beliounis (Ben Younes) village (the so-called panorama de la Mujer Muerta). For location, see Fig. 5.9. Together with the homologous Gibraltar Rock, the J. Moussa Group defines the Tariquide sub-zone of the Predorsalian distal zone (Durand-Delga et al., 2005, 2008). In contrast, the J. Fahs belongs to the proximal Internal Dorsale domain (unit D3) of the Alboran block paleomargin. The J. Moussa massif and the Ras Leona sliver overthrust the Predorsalian Cenozoic formations (BS, BP). The houses of Beliounis are built on a middle Pleistocene uplifted terrace. BP: Brown tobacco pelites (Burdigalian); BS: Beliounis sandstones (Aquitanian-lower Burdigalian); F: fault; LJ: Lower Liassic limestones; RGP: Maastrichtian-Paleocene red and grey pelites; UJ: Upper Jurassic radiolarites; UT: Upper Triassic dolomites

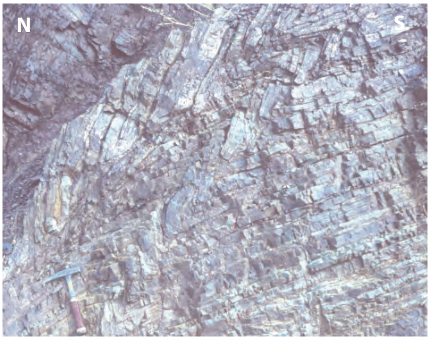


Fig. 5.22 Jurassic radiolarite/calci-radiolarite from the Predorsalian Beni Derkoul outcrops. Location: see Fig. 5.18. Note the S-vergent chevron-like folds that affect these well bedded sediments, associated with a north-dipping shear plane (*upper left*)

sequence including more distal terms than the coeval facies from the neighbouring Dorsale, such as thick Upper Jurassic radiolarites/calciradiolarites (Fig. 5.22). Therefore, they do not originate from the Dorsale, but from a specific domain, i.e. the Predorsalian domain.

The "Bokkoya Tertiary Sole" allows us to separate two periods of olistostrome emplacement, (i) the Late Oligocene, with a source area in the External Dorsale, and (ii) the Burdigalian, with a source area in the Internal Dorsale. Between both olistostrome accumulations, the sequence includes Aquitanian marls with interbedded Numidian-facies sandstones. The original substratum of this "Tertiary Sole" is unknown.

5.3 Maghrebian Flyschs

References: The main recent works on the Maghrebian Flyschs stratigraphy in the Gibraltar Arc are those by Thurow & Kuhnt (1986), Durand-Delga & Olivier (1988), Esteras et al. (1995), Olivier et al. (1996), Durand-Delga et al. (1999), Puglisi et al. (2001), Zaghloul (2002), El Kadiri et al. (2003, 2006), Zaghloul et al. (2007). The publications by Chalouan et al. (2006b) and Crespo-Blanc & Frizon de Lamotte (2006) more specifically address the Flysch nappe structure. Correlations at the Maghrebide scale and paleogeographic reconstructions are addressed by Bouillin (1986), Hoyez (1989), Durand-Delga et al. (2000), Guerrera et al. (2005), and De Capoa et al. (2007).

As defined in Sect. 5.1.2.2, the Maghrebian Flyschs form relatively thin, but extensive thrust-nappes over the External Zones and restricted back-thrust elements (e.g. J. Zemzem) over the Ghomaride-Malaguide complex (Figs. 5.5, 5.6, 5.7 5.8). Turbidite sequences ("flyschs") are dominant in these nappes, but clay-dominated sequences occur at the bottom of each nappe ("pre-flysch" sequences).

From an historical point of view, it is worth noting that the occurrence of large Flysch inliers over the Kabylian Oligocene-Miocene cover have been used as an argument for an internal origin of the Flyschs. This "ultra-kabylian" hypothesis, warmly debated in the 60s–70s, has been abandoned since.

5.3.1 Mauretanian Nappes

Above a thin Jurassic series including Upper Jurassic radiolarites, the *Tisiren nappe* series (Fig. 5.23A) begins with a pre-flysch sequence consisting of *Aptychus* and *Calpionella* marly limestones from the Berriasian-Valanginian. This sequence played the role of a décollement level at the bottom of the competent Flysch mass. The first turbidite sequence consists of graded siliciclastic layers interleaved with argillaceous-pelitic horizons where palynomorphs indicate a Hauterivian-Barremian age (Fig. 5.24). Then, a dominantly pelitic sequence occurs (Barremian-early Aptian interval), followed upward by a new turbiditic cycle (late Aptian-middle Albian). The lateral extent of these facies is considerable, i.e. from Western Betics (Los Noguales) to Algeria (Guerrouch), and to Sicily (Monte Soro).

The *Beni Ider nappe* series represents the detached ("diverticulated") upper part of the Mauretanian basin infilling. It includes a lower "pre-flysch" series, which operated as a décollement level between the Tisiren and Beni Ider successions. The "pre-flysch" series (Fig. 5.23B) begins with upper Albian-Cenomanian-Turonian spongolites and black shales. These levels are followed upward by Upper Cretaceous coloured pelites and calciturbidites, which most often include breccias with carbonate elements from the Dorsale domain. The Campanian layers also include reef fragments and Rudists of unknown origin, whereas the Maastrichtian flysch contains Permian-Triassic (Verrucano) fragments likely reworked from Ghomaride or similar sources. Calciturbidite flows, emplaced during the Paleocene, include reworked *Microcodium*. Thick sandstone layers with local nummulite accumulations correspond to the Early Eocene, and nummulitic turbidites represent the Middle-Late Eocene. The Eocene-Oligocene transition is marked by the emplacement of chaotic breccias and olistoliths within greenish-reddish pelites ("flysch coloré").

The series is topped by a thick sandy-micaceous turbidite accumulation (Fig. 5.23C), referred to as the "Flysch gréso-micacé" (Sandy-Micaceous Flysch) or "Flysch à Lépidocyclines" throughout the Maghrebides, or "Algeciras Flysch" in Andalucia, and dated by planctonic foraminiferans and nannoplankton from Late Oligocene to middle Burdigalian. One can recognize two main turbidite sequences (I, III) and a dominantly pelitic sequence (II) in between. Sequence I ends with an olistostrome where Dorsale fragments are reworked (Jurassic and Eocene limestones), as well as low grade metamorphic elements (Ghomaride?). Sequence II includes by place (Anjra unit) Numidian-like layers. Eventually, the flysch series ends with brown pelites and breccias with schist elements (sequence IV), which closely compare with the Sidi Abdeslam Fm (Sect. 5.2.2.3).

Each Mauretanian nappe displays a particular structural style in relation with its particular mechanical stratigraphy. Weak layers dominate in the Beni Ider nappe, allowing external-verging folds to form. Contrastingly, the stiff Tisiren material results in the formation of thrust sheets over the previously detached Beni Ider nappe. It is worth noting that the Tertiary cover of the Mauretanian Flyschs does not exist along the Bokkoya transect, suggesting that they have been back-thrust over the In-

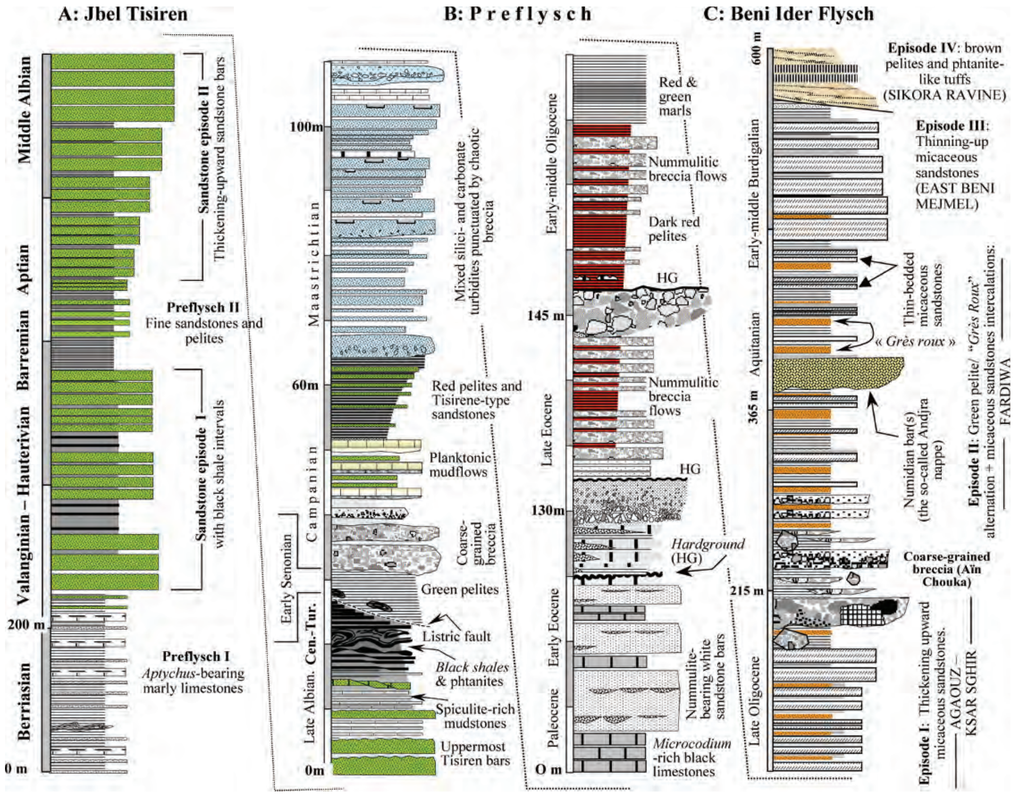


Fig. 5.23 Stratigraphy of the Mauretanian Flyschs from the Rif Belt. (**A**): J. Tisiren nappe, after Durand-Delga et al. (1999), modified. The Jurassic base of the nappe (radiolarites, basalts, locally Liassic limestones) is not shown. – (**B**) ("Préflysch") and (**C**) ("Flysch") of the Beni Ider nappe (detached from the *top* of the Tisiren series), after El Kadiri et al. (2003, 2006), Zaghloul et al. (2007)

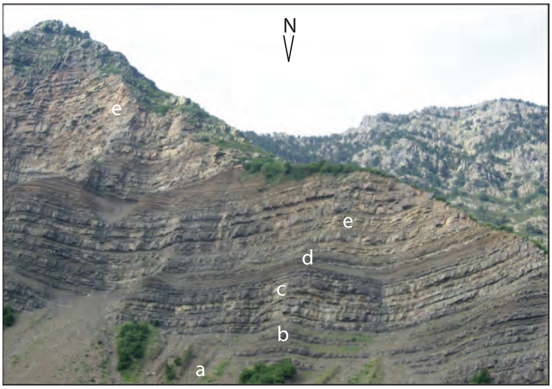


Fig. 5.24 Southern slope of the J. Tisiren, viewed from the Ketama-Bab Taza road (see Fig. 5.7 for location). The turbiditic sequence represents a 30 Ma interval, with Upper Berriasian-Valanginian (**a, b**), Upper Valanginian-Hauterivian (**c, d**), and Upper Hauterivian, Barremian, Aptian and lower Albian (**e**). Stratigraphic dates are based on nannofloras from the most pelagic episodes (Durand-Delga et al., 1999)

ternal Zones, similar to the J. Zemzem and Riffiene klippes south of Ceuta and to their widespread equivalents over the Kabylias.

5.3.2 Massylian Nappes

The name of the *Chouamat-Meloussa nappe* was based on two localities of the Central Rif and Tanger (Tangier) area, respectively. This nappe compares with the Massylian nappe *sensu stricto* of Algeria, and occurs as sheared sheets beneath the Mauretanian or Numidian nappes. This supports the idea that the Massylian basin was located externally with respect to the Mauretanian, and that the Numidian represents the detached upper part of the Massylian sedimentary pile.

The Chouamat-Meloussa series begins with an Aptian-Albian siliciclastic flysch, relatively fine grained when compared with the coeval Tisiren flysch. This 700 m-thick flysch formation is followed upward by thin (20 m) black cherts and calcareous microbreccias dated as Cenomanian-Turonian, and then by Senonian pelites and microbreccias (100 m). This specific succession can be correlated to the north with the Mauretanian one, and to the south with the Intrarif (Ketama unit; see below).

The Numidian nappe is named after the particular lithologic facies of the "numidian sandstones", defined as early as 1890 in the Kabylian coast. Such sandstones make up most of the Numidian nappe, which can be easily identified from Sicily to Andalucia (Algibe sandstones). In the Rif Belt, the Numidian nappe comprises extended, but relatively thin thrust elements or klippes. They include tilted, folded and occasionally overturned beds (J. Zinat) abruptly truncated along the nappe sole thrust (Fig. 5.25). The Numidian allochthons overlay the varied Intrarif units (either directly or through Chouamat-Melloussa slivers), except the J. Zemzem

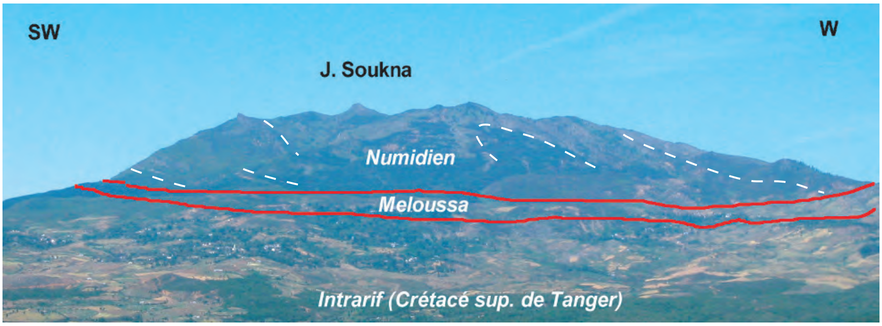


Fig. 5.25 A typical Maghrebian Flysch klippe as outcropping at J. Soukna, viewed from Chaouen (see Fig. 5.7 for location). The main part of the klippe corresponds to Aquitanian siliciclastic turbidites whose NE-dipping layers (SW-vergent folds) are deformed by a SW-vergent anticline, troncated by the sole thrust of the nappe. The Cretaceous pelites of the Meloussa nappe form a plastic cushion at the *bottom* of the competent Numidian nappe, and on *top* of the Intrarif Upper Cretaceous shales (Tanger unit)

klippe, which overlies the Internal Zones and probably originates from a relatively internal part of the Numidian basin (Anjra unit).

The stratigraphic pile (Fig. 5.26) begins with a “pre-flysch” series, namely the *Argiles sous-numidiennes* (Infra-Numidian Clays). These are vari-coloured clay deposits including scarce bioclastic layers and Fe-Mn rich concretions referred to as *Tubotomaculum* (likely epigenised crustacean burrows). They were deposited at depth during the Late Lutetian to Late Oligocene times, according to both the pelagic foraminiferans and nannoplankton associations. The weakness of this formation greatly favoured the detachement of the more than 1000 m thick Numidian pile.

The Numidian sandstones sensu stricto consist of thick yellowish, poorly cemented sandstsones, which include scattered, almond-sized quartz pebbles. Graded bedding is poorly marked, the sandstones layers are channelized, often amalgamated, and organized into upward thinning sequences interleaved with reddish pelites, suggesting a system of fluxoturbidites emplaced in deep lobate fans. The origin of the highly mature sediment (quartz sand) is to be found in the reworking of previous sandy deposits, possibly the Jurassic-Cretaceous Saharan continental sandstones, which itself has been fed by the erosion of the Cambrian-Ordovician sandstones overlying the Saharan crystalline shields. The Saharan sands would have entered the Ligurian-Maghrebian Ocean through some gates (e.g. in Tunisia, according to Hoyez, 1989), being then distributed by longitudinal turbidity currents.

The age of the Numidian sedimentation, latest Oligocene-Aquitanian, is constrained only by the age of the bounding levels. Indeed, the turbidite sequence is overlain by brownish, siliceous supra-numidian clays dated from the latest Aquitanian and early-middle Burdigalian. Flint layers of the same age are found on top of the Beni Ider series, and these siliceous levels are possibly related to an early volcanic event (Sect. 5.5).

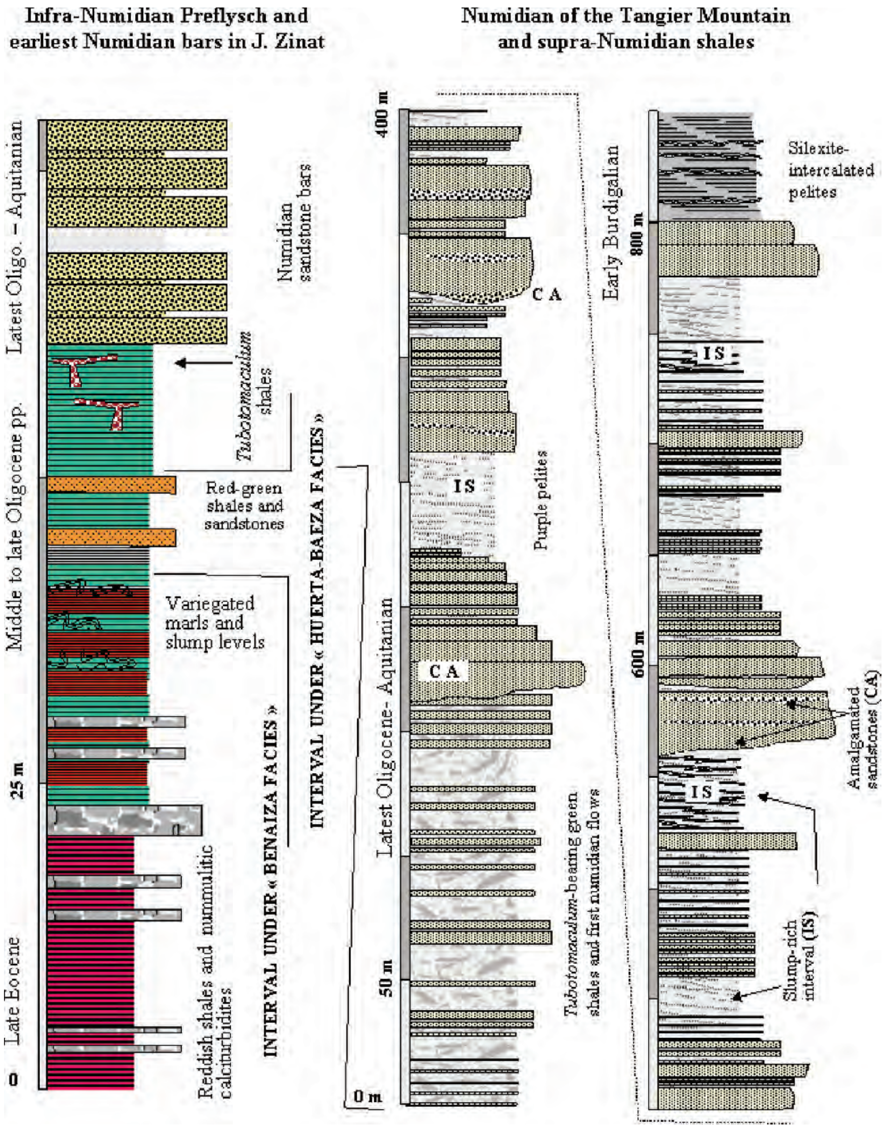


Fig. 5.26 Stratigraphy of the Numidian nappe from the J. Zinat (*left*) and Tangier (Tanger) Mountain (two columns on the *right*; note the different vertical scale used in the *left*-hand column). See Figs. 5.6 and 5.7 for location

As a whole, the Numidian flysch appears to be coeval with the Beni Ider sandy-micaceous flysch. Thus, it is not suprising to find mixed series such as the Talaa-Lakrah (Tanger) and Bolonia (Andalucia) series. Their stratigraphy compares with the Beni Ider one, except the occurrence of interbedded Numidian-type sandstones layers with almond-like quartz pebbles, sometimes located within sand-balls. This suggests that southeastern, Numidian inputs were converging with northern ones (Beni Ider).

5.3.3 Paleogeography

The Maghrebian Flyschs nappes, which display turbiditic sequences of Cretaceous-Tertiary age are detached from an oceanic or thinned continental crust domain. Ophiolite slivers occur at the bottom of the more or less recrystallized flyschs from southern Apennine Calabria, Sicily, and even beneath the Lesser Kabylia allochthon (Rekkada-Metletine serpentinites, gabbros, pillow basalts and radiolarites dated from the Jurassic). Interestingly, the Achaiche Mauretanian unit, also overthrust by the Lesser Kabylia massif, displays a continental margin-type pre-flysch series with Paleozoic metapelites, Triassic sandstones, Liassic carbonates intruded by thick sills of basalts, limestones and radiolarites of Dogger-Malm age, and Berriasian pillow basalts. Further west, Middle-Upper Jurassic pillow basalts and radiolarites associated with siliceous micrites crop out beneath the Mauretanian nappe south of the Bokkoya (Izroutene), whereas an agglomerate of Middle Jurassic limestones and variolitic pillows occurs at the bottom of the J. Chouamat Massylian nappe. Pillow basalts are also found as olistoliths within the Bokkoya Tertiary sole. In contrast, no basalt occurs in the Ouareg sheet beneath the Tisiren nappe close to Targuist, the series of which displays only Toarcian-Bajocian calciturbidites and Middle-Upper Jurassic radiolarites.

It is suggested that the Ligurian-Maghrebian basin was a true oceanic basin in its eastern, wider part, whereas it was floored mostly by thinned continental thrust in its narrower western part. However, the tomographic studies image a deep cold slab, about 600 km wide E-W and N-S below the south Balearic basin, which would correspond to a former Ligurian-Maghrebian oceanic substrate subducted down to the base of the asthenosphere (Sect. 5.7). This indicates additionally that not only most of the Ligurian-Maghrebian crust, but also a large part of the overlying sediments disappeared by subduction: the Flysch nappes and their scarce basal slivers are tiny remnants of the lost oceanic domain.

The Mauretanian sub-basin was located on the northern side of the Maghrebian basin, as indicated by the detrital elements reworked from the AlKaPeCa domain in the Beni Ider breccias and coarse turbidites. The sandy-micaceous flysch was probably fed by the erosion of the Ghomarides-Malaguides and equivalent Kabylian terranes, like the “post-nappe” cover of these mostly Paleozoic domains. Conversely, the Massylian sediments accumulated on the southern side of the Maghrebian basin, close to the African margin. This is consistent with the similarities between the Cretaceous series of the Massylian nappes and Intrarif units (see below), and also with the African source of the Numidian turbidites.

5.4 External Zones

References: The internal parts of the External Zones (Intrarif, Mesorif) have been the subject of a number of publications in the last decades: Andrieux (1971), Vidal (1977), Leblanc (1979), Leblanc & Wernli (1980), Wildi (1981, 1983), Leblanc & Feinberg (1982), Septfontaine (1983), Monié et al. (1984), Frizon de Lamotte (1985, 1987), Morley (1987, 1992), Asebriy et al. (1987), Ben Yaïch (1991), Ben Yaïch et al. (1991), Favre (1992), Michard et al. (1992), Asebriy (1994), Benzaggagh (1996), Elazzab et al. (1997), Abdelkhaliki (1997), Azdimousa et al. (1998), Toufiq et al. (2002), Zakir et al. (2004), Zaghloul et al. (2005), Crespo-Blanc & Frizon de Lamotte (2006), Negro et al. (2007), Michard et al. (2007).

As for the more external parts (Prerif and foredeep) and more recent, post-nappe deposits, the main recent references are the followings: Guillemin & Houzay (1982), Feinberg (1986), Wernli (1987), Benson et al. (1991), Kerzazi (1994), Flinch (1996), Benzaggagh (1996), Zizi (1996, 2002), Samaka et al. (1997), Lamarti-Sefian et al. (1998), Plaziat & Ahmamou (1998), Bernini et al. (1999), Krijgsman & Langereis (2000), Litto et al. (2001), Zouhri et al. (2001), Münch et al. (2001), Bargach et al. (2004), Chalouan et al. (2006a).

As reported above (Sect. 5.1.2.3), the Rif External Zones are divided into three zones, according to structural and stratigraphic criteria, i.e. from NE to SW the Intrarif, Mesorif and Prerif (Figs. 5.7, 5.8).

5.4.1 Intrarif

The Intrarif zone includes the most distal units derived from the African paleomargin. These units crop out immediately beneath the Maghrebian Flyschs and Dorsale units.

The *Ketama unit* consists mainly of Lower Cretaceous series where siliciclastic turbidites predominate. These terms are intensely folded and affected by low grade metamorphism (Sect. 5.5). The Ketama unit extends into the Central Rif range, being limited eastward by the Nekor Fault, the nature of which is discussed in Sect. 5.4.2. The Ketama unit involves, at least, two sub-units. The southern unit begins with Sinemurian massive limestones (Fig. 5.27A), suggesting a continental substrate. Then the syn-rift series involves ammonite-rich marly limestones (Middle Liassic), silty marls (Toarcian), hemipelagic “calcaires à filaments” (Aalenian), and Posidonomya marls (Dogger). The overlying post-rift series consists firstly of Upper Jurassic “ferrysch” (a rythmic marly-clastic formation), followed upward by Tithonian-Berriasian pelagic limestones. Then a pelitic-sandy sedimentation corresponds to the Valanginian, Barremian and Aptian-Albian times, being characterized by thick quartzose turbidites (Fig. 5.28). Vraconian belemnite marls and spongolites are preserved to the north.

The *Tanger unit*, partly detached from the underlying Ketama, and the *Aknoul nappe*, totally detached and thrust over the Mesorif and Prerif (and even over the Middle Atlas foreland in the easternmost Rif), expose the Upper Cretaceous-Eocene

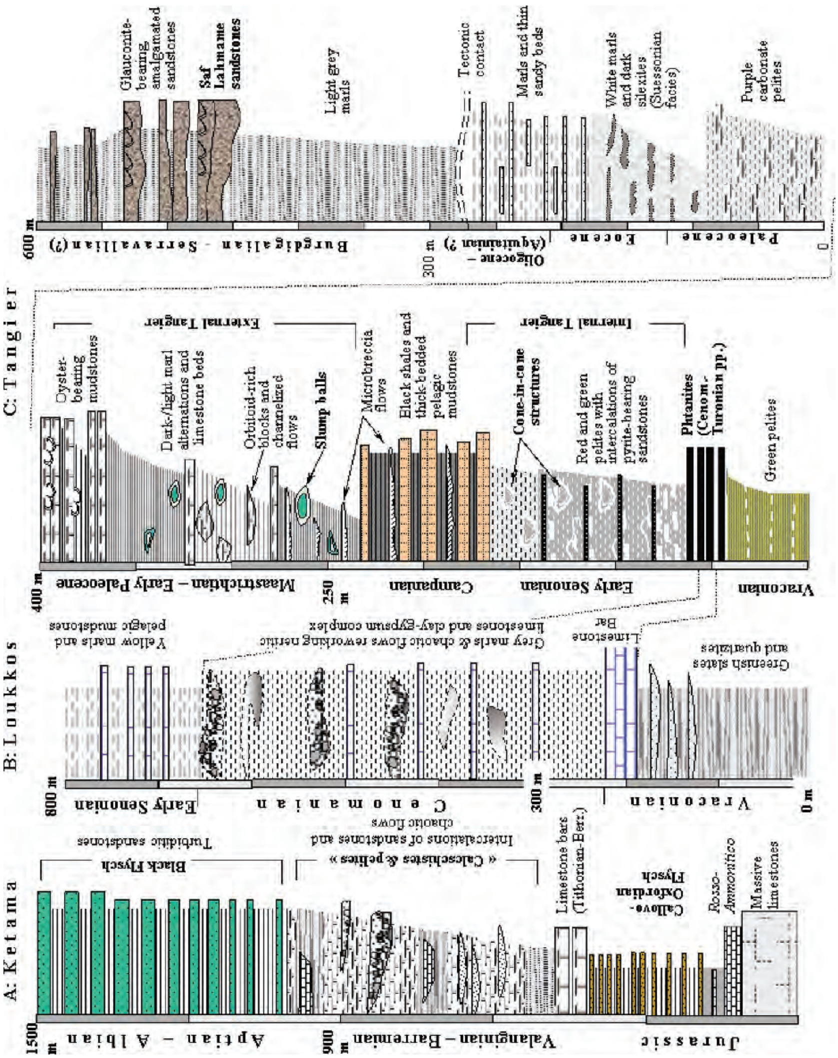


Fig. 5.27 Stratigraphy of the Intrarif "rooted units". (**A**): Ketama unit, after Frizon de Lamotte (1985). – (**B**): Loukkos unit, after Lespinasse (1975). – (**C**): Tanger unit; Saf Lahmane Neogene sequence after Zaghloul et al. (2005). Saf Lahmane is located 25 km south of Tanger (Tangier)

Fig. 5.28 Large load casted flute casts at the bottom of an overturned siliciclastic turbidite strata from the Lower Cretaceous Ketama unit, 20 km west of Ketama town. Two superimposed paleocurrent directions can be distinguished (*arrows*)

marly-pelitic formations of the Intrarif zone (Fig. 5.27C). The Tanger series spans the Cenomanian-Maastrichtian interval in Central Rif, whereas it is diverticulated in Western Rif into an Internal Tanger unit (Cenomanian-Senonian) and and External Tanger unit (Campanian-Paleocene). Some facies contrast with the dominant pelites, such as the Cenomanian-Turonian phtanites, the lower Senonian *cone-in-cone* nodules, the Campanian-Maastrichtian calcareous microbreccias, and the *Ostrea* marly limestones at about the K-T boundary.

The *Loukkos unit* conventionally belongs to the Intrarif, although it forms a transition with the western Mesorif. Compared to the Tanger unit, the Loukkos unit (Fig. 5.27B) is typified by thicker Cenomanian deposits, a higher proportion of carbonates, and frequent diapiric intrusions of the Triassic clay-gypsum complex. As for the younger formations of the former Intrarif basin, they have been detached and diverticulated, except locally (Saf Lahmane; Zaghloul et al., 2005), to form the *Habt nappe* in the west, and the *Ouezzane and Tsoul nappes* south and southeast. Indeed, Intrarif Eocene sediments consist of white siliceous marls and marly limestones, which acted as a décollement level. Sand input increased during the Middle Eocene. During the Oligocene, olistostromes occurred in part of the Habt (Rirha and Meliana units), whereas a siliciclastic flysch, namely the *Asilah-Larache Sandstone*, accumulated in the external Habt sub-basin. The latter turbidite formation is dated from the Late Oligocene-Burdigalian, and thus corresponds to a lateral facies of the Numidian. Indeed, it seems that the J. Berkane Numidian klippe (Eastern Rif) is in stratigraphic contact with the underlying Aknoul nappe, suggesting they have been transported altogether. The youngest Intrarif levels at Saf Lahmane are dated from the late Burdigalian-Middle Miocene (Serravallian?), and then resemble the younger pre-nappe levels of the Mesorif zone.

5.4.2 Mesorif

The Mesorif zone displays different characteristics in Western-Central Rif and Eastern Rif, east of the Nekor Fault, respectively.

5.4.2.1 Western and Central Rif

In the central part of the Rif range, the Mesorif zone is also termed *Zone des Fenêtres* ("Window Zone") as it is characterized by antiforms with Lower-Middle Miocene rocks in the core (e.g. Tamda, J. Kouine), and mainly Mesozoic thrust units above them. The allochthonous units have two possible origins, either infra- or supra-Ketama (see Figs. 5.7, 5.8 for location).

The *Tifelouest-Tafraout-Afress-Rhafsaï* group of infra-Ketama units corresponds to folded duplexes at the very front of the Ketama massif, being rooted beneath the massif. The *Senhadja nappe* is also part of the infra-Ketama units, but appears as unrooted klippes on top of the autochthonous Miocene turbidites and olistostrome. The Bou Haddoud nappe has the same infra-Ketama origin as the Senhadja nappe, but extends more to the south over the Prerif units.

The Mesozoic series typical of the Tifelouest group are found also in the Izzarene forest of Western Rif, as well as in the "Zone des Sofs", i.e. the rocky ridges that underline the Mesorif-Prerif boundary. Everywhere within this fragmented domain, rapid facies and thickness changes in the Lower and Middle Jurassic formations indicate the synsedimentary activity of normal faults, and characterize the paleomargin syn-rift evolution (Fig. 5.29). The post-rift sequence begins with the Callovian-Oxfordian "ferrysch", followed by Kimmeridgian and Tithonian-Berriasian micrites, and Neocomian pelites. The syn-orogenic, Cenozoic sedimentation begins with unconformable, blocky marls, probably Late Oligocene in age (M. Durand-Delga, written comm., 2007) as they contain both Nummulites and Lepidocyclines. These nummulitic marls are unconformably overlain by a chaotic complex (Oligocene-Aquitanian?) followed upward by the Aquitanian-Serravallian turbidites cited above. The Mesozoic formations are strongly folded and foliated, whereas the Lower-Middle Miocene formations are moderately deformed and only anchimetamorphic (Sect. 5.5). The occurrence of foliated Eocene pebbles in the Oligocene-Aquitanian chaotic complex yield evidene of a Late Eocene-Oligocene phase of synmetamorphic deformation. In the Zoumi unit of Western Rif, the Miocene formations compare with those of Central Rif, but the older formations are not metamorphic.

The Senhadja "nappe" includes hectometre- to kilometre-sized blocks (Taïneste, Merzouk, Azrou Akchar) on top of the Middle Miocene turbidites. The reworked Mesozoic layers compare with those of the Tifelouest, but they are accompanied by Triassic elements (reddish sandstones and pelites, spilitic dolerites and gabbros) and even by Paleozoic material (quartzites, phyllites). Therefore, the basement itself is clearly involved in the External Rif deformation, likely due to inversion of the paleomargin normal faults.

As for the supra-Ketama units, they correspond mainly to the already quoted *Aknoul nappe*, detached from the Ketama unit on top of the Cenomanian undercompacted clays, and to the J. Berkane *Numidian klippe* transported on top of the latter nappe. Note that the underlying Bou Haddoud unit displays a remarkable, continuous "K-T" section, dated precisely with planktonic Foraminifera (Toufiq et al., 2002).

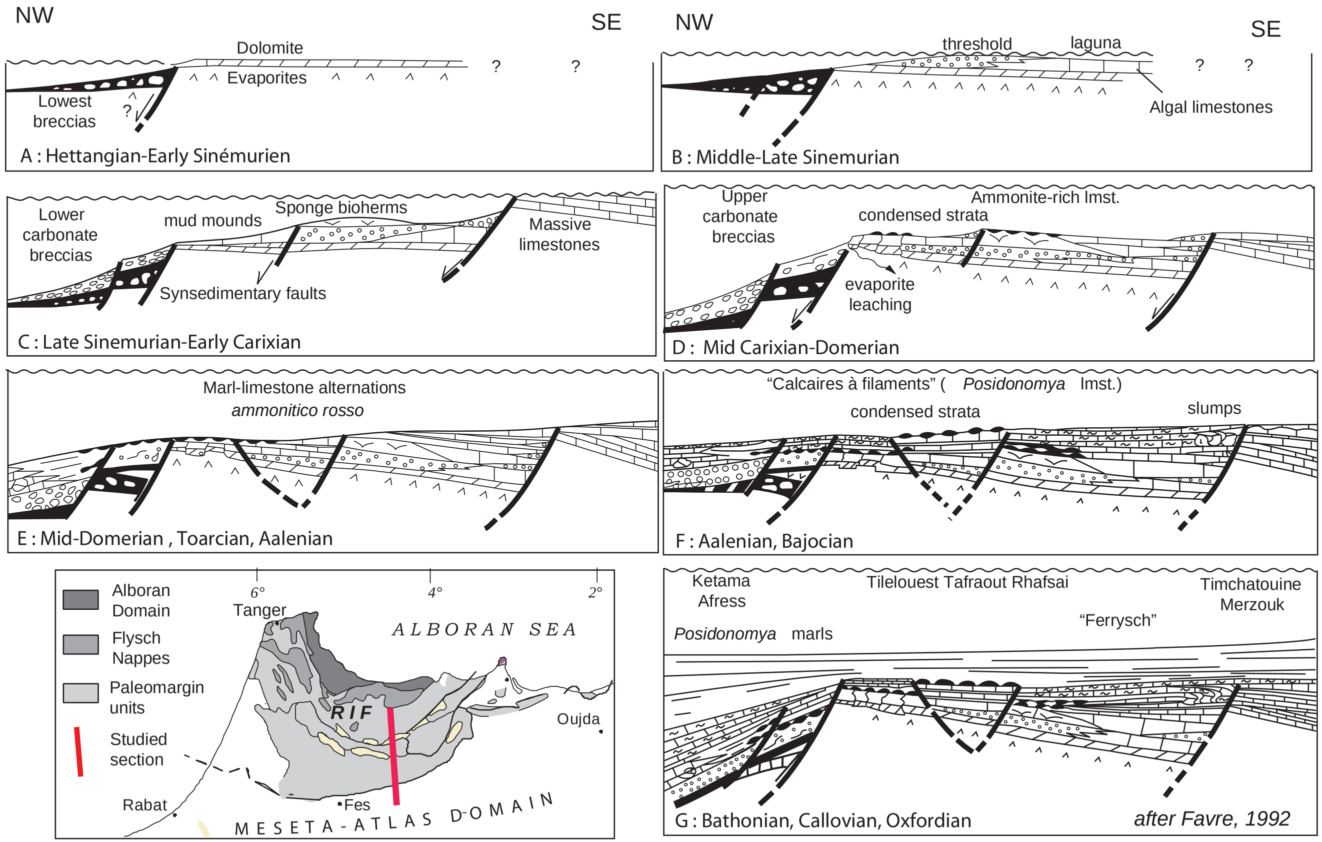


Fig. 5.29 Tectonic and sedimentary evolution of the African paleomargin during the Jurassic at the Mesorif-Intrarif border (Tifelouest and Ketama units), after Favre (1992), modified

5.4.2.2 Eastern Rif

The structural setting of the Central Rif is disturbed eastward by a large tectonic corridor, referred to as the Nekor Fault ("accident du Nekor"), which is a NE-trending left-lateral wrench fault oblique to the general trend of the Rif-Tell belt (Figs. 5.5, 5.7). The importance of this fault is well indicated by its association with a low-velocity zone at 5 km depth, coinciding with high conductivity and low-gravity structure, interpreted as a fault gouge zone and fluid-filled subsurface rock matrix (Serrano et al., 2003). East of the fault, the Mesorif can be easily recognized in the *South Temsamane* sub-zone (unit I, Fig. 5.30). The Mesorif zone can be followed northward to the Mediterranean coast in the *North Temsamane* sub-zone, through several steps of increasing deformation and metamorphic grade (Sect. 5.5.3). The Intrarif zone disappears undersea, and crops out again only in the Oran region east of the Algerian border.

The South Temsamane stratigraphy is complete and easily recognized from the Liassic to Upper Cretaceous levels, up to the unconformable Lower-Middle Miocene turbidites. In contrast, the stratigraphic formations are less continuous and less easily dated in the North Temsamane sub-zone, which consists of more or less diverticulated units (II–VII from S to N), duplicated and folded together during the pre-Miocene synmetamorphic event. Their metasediments are deformed into overturned folds stretched along their WSW-trending axes, which is consistent with a sinistral throw during the Alboran Terrane-Africa oblique collision (Sect. 5.7).

Two isolated massifs involving metamorphic units occur in Eastern Rif, i.e., (i) the Tres-Forcas (Trois-Fourches) massif (unit VIII), whose lowest unit likely continues the North Temsamane system upward; and (ii) the Khebaba unit (unit IX), which overlies the South Temsamane, being located beneath the Aknoul nappe. The Khebaba massif structural setting thus compares with that of the Senhadja klippes, but the Khebaba shows a metamorphic grade comparable with that of the northernmost Temsamane unit (Ras Afraou; Sect. 5.5.3).

In the North Temsamane domain, gabbroic sills intrude the Jurassic-Cretaceous formations, being affected also by the regional syntectonic recrystallization. These metagabbros and metadolerites testify to the importance of the paleomargin thinning in the distal Mesorif area during the Late Jurassic-Early Cretaceous.

Such crustal thinning is still better documented by the occurrence of a serpentinite massif, namely the *Beni Malek massif*, in the Nekor Fault corridor. The massif forms a kilometre scale lense at the bottom of the Ketama unit, and overlies a metasedimentary unit (Igarmaouas unit) intermediate between the Ketama and North Temsamane complexes (Fig. 5.30C). The ultrabasites (altered spinel lherzolite) are partly draped by a cover sequence consisting of limestones with ophiolitic clasts (Late Jurassique-Early Cretaceous?). Laterally, the massive serpentinites are replaced by greenschists, which correspond to metabasites and recrystallized serpentinite-gabbro sands. The Beni Malek unit is interpreted as a sliver of oceanic crust originated from an Alpine-type oceanic crust, where tectonic denudation of mantle rocks plays a major role with respect to magmatism. Modelling of the regional magnetic anomaly suggests that the serpentinite and greenschist sliver

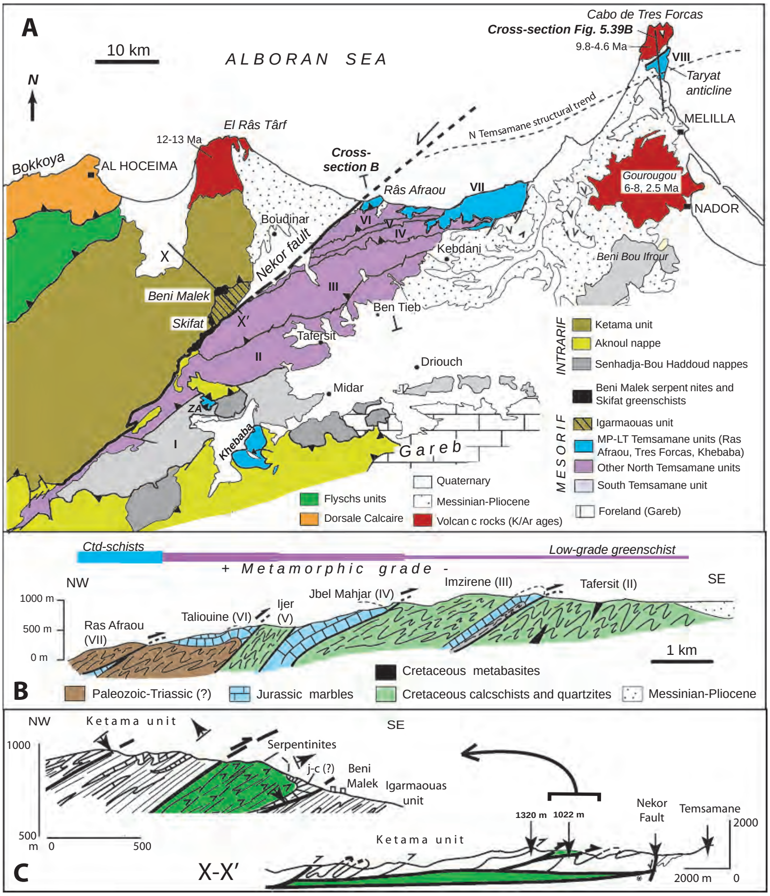


Fig. 5.30 Structure and metamorphism of the Eastern Rif. Schematic structural map (**A**) and cross-section (**B**) after Frizon de Lamotte (1985), Negro et al. (2007), and Michard et al. (2007), modified. Khebaba massif after Darraz & Leblanc (1989). K/Ar ages (Ma) of post-orogenic volcanoes after El Azzouzi et al. (1999) and Münch et al. (2001). XX': trace of lower cross-section (**C**); ZA: Zaouyet Sidi Hadj Ali. – (**C**): Geological cross-section of the Beni Malek ultrabasite massif (YY') after Michard et al. (1992), and 2D interpretation of the regional aeromagnetic anomaly (XX') suggesting the occurrence of a cryptic ultrabasic body, after Elazzab et al. (1997). j-c(?): carbonates with serpentinite clasts (Upper Jurassic-Lower Cretaceous?)

is derived from a deeper and larger serpentinite body obducted onto the Mesorif units (Fig. 5.30C). This would indicate that a minor, intracontinental (intra-margin) suture zone occurs between an Intrarif distal block and a Mesorif-Prerif proximal margin. Serpentinite outcrops are known on top of the Tres-Forcas massif. In fact, the trace of this suture can be followed eastward at least up to the Oran region, and likely up to the Chéliff Mountains south of Algiers. To the west, the suture is likely buried beneath the Intrarif thrust, whose lateral ramp corresponds to the Nekor Fault.

5.4.3 Prerif Zone and Foredeep

The main, *external Prerif Zone* comprises Upper Cretaceous-Eocene and Middle-Upper Miocene marly-argillaceous formations detached from their Jurassic-Early Cretaceous basement and slid towards the foredeep (Figs. 5.7, 5.8). The Triassic clay and salt complex formed diapirs and rock glaciers within the Cretaceous formations. The diapiric complex frequently includes ophite bodies of Late Triassic-Early Liassic age. Subsequently, the Triassic diapirs and the resedimented Triassic complex are incorporated within the Prerif nappes (Figs. 5.31, 5.32).

The Jurassic-Lower Cretaceous formations themselves are detached from their Paleozoic basement and now form an alignement of carbonate slivers in the internal Prerif Zone, also labelled the “sof line”. The “sof” Mesozoic series compares with the Mesorif as they exhibit Upper Jurassic “ferrysch” and Lower Cretaceous marly limestones with turbidite intercalations. The Upper Cretaceous-Eocene marls are followed upward by Middle-Upper Eocene to Middle Miocene detrital formations including, in particular, the Eocene-Oligocene Sidi Mrayt sandstones, which were fed from the external foreland through the Meliana paleo-canyon. The latter stratigraphic levels lack in the External Prerif, suggesting a flexural uplift of the proximal margin south of the Mesorif, which played the role of an early foredeep during the Early-Middle Miocene.

The *foredeep* subsidence began during the Middle-Late Miocene at the front of the Rif-Tell tectonic prism, through flexural bending of the African plate due to tectonic overburden, and/or to slab pull from the subducting Maghrebian Ocean lithosphere (Sect. 5.7). In the central Fes-Meknes region, the *Rides prérifaines* are late anticlines formed after the synsedimentary emplacement of the Prerif nappe within the foredeep. Their varied axial trends (E-W, NE-SW and N-S) are controlled by the varied normal faults inherited from the paleomargin. The Mesozoic series of the Rides prérifaines compares with that of the Middle Atlas foreland (sandy carbonates of Dogger age, Late Jurassic-Early Cretaceous gap). It is unconformably overlain by Middle-Upper Miocene molassic sandstones, then by Upper Miocene sandy marls within which the front of the nappe is interleaved (Fig. 5.33; see also Chap. 6).

Further west in the Gharb Basin, industrial boreholes evidence the Miocene molasse onlapping onto the Caledonian-Variscan Sehoul Block barely covered by thin Cretaceous deposits. Near Rabat, the molasse directly overlies the Paleozoic basement. The lack of Jurassic formations and the reduction of Cretaceous sediments can be ascribed to the uplift of the Atlantic rift shoulder (Chap. 4). The molasse transgression onto the foreland occurred earlier in the eastern areas (Langhian) than in the western (Tortonian). Subsequently, sandy marls accumulated in the foredeep during the Tortonian. At that time, the Prerif formations, already overlain by the Intrarif nappes (Ouezzane, Tsoul, Aknoul) detached and slid on top of the foredeep sediments under submarine, synsedimentary conditions (olistostromes). The cover of the nappe consists again of terrigeneous muds of Late Tortonian-Messinian age, labelled the *Miocène post-nappe* and followed upward by similar deposits of Pliocene age. The regression occurred after the middle Pliocene (Saiss “Sables fauves”). The collapse of the orogenic prism resulted in the transgression of the post-nappe

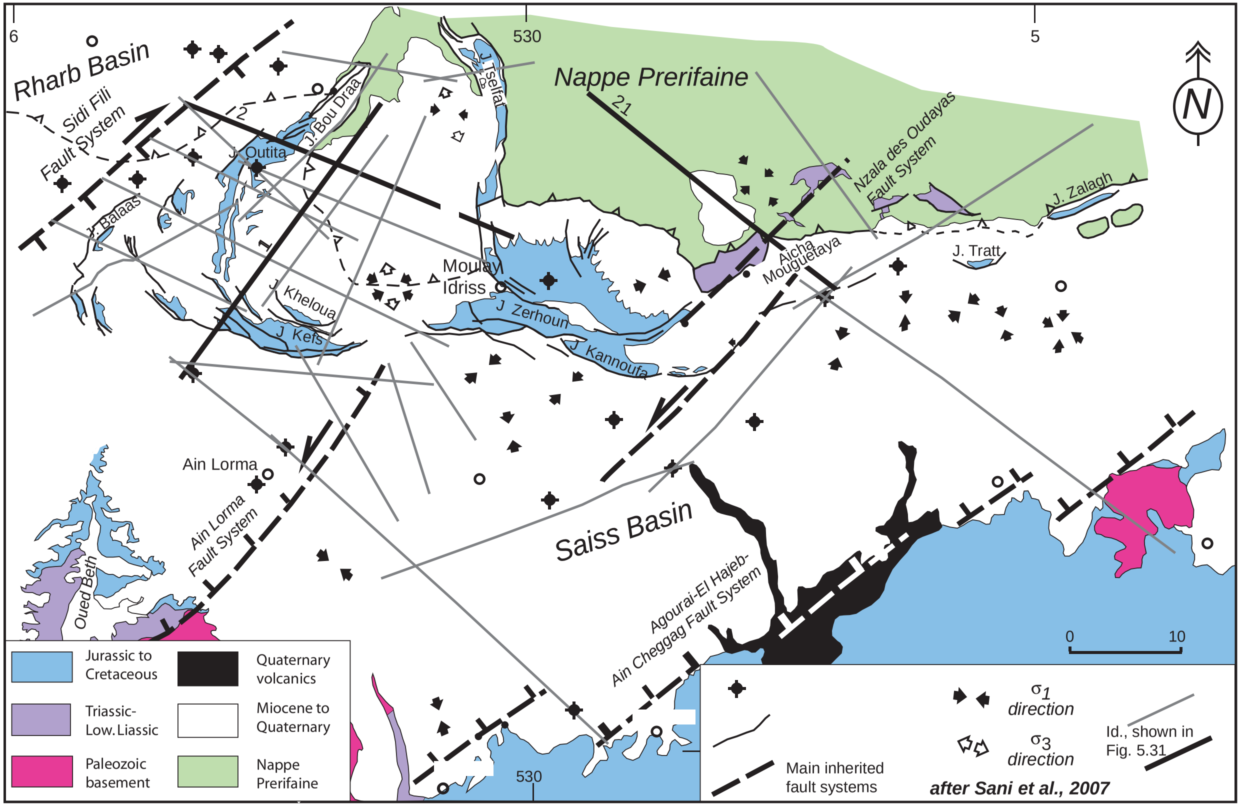


Fig. 5.31 Structural map of the Prerif front in the area of the Rides Périfaines, with location of the ONHYM seismic profiles and industrial wells, after Sani et al. (2007), modified. Note the importance of the fault systems inherited from the Variscan evolution of the basement and active during the Triassic-Jurassic evolution. These faults heavily control the Neogene deformation

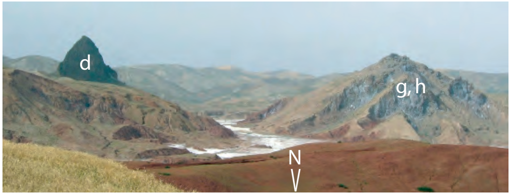


Fig. 5.32 The Late Triassic complex of the External Prerif, 25 km NW of Fes city. The complex includes red-brown clays, evaporites (*g gypsum; h halite*), and a neck of spilitic dolerites (*d*). The leached salt turns the bed of oued Mellah white ("mellah"=salted, in Arabic). In the background, Cretaceous and Miocene marls

deposits onto the Mesorif domain of Central-Eastern Rif. These deposits are now preserved in post-nappe synclines, which formed contemporaneously with the Rides prérifaines anticlines. The Rif foredeep connects offshore with the Betic one in the submarine accretionary prism of the Gulf of Cadiz (Fig. 5.5).

5.5 Metamorphism

In the Gibraltar Arc, metamorphism characterizes essentially the Internal Zones (Alboran Domain), and more particularly the Ghomarides-Malaguides and Sebtides-Alpujarrides. In the Dorsale units, very low-grade, post-Eocene metamorphism occurs only along the internal border of the range and along the Jebha Fault. In the two former complexes, as well as in their Kabylian equivalents, one can clearly separate, (i) Variscan metamorphism, which is the main recrystallization episode in the Ghomarides-Malaguides and the Kabylian upper units (i.e. the "upper plate" of the broad metamorphic structure), and (ii) Alpine metamorphism, which hardly affects the bottom of the upper plate, but deeply concerns the lower (Kabylian lower units, Sebtides-Alpujarrides, and Nevado-Filabrides in Central-Eastern Betics). As a whole, these upper and lower plate units define a dismembred metamorphic core complex, which records a tectonic evolution involving, initially, the subduction of the lower plate, and subsequently its exhumation (Sect. 5.7).

Metamorphic recrystallization also affects part of the External Zones in the Central and Eastern Rif, and their Tellian equivalents. Although the metamorphic grade is moderate or low in these units, it is a useful tool to delineate the intracontinental suture zone, which is also marked by the Beni Malek massif of serpentinites and greenschists.

5.5.1 Ghomaride Metamorphism

References: A limited number of works address the metamorphism of the Ghomaride, e.g. Chalouan & Michard (1990), Michard et al. (2006), Negro et al.

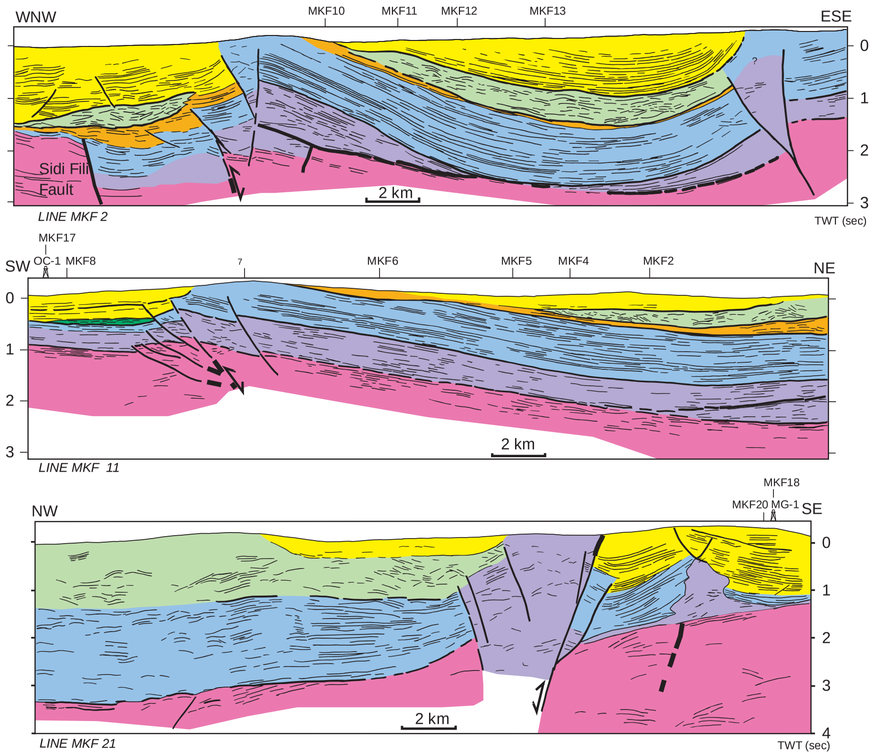


Fig. 5.33 Interpreted seismic profiles across the Rides prérifaines, after Sani et al. (2007), modified. See Fig. 5.31 for location; (**A**): Profile MKF2; (**B**): Profile MKF11; (**C**): Profile MKF21. The colours indicate the different units as follows. *Light red*: undefined Paleozoic; *violet*: Triassic units, evaporites; *light blue*: Jurassic units (Lower-Middle Liassic, Aalenian-Bajocian); *light brown*: Lower-Middle Miocene foredeep units (molasses) below the Prerif nappe; *light green*: Prerif nappe complex; *yellow*: Upper Miocene-Pliocene foredeep deposits. Note the crossed orientation of profiles (**A** and **B**), which corresponds to the curved trend of the outcropping anticlinal ridges (Fig. 5.31). Profile (**C**) illustrates a Triassic diapir first emplaced during the Cretaceous and then pinched and reactivated during the Plio-Quaternary shortening event (cf. Roca et al., 2006)

(2006). A very low grade recrystallization of the overlying Dorsale units was described locally by Olivier et al. (1992). In the Betics, the Malaguide metamorphism has been discussed by Platt et al. (2003b) and Negro et al. (2006). A review of the Kabylian upper plate metamorphism is given in Michard et al. (2006).

The Ghomaride Mesozoic-Cenozoic series are virtually devoid of metamorphic recrystallization. Contrastingly, two superimposed metamorphic histories can be distinguished in the Paleozoic terranes, Variscan and Alpine, respectively, although the latter concerns only the lower part of the lowest nappe (Aakaili).

The Variscan evolution itself is divided into two episodes (Sect. 5.2.2.1). The first affects the Lower and Middle Paleozoic series, but not the Carboniferous. This Eo-variscan episode is characterized by low grade greenschist-facies recrystallizations coeval with the formation of superimposed, NE-trending folds (P1 and P2 folds in Fig. 5.16). The second, Hercynian-Alleghanian episode affects the whole Paleozoic pile, being mainly identified in the Carboniferous formations by very low grade recrystallizations and NW-trending folds. The overlying, unfolded Triassic deposits show only a diagenetic evolution.

However, K-Ar datings of Paleozoic samples from the Ghomaride nappes fail to yield any exact Variscan date. In contrast, they reveal a progressive variation of the measured ages from 259 ± 5 Ma in the uppermost nappe to 25 Ma at the bottom of the lowest, just in contact with the Sebtides schists (Fig. 5.34). This suggests that a heating event affected the base of the Ghomaride pile, contemporaneously with the underlying Sebtides, i.e. at about 23 Ma (Sect. 5.5.2). This thermal event has been calibrated recently by Raman spectroscopy of carbonaceous material: T reached $\sim 500\,^{\circ}$C at the bottom of the Aakaili nappe whereas it remained below $\sim 300\,^{\circ}$C in the upper part of the nappe and overlying units (Fig. 5.34). Similar observations were made in the Malaguides. This Oligocene-Miocene thermal event caused biotite and andalusite growth at the very bottom of the Ghomaride-Malaguide complex, and can be ascribed to the major extension phase responsible for the opening of the Alboran Basin (Sect. 5.7).

5.5.2 Sebtide Metamorphism

References: Many recent publications concern the Sebtide-Alpujarride metamorphism. To concentrate on the Sebtide metamorphism, we must refer to the following papers, which include many Betic references: (i) for the Lower Sebtides (Beni Bousera and Filali), Saddiqi et al. (1988), Kornprobst et al. (1990), Saddiqi (1995), Kumar et al. (1996), Bouybaouene et al. (1998), Azañón et al. (1998), El Maz et Guiraud (2001), Haissen et al. (2004), Negro et al. (2006); (ii) for the Upper Sebtides (mainly Beni Mezala), Bouybaouene (1993), Zaghloul (1994), Goffé et al. (1996), Michard et al. (1997), Vidal et al. (1999), Agard et al. (1999), Negro (2005), Janots et al. (2006), Michard et al. (2006), Negro et al. (2006).

The age of the Sebtide-Alpujarride metamorphism has been repeatedly investigated. Let us particularly refer to Monié et al., 1991, 1994), Sosson et al. (1998),

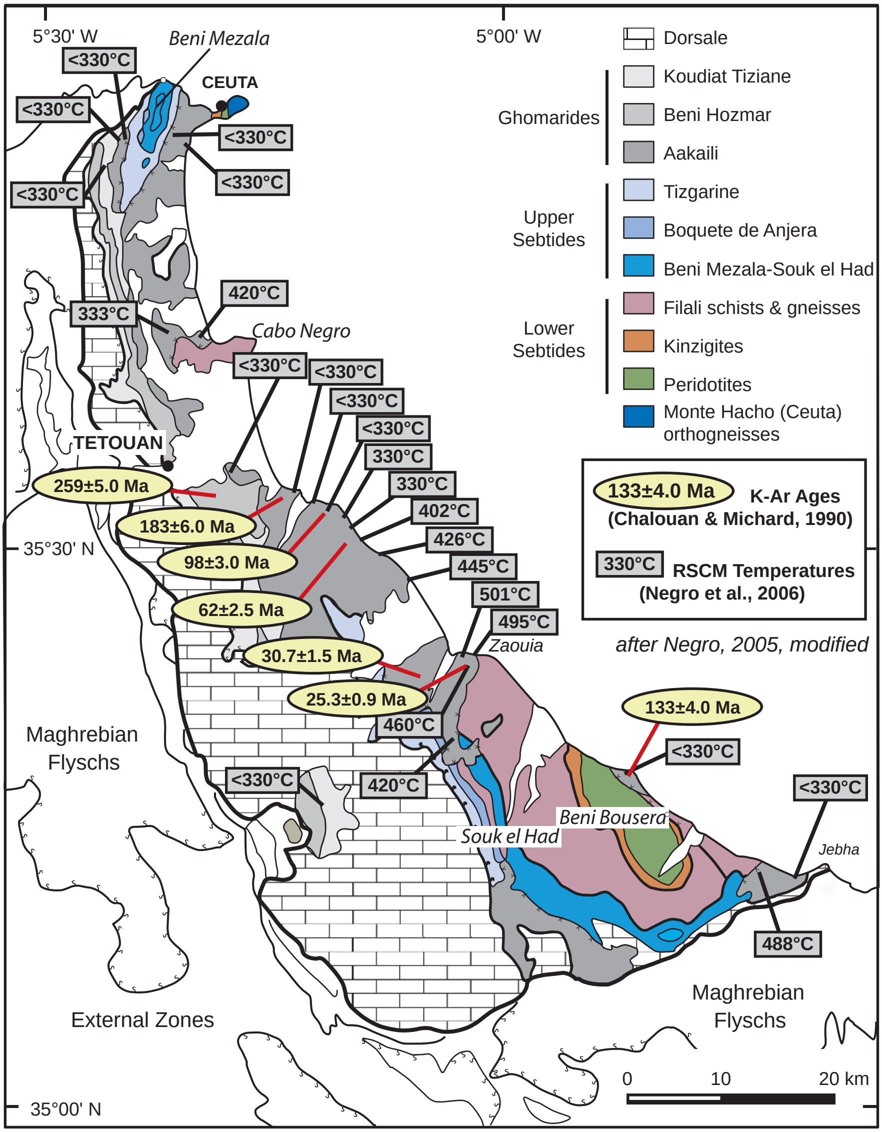


Fig. 5.34 The Alpine thermal event in the Ghomaride nappes: progressive variation of the K-Ar white mica ages (after Chalouan & Michard, 1990) and maximum T recorded in the Paleozoic rocks (RSCM method; Negro et al., 2006)

Argles et al. (1999), Platt & Whitehouse (1999), Blichert-Toft et al. (1999), Sánchez-Gómez et al. (1999), Zeck et Whitehouse (1999), Montel et al. (2000), Sánchez-Rodriguez & Gebauer (2000), Esteban et al. (2004), Platt et al. (2003b, 2005), Janots et al. (2006), Platt et al. (2006). A review of the Kabylian dates can be found in Michard et al. (2006).

In the Sebtides, Alpujarrides and equivalent Kabylian units, the most conspicuous metamorphic record results from the Alpine evolution, and the previous evolution of the corresponding crustal and mantle rocks is seldom recorded, because of the high grade of the Alpine metamorphism.

5.5.2.1 Alpine Recrystallizations in Northern Upper Sebtides

The Federico units of the Beni Mezala post-nappe antiform (Figs. 5.9, 5.14, 5.35A) offer the best opportunity to calibrate the Alpine metamorphism as most of its rock material is Permian and Triassic, and consists mostly of metapelites (Sect. 5.2.1.2). In the uppermost unit (Tizgarine), the cookeite-pyrophyllite-low Si phengite assemblages correspond to low pressure, low temperature conditions (Fig. 5.35C). In the underlying, Boquete Anjera unit, sudoite, Mg-chlorite and phengite occur in the quartz veins whereas chloritoid is abundant in the matrix; the corresponding metamorphic conditions can be estimated at ~7 kbar, 300–380 °C, i.e. intermediate pressure and temperature. Deeper in the antiform, the upper Beni Mezala unit (BM2) displays Mg-carpholite relics within chloritoid or kyanite-bearing intrafolial quartz veins. This corresponds to blueschist facies (HP-LT) conditions, between 8 kbar, 380 °C and 13 kbar, 450 °C. Eventually, in the lower Beni Mezala unit (BM1), Mg-carpholite relics associated with talc-phengite assemblages within quartz-kyanite segregations indicate that eclogite facies conditions have been reached, between 13–15 kbar, 450 °C, and 15–18 kbar, 550 °C. The underlying Benzu schists compare with the Filali recrystallized Paleozoic basement, and show garnet-chloritoid-phengite-quartz assemblages equilibrated at $P > 14$ kbar, $T \sim 550$ °C.

The earliest retrograde path is constrained in BM1 by tremolite-talc and phlogopite-chlorite associations, and appears virtually isothermal (Fig. 5.35C). Subsequently, further exhumation occurs at decreasing T, being accompanied by paragonite, muscovite, chlorite, kaolinite, cookeite and margarite growth. Microstructural observations indicate that these crystallizations were contemporaneous with a ductile, top-to-the-north extensional shearing (Fig. 5.35A).

The reported data indicate that the Sebtide units were buried by subduction (HP-LT conditions) down to 60 km at least for some of them (eclogite facies), then tectonically exhumed, forming a pile of tectonic units separated by subtractive contacts (P and T gaps). The corresponding geodynamic setting is discussed hereafter (Sect. 5.7).

5.5.2.2 Alpine Recrystallizations in the Southern Upper Sebtides and Lower Sebtides

The Federico units on top of the Beni Bousera antiform (Fig. 5.9) show Alpine mineral assemblages similar to those of their northern equivalents (Beni Mezala), except for the lowest, Souk-el-Had unit. The latter unit contains phengite-chlorite-phlogopite-kyanite assemblages, and late growth of andalusite and cordierite, which

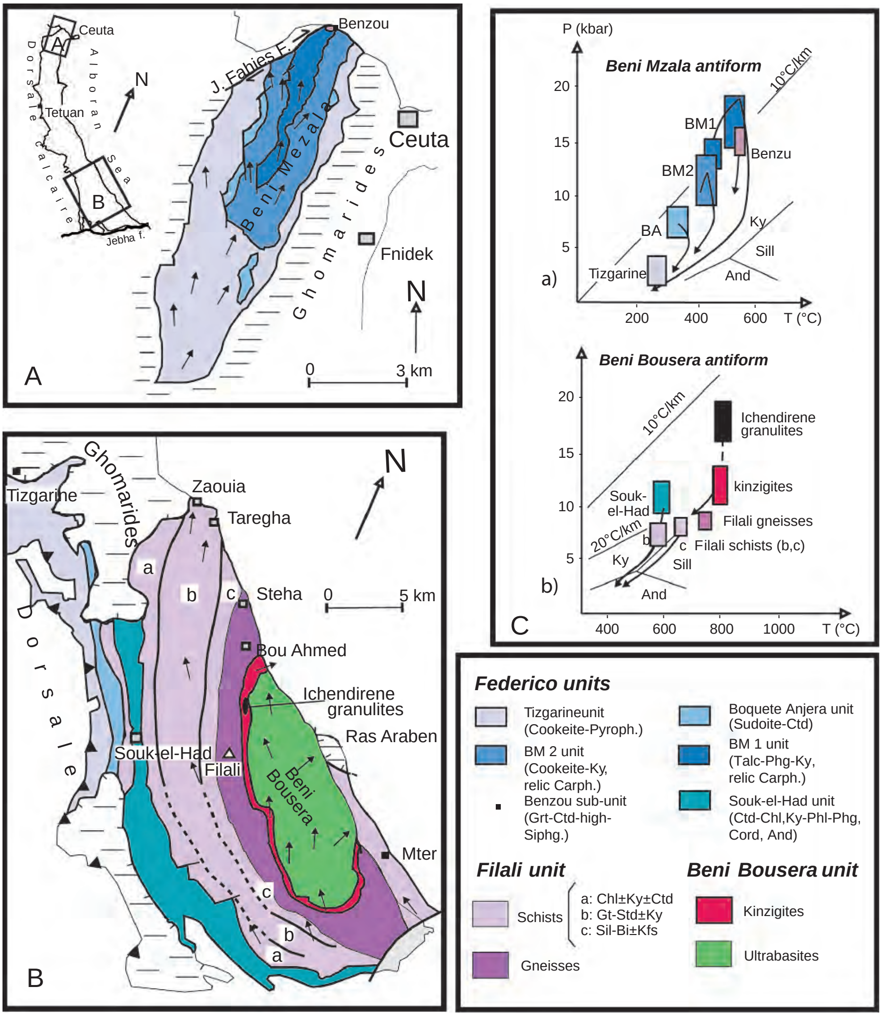


Fig. 5.35 Metamorphic structure and P-T conditions of the Sebtide units, after Bouybaouene (1993), Saddiqi (1995), Michard et al. (1997, 2006), and Negro et al. (2006). (**A**): Beni Mezala antiform. – (**B**): Beni Bousera antiform; a–c: Filali schist metamorphic zones. – (**C**): Corresponding P-T estimates. *Arrows*: sense of shear inferred in late metamorphic structures

suggests an evolution at 12 kbar, 550–600 °C (Figs. 5.35B, C), i.e. under a higher geotherm than the northern unit.

As for the underlying Filali mica-schists (Fig. 5.35B), their upper part displays chlorite-chloritoid-muscovite ± biotite ± kyanite assemblages (zone “a”), changing downward into garnet-biotite-staurolite-kyanite (zone “b”), and finally, garnet-biotite-sillimanite assemblages (zone “c”). Andalusite is almost ubiquitous, and shows syn- to post-kinematic character (Fig 5.36A). The coexistence of the three

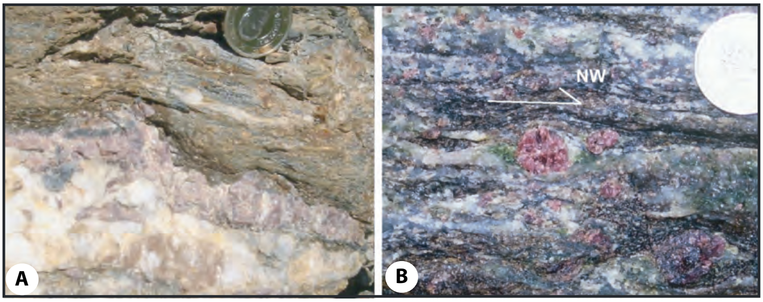


Fig. 5.36 High grade rocks from the Lower Sebtides (see Fig. 5.35 for location). (**A)**: Quartz segregation with pink andalusite crystals, zone (b) of the Filali schist unit south of Taregha. – (**B)**: Kinzigites from the Beni Bousera unit top slivers east of Bou Ahmed, with large garnet crystals whose asymmetric quartz-phengite-green spinel pressure shadows indicate top-to-the-NW, non-coaxial shear strain. The mylonitic matrix shows quartz-retromorphic kyanite-sillimanite ribbons and biotite-graphite foliae. Coin diameter: 15 mm. Photos by O. Saddiqi

Al-silicate polymorphs prooves the complexity of the metamorphic history, where kyanite growth seems to have preceded that of the high-temperature polymorphs. The assumed metamorphic conditions range from 7 kbar, 580 °C in zone "b" to 8 kbar, 680 °C in zone "c", and finally reach 8 kbar, 780 °C in the gneisses.

A dramatic pressure gap (which implies a tectonic omission) occurs between the Filali and Beni Bousera units. The kinzigite (or granulite) garnet-sillimanite ± kyanite-graphite assemblages characterize P-T conditions at 9–13 kbar, 800–850 °C (Figs. 5.35C, 5.36B). However, the Ichendirene metabasite lense contains a primary assemblage pyrope-jadeite rich pyroxene-kyanite-rutile-plagioclase-quartz, which corresponds to peak metamorphic conditions at 16–20 kbar, 760–820 °C (Fig. 5.35C). The peak conditions in the underlying spinel-garnet harzburgites are similar, 18–20 kbar, 850–900 °C. In contrast, within the peridotite massif itself, still higher P-T conditions are recorded by the garnet-corindon-bearing pyroxenites, $P > 20$ kbar, 1200–1350 °C.

Thus, the P-T conditions recorded in the Lower Sebtides typifie a higher geotherm than those from the Upper Sebtides. The reason for this contrast probably arises from the different location of these complexes within the subduction zone (Sect. 5.7), together with the different nature of the dominant rock material (Permian-Mesozoic sediments *versus* crustal basement and mantle rocks).

5.5.2.3 Dating the Sebtide-Alpujarride Metamorphism

The occurrence of Sebtide-Alpujarride pebbles and minerals reworked in the earliest Burdigalian detrital formations from Andalucia, and in the early-middle Burdigalian formations from the Rif (Sect. 5.2.2.3) testifies that part of the Sebtide-Alpujarride were already exhumed up to the surface at that time (ca. 20–18 Ma). Previously, during sedimentation of the earliest post-nappe formations (Fnideq, Alozaina) upon the Ghomarides-Malaguides during the latest Oligocene-Aquitanian (23–20 Ma), the

Sebtides-Alpujarrides (which belong to the lower plate) were totally hidden beneath the Ghomaride-Malaguide upper plate. In other words, the metamorphic units have been exhumed during the late Oligocene-Burdigalian interval (23–18 Ma), which corresponds to the beginning of the Alboran Sea opening (Sect. 5.7).

Recent isotopic datings are consistent with these stratigraphic constraints; they also help to precise the age of the thermal peak, but fail to attain with certainty that of the pressure peak itself. In the southern Sebtides (Fig. 5.37), most of the results concentrate in the range 30–25 Ma to 18 Ma, and more particularly between 23 and 20 Ma, whatever the isotopic method. The only exceptions are a 66 Ma Lu-Hf age, probably devoid of geologic meaning, and the ages close to 300 Ma which correspond to Variscan relic minerals. This is a good indication that the thermal peak (T > 600 °C, admittedly the closure temperature for U-Pb zircon method) occurred at about 25–23 Ma. Subsequently, a rapid exhumation down to T < 300 °C took place at ∼20 Ma (K-Ar biotite dates).

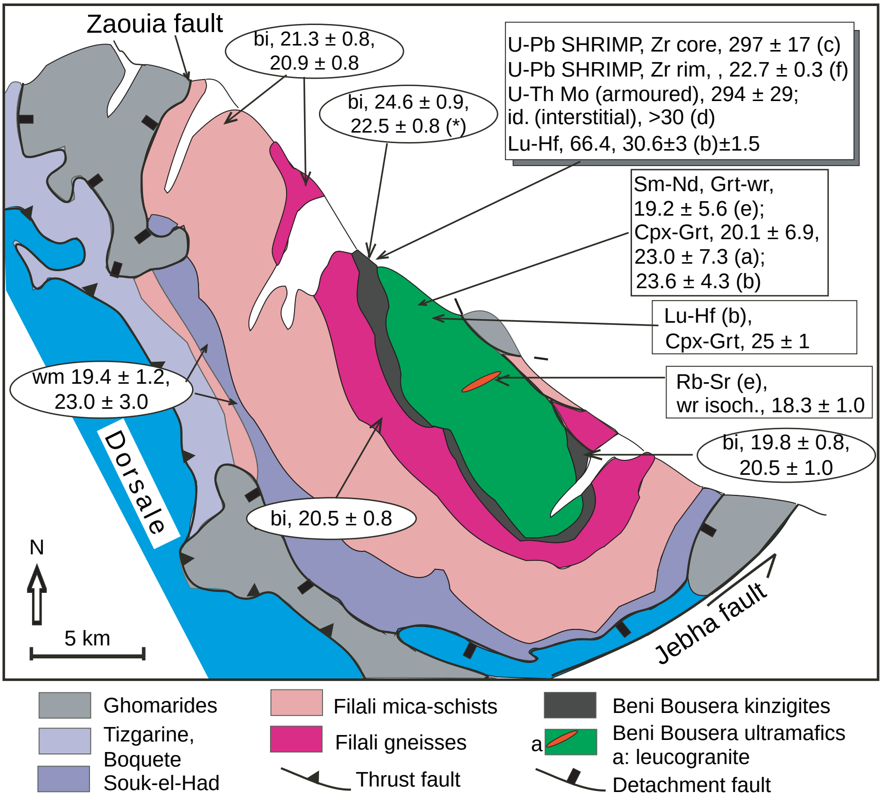


Fig. 5.37 Isotopic datings of the Sebtide units from the Beni Bousera antiform. The ellipses show the K-Ar and ^{39}Ar-^{40}Ar (*) results obtained on biotite (bi) and white micas (wm) by R. Montigny, analyst, in Michard et al. (1991). Other dates (framed in rectangles): (a) Kumar et al. (1996); (b) Blichert-Toft et al. (1999); (c): Sánchez-Rodriguez & Gebauer (2000); (d): Montel et al. (2000); (e) : Polvé (1983); (f): Platt et al. (2003a). Mo (arm)/(interst): armoured/interstitial monazite

Dates from the Alpujarrides are fairly consistent with the Sebtides ones. They show a striking concentration between 22 and 19 Ma, besides of some Variscan ages, already quoted above (Sect. 5.2.1.1). Partial melting occurred in the Ojen gneisses beneath the Ronda peridotites at about 18–20 Ma according to the age of cordierite-bearing leucogranite dykes in the peridotites.

The age of the peak of pressure was precised via ^{39}Ar-^{40}Ar dating of the phengites from the low grade units (i.e. where the measured age equal the crystallization age). A minimum age of ca. 25 Ma was obtained in Central Alpujarrides, and two minimum ages in the Beni Mezala, ca. 23 and 27 Ma. A much higher figure, 48 Ma, was ultimately obtained from Eastern Alpujarrides, together with younger dates up to 20 Ma. According to Platt et al. (2005), the Alpujarride burial would have begun during the Early Eocene. However, the 48 Ma date possibly corresponds to excess argon, as it seems likely that the pressure peak should be close to the thermal peak, i.e. at ca. 30–25 Ma (Michard et al., 2006). We recall that the Malaguides thrust are dated stratigraphically at about 28 Ma ± 1 Ma (Sect. 5.2.2.3).

5.5.3 External Zone Metamorphism

References: Metamorphism is very limited in the External Zones, and the number of related publications alike: Monié et al. (1984), Frizon de Lamotte (1985), Leikine et al. (1991), Favre (1992), Michard et al. (1992), Azdimousa et al. (1998), Negro et al. (2007), Michard et al. (2007). The equivalent zones in the Algerian Tell are described by Guardia (1975), and Kirèche (1993).

In the External Rif, metamorphism affects only parts of the central and eastern regions, namely the Ketama unit (Intrarif), the Tifelouest group of units (internal Mesorif), and the North Temsamane units on the east side of the Nekor Fault (Figs. 5.7, 5.30A). Additionally, two isolated massifs, i.e. the Tres Forcas and Khebaba massifs are related to the North Temsamane metamorphic zone.

5.5.3.1 West of the Nekor Fault

The Ketama rock materials are recrystallized under low grade greenschist facies conditions, with temperature seemingly lower in the north (~200–250 °C) than in the south (~300 °C) under pressure close to ~3 kbar. Recrystallization was coeval with S- or SE-vergent recumbent or overturned folds (Fig. 5.38A), refolded by upright folds. Tentative K-Ar datings did not yield reliable results due to the low metamorphic grade and abundance of clastic muscovite grains. In contrast, apatite fission track analysis indicates that the post-metamorphic cooling down to ca. 100 °C occurred at about 14–15 Ma.

A closely similar tectonic-metamorphic evolution can be observed, and dated stratigraphically in the Tifelouest units where syntectonic recrystallization affects the series up to the Late Oligocene blocky marls inclusively (cf. Sect. 5.4.2.1).

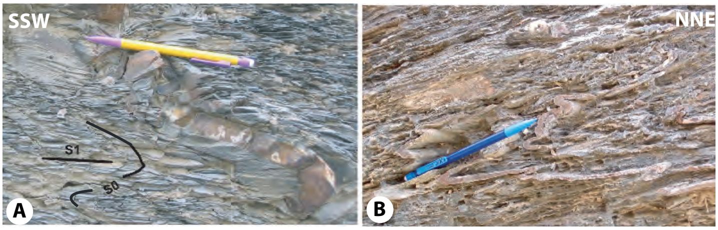


Fig. 5.38 Synmetamorphic structures from the External Rif. (**A**): Slaty cleavage S_1 and minor folds in the inverted limb of a major fold; Aptian-Albian upper part of the Ketama unit, 33 km north of Taounate (see Fig. 5.7 for location). – (**B**): Sub-horizontal foliation and associated recumbent minor folds in the eastern part of the Ras Afraou unit (north of Kebdani; see Fig. 5.30 for location). The rock material includes metapelites, quartzites and rare carbonate layers, probably Paleozoic in age

The metasedimentary duplexes are unconformably overlain by unmetamorphosed mélange with foliated Mesozoic elements, followed upward by the Lower-Middle Miocene turbidites. Therefore, the metamorphism of the Tifelouest and (probably) Ketama units probably occurred during the Late Oligocene (~28–23 Ma). Then, the tectonic prism thickened and migrated southward onto the Mesorif. As a result, upright folds with axial-plane cleavage developed in the Mesorif units (crenulation cleavage in the foliated Tifelouest material, and spaced pressure solution cleavage in the Lower-Middle Miocene deposits). This very low grade metamorphic event took place during the Serravallian-early Tortonian interval, before the transgression of the late Tortonian-Messinian post-nappe sediments, and probably resulted of the collision of the Alboran block against Africa (Sect. 5.7). In contrast, the earlier, Late Oligocene metamorphism depends on another tectonic event, which affected dominantly the regions east of the Nekor Fault.

5.5.3.2 East of the Nekor Fault

The South Temsamane units only display an anchimetamorphic evolution as in the Central Rif windows. In contrast, the North Temsamane units exhibit greenschist facies recrystallizations the grade of which increases upward in the tectonic pile, indicating a post-metamorphic stacking event. Usual mineral assemblages are chlorite-phengite-quartz-albite in metapelites, and tremolite-epidote-albite-chlorite-sphene in metabasites (Unit VI). Chloritoid appears in the highest, Ras Afraou unit (Unit VII), in association with Si-rich phengite. There, peak P-T conditions are estimated at 7–8 kbar, $350 \pm 30\,°C$, which corresponds to medium pressure, low temperature (MP-LT) metamorphism. The coeval, ductile structures include S-vergent overturned folds and gently dipping foliation (Fig. 5.38B) associated with strong SW- to SSW-trending stretching lineation and top-to-the-WSW shear indicators.

The Tres-Forcas massif (Taryat anticline) crops out beneath the late Tortonian-Messinian deposits (Fig. 5.39B, C). The massif is cored by likely Paleozoic

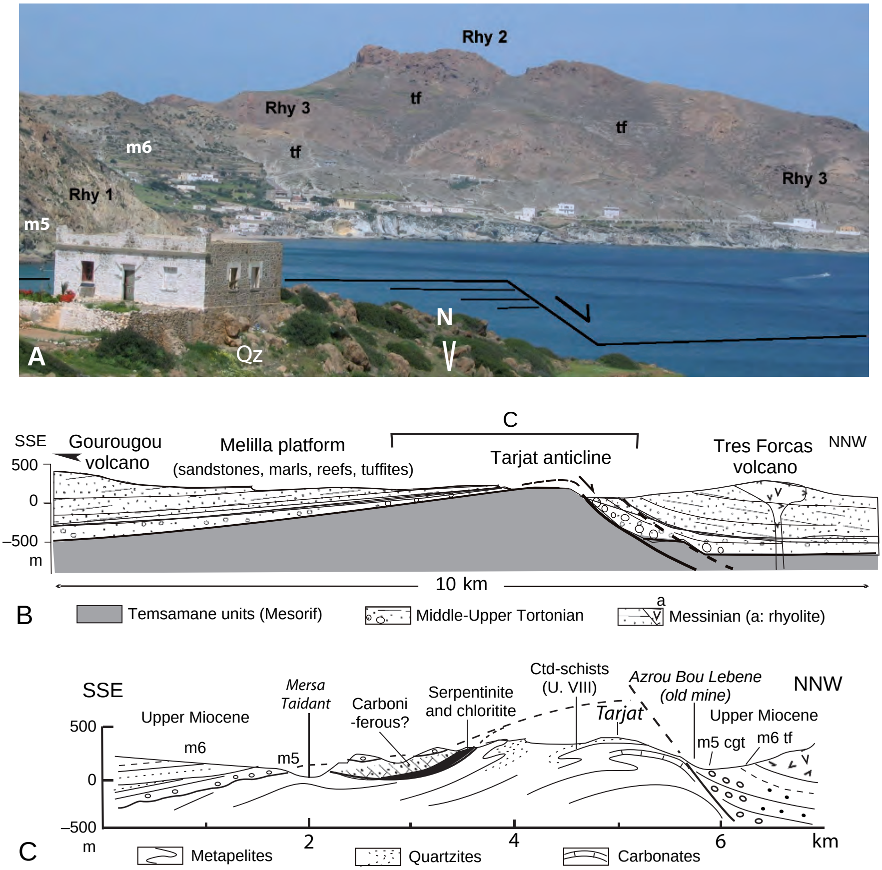


Fig. 5.39 The Cape Tres-Forcas volcano and its tectonic setting (see Fig. 5.30 for location). (**A**): The volcanic massif viewed from the ancient Azrou n'Bou Lebene mine. Qz: cataclastic quartzite layer from the top of the metamorphic "Temsamane VIII" unit; m5: Middle-upper Tortonian; m6: Messinian; tf: volcanic tufs and breccias, lahars; Rhy: rhyolites (2: with amphibole-biotite; 3: with pyroxene). – (**B**): Cross-section from the Tres-Forcas volcanic-sedimentary complex (Upper Tortonian-Messinian; age of rhyolites: 9.8–4.6 Ma) to the Melilla platform where the youngest lava flows (6.4–5.7 Ma), originated from the Gourougou strato-volcano, are interbedded. Structural data after the geologic map of Morocco, sheet Melilla (1983), and personal observations. K/Ar and stratigraphic dates after Hernandez et al. (1987), Cornée et al. (1996), and Münch et al. (2001). – (**C**): Cross-section of the Taryat anticline (Temsamane Unit VIII), after Michard et al. (2007)

quartz-phyllites, quartzites and marble recrystallized under P-T conditions hardly higher than those of the Ras Afraou unit (~8 kbar, 400 ± 30 °C). This "Unit VIII" of the Temsamane system is overlain by slivers of serpentinites, chloritites and jaspes, homologous of the Beni Malek unit, and finally topped by pelites and sandstones of probable Carboniferous age.

The Khebaba-Zaouyet Sidi Hadj Ali unit consists of dismembered terranes including Paleozoic rocks (Devonian metapelites, Early Carboniferous flysch) and their former cover series (Permian-Triassic red beds, dolomites, marbles, and gypsum). These rocks are recrystallized under MP-LT conditions similar to those of the Tres-Forcas metamorphic unit. The structural position of the Khebaba massif compares with that of the Senhadja klippes as the massif is topped by the Aknoul nappe and lies over the south Mesorif.

Preliminary ^{40}Ar-^{39}Ar phengite datings (Monié et al., 1984) yielded an age of 28.6 ± 1 Ma for the peak metamorphism of Unit V, with a low grade stage at 8 Ma. Negro (2005) obtained three groups of ^{40}Ar-^{39}Ar results from the Ras Afraou and Tres Forcas phyllites. A group of minimum ages at 23–20 Ma characterizes the high-Si phengite grains preserved in the intrafolial quartz segregates; it is referred to the peak pressure metamorphism, whose age would be close to 28–23 Ma. A second group of results at 15–10 Ma characterizes the phengite lamellae from the foliation; it is referred to the ductile deformation associated to the southwestward exhumation of the metamorphic units (Negro et al., 2007). Eventually, the dates of 10–6 Ma obtained from the illite-kaolinite-bearing retromorphic samples may represent late brittle-ductile deformation.

5.5.3.3 Interpretation: The External Maghrebide Suture Zone

As the metamorphic grade and reddish colour of some of the Ras Afraou and Khebaba metapelite outcrops evoke those of the Upper Sebtide Permian-Triassic levels in the Boquete Anjera unit, certain authors assumed that the Ras Afraou, Tres-Forcas and Khebaba units originated possibly from the Sebtide domain, having been thrust above the Dorsale and Flyschs domains (e.g. Suter, 1980a, 1980b; Negro et al., 2007). However, new examination of these units led Michard et al. (2007) to favour an external origin, as admitted by Faure-Muret & Choubert (1971a, 1971b) and Frizon de Lamotte (1985). This is strongly suggested by the continuous metamorphic gradient observed from the South Temsamane to North Temsamane units (Fig. 5.30). Moreover, the lack of Dorsale or Flysch slivers at the bottom of the Ras Afraou unit, and the fact that the Khebaba unit is overlain by the Aknoul nappe obviously contradict an origin from the Internal Zones.

Therefore, the hypothesis arises that the External Rif MP-LT metamorphism is related to a N-dipping subduction zone extending between the Intrarif and Mesorif zones of the African paleomargin (Michard et al., 2007). This is supported by the occurrence of serpentinite remnants in the Beni Malek and Tres-Forcas massifs. West of the Nekor Fault, which is a lateral ramp for the Ketama SE-verging thrust, the “Mesorif suture zone” is probably hidden beneath the Intrarif. However, a volcanogenic level with gabbro and diabase clasts occurs at the bottom of the Ketama series (Zaghloul et al., 2003), suggesting a western continuation of the former thinned crust zone. This hypothetic intra-margin (intracontinental) suture continues eastward at least up to the Oran coastal massifs where serpentinites and chloritoid-bearing metapelites occur beneath the most internal Tell units (Guardia, 1975; Fenet, 1975). Accordingly, the Mesorif suture zone can also be referred to as the “External Maghrebide Suture Zone” (Michard et al., 2007).

5.6 Syn- to Post-Orogenic Magmatism

References: The Cenozoic magmatism of the Gibraltar Arc and Maghrebide belt has been repeatedly considered in the last decades : Hernandez et al. (1987), Turner et al. (1999), Zeck et al. (1999), El Bakkali et al. (1998), El Azzouzi et al. (1999), Maury et al. (2000), Münch et al. (2001), Coulon et al. (2002), Savelli (2002), Duggen et al. (2004), Gill et al. (2004), and Pecerillo & Martinotti (2006). The earliest magmatic intrusions were described by Hernandez et al. (1976), Torres-Roldan et al. (1986), and Cuevas et al. (2006). Thermal spring geochemistry has been recently addressed by Tassi et al. (2006).

In the eastern part of the Gibraltar Arc and the Algerian-Tunisian Maghrebides, magmatism essentially developed during the Miocene. However, the magmatic climax was preceded by discrete magmatic events as early as the Paleocene, and followed by a very recent volcanic activity, which extends widely outside of the Maghrebide belt, in the Atlas and Anti-Atlas domains (see Chap. 4).

5.6.1 Early Magmatic Events

Eocene alkaline magmatism took place here and there in the African margin (e.g. Tamazert; Chap. 4). In the eastern Prerif, basanites emplaced at Sidi Maatoug, northeast of Taza (Fig. 5.7). These lava flows are dated by the occurrence of pyroxene crystals reworked from associated ashes in the neighbouring lower-middle Paleocene sediments. Moreover, they yielded a 57 ± 7 Ma Rb-Sr date, and basanite pebbles occur in the Oligocene formations close to the volcanic body (Hernandez et al., 1976). Such volcanism can be assigned to a local extensional/transtensional setting.

Another early magmatic event, probably more significant, concerns the emplacement of an andesitic dyke swarm in the western Malaguides. The dykes were first dated by K/Ar at ~23 Ma (Torres-Roldan et al., 1986). Recent datings by ^{40}Ar-^{39}Ar yielded both older and younger ages: 30 ± 0.9 Ma (Turner et al., 1999) and 33.6 ± 0.6 Ma (Duggen et al., 2004), and on the other hand several ages in the range 19.8–17.4 Ma. The older ages are consistent with the dyke emplacement in the Malaguide terrane prior to its thrust deformation (~28 Ma; see Sect. 5.2.2.3). The K/Ar system was likely disturbed by the heating event that affected the Malaguide nappe stack at about 25–23 Ma. The dominant N15E trend of the less deformed dykes suggests a dominant E-W extension of the Malaguide-Ghomaride terrane during the Early Oligocene. Moreover, the major and trace elements data and Sr-Nd isotopes data favour derivation of the Malaga dykes through the subduction process (Duggen et al., 2004).

As for the leucogranitic dykes, which intrude the Ronda and Beni Bousera peridotites and their country rocks, they are younger, 20–19 Ma (Sect. 5.5.2.2), and generally less deformed. They are assigned to partial melting of the crustal unit beneath the peridotite slab (Ojen, Monte Hacho) during the thermal peak, at about 21 Ma (see above). This early and limited magmatism is coeval with the initial opening of the most active Alboran Basin.

5.6.2 The "Orogenic", Post-Collisional Magmatism

The so-called "orogenic magmatism" of the Rif Belt is part of a large belt, which stretches from Tunisia to eastern Morocco, then forms a number of volcanic centers in the Alboran Sea (in particular the Alboran Island itself), and finally extends to eastern Betic Cordilleras ("Trans-Alboran magmatism"; Fig. 5.40). This widespread magmatism is labelled "orogenic" as it generally displays petrologic and geochemical signatures typical for supra-subduction zone (SSZ) calc-alkaline magmas. However, it differs from a genuine SSZ orogenic magmatism by its moderate

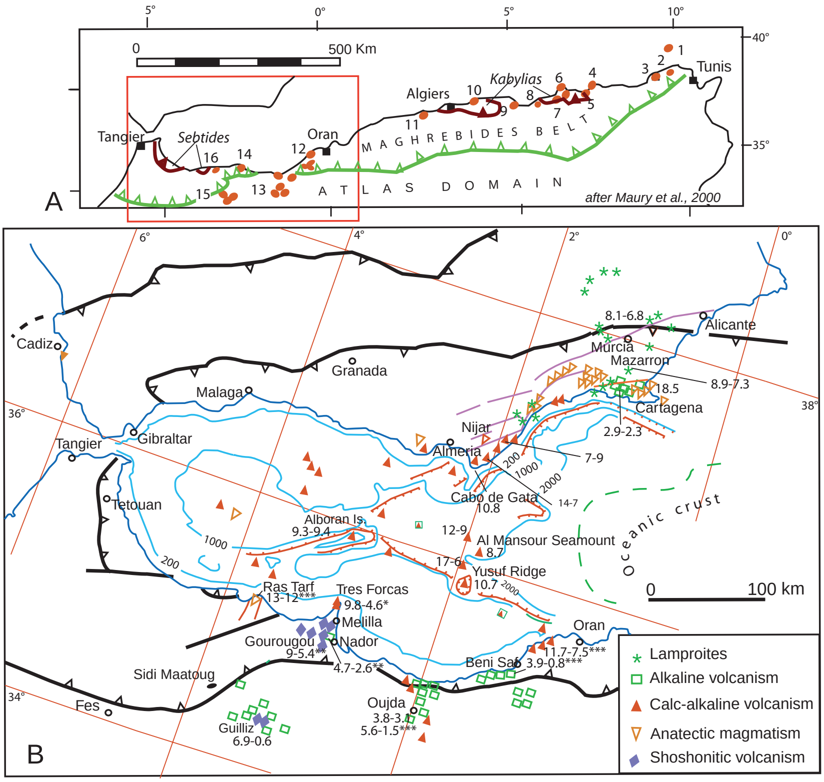


Fig. 5.40 Distribution of the late- to post-orogenic magmatism in the Maghrebides and Gibraltar Arc. (**A**): Maghrebian magmatic belt, after Maury et al. (2000), modified. G: granodiorite (with gabbro at Cap de Fer) and associated volcanics; Vo/m/po: orogenic/mixed/post-orogenic volcanos. 1 La Galite (G); 2 Mogods (Vpo); 3 Nefza (m); 4 Cap de Fer-Edough (G); 5 Filfila (G); 6 Cap Bougaroun (G); 7 Beni Touffout (G); 8 El Aouana (Vo); 9 Béjaïa-Amizour (G); 10 Algérois (G); 11 Cherchell (G); 12 Oranie (Vo, Vm, Vpo); 13 Oujda (Vpo); 14 Gourougou-Trois Fourches (Vo, Vm, Vpo); 15 Guilliz (Vm, Vpo); 16 Ras Tarf (Vo). – (**B**): Eastern Rif, Eastern Betics and Trans-Alboran domain, after Hernandez et al. (1987), El Azzouzi et al. (1999) and Duggen et al. (2004). Ages (Ma) reported mainly after Duggen et al. (2004, with reference therein), except (*), after Hernandez et al. (1987), (**), after Münch et al. (2001), and (***), after Maury et al. (2000)

volume, and because it occurs after the orogenic paroxysm. In fact, the Maghrebide and Trans-Alboran magmatism postdates the Internal Zones overthrusting on the External Zones, and straddles the limit between these zones. It begins as early as 15–16 Ma in Tunisia and eastern Algeria, then reaches western Algeria, eastern Morocco, Alboran and eastern Cordilleras around 13–10 Ma (Figs. 5.40, 5.41), being contemporaneous with the Serravallian-Tortonian sedimentation in the Alboran Basin (cf. Fig. 5.8). In other words, this "orogenic magmatism" is indeed coeval with the late orogenic extension of the Internal Zones. It is described by Lustrino

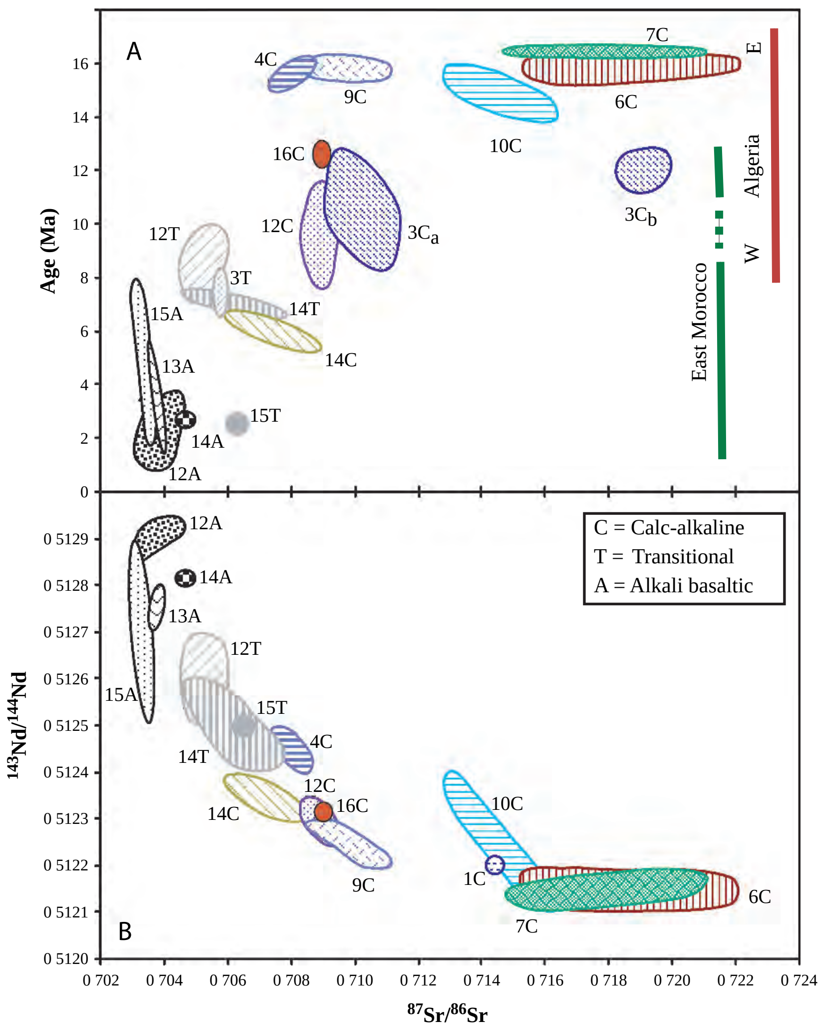


Fig. 5.41 Geochemical evolution of the Maghrebide Belt magmas through time, after Maury et al. (2000), modified. (**A**): Correlations between Sr isotopic ratios and age. – (**B**): Correlations between Sr and Nd isotopic ratios. Numbers refer to the areas shown in Fig. 5.40A. 3a/3b: cordierite-free/cordierite-bearing shoshonitic lavas (Nefza)

& Wilson (2007) as a part of the much larger circum-Mediterranean *anorogenic* Cenozoic igneous province.

The magmatic belt includes granodiorite massifs (e.g. Beni Bou Ifrour), but in most cases magmatism corresponds only to volcanic complexes of andesites, dacites, rhyodacites and rhyolites flows, dykes and sills. The lavas are generally K-enriched, up to shoshonitic compositions (Oran region, Gourougou south of Melilla). Their Sr-Nd-Pb-O isotope compositions suggest a lithospheric mantle origin with a strong crustal imprint (Fig. 5.41). The magmatic evolution ends at 10–7 Ma with the emplacement of transitional, calc-alkaline/alkaline basalts, andesitic basalts and trachy-andesites (Gourougou, Guilliz), and lamproites (Murcia).

As a whole, the detailed geochemical data now available give evidence of two distinct sources, i.e. (i) a main source in a subcontinental lithospheric mantle modified by the subduction of a lithospheric slab, and (ii) the overlying continental crust locally affected by partial melting (garnet-cordierite granites), which have contaminated the ascending calc-alkaline magmas. Many authors (e.g. Maury et al., 2000) hypothesize a direct link between this magmatic evolution and the African plate subduction, ending with the oceanic slab tear-off (Sect. 5.7). The upward flow of enriched asthenospheric mantle through the tear would have triggered melting of the lithospheric mantle already metasomatised during a previous subduction episode. Later, partial melting would have occurred at the uprising asthenosphere-lithosphere boundary, thus generating basalts having transitional character.

5.6.3 Recent Alkaline Magmatism of the Morocco Hot Line

An alkaline, intraplate-type volcanism took place at both the east and west tips of the Maghrebian magmatic belt in the most recent period, from 6 to 0.8 Ma. In contrast with the earlier magmas, the new ones also occur outside of the Maghrebide belt. The volcanic centers now occur along a NE-trending belt extending from the Trans-Alboran zone (Fig. 5.40B) to the Middle Atlas domain (Chap. 4) and to the Anti-Atlas (J. Siroua volcano). The magmatic rocks are alkaline basalts, basanites, hawaites, and nephelinite, lacking crustal contamination (Fig. 5.41). Their geochemical signature compares with that of the intra-oceanic island basalts, and suggests the role of an asthenospheric "hot line", labelled the *Morocco Hot Line* (MHL) in the Chap. 4 of the present volume. Such interpretation is consistent with the gravimetric and geodetic modelling of the lithosphere (cf. Chap. 1), and the lack of well defined age versus position trend. The MHL could extend from the Canary Islands to southeast Spain at least.

It is worth noting that CO_2-rich thermal springs with ^{3}He anomalies are likely related to this hot line. They are mainly distributed along a NE-SW trend from Nador to Taza, and from Fes (Moulay Yacoub) to Oulmes south of the Rif frontal thrust. The contemporary presence of ^{3}He anomalies and minor recent basalt outcrops indicate that CO_2 originates from mantle degassing or deep hydrothermal systems in these thermal discharges (Tassi et al., 2006).

5.7 Mountain Building

The future Rif-Betic or Gibraltar Arc orogenic domain was created by the breaking down of Pangea and opening of the Central Atlantic and western Tethys Oceans (Figs. 1.5, 1.9). The oceanic corridor between Iberia and Africa was always narrow (200–300 km). However, oceanic lithosphere occurred everywhere north of the Maghreb margin and west of the Adria plate (the "African promontory"; Fig. 1.9A), being most probably connected with the Alpine oceanic lithosphere. At about 75 Ma, the latter ocean began to close by subduction of the European plate beneath the Adria margin. Further to the southwest the scenario in the Ligurian-Maghrebide Ocean is still a matter of debate.

In this section, we discuss those geodynamic processes, which resulted in the building of the Rif mountains. This cannot be done without taking into account the entire Gibraltar Arc. After an abridged synopsis of the successive events in due chronological order (Sect. 5.7.1), and the presentation of the kinematic data and paleomagnetic rotations (Sect. 5.7.2), we first present the most recent orogenic period, which is also the best understood, i.e. the Oligocene-Neogene interval (Sect. 5.7.3). Based on this reconstruction, which is currently accepted, we then discuss the earlier and much debatable stages of the Rif-Betic orogeny (Sect. 5.7.4). The last Sect. 5.8 presents the present-day stress field in the region, which has important environmental consequences and completes the orogenic evolution up to contemporary times.

5.7.1 Abridged Orogenic Chronology and First Interpretations

References: The recent stratigraphic and radiochronologic references are indicated in the preceding Sects. (5.3–5.6), and some of them recorded in the legend of Fig. 5.42. Synthetic tables are proposed by Duggen et al. (2004), and Jolivet et al. (2006) concerning the metamorphic, magmatic and tectonic events.

Mountain building of the Betic Cordilleras and Maghrebides is basically the final product of the Africa-Eurasia plate convergence since ca. 70 Ma ago (see Chap. 1, Fig. 1.10). Africa-Europe convergence in the west, along the Gibraltar transect, is lesser than in the east, along the Sicilian transect, by approximately a factor 2. In the Gibraltar transect, the orogenic domain suffered in total 250 km of N-S shortening from Late Cretaceous to Tortonian, and 50 km of further NW shortening until present day. Convergence is directed WNW since ~3 Ma (Fig. 1.10).

Before discussing the geodynamic processes that formed the belt, it is convenient to summarize the chronology of the orogenic events (Fig. 5.42), as recorded by the stratigraphic, metamorphic and magmatic data reported above (sect. 5.2 to 5.6). At each step, the suggested geodynamic interpretation will be shortly noted hereafter in *italics*.

The earliest sedimentary events recording significant tectonic movements are observed during the ***Late Cretaceous*** in the Alboran (AlKaPeCa) Internal Domain. In the Dorsale units, Senonian hemipelagic deposits (Couches rouges) overlie, through an unconformity, the Tithonian-Berriasian limestones. Chaotic breccias interbedded within the “scaglia”-type deposits (Hafa Ferkenich in the Rif Internal Dorsale, Djurdjura in Algeria; Raoult, 1974) suggest an *increasing tectonic activity*.

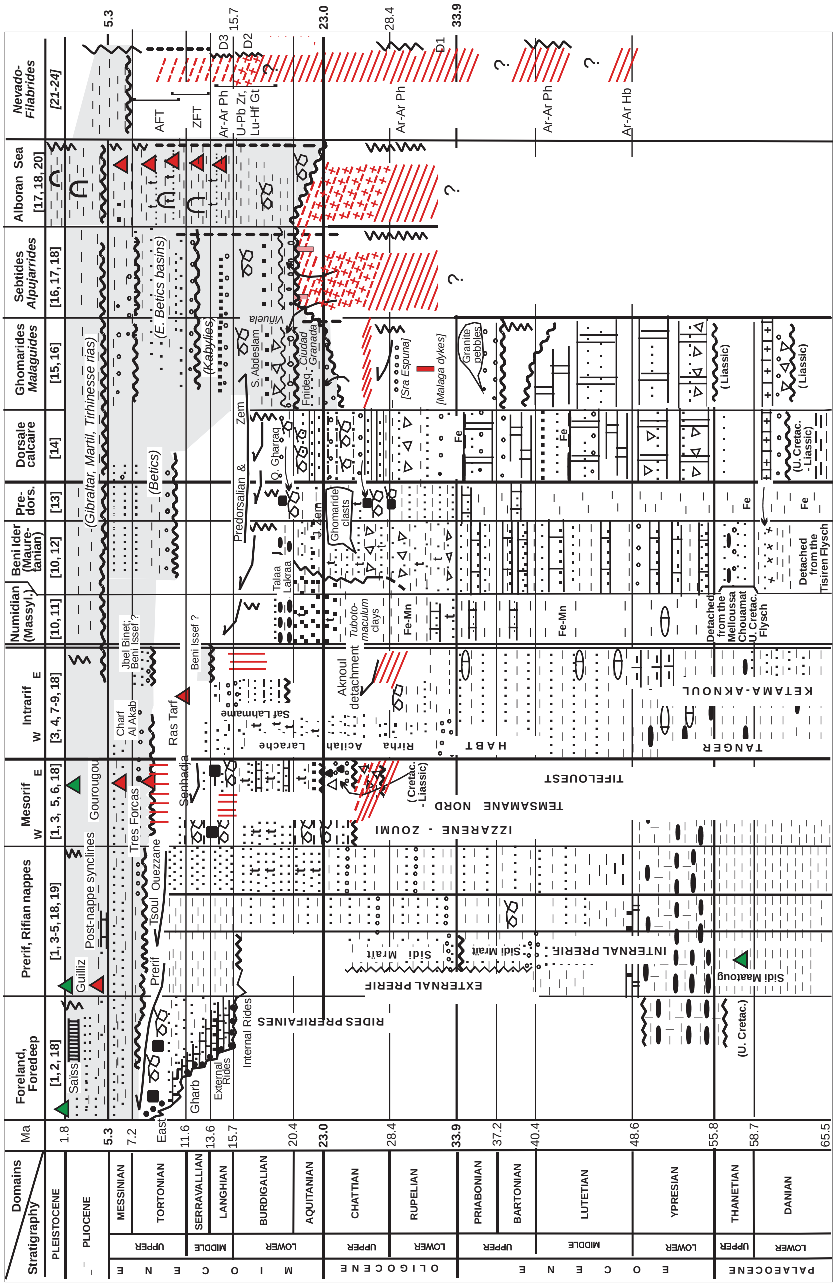
By the end of this period, during the ***Paleocene-Eocene***, the Ghomaride-Malaguide-Kabylides continental domain and its proximal margin (Internal Dorsale) first emerged, before being converted into a shallow marine platform fringing emerged lands. *The interpretation of this major event is not straightforward, but it can record the overriding of a subduction zone if the uplifted block belonged to the active margin*. At the northern (northwestern) boundary of the Maghrebian Flysch basin (Mauretanian basin), sandy turbidites and coarse olistostromes first appeared, whereas the southern part of the oceanic trough (Massylian basin) had pelagic sedimentation. *This suggests that a subduction trench was formed by the Middle-Late Eocene at the Dorsale-Maghrebian ocean transition* (Fig. 5.23).

In contrast, the African paleomargin itself remained virtually quiescent throughout Late Cretaceous to Eocene time. In the Prerif zone, Triassic evaporites and associated rocks are resedimentated in the Upper Cretaceous marls. Such phenomenon suggests the ascent of diapirs up to the Cretaceous seafloor. This was probably a result of an extensional/transtensional regime, in agreement with the Paleocene alkaline basalts erupted in the eastern Prerif (Sidi Maatoug).

Contractional events first occurred during the ***Late Eocene-Oligocene***, being mostly concentrated in the Internal (Alboran) Domain. These are: (i) stacking of the Ghomaride-Malaguide nappes, locally dated at about 28 Ma, and associated with deep erosion and coarse conglomerate deposits; (ii) turbiditic and olistostrome-rich sedimentation with internal alimentation in the Dorsale-Predorsalian-internal Flysch Trough (Beni Ider-Algeciras Flyschs); (iii) metamorphism in the Sebtide-Alpujarride units, the peak of which occurred at ~30 Ma, although some earlier isotopic ages (48 Ma) have been obtained; and (iv) emplacement of andesitic basalt dykes in the Malaguides (33–30 Ma). The latter two events are indicative of an ongoing subduction during the Oligocene. *The Dorsale and Ghomaride-Malaguide units were located in the arc domain above the subduction zone where the Sebtide-Alpujarride units were buried*.

During this Eocene-Oligocene period, coarse turbiditic influx increased in the northern part of the Flysch Trough (Fig. 5.23), which was progressively consumed, whereas its southern part was almost devoid of contemporary turbidites (Fig. 5.26). Likewise, most of the External Rif remains undisturbed, but *the Intrarif-Mesorif boundary acted as a minor suture zone* (from the Nekor-Beni Malek to the Oran massifs at least). This view is supported by the occurrence of the Tifelouest duplexes sealed by Oligocene-Miocene chaotic breccias and turbidites, and by the age of the greenschist to MP-LT metamorphism of the Ketama–Temsamane massifs, dated at ca. 28–23 Ma.

At the transition from ***Oligocene to Miocene*** (latest Oligocene-Early Burdigalian, 25–18 Ma), the upper plate, i.e. the Ghomaride-Malaguide-Kabylian range, was eroded, faulted, and progressively submerged. Contemporaneously, the buried Sebtide-Alpujarride units were exhumed under increasing geothermal gradients.


Domains
Stratigraphy
Ma
Foreland, Foredeep
[1, 2, 18]
Prerif, Rifian nappes
[1, 3-5, 18, 19]
Mesorif
W E
[1, 3, 5, 6, 18]
Intrarif
W E
[3, 4, 7-9, 18]
Numidian (Massyl.)
[10, 11]
Beni Ider (Mauretanian)
[10, 12]
Pre-dors.
[13]
Dorsale calcaire
[14]
Ghomarides Malaguides
[15, 16]
Sebtides Alpujarrides
[16, 17, 18]
Alboran Sea
[17, 18, 20]
Nevado-Filabrides
[21-24]
PLEISTOCENE
PLIOCENE
MESSINIAN
TORTONIAN
SERRAVALLIAN
LANGHIAN
BURDIGALIAN
AQUITANIAN
CHATTIAN
RUPELIAN
PRIABONIAN
BARTONIAN
LUTETIAN
YPRESIAN
THANETIAN
DANIAN
UPPER
MIDDLE
LOWER
MIOCENE
OLIGOCENE
EOCENE
PALAEOCENE
1.8
5.3
7.2
11.6
13.6
15.7
20.4
23.0
28.4
33.9
37.2
40.4
48.6
55.8
58.7
65.5
Saïss
Gharb
East
External Rides
Internal Rides
RIDES PRERIFAINES
Guilliz
Post-nappe synclines
Prerif
Tsoul
Ouezzane
Tres Forcas
Gourougou
Senhadja
EXTERNAL PRERIF
INTERNAL PRERIF
Sidi Mrait
Sidi Maatoug
(U. Cretac.)
IZZARENE - ZOUMI
TEMSAMANE NORD
TIFELOUEST
(Cretac. - Liassic)
Charf Al Akab
Ras Tarf
Jbel Binet; Beni Issef ?
Beni Issef ?
Saf Lahmame
Larache
Aciliah
Rirha
HABT
TANGER
KETAMA-AKNOUL
Aknoul detachment
Tuboto-maculum clays
Fe-Mn
Detached from the Melloussa Chouamat U. Cretac. Flysch
Talaa Lakraa
Ghomaride clasts
Zem
Predorsalian & Zem
O. Gharraq
Detached from the Tisiren Flysch
Fe
(U. Cretac. - Liassic)
(Gibraltar, Martil, Tirhinesse rias)
(Betics)
(E. Betics basins)
(Kabylies)
S. Abdeslam
Viñuela
Fnideq, Ciudad Granada
[Sra Espuna]
[Malaga dykes]
Granite pebbles
(Liassic)
AFT
ZFT
Ar-Ar Ph
U-Pb Zr, Lu-Hf Gt
Ar-Ar Hb
D1
D2
D3
?

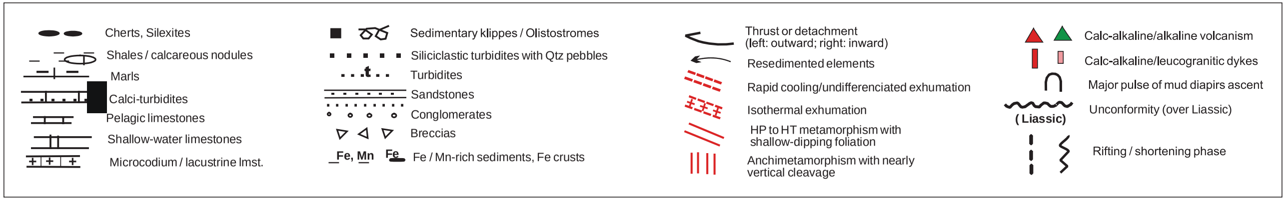


Fig. 5.42 Chronology of the stratigraphic, metamorphic and magmatic events recorded in the Rif Belt and Betic Cordilleras (*names in italics*) during the Cenozoic, after Chalouan et al. (2001), substantially modified. Time scale from the International Commission on Stratigraphy (2004). Main data sources, in addition to Suter (1980a, 1980b): 1: Feinberg (1986), Wernli (1987); Kerzazi (1994); 2: Faugères (1978), Plaziat & Ahmamou (1998), Zizi (1996), Zouhri et al. (2001), Litto et al. (2001); 3: Ben Yaïch (1991), Toufiq et al. (2002); 4: Morley (1992), Leblanc (1979); 5: Cornée et al. (1996), Samaka et al. (1997), Münch et al. (2001); 6: Frizon de Lamotte (1985), Ben Yaïch et al. (1989), Favre (1992), Negro et al. (2007); 7: Zaghloul et al. (2005); 8: Abdelkhaliki (1997), Zakir et al. (2004); 9: Didon & Feinberg (1979), Septfontaine (1983), Lamarti-Sefian et al. (1998); 10: Hoyez (1989), Guerrera et al. (2005); 11: Esteras et al. (1995), El Kadiri et al. (2006); 12: Didon & Hoyez (1978), Zaghloul (2002), Puglisi et al. (2001), Zaghloul et al. (2007); 13: Mourier et al. (1982), Olivier (1984), Durand-Delga & Maaté (2003); 14: Wildi et al. (1977), Olivier et al. (1979), Durand-Delga (1980), Nold et al. (1981), Olivier (1981–1982), Ben Yaïch et al. (1986), El Kadiri et al. (1992), Hlila et al. (1994), El Kadiri (2002a, 2002b), El Kadiri et al. (2005); 15: El Kadiri et al. (1992), Martín-Martín et al. (1997), Martín-Algarra et al. (2000), Negro et al. (2006); 16: Chalouan & Michard (1990), Feinberg et al. (1990), Durand-Delga et al. (1993), Lonergan & Mange-Rajetsky (1994), Serrano et al. (2006); 17: Michard et al. (1991, 1997), Sánchez-Rodriguez & Gebauer (2000), Platt et al. (2003b), Negro (2005); 18: Hernandez et al. (1987), El Bakkali et al. (1998), El Azzouzi et al. (1999), Maury et al. (2000), Duggen et al. (2004); 19: Hernandez et al. (1976); 20: Platt et al. (1988), Comas et al. (1992, 1999), Chalouan et al. (1997), Soto & Platt (1999), Soto et al. (2003), Sautkin et al. (2003), Talukder et al. (2003); 21: Johnson et al. (1997); 22: Martínez-Martínez et al. (2002); 23: Augier et al. (2005a, 2005b); 24: Platt et al. (2006)

Turbiditic sedimentation then invaded the entire, residual Flysch Trough and the Intrarif domain. By the end of this period, during the Middle Miocene (16–11 Ma), shortening encroached the Flysch Trough, Intrarif, and Subbetic domains.

Nevertheless, extension in the central part of the Alboran Domain continued during the whole ***Miocene*** (Late Burdigalian-Early Tortonian, 18–9 Ma), resulting in the continued subsidence and the development of the Alboran Sea in the west. This protracted extension at the core of the Alboran Domain was accompanied by conspicuous calc-alkaline volcanism. It was also coeval with the outward displacement of the turbiditic depocenters and thrust contacts into the Mesorif domain, and then into the Prerif domain. Obviously, plate convergence is not the only process at work, as shortening is combined with extension in the Alboran Domain. *This is interpreted through the retreat of the Maghrebian subduction zone with tearing and break-off of the plunging slab* (see below).

Finally, during the ***Late Tortonian-Pleistocene times*** (9 Ma to Present), the Mediterranean Sea overlay large parts of the orogen, due to *late orogenic collapse of the tectonic prism*. However, *contractional processes mostly related to plate convergence* also operated at some places (e.g. Alboran Ridge, post-nappe synclines of the Mesorif), being still active in the Rides prérifaines. Alkaline basalts emplaced in the eastern Rif, in relation with the Morocco Hot Line (Sect. 5.6.3) superimposed onto the Maghrebide-Betic structures.

5.7.2 Kinematic Data and Paleomagnetic Rotations

References: Kinematic data (transport direction and sense of shear) have been collected throughout the Gibraltar Arc by many authors and for different time intervals. They are presented synthetically in the following works (with references therein): (i) for the Sebtide-Alpujarride units (late Oligocene-Miocene deformations), by García-Dueñas et al., 1992, 1995), Michard et al. (1997), Chalouan et al. (1995, 1997), Balanyá et al. (1997), Martínez-Martínez & Azañón (1997), Azañón & Crespo-Blanc (2000), and Booth-Rea et al. (2004); (ii) for the Nevado-Filabride Complex (Middle-Late Miocene deformation) by Martínez-Martínez et al. (2002) and Augier et al. (2005a); (iii) for the External Zones of the entire orogenic arc (Early-Late Miocene to Pleistocene deformation) by Frizon de Lamotte et al. (1991), Crespo-Blanc & Campos (2001), Aït Brahim et al. (2002), Platt et al. (2003a) – a paper discussed by Michard et al. (2005), Zakir et al. (2004), Bargach et al. (2004), and Crespo-Blanc & Frizon de Lamotte (2006). Concerning paleomagnetic investigations, a number of references are given in the caption to Fig. 5.43. Other pertinent works are those by Villalain et al. (1994), Platt et al. (2003a), Villasante-Marcos et al. (2003), Osete et al. (2004), Krijgsman & Garcès (2004), and Cifelli et al. (2008).

Kinematic indicators are widespread in the Sebtides (Fig. 5.35), and come from the top of the Beni-Bousera peridotites and the Federico units (Upper Sebtides) in the Beni Mezala area. They are associated with retrogressive mineral phases, and then related to the exhumation processes. They show a remarkably constant top-to-NW

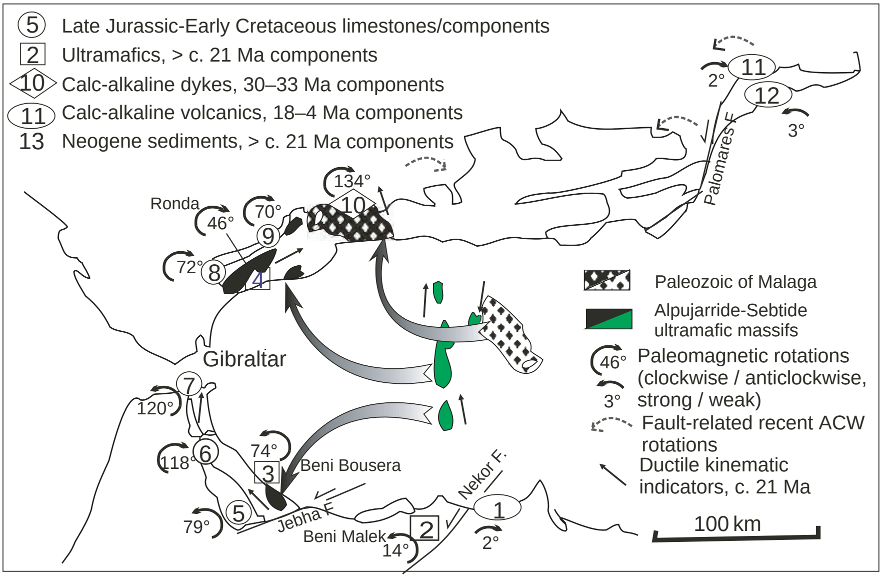


Fig. 5.43 Paleomagnetic rotations (mean site values) in the Betic-Rif Belt, after Feinberg et al. (1996) and Chalouan & Michard (2004), modified. The inferred approximate N–S strike of the pre-Early Miocene orogen is shown schematically by the alignement of the three ultrabasic bodies in the mid-Alboran area. This also restores the main ductile shear directions of the west Gibraltar Arc, as reported by Michard et al. (1997), Balanyá et al. (1997), Martínez-Martínez et al. (2002), into an broadly N-S trend. 1: Najid et al. (1981); 2: Elazzab & Feinberg (1994); 3: Saddiqi et al. (1995); 4: Feinberg et al. (1996); 5–9: Platzmann (1992), Platzmann et al. (1993), Allerton et al. (1994); 10: Platzmann et al. (2000), Calvo et al. (2001); 11, 12: Calvo et al. (1997); 13: Mattei et al. (2006)

(Beni Bousera) and top-to-N (Beni Mezala) direction of shear. In the northern branch of the Gibraltar Arc, i.e. in the western Alpujarrides, coeval kinematic indicators point to top-to-N to top-to-NE shear. This strongly suggests that the corresponding exhumation tectonics originally occurred with a top-to-N sense of shear, and that the metamorphic units carrying the stretching lineations were bended during a further stage. Assuming that the recorded exhumation phase occurred at ca. 23–20 Ma, the curvature of the arc would have developed after the lowermost Miocene.

Paleomagnetic data permit to precise the latter conclusion (Fig. 5.43). Rotations of the Late Jurassic and Late Cretaceous limestones are mostly anticlockwise in the Rif Dorsale units, and clockwise in the Betic Dorsale and Penibetic (Internal Sub-Betic) zones. Similar opposite rotations are observed both in the Beni Bousera and Ronda peridotites, and within their leucogranite dykes. This observation enables us to demonstrate that rotations occurred after the cooling of these rocks (ca. 20 Ma). The moderate anticlockwise rotation measured in the Beni Malek ultramafics and overlying metasediments can be assigned to the sinistral deformation along the Nekor fault zone. At first sight, opposite and low rotations inferred in calc-alkaline volcanic rocks (cf. sites 1, 11, and 12; Fig. 5.43) suggest that rotation within the arc was completed at ca. 13–7 Ma. However, paleomagnetic results from

Fig. 5.44 An example of Miocene normal fault from the southern Alboran Sea shore: the Cape of Tres Forcas NW-dipping fault, cutting across the Late Tortonian-Messinian basal conclomerates at the north boundary of the Tarjat anticline (see Fig. 5.39 for location). The fault operated during the Messinian subsidence of the Cape Tres-Forcas block, with alternating normal and sinistral movements

Neogene sedimentary sequences from the Betic Chain show also that vertical axis rotation continued after the Late Miocene (Mattei et al., 2006; Cifelli et al., 2008). In summary, bending of the Gibraltar Arc occurred mostly during the Early-Middle Miocene, i.e. during the climax of extension, but continued afterwards.

Transport direction along Miocene thrusts display a broadly radiating pattern in the External Zones of the Arc (Fig. 5.5). They show a dominant westward component, thus suggesting a westward migration of the Gibraltar Arc. Such sense of displacement is clearly recorded in the N-Temsamane domain (Fig. 5.30) and results congruent with the one observed in the low-angle detachment that separates the Alpujarride Complex from the underlying Nevado-Filabride Complex. There, top-to- west shearing evolved from ductile to brittle conditions between ca. 12 and 8 Ma, contemporaneously with subsidence in the nearby Serravallian-Tortonian basins of the Betic Cordillera. The coeval, extensional directions associated with brittle normal faults all around the Alboran basin itself are roughly centripetal (Figs. 5.5, 5.44). *These observations are consistent with the formation of an orogenic arc, by simultaneous frontal thrusting and back, centripetal extension, in the context of a west-directed slab retreating process*.

5.7.3 Oligocene-Neogene: The Slab Retreat Process

References: The role of the slab retreat process in the opening of the Mediterranean basins and dispersal of the AlKaPeCa units was advocated by Rehaut et al. (1984), Frizon de Lamotte (1985), Malinverno & Ryan (1986), Royden (1993), Lonergan & White (1997), Doglioni et al. (1998), Frizon de Lamotte et al. (1991, 2000), and Jolivet & Faccenna (2000), among others. Recent papers describing the Gibraltar E-dipping slab and associated accretionary prism, based on marine geophysics, seismology and/or tomography are those by Torelli et al. (1997), Calvert et al. (2000), Gutscher et al. (2002), Faccenna et al. (2004), and Spakman & Wortel (2004). The occurrence of edge delamination is particularly advocated by Seber et al. (1996), Calvert et al. (2000), and Fadil et al. (2006). A detailed kinematic reconstruction of the movements of the Mediterranean microblocks (except Alboran) is proposed by Schettino & Turco (2006). Other seismological and geodetical references are reported in Chap. 1, Sect. 1.4.

Recently acquired seismological data offer convincing evidence for active east-dipping oceanic subduction below the western Aboran Sea. An east-dipping slab of subducted lithosphere is clearly imaged by tomography in the Western Mediterranean mantle (Fig. 5.45). Beneath the Betic-Rif-Alboran region, a positive (fast) P-wave velocity anomaly (i.e. a cold, dense body respective to the warmer ambient mantle), interpreted as a subducting slab, is found from the base of the crust across the entire upper mantle, down to the 660 km discontinuity. The deeper part of the anomaly extends beneath SE Spain, north of the Alboran basin. The upper part of the slab connects with the Atlantic oceanic lithosphere west of the Strait of Gibraltar. Evidence for current activity of this subducting slab is offered by seismicity distribution and seismic-reflection profiling (Fig. 5.46). Intermediate depth earthquake hypocentres are concentrated in the Alboran Sea, where plate curvature increases abruptly, probably due to P and T dependant dehydration reactions in the plunging slab. The deepest part of the slab coincides with the locus of deep earthquakes beneath southern Spain. On the Atlantic side, multichannel seismic profiling has imaged the "olistostrome" of the Gulf of Cadiz as an eastward-thickening accretionary

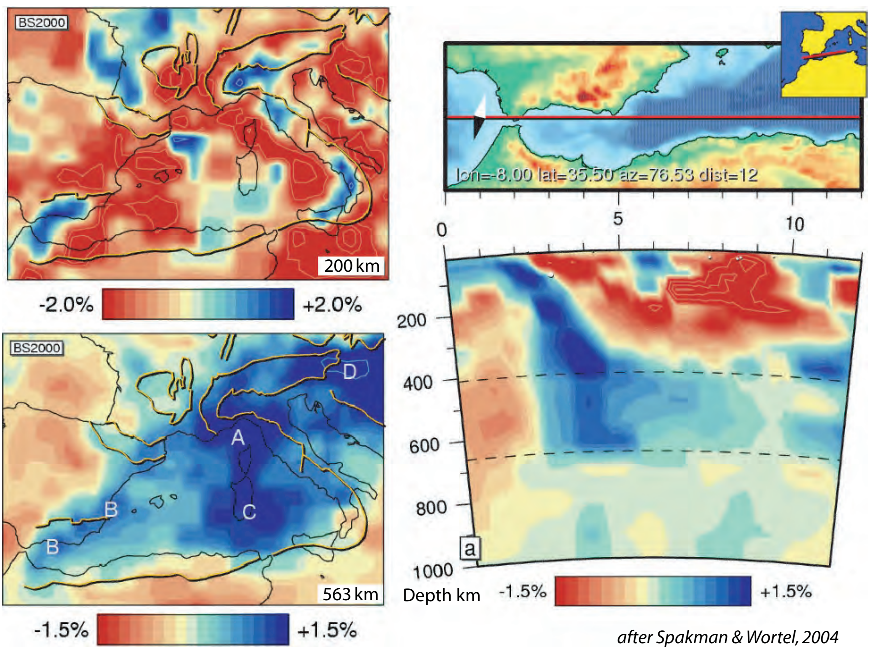


Fig. 5.45 V_p tomography of the Western Mediterranean area, after Spakman & Wortel (2004). *Left*: horizontal sections at 200 km and 563 km depth. *Right*: E-W vertical section across the Alboran-southern Algerian basin down to 1000 km depth, with location map. *Dashed lines* represent mantle discontinuities at 410 and 660 km depth

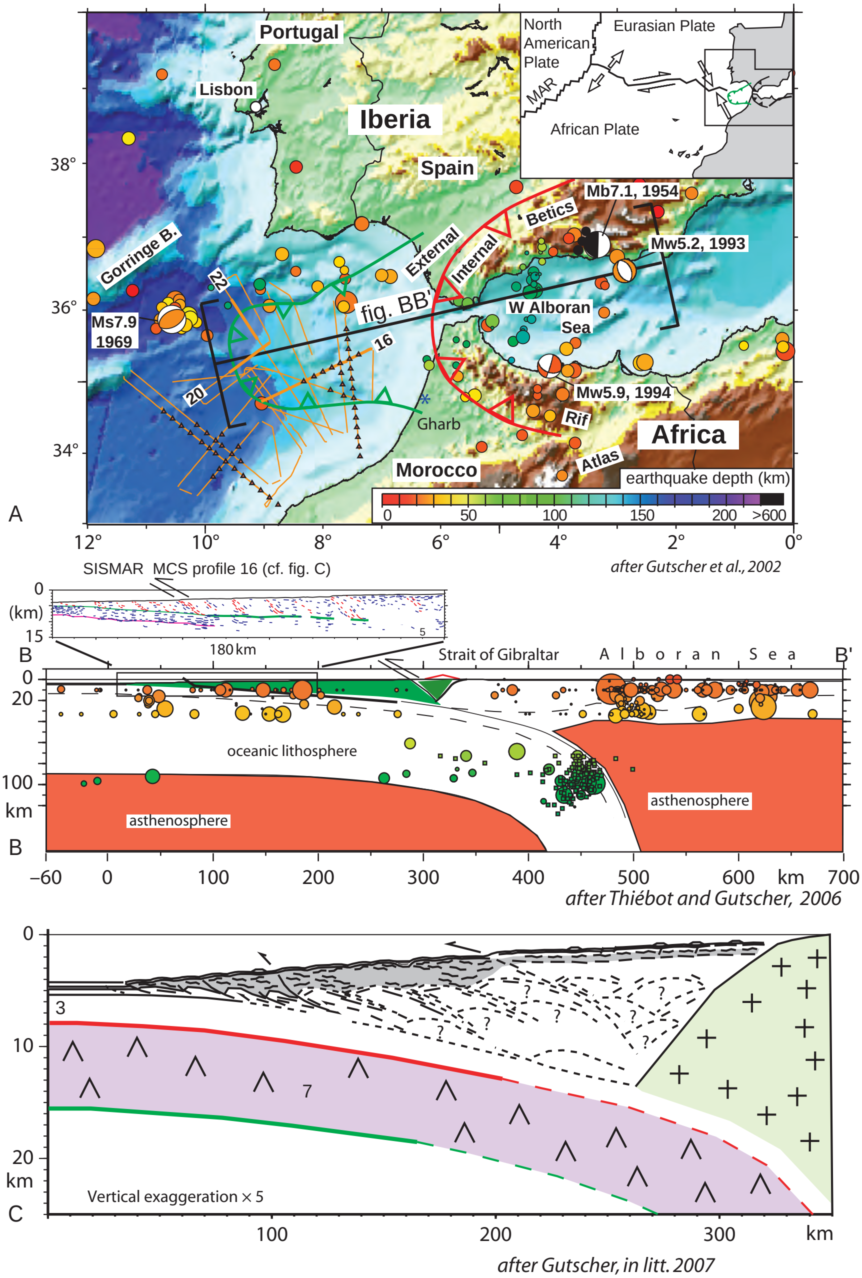


Fig. 5.46 (continued)

wedge ("Atlantis wedge"), overlying eastward dipping undeformed sediments and basement. Although some authors suggest that subduction is not active at present (e.g. Mauffret et al., 2007), the east-dipping reflectors deforming the seafloor indicate active ramp thrusts, thus suggesting that subduction is still active.

The shape of the Betic-Alboran subducting slab at a depth of 200 km correlates with the arcuate shape of the Gibraltar Arc and is also consistent with the spatial distribution of calc-alkaline volcanism in the back-arc domain. These observations offer a coherent timing for the kinematic evolution of the slab roll-back during the Miocene (Fig. 5.47). The subduction trench progressively turned from E-W to roughly N-S trend, through tearing and detachment along both the Iberia and the Maghreb margins. Indeed, no north-dipping slab is observed by tomography along the Algerian-Moroccan coast. The influx of asthenospheric mantle in the lithosphere tear-zones resulted in calc-alkaline magmatism, which tracks the westward mantle tearing (16–7 Ma in Algeria *versus* 12–6 Ma in Morocco and Alboran).

Fig. 5.46 (continued) The active subduction beneath Gibraltar. (**A**) Geographic map with shaded relief and location of epicentres (1973–Present; M > 3) and some focal mechanisms, after Gutscher et al. (2002). Red thrust teeth symbols indicate Gibraltar Arc; green thrust teeth symbols indicate active "Atlantis" accretionary wedge. SISMAR seismic reflection profiles in orange; position of bottom seismometers as small, lined up red triangles. Seismicity sampling box for B is also indicated. – (**B**): Simplified lithospheric cross-section showing the distribution of earthquake hypocenters (sampling box in **A**), and the geometry of the upper plate (Gibraltar block), lower plate and accretionary prism, after Thiébot & Gutscher (2006). – (**C**): Detailed cross-section of the accretionary prism along the SISMAR MCS Profile 16 (for location, see **A** and **B**), after M.-A. Gutscher (pers. comm., 2007)

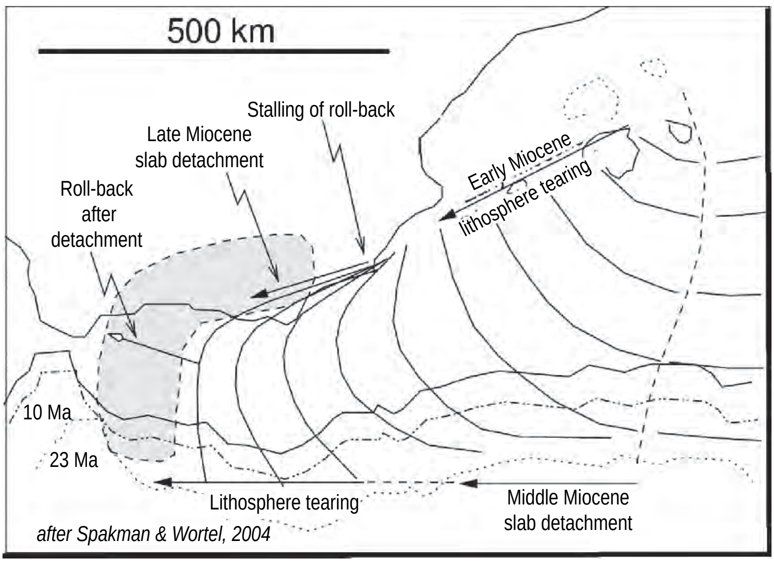


Fig. 5.47 Cartoon of the Miocene slab roll-back and lateral detachment and tearing of the Maghrebian-Ligurian oceanic slab in the Alboran area, after Spakman & Wortel (2004). The grey pattern shows the present-day position of the slab at 200 km depth (see Fig. 5.45). Relative position of Africa with respect to fixed Iberia is shown at 23 Ma, 10 Ma and Present

In this generally accepted scenario, the Alboran Basin spread in a back-arc position, similar to the Provençal and Valencia Basin (Fig. 5.48D, E). Mauffret et al. (2007) proposed that the western part of the Alboran Basin could have developed in a fore-arc position, but this is just open to discussion. It must be emphasized that

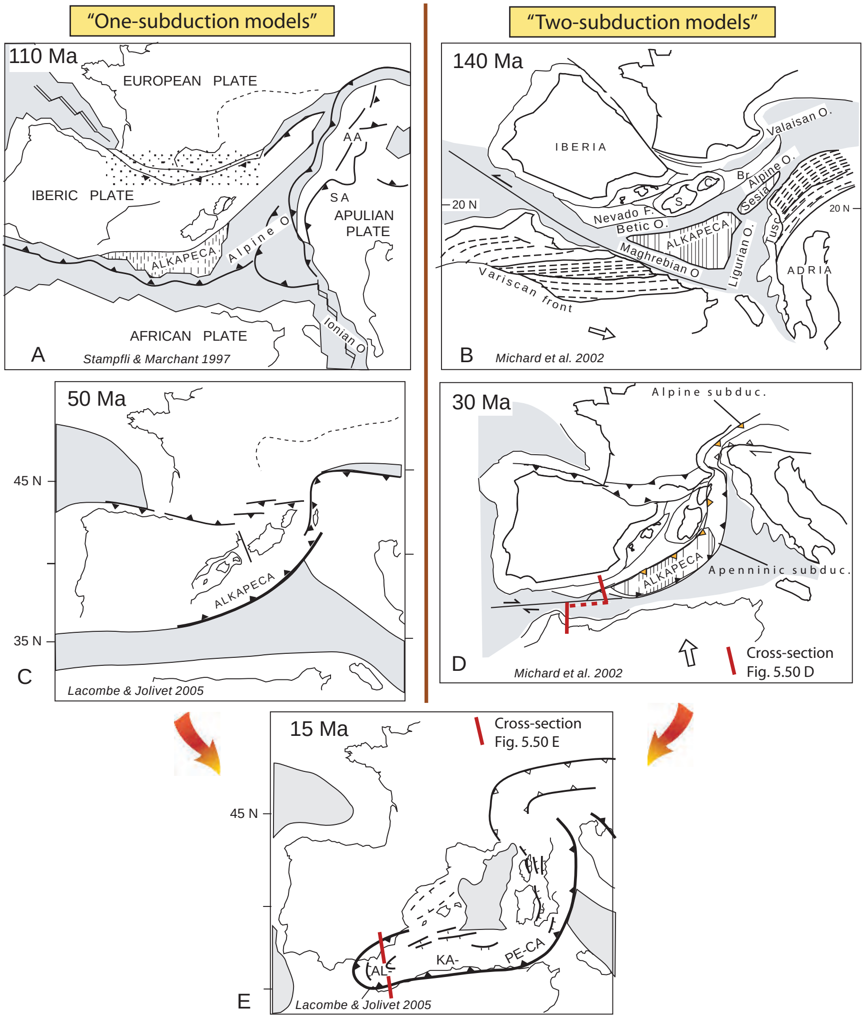


Fig. 5.48 Alternative scenarios for the Mesozoic setting of AlKaPeCa continental blocks and the successive positions of the subduction zone(s), after Michard et al. (2006), modified. *Left*: single subduction hypothesis; AlKaPeCa blocks were parts of SE Iberia during Early Cretaceous (**A**); during Tertiary convergence (**C**), the SE dipping Alpine subduction coexists with a NW-dipping Ligurian-Maghrebian subduction south of Corsica. – *Right*: two-subductions hypothesis; during the Mesozoic (**B**), AlKaPeCa formed a microcontinent separated from Iberia by a Betic oceanic arm; during the Eocene (**D**), a hypothetic SW extension of the Alpine subduction closes the Betic Ocean, being followed by the Ligurian-Maghrebian subduction at *ca*. 30 Ma. The Oligocene-Miocene evolution by back-arc extension and slab roll-back (**E**) is identical in both scenarios

the slab subduction does not result from E-W plate convergence but from passive sinking of unstable lithosphere into the mantle, within the narrow oceanic corridor between Iberia and Africa. At the scale of the Western Mediterranean, the subducted Ligurian lithosphere is now found at the base of the upper mantle, and as narrow, steeply dipping slabs in the Gibraltar Arc and its mirror image, i.e. the Calabrian Arc.

Slab roll-back was able to transport the fragments of the Betic-Rif Internal Zones (which were probably concentrated along the Balearic margin by the Eocene-Early Oligocene; see below) towards their present-day position with a predicted, dominant ENE-WSW trending extension. Slab roll-back also accounts for the rotations of the Alpujarride-Sebtide structures and paleomagnetic components during the Early Miocene, as described above (Fig. 5.43).

5.7.4 Late Cretaceous-Eocene: The Debatable Issues of the Early Subduction-Collision Tectonics

References: Tectonic scenarios for this period are still warmly debated. Correlations with the Alps in terms of plate tectonics have been repeatedly addressed since the late 70s: see reviews and references in Bouillin (1984) and Michard et al. (2002). Certain authors emphasized the role played by collisional processes and late-orogenic extension in the origin and evolution of the Betic-Rif orogen (e.g. Platt & Vissers, 1989; Turner et al., 1999; Houseman & Molnar, 2001). The occurrence of subduction is now widely accepted, although two competing models are proposed: (i) a single, NW-dipping subduction (e.g. Jolivet & Faccenna, 2000; Stampfli & Marchant, 1997; Stampfli et al., 2002; Lacombe & Jolivet, 2005; Jolivet et al., 2006), and (ii) two opposite subductions operating successively (e.g. Andrieux et al., 1989; Doglioni et al., 1999; Frizon de Lamotte et al., 2000; López Casado et al., 2001, Michard et al., 2002; Chalouan & Michard, 2004; Guerrera et al., 2005; Pecerillo & Martinotti, 2006). A discussion of these alternative models is proposed in Michard et al. (2006), and summarized hereafter.

5.7.4.1 Plate Tectonic Setting

During the latest Cretaceous-Paleogene times, NW-Africa moves northward relative to Eurasia by 200–250 km in the Morocco-Iberia transect (cf. Sect. 1.1, Fig. 1.10). Taking into account the shortening between Iberia and stable Europe (ca. 100 km across the Pyrenean belt), 100–150 km must have been consumed south of Iberia. As shortening across the Atlas is negligible up to the Neogene (Chap. 4), most of this shortening must have been accommodated by subduction of the Ligurian-Maghrebian oceanic lithosphere. During the Early Cretaceous, the Ligurian-Maghrebian (or, shortly, Ligurian) ocean was a triangular area connecting Central Atlantic, Alpine and Ionian oceans (Chap. 1, Fig. 1.9). When the Alpine ocean closed during the Eocene by subduction of the European plate below Adria, subduction affected most likely the Ligurian lithosphere, and thus may have dragged down some parts of the nearby continental blocks, thus generating HP-LT metamorphism in some of the AlKaPeCa units.

However, the location and number of subduction zones in the Ligurian area during the Late Cretaceous-Paleogene are questionable, as well as the location of the AlKaPeCa units (and particularly the Alboran Domain) at the onset of convergence. Several authors proposed that AlKaPeCa was part of the southeastern Iberian margin, close to Sardinia and the Balearic Islands (Fig. 5.48A). In contrast, others (included the authors of the present chapter) suggested that AlKaPeCa formed a microcontinent within the Ligurian oceanic area (Fig. 5.48B). In the latter hypothesis, a southwestward projection of the SE-dipping Alpine subduction could have been responsible for the consumption of the oceanic lithosphere between AlKaPeCa and Iberia (Fig. 5.48D). Metabasites and meta-serpentinites located toward the top of the Nevado-Filabride Complex can be taken as evidence for this interpretation. In consequence the Nevado-Filabrides could correspond to the distal part of the Iberian margin (as suggested recently by Platt et al., 2006). In contrast, in the "single-subduction hypothesis", the Ligurian subduction zone could have been located at the southeastern boundary of the AlKaPeCa units as early as the Late Cretaceous. As shown in Fig. 5.48C, this implies a flip-like reversal of the Alpine subduction from a SE dip along the Corsica-Northern Apennine transect to a NW dip along the AlKaPeCa-Southern Apennine/Maghreb transects. In both hypotheses, the future Alpine terranes of the western Mediterranean were located next to the Iberian margin, immediately before the slab retreat process responsible for their migration and opening of the Mediterranean Sea (Fig. 5.49).

5.7.4.2 The Nevado-Filabrides Issue

References: A new scenario was proposed by Platt et al. (2006) based on the young dating (ca. 18–14 Ma) of the Nevado-Filabride high-pressure (HP) metamorphism (under eclogite facies conditions). However, and due to the nature of this metamorphism, other scientists inferred older ages, ranging from Middle Eocene to Early Miocene, as for example Monié et al. (1991), López Sánchez-Vizcaíno et al. (2001), Puga et al. (1999, 2002a, 2002b), De Jong (2003), and Augier et al. (2005b).

None of the above tectonic scenarios attain a consensus, partly due to the lack of unequivocal information on the Nevado-Filabride Complex, its stratigraphy, the age of the high pressure (HP) metamorphism, the initial relationships with respect to the Alpujarride-Malaguide complexes in the Betic Cordillera, and its former position in the western Mediterranean. Did the Nevado-Filabride separate from the Iberian margin (e.g. Fig. 5.48B) or from the African one (Bouillin, 1984)? Or did it form an independent terrane located either north or south of the Alpujarride-Malaguide complex? Moreover, it was assumed that the Nevado-Filabride HP metamorphism (with eclogite facies rocks) was older or simultaneous with the Alpujarride one, but most recent datings at ca. 18–14 Ma (U-Pb in zircon, López Sánchez-Vizcaíno et al., 2001; Lu-Hf in garnet, Platt et al., 2006) now conflict with others radiometric ages at ca. 40–30 Ma (*in situ* laser $^{40}Ar/^{39}Ar$ on phengite, Augier et al., 2005b). This, of

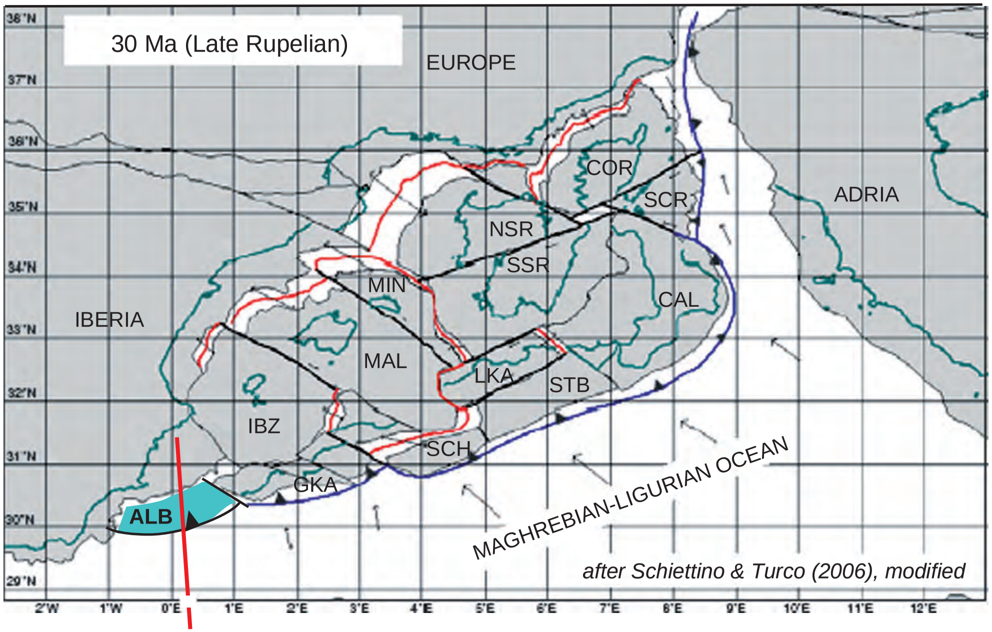


Fig. 5.49 Plate tectonic reconstruction of the Western Mediterranean microplates at 30 Ma, i.e. by the beginning of their disruption, after Schettino & Turco (2006), modified. Red lines are extension centres. *Straight black lines* are strike-slip faults. *Curved black lines* with teeth are subduction zones (note the hypothetized flip in front of southern Corsica). *Arrows* represent direction and magnitude of motion relative to Eurasia. *Thick red line* across Alboran corresponds to cross-section in Fig. 5.50. ALB: Alboran; CAL: Calabria; COR: Corsica; GKA: Greater Kabylia; IBZ: Ibiza; LKA: Lesser Kabylia; MAL: Mallorca; MIN: Menorca; NSR: Northern Sardinia; SCH: Sardinian Channel; SCR: Southern Corsica; SSR: Southern Sardinia; STB: South Tyrrhenian Block

course, results in very distinct reconstructions of the pre-Oligocene evolution of the Alboran Domain.

In particular, Platt et al. (2006) suggest that two phases of continental subduction operated successively; with (i) a Paleogene, north-dipping Alpujarride subduction beneath the Malaguides, then (ii) an Early Miocene, south-dipping Nevado-Filabride subduction underneath the Malaguide-Alpujarride stack. Assuming this interpretation, Early Miocene subduction of the Nevado-Filabrides is coeval with the pervasive extension of the upper terranes of the Alboran Domain (Alpujarrides and Malaguides) and with the slab retreating described above. This scenario corresponds to the Burdigalian-Langhian interval, very close to the onset of continental to shallow-marine sedimentation observed in some of the sedimentary basins placed close to the Nevado-Filabrides. Therefore, it probably indicates that subduction was followed immediately by ultra-rapid exhumation of this terrane and concomitant subsidence during the Middle Miocene. However, one cannot rule out confidently that some of these ages could have been rejuvenated during the isothermal exhumation episode.

5.7.4.3 Internal Paleogeography of the Alboran Domain

The respective location of the Sebtide-Alpujarride, Ghomaride-Malaguide and Dorsale units in the initial paleogeography are seldom addressed in the tectonic scenarios for Betic-Rif mountain building. However, this paleogeographic problem is quite important because its solution puts strong constraints on the interpretation of the Sebtide-Alpujarride metamorphism. The Dorsale domain was certainly located at the south border of the continental domain as it represents its passive margin north of the Maghrebian Flyschs oceanic arm. If we assume with Durand-Delga (1980, 2006) that the Sebtide-Alpujarride domain was located north of the Ghomaride-Malaguide and formed an uplifted, deeply eroded part of the continental domain, then a south-dipping subduction must be hypothetized to explain the present-day structure (Fig. 5.50A, B), and the Sebtide-Alpujarride metamorphism could then be nearly coeval with that of the Nevado-Filabrides. In contrast, we may assume with Wildi et al. (1977) that the Sebtide-Alpujarride units represent the former Paleozoic-Middle Triassic basement of the Dorsale units, as the latter are detached on the Carnian evaporites. In that case, the Sebtide-Alpujarride units were located south of the Ghomaride-Malaguide domain (Fig. 5.50C), and the corresponding subduction would dip northward, being distinct from the Nevado-Filabride (Fig. 5.50D). The present-day structural setting (Fig. 5.50E) seems more easily obtained (after the slab roll-back evolution) in the frame of the latter hypothesis than of the former. Discussing more deeply the above scenarios, either in map view or in cross-section, would clearly be beyond the scope of this chapter. They must be regarded as working hypotheses, not as the only possible answers to the Rif-Betic conundrum.

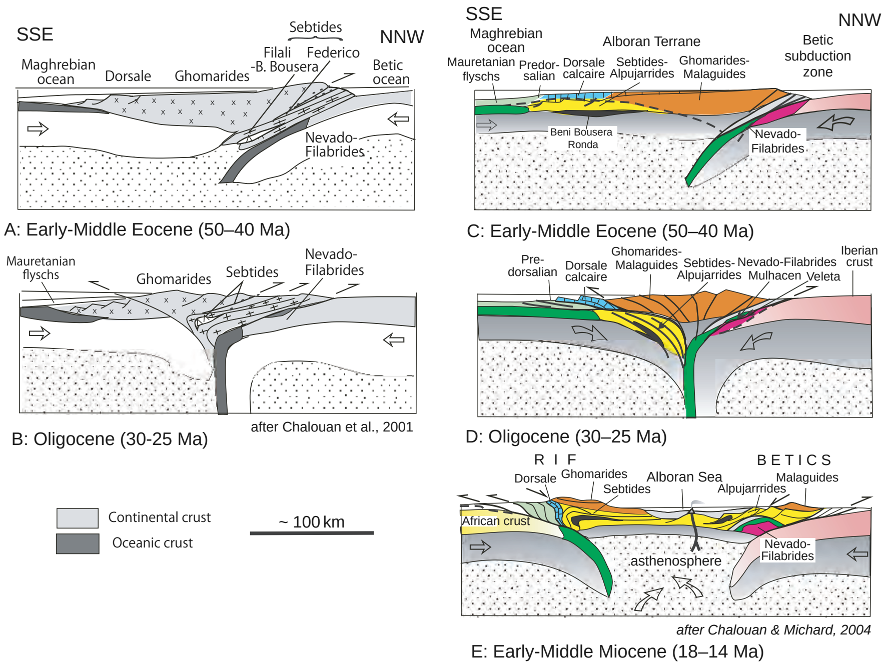


Fig. 5.50 Alternative hypotheses for the Betic-Rif mountain building in cross-section, after Chalouan & Michard (2004). (**A, B**): In this scenario (Chalouan et al., 2001), the Sebtide-Alpujarride domain (future *lower* plate) is initially located north of the Ghomaride-Malaguide domain (future *upper* plate), and then enters in a south-dipping subduction zone. – (**C, D**): In this scenario, the Sebtide-Alpujarride complex forms the basement of the Dorsale units which detach on the Carnian evaporites and form the accretionary prism together with the Flysch units during the north-dipping Oligocene-Miocene subduction

5.7.4.4 The Peridotite Emplacement

References: Literature concerning the Gibraltar Arc peridotites abounds. The most significant and essentially recent references are given directly in the following text.

The Gibraltar Arc peridotites are among the largest infracontinental mantle rock outcrops worldwide. Their emplacement mechanisms have been certainly more debated than if they were part of an ophiolite. Ophiolititic peridotites would have been regarded as dismembred thrust unit obducted during the early collisional stages of the Betic-Rif orogeny, comparable to the Alps and Apennine ophiolites. This manner of thinking was first adopted by Kornprobst (1974), and by Reuber et al. (1982) who interpreted the Beni Bousera peridotites as “cold” slivers of Jurassic infracontinental mantle included in the Cenozoic nappe stack. However, this “cold” tectonic interpretation was challenged as early as the eve of the 70s by a “hot” diapiric interpretation suggesting that the Ronda peridotites originate from a Neogene asthenospheric mantle diapir (Loomis, 1972; Obata, 1980). This “hot” emplacement theory was based, at that time, on poor geochronologic, petrologic and structural data, but still inspires a trend of thinking today.

The diapiric emplacement theory essentially rests on the late Oligocene-Early Miocene multimethod isotopic ages obtained from the ultrabasites and (mostly) from their country rocks, whose recrystallization would be allegedly caused by the peridotite emplacement. Moreover, the theory takes advantage of the occurrence of coeval mantle uplift recorded in the Alboran Basin area. In recent years, evidence was derived from the occurrence of a high temperature, low-pressure plagioclase peridotites associated with a recrystallization front toward the bottom of the Betic peridotite massifs (Van der Wal & Vissers, 1993; Lenoir et al., 2001; Tubía et al., 2004). According to Tubía et al. (2004), the Ronda peridotites would have a dual origin, including a cold sub-continental lithosphere (garnet and spinel domains), uplifted during the Mesozoic (cf. Reuber et al., 1982), and a hot (virtually intrusive) Miocene asthenosperic diapir.

However, Vissers et al. (1995) recognized that "there is no immediate causal relationship between the emplacement of the peridotites and the thermal metamorphic event [in the crustal rocks]: both rock bodies were affected by a thermal pulse at different levels". On the other hand, the diapiric emplacement theory has to consider the fact that the peridotites are included as relatively thin thrust sheets (less than 3 km thick) within the Alpujarrides crystalline nappe stack (Lundeen, 1978). Montel et al. (2000) suggested that a compressional event would have brought the diapir head on top of the adjoining continental crust immediately after the alleged diapir ascent (dated at 20–30 Ma). Platt et al. (2003b) and Tubía et al. (2004) recognized that the 25–20 Ma interval corresponds to a well documented regime of extensional tectonics in the Alboran domain, and then proposed that interleaving of the peridotite sliver within the crustal nappe stack would result from a complex sequence of extensional detachements. However, others authors emphasized that extensional tectonics alone cannot account for the early tectonic evolution of the Alpujarride-Sebtide nappe stack, particularly the occurrence of HP-LT recrystallization below and above the peridotites (Michard et al., 1991, 1997; Torné et al., 1992; Sánchez-Gómez et al., 2002; Chalouan & Michard, 2004). The emplacement model proposed by Sánchez-Gómez et al. (2002) includes the following steps during the Betic-Rif orogeny (i.e. after the Paleozoic-Mesozoic evolution of the Tethyan margin lithosphere): (i) pre-Miocene subduction-collision and HP-LT metamorphism; (ii) early exhumation of the Alpujarride crust, which produced its thinning and the rise of the peridotites up to c. 18 km depth; (iii) contractive emplacement of a peridotite slab onto the crust, recorded in the footwall of the thrust by a HT melange, overturned folds with HT crenulation cleavage, and a complete inversion of the Ojen-Blanca footwall unit in the HT-LP conditions of the plagioclase peridotites; and (iv) Miocene collapse of the Alboran Basin, with dismembering of the former peridotite slab, granite generation at 22–18 Ma, and plastic flow of serpentinites between the peridotite bodies. This scenario is compatible with the tectonic scheme of Fig. 5.50. Moreover, a close parallelism with the emplacement history of the Lower Penninic infracontinental peridotites of Western Alps can be remarked (e.g. Geisspfad Complex; Pastorelli et al., 1995; Bianchi et al., 1998).

5.8 Neotectonics, Seismicity and Present-Day Stress Field

References: The references concerning the Gibraltar Arc active tectonics are indicated within the following text. They are implemented by those associated with Chap. 1, Sect. 1.4.

The active deformation of the Gibraltar Arc has been evidenced years ago, based on the differential uplift of the Quaternary terraces (Cadet et al., 1977; Brückner & Radtke, 1986; Hillaire-Marcel et al., 1986; Zazo et al., 1999). Such very recent deformations and the present-day stress field itself result mainly of the ongoing Africa-Eurasia plate convergence (see Chap. 1, Sect. 1.4), with little (if any) contribution of the subduction process described above (Sect. 5.7.4). The stress field evolution from the late Miocene-Pliocene to the present-day, and the associated active fault structure have been discussed repeatedly in the Maghrebian region (e.g. Morel, 1989; Morel & Meghraoui, 1996; Aït Brahim et al., 2002; Bargach et al., 2004; Meghraoui et al., 2004).

Detailed structural observations demonstrate that very recent deformation occurred at the front of the Rif Belt. Recent shortening is clearly visible in the Rides Prérifaines (Fig. 5.31), where Plio-Quaternary conglomerates are locally verticalized (e.g. J. Tratt next to Fes city) and display pressure solution imprints recording the entire progression of the folding process (Bargach et al., 2004; Chalouan et al., 2006a). Further to the west, in the northern Gharb Basin near the Moulay Bouselham lagoon (Figs. 5.7, 5.46), high-rate (up to 14 mm/yr) uplift has affected the lagunal deposits since 2400 BP (Benmohammadi et al., 2007). This local deformation is likely related to argilokinetic deformation of the Prerif front, also evidenced offshore by mud volcanoes. Around the Strait of Gibraltar region in southern Iberia, uplift rates have been calculated for the Late Interglacial period, using marine terraces by Zazo et al. (1999). The deformation rates deduced from these studies must be regarded with caution, due to possible chronological bias (see Plaziat et al., Chap. 8, this vol.). Nevertheless, these authors showed that to the west of the Strait of Gibraltar, the terraces suffered differential uplift, whereas to the east (along the northern margin of the Alboran Sea) they were affected by subsidence. Coastal uplift would result from the Africa-Iberia convergence accommodated by conjugate NE-SW sinistral and NW-SE dextral strike-slip faults. These active faults connect in depth with scattered seismic swarms and merge into a shallow (12–9 km depth) brittle-ductile transition (Fernández-Ibáñez & Soto, 2008).

Available GPS measurements yielded further evidence for active deformation related to the ongoing plate convergence in the Alboran region (cf. Chap. 1, Fig. 1.17). Recent geodetical observations in permanent stations (Reilly et al., 1992; Fernandes et al., 2003, 2004; Fadil et al., 2006; Stich et al., 2006; Serpelloni et al., 2007) depict a broad view of the present-day velocity field of the region. Although its precise reconstruction deserves further studies, preliminary data suggest a W-to-SW motion of the front of the Gibraltar Arc in the Rif region. Most of the stations surrounding the Alboran Sea region show a distinctive motion with respect to the overall pattern of Africa-Eurasia convergence inferred from global geodetic models. These motions reflect independent motions of crustal blocks, bounded by active faults, which suggest active delamination processes (see also Chalouan et al., 2006a).

However, the dense seismic activity of the Alboran area (cf. Fig. 1.16) yields probably the most robust evidence for the ongoing deformation of the Gibraltar Arc (e.g., Grimison & Chen, 1986; Udías & Buforn, 1991; Rebai et al., 1992; Buforn et al., 2004; Stich et al., 2006; Serpelloni et al., 2007). Fernández-Ibáñez et al. (2007) have recently compiled four types of stress indicators (wellbore breakouts, earthquake focal plane mechanisms, young geologic fault slip data, and hydraulic fracture orientations) to reconstruct the present-day stress field of the Gibraltar Arc. They indicate a regional NW-SE compressive stress field resulting from Africa-Eurasia plate convergence (Fig. 5.51A). In some particular re-

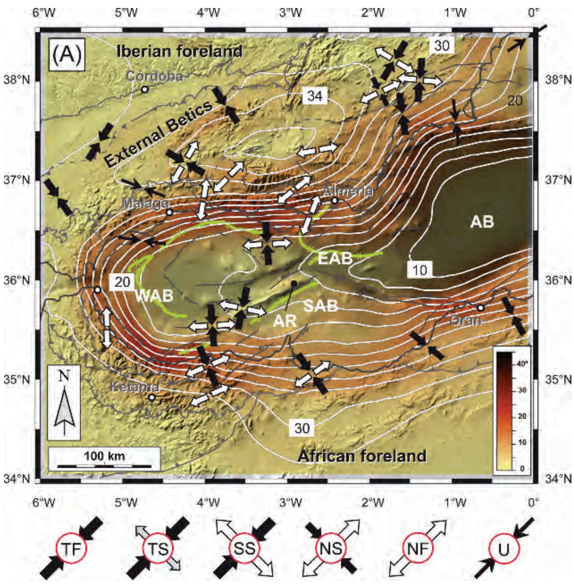


Fig. 5.51 The Present-day stress field of the Gibraltar Arc, after Fernández-Ibáñez et al. (2007). (**A**): Gradients of crustal thickness variation in the Gibraltar Arc and synthetic stress regime map. Moho contour lines (in km) are taken from Torné et al. (2000). AB, Algerian Basin; AR, Alboran Ridge; EAB, East Alboran Basin; SAB, South Alboran Basin; WAB, West Alboran Basin. Major sedimentary depocenters (*in green*). NF: normal faulting; NS: predominantly normal faulting with strike-slip component; SS: strike-slip faulting; TS: predominantly thrust faulting with strike-slip component; TF: thrust faulting; U: unknown stress regime. – (**B**): Tectonic sketch of the Gibraltar Arc showing the active fault structure, the associated stress rotations, and suggested mode of deformation partitioning within the Africa-Eurasia plate boundary. S_{Hmax} orientation and stress rotation with respect to the regional stress field are taken from Fig. 5.51A (*blue circles*, clockwise rotation; *red circles*, anticlockwise rotation; *white*, no rotation). Large arrows correspond to the Africa-Eurasia relative motion according to NUVEL-1A (DeMets et al., 1994). AD, Alboran Domain; ALF, Alpujarras fault zone; AF, Alhoceima fault zone; AMF, Alhama de Murcia fault; ARF, Alboran Ridge fault; CF, Carboneras fault; JF, El-Jebha fault; MF, Maro-Nerja fault zone; NF, Nekor fault; PF, Palomares fault; YF, Yusuf fault

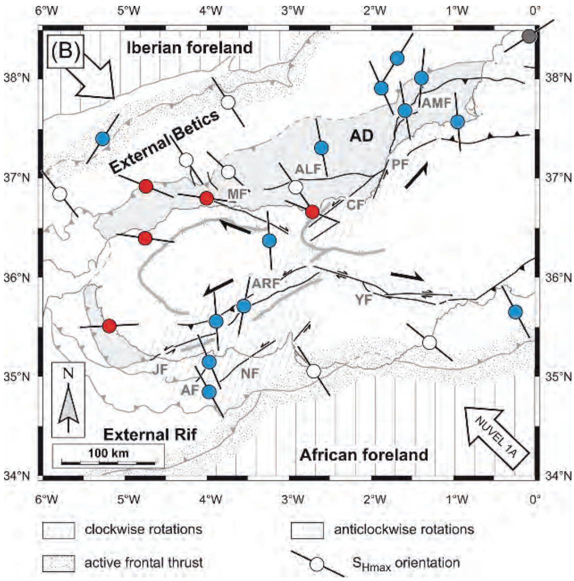


Fig. 5.51 (continued)

gions, deviations of S_{Hmax} are observed with respect to the regional stress field. They are gentle-to-moderate (22–36°) anticlockwise rotations located along the North Alboran margin and moderate-to-significant (36–78°) clockwise rotations around the Trans-Alboran Shear Zone (TASZ) (Fig. 5.51B). This is a broad fault zone, with a probable weak strength, and composed of different left-lateral strike-slip fault segments running from the eastern Betics to the Alhoceima region in the Rif and resulting in a major bathymetric high in the Alboran Sea (the Alboran Ridge fault zone). Some of these stress rotations appear to be controlled by steep gradients of crustal thickness variation across the North Alboran margin (Fig. 5.51A) and/or differential loading imposed by thick sedimentary accumulations in basin depocenters parallel to the shoreline. Other stress perturbations may be related to active left-lateral, strike-slip deformation within the TASZ that crosscuts the entire orogenic arc on a NE–SW trend and represents a key element to understand present-day deformation partitioning in the Western Mediterranean (Fig. 5.51B).

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Chapter 6
Atlantic Basins

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6.1 Introduction

References: For an introduction to the classical geology of the Atlantic coastal basins of Morocco, the reader is referred to Ambroggi (1963), Duffaud et al. (1966), Tixeront (1973), Uchupi (1976), and Brown (1980). The stratigraphy and structure of the Morocco passive margin has been described in the context of the DSDP and

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IPOD international research projects (e.g. Lancelot & Winterer, 1980; Vincent et al., 1980; Von Rad et al, 1982; Jansa & Weidmann, 1982; Stets & Wurster, 1982; Hinz et al., 1984; Winterer & Hinz, 1984; Jansa et al., 1984; Ruellan & Auzende, 1985; Heyman, 1989). The date and mechanism of the opening of the Central Atlantic Ocean, and the lithospheric structure of the margin are documented by Steiner et al. (1998), Olsen et al. (2003), Contrucci et al. (2003), Knight et al. (2004), Sahabi et al. (2004), and Maillard et al. (2006). Several unpublished thesis and published papers reviewed the stratigraphy and/or structure of certain segments of the margin based on industrial and research-generated geophysical and geological data. They are referenced in the next sections. Papers with the most general stratigraphic/structural

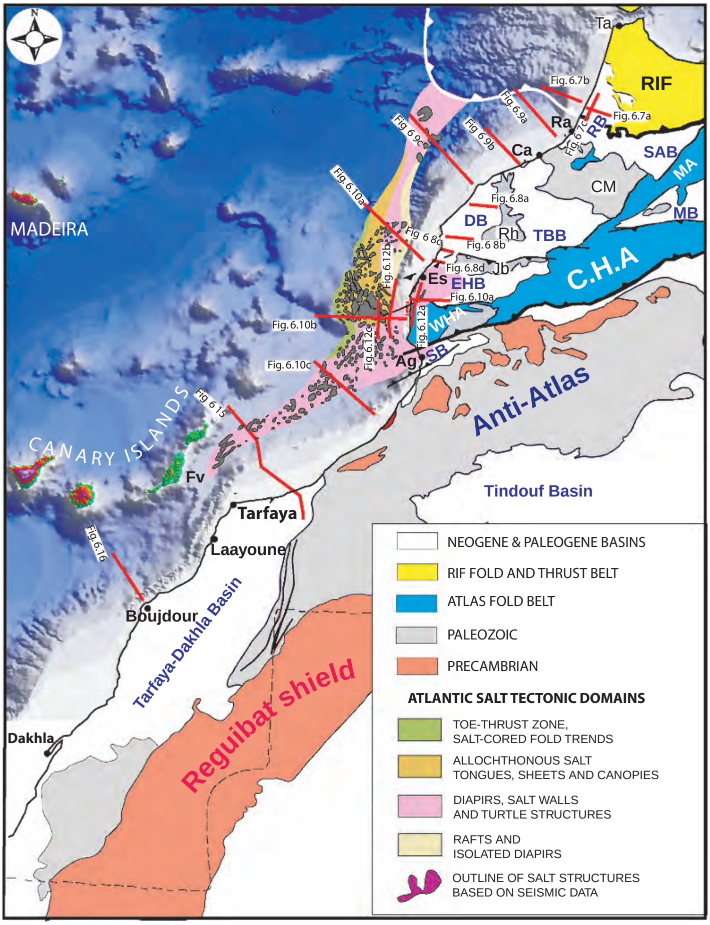

Fig. 6.1 Location map partly based on ONHYM internal reports and Tari et al. (2003). Bathymetry is a courtesy of Vanco. Ag., Agadir; Ca., Casablanca; C.H.A, Central High Atlas; CM, Central Morocco; DB, Doukkala Basin; EHB, Essaouira-Haha Basin; Es., Essaouira; Fv: Fuerteventura; Jb, Jebilet; MA, Middle Atlas; MB, Missour Basin ; Ra, Rabat; RB, Gharb Basin; Rh. Rehamna; SAB, Saiss Basin, SB, Souss Basin; Ta, Tangier; TBB, Tadla-Bahira Basin

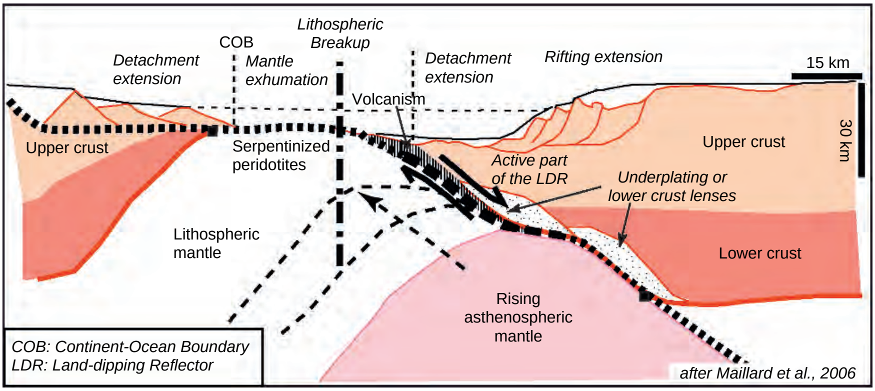

Fig. 6.2 2D-model of the opening of the central Atlantic Ocean between the conjuguate margins of Morocco and Nova Scotia at the pre-rupture stage (~ 195 Ma), after Maillard et al. (2005), modified. The model is build from seismic reflection and refraction data of SISMAR and SMART cruises. Note the asymmetry of the margins at the continent-ocean boundary (COB). Black squares are homologous points along the large detachment fault, whose latest active part corresponds to the imaged land dipping reflector (LDR)

purpose are those published by Le Roy et al. (1998, 2004), Tari et al. (2000, 2003), Le Roy & Piqué (2001), Zühlke et al. (2004), Tari & Molnar (2005),

The Atlantic margin of Morocco extends over nearly 3,000 km from Tangier in the north to Lagouira (La Gouira) in the south (Fig. 6.1). It is one of the oldest existing passive margins, and is conjugate to the Nova Scotia margin of North America (Fig. 6.2). Because of its key position within the Central Atlantic Ocean, and its promising hydrocarbon potential (Morabet et al., 1998; Davison, 2005) it was, and still is the object of extensive research programs and industrial exploration studies. These works reveal that the Moroccan Atlantic margin shows a broad homogeneity in its geodynamic evolution, with the exception of the segments adjacent to the Rif and Atlas ranges. This evolution can be subdivided into two main phases: a rifting phase, which started in the Late Permian-Triassic, followed by a drifting phase, which initiated in the Early Liassic around 195 Ma (cf. Chap. 1). In the present chapter we first review the main stratigraphic and structural features which characterize each of these two phases within the Moroccan Atlantic margin as a whole. Second, we describe the individual offshore and onshore basins which compose this margin, starting from north to south. The description is based essentially on regional seismic transects and borehole data.

6.2 Rifting

References: The following section relies on the papers by Hinz et al. (1982), Fietchner et al. (1992), Flinch (1993), Medina (1995), Tari et al. (2000, 2005), Le Roy & Piqué (2001), Roeser et al. (2002), Olsen et al. (2003), Maillard et al. (2006), Davison (2005), Hafid (2000, 2006), El Arabi E.H. (2007). Concerning the Central Atlantic Magmatic Province (CAMP), see Sebai et al. (1991), Hames et al. (2003), Olsen et al. (2003), Youbi et al. (2003), Marzoli et al. (2004), Knight et al. (2004), Sahabi et al. (2004).

Before the opening of the Atlantic Ocean, a Late Permian to Triassic/Early Jurassic continental rifting phase resulted in the formation of what was to become the Moroccan Atlantic margin, i.e. a system of predominantly NE-trending half-grabens, linked by E-W striking transfer faults on top of the Variscan continental crust (cf. Chaps. 1 and 3). The Moroccan margin was located in the upper plate of the asymmetric rift that opened between this eastern margin and the conjugated Canadian margin of Nova Scotia (Fig. 6.2). In the onshore basins, where they have been mapped from seismic and surface studies, most of these structures are bounded by northerly striking normal faults that mimic the structural grain of the underlying basement (compare Fig. 6.3 with Fig. 3.16). The half-grabens were filled by at least 2000 m of continental red beds overlain by basalt flows and silts. This infilling crops out in the so-called "Couloir d'Argana" (cf. Chap. 4, Fig. 4.6), where four tectonostratigraphic sequences (TS) have been recognized (Olsen et al., 2003). The lowermost sequence (TS I) is dated from the Late Permian by its Vertebrate fossils; the intermediate sequences (TS II, TS III) are assigned to the Carnian and Norian based on palynology (although the occurrence of the Middle Triassic is likely at the bottom of TS II as in Central High Atlas; El Arabi E.H., 2007), and the basaltic sequence (TS IV) is dated from the Rhaetian-Hettangian to lowermost Sinemurian by isotopic results. This stratigraphic record reflects the combined effects of tectonic pulsations (rifting episodes) and climatic changes on sedimentation and can be correlated with sediments of the equivalent basins of North America. In the offshore

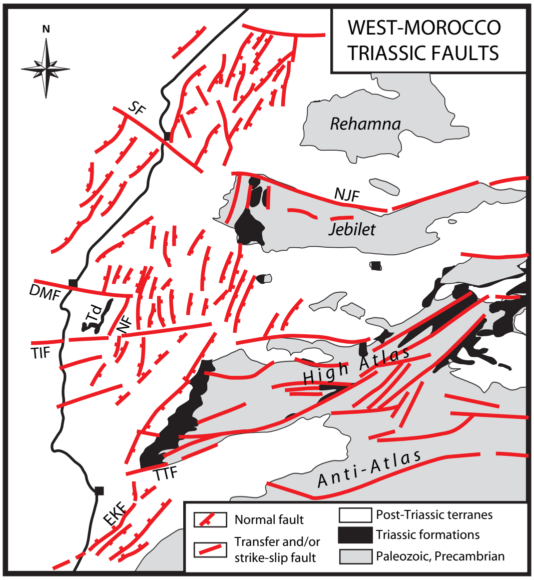

Fig. 6.3 Pattern of upper Triassic normal and transfer faults from Western Meseta to Souss Basin, after Hafid et al. (2005). DMF: Diabet-Meskala fault; ELF: El Klea fault; NF: East Necnafa fault; NJF: North Jebilet Fault; SF: Safi fault; Td: Tidsi salt diapir; TIF: Tarhzhout-Ichech fault; TTF: Tizi n'Test fault

beyond the shelf break, the early syn-rift half-grabens are generally deeper than recorded by most of the old seismic data and have been rarely drilled. Evidence from recently acquired sections suggests that they trend similarly and yield thicknesses comparable to their onshore equivalents (Tari et al., 2005).

Syn-rift salt presently forms extensive diapiric provinces along both Eastern and Western North Atlantic margins (Fig. 1.8). The eastern diapiric province extends along the Moroccan Atlantic margin over some 900 km. Its northern edge is offset and overlapped by the Rif-Betic thrust belt and its eastern limit makes an inland incursion into the Essaouira-Agadir Basin (Fig. 6.1). Seismic evidence shows that it extends southward into the Tarfaya-Laayoune Basin, and its western limit roughly coincides with the S1 magnetic anomaly (Fig. 1.8) which is interpreted by several studies as marking the ocean-continent boundary (OCB; Fig. 6.2). As illustrated below, mobile syn-rift salt strongly influenced the post-rift sedimentary and structural evolution in this province. The original thickness of the salt layer probably exceeded 1.5 km judged by the size and frequency of the salt diapirs (Davison, 2005).

Seismic, surface and borehole evidence indicates that toward the end of salt deposition high magmatic and volcanic activities occurred in the Moroccan Atlantic basins and resulted in extensive basalt flows that were emplaced at the Triassic/Jurassic boundary (Figs. 4.6, 4.7). These volcanics immediately pre-date the spreading stage of the Central Atlantic and characterize the Central Atlantic Magmatic Province (CAMP; cf. Chap. 1, Fig. 1.6). In Morocco, CAMP magmatic events range in age from 203 to 197 Ma (Knight et al., 2004; Marzoli et al., 2004). Recent studies suggest that the CAMP volcanics were extruded or intruded over a very large area in both sides of the central Atlantic during a brief magmatic activity (less than 1 Myr) that centred around 200 Ma, i.e. at the Rhaetian/Hettangian boundary. In Morocco, they consist essentially of subaerial lavas and pyroclastics interbedded with shallow-water clastic/evaporitic sequences and yielding a geochemistry that is compatible with a lithospheric mantle source (Youbi et al., 2003).

6.3 Drifting and Further Tectonic Interferences

References: The age of earliest inception of Atlantic drifting was discussed by Jansa (1981), Steiner et al. (1998), Medina et al. (2001), Sahabi et al. (2004), Davison (2005). The main recent references for the further evolution of the margin are Le Roy et al. (1997, 1998), Le Roy & Piqué (2001), and Hafid et al. (2000, 2006). Detailed data on the Jurassic and Cretaceous sediments can be found in Bouaouda (1987, 2004), Stets & Wurster (1982), Einsele et al. (1982), Du Dresnay (1988), Stets (1992), Algouti et al. (1999), Labbassi et al. (2000), Nouidar & Chellai (2002), Kolonic et al. (2005).

The onset of spreading in the central Atlantic is not well constrained. It is placed in the Early Toarcian based on field observations of Middle Aalenian to Bajocian sediment lying unconformably on oceanic basalts on Canary Islands (Fuerteventura; Steiner et al., 1998). Using a spreading rate of 43 mm/yr and considering the 30 km distance that separates this outcrop from the estimated position of the ocean–continent boundary, Davison (2005) estimated that the oldest oceanic crust of the Atlantic margin of Morocco would have formed 0.7 Myr before the ocean crust on eastern Fuerteventura. Hence, the oldest possible age of this crust would be 178.7 Ma, and the youngest would be about 170 Ma according to this author. Based on the adjustment of North American and Northwest African marginal magnetic anomalies, which he considers to mark the end of salt deposition, Sahabi et al. (2004) estimated that the first expansion occurred earlier, at about 195 Ma (Sinemurian).

The onset of drifting in the central Atlantic coincides with the establishment of a widespread carbonate platform that extended from Portugal to Guinea Bissau (Jansa, 1981). In the Moroccan margin this platform involved shallow marine limestones, dolomite and anhydrite that were deposited near-shore covering the syn-rift sequence (Fig. 6.4) as well as deeper water carbonates further offshore (Fig. 6.5). Predominantly open marine transgressive conditions prevailed until carbonate deposition terminated with the early Berriasian global sea-level fall which continued throughout Valanginian to Hauterivian time. This resulted in certain areas in the development of prograding delta systems (e.g. Tan-Tan delta). From the Barremian to the end of the Cretaceous sedimentation was associated with a steadily rising sea level which resulted in deposition of extensive marine shales and carbonates (Figs. 6.4 and 6.5).

The Atlas and Rif Mountains are Alpine orogenic belts which resulted from the converging motion between Africa and Eurasia from Late Cretaceous times onward. A synopsis of their geological history is given in chaps. 4 and 5, respectively. Both of

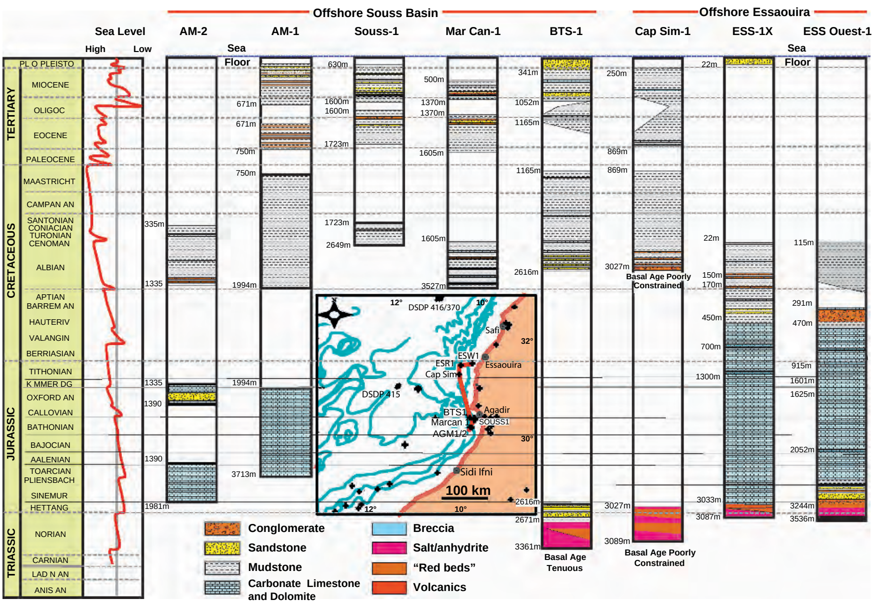

Fig. 6.4 North-South chronostratigraphic correlation of logs of industrial wells drilled in the Essaouira-Agadir shelf segment of the Atlantic Margin of Morocco

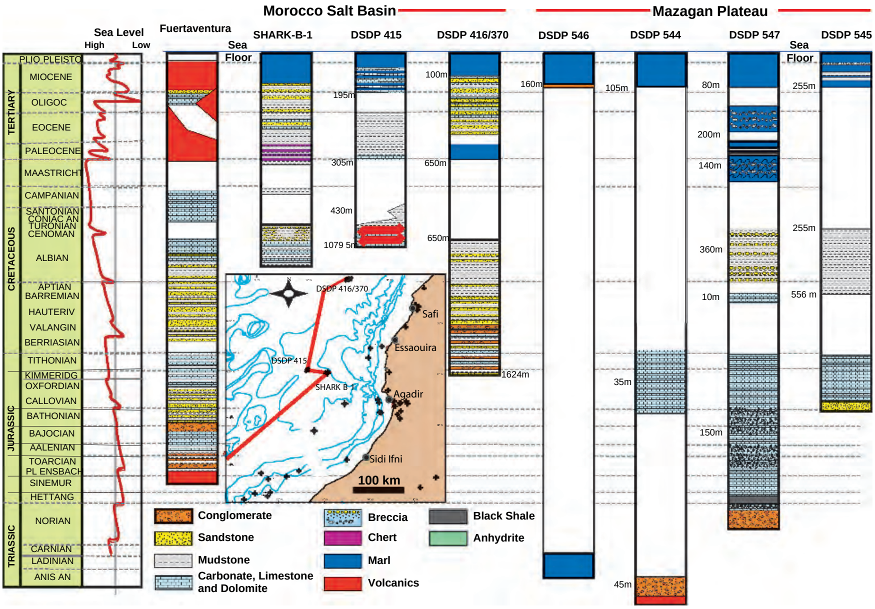

Fig. 6.5 North-South chronostratigraphic correlation of logs of DSDP wells and SHARK-B-1 well drilled in the El Jadida-Agadir deep basin of the Atlantic Margin of Morocco with the outcrop section of Fuerteventura, Canary Islands

these fold belts extend westward onto the Atlantic margin (Figs. 1.2, 4.1 and 5.5) introducing thereby several interferences in the classical evolution scheme of a typical passive margin. Below the main features that have resulted from these interferences are shown on selected seismic transects.

6.4 Regional Transects Across the Atlantic Margin and Coastal Basins

In the following sub-sections we present the specific structure of the Moroccan Atlantic transects from North to South (see Fig. 6.1 for location).

6.4.1 Gharb-Mamora Segment

References: The recent literature concerning the structure of this segment includes the papers by Flinch (1993, 1996), Litto et al. (2001), and Zizi (2002). The Neogene stratigraphy has been studied by Feinberg (1986), Wernli (1988), Benson et al. (1991), Kerzazi (1994), and Sefiani et al. (2002).

The Gharb (Rharb)-Mamora and Saiss Basins form a roughly E-W trending Late Miocene to Pleistocene trough that covers the western part of the so-called “Couloir

Sud-rifain” (South Rifian Corridor). This trough evolved as a foredeep that separates the Rif fold-thrust belt from its foreland to the south (Meseta and Middle Atlas) (Fig. 6.1). It was a 500 km long E-W marine channel that connected the Atlantic with the Mediterranean during the Miocene. Following the closure of the strait in Taza area at the end of Miocene, it became an Atlantic gulf. On the southeast border of the Gharb Basin the foredeep is bounded by the “Rides Prérifaines” (Fig. 5.7) that are made up of Jurassic sediments deposited in extensional systems inverted during the Pliocene-Pleistocene (Figs. 5.31, 5.32).

6.4.1.1 Stratigraphy

The stratigraphy of the Gharb-Mamora and Saiss Basins can be subdivided into pre-foredeep and foredeep successions (Flinch, 1993). The pre-foredeep corresponds to the infra-nappe formations that include Triassic sandstones and shales unconformably overlying the Hercynian basement and overlain by Upper Cretaceous sediments in the onshore Gharb-Mamora Basin and by Jurassic to Upper Cretaceous sediments to the West in the Atlantic margin. To the east the Mesosoic section underlying the Saiss Basin essentially consists of Jurassic carbonates and evaporites. The pre-foredeep section is topped by the so-called “Molasse de base” which consists of Middle Miocene (?)-Tortonian conglomerates, sandstones, sandy limestones and marls that transgress Paleozoic or Triassic formations of Meseta (Fig. 6.6).

The foredeep succession consists of Upper Miocene (mainly Messinian) to Pliocene sediments that unconformably overlie the pre-foredeep succession by a

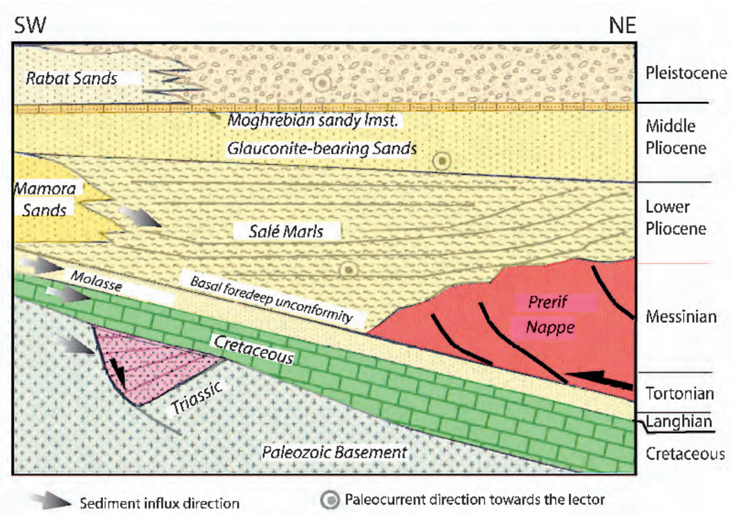

Fig. 6.6 Idealized Rif foredeep subdivision into local stratigraphic units, after Flinch (1993), slightly modified

basal unconformity, and can be up to 3000 m thick in the northern parts of the basin. They consist of a deep-water pelagic facies overlain by littoral deposits and alluvial-fluvial deposits. The Prerif Nappe was emplaced within the lower part of the foredeep succession (Fig. 6.6). It consists of Triassic to Miocene sediments whose stratigraphy is obscured by complex, compressional and gravity-driven tectonics (Feinberg, 1986). The deep-water foredeep sediments are represented by Upper Tortonian-Messinian to Lower Pliocene pelagic blue marls (“Marnes de Salé”) that grade southwardly towards the southern basin margin into shallower coarse sediments (“Sables de Mamora”) and are overlain by Middle Pliocene glauconitic or ferruginous sandstones (“Sables verts glauconieux”, “Sables fauves” in the Saiss Basin) that grade eastward into lacustrine Upper Pliocene carbonates (“Calcaires du Saïss”), and westward into alluvial sediments.

6.4.1.2 Structure

Two regional NW-SE onshore and offshore sections and a N-S onshore section across the Gharb Basin (Fig. 6.7) allow us to make the following remarks with respect to the structural styles: (1) the basin fill shows three major seismic units corresponding to the Pre-Nappe, the Nappe and Post-Nappe units (2) the Pre-Nappe unit which consists of Paleozoic rocks and their thin Mesozoic-Cenozoic Infra-Nappe cover is characterized by a gentle northerly dip. The upper part of this unit shows normal faults and its thickness increases towards the Atlantic (3) the Prerif Nappe unit is characterized by a northward thickening wedge-like geometry with southward vergent imbricates (4) the top of the Nappe is offset by extensional faults that are locally associated with frontal thrusts and that controlled the deposition of the Post-Nappe sediments.

The northward dip of the Pre-Nappe unit reflects the flexural response to loading the northern margin of the basin by the Rif frontal thrusts, and to slab pull from the subducting Tethyan lithosphere (see Chap. 5). The normal faults and related tilted blocks that are seen in the upper part of this unit are probably due to the Triassic rifting as it is strongly suggested by similar structures seen under the shelf margin offshore the Gharb Basin associated with other seaward-dipping normal faults (Figs. 6.7b and 6.8). These Triassic structures are overlain by a relatively thin Jurassic and Cretaceous succession similar to that of the El Jadida margin (see below). The Prerif Nappe unit is a tectono-sedimentary complex consisting of chaotic mixture of Triassic to Neogene sediments. The emplacement of this allochthonous complex within the basin is classically understood to be the result of several phases of deformation and gravitational tectonics due to unstable uplifted portions of the Rif uplands to the north (Michard, 1976; Feinberg, 1986). A more recent work based on subsurface data interprets the Prerif Nappe rather as a frontal accretionary wedge of the Rif thrust belt that is deformed from below due to late thrusting which involves the underlying passive margin successions (Flinch, 1993).

The supra-Nappe sediments involve compressional-extensional and extensional satellite basins trending parallel and perpendicular to the arc. Contrary to

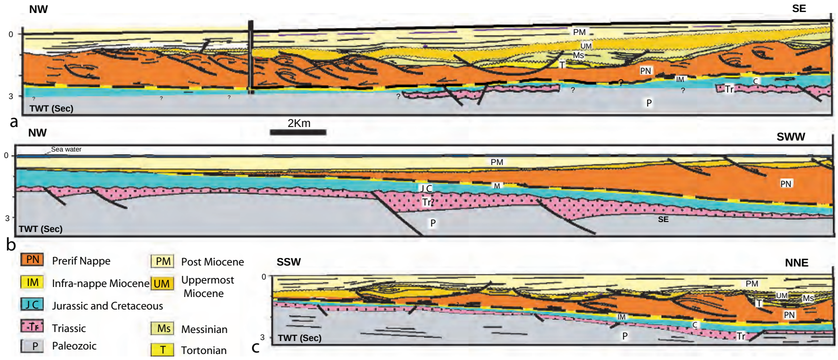

Fig. 6.7 Line drawings of seismic transects across the Gharb Basin onshore (a and c, after El Moussaid, 2007) and offshore (b, interpretation of M. Dakki). See Fig. 6.1 for location

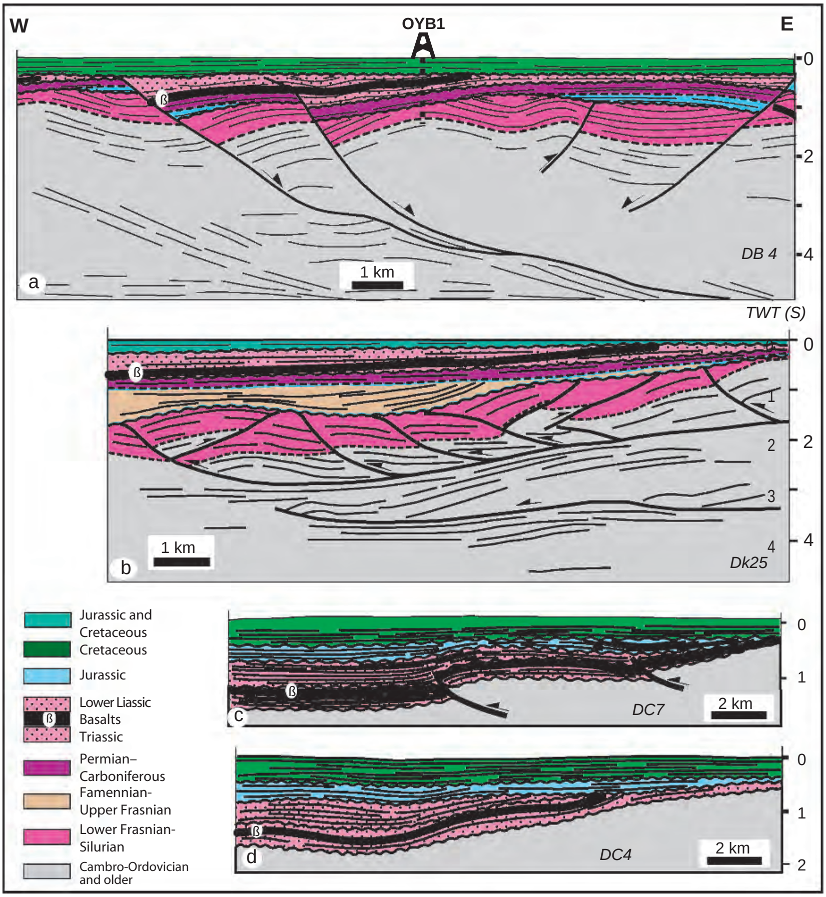

Fig. 6.8 Line drawings of seismic transects across the onshore Doukkala (**a, b**) and Abda (**c, d**) Basins, after Echarfaoui et al. (2003). See Fig. 6.1 for location

Flinch (1993) who interpreted the compressional and extensional structures as coeval, Litto et al. (2001) suggested two deformation episodes: an extensional Tortonian-Messinian episode followed by a compressional Plio-Quaternary episode (Fig. 6.6). The study of the post-nappe synclines at the north-eastern border of the Gharb Basin also leads to controversial interpretations (cf. Chap. 5). However, the association of compression and gravity-driven structures must be kept in mind, whatever their relative chronology. Finally, in the offshore, the rifting/drifting structures of the margin are generally difficult to image on the seismic data, as Mesozoic strata become involved northward in the Rif thrust sheets.

6.4.2 Rabat-Safi Segment

References: This segment has been particularly investigated by the Cyamaz Group (1984), Ruellan & Auzende (1985), Jaffrezo et al. (1985), and more recently by Echarfaoui et al. (2002a, b), as well by ONHYM (unpublished report, 1996). Le Roy et al. (2004) have dedicated a study to the Miocene-Pleistocene evolution of the continental shelf offshore El Jadida.

The basement of this segment is made up of folded Paleozoic rocks of the Coastal Block of the western Hercynian Meseta which crop out in its northern onshore part. This onshore part is mostly occupied by the Doukkala Basin where as its northern and southern offshore parts are termed by petroleum geologists the Casablanca Offshore and the Safi basin, respectively (Morabet et al., 1998; Tari et al., 2003) (Fig. 6.1).

6.4.2.1 Stratigraphy

Seismic and borehole data indicate that the Doukkala Basin is underlain by Triassic, Jurassic and Cretaceous formations that unconformably overlie a Paleozoic basement that was essentially structured during an early Variscan deformation phase (Echarfaoui et al., 2002) (Fig. 6.8a & b). Triassic Atlantic rift related extension structures opened on top of this basement along northerly striking normal faults that sometimes used Hercynian reverse faults (Fig. 6.3). As in the rest of western Morocco, these half-graben structures were filled with fluvio-lacustrine deposits containing evaporites and interbedded volcanics. The sequence attains *ca.* 1,400 m in the troughs and consists of four units: (1) a thin basal alluvial fan conglomerate changing laterally to fluvial sandstones (2) a lower shale unit consisting of an argillaceous series with sandstones, gypsum, anhydrite and interbedded salt (3), a basalt unit consisting of vuggy to tight volcanic flows and (4), an upper shale unit consisting of massive saline series containing shale interbeds and capped by red-brown shales (ONHYM, 1996). The Jurassic-Cretaceous cover that unconformably overlies this syn-rift series is thin relative to its equivalent further south in the Essaouira-Agadir Basin. The Jurassic is represented by supratidal dolomite and anhydrite that thicken gradually westward and change into more open marine facies. The Jurassic carbonate platform was drowned by the Cretaceous clastics that are encountered across most of the basin and extend into the offshore.

6.4.2.2 Structure

Figure 6.9a is a line drawing of a seismic section that illustrates one of several large Triassic syn-rift structures (the Oulad Yacoub graben), mapped from existing seismic data in the onshore Doukkala Basin (Fig. 6.3). As can be noted in this section, some of the normal listric faults that bound the rift structures are rooted in older Hercynian reverse faults and the post Triassic cover is virtually undeformed.

Farther south in the Abda Basin (southern part of the Doukkala Basin *sensu lato*, east of Safi), Fig. 6.8c and 6.8d show a clear flexuration of the Upper Triassic to Lower Jurassic units with westward divergent wedges and mild reverse faults. This suggests that the Abda Corridor was subjected, during this period, to a compressional uplift of the Paleozoic underlying the Mouissat-Jebilet area to the east of the "West Meseta flexure or Lineament". This uplift is related regionally to the "Dorsale du Massif Hercynien Central" (cf. Chap. 4). It can be explained either as due to the uplift of the Atlantic rift shoulder or simply as a local accommodation of extension to strike-slip movements along E-W-oriented transfer faults. A third option is the local compression generated in the rift margin by differential cooling near and away from the mid-oceanic upwelling ridge (Echarfaoui et al., 2002).

No industrial wells were drilled in the Rabat-Safi offshore segment. The only direct lithostratigraphic information available for this area is obtained from DSDP leg 79 holes (544, 545, 546, 547; Fig. 6.5) and from Cyana submersible sampling. This information indicates that the basement of this segment consists of Precambrian and Paleozoic rocks overlain by a thick Upper Triassic–Lower Liassic evaporitic series.

Figure 6.9 shows line drawings of three seismic transects that illustrate the structure of the Rabat-Safi offshore. The northern transect (6.9a) is a NW-SE section across Rabat offshore south of the external limit of the Prerif nappe of which it slightly crosses the edge at the NW extremity (Fig. 6.1). The basement, which consists of Paleozoic rocks probably similar to those known from the nearby coastal zone, is not resolved in this section. However, its upper surface shows vertical offsets due to the presence of syn-rift structures bounded by both eastward- and westward-dipping normal faults. They are probably filled with siliciclastic and evaporitic series that characterize the West Moroccan Triassic as we have seen above. The Jurassic shows a westward dipping carbonate ramp structure whose base seems to have sealed the normal faults. Its thickness decreases both eastward and westward, and its upper surface is deeply cut by the basal Cretaceous unconformity.

Transect 6.9b shows the shelf area of the Casablanca offshore with a relatively thin Mesozoic cover overlaying a deformed Paleozoic basement characterized by broad compressive structures similar to those seen in the Coastal Block outcrops to the east and in the subsurface of the Doukkala Basin to the south (Fig. 6.8a). Further south, the much longer transect (Fig. 6.9c) illustrates the transition from the shelf of the Mazagan Plateau to the oceanic basin. The syn-rift tilted blocks that characterize this transition are similar to those seen in the Rabat offshore but here the distension clearly continues into the Jurassic and the post-rift sedimentation and structure are strongly controlled in the deep basin by halokinesis. This is attested by the presence in the western part of the section of three large salt domes surrounded

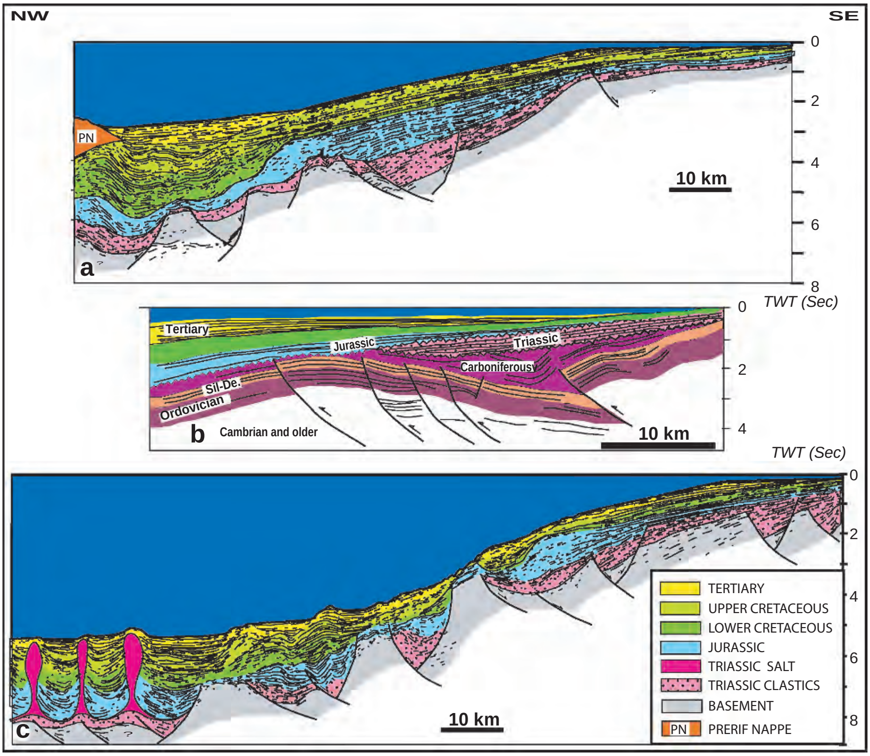

Fig. 6.9 Line drawings of seismic transects across the Rabat offshore (**a**), Casablanca offshore (**b**) and Mazagan Plateau (**c**). (**a**) and (**c**) after Tari et al., 2003; (**b**): interpretation of A. Ait Salem. See Fig. 6.1 for location

by Jurassic to Tertiary, strongly subsiding mini-basins filled with a 4–5 km thick stratigraphic sequence of Jurassic to Pliocene age. The localisation of these domes above tilted Triassic blocks suggests a genetic relationship between both types of structures. Salt growth started at least from Jurassic time onward (see below).

Transect 6.10a crosses the Safi offshore and shows deep seated extensional systems similar to those drilled by the DSDP program on the Mazagan plateau (e.g. Winterer & Hinz, 1984) or mapped on the southern continuation of that plateau (Ruellan, 1985; Ruellan & Auzende, 1985). Above these syn-rift structures one can see the characteristic halokinetic elements of a passive margin, such as extensional salt structures beneath the shelf and the upper slope, and compressional features downdip on the lower slope. The upper slope shows several raft-like features sliding downdip on the salt accommodating significant extensional strain. Updip correlation of the Mesozoic stratigraphy encountered in DSDP well 416 (Lancelot & Winterer 1980) indicates that the stratigraphic sequence within the rafts includes Jurassic shallow and then deeper water carbonates, and Lower Cretaceous shales. The overlying beds display progressive growth which dates the inception of rafting as Middle Cretaceous. The major unconformity truncating the rafted sequence is interpreted to be intra-Tertiary, possibly mid-Oligocene in age. The rafted domain up dip is separated from the allochthonous salt tongues and sheets by a narrow zone of turtle structures. The most important observation is that all the salt structural domains appear to be linked to the salt detachment. Based on the expression on the present-day seafloor, the salt tongues at the basin-ward edge of the salt system are still active. Most of this salt basin lies in water depths greater than 2000 m.

6.4.3 Essaouira-Agadir Segment

References: This segment was repeatedly studied onshore and offshore. The main recent references dedicated to this particular transect are Bouaouda (1987), Heyman (1989), Stets (1992), Broughton & Trepanier (1993), Medina (1994, 1995), Amrhar (1995), Mustaphi et al. (1997), Labbassi et al. (2000), Frizon de Lamotte (2000), Hafid (2000), Mridekh et al. (2001), Mridekh (2002), Hafid et al. (2000, 2006), and Hafid (2006).

The most special feature of the Essaouira-Agadir segment is its localization in an area where the High Atlas fold belt intercepts the Atlantic passive margin. The geological evolution of this large area can be subdivided into three main stages that are : (i) the Syn-Rift stage from Middle-Late Triassic to Early Liassic; (ii) the Pre-Atlasic Post-Rift stage from late Early Jurassic to Mid-Late Cretaceous, and (iii) the Atlasic Post-Rift stage from Mid-Late Cretaceous to the Present.

The onshore part of this segment which corresponds to the Essaouira-Haha and Souss Basins are treated with the Western High Atlas (cf. Chap. 4.). Figure 4.5 shows a synthetic stratigraphic log of the Essaouira Basin. Except for the salt tectonics, which will be analysed for the entire onshore/offshore area, we focus here on the offshore zone which corresponds to the Essaouira and Agadir offshore basins of petroleum geologists.

6.4.3.1 Syn-rift Sequence

As stated above, during the syn-rift stage, extensional depocenters opened on top of a peneplained Paleozoic basement, being controlled by N- to NNE- striking normal faults laterally offset by E-W trending transfer faults (Fig. 6.3). The formations that have filled these depocenters can be subdivided, both in the surface and in the subsurface, into at least three tectonostratigraphic sequences (Figs. 4.5, 4.33) that are bounded, in the onshore, by mainly eastward dipping faults, and in the offshore by mainly westward dipping faults (Fig. 6.10b).

Seismic profiles show that the uppermost sequence was deposited, probably during the transitional phase between rifting and drifting in a continuous, considerably less faulted, salt-rich sag basin with extensive basalt flows (Fig. 4.16). Most of the salt presently encountered at different structural levels within the studied Western

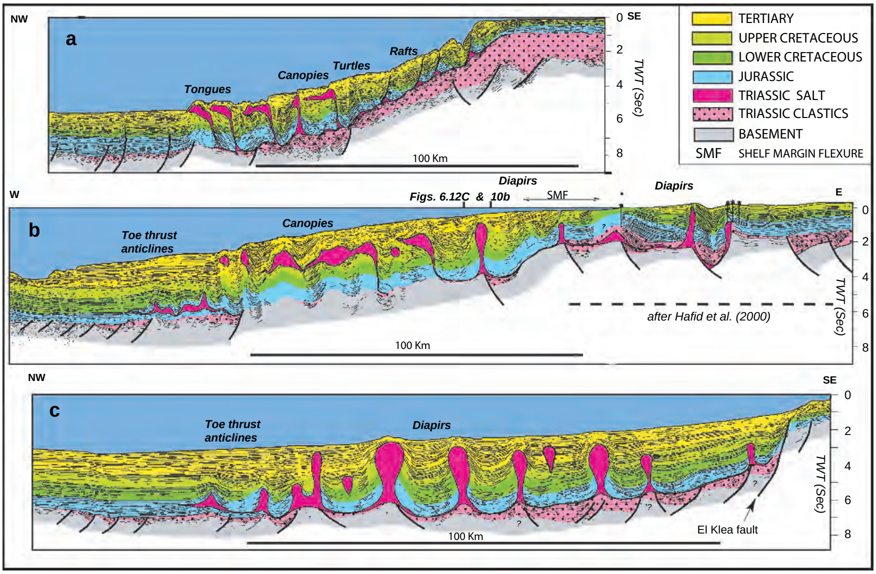

Fig. 6.10 Line drawings of seismic transects across the Safi offshore (**a**), the Essaouira on-shore/offshore (**b**) and the Ifni offshore (**c**) basins. After Tari et al. (2003), modified. See Fig. 6.1 for location

Morocco basins was deposited in this shallow evaporitic sag basin. The occurrence of a possible direct connection of this Atlantic salt basin with the Tethyan salt basin east of the West Moroccan Arch (Fig. 4.7A) is presently in debate.

6.4.3.2 Pre-Atlasic Post-rift Sequence

The post-rift stage started with the onset of spreading in late Early Jurassic times by the establishment of a widespread carbonate platform where shallow marine limestones and dolomites were deposited near-shore over the syn-rift sequence and deeper water carbonates farther offshore. Open marine transgressive conditions prevailed until the carbonate platform ended by the early Berriasian global sea-level fall which continued throughout Valanginian to Hauterivian time. By the end of this, period the Barremian sedimentation was associated with a steadily rising sea level which resulted in deposition of extensive marine shales and carbonates (Figs. 4.35, 6.4 and 6.5).

6.4.3.3 Atlasic Post-rift Sequence

A NNW-SSE compression affected the Essaouira-Agadir Atlantic margin segment starting in Late Cretaceous and reaching its paroxysm in Neogene times. The effects of this compression at the scale of the entire Atlas System are reviewed in Chap. 4. Here we focus on the perturbations which overprinted the classical profile of the passive margin. The most important of these perturbations are illustrated by two regional transects: an E-W transect that extends from the Chichaoua onshore platform to the DSDP 415 well in the deep offshore basin (Fig. 6.10b) and an N-S transect that crosses the Cape Tafelney offshore basin (Fig. 6.12b).

The near-shore platform appears very distinctly in these profiles separated by a shelf margin flexure (SMF) from a southward-dipping post-Jurassic flexural basin: the Cape Tafelney Basin. The SMF is characterized by steep dips and deeply eroded layers suggesting a pronounced Neogene compressional flexure coeval with the Tertiary formation of the Atlas Mountains. The Cape Tafelney Basin is bounded to the north and south by two E-trending transfer faults which correspond to the Diabet-Meskala fault and the offshore branch of the El Kléa fault, respectively (Fig. 6.3, Fig.6.10c and Fig.6.12b). The Cape Tafelney Basin was filled mostly by thick northward wedging Upper Cretaceous to Tertiary marly/clastic sediments. The folds trend NE-SW and form the Cape Tafelney High Atlas (CT in Fig. 6.11) that corresponds to the tip of the Atlas System at its intersection with the Atlantic margin. The SMF is interpreted as the expression of a complex set of lateral ramps between a salt-based fold belt on top of an oceanic and/or transitional crust (the Cape Tafelney High Atlas) from a coastal to onshore, basement-controlled fold belt characterized by complex inversions of Triassic-Upper Jurassic extensional systems (the onshore Western High Atlas). The basal décollement level of the Cape Tafelney folds is located above syn-rift salt and “ramps” going down into the Paleozoic basement of the coastal basins to the east (Fig 6.10b). On a regional scale, this décollement level can be linked, in the North, to the J. Hadid-J. Kourati reverse fault which connects with the North Jebilet reverse fault and, farther east, to the North Atlas frontal ramp. In the South, it connects the Agadir-Souss basin to the South Atlas front (El Kléa and Tizi n’Test faults) (Fig. 6.11 and Fig. 4.35a). All these major faults were interpreted by Hafid et al. (2000) and Hafid (2006) as merging into a mid-crustal decoupling level revealed by previous geophysical studies (Giese & Jacobshagen, 1988).

6.4.3.4 Salt Tectonics

As stated above, the Essaouira-Agadir Basin is the only segment of the Moroccan Margin where the Atlantic salt basin makes an incursion into the onshore. Figure 6.11 shows the structural map of the main mobile salt bodies as roughly mapped in this segment from the available seismic and well data. The patchy distribution of these bodies suggests that the thickest salt is located above half-graben structures (Tari et al., 2003). Onshore, the Necnafa area contains the main salt domes and diapirs (e.g. Tidsi) and also the thickest Jurassic and Cretaceous sections in the basin. It was shown that this local thickening, which was well mapped since the early work of Duffaud et al. (1966), can be explained by differential subsidence due to salt withdrawal (Fig. 6.10b) (Hafid et al., 2000a; Hafid, 2006). Seismic evidence,

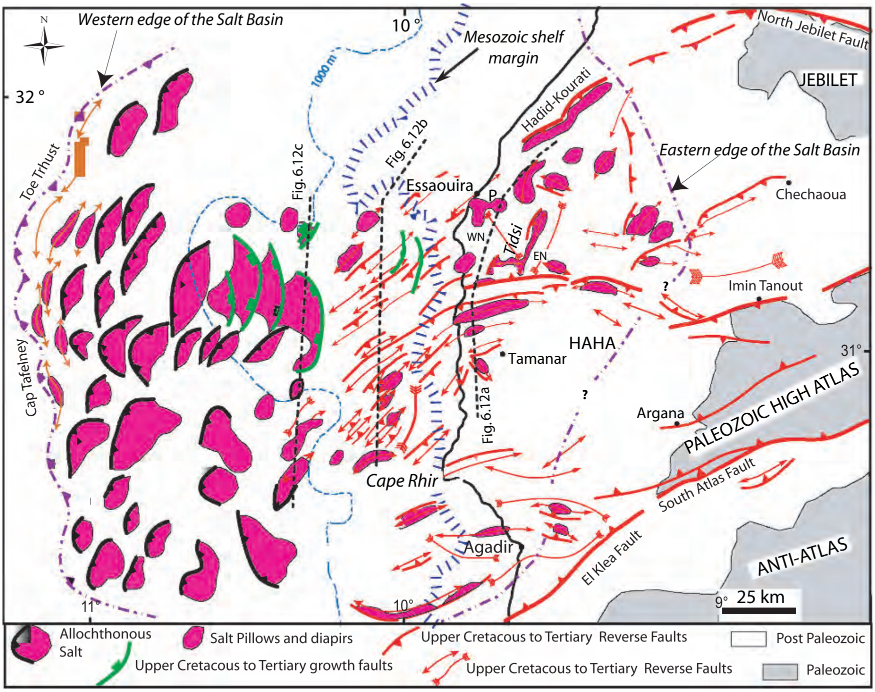

Fig. 6.11 Structural sketch and salt map of the western termination of the Jebilet-High Atlas system and of the Essaouira-Cape Tafelney segments of the Moroccan Atlantic Margin, after Hafid (2006), modified. EN: East Necnafa; P: Palmera; WN: West Necnafa

however, is not good enough to exclude the possibility that for the Jurassic, this thickening can also be partially due to renewing the extension as suggested by some earlier works (Medina, 1994; Amrhar, 1995). In areas where salt is thinner (e.g. Meskala), there is some evidence that salt started to move before the onset of the drifting stage. On the other hand, in some other areas (e.g. OTA well), salt is still interbedded within the section.

The presence of a thick salt layer within the section also controlled the structural styles of the Atlas compression in several important ways:

- In areas where the original salt layer was not very thick (e.g. eastern onshore Essaouira, Fig. 4.33A), salt was squeezed into the cores of the anticlines.
- In the case of inverted Triassic-Liassic half-grabens, such as the Kourati-Hadid structure, salt was obliquely injected upward along the reverse fault planes.
- In the case of well-developed pre-existing diapirs (e.g. Tidsi diapir), compression simply reactivated them as compressional folds sometimes associated with westerly verging reverse faults (Figs. 4.33B and 6.10b).
- In the Essaouira offshore the syn-rift salt acted as a detachment level above which the Cape Tafelney folds formed by thin-skin shortening (Hafid et al., 2000). Salt was squeezed upward along the reverse fault planes into inclined salt anticline

cores, but it seems to stay in normal stratigraphic contact with the overlying strata, except where pre-Atlasic well-formed diapirs existed (e.g., south end of Fig. 6.12b).

- The previously formed slope structures were reactivated by the Late Cretaceous to Tertiary Atlasic compression which squeezed the salt upward to form large west-verging salt tongues. These salt tongues coalesce downdip and form large salt canopies and salt sheets (Figs. 6.12b, c).
- With compression starting in the Late Cretaceous, the rising relief of the High Atlas supplied many clastics that bypassed the shelf margin and locally initiated by differential loading Late Cretaceous to Tertiary extensional structures (Fig. 6.11). These raft structures slid downdip either on shales (near shore raft system in Fig. 4.34) or on allochthonous salt. Figure 6.12c is a strike section along this latter raft system. The basal salt décollement surface extends over several tens of kilometres, and its roof is dissected by gravity-driven listric growth faults that limit supra-salt minibasins which accumulated a thick Upper Cretaceous-Neogene clastic section. The inception of rafting is slightly earlier (Mid-Cenomanian to Campanian) in the shale-based near-shore raft compared to its western neighbour which outdated it and appears to remain active.
- On the lower slope of the Essaouira offshore west of the salt tongues, prominent toe thrust–related, salt-cored folds (Fig. 6.10c) are located, which represent the main exploration target in this area (Tari et al., 2000, 2003). Newly acquired high quality seismic data show that these salt-cored folds, which represent the westernmost salt structures in the Essaouira offshore, are underlain by basement tilted blocks that are bounded by westerly dipping normal.

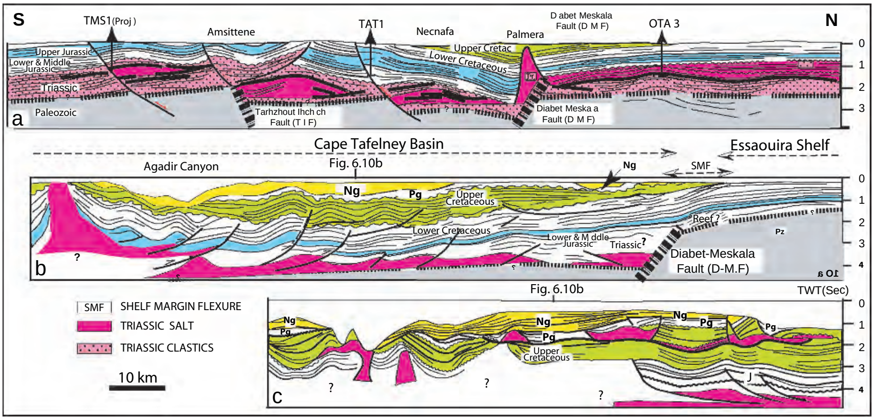

Fig. 6.12 Line drawings of N-S seismic transects across the Essaouira-Haha Basin, after Hafid (2006), modified. See Fig. 6.1 and Fig. 6.11 for location. (**a**) Near-shore/onshore profile showing, from north to south: the platform (OTA well), the West Necnafa depression and the outcroping Amsittene and Timsiline (TMS) anticlinal structures.- (**b**): Essaouira-Haha offshore zone showing from north to south the wide Essaouira shelf, the shelf margin flexure and the folded Cape Tafelney Basin.- (**c**): Upper slope of the Essaouira-Haha offshore showing extensive allochthonous salt and an extruded diapir

- To the west of the western termination of the offshore salt basin there is a zone of non-deformed flat Meso-Cenozoic sediments penetrated by the DSDP 415 borehole (Fig. 6.10c).

6.4.4 Tarfaya-Dakhla Basin

References: The recent recent works dealing with this transect are those by Ranke et al. (1982), Einsele et al. (1982), El Khatib (1995), El Khatib et al. (1995), Le Roy et al. (1997, 1998), Abou Ali et al. (2005). The Cenomanian-Turonian black shales have been particularly studied by Lüning et al. (2004) and Kolonic et al. (2005).

The Tarfaya-Dakhla Basin is the southernmost Atlantic basin of Morocco. It stretches over more than 1000 km along the western margin of the Sahara. It is bounded to the southeast by the Mauritanide thrust belt (Adrar Souttouf, Dhoul, Zemmour) and the Precambrian Reguibat Arch, and to the northeast by the Paleozoic outcrops of the Anti-Atlas (Figs. 1.11 and 6.1). Two well correlations (Figs. 6.13 and 6.14) and two regional cross-sections based on seismic data (Figs. 6.15 and 6.16) allow us to illustrate the stratigraphy and structure of the Atlantic-related formations in the northern parts of this very large area.

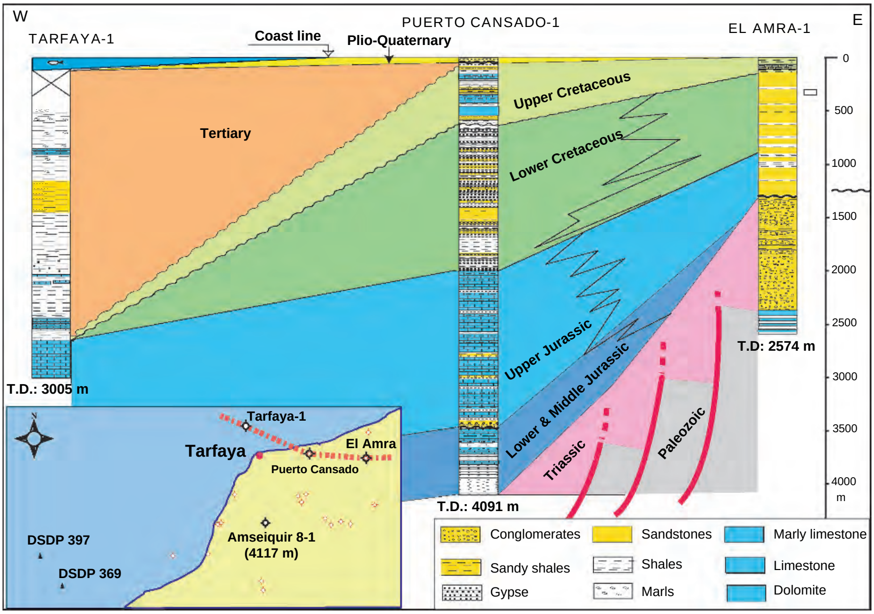

Fig. 6.13 E-W lithostratigraphic correlation between Tarfaya-1 offshore well, Puerto Cansado-1 and El Amra-1 onshore wells, Tarfaya Basin (after ONHYM 2004, modified)

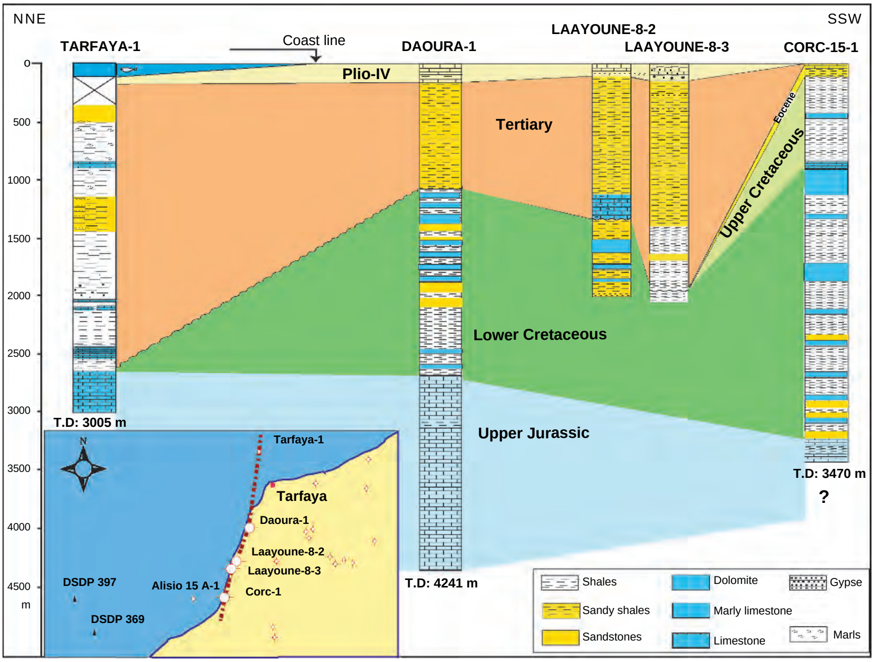

Fig. 6.14 N-S lithostratigraphic correlation between Tarfaya-1 offshore well and Daoura-1, Laayoune-8-2, Laayoune-8-3 and CORC-15- onshore wells, Tarfaya Basin, after ONHYM 2004, modified

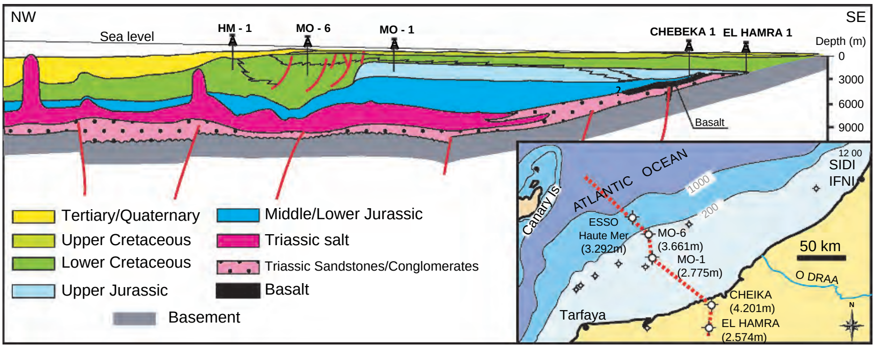

Fig. 6.15 SE-NW geological cross-section based on seismic data, Tarfaya Basin, after ONHYM 2004, modified. See Fig. 6.1 for location

6.4.4.1 Stratigraphy

The Upper Triassic to Lower Jurassic syn-rift megasequence is encountered only in the subsurface. It unconformably overlies the Upper Precambrian basement of the Reguibat Massif and probably also the Paleozoic of the Mauritanides in the South.

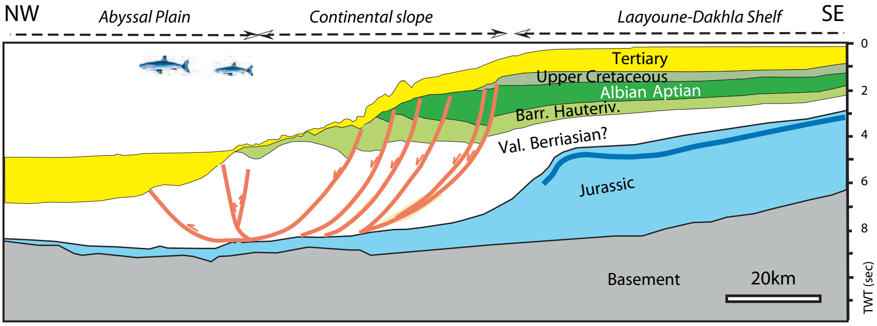

Fig. 6.16 Line drawing of the seismic section 1266 BJ-2, Boujdour offshore basin, after ONHYM 2004, modified. See Fig. 6.1 for location

It consists of red conglomerates and sandstones and thin bedded shales and carbonates. Doleritic basalt lava flows are interbedded in the section which was drilled with a maximum thickness of 2165 m in the Chebeika well (Fig. 6.13). Evaporites were detected by drilling and seismic profiling only in the offshore part of this segment where they form the southern extremity of the Moroccan Atlantic salt basin (Figs. 6.1, 6.15).

The post-rift megasequence is also known (partially) only from wells. It starts with Middle to Upper Jurassic carbonates which grade laterally into detrital facies towards the continent in the east and into marly facies towards the basin in the west. The thickness of the section is variable and increases northward and westward (up to 600 m of Middle Jurassic in the Puerto-Cansado well and up to 1700 m of Upper Jurassic in the Ameisquir-1 well) (Figs. 6.1 and 6.13). Prominent shelf edge reefal mounds formed during the Liassic and Dogger. They attain a maximum thickness of 1500 m. The Dogger reefs migrated seaward with respect to the Liassic ones. This reflects the progressive flooding of the area by the waters of the nascent Atlantic Ocean and the establishment of a carbonate platform. The transgression reached its maximum in the Oxfordian-Kimmeridgian and the carbonate platform prevailed up to the Portlandian. To the west of the shelf margin, the Jurassic thins into a pelagic basinal sequence that underlies the continental slope and the deep basin.

From Late Portlandian to Early Cretaceous an important regression resulted in the deposition of a thick (up to 2334 m in the Corc-15-1 well) widespread continental to transitional (lagoonal and deltaic) section that unconformably overlies the carbonate platform sediments. The Upper Cretaceous is characterized by thinner (about 750 m) shallow marine to lagoonal sediments that are deeply truncated at the shelf edge by the base Tertiary regional unconformity (Figs. 6.13 and 6.15). Organic-rich black shales occur at about the Cenomanian-Turonian boundary. They developed in relation with upwelling along the Atlantic coast during the C/T Oceanic Anoxic Event (Lüning et al. 2004; Kolonic et al., 2005). Paleocene-Eocene thin sandy to marly sediments overlie unconformably the Upper Cretaceous strata and, in turn, are overlain by a Miocene sequence which abruptly thickens westward beyond the shelf edge (up to 1 km in thickness).

6.4.4.2 Structure

Figure 6.14 is the interpretation of an E-W regional seismic transect that illustrates the transition from the onshore Tarfaya Basin to its offshore. It shows that the onshore basin is characterized by syn-rift structures in grabens and half grabens overlain by virtually undeformed Jurassic to Cretaceous formations which form a slightly westward dipping monocline structure. The syn-rift deposits were subdivided into four tectonically controlled sequences comparable to those defined in northern basins, and the rifting also migrated westward (Abou Ali et al., 2005).

The Alpine phases responsible for the folding of the Atlas belt to the north are reflected here by uplifts that resulted, from Upper Cretaceous onward, in periodic influx of detrital material into the basin. However, salt tectonics, as it was the case in the northern basins, played an important role in the evolution of the slope and the abyssal plain in the northern part of this segment. Important salt diapirs formed to the west of the shelf edge, and halokinesis greatly influenced sedimentation around salt bodies, among which some reached the sea floor (Fig. 6.15). Likewise, shale-related slope tectonics, characterized by the up dip extension and downdip compression, controlled Cretaceous sedimentation. This control seems to increase southward as illustrated by a section from the Boujdour margin (Fig. 6.16). Finally our sections show that a major regional Tertiary unconformity cuts deeply into the Mesozoic-Cenozoic section.

6.4.4.3 Comparison to the Northern Segments of the Atlantic Margin

Based on the data analysed in the present chapter and on previous studies we can differentiate two end-member passive margin types in the Atlantic margin of Morocco:

(1) The Mazagan Plateau-type passive margin is best described as an extensional margin that began with Triassic extension along west-dipping, most likely listric faults which seem to have lasted well into the Jurassic. The most landward block of this system could have formed the foundation of the platform margin. More seaward blocks drowned and were covered by hemipelagic and pelagic sedimentation in the progressively deepening passive margin (Fig. 6.9c) (Winterer & Hinz, 1984).

(2) The Tarfaya-type passive margin is also characterized by Triassic extension and Triassic-Lower Liassic evaporites, but it is overlain by massive Jurassic carbonate build-ups forming a thick carbonate platform with a transition into a coeval, thin basinal sequence. This platform evolved during a rapid sea level rising from a ramp type in the Lower and Middle Jurassic to a shelf margin platform with prominent build-ups in the Upper Jurassic (Fig. 6.15) and is later overlain by Cretaceous deep water sequences, i.e. low-stand wedges (Figs. 6.15 and 6.16).

The Essaouira-Western Atlas passive margin whose rifting/drifting structures, as we saw, were obliterated by the Atlas compression appears more likely to be comparable to the Mazagan-type margin and not, as suggested by Lancelot & Winterer (1980), with a Tarfaya-type margin. This conclusion is supported by the relatively minor Jurassic to Early Cretaceous thickness changes across the platform margin. Unfortunately, inadequate deep resolution does not permit to detail the basement structure that underlies much of the Essaouira offshore.

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Chapter 7
The Cretaceous-Tertiary Plateaus

S. Zouhri, A. Kchikach, O. Saddiqi, F.Z. El Haïmer, L. Baidder and A. Michard

7.1 General

References: The tectonic considerations hereafter are summarized after the synopsis of the Phanerozoic evolution of northern and central Africa by Guiraud et al. (2005). The role of the Trans-Saharan Seaway in the Late Cretaceous faunal exchanges has been also addressed recently by Courville (2007).

Morocco displays two sets of Cretaceous-Tertiary plateaus with unequal extension (Fig. 7.1). A first, western group includes the Oulad Abdoun Plateau, also referred to as the Plateau des Phosphates, the Ganntour Plateau and the Meskala Plateau north of the Atlas Mountains, and finally the coastal Laayoune-Bou Craa Plateau south of the Atlas. All these plateaus overlie a Variscan basement, and appear in close connection with the Atlantic Coastal Basins described in Chap. 6.

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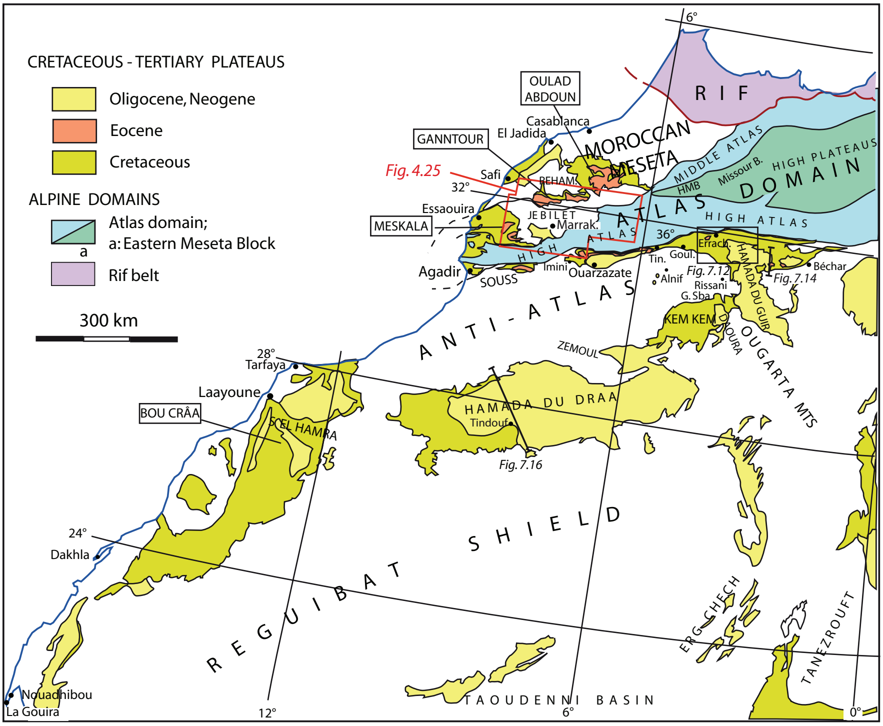

Fig. 7.1 Schematic geological map of the Cretaceous-Tertiary plateau area, after the Geological map of Morocco, 1:1, 000,000 (Geological Survey of Morocco, 1985). Errach.: Errachidia; G. Sba: Gara Sba; Goul.: Goulmima; Marrak.: Marrakech; Tin.: Tineghir

A second and eastern group corresponds to the so-called "hamadas", which extend essentially over the Sahara cratonic domain from the Guir Hamada and associated plateaus (Meski, Boudenib) to the east, to the Kem-Kem plateau, and eventually to the Draa Hamada to the southwest. The small plateaus on the northern slope of the Anti-Atlas (Imini, Goulmima) also belong to the hamada paleogeographic domain.

Both these plateau groups escaped the main effects of the Triassic rifting and those of the Alpine shortening. Their sedimentary sequence spans essentially from the late Early Cretaceous to the Neogene. Thus, neither the High Plateaus nor the Missour and High Moulouya basins of Eastern Meseta nor the Middle Atlas "Causses", whose more or less detached stratigraphic sequences include Triassic evaporites and thick Lower-Middle Jurassic limestones, are considered in this chapter. They belong to the Atlas realm, within which they simply form extended rigid blocks (see Chap. 4). Admittedly, this classification is partly conventional as the Moroccan (Western) Meseta also suffered significant Neogene faulting (e.g. North Jebilet Fault) and uplift (see Sect. 7.4).

During Cretaceous-Cenozoic times, sedimentation above the Meseta Block and the Sahara cratonic realm has been controlled by two mechanisms, i.e. faulting and active rifting episodes of the African plate linked to the Gondwana break up, and eustatic high stands related to strong warming of the global climate. By the beginning of the Early Cretaceous, a large, N-S trending Trans-Saharan Fault Zone (TSFZ) developed on the Algeria-Niger confines, in relation with the opening of the South Atlantic Ocean (Fig. 7.2A). The Saharan regions of Morocco were part of the West African Block. At that time, the marine transgression was limited to a wide gulf in the southern Algerian-Tunisian regions. In contrast, a large continental deltaic system prevailed south and west of this marine embayment, which nourished the deposition of turbidites on the North African margin (see Chap. 5).

Then, by late Barremian-early Aptian times (Figs. 7.2B, 7.3A), an active rifting episode linked to the opening of the Equatorial Atlantic Ocean reactivated large faults SE of the Hoggar massif and in the Benoue Trough. The continental fracturation, and the subsequent thermal relaxation controlled the outline of the transgression upon the Meseta, and especially above the Sahara domain during the Cenomanian-Turonian high stand. At that time, the Equatorial Atlantic Ocean and the western Tethys were connected by a Trans-Saharan Seaway situated east of the Hoggar massif (Fig. 7.3B). Shortly later, during the Late Cretaceous (late Santonian), a first compression episode linked to the onset of the Africa-Eurasia convergence was recorded within the intraplate realm itself. As a result, the Saharan seas developed a new outline during the Campanian-Maastrichtian high stand: the Hoggar massif was now bypassed along its western side, i.e. the eastern Taoudenni

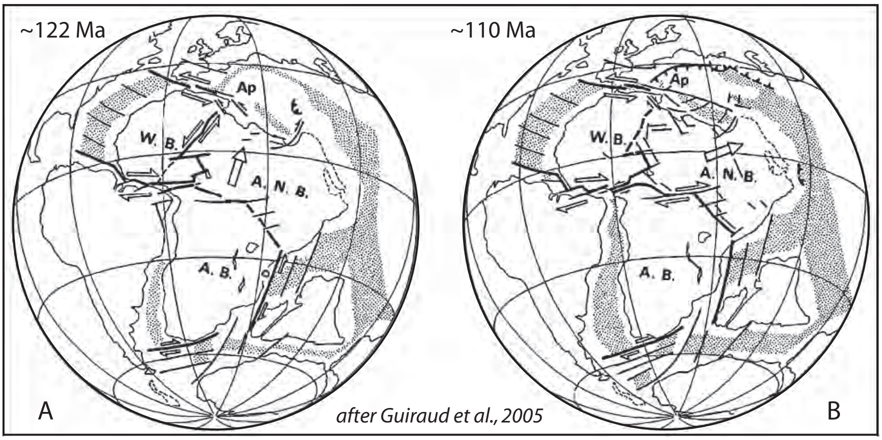

Fig. 7.2 Fracturation of the African plate during the Cretaceous Gondwana break-up, after Guiraud et al. (2005), modified. – **A**: Late Barremian (122 Ma). – **B**: Early Albian (110 Ma). Dotted: oceanic crust. AB/ANB/WB: Austral/Arabo-Nubian/Western Block. Ap: Apulia, bounded to the north by subduction planes (teeth)

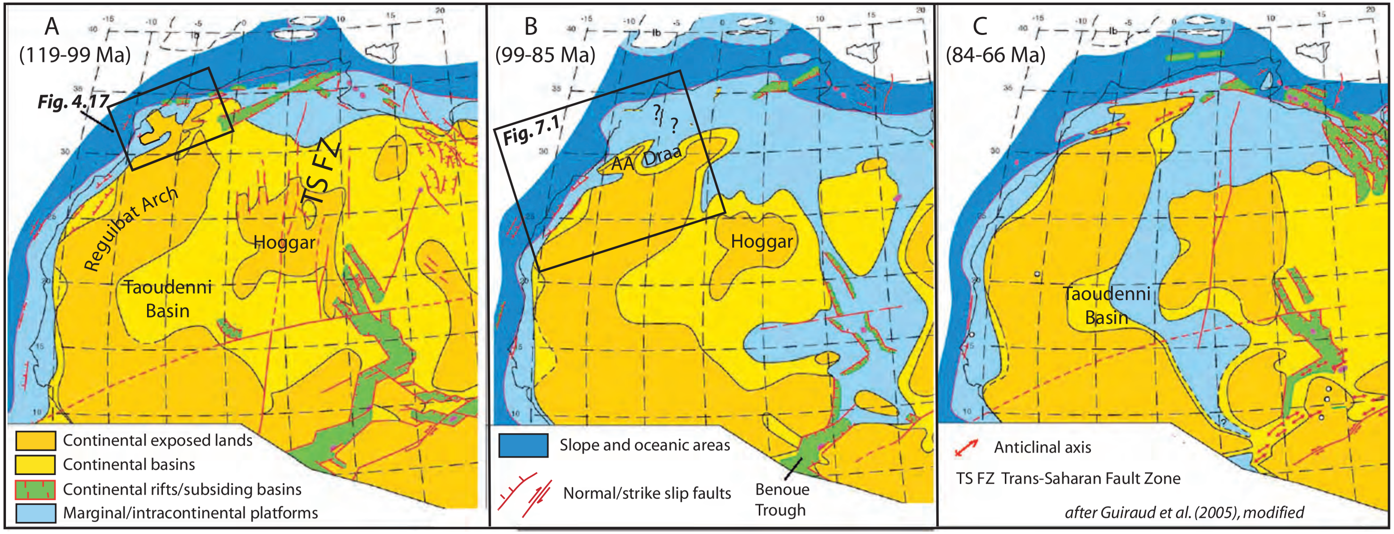

Fig. 7.3 Cretaceous paleogeography of NW Africa, after Guiraud et al. (2005), modified. – **A**: Aptian-Albian (119–99 Ma). Detail paleogeography of Morocco modified after Chap. 4, Fig. 4.17. – **B**: Cenomanian-Early Senonian (99–85 Ma), modified (the Atlas domain is regarded as entirely flooded). – **C**: Late Senonian-Mastrichtian (84–66 Ma). TSFZ: Trans-Saharan Fault Zone

Basin (Fig. 7.3C). During the latest Cretaceous, the sea progressively left the inland areas, whereas the Atlantic plateaus remained in shallow marine conditions up to the Middle Eocene (cf. Fig. 4.19).

7.2 The Atlantic Plateaus and Their Phosphate Deposits

References: The classical stratigraphic literature on the "Plateau des Phosphates" is referenced in Michard (1976). Among the more recent titles, the most important is certainly the third volume of the *Géologie des Gîtes minéraux* marocains (2d ed., 1986), entitled *Phosphates* and essentially prepared by H.M. Salvan, A. Boujo, and M. Azmany-Farkhany. Other salient works relative to the paleogeography, petrology and exploitation of the phosphorite are those by Einsele et al. (1982), Prévôt (1990), Trappe (1992), Moutaouakil & Giresse (1993), Gharbi (1998), Kchikach et al. (2002, 2006). References concerning the Vertebrate paleontology are included in the corresponding Sect. 7.2.4. The history of the phosphorite discovery is evoked in Chap. 9.

7.2.1 Stratigraphy

The stratigraphic sequence of the Atlantic Plateaus changes from west to east and from south to north. The well-developed Triassic deposits of the Doukkala Basin are preserved locally on the western and central Jebilet Massif (cf. Fig. 4.25). They are also known in borehole from the Bahira Basin. In contrast, the sedimentary cover of the Rehamna massif begins with Upper Jurassic-Lower Cretaceous red beds including locally a thin intercalation of Lower Cretaceous (Valanginian) marine limestone (Figs. 7.4, 7.5). These red and pink beds are followed upward by shallow marine, marly limestones dated as Cenomanian-Turonian. The Turonian limestones (e.g. Settat plateau) in turn are followed upward by regressive Senonian (Coniacian-Campanian) gypsum-bearing marls. This 100–400 m thick platform sequence constitutes the substratum of the Oulad Abdoun plateau northeast and east of the Rehamna massif, and south of the Central Massif (cf. Landsat image Fig. 3.17). They also form the northeastern Laayoune – Seguiet-el-Hamra Plateau in the Southern Provinces, on top of the Smara – Zemmour Paleozoic series west of the Reguibat Arch (cf. Landsat image Fig. 1.12). In the latter case, the thin Cenomanian-Turonian marls of the inner shelf quickly change westward into the thick black shale accumulations of the Tarfaya-El Aioun (Laayoune) Basin (see Chap. 6, Fig. 6.15).

The Early Cretaceous-Senonian sequence, about 150–200 m thick, forms the base of the thinner, but much more interesting phosphorite-bearing sequence ("Série phosphatée"). The thickness and facies of the latter sequence changes laterally from the western to eastern parts of the former Atlantic gulf (cf. Fig. 4.19). The

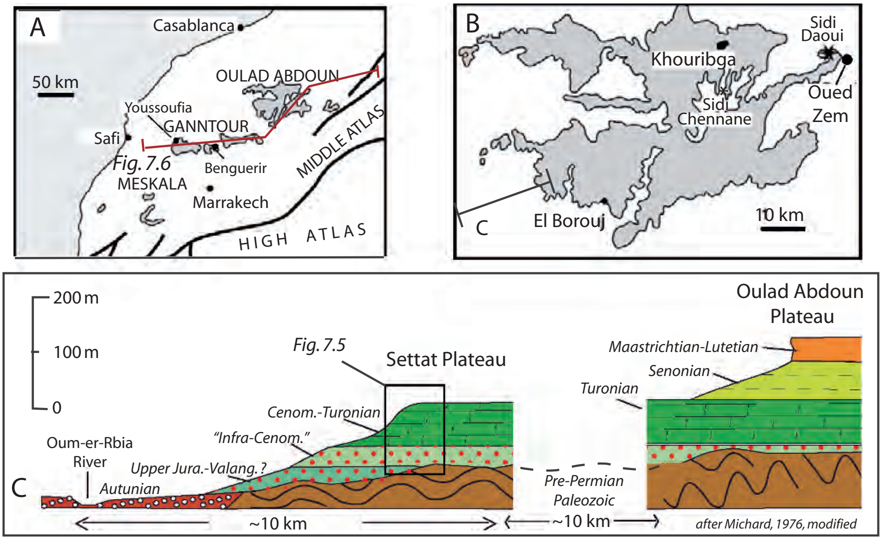

Fig. 7.4 Extension of the Phosphate Series in Western Meseta (**A**), main exploitation centers (**A, B**), and schematic cross-section of the Settat-Oulad Abdoun Plateau (**C**). – **A, B**: from Pereda Suberbiola et al. (2004), modified. – **C**: from Michard (1976), modified. Valanginian marine intercalations have been mapped both to the northwest and southeast of the Rehamna Massif (Gigout, 1954; Bolelli et al., 1959). Triassic red silts and basalts only occur in the Doukkala Basin, west of the Rehamna Massif

2D restoration of these stratigraphic variations has been obtained for the Western Meseta Atlantic embayment where the main phosphate sedimentation occurred (Fig. 7.6). The western deposits are thicker, and include black shales which progressively disappear eastward. The Maastrichtian deposits which transgress upon the lower Senonian deposits are thin and phosphate-rich in the northern Oulad Abdoun and, in contrast, become thick and marly-phosphatic southward. The

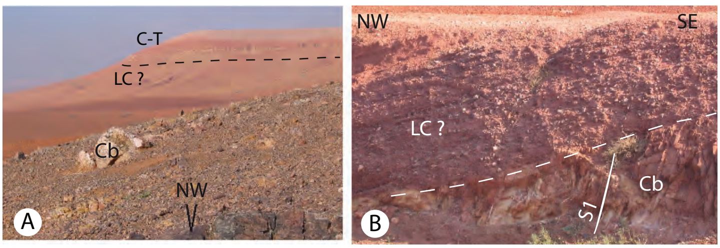

Fig. 7.5 Transgression of the Mesozoic sequence on the Variscan basement at the southwest border of the Settat Plateau, north of Mechra-ben-Abbou, Oum-er-Rbia Valley (see Fig. 7.4C, left). – **A**: Panorama. The escarpment of the Turonian plateau in the background is about 10 m high. – **B**: Detail of the basal unconformity in the 2 m-high roadcut. Cb: Folded Middle Cambrian metagreywackes (S_1: slaty cleavage); LC?: Lower Cretaceous continental/shallow marine deposits, only dated west of the Rehamna massif (Valanginian limestone intercalation); C-T: Cenomanian-Turonian marls and limestones

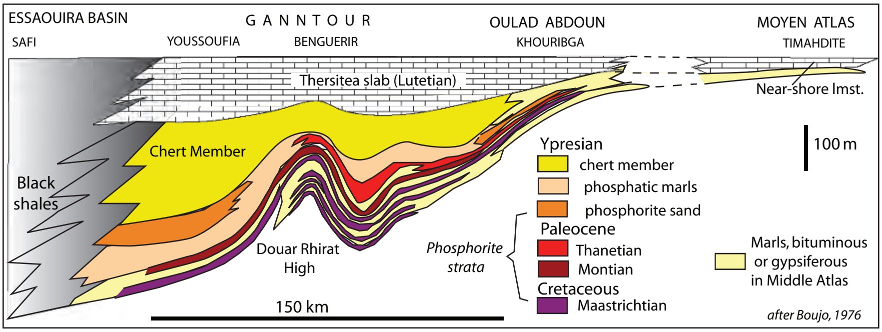

Fig. 7.6 Succession and lateral variation of the sedimentary facies of the Phosphate Series in Western Meseta, after Boujo (1976), modified. For location, see Fig. 7.4A

Montian-Thanetian beds are more uniform, consisting of unconsolidated phosphorites, but again conspicuous lateral changes occur in the Ypresian beds, where phophorite concentrations are the most important. The thick, late Ypresian cherty beds of the Ganntour Basin change north-eastward into marly phosphorite beds. The Lutetian dolomitic limestone slab ("Dalle à Thersitées") which caps the phosphate beds becomes progressively thinner eastward, being overlain by marls, gypsum and continental conglomerates (Late Eocene-Oligocene) in the Middle Atlas (Timahdite syncline; cf Chap. 4, Fig. 4.11).

The typical phosphorite sequence of the Oulad Abdoun Basin (Fig. 7.7) barely exceeds 40 m, including its Lutetian cap. The sequence begins with Maastrichtian calcareous bone-beds, followed upward by phosphatic marls (miners' "Couche III"), whereas the top of this stage consists of marly limestones and marls. The Montian (Danian) shows its usual facies of uncemented, sandy phosphorites ("Couche IIa, IIb") overlain by phosphatic limestones. The Thanetian-lowermost Ypresian consists of phosphatic limestones with nodular flints and coprolites, which gives a useful guide horizon ("Intercalaire couches I/II"). The sequence continues upward in the Ypresian with alternating beds of marly and phosphatic limestones, coarse sandy phosphorites ("Couches I, 0, A, B") with chert horizons and scarce silt or pelite layers (Gharbi, 1998).

7.2.2 Petrology and Formation of the Phosphorite Deposits

In view of both their sedimentary originality and economic importance, the sandy phosphorite deposits of the Meseta plateaus deserve some petrological comments. These very peculiar sands consist of various types of phosphate grain, coated or not (Prévôt, 1990; El Moutaouakil & Giresse, 1993). The inner part of the coated grains most often consists of phosphatic muds with laminar or concentric stromatolitic laminations, which preserve cyanobacterian nannostructures. Foraminifera, diatoms and/or radiolarians totally transformed into apatite are also

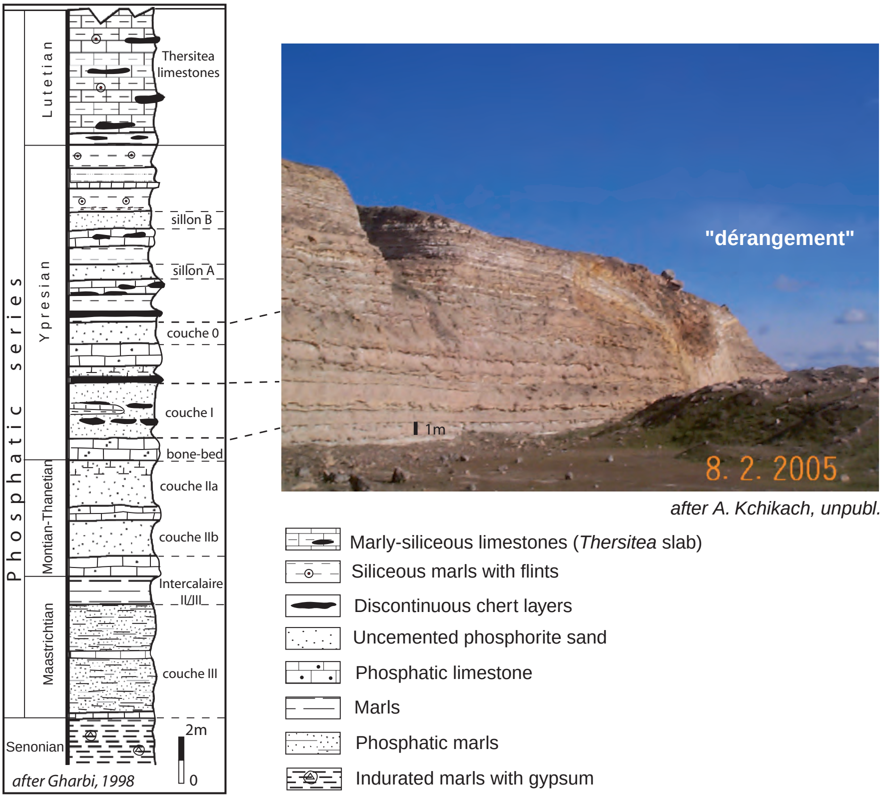

Fig. 7.7 The condensed phosphate series of the Oulad Abdoun Plateau at Sidi Daoui, after Gharbi (1998) and OCP documents. *Right*: view of the front of an exploitation showing a conspicuous "dérangement" (collapse) in the background. Photo A. Kchikach

frequent, as well as bone fragments. The coating consists of isotropic apatite. Other grains are lithoclasts of phosphatic muds with pellets, quartz grains and organic material cemented by cryptocrystalline apatite. The phosphorite sand often includes abundant coprolites transformed into apatite, as well as Vertebrate teeth and bones.

Apatite constitutes 50–98% of the sediments, with a low CO_2 content. The origin of this extraordinary concentration of phosphate was controlled both by paleogeographic and eustatic phenomenon (El Moutaouakil & Giresse, 1993, with references therein). According to the latter authors, who particularly studied the Ganntour-Oulad Abdoun deposits, phosphorite genesis occurred in a shallow water gulf whose communication with the ocean was progressively reduced from the Maastrichtian to the Lutetian. The occurrence of dolomite, attapulgite and traces of halite indicate an obvious restriction of the communications toward the open ocean. Strong upwelling currents favoured an important biogenic fixation of phosphorus by the phyto- and zooplankton. Intertidal stromatolitic mats, especially extensive during the marine low stands, participated in this process. Mineralization of phosphorus resulted in the formation of phosphatic muds, which were fragmented and reworked as tiny chips by the tempests during high stand periods, and transported toward accumulation areas by tidal currents. As many as eight eustatic cycles have been identified based on minor transgressive-regressive sequences.

7.2.3 Mining

Morocco phosphorite deposits represent about three quarters of the phosphate reserves of the world. The largest deposits exist in the Oulad Abdoun Basin (Khouribga center), with 44% of the Moroccan reserves compared to the Ganntour Basin (Youssoufia, Bengerir, 36%), the Meskala Basin (Chichaoua, 19%), and Bou Craa (1%).

The Oulad Abdoun Basin offers the easiest conditions for open air mining (Fig. 7.8). Locally, the tabular layering suffers perturbations by roughly conical structures filled with coarse breccias, the so-called "dérangements" (Figs. 7.7, 7.8). These sterile bodies are formed by accumulations of silicified limestones, or by blocks of limestones within an argillaceous matrix. They formed through the collapse of the phosphate series above karstic caves opened by dissolution of the gypsum and/or limestones formations in the Upper Cretaceous substratum. As these collapses seriously disturb the exploitation, several exploration works by wells and boreholes attempted to localize them beneath the Quaternary cover. However, mapping by Time-Domain Electro-Magnetic method offers promising results (Fig. 7.9), and would permit (especially when combined with other methods) the definition of these structures before the mining front reaches them.

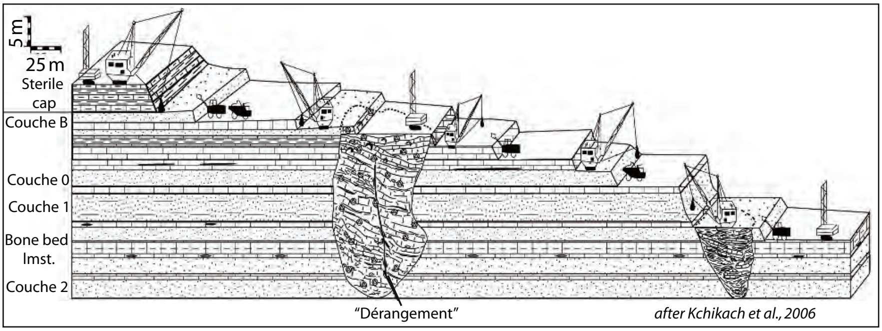

Fig. 7.8 Chain of exploitation of the Oulad Abdoun phosphorites (Sidi Chennane area), after Kchikach et al. (2006). The chain begins with the removal of the sterile cap (*left*) up to the "sillon B", and then the phosphorite layers ("sillons", "couches") are mined step by step. The "dérangements" (karstic collapses) hampers the mining process

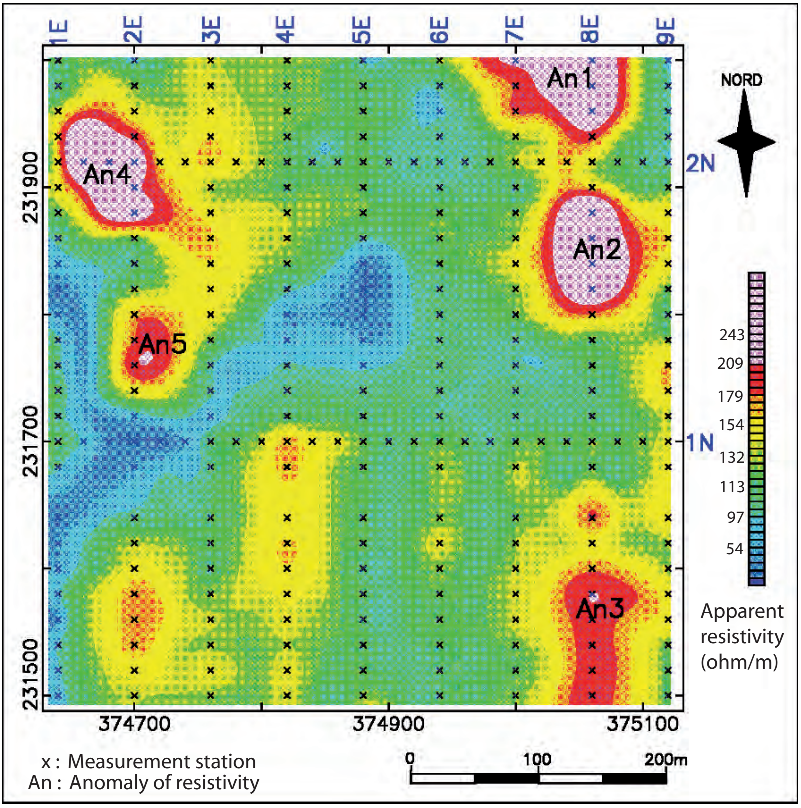

Fig. 7.9 Apparent resistivity map of an area of the Phosphate Plateau close to Sidi Chennane, based on sampling of the Time-Domain Electromagnetic Sounding method at 3 µs, after Kchikach et al. unpublished (cf. Kchikach et al., 2006). The main anomalies correspond to hidden "dérangements", as confirmed by mechanical boreholes (A1) or by further exploitation (A2)

7.2.4 The Vertebrate Fossils of the Oulad Abdoun Phosphorites

The pioneering work by Arambourg (1935, 1952) illustrated the richness of the Moroccan phosphorite deposits in Vertebrate fossils. The scientific importance of these deposits was recently confirmed with the discovery of the oldest mammals and birds on Africa. Such fossils occur within the entire phosphate series from the Maastrichtian to the base of Lutetian, and they are particularly abundant and varied in the Oulad Abdoun Basin. They constitute one of the most important references for the history of biodiversity during the end of Cretaceous-early Tertiary times, and thus deserve a summary in this chapter.

Apart from the Teleostean and Selachian groups, which are extremely abundant and diversified, the fauna includes several major taxonomic groups. The mammals from the "Plateau des Phosphates" include the most ancient herbivorous ungulate of the Proboscidean group (Fig. 7.10A; Gheerbrant et al., 1996, 1998, 2002,

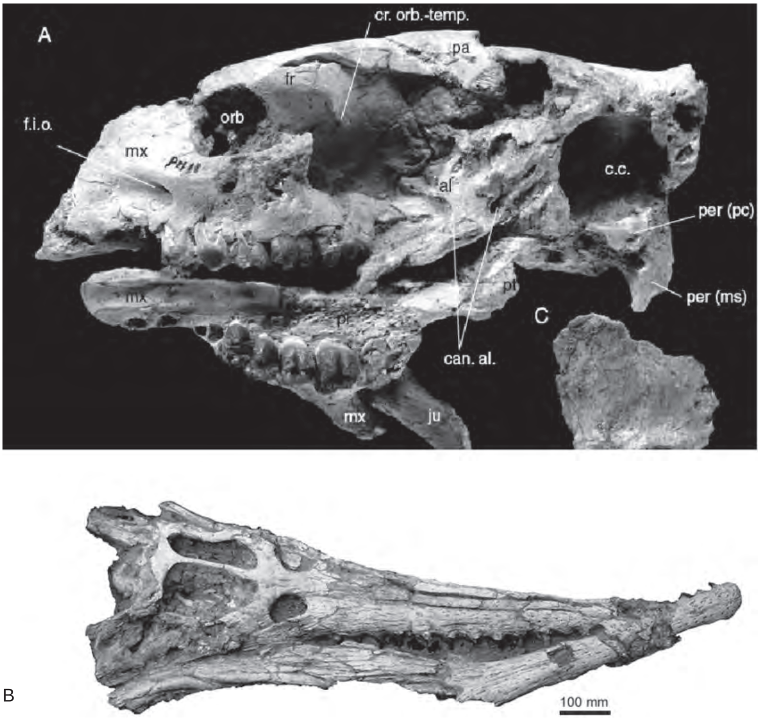

Fig. 7.10 Examples of Vertebrate fossils extracted from the Oulad Abdoun phosphorites. – **A**: Proboscidean skull, *Phosphatherium escuilliei*, after Gheerbrant et al. (1996). The body was probably carried into the sea by a river flood. The skull is almost complete, except the premaxillaries and nasal bones. The varied skull bones are clearly identified (see original publication). "C" is an isolated squamosal. This is the oldest known Proboscidean, found at the Paleocene-Eocene boundary in the Grand Daoui zone. A living reconstitution of this "elephant without trunk", weighty about 15 kg, is shown in *Maroc, memoire de la Terre*, Edit. Muséum Nat. Hist. Nat. Paris, 1999, p. 163). – **B**: Skull and mandible of a Crocodyliform from the Oulad Abdoun Paleocene phosphorites: *Arambourgisuchus khouribgaensis*, after Jouve et al. (2005)

2003). Most of these fossils occur in the bone-bed at the bottom of the "Intercalaire Couches II/I" (Fig. 7.7) dated as lowermost Ypresian by its Selachian fossils (Gheerbrant et al., 2003). Besides the Proboscideans, two condylarths and three other taxa have been found (Gheerbrant et al., 2001, 2006). The birds are almost exclusively marine or coastal Procellariiformes, Pelecaniformes and Anseriformes, and include at least 10 species (Bourdon, 2006a, b). They are dominated by pseudoteeth birds, big, long-flight pelagic birds, up to 5 m wingspan, which thus evoke the modern albatross. The only continental bird from the Oulad Abdoun compares in size with the European white stork. Selachians (sharks and rays) are the most abundant vertebrate from the Phosphate Series, and these have been used to define a stratigraphic scale (Arambourg, 1952; Noubhani & Cappetta, 1997). Crocodyliformes are well represented by Dyrosaurids and Eusuchians in the Paleocene levels

(Hua & Jouve, 2004; Jouve, 2004, 2005; Jouve et al. 2005, 2006), whereas they are scarce in the Cretaceous levels where they are dominated by other reptilian such as the Mosasaurs. Chelonians are represented by Pleurodires and Cryptodires with several new taxa (Gaffney et al., 2006; Gaffney & Tong, 2003; Hirayama & Tong, 2003; Tong & Hirayama, 2002, 2004). Several almost complete Mosasauridae skeltons, up to 15 m long, were recently discovered from the Oulad Abdoun Maastrichtian phosphorites (Fig. 7.10B; Bardet et al., 2004; 2005a, b). Other Squamata are represented by a varanoid and a puzzling snake. Bones from a small sauropod titanosauriform (last represent of Dinosauria in Africa) is known from the Oulad Abdoun Maastrichtian beds (Pedera Suberbiola et al., 2004). The flying reptilians or Pterosauria are also represented by one of the last species of the group; it had a 5 m large wingspan.

This extraordinary faunal succession documents the Vertebrate evolution along a 25 My interval, involving two major climatic-biological crisis, the famous K-T crisis and the less known, but very important Paleocene-Eocene crisis. The excellent bone preservation in the phosphate-rich matrix accounts for the great value of these fossils for scientists and amateur collectors, and also for local diggers.

7.3 The Saharan and Sub-Saharan Plateaus

References: The Saharan plateaus are described by Fabre (2005), based on personal observations and previous classical works. The stratigraphy of the Errachidia-Boudenib hamadas has been studied by Ferrandini et al. (1985), Herbig (1988), and Ettachfini & Andreu (2004). The geomorphological evolution and associated paleo-alterations (carbonation, silicification) of these plateaus are described by Elyoussi et al. (1990), Ben Brahim (1994), Thiry and Ben Brahim (1997). Recent observations on the small plateaus of the Anti-Atlas northern slope (Imini, Goulmima) are reported in Rhalmi (2000) and Gutzmer et al. (2006). Paleontological references concerning the Vertebrate fossils are given in Sect. 7.3.4, and the current works on apatite fission tracks are mentioned in Sect. 7.4.

7.3.1 The Imini and Goulmima Plateaus

A narrow, but elongated zone of Cretaceous (-Tertiary) plateaus occur on the northern fringe of the central and eastern Anti-Atlas, from the Imini manganese district to Goulmima (Fig. 7.1). These Sub-Saharan plateaus correspond to the southern limb of the Ouarzazate and Errachidia-Boudenib basins, respectively, the northern limb of which is included in the folded South Sub-Atlas Zone (see Chap. 4, Figs. 4.36, 4.40, 4.41). These plateaus are more or less tilted and mildly folded in relation with the nearby Atlas fold belt.

From the stratigraphic point of view, the western part of this narrow domain is characterized by the absence or great reduction of the Lower Cretaceous deposits,

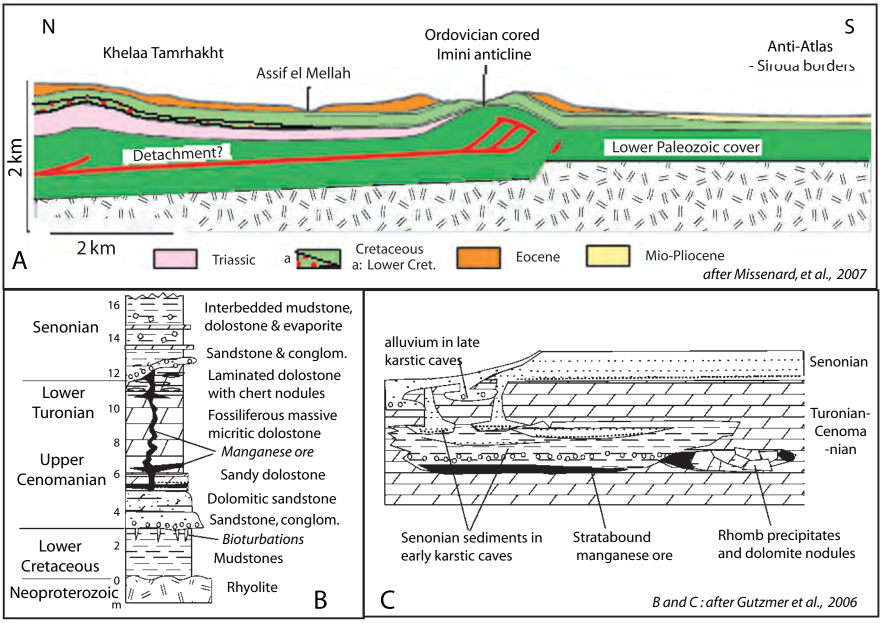

Fig. 7.11 The weakly deformed Cretaceous-Tertiary plateau of the Imini district west of Ouarzazate. – **A**: General cross-section, after Missenard et al. (2007). – **B**: Stratigraphic column, after Gutzmer et al. (2006), modified. – **C**: Localization of the stratabound manganese ore in a Late Cretaceous paleokarst, after Gutzmer et al. (2006), modified

and the direct (or almost direct) onlap of the Cenomanian-Turonian carbonates onto the Paleozoic or Neoproterozoic formations (Fig. 7.11A, B). The latter carbonate formations display clear evidence of deposition close to the shoreline of the Atlas Cretaceous gulf (cf. Chap. 4, Fig. 4.17). This particular paleogeographic location has favoured the rich manganese mineralization of the Cenomanian-Turonian dolomitic carbonates of the Imini district. The Anti-Atlas basement volcanics were the source for Mn, Pb and Ba. The stratabound ores occur in dolostone breccias and ferruginous clays which, according to Gutzmer et al. (2006), represent the earliest phase of internal sedimentation in a karstic cave system (Fig. 7.11C). The karst developed during a long period of weathering, prior to the deposition of the "Senonian" terrestrial red beds which filled the caves. The stratigraphic record ends with unconformable continental silts and conglomerates of Mio-Pliocene age. They were deposited during the mild deformation that formed the Imini anticline, 25 km south of the South Atlas Fault itself.

The J. Saghro Cretaceous-Eocene series crop out in the Tineghir area, at the eastern tip of the Ouarzazate basin. Further east, the Cretaceous plateau widens toward Goulmima and Errachidia, where it merges with the "Hamada de Meski" (Sect. 7.3.2). Lower Cretaceous continental sediments display a greater thickness than at Imini (about 200 m), and the Cenomanian-Turonian fossils record open marine sedimentation (Sect. 7.3.4). By contrast, Eocene deposits are lacking in the

Goulmima region. They occur close to Errachidia under continental facies which constitue the oldest Cenozoic deposits of the hamadas. The Goulmima Hamada is affected by a conspicuous ENE anticline (Tadighoust anticline) located above a Triassic normal fault (Saint-Bézar et al., 1998) and related to the detachment of the Mesozoic series (see Chap. 4, Fig. 4.35D). This compares with the structure of the Imini anticline (Fig. 7.11A).

7.3.2 The Hamada System in the Errachidia Region

The northern tip of the Hamada System is exposed close to Errachidia, immediately south of the South Atlas Fault (Fig. 7.12). From bottom to top, three main superimposed plateaus or "hamadas" are observed:

– the Hamada of Meski-Aoufous (Lower Hamada) is a table of Upper Cenomanian-Lower Turonian limestones above Lower Cretaceous continental redbeds that unconformably overlie the Paleozoic series of the eastern Anti-Atlas;

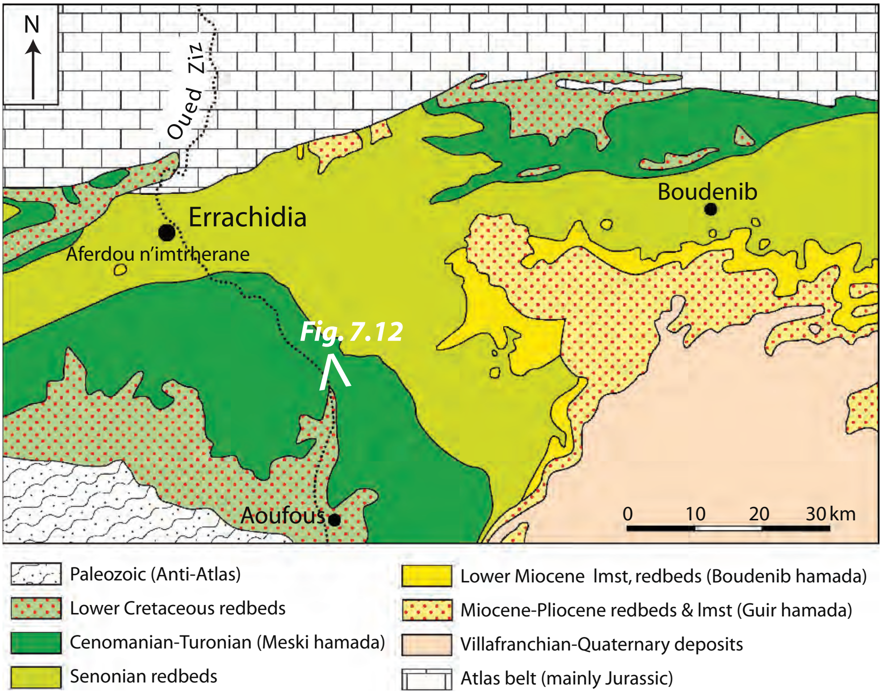

Fig. 7.12 Geological setting of the hammadas east of the Anti-Atlas (see Fig. 7.1 for location), after the Geological map of Morocco, 1:1, 000,000 (1985). Asterisk: Location of Fig. 7.13. North of Errachidia and Boudenib, the Cretaceous-Tertiary formations are clearly folded, and belong to the frontal zone of the Atlas belt (see Chap. 4, Fig. 4.35)

– the Hamada of Boudenib (Hamada à *Clavator*, Middle Hamada), whose escarpment is made of Upper Cretaceous (Senonian p.p.) redbeds topped by Lower Miocene lacustrine limestones;
– the Guir Hamada or Upper Hamada, which displays an escarpment of reddish continental siltstones (Upper Miocene?) topped by lacustrine limestones and conglomerates of probable Pliocene age.

It must be noted that limited plateaus of Eocene continental deposits are described west of Bechar, and referred to as the "Hamada à *Ceratodes*" (Fig. 7.1). The undated hill 10 km SW of Errachidia (Aferdou n'Imtrherane) is currently assigned to the Lower Miocene (Fig. 7.12), but could alternatively be an equivalent of the Eocene Hamada. If correct, the Eocene transgression of the Ouarzazate Basin would have been limited to the east between Tineghir and Errachidia, i.e. near Goulmima.

The gentle slope of the Hamada of Meski escarpement (Fig. 7.13) corresponds to reddish siltstones and sandstones with gypsum and rare limestone intercalations. These continental deposits represent the Early Cretaceous ("Infra-Cenomanian"), referred to as "Continental intercalaire" in the Kem Kem and more southern Saharan areas. The overlying table includes four members corresponding to progressive environmental changes (Ferrandini et al., 1985). The lowest, Cenomanian coastal member (Ostreidae wackstones with benthic foraminifers) displays an eastward increasing brackish tendency. The second member includes reef limestones with rudists, stromatoporidae and dasycladaceae typical of a proximal platform environment. The third member consists of *Exogyra* and *Nerinea* bioclastic limestones topped by a hard ground which marks the end of Cenomanian. The last member consists of siliceous, weakly dolomitic micrites and ostreidae limestones dated from the Turonian and corresponding to open marine conditions.

The lower part of the Hamada of Boudenib escarpement consists of argillaceous sandy redbeds with gypsum and salt. These poorly dated formations conformably overlie the Turonian limestones, and thus probably correspond to Senonian sediments deposited in a confined environment. The Aquitanian succession of the

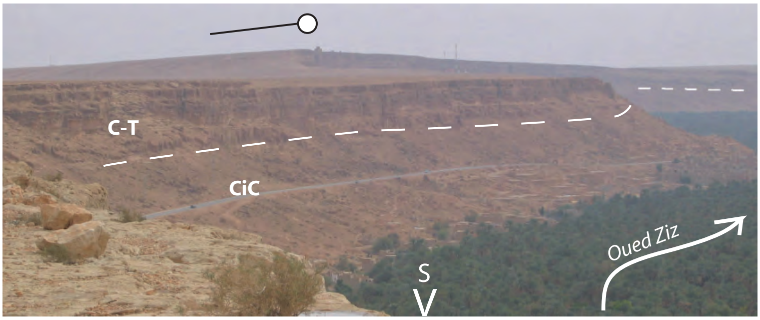

Fig. 7.13 View of the Meski Hamada escarpment from Ait Chaker (south of the Meski "Blue Spring"; location in Fig. 7.12) looking southward on the Oued Ziz Valley. CiC: Lower Cretaceous continental redbeds. C-T: Cenomanian-Turonian marine limestones. Note the widely open fold related to the Atlas Orogeny

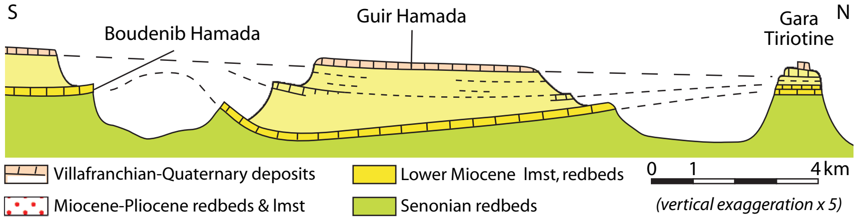

Fig. 7.14 Cross-section of the eastern border of the Guir Hamada, half-way between Boudenib and Bechar: an example of mild Mio-Pliocene deformation south of the South Atlas Fault. After Lavocat (1954), modified

plateau (formerly classified as Upper Oligocene) starts with *Melanatria* calcareous sandstones, and continues upward with lacustrine limestones with abundant gastropods of the genus *Clavator*, associated with sands and conglomerates. Although these Lower Miocene deposits overlie conformably the Upper Cretaceous redbeds, an important hiatus occurs in between, as indicated by the local occurrence of the Eocene "*Ceratodes* Hamada". An unconformity is observed west of Bechar.

Sediments of the Guir Hamada form the uppermost cuesta of the Hamada System. The escarpment consists of continental siltstones and sandstones with limestone intercalations. The plateau corresponds to lacustrine/palustrine silicified limestones (Fig. 7.14). The sequence is poorly dated and was assigned to the "Mio-Pliocene". On top of the Hamada, continental sedimentation continues into the Villafranchian, whose silty-conglomeratic deposits are cemented by thick diagenetic calcretes. Diagenetic silicifications are widespread throughout the Boudenib and Guir Hamada formations (Fig. 7.15). Their development was likely related to wet climatic stages (Thiry & Ben Brahim, 1997; see also Chap. 8).

Slightly different dipping planes and local unconformities between the plateaus and underlying redbeds (Fig. 7.14) suggest syntectonic sedimentation at the southern rim of the uprising Atlas belt. Open folding of the Meski Hamada next to Aoufous (Fig. 7.13) is probably related to far-field stress induced by the Atlas orogenic episode. Further south along the western border of the Guir Hamada, the Mio-Pliocene sediments overlie directly the Lower Cretaceous redbeds (e.g. east of Rissani), and then the Tafilalt Paleozoic series (see Geological map of Morocco, scale 1/200,000, sheet Tafilalt-Taouz). The Miocene-Pliocene deposits also disappear, and then the Plio-Quaternary formations of the uppermost Hamada overlie the south-eastern Tafilalt Massif directly. Still further to the SW in the Kem Kem area, the Cretaceous and Miocene-Pliocene formations again develop, suggesting a regional deformation of the Anti-Atlas basement and Mesozoic cover during the Neogene phase of the Atlas Orogeny.

7.3.3 The Draa Hamada

The Draa Hamada is the largest of the "Grandes Hamadas" System, which also includes the Guir Hamada and the smaller Daoura Hamada (Fig. 7.1). The Draa

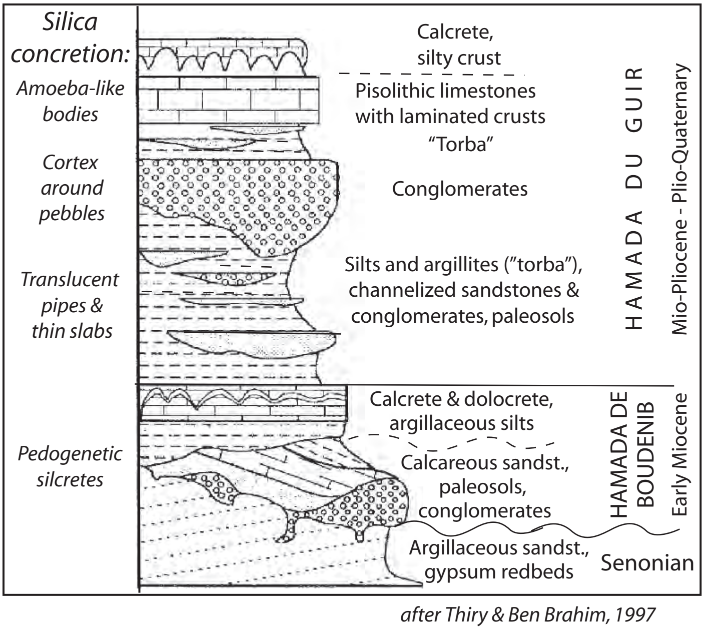

Fig. 7.15 Schematic sedimentary column of the continental hamadas in the Boudenib area, showing the associated calcrete and silcrete formations, after Thiry & Ben Brahim (1997), modified

Hamada extends onto the Tindouf basin between the southernmost Anti-Atlas aureoles (Carboniferous series of the J. Ouarkziz massif) and the Reguibat Precambrian arch. Most of this Saharan plateau consists of Cenozoic continental deposits, loosely assigned to the Miocene-Pliocene and passing upward to Quaternary formations (see Chap. 8). However, Cretaceous sediments also occur in the western part of the basin, including a thin wedge of Cenomanian-Turonian shallow marine limestones (Fig. 7.16). The overlying sandstones and conglomerates with silicified wood have been formerly considered as Paleogene, but they are currently assigned to the "Mio-Pliocene". The synformal geometry of the Draa Hamada testifies for large wavelength deformation of the northwest Saharan lithosphere during the Late

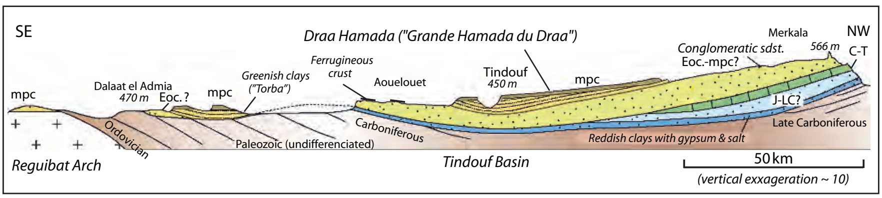

Fig. 7.16 Schematic cross-section of the Great Draa Hamada, after Gevin (unpubl. Report, 1974), in Fabre (2005), modified. Vertical scale strongly exaggerated. See Fig. 7.1 for location

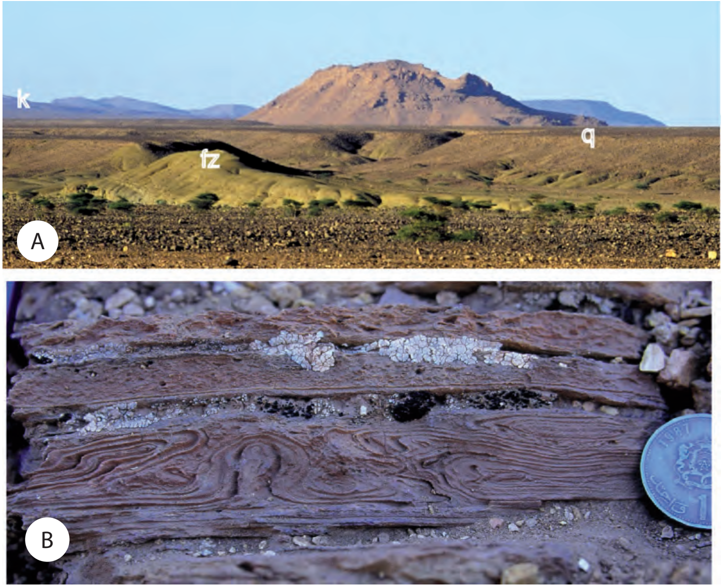

Fig. 7.17 Mio-Pliocene deposits in the Eastern Anti-Atlas. – **A**: The northern Tourt, one of the two Miocene-Pliocene hills 15 km SE of Alnif, as seen from the south (k: Cambrian; fz: Fezzouata Fm.; q: Quaternary terrace). – **B**: Seismites in the silty lacustrine/palustrine limestones of the higher beds of the hill (Upper Miocene-Pliocene). Diameter of the coin: 3 cm. Photos L. Baidder

Cretaceous (cf. Anti-Atlas swell and Draa/Tindouf embayment, Fig. 7.3B), reactivated during the Neogene.

North of the Kem Kem plateau, the twin pink hills named "Tourt" close to Alnif (Fig. 7.17A) give evidence of the large extension of the Miocene-Pliocene deposits between the Draa and Guir Hamadas, respectively. They overlie unconformably the Lower Ordovician Fezzouata formation, and consist of coarse sandy carbonates at the bottom, homologous to the Lower Miocene Boudenib limestones, topped by micritic, lacustrine/palustrine silty limestones homologous to those of the Guir Hamada. In the latter deposits, slump balls and convolute bedding have recorded a significant, synsedimentary seismic activity (Fig. 7.17B). This can be regarded either as a far field effect of the Atlas Orogeny, or more probably as echoing the coeval volcanic activity responsible for the emplacement of nephelinites and basanites in the neighbouring J. Saghro (cf. Chap. 4).

7.3.4 The Vertebrate Fossils of the Kem Kem and Goulmima Plateaus

In common with the Vertebrate fossils of the Phosphate Plateau, those of the Sub-Saharan Plateaus from the Errachidia-Tafilalt region deserve special mention. The rich fossil faunas from the Kem Kem and Goulmima plateaus occur in continental and marine beds, respectively. Together, they document the crucial period of the Gondwana break-up (see above, Fig. 7.2).

The Kem Kem fossiliferous beds are exposed along the escarpment of the Cenomanian-Turonian plateau south of the Tafilalt Paleozoic region (Fig. 7.18). The escarpment consists mainly of red sandstones (200 m thick "Continental intercalaire"), overlain by coloured gypsum marls, which finally pass up into Cenomanian marly limestones. The particular interest of the Vertebrate fossils of the "Continental intercalaire" in this region dates from the years 1940–1950 when R. Lavocat collected a rich fauna of "fishes", turtles, crocodiles, dinosaurs and pterosaurs (e.g. Lavocat, 1951, 1954). Indeed, the Kem Kem Dinosaurian and Pterosaurian assemblage is one of the richest of Africa. In 1995, an international team discovered the skull of one of the greatest carnivorous dinosaurs, *Carcharodontosaurus saharicus*, a Theropod which was probably as tall as the famous *Tyrannosaurus* (about 15 m long; Buffetaut, 1989a; Russel, 1996; Sereno et al., 1996; Amiot et al., 2004). Theropods are also represented by four other families in the Kem Kem beds: (i) the Spinosaurids, probably fish-eating animals, with *Spinosaurus aegyptiacus* (Buffetaut 1989a, 1989b, 1992; Dal Sasso et al., 2005) and *Spinosaurus maroccanus* (Russel, 1996); (ii) the Abelisaurids (Russel, 1996; Mahler, 2005); (iii) the Noasaurids, with *Deltadromeus agilis* (Sereno et al., 1996), probably a running hunter; and (iv) the Dromaeosaurids (Amiot et al., 2004). The giant herbivorous sauropods are represented by *Rebbachisaurus garasbae* (found at Gara Sba and dedicated to the Aït Rebbach tribe; Lavocat 1951, 1954; Russel, 1996), whose dorsal vertebras measure 1 m, and which probably reached 20 m in length. Other sauropods, belonging to the Titanosaurid and Dicraeosaurid families (Russel, 1996; Sereno et al., 1996; Russel et Paesler, 2003) are also known. In contrast, the Ornithischians are only known by their foot casts yet. It is notable that remarkable dinosaur tracks and skeletons are also known

Fig. 7.18 View of the "Continental intercalaire" of the Kem Kem Plateau in the Gara Sba area (east of Oum Jerane; see Fig. 7.1 for location). The age of these rich fossiliferous horizons is debatable, Albian (?) or Cenomanian (Sereno et al., 1996). The whitish blocks are Turonian limestones from the overlying plateau (not shown). Photo S. Zouhri

from the Atlas synclines, being dated at ca. 160 Ma. This is the case of the tall sauropod *Atlasaurus imelakei* from the Bathonian-Callovian (Guettioua Fm) of the Tilougguit syncline of the central High Atlas south of Beni Mellal (Monbaron et al., 1999).

In addition to the dinosaurs, the pterosaurs were particularly diversified in the Tafilalt region. Three Pterodactyloid families have been recognized (Kellner et Mader, 1996, 1997; Mader & Kellner, 1997, 1999; Wellnhofer & Buffetaut, 1999). All ecological niches were occupied. Many species of "fish" populated the rivers, among them fresh water sharks, a giant coelacanth, dipneusts and bone fishes actinopterygians, whose remains are remarkably preserved, sometimes including the muscle fibres (Russel, 1996; Cavin & Dutheil, 1999; Dutheil, 1999; Cavin et al., 2001). According to the latter authors, the Cenomanian-Lower Turonian ichtyofauna of the Tafilalt and neighbouring areas includes 680 taxa from both fresh water and marine environments. At least five species of crocodile occupied the river shorelines, including a giant (Buffetaut, 1976, 1989a, 1994; Sereno et al., 1996). Four fresh water turtle families have been described (de Broin, 1988; Tong & Buffetaut, 1996; Gmira, 1995; de Lapparent de Broin, 2000; Gaffney et al., 2002). Snakes and varanus are also present in the area (Rage & Escuillé, 2003). Besides of the Vertebrates, insects and angiosperms complete this uncommon, 110–95 My old terrestrial environment.

The Goulmima plateau enables paleontologists to document a slightly younger period in a marine environment. The Vertebrate fossils are preserved within calcareous nodules included in the Lower Turonian marly limestones, also dated by their ammonites. The environment corresponds to an open marine platform, probably rich in plankton. Numerous new fish species have been described (one of them named *Goulmimichthys*), together with more cosmopolitan species. The reptilian fauna (Bardet et al., 2003a, b) includes mosasaurs and plesiosaurs, which were typical predators of the Cretaceous seas. Among the plesiosaurs, both the Elasmosaurids (characterized by their long neck) and the Polycotylids (shaped as good swimmers; Fig. 7.19) have been described. The mosasaurs from Goulmima belong to the earliest forms of this group, whose spectacular radiation occurred around 90 Ma, and which populated the most varied marine niches up to the K-T crisis.

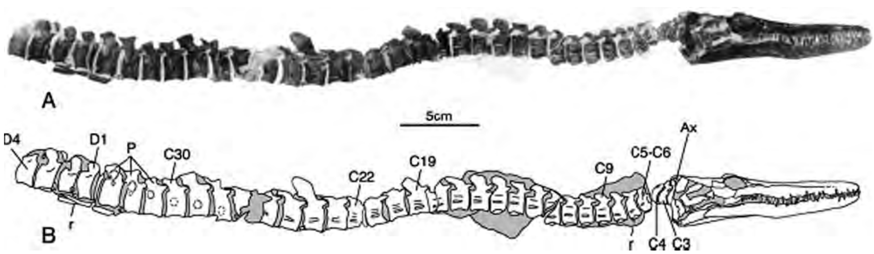

Fig. 7.19 *Thililua longicollis*, a small, but nicely preserved plesiosaur skelton (**A**) from the Turonian beds of Goulmima. – **B**: Interpretation: Ax: axis; **C**: cervical vertebra; **D**: dorsal; P: pectoral; r: rib. After Bardet et al. (2003a)

7.4 Vertical Movements in the Plateau Domains

From the tectonic point of view, two questions arise concerning the Cretaceous-Tertiary plateaus, (i) why does the Triassic-Jurassic sequence is lacking (no deposition or erosion?), and (ii) how much the plateaus were affected by the Atlas deformation during the Late Eocene-Neogene?

7.4.1 Atlantic Plateaus

The lack of Triassic-Jurassic deposits at the bottom of the Phosphate Plateau and Ganntour has been taken for years as evidence for a subaerial exposure of the Western Meseta basement during this time interval. This was the concept of "Terre des Almohades" of Choubert & Faure-Muret (1960–62) or "Dorsale du Massif hercynien central" of du Dresnay (1972, in Michard, 1976), currently referred to as the West Moroccan Arch (WMA; see Chap. 4, Figs. 4.4, 4.7). In the classical view, this land was emergent up to the late Early Cretaceous, except locally (Fig. 4.17). However, new studies based on low-temperature thermochronology of basement rocks (mostly granites) make compulsory to revisit this concept (Ghorbal et al., 2007; Saddiqi et al., 2008).

In fact, the WMA suffered contrasted vertical movements during the Triassic-Jurassic interval. The results obtained by Saddiqi et al. (2008) on the Jebilet and Rehamna granites are summarized hereafter (Fig. 7.20). These Variscan granites, whose emplacement occurred at about 330 Ma and 300–290 Ma, respectively (see Chap. 3, Fig. 3.23), yielded apatite fission-track (AFT) ages around 192 Ma and 152 Ma, respectively. Their low-temperature T-t paths have been modelled using, in addition to the age dataset, the fission-track length measured in the same samples, and taking into account the available geological constraints. The Jebilet T-t path presented here (Fig. 7.20C) clearly indicates that the studied granite suffered a weak heating ($T < 75\,^{\circ}C$) after its Permian exhumation (around 260 Ma). In other words, the Jebilet basement subsided by about 1500 m during the Triassic-Early Jurassic (admitting a slightly improved geotherm in relation with the context of rifting and magmatic activity). The Rehamna granite, which emplaced later and at greater depth, was not completely exhumed at 260 Ma, and then reached a slightly higher temperature (ca. 90 °C) at 180 Ma.

Both the Jebilet and Rehamna massifs reached again the surface around 120–95 Ma, i.e. before the widespread Cretaceous onlap. Thus, the major role of the Late Jurassic-Early Cretaceous uplift is clearly illustrated here, as in the Zaer Massif (Ghorbal et al., 2007). This uplift event was responsible for the active denudation of the Variscan basement (see Fig. 7.5B above) of the former Western Meseta platform domain, which changed to a complex of emerged islands and peninsula (Fig. 4.17).

The last parts of the modelled T-t paths show the slight burial heating related to the Cretaceous Eocene sedimentation and the more rapid exhumation during

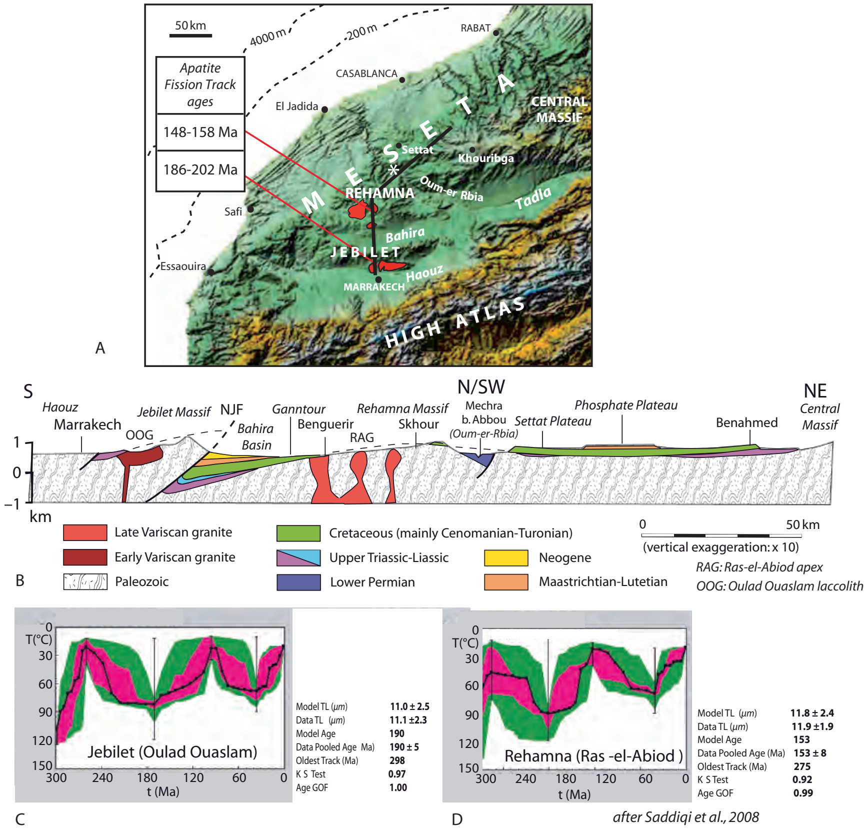

Fig. 7.20 Vertical movements in the south-western Moroccan Meseta (Rehamna and Jebilet Massifs), based on stratigraphy and apatite fission-track (AFT) data, after Saddiqi et al. (2008). **A**: Location of the studied massifs and corresponding AFT ages from their Variscan granites. Bold black line: location of cross-section **(B)**. – **B**: Cross-section showing the different types of granites and the geologic constraints used in the modelling experiments (Permian erosion, Triassic-Liassic onlap, Cretaceous unconformity, post-Lutetian uplift). – **C**, **D**: Examples of modelling experiments (AFTsolve program) for the Jebilet **(C)** and Rehamna **(D)** granites. The importance of the pink areas respective to the green ones in the T-t diagrams shows the excellent fit of the modelled T-t paths with the AFT data, as also shown by the quantitative indications on the right. TL: track lengths

the Late Eocene-Neogene interval. This final exhumation is coeval with the Atlas shortening and, indeed, the Jebilet massif can be regarded as part of the Atlas system, being bounded to the north by an important reverse fault. However, it must be kept in mind that the uplift of all the deformed regions south of the Maghrebide collisional belt (Meseta, Atlas and Anti-Atlas domains) does not only rely on crustal shortening, but was greatly enhanced by the development of a hot mantle anomaly underneath (e.g. Missenard et al., 2006; Chap. 1, Sect. 1.5; Chap. 4, Sect. 4.5, 4.6).

7.4.2 Saharan Plateaus

The behaviour of the future Hamada domain (i.e. the central and eastern Anti-Atlas and Ougarta Belts, and the Tindouf, Reggane and Bechar Basins) during the Triassic-Early Jurassic was recently questioned. In the eastern Anti-Atlas, Robert-Charrue (2006) argued that the Triassic extension would have been able to reactivate, again as normal faults, the N-dipping Cambrian paleofaults inverted during the Variscan compression. Actually, the Triassic-Liassic extension allowed gabbroic magmas of the CAMP (cf. Chap. 1) to emplace as huge dykes into the crust (e.g. Chap. 2, Fig. 2.2), and to form a number of sills in the already folded Paleozoic series (Hollard, 1973). However, the Triassic-Early Jurassic facies of the Atlas southern border strongly suggest the proximity of emerged lands southward (see Chap. 4, Figs. 4.7, 4.8). Accordingly, we may assume that the Anti-Atlas formed at that time the emergent shoulder of the High Atlas rift.

The entire Sahara domain was deformed significantly during the Early Cretaceous, particularly during the late Barremian-early Aptian, and then during the Late Cretaceous (see above, Sect. 7.1). Detail paleomagnetic study ("fold test") of Triassic-Early Jurassic sills intruded in the folded Paleozoic series of the Reggane-Ahnet region demonstrated that folding has been actually initiated during the late Paleozoic, but was significantly rejuvenated after intrusion, likely during the Cretaceous (Smith et al., 2006). A larger-scale consequence of this intraplate deformation was the shift of the Trans-Saharan Seaway to the west of the Hoggar Massif during the Senonian (see above, Fig. 7.3.C).

The Anti-Atlas was clearly affected by the Atlas Orogeny during the Cenozoic. Varied structures have been mentioned above (e.g. Figs. 7.11, 7.13, 7.14), being related to the Atlas shortening. Some Variscan reverse faults were reactivated in the Paleozoic basement at the northern border of the Saghro Massif, as documented by apatite fission track results (Malusà et al., 2007). In contrast, at the southern border of the Eastern Anti-Atlas, Baidder (2007) described a set of post-Cretaceous, ENE-trending normal faults. As a matter of fact, the interpretation of the post-Cretaceous tectonic regime in the Sub-Saharan regions is still widely open to investigation. The regional uplift, particularly that of the Anti-Atlas, was strongly enhanced by the mantle anomaly quoted above, i.e. the Morocco Hot Line, also responsible for the high elevation of the Atlas Mountains. Nevertheless, the trend of the Anti-Atlas antiform, parallel to the South Atlas Front, and the chronology of its formation (Late Cretaceous-Neogene) strongly suggest that its uplift was also controlled by the stress-induced lithosphere buckling due to the Africa-Europe convergence (Frizon de Lamotte et al., 2008).

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Chapter 8
The Quaternary Deposits of Morocco

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8.1 Introduction

Quaternary deposits are widespread in Morocco, being especially important both on the Atlantic coast as well as in the continental basins (e.g. Saiss, Tafilalt, etc.). The exceptional outcrops of the Atlantic margin have enabled precise analyses of the glacio-eustatic evolution of sea level during the last two million years. This is an important base for the study of the recent tectonic evolution of Morocco. The present chapter offers a synopsis of the state of the art, exposes the difficulties in dating the successive Quaternary stages, and takes into account the improvements acquired during the last two decades.

8.1.1 The Pleistocene Lower Limit, and Correlations With the Global Climate Chart

The base of the Quaternary = Pleistocene epoch is conventionally placed near 1.8 Ma, but it may soon be placed at 2.6 Ma (base of Gelasian, currently the youngest Pliocene stage) in order to take into account the climatic evolution (glaciations) and the earliest evidence of man (stone artefacts). The Pleistocene thus would extend

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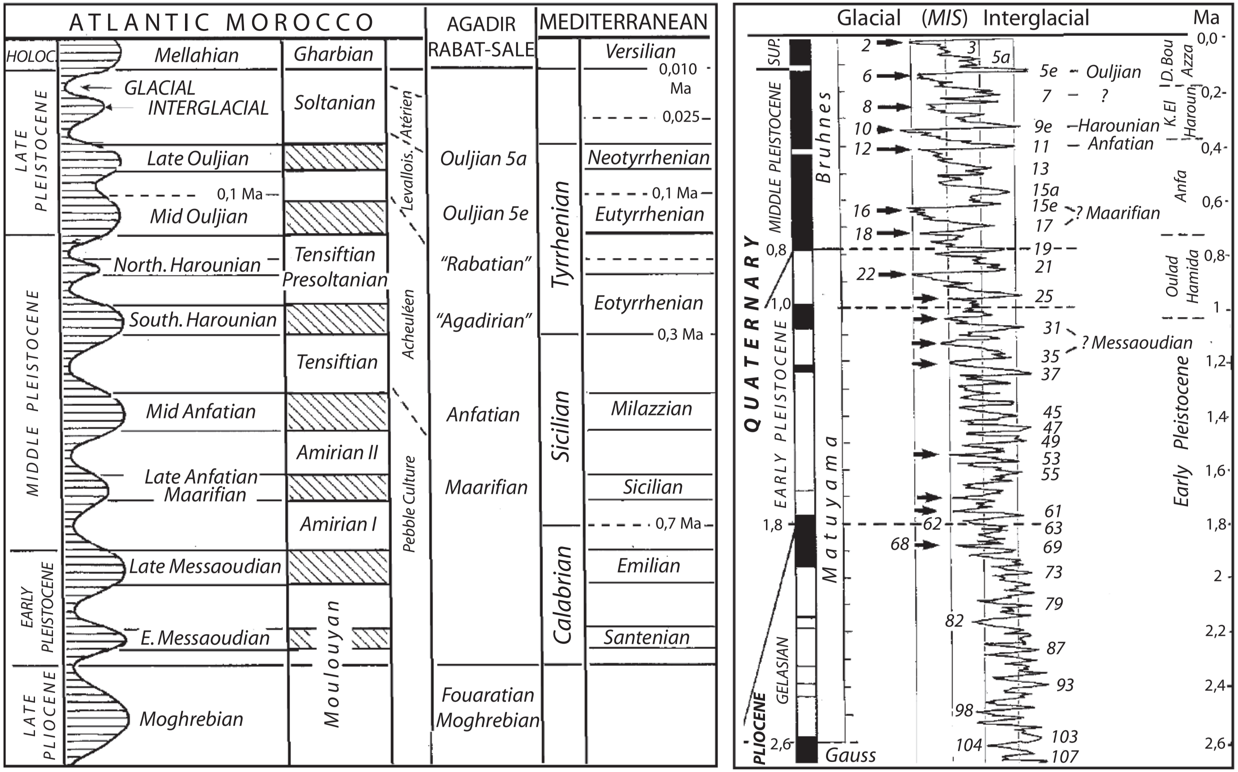

Fig. 8.1 Classic stratigraphy of the Quaternary deposits of Morocco (*left*) and new version, based on the marine isotope stages (MIS, *right*), with correspondence between the marine/continental Moroccan "stages" and those defined in Mediterranean (*left*), Horizontal/oblique pattern: transgression/regression

at the expense of the Pliocene, according to the climatic deterioration recorded by continental floras and faunas (Villafranchian mammals) and, in the marine domain, by the extreme variations of the oxygen isotopes (Marine Isotope Stages, MIS) used for worldwide correlations of sea level evolution. The present Quaternary limits include MIS 1 to MIS 61 (1.765 Ma), but in the near future may extend up to MIS 104 = 2.6 Ma (Fig. 8.1, right). The remarkable Atlantic sequence of raised beaches was recently correlated with this modern chronology while the classic terminology of the "Moroccan stages" (Fig. 8.1, left) was not abandoned. The reliability of the correlation scale of course weakens with increasing time, and the transfer of the Moghrebian stage to Quaternary *versus* Pliocene series remains questionable.

Another disputed limit occurs in the marine Moroccan series, "one of the most complete Pleistocene successions of the world" (Howell, 1962 in Stearns, 1978), which is the Early/Middle Pleistocene boundary. This limit is currently located at 0.780 Ma, i.e. *after* the significant change that occurred between 1 and 0.9 Ma from the 41,000 year periodicity of warm to cold episodes, characteristic of the Pliocene-Early Pleistocene times, to the 100,000 year periodicity with longer and more intense glacial episodes. Based on field data such as the Moroccan record itself, Quaternarists plan to move the limit back.

The classic description of the Moroccan Quaternary Atlantic sequence was summarized in a stratigraphic chart of regional "stages", which compete with the Mediterranean "stages". This chronostratigraphic terminology has been questioned since 1994 based on lithostratigraphic studies (sequence stratigraphy) correlated with the global, glacio-eustatic chronology (MIS). We propose (Fig. 8.1) the necessary correspondence key between the modern and ancient nomenclatures. However, it is essential in every case, and especially in the case of the Moroccan continental "stages", to justify the use of the ancient terms, and to use them with greatest caution.

8.1.2 Brief History of Research

G. Lecointre is the pioneer in Moroccan Quaternary research. As early as 1926 he recognized 4 down-stepping sea levels south of Rabat, defined by their respective molluscan faunas and local altitude, but he excluded a general levelling (contrary to Ch. Depéret) as he considered that the epirogenic variations were likely to have modified the depositional altitude. The many subsequent works based on the series around Casablanca and Rabat (by Antoine, Arambourg, Choubert, Ennouchi, Gigout and Lecointre himself) have been synthesized by Biberson in order to help correlate prehistoric research (1961a, b). The fortuitous encounter of remote Palaeolithic artefacts and fossil mammals was followed by active researches, which resulted in the discovery of the "man of Sidi Abderrahman" (Arambourg & Biberson, 1955), now considered as a *Homo erectus*. This drew attention to the Atlantic littoral series (Fig. 8.2), initiating the difficult question of their correlation with the continental deposits: synchronism of pluvial and glacial episodes,

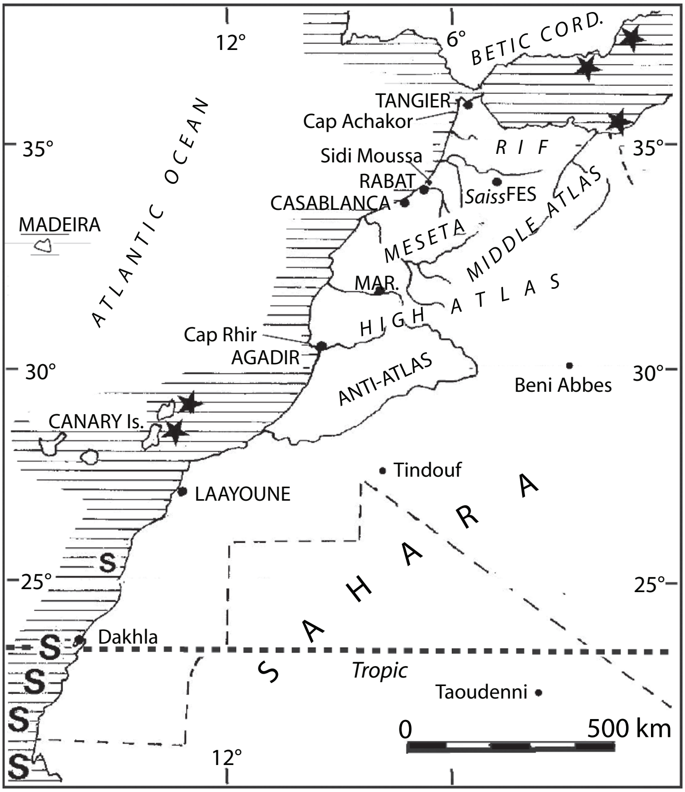

Fig. 8.2 Location of the localities and regions cited in this chapter, and extension of two fossil species characteristic of warmer climates than present. ★: *Strombus latus* from the Late Pleistocene (MIS 5.5 = 5e). **S**: *Senilia senilis* from the Holocene optimum (MIS 1). After Meco et al. (1995) and Rosso & Petit-Maire (1978)

calcrètes chronology, paleoclimatic significance of shoreline dune growth, etc. (e.g. Beaudet, 1971; Weisrock & Rognon, 1977, Weisrock, 1980–1993; Weisrock et al., 1999).

Since 1965, geochronologic studies based on the Th^{230}-U^{234} and ^{14}C dating of the marine shells (e.g. Stearns & Thurber, 1965), as well as biostratigraphic studies applied to the classic series or to new collections (Brébion, 1973, 1978, 1980; Brébion et al., 1984; Weisrock et al., 1999) have been used in the frame of the six Moroccan “classic stages”. On the other hand, the researches initiated by the prehistorians of the Casablanca Program demonstrated that at least 20 sea level units may be identified (Texier et al., 1994; Raynal et al., 2001; Lefèvre & Raynal, 2002; Texier et al., 2002; Sbihi-Alaoui & Raynal, 2002). The progressive elaboration of their sequence stratigraphy and the increase of the number of lithostratigraphic units suggest that this new state of knowledge is not yet finalized. The sequence chart appears to be well established up to about 1 Ma, but older formations down to the Messinian recognized in the Mediouna Formation (Fm) still require further study.

For example, the name of Maarifian stage may apply to 8 interglacial high sea levels spread over more than 500,000 years.

In the most recent years, detail seismic surveys have been used to decipher the most recent evolution of some parts of the Moroccan continental shelf. For example, Le Roy et al. (2004) described a succession of terraces offshore El Jadida, and the extension of the Oum-er-Rbia River prodelta. Likewise, Baltzer et al. (submitted) imaged the gravity-driven deformation of the Quaternary sediments shed over the continental slope offshore Dakhla in the Southern Morocco (see Chap. 1, Fig. 1.22). However, in the following we focus on the emerged marine sequences that permit to define precisely the stratigraphic chart of the Moroccan coastal deposits.

8.2 The Littoral Sequence Around Casablanca

In the Casablanca region, each sea level high stand is preserved as beach deposits at the foot of a cliff entrenched in previous marine and eolian deposits. Generally, the beach deposits progressively merge upward into lithified dune sands (eolianites), which yielded important volumes of building stone extracted from large quarries. The latter excavations favoured the precise identification of the lithostratigraphic units following their basal, subaerial/marine erosional surface, their interglacial, marine terrace deposits, and the subsequent continental deposits that begin before the end of the interglacial episode. These eolian build-ups form long bands parallel to the shoreline, amalgamated close to the coast and more spaced inland, and extending up to 10 km from the coast (Figs. 8.3, 8.5; see also in Chap. 4 a satellite view of the paleodunes south of Essaouira, Fig. 4.32).

The Casablanca area offers the richest record for the Moroccan Quaternary stratigraphy, not only due to numerous large quarries, but also to the regional epirogenic regime. The 115 m altitude of the oldest Pliocene transgression suggests a limited uplift prone to register all the major eustatic variations of sea level. However, it is uncertain whether the rate of uplift was constant because the last interglacial (isotopic substage 5e = Eemian = MIS 5.5) is 4 m high instead of the 6–8 m worldwide standard. The corresponding mild subsidence over the last 100,000 years seems to have ceased as the Holocene transgression ("*Mellahian*" = Flandrian = MIS 1) is located between 0 and 2 m as everywhere, and especially in the stable Mediterranean areas. This Holocene transgression curiously yielded slightly younger ages (3.7–2.2 ka, ^{14}C age) than the corresponding Mediterranean high stand (6–4 ka).

Going back in time, the second high sea level is the *Ouljian* stage (MIS 5.5, ~ 125ka), most frequently characterized by its mollusc fauna, and best known from Agadir (Fig. 8.4). It is referred to an especially warm interglacial episode. The equivalent fauna in the Mediterranean realm is known under the name of "Senegalian fauna", and characterized by *Strombus bubonius* (or *Strombus latus*), which currently lives from Angola to Mauritania and Cabo Verde Islands. However, during MIS 5.5 this warm fauna occurs in the Canary Islands, but is lacking from the

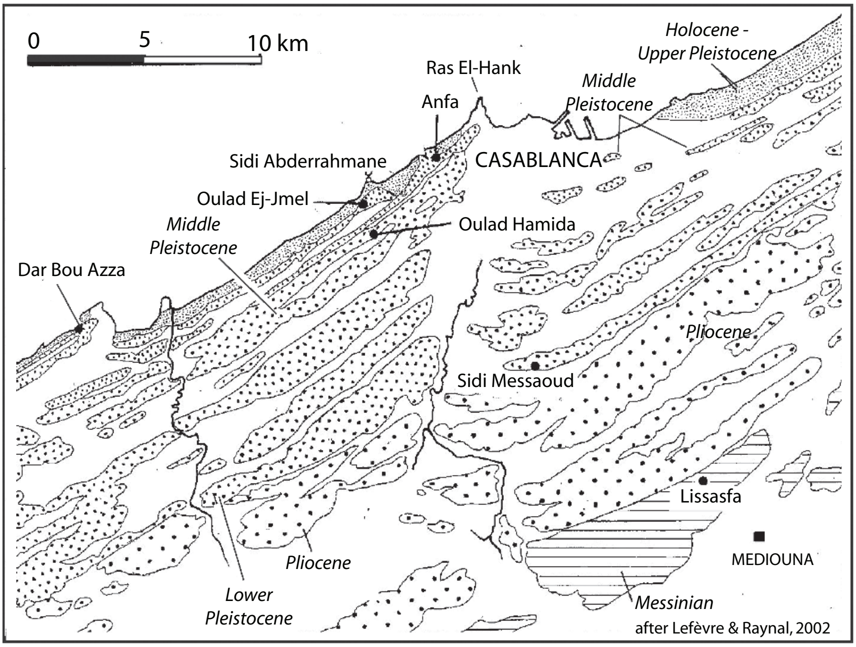

Fig. 8.3 Schematic map of the Miocene to Quaternary deposits of the Casablanca region, after Lefèvre & Raynal (2002)

Fig. 8.4 Ouljian deposits from the Agadir coast south of Tamghart (see Fig. 8.7 for location). – **A**: Pebbly beach deposit with sand matrix and shell fragments overlapping Cretaceous sandstones. The Ouljian beach deposit is overlain by eolianites (calcarenites) with cross-bedding and vertical root channels. – **B**: Storm deposit of piled shells from sandy shore (*Solen, Natica, Cardita*…) and rocky environments (*Patella*)

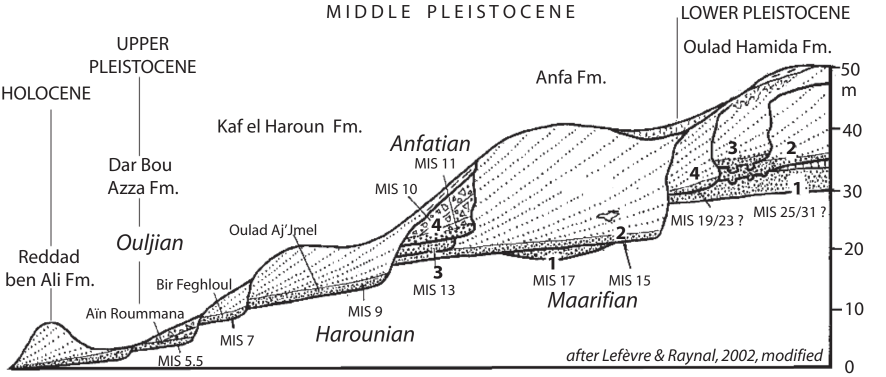

Fig. 8.5 Cross-section of the morpho-sedimentary units of the Casablanca littoral Lefèvre & Raynal (2002). Correlations are suggested with the high glacio-eustatic levels from MIS 5.5 to MIS 31 (?) by reference to their respective altitude resulting from the epirogenic uplift of the coast

Atlantic coast of Morocco (Fig. 8.2). This surprising distribution seemingly results from deep cold water upwelling along the Moroccan shoreline, a hydrologic process that also bounded the *"Arca" senilis* extension from Senegal to southernmost Morocco during the Holocene optimum.

The third high sea level, identified between years 1994 and 2000, has not received any "stage" name. Its marine deposits are referred to MIS 7, ~ 200 ka, which is the last glacio-eustatic high stand of Middle Pliocene. The fourth interglacial transgression is better known, and was labelled *Harounian* at Casablanca. Its fauna and 9–13 m altitude suggest a correlation with MIS 9, ~ 325 Ma. In the 80s, erroneous datings on complementary sections close to Rabat and Agadir led to the splitting of the Harounian into the "Rabatian" and "Agadirian", two stages excluded from the chart according to our own studies (Fig. 8.6).

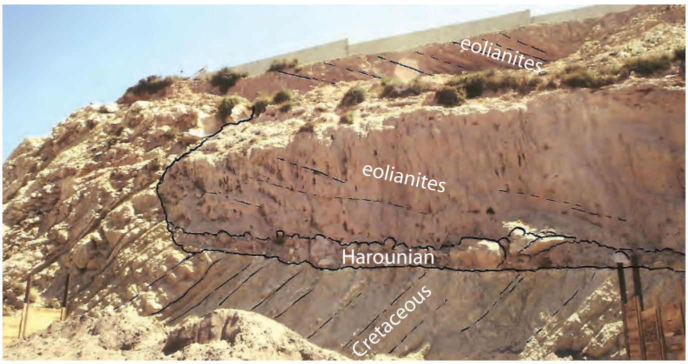

Fig. 8.6 The Harounian marine terrace from the Agadir port, preserved at the foot of a cliff notch in the Cretaceous beds. The pebbly marine deposits are overlain by eolian calcarenites that record the continental episode subsequent to the Harounian culmination of sea level

The fifth high stand deposit is the youngest of a set of marine deposits whose marine erosional platforms are located around 20 m a.s.l., but locally attain 26 m. This is the *Anfatian* “stage”, whose malacofauna includes the last representative Early Pleistocene species (*Acanthinus plessisi*, *Calyptraea Trochita trochiformis, Sphaeronassa mutabilis*). The Anfatian is assigned to MIS 11, one of the clearest interglacial episodes, named Holsteinian in northern Europe, and dated at 400 ka. The slightly older, sixth high sea level deposit stands at 22 m a.s.l., and displays a relatively cool malacofauna (*Nucella lapillus*, *Littorina littorea*, typical for the present-day French-Cantabric province). Thus, this (unnamed) level can be assigned to MIS 13, the coolest of the Mid Pleistocene interglacial episodes according to the oxygen isotopes. The seventh level (also unnamed) also culminates at 22 m a.s.l.; it displays climatically intermediate malacofauna, and may be referred to MIS 15a. The eighth marine deposit, locally observed below the preceding ones, records another cool interglacial stage that may be assigned to MIS 15e or to MIS 17 (to which the authors of the 2002 publications refer). Both last units are correlated with reserve to the *Maarifian* defined on contradictory data from two nearby quarries at 50–60 m and 20–22 m altitudes, respectively.

The following (older) set of units is difficult to subdivide, although it corresponds to an obvious erosional platform 10 m higher than the Anfatian-Maarifian one. This important vertical gap may record a sedimentary hiatus or an increased rate of uplift. The set includes five marine deposits, and was named the “Oulad Hamida morpho-sedimentary unit, or Fm”. It has been referred, according to the different authors, to 3 or 5 interglacial stages, the most recent ones (9th and 10th?) being assigned to MIS 19, which is the base of Middle Pleistocene (magnetic reversal Matuyama/Bruhnes, 0.78 Ma). The 11th (?) marine deposit is 3 m higher; it is correlated with doubt with MIS 21. As the partition of the following marine levels (12th, 13th?) is somewhat doubtful, they may perhaps represent the first interglacial stage of the extended Middle Pleistocene (MIS 25), at the turn in the periodicity of cold and warm climates, shortly after 1 Ma.

Four older marine units have been identified and referred to the Early Pleistocene (Fig. 8.3). One of these units must be the type-section of the classic *Messaoudian* “stage”. The next three marine units post-dating the Miocene Mediouna Fm would be the only deposits recording the Pliocene. Thus, it is likely that hiatuses (non-deposition or erosion) have affected the lowest part of the Plio-Quaternary series. The *Fouaratien* and *Moghrebian* classic names should be used for these post-Miocene units, although it is difficult to justify their respective assignations. Considering the Moghrebian mollusc fauna, this “stage” seems to be essentially Pliocene in age. The Fouaratian, defined near Salé, yields the first Early Pleistocene malacofauna (*C. trochiformis, A.* p*lessisi*) and Villafranchian vertebrates. The downward extension of the Pleistocene (up to 2.6 Ma) may solve this dilemma, by the introduction of these puzzling terms in the Quaternary times.

8.3 Other Marine Sequences

8.3.1 From Rabat to the Mediterranean Shoreline

Compared to the region of Casablanca, the Rabat-Salé coast has few large quarries opened in the calcarenite belt. Interest therefore was transferred to the shoreline cliffs (Aberkan, 1989), valued by the prehistory investigators. Anfatian is identified by its molluscs below the level that yielded *Homo erectus* at Sidi Bouknadel, and whose fossil rodents suggest an age < 400 ka. Complex relative relationships of the continental deposits (more or less karstified superimposed eolianites with paleosoils and calcretes), uncertain neotectonic movements and subsidence at the margin of the Gharb Basin hamper interpretation.

The Sidi Moussa sequence (Plaziat et al., 2006) demonstrates that the Ouljian includes two sea level high stands, close to each other, whose ages fall within MIS 5.5 (last interglacial optimum). The marine and eolian deposits show synsedimentary deformations of seismic origin. Further north, the Cape Achakar series display three entrenched marine units (Alouane, 1997). The most recent is unquestionably Ouljian in age but, remarkably, it is not the unit dated by Stearns and Thurber (1965) at 125 ± 10 ka on a "pre-Ouljian shell" with 70% of calcite (which indicates significant diagenesis). This incorrect date was taken until now as the best dating of the last interglacial, a paradox that draws attention on the need for a complete reappraisal of the Th^{230}-U^{234} – based correlations of the ancient stratigraphic nomenclature.

Along the Rif Mediterranean coast, a recent tectonic mobility has been recognized for long (e.g. Cadet et al., 1977). Accordingly, the varying terracing and the stepping of the marine cliff notches cannot be assigned to eustatic events without cautious geochronological evaluation.

8.3.2 The Atlas Coast North of Agadir

The Quaternary outcrops between Agadir and Cape Rhir strongly contrast with those around Casablanca (Fig. 8.7). With the exception of the highest terrace (~ 200 m a.s.l.) assigned to the Moghrebian (Pliocene), the Quaternary outcrops are narrow marine terraces, frequently restricted to notch deposits (the "cave" deposits of Agadir Port) at the foot of shoreline cliffs cut across steep Cretaceous or Jurassic substrate. The field studies by Ambroggi (1963) and Weisrock (1980–1993) have been completed by biostratigraphic data based on the rich malacofauna (from Brébion, 1979, to Weisrock et al., 1999), and by Th^{230}-U^{234} dating of shells. A recent French-Moroccan research program managed by J.-L. Reyss, the results of which are mostly unpublished, also relies on the same approaches. However, the latter program develops a critical methodology for the interpretation of mollusc datings, in order to check the previous dates. The Ouljian stage, whose ancient dates are younger than that of the last interglacial (Hoang et al., 1978), was characterized

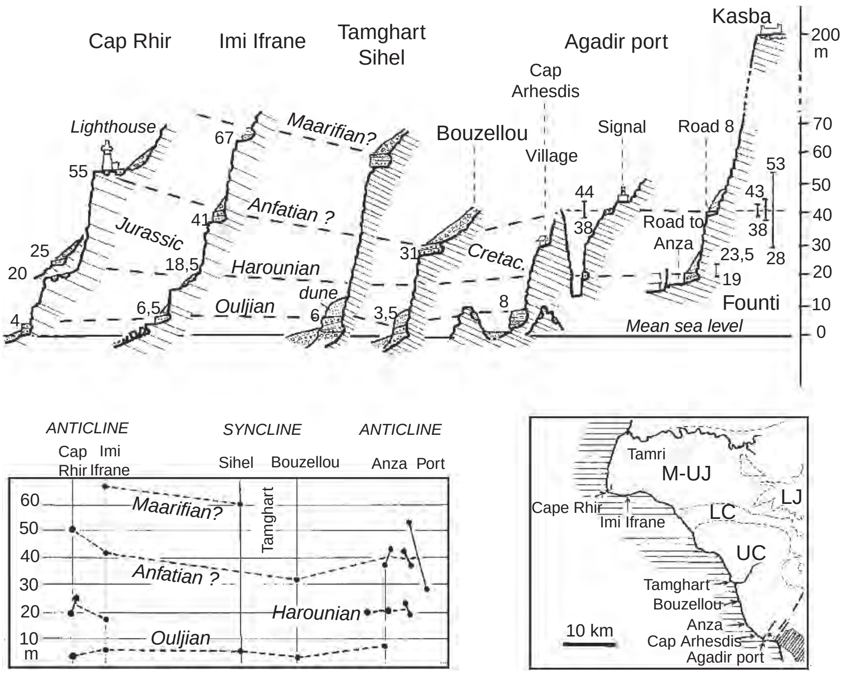

Fig. 8.7 The deformed terraces between Agadir and Cape Rhir.– *Lower inserts*: Diagram of altitude variations (*left*) and location map (*right*). Abbreviations: LJ: Lower Jurassic; M-UJ: Middle-Upper Jurassic; LC: Lower Cretaceous; UC: Upper Cretaceous. Local altitude variations are related to minor block faulting, whereas the regional synclinal deformation, which affects the older terraces (Maarifian and Anfatian) corresponds to that observed in the Mesozoic substrate

by 78 new measurements (Choukri et al., 2001–2007). The results span from 132 to 40 ka (Fig. 8.8), which indicates variable rejuvenation. A number of analyses from selected localities revealed the role of the diagenetic processes (leaching of organic matter in bio-minerals, post-burial incorporation of uranium) in the observed rejuvenations. The only reliable dates are the oldest ones, which resulted in virtually full agreement with the assumed age of the studied Ouljian deposits. This terrace eventually fits the MIS 5.5 age, not the MIS 5.3 (5c) or MIS 5.1 (5a) as previously suggested.

Similarly, the southern Harounian or "Agadirian" deposits now characterized by 28 dates from 314 to 80 ka must belong to MIS 9 as common with the Harounian of Casablanca. However, the higher and variable altitude of the southern Harounian, 20–23 m a.s.l. instead of 9–13 m around Casablanca, indicates more active neotectonic deformation along the Atlas coast relative to the Meseta. This corroborates the probable role of co-seismic ramp deformation (Meghraoui et al., 1998) induced by the blind thrust of the active South Atlas Front (see Chap. 4, and Sébrier et al., 2006).

The subsequent 30–50 m high marine terrace is known as Maarifian, but is probably Anfatian: an age somewhat older than 300 ka based on 15 analyses suggests a

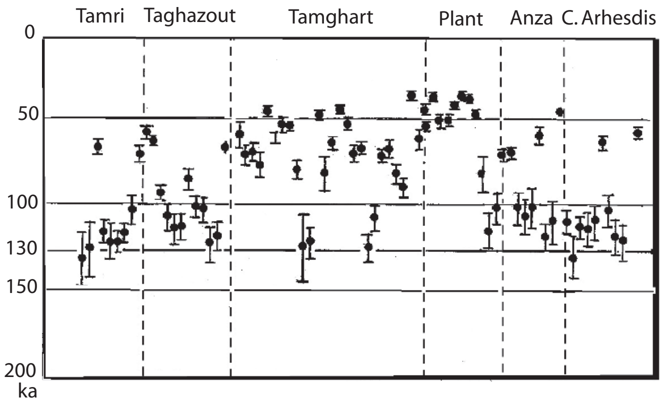

Fig. 8.8 Distribution of Th^{230}-U^{234} dates measured on mollusc shells from various Ouljian terraces (MIS 5.5 = 5e) between Agadir and Tamri, north of Cape Rhir, after Choukri & Reyss (unpublish. report). The date dispersion results from diagenetic processes that affected the shells after their fossilization in the Ouljian deposits, ~125 ka ago

correlation with MIS 11. The fourth terrace (59–67 m a.s.l.) identified locally close to Cape Rhir can be regarded as Maarifian, although without significant data except altitude.

8.4 Continental Sedimentation

8.4.1 Problems of Correlation With the Marine Sequences

Moroccan continental "stages" were introduced into ancient terminology in order to name the periods between the interglacial culminations of sea level (Fig. 8.1, left). This was based on the littoral eolian deposits associated with colluvium, and on the conventional belief that alluvial terraces could be referred to pluvial episodes and finally, correlated with glacial pulses and low stands of sea level. A subsequent revaluation of this classic belief led to the twofold equation: terraces = interglacial episodes = pluvial (Beaudet, 1971). This was the first step toward the complete desynchronization of continental sedimentation and erosion phases with respect to climatic maxima, either cold or warm (Weisrock & Rognon, 1977), eventually ending with the concept of prevalence of the destabilization phases (Lefèvre, 1989). According to this new way of thinking, the warmer and damper interglacial episodes would be the most favourable for the vegetation and pedogenesis (Fig. 8.9). Conversely, the restricted vegetation cover linked to glacial cooling would favour erosion and deposition of alluvium and colluvium, stabilized during the arid periods of peri-glacial stages (under temperate or cold local conditions). The return of heavy rains during the subsequent warming stage would be responsible for terrace cuts

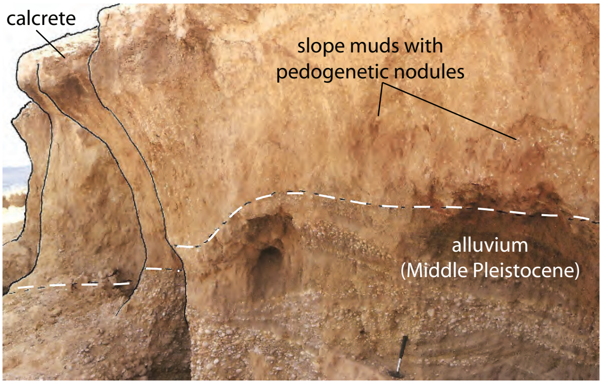

Fig. 8.9 Carbonate crust on top of Middle Pleistocene alluvium from the Saiss Basin north of the Middle Atlas El Hajeb Causse. The pebbly alluvium contains onconliths. Pedogenetic nodules develop within the overlying muddy colluvium, and eventually merge into the thick calcrete. Note the hammer at the bottom

by waters relatively poor in sedimentary suspension (due to the increased vegetation cover). Hence the value of the classic “continental stages” must be questioned according to the local conditions.

8.4.2 Calcretes and Travertines

Calcrete (“croûtes calcaires”, “caliches”) are ubiquitous in Morocco, being a nuisance for farming, and subject of early researches by pedologists. According to the modern ideas (e.g. Verrecchia & Freytet, 1987; Verrecchia et al., 1995), their origin and climatic significance should be reconsidered. Origin is always complex: lateral sedimentary nourishment, pedologic processes involving dissolution of carbonate grains and lateral and downward ion migration and precipitation, subaerial biogenic coating forming laminar horizons (“croûtes zonaires”) comparable to stromatolites. We recall the paradoxical occurrence of calcretes on the Anti-Atlas granite massif. Carbonate crusts and feldspath epigenesis result from the pedogenetic reworking of a sheet of eolian deposits enriched in minute marine bioclasts from the marine plateau emerged during the glacial epochs.

Age diversity of the innumerable calcretes intercalated in the Pliocene to Quaternary glacis and alluvial terraces suggests that favourable conditions for their growth have been very frequent, being characterized by a subarid climate with contrasted seasons and rainfall. The surficial crusts are polygenetic, their armouring properties favouring their longevity. Generally speaking, calcretes are far from being evidence of hyperaridity. They reflect the complex variations of the Atlantic and Mediterranean rainfall before and after the most arid episodes.

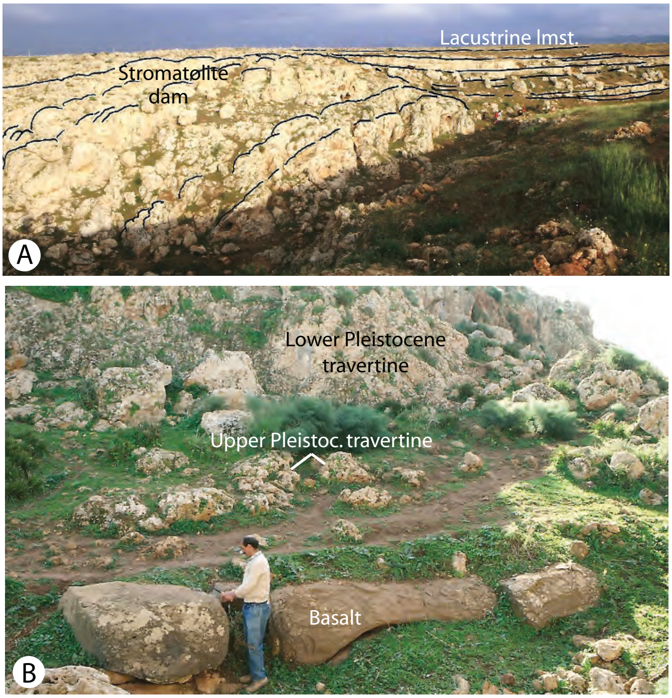

Fig. 8.10 The Saiss Limestones in the Fes region, at the northern border of the Middle Atlas "Causse". – **A**: Natural cross-section of Early Pleistocene stromatolite build-ups, behind which horizontal lacustrine beds formed upstream. – **B**: Basalt flow emplaced in a gully cut in the Early Pleistocene travertine dam, and overlain by upper Pleistocene travertines. The basalt has been dated at 0.9 Ma

Another type of carbonate deposit is also well developed in Morocco, i.e. the spring travertines, fluviatile stromatolitic dams, and associated lacustrine and palustrine deposits (Fig. 8.10). The proximity of the Atlantic Ocean and the widespread occurrence of karst explain the frequency and repeated activity of highly mineralized resurgences. Their study is currently limited to lithostratigraphical monographs (Weisrock, 1993; Ahmamou, 2002), which lack precise dating. The travertine dams of the northern flank of Middle Atlas (Fig. 8.10) extend laterally for several kilometres. They developed from the Late Pliocene to the Holocene, as shown by the occurrence of an entrenched and interbedded basalt flow dated at 0.9 Ma, and by Acheulean artefacts (Onoratini et al., 1992). Such carbonate deposits illustrate a new type of fluvio-lacustrine complex that record the long-term variations of pluvial episodes.

8.4.3 The Saharan Quaternary Record

This vast, presently hyperarid area indeed has been characterized by quite varied geographies during the Quaternary pluvial episodes (e.g. Conrad, 1971; Fabre, 2005). During the Plio-Villafranchian, an endoreic system developed with the Ziz-Daoura and Guir-Saoura rivers flowing into a large lake rich in fish, *Cardium* and ostracods, that extended mainly in Algerian Sahara (Fig. 8.11). After a long arid period, another pluvial episode is recorded, which permitted the development of vegetation along the rivers, and the alimentation of deep aquifers whose emergence formed several lakes. Another long essentially arid period was followed during the Holocene by the last humid episode that favoured the Neolithic cultures in the Saharan settlements.

The Sahara Quaternary deposits include three types of crusts, i.e. calcretes, silcretes and ferricretes. The latter two developed mostly at the fringe of the Hoggar crystalline massif. Generally speaking, crusts grow within the colluvium or alluvium sediments, at the top or near the exurgence of water tables. They are laterally

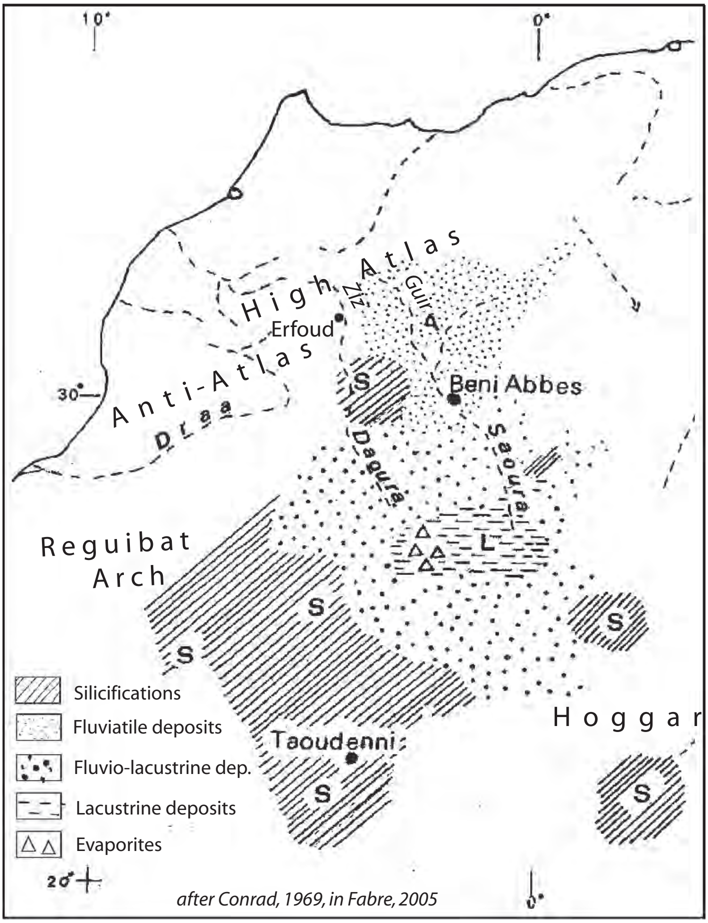

Fig. 8.11 Schematic map of the Plio-Villafranchian deposits in the west Saharan domain, after Conrad (1969), in Fabre (2005), modified

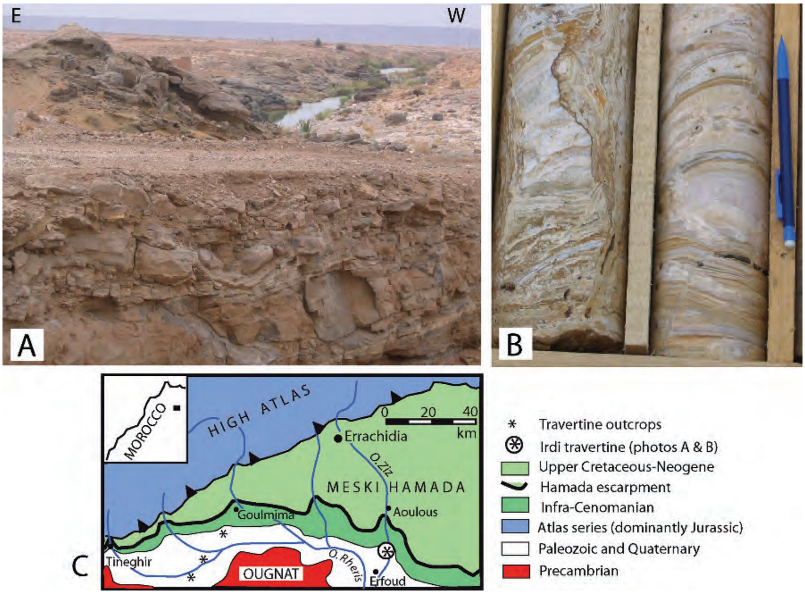

Fig. 8.12 The Middle-Upper Pleistocene Irdi travertine (Boudad et al., 2003) south of the Meski Hamada. **A**: Roadcut outcrops at the bridge over the Oued Ziz north of Erfoud. – **B**: Examples of cores drilled in the latter outcrops and exhibiting typical laminated facies affected by dislocations and karstic veins. – **C**: Location map showing the regional distribution of the main travertine outcrops close to the erosional limit of the porous Infra-Cenomanian sandy strata overlying the impermeable Paleozoic substrate

staged along the slope of depressions as a function of the chemical components of the underground waters. The Irdi travertine north of Erfoud (Fig. 8.12) is a typical example of Quaternary carbonate formation, more than 6 m thick and extending over 10 km^2. The Irdi travertine (Boudad et al., 2003) exhibits massive carbonate facies containing clastic grains, interbedded with laminated facies, most often deformed and dislocated, probably due to the regional seismic activity. The carbonate fraction was provided by groundwaters originating from the High Atlas and percolating through the Infra-Cenomanian confined aquifer. Two growth stages have been dated (Th^{230}-U^{234}) around 260 ka (MIS 8) and between 20 and 10 ka, which would correspond to the lower and upper limits of a long inactive period likely due to an excessively arid climate.

8.5 The Arrival of Man

The considerable chronological interest of the discovery of primitive artefacts (Pebble culture, Acheulean and Levalloisian transitional technologies) in the Casablanca quarries where the marine lithostratigraphic scale was defined merits emphasis (Biberson, 1961b; Raynal et al., 2001; Sbihi-Alaoui & Raynal, 2002). The skull, bones and teeth of early men from Casablanca, Rabat and Salé are also of major importance though fragmentary, but complete skulls were found in the J. Irhoud caves from west Jebilet (Ennouchi, 1962). The oldest human remains are associated with an early Acheulean industry of the Early – Middle Pleistocene, ∼ 1 Ma (Debénath et al., 1982). They seem not to differ from those of the man from Sidi Abderrahmane discovered in 1955 in a deposit of the “Littorina cave”, and dated at about 400 ka, similar to the man of Salé. This human species, at first named *Atlanthropus mauritanicus* is now regarded as a *Homo erectus* merging into the true *Homo sapiens*. The man of J. Irhoud, probably living at the end of the Middle Pleistocene (Geraads, 2002) is now also admitted to be a transitional human type, not a Neanderthal, a group deemed to be absent from the Maghreb. The Levalloisian technology, which appeared long ago, continued into the Aterian, referred to *Homo sapiens*, mainly known since 50 ka. The Ibero-Maurusian culture succeeded (12–10 ka) until a Caspian type of Neolithic, which is a culture probably introduced from the Middle East.

Even the Sahara was settled by man since the eve of the Middle Pleistocene (cf. the founding of manufactured pebbles, then Acheulean tools). After this first anthropical optimum, the great desert of Late Pleistocene times favoured great erg development between 100,000 and 30,000 a (Alimen, 1987, in Fabre, 2005). During the Holocene, a humid episode from 9,000 to 6,000 BP favoured new, partly nomadic human settlements documented by Mesolithic and Neolithic artefacts. This is the time when the Tassilis, NE of the Hoggar, became “the best prehistoric museum worldwide” (Cornevin, 1982, in Fabre, 2005).

A post-Paleolithic prehistory of Morocco would require detailed analysis of the climatic evolution through latest Pleistocene-Holocene times. The insufficient and heterogeneous nature of data discourages any modern synthesis. However, work in progress (cf. El Hamouti et al., 1991; Baudin et al., 2007) in close relation with the deciphering of the climate history at the Sahara margins will hopefully succeed in this task.

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Chapter 9
Major Steps in the Geological Discovery of Morocco

Y. Missenard, A. Michard and M. Durand-Delga

9.1 Introduction

This chapter aims at giving recognition to all those, thanks to whom the geological knowledge of Morocco has become as precise as shown in the previous chapters. That knowledge is indeed the achievement of many predecessors, pioneers and patient observers, anonymous cartographers or brilliant synthetic thinkers, Ph.D. student shooting stars or great managers of geological surveys of the past hundred years. The reference lists of the above chapters are limited to the most recent publications. In the present chapter, we wish to go back to the earliest relevant milestones. It would of course be impossible to quote them all, but we want to record the major steps of the research on the substratum of Morocco.

Ph. Morin's bibliography provides the complete list of geological works carried out in Morocco until 1976. Many references prior to the 1930s are forgotten today as the conceptions and means of circulation of scientific works have evolved. Others are quoted over and over although the original documents are not available for consultations because of their scarcity or their disappearance. Thanks to computers, much literature now is at hand. In 2002, the BNF (French National Library) started digitising over 300 books reporting journeys to Africa. Many publications, maps and illustrations are now available at http://gallica.bnf fr/VoyagesEnAfrique/. These documents cover the period between the 18th and 20th century. The French Ministry for National Education and Research also supported an ambitious digitising work

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giving access to many documents at http://www.persee.fr. It is in part thanks to the present easy access to the ancient documents that this brief final chapter was written.

9.2 The Topographic Prerequisite and The First Heartbeats of Geology

Topographic maps supply the *sine qua non* base for any geological survey. However, the creation of these maps, based on physical geography rather than geology, meets with many problems, such as the difficult access to remote and dangerous areas, the limits on the precision of measurements, the lack or incoherence of geographic landmarks, unsteady toponymy etc. In *Le Maroc Physique* (http://gallica.bnf.fr/ark:/12148/bpt6k103614q), L. Gentil delivered in 1912 a summary of the Moroccan cartographic works. We will list here the major steps that led to the topographic maps as we know them today.

The first navigators of historical times who ventured beyond the Mediterranean Sea were the Phoenicians, in the 11th century B.C. They passed the Strait of Gibraltar and travelled along the Moroccan Atlantic coasts, as testified by the word Gadir (Gadis or Gadix), that means “defensive enclosure”, recorded in the Berber name of Agadir (and in Cadiz as well in southern Spain). Unfortunately, there is no written evidence of the Phoenician’s journeys. The Greeks would not venture as far. As described in the *Timaeus* of Plato, they crossed – and named the “Pillars of Hercules” (i.e. the Gibraltar and J. Mousa Rocks) about 9,000 years before Plato’s time (i.e. at about 11 ka B.P.), and reached some islands north of Tangier, that are now drown at shallow depth and that could be the mysterious “Atlantis” (Collina-Girard, 2001). According to L. Gentil, the Greeks also created the term “Atlas” that was a distortion of the Berber word “Adrar”, the hill, or the assimilation of the name of the mythological Giant once condemned to bear the sky on his shoulders. The first maps appear in 300 B.C., with Dicearchus’ world maps. He was the first to use scales for an accurate transfer of the measured points. The Romans between the 2nd and 1st century B.C. made major military expeditions, some of them even crossed the Atlas. Many names of rivers appeared at that time: the Malua, our Moulouya River, and the Guir. Claudius Ptolemy, in 140 A.D. compiled the available coordinates for the some 8,000 astronomically georeferenced places. However the diversity of the sources led to important errors, in particular about the size of the Mediterranean Sea, which was largely overestimated. That mistake would not be corrected until the 17th century.

In the Middle Age, the return of past concepts including the negation of the globe sphericity could lead to think that the quality of the works decreased. However, it was then that navigators drew up the “portulan” maps, which were not scaled, but were of great positioning quality. Among those is the famous 1375 “Catalan map”, probably the work of Cresques, a Jew from the Balearic Islands (L. Gentil, 1912). Although these maps were elaborated by navigators for their own use, they deliver many details concerning the land, for example the location of the Atlas Mountains.

The coasts were drawn with extreme precision and these maps would be the base for future works. The Arab geographers also contributed significantly to enhancing the knowledge of the country. Among them were Maçoudi and Ibn Hankal, (10th century); Albitouni and El Bekri (11th century); Idrissi, the Spanish Moor who summed up the whole knowledge of his time in 1154; Ibn Said in the 13th century and Ibn Batouta who journeyed to Mecca in 1375. Whereas these scholars brought only little precise geographic information, Abdul Hassan – born in Marrakech in the late 12th century – delivered many astronomical positions; he used the portulans and corrected Ptolemy's mistake concerning the size of the Mediterranean Sea five centuries before the Europeans.

On the whole, the 15th and early 16th century maps suffered a regression in comparison to earlier maps. On observing the maps delivered by Linio Sanuto in 1558 (*Fessa tractae*: http://gallica.bnf fr/ark:/12148/btv1b77594004), one can for instance note that mountains are rather randomly displayed. In contrast, in 1570, Ortelius draws a map teeming with details and on which the major rivers are shown, namely the Moulouya, Tensift, Souss, Draa and Ziz rivers. The spring of the Souss and the Draa rivers are separated by a region called Sahara, now known as the Siroua Massif. The name Rif appears for the first time. The map *Fezzae et Marocchi regna Africae celeberrina/describebat Abrah. Ortelius* (http://gallica.bnf fr/ark:/12148/btv1b7759389j), dated from 1636 and inspired from Ortelius, illustrates the huge advances made. P. Van der Aa will base his work on that map producing a slightly modified map in the 18th century (http://gallica.bnf.fr/ark:/12148/btv1b77593906). The current Siroua region is named Serra de Haha on his map, which might be the origin of the word Sahara. A mountain chain running East-West in the south of Marrucos (Marrakech) corresponds to the Atlas, another is shown south of the Souss, probably the Anti-Atlas. The relief of the Rif and Middle Atlas also appear.

In the 17th century, the advances accomplished by Galileo and Cassini permitted a significant step forward and Guillaume Delisle published in 1700 corrected documents that erased Ptolemy's mistake concerning the Mediterranean Sea. Wadis take their final position. The Atlas accurately appears North of Agadir. The Anti-Atlas, which still has no name, is represented for the first time. Bourguignon d'Anville publishes, in 1749, a map where he withdraws all items that are uncertain. Although very basic, that map is much more precise than all the previous ones. However, it will be in the 19th century that mapping is finally completed by real scientific data. In 1811, James Grey Jackson publishes a famous 1:4, 500,000 map entitled *Accurate map of West Barbary including Suse and Tafilalt forming the Dominions of the present Emperor of Marocco* (http://gallica.bnf.fr/ark:/12148/btv1b2300120p/f2.item). Drawing of the coast is accurate, but the Atlas shape is not, especially in the northern regions. These mistakes concerning the relief of Morocco would be corrected later by the Spaniard Baolia, also known as Ali Bey el Abbassi. He determined the position of his route using astronomy and published a 1:1, 950,000 map which represents exactly the overall relief of Morocco.

The precision of maps increases along with scientific works. In his work *Die Erdkunde im Verhältnis zur Natur und zur Geschichte des Menschens : oder allgemeine vergleichende Geographie, als sichere Grundlage des Studiums und Unterrichts in*

Fig. 9.1 Excerpt of the geographic map of the "Sultanat Marocco" by J. G. Radefeld (1844). Insert: overview of the original map, which also shows a plan of the city of "Marocco" (Marrakech). Courtesy of David Rumsey, http://www.davidrumsey.com

Physikalischen und historischen Wissenschaften (2nd edition, 1822–1859), Ritter distinguishes the Lesser Atlas (Rif), the High Atlas, and a plateau that corresponds to the Middle Atlas. E. Renou publishes a 1:200,000 map with text: *Description géographique de l'empire du Maroc* (http://gallica.bnf.fr/ark:/12148/bpt6k1037743). In his 500 pages work, the author delivers an explanation for his choices as to the incorporation or omission of most ancient routes as well as a comparison with his own drawings. However, the Anti-Atlas completely disappears from his work, and the Middle Atlas orientation is erroneous. In 1848, Beaudoin proposes a more precise 1:1,500,000 map. His drawings of the Rif region will be a reference until as far as the beginning of the 20th century. The map of the German captain Radefeld, published in 1844 is the best document to assess the state of the knowledge in the mid 19th century. We believe this map to be the most realistic of its time. It is shown here (Fig. 9.1) from the website http://www.davidrumsey.com/maps695.htm.

Thus, shortly before the half of the 19th century, trustworthy topographic documents are at hand. Inspired by Elie de Beaumont's theory, geologists immediately rush to consult them. The interpretation of the High Atlas as "one of the three or four great islands of Paleozoic Africa" is suggested by the first *Carte géologique de tout le globe* which was presented to the Geological Society of France by Ami Boué in 1844 (Durand-Delga, 1997). This map links up Morocco and Southern Spain (Sierra Nevada), because "a similar geological nature often characterizes belts having parallel orientations [...]".

In 1845, Henri Coquand (1811–1881) believed to be "the first geologist setting foot on the Moroccan land". One also owes him the "*Description géologique de la partie septentrionale de l'Empire du Maroc*" published in 1847 in which Coquand sets the milestones of the knowledge concerning the "Petit Atlas [i.e. the Rif]... which seems to extend continuously to Tunis." Unfortunately, Coquand is blinded by Elie Beaumont's concepts on the geometric structure of mountain belts and thinks that uplifts from different eras generated the N-S trending "ancient mountains" of Northern Rif and those, E-W trending, of Eastern Rif, which he trusts to date as Tertiary...

9.3 The Expansion of Moroccan Geology During the Age of Expeditions

The German Rohlfs was the first to give a description of the Moroccan soil based on true scientific expeditions. He published two volumes of his journeys (1868 and 1873) and came back with a map and observations on the nature of the soil (http://gallica.bnf fr/ark:/12148/bpt6k49687h, http://gallica.bnf fr/ark:/12148/bpt6k 1042709). However, the geologic knowledge of Morocco began with the expedition of the British botanist J. Hooker in 1871.

The "Journal of a tour in Morocco and the great Atlas" was published in 1878 by J. Hooker, J. Ball and G. Maw (http://gallica.bnf.fr/ark:/12148/bpt6k98288j). This imposing volume presents the different routes followed by the explorers. The geological observations are gathered in an "Appendix" written by George Maw, the geologist of the expedition. Many pictures illustrate the text, including a geological section of the Atlas, the first ever to be published, as well as an amazingly precise map of southern Morocco produced by J. Ball. The manuscript contains many observations concerning local customs and society organisation and, of course, abundant botanical observations.

G. Maw studied particularly the Tangier and Safi regions, but the plain of "Marocco" (Marrakech) and the High Atlas gained a major place in his work, being presented in a cross-section (Fig. 9.2) which is the first attempt of this type in Morocco. Maw notes the presence of a coloured limestone covering the rocks everywhere in the plain; this "singular deposit" corresponds to the now well known calcretes. North of the plain, the author describes 2000–3000 feet high hills (the Jebilet). He estimates the High Atlas elevation at 12–13,000 feet above sea level (3600–3900 m), which is barely underestimated as far as the J. Toubkal (4167 m) is concerned. The High Atlas approaches are described as a series of hills, 2000–3000 feet below the chain. The author distinguishes several lithological sets in the hills: surface deposits (conglomerates, moraines), red sandstones and limestones assigned to the Cretaceous (mostly Triassic, in fact), "eruptive basalt" (Triassic-Early Jurassic), "grey clays" (the Paleozoic slates), metamorphic rocks and "porphyrites". The author describes each site with frequent comparisons to English sites. Although some of his deductions proved wrong, for example the occurrence of true moraine deposits, others, such as the discordance between the red sandstones and "grey clays" or "prophyrites" (corresponding to the Paleozoic/Mesozoic unconformity), were later confirmed. The inclination of the "grey clays" is described as perpendicular to the general orientation of the Atlas, which is in accordance with our knowledge concerning the structure of the Hercynian belt there. Finally, G. Maw presents an essay on the geological history of the Atlas, facing serious difficulties due to lack of chronological elements. Overall, G. Maw deserves credit for being the first to have suggested a plausible stratigraphy in which we may recognize (under different names) the Quaternary, Tertiary, Mesozoic, Paleozoic and Precambrian systems. We also owe the name of "Anti-Atlas", inspired from the Anti-Lebanon mountain range, to the members of the J. Hooker's expedition.

In 1879, during his journey to *Timbouctou – voyage au Maroc, au Sahara et au Soudan* (http://gallica.bnf fr/ark:/12148/bpt6k1048451, http://gallica.bnf.fr/ark:/12148/bpt6k104232p for the French translation), Dr. Oskar Lenz crossed the High Atlas through the Bibawan pass, the current path through the Argana Corridor. Unfortunately, he only brought back little geological references, reporting only the presence of azoic red sandstones on the Atlas and the "Paleozoic plateaus" (Carboniferous and Devonian) that build up the northern part of the Western Sahara. Nonetheless, his work is instructive in its description of the difficulties linked to such expedition. The author foresees that "a few decades later things may have evolved so that tourists may take tours through the Atlas as they do for the Himalayans, the Caucasus, etc. People would then smile when told that this path was considered as problematic, although it indeed is the case today and it would remain so for a while".

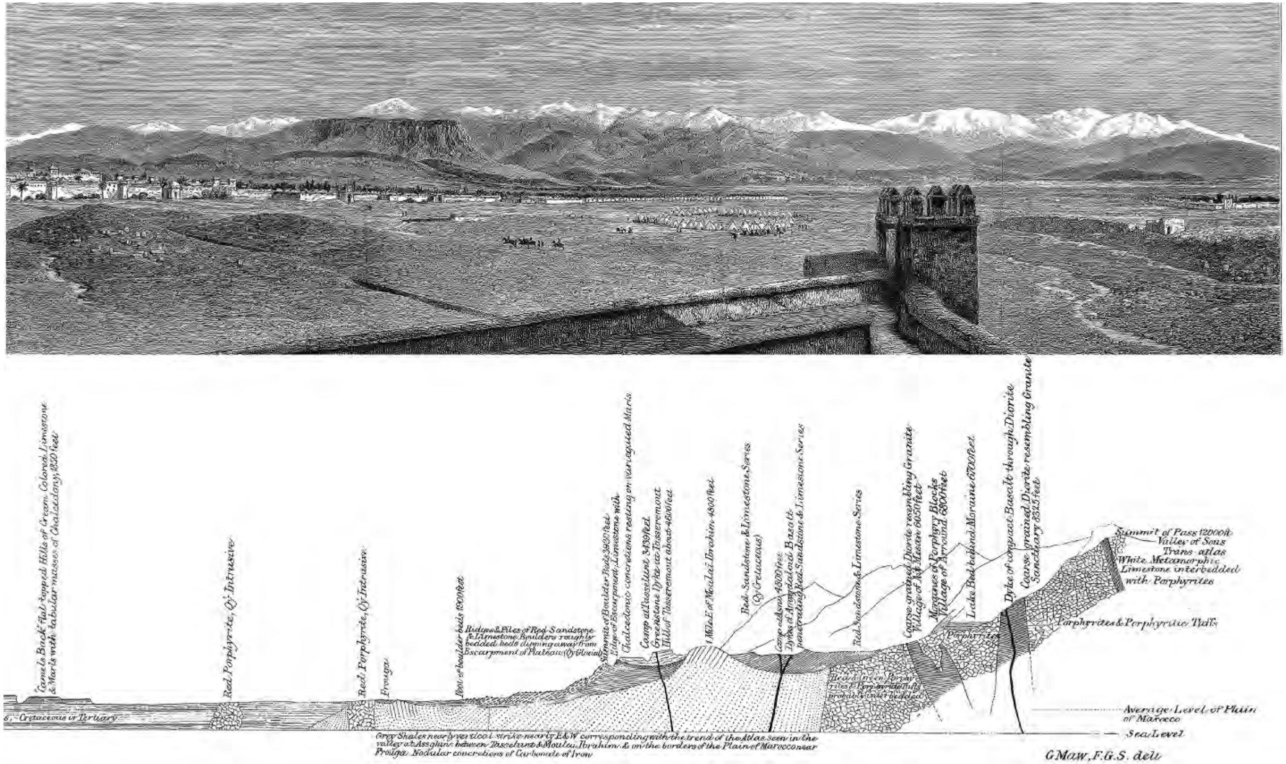

Fig. 9.2 The first geological cross-section of the High Atlas, drawn by G. Maw, and published in the "Journal of a tour in Morocco and the great Atlas" by J. Hooker, J. Ball and G. Maw (1878). The naturalists followed the most direct road toward the south, via Moulay Ibrahim and Asni, then they climbed up to the crest line ("Summit of Pass 12000 ft"). The main geological units have been correctly recognized by G. Maw: the "Grey Shales", to the north, correspond to the Early Carboniferous slates (Moulay Ibrahim); the "Porphyrites and Diorites" of the axis of the mountain range correspond to the Precambrian basement; and the "Red Sandstones and Limestones series", overlying unconformably both the Paleozoic and Precambrian units, correspond mainly to the Triassic deposits. However, the Late Cretaceous-Eocene age of the "escarpment limestones" close to "Tasselunt" (Tahanaout) and south of Asni was not recognized

Ten years later, two expeditions brought back more complete geological observations. The first expedition is that by the viscount Charles-Eugène de Foucauld, and the second, that of Joseph Thomson. De Foucauld's book, *Reconnaissance du Maroc* (http://gallica.bnf fr/ark:/12148/bpt6k200835s), describes his journey in 1883/1884 during which de Foucauld wandered through nearly all Morocco, disguised as a rabbi. He comes back with impressive quantities of observations concerning the flora and fauna, architecture, meteorology, and society. He distinguishes four mountain systems on a geographical point of view, namely the Rif, Middle Atlas, High Atlas, and Anti-Atlas. He highlights the curvature of the Rif belt and its independence relative to the Atlas, but he adds another "Mesozoic chain" wrongly south of the Anti-Atlas, which corresponds, in fact, to the J. Bani Hercynian folds.

Joseph Thomson wandered through Southern Morocco and the Atlas a few years after Charles de Foucauld. His narration, published in 1889 and entitled *Travels in the Atlas and Southern Morocco – a narrative of exploration* (http://gallica.bnf.fr/ark:/12148/bpt6k105356g), is teeming with varied information, especially with geological data. The Central High Atlas close to Demnate is described for the first time as a folded mountain belt consisting of red sandstone and limestone series. The stratigraphic units described by G. Maw are generally kept in use, and Thomson's main contribution is certainly the geological map which concludes his work. This map, of course still simplistic, is the first geological map of Morocco ever delivered.

The last decade of the 19th century marks the beginning of a new era of scientific activity. This is also the moment when the Académie des Sciences of Paris organises its "Mission d'Andalousie", which provided the scientific community with abundant new data on the Betic Cordillera (Bonnin and et al., 2002). Scientific expeditions are increasingly frequent. Travellers like Theobald Fischer, Dr. Weisgerber or R. de Segonzac bring back varied observations, including on geology. The geographer P. Schnell publishes his *Marrokanisches Atlas* in 1892. In that work, he describes the High Atlas as a "stranger coming from the north and part of the great Mediterranean fold zone". Topographic mapping keeps improving, and M.R. de Flotte de Roquevaire publishes in 1897 and 1904 high quality maps at scale 1:1,000,000.

9.4 The Era of Systematic Exploration, Mining Research, and First Academic Synthesis

It was not until 1903–1905 that geologists such as A. Brives, L. Gentil, P. Lemoine, and E.F. Gautier carried out expeditions in Morocco exclusively on geological purpose. At that time, the fundamentals of geology (i.e. the succession of orogenies, thrust nappe concept, structure of the Earth, etc.) were known and taught, especially thanks to E. Suess' monumental synthesis *Das Anlitz der Erde*, 1883–1909, translated in French (*La Face de la Terre*) by Emmanuel de Margerie in 1897. Abel Brives and Louis Gentil were coming from Algiers. It is notable that the systematic exploration of the Algerian territories had been initiated as soon as 1842 under the aegis of the Mining Survey created there by the French authorities (Durand-Delga, 2005). However, the earliest Saharan expeditions, by G.B.M. Flamand, then René Chudeau, lagged until the turn of the century, as reported by Ph. Taquet (2007).

A. Brives visited the "Bled Maghzen" with the Sultan's officials. L. Gentil, who was able to speak Arabic, ventured alone in "Bled Siba", disguised in a countryman, almost the way Ch. De Foucauld had chosen before him. In 1905, A. Brives publishes his *Contribution à l'étude géologique de l'Atlas marocain* in the Bulletin de la Société géologique de France. L. Gentil begins his expeditions in 1904 and the controversies concerning the interpretation of the Atlas arise right away! The *Annales de Géographie* regularly echoes the research progress. For instance one may find in vol. 14, 1905, # 75, p. 285, available at http://www.persee fr, a report by M. Zimmermann concerning a joint mission of De Segonzac, Gentil and De Flotte de Roquevaire. In 1908, A. Brives mentions phosphatic beds probably "Suessonian" (Early Eocene) in age, SW of Marrakech, and in 1909, he suggests the occurrence of such beds north of the Oum-er-Rbia River (cited by Boujo & Salvan, 1986). According to the latter authors, it was only in 1917 that a military engineer, former Director of the phosphate mines of Gafsa (Tunisia), recognized a phosphate deposit in the sand extracted, for building purpose, from quarries near Oued Zem town! Afterward, industrial trenches have been opened in the whole area, the Office Chérifien des Phosphates was founded in 1921, and the railway will reach Khouribga in 1922. During the same period, the exploration of the Prerif regions for oil, under way since 1910, was strongly boosted by a first success in the Prerif Ridges in 1919 (Jabour et al., 2004).

The state of the art synthesis will be Louis Gentil's fulfilment. After a first study that concentrated on northern Morocco (1906), L. Gentil published *Le Maroc Physique* (1912; op. cit. in Sect. 9.2 with web ref.), based on the author's observations and on those of his predecessors. This work is a real turning point. Additionally, L. Gentil published in 1920, "by order of M. le Maréchal Lyautey", a remarkable *Carte géologique provisoire du Maroc* at scale 1:1, 500,000.

In 1912, *Le Maroc Physique* presents the structural domains of Morocco quite clearly (Fig. 9.3). The precision of the text is outstanding, but its illustration is very poor (a drawback characteristic of that time). Figure 9.4 is a map of the western High Atlas published "after Louis Gentil" by E. Suess in *La face de la Terre*, vol. 3, 1918. This French edition of the famous *Das Antlitz der Erde* (Suess, 1908) contains numerous additional illustrations on Emmanuel de Margerie initiative. The High Atlas is regarded by L. Gentil (and by Suess after him) as a "vast anticlinal dome, whose Jurassic cover displays upright or recumbent folds over large areas" [translated by the Authors]. At the bottom of the Jurassic beds, he describes a series of "Permo-Triassic red beds overlying unconformably the strongly folded Lower Carboniferous". Folding of the Carboniferous beds is assigned to a belt formed "between the end of Dinantian and the deposition of the overlying red beds, [...] contemporaneous with the European Hercynian belt, [and directed NNE-SSW] similar to the Variscan structures". Afterward, the Paleozoic belt has been "the prey of continental erosion, which may have begun since the end of Carboniferous, and continued during the Permian and part of the Triassic. [...] large fractures occurred, resulting in

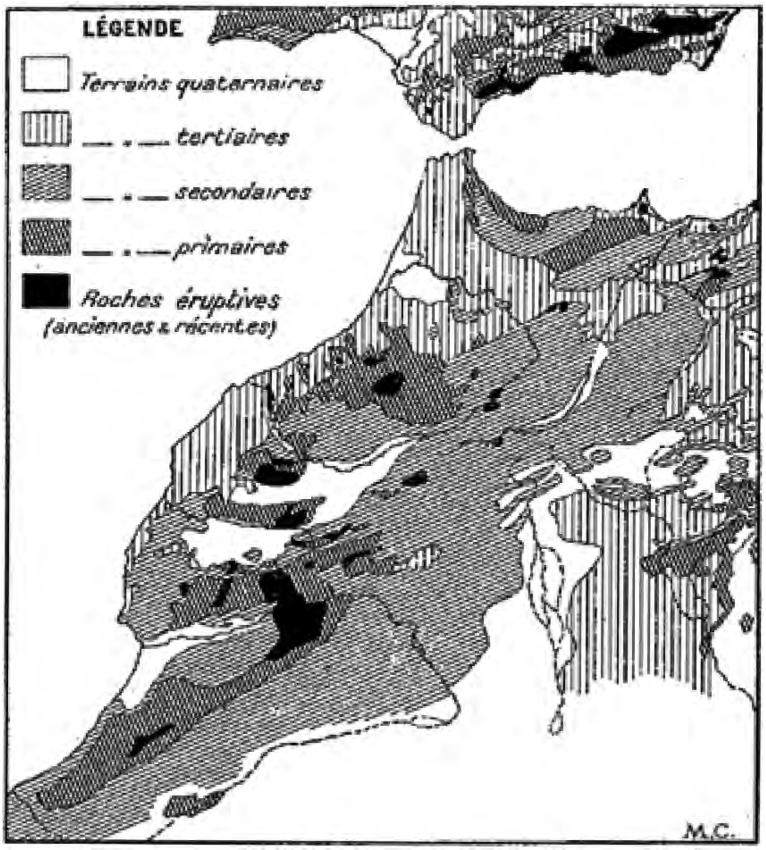

Fig. 9.3 The first geological map of Morocco, by Louis Gentil (1912). This map already shows the main structural domains of the country (compare with Fig. 1.20A, Chap. 1), but still includes important errors. In particular, the Anti-Atlas is indicated with "Primary terranes" in its axis and "Secondary terranes" in its envelope, instead of Precambrian and Paleozoic, respectively. Likewise, the occurrence of "Tertiary terranes" in the western Meseta Plateau is not recognized yet

the subsidence of multiple blocks. It clearly seems that the most important of these fractures took place on the borders of the present belt". This refers to our rifting episode, and L. Gentil also notes that it was associated with "tremendous volcanic eruptions". The lack of Mesozoic cover rocks in the axis of the Marrakech Atlas is discussed. The author rejects the hypothesis of a Jurassic island as coastal deposits are lacking around the massif. He also discards the hypothesis of denudation by nappe tectonics, as "the belt only offers Jura-type folds which deeply contrast with the more violent shape of Alpine folds". According to L. Gentil, the Cenomanian transgression flooded most of the incipient High Atlas belt. An important hiatus occurred before the end of the Eocene, "not unrelated to the Tertiary movements that probably began before the Eogene period [...]. Anticlines and synclines formed

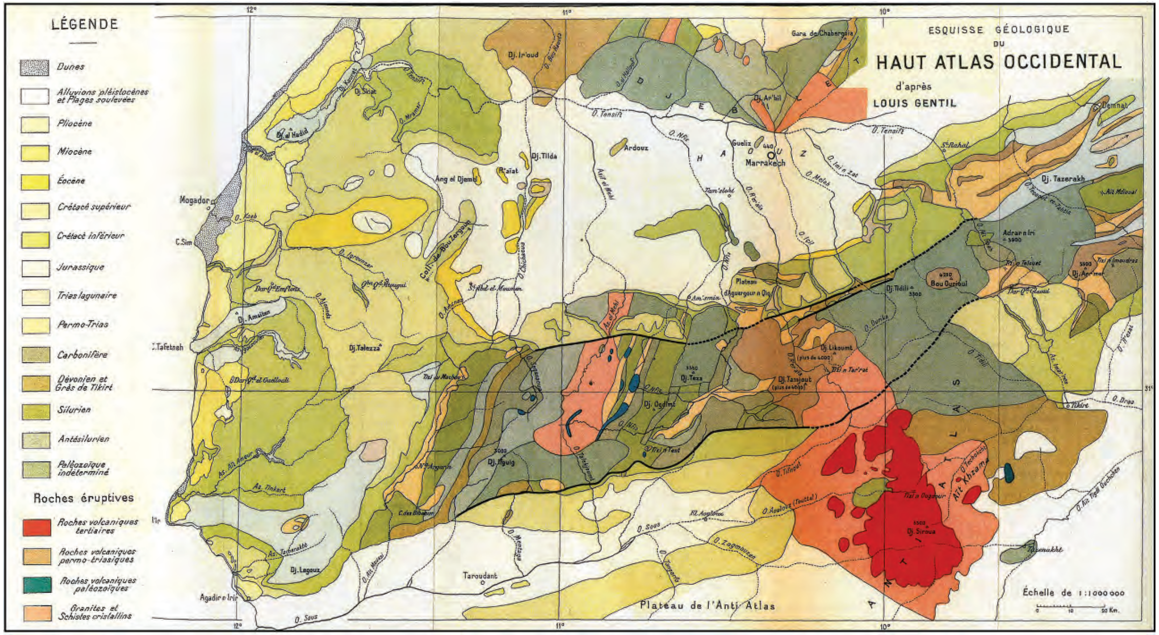

Fig. 9.4 The first geological map of the Western High Atlas, published by E. Suess "after Louis Gentil" in *La Face de la Terre*, the French edition of *Das Antlitz der Erde* translated and edited by Emmanuel de Margerie in 1918, i.e. ten years after the original work of Suess (1908). The original map, scale 1:1,000,000, here reduced by about half, was specially prepared for the French edition, in order to account for the important geological achievements obtained since the previous Gentil's synthesis (1912). The obliquity of the tectonic grain of the Atlas range respective to its boundary faults is clearly shown. The Triassic Argana Corridor is represented except at its southern extremity toward the Souss Basin, as at that time, the road from Argana to Taroudant crossed the crest line at the Bibaoun Pass. All the post-Cambrian geological epochs are identified (except the Oligocene, but this did not changed up to now!), but their geographical extension is still uncertain (e.g. the "Lower Cretaceous" shown in the O. Zat Valley corresponds in fact to the Carnian Oukaimeden Sandstones). The Anti-Atlas domain is still poorly known (the allegedly Devonian "Grès de Tikirt" are in fact Cambrian in age, and the "Undetermined Paleozoic" of Tazenakht corresponds to the Zenaga Precambrian inlier)

then, and gave to the belt its definitive direction and orographic outline”. The Atlas structure includes “stepped fractures […], forming fault assemblages broadly parallel to the Atlas axis”. The large Siroua volcano is also described.

Concerning the other Moroccan Mountains, geologic descriptions are more succinct in L. Gentil’s book. The Anti-Atlas carbonate series are erroneously ascribed to the Cretaceous. In line with de Foucauld, the author considers the J. Bani as a Tertiary fold belt formed by the “cover slide, maybe favoured by a plastic layer included in Triassic gypsum beds”. This interpretation, erroneous from the point of view of stratigraphy, is correct for that of folding mechanisms. The Middle Atlas is described as “a succession of more or less parallel crests, whose orientation swings between NE-SW and ENE-WSW trends”; the folds “plunge beneath the Miocene deposits of the Middle Moulouya valley and Taza area”. Last but not least, concerning the Rif, L. Gentil confirms E. Suess interpretation of the Gibraltar Arc, and postulates the occurrence of “imbricated folds or even nappes, thrust toward the South Rifian Corridor depression”. L. Gentil is the promoter of the innovative nappe concept in the Rif belt. He recognizes the thrusting of the Haouz range (our Dorsale calcaire), that of J. Mousa, compared with the Gibraltar Rock, and that of the “Nummulitic flyschs” (in fact, the Mesozoic-Cenozoic Maghrebian Flyschs) on top of the Tangier “Cretaceous domes […], which may be autochthonous”. The South Rifian Corridor is regarded as the gateway between the Atlantic Ocean and Mediterranean during the Miocene. The strait of Gibraltar would have opened at the very beginning of the Pliocene (as already assumed by J. Bergeron).

An interesting section of the *Maroc physique* deals with epirogenic movements. The author suggests that “positive epirogenic movements” are recorded in the Moroccan Meseta and South Rifian Corridor, where they resulted in the closure of the seaway by the end of the Miocene. The regional uplift is estimated at about 1000 m. To our knowledge, the uplift described by L. Gentil disappeared from the literature until recent years, when the occurrence of thinned lithosphere beneath the Atlas domain was recognized. Remarkably, the resulting uplift appears to be close to 1000 m in the recent geophysical models (see Chap. 1).

9.5 From the Age of Monographs to the Present-Day Thematic Researches

L. Gentil’s work opened an age of detail regional studies in the varied domains of Morocco. The text that J. Savornin published in 1924 (www.persee fr, Annales de Géographie, Numéro 183) allows us to measure the achievements realized 12 years after the earliest modern works by A. Brives and L. Gentil. The varied series are more securely dated now, based on new fossil collections. The “Institut scientifique chérifien” (founded in 1920), and the “Service géologique du Maroc” (founded in 1921) will now play a major role in the development of Moroccan geology (see the last catalogue of the Service Géologique, Anonymous, 2004).

A monograph of the northwestern Moroccan Meseta is presented by Georges Lecointre (Lyautey's nephew, according to Willefert, 1997) as a "Doctorat ès-Sciences" thesis at Paris (Sorbonne) in 1926, resulting in a much better understanding of the Hercynian evolution there. Further south, J. Barthoux undertakes the study of the Rehamna and Jebilet (1924). Dr. P. Russo (senior Moroccan delegate at the 13th International Congress of Geology, Brussels 1922) maintains the tradition of explorations, as his itineraries concern the entire countries north of the South-Atlas Fault, which he describes as "the limit between the Mediterranean and Saharan terrains". P. Despujols, the head of the new Geological Survey of Morocco, publishes in 1933 the first history of geological researches in Morocco, but the number of new works increases rapidly until 1940. Most of these works are published in the *Notes et Mémoires du Service géologique du Maroc* (initiated in 1928; cf. Morin, 1965, and www.mem.gov.ma), the *Bulletin de la Société géologique de France*, or the *Comptes Rendus de l'Académie des Sciences de Paris*, and some others in the *Bulletin de l'Institut Scientifique de Rabat*. The overwhelming number of works would make tedious an exhaustive presentation in this chapter, and thus, we mention only the most important syntheses hereafter.

Before World War II, Paul Fallot investigated the geology of the Internal Rif and Betics (*Essai sur la géologie du Rif*, 1937), and Henri Termier that of the Central Massif and neighbouring Middle Atlas. With its 1600 pages and countless cross-sections, maps, plates and engravings (Fig. 9.6), Termier's monumental memoir, published in 1936, is a milestone work for the knowledge of the Paleozoic terranes of Morocco (it is also an extreme example of the gigantism which characterized the former French "Doctorat d'Etat" thesis!). We have also to mention the work of Gabriel Lucas (1942) in the Oran area and Eastern Morocco, where Brichant has discovered the Jerada coal deposits as early as 1928; the work of Jacques Bourcart (1927, 1931) in the western Anti-Atlas where he recognizes the Cambrian Archaeocyath limestones and the Acadian; the discoveries of Louis Neltner (1929, 1938), who recognized the unconformity between "Archaean" and "Algonkian" terrains in the Anti-Atlas (i.e. the Paleoproterozoic and Neoproterozoic groups, respectively); that of Edouard Roch and Léon Moret (1931) in the High Atlas, etc. In the Western

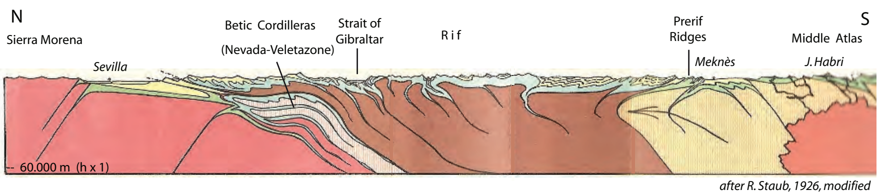

Fig. 9.5 The first cross-section of Morocco at lithosphere scale (northern part), as published by Rudolf Staub (1926). The Betic nappes are considered as rooted in the African margin, and the Sierra Nevada window is assumed to be an equivalent of the Tauern window, by extrapolation of the Eastern Alps structure (cf. Austrides and Pennides, respectively). Note the exaggerated thickness of the continental crust, and the enormous magmatic chamber related to the Middle Atlas volcanism (pink colour on the right within the Meseta-Atlas crust)

Fig. 9.6 An example of the engravings included in H. Termier's masterpiece (1936). They are the work of the captain of the military detachment which accompanied Termier in the remote and barely secured mountains he had to study. The Jebel Baddouz is located in the Khouribga-Oulmes anticlinorium of the western Central Massif (see Fig. 3.17 for location). The original illustration includes the following legend, written above the landscape: "1. Schistes à Trinucleus ornatus; 2: Quartzites (masse principale; 3: Quartzites à Homalonotus Brongniarti" (sic). These series range from the Upper Llandeilo-Middle Caradoc (1) to the Upper Caradoc (3) (Razin et al., 2001)

Rif, Fernand Daguin (1927) defines the Prerif nappe, and Jean Marçais (1936, 1938) describes the "Rifian nappe" above the Miocene formations of the Central Rif. A new geological map 1/1,500,000 is published in 1936 by Boris Yovanovitch, a petroleum geologist and hydrogeologist, but this map is soon overshadowed by those of the Geological Survey, at scale 1:500,000.

These new scientific achievements, disseminated within the international geological community, nourished several attempts of interpretation of the Moroccan fold belts, among which it is worthy citing those by Emile Argand (1924), Rudolf Staub

(1926), and Hans Stille (1927). These syntheses were fragile and controversial, as hypotheses have often to make up for insufficient data. This can be exemplified with the controversies concerning the nature, either real or fallacious, of the Gibraltar Arc (Durand-Delga, 2006). In 1926, R. Staub produced a cross-section of Morocco from which the Rif part is presented here (Fig. 9.5). This first lithosphere scale transect (to be compared with the recent Transmed I Profile by Frizon et al., 2004; cf. Chap. 1, Fig. 1.20B) is based on the assumption that the Rif structures do not bend to connect with the Betic, but continues to the west into the Atlantic – which is partially correct for the External Zones, but completely incorrect for the Internal Alboran Domain.

World War II obviously slowed down geological research. However, during this period, H. Termier reorganized the Geological Survey, now separated from the Mining Administration. Two new synopsis of Moroccan geology were published soon after war, a short *Schéma structural du Maroc* by de Georges Choubert & Jean Marçais (1952), and the voluminous *Histoire stratigraphique du Maroc* by E. Roch (1950) which was intended to serve as an explanatory notice for the geological maps 1:500,000 established since 1950 under the direction of G. Choubert. Indeed, a fever of production seized the Geological Survey as the 19th International Geological Congress, to be held in Algiers, 1952, approached (Durand-Delga, 2005). The congress was an opportunity to publish several *Monographies régionales* and a set of field trip guide books which expose the state of geological knowledge concerning every region of Morocco.

Shortly after the congress, the 1:500,000 maps are achieved and published sheet after sheet. A new mapping program is launched, including a set of 1:200,000 maps in Southern Morocco, and a number of 1:50,000 maps in the Rif belt, edited by G. Suter. The Geological Survey is fully active, under the impulse of G. Choubert and A. Faure-Muret. Modern geological methods (e.g. geochemistry, isotopic dating, structural analysis, etc.) are available, but the plate tectonic theory is not. In 1960–62, a collective book, the *Livre à la mémoire du Professeur Paul Fallot*, offers a synthetic presentation of the geology of the Atlas and Rif domains. This contribution is particularly innovative in the latter domain, as field studies in the Rif have greatly benefited of the recent reunification of the Moroccan kingdom, now independent (1956). Another synopsis of Moroccan geology appears in 1976, prepared by A. Michard with the friendly help of Jacques Destombes, Renaud du Dresnay, Henri Hollard, Gabriel Suter, and Solange Willefert, from the Geological Survey. Entitled *Eléments de Géologie Marocaine*, this book summarizes the geological knowledge concerning the country in its 1975 extension, i.e. without the Southern Provinces. Since that time, the description of the geological evolution of Morocco has considerably progressed, based on the new concepts and methods of earth sciences (plate tectonics, trace geochemistry, robust isotopic datings, seismic profiling, seismic tomography, sequential stratigraphy, etc.). A new geologic map at 1:1,000,000 has been published, and an ambitious National Plan of Geological Mapping at 1:50,000 has been launched. Seismic profiling and boreholes allowed petroleum geologists to gain better views of the Moroccan basins, onshore and offshore. A lot of works have been published by two generations of Moroccan geologists, graduated in foreign universities, then in the new Moroccan universities. This is why it was compulsory to undertake a new synopsis of the geology of that beautiful country.

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Chapter 10
Conclusion: Continental Evolution in Western Maghreb

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In this conclusive chapter, our aim is to summarize the lessons which can be drawn from the preceding chapters with regard to continental evolution. Obviously, only part of the vast geophysical/geochemical/historical problem of continental evolution can be enlightened through the Moroccan case study, as Morocco extends only on a very restricted, marginal part of Africa. Moreover, Morocco is located in a particular area of the continent, at its northwest "corner" (Fig. 10.1). This restricted region of the large African continent is bounded by the Atlantic passive margin to the west, as old as Early Jurassic, and by the Oligocene-Neogene West Mediterranean Sea to the north, a young thinned continental crust/oceanic basin born within the Alpine collisional domain. The geological evolution of the Moroccan lithosphere is deeply marked by this situation.

Second, this chapter aims at evoking the most important avenues of research, which are currently opened or should be traced in the near future in order to get a best and fruitful knowledge of the Moroccan subsurface geology and deep lithosphere structure. In this field, the interpretation of topography still occupies a significant place.

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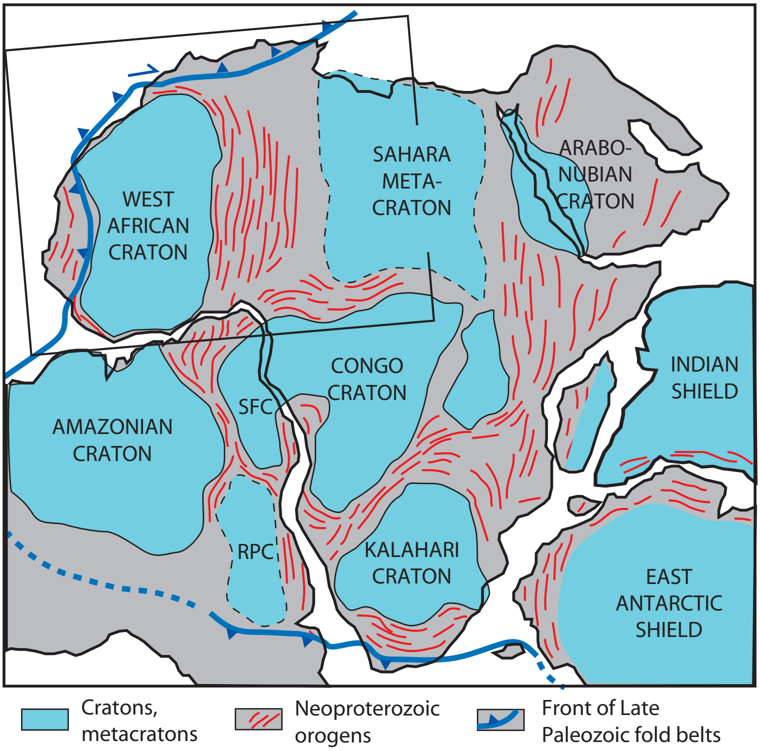

Fig. 10.1 Sketch map of the African Continent in the framework of the supercontinent Gondwana, after Meert & Lieberman (2007), modified. The Neoproterozoic orogens criss-cross the supercontinent, being associated with its progressive amalgamation. In Northwest Africa, the "South Variscan Front" indicates the front of the Variscan displaced terranes (Mauritanide thrust nappes, Meseta Domain). This front runs west and north of the Anti-Atlas and Ougarta foreland fold belts, respectively (not shown; see next figure). The Alpine belts of North Africa (Atlas and Maghrebides) are not differentiated north of the South Atlas Fault. Likewise, the Andes Orogen of South America is not shown south of the South Gondwana Paleozoic Orogen

10.1 Continental Building/Break Up Alternations

In line with L. Gentil (1912), G. Choubert and J. Marçais (1952) outlined the geological structure of Morocco as consisting of a succession of east-west fold belts bordering the Saharan platform, and being superimposed through time, the youngest being located progressively further to the north. Nowadays, this ancient paradigm of "continental growth" has to be substituted by a more complex geodynamic conception, which recognizes alternating phases of continental building and break-up controlled by plate tectonics. This is well illustrated in Morocco since about 2 Ga (Fig. 10.2).

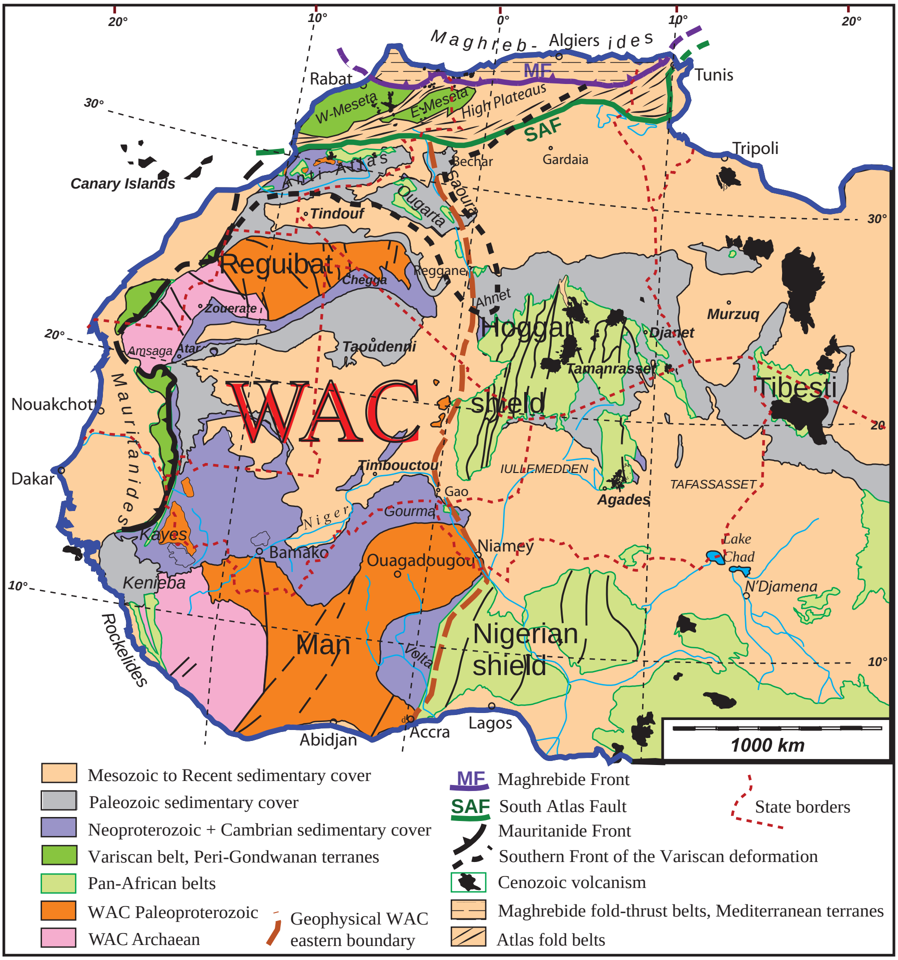

Fig. 10.2 Structural map of NW Africa. Location: see Fig. 10.1. After Fabre (2005), Liégeois et al. (2005), Ennih & Liégeois (2008). State borders of the Maghreb and Saharan countries are indicative. The Cretaceous-Cenozoic intraplate deformations (e.g. Reggane area; Smith et al., 2006) are not shown for lack of regional data

10.1.1 From 3 Ga to 800 Ma (cf. Chap. 1)

The southernmost lowlands of Morocco correspond to the Archean-Paleoproterozoic lithosphere of the West African Craton (WAC), which extends widely in the neighbouring Saharan regions of Mauritania and Algeria (Fig. 10.2), and to the south up to the Ivory Coast (a chip of the WAC forms the Sâo Luis Craton of North Brazil). The WAC subsurface is not homogeneous: its northern part displays a large basement rise, the Reguibat Shield, and two synformal basins, namely the Tindouf-Reggane and Taoudenni Basins. South of the latter basin, the southern counterpart of the Reguibat Shield, namely the Man Shield, takes place. These basins are filled up with virtually undeformed Neoproterozoic to Neogene deposits, as it is typical for a cratonic area.

The WAC lithosphere itself is also heterogeneous. The western half of the Reguibat Shield consists of 3 Ga-old Mesoarchean terranes, whereas its eastern half is made up of 2.1–2.0 Ga-old Paleoproterozoic terranes. This involved the collage of two wandering continents and building of a larger continent through a collisional tectonic phase referred to as the Eburnian Orogeny. The corresponding NW-trending "suture zone" is seemingly devoid of ophiolite.

The WAC behaviour during the Mesoproterozoic remains mysterious (simply drifting and weathered during 1 Ga?) as no geological event at all is recorded during that period. By the early Neoproterozoic, the WAC was probably included in the Rodinia supercontinent, whose rifting may have triggered the subsidence observed in the Taoudenni Basin and southwest Anti-Atlas. The provenance of the clastic input in the WAC shallow water depocenters has not been studied until now (uplifted parts of the WAC or neighbouring Mesoproterozoic belts?). The break-up of Rodinia resulted in the insulation of the WAC, which was surrounded by oceanic domains until the Pan-African convergence took place.

10.1.2 From 760 to 560 Ma (Cf. Chap. 2)

The 760–560 Ma-old Pan-African Belt has surrounded continuously the WAC to the north (Anti-Atlas and Ougarta inliers), east (Western Hoggar or Pharusides, Iforas, Gourma), southeast (Dahomeyides), southwest (Rokelides) and northwest, as Pan-African terranes with both felsic and ophiolitic lithologies are included in the Mauritanide nappes finally emplaced during the Variscan orogeny (see below). This is consistent with a major collisional event responsible for the agglutination of the wandering continents into a single Pannotia supercontinent (Gondwana + Laurentia + Baltica + Siberia). For the first time in the geological record, the plate tectonics with subduction and collision is well illustrated by obducted ophiolites and oceanic arc (Siroua and Bou Azzer inliers in the Anti-Atlas), and HP-LT coesite-bearing metamorphism in the eastern suture zone (Mali). The putative "northern continent" involved in the collision along the Anti-Atlas segment is unknown. It could be hidden beneath the Atlas and Meseta system but could also be included in some of the Peri-Gondwanan terranes now dispersed in Europe and America.

This phase of continental building is also characterized by a huge magmatism, intrusive and effusive, firstly calc-alkaline, then potassic and finally alkaline. Part of the magmas is anatectic, resulting in the recycling of older continental crust, but another part is mantle-derived, thus adding fresh material to the continent. Besides of the nappe tectonics and of the folding of the detached foreland cover (e.g. Bou Azzer, Kerdous), the tectonic regime coeval with the main magmatic accretion was dominated by wrench faulting, either transpressive or transtensive, in a large metacratonic zone fringing the WAC.

10.1.3 From 550 to 270 Ma (cf. Chap. 3)

Pannotia was no more that a transient, unstable continental configuration. As far as NW-Gondwana is concerned, extensional tectonics affected the Pan-African Belt during the late Ediacaran, then rifting increased during the Cambrian and continued during the Ordovician, basically along the western and northern deformed borders of the WAC. This resulted in the formation of thinned crust/oceanic domains at the emplacement of the present-day Mauritanide Belt and Meseta Domain. However, the WAC itself subsided contemporaneously, and extensional faulting operated until the Devonian-Early Carboniferous. By the end of this time, 5–10 km of dominantly clastic deposits had been accumulated over the northern WAC and former Pan-African Belt. The source of this huge clastic sedimentation may be found in the large Tuareg (Hoggar) Shield, still affected by magmatism and uplift during the Cambrian.

The driving force responsible for this continuous or recurrent Paleozoic extension is a matter of debate: plate divergence or back-arc extension associated with subduction? In every case, this resulted in the dismembering of the metacratonic margin of the WAC into terranes which drifted toward the NW (present coordinates): Avalonia and the Hun/Armorica terranes. Was the Meseta Domain part of the latter, or did it remain at short distance from Africa? This is also controversial.

A new phase of continental building corresponds to the Variscan Orogeny (350–270 Ma). More or less exotic terranes were accreted then to NW Africa: the Mauritanide Belt on the west border, and the Meseta Variscan Domain on the north border. The latter domain includes a northward, Caledonian-Sardic terrane (Sehoul Block) characterized by Ordovician felsic magmatism and metamorphism, and brought against the undeformed Central Meseta during the Late Silurian. Late Devonian-Tournaisian shortening events are recorded within the Meseta Domain. However, the final collage of both the Mauritanide and Meseta terranes against Africa occurred during the Late Carboniferous-Early Permian, in the framework of the building of the Pangea supercontinent. Thus, the small part of the Variscan Belt which remained in Africa after the Pangea break-up outlines a new "orogenic aureole" around the WAC to the North and West. It is notable that this "aureole" is widely open eastward, as the Meseta Variscides are accreted, not only to the WAC, but along the entire Saharan Precambrian platform. Remarkably, the Ougarta Pan-African segment was reactivated during the Variscan collision, giving birth to an intracontinental belt. The latter belt is connected to the Anti-Atlas, and together these belts represent the foreland fold-belt of the main Variscan segments, i.e. the Mauritanides and the Mesetan Variscides.

Overall, the Variscan Orogeny brought back to Africa continental fragments formerly rifted from Gondwana. They were accreted to the continent together with metasediments whose clastic elements had been mostly taken from the Saharan massifs. Only few mafic and felsic magmatic rocks, either pre-, syn- and post-orogenic added some fresh material to the previous continental mass.

10.1.4 From 250 Ma to Present (cf. Chaps. 1 and 4–8)

The break-up of Pangea disconnected definitively Africa from the Americas, resulting in the formation of the Atlantic margin of Morocco. This is one of the oldest passive margin preserved worldwide, being only perturbed offshore in the western continuation of the Atlas belt and along the Gibraltar Arc. Rifting began as early as the Late Permian, whereas spreading began during the Early Liassic. The 200 Ma-old magmatic event of the giant Central Atlantic Magmatic Province (CAMP) added a significant mass of basaltic trapps and gabbroic intrusions to the Moroccan crust.

Unlike the wide Atlantic spreading, the coeval, Jurassic-Early Cretaceous Tethyan opening hardly separated Africa from Europe. It was characterized by a strong left-lateral movement parallel to the NW Africa margin. The latter margin encroached obliquely onto the Variscan domain: it is located within the Variscan domain in the Moroccan transect and south of it, i.e. within the metacratonic domain, in the eastern Maghreb. The opening of the western Tethys Ocean (Ligurian-Maghrebian Ocean) resulted in the splitting up of the Variscan Belt between Africa and Europe (Iberia). The future Internal Zones of the Betic-Maghrebide belts, i.e. the AlKaPeCa Domain, was either the southern distal part of the Iberian margin or an isolated block within the Tethyan realm.

The Tethyan rifting also affected the Variscan and metacratonic domains south of the African passive margin itself. From the Triassic to the Middle Jurassic, extensional faulting created a mosaic of elongated basins (troughs) and highs (platforms) which prefigured the future Atlas Mountains and Meseta/Plateaus areas. Rifting aborted in this domain during the Middle-Late Jurassic, after the emplacement of a tiny volume of gabbroic rocks (Central High Atlas). A conspicuous shallow structure, namely the West Moroccan Arch (WMA), developed between the Atlantic Coastal Basins and the Tethyan Atlas Gulf. This block acted firstly as a common shoulder between both rift basins, then as a poorly subsident high from the Late Triassic to the Middle Jurassic, before suffering uplift and erosion during the Late Jurassic- Early Cretaceous. From this point of view, the western half of the Maghreb is very different from its eastern half (Eastern Algeria, Tunisia) characterized by a continuous subsidence since the Liassic.

The North Africa passive margin (i.e. the future External Maghrebides) was particularly wide. In its western part, it is characterized during the Jurassic-Early Cretaceous by the occurrence of an elongated zone of thinned crust/oceanic crust between the proximal margin (Prerif, Mesorif) and a distal continental block (Intrarif). This type of wide, dislocated passive margin compares fairly well with that of western Apulia.

From the Late Cretaceous onward, the Africa-Europe convergence tended to close the Tethys oceanic hiatus. Subduction was initiated along the Iberian plate margin (possibly two successive subductions, one beneath Iberia, the other one beneath the Alboran Domain?). In North Africa, few if any shortening did affect the Atlas and External Maghrebide domains before the Middle-Late Eocene. However, large wavelength folding affected probably the entire lithosphere during the Senonian-Eocene interval, being responsible for the individualization of E-trending rises: (1) between the Maghrebide and Atlas Domains (the so-called North Moroccan Bulge, cf. Chap. 4); (2) on the site of the Anti-Atlas and (3) on the site of the Reguibat Shield (Frizon de Lamotte et al., 2008). The general inversion of the Atlas paleofaults began during the Middle-Late Eocene, i.e. before the docking of the AlKaPeCa terranes (Internal Maghrebides) against Africa. The climax of Alpine shortening occurred during the Neogene in the External Maghrebides. It is the direct result of the development of an accretionary prism in front of the moving AlKaPeCa and finally of the "collision" of AlKaPeCa against Africa, which occurred at 18 or 15 Ma. The collision-related deformation propagated progressively from the Mediterranean coast southward up to the South Atlas Front. At the moment, the deformation is clearly concentrated in the Maghrebides, but diffuse deformation also occurs in the Atlas System and even in the Anti-Atlas. On the other hand, the convergence had far-distant intraplate consequences in Africa, marked by swells (Hoggar, Tibesti) and volcanism in several areas including the metacratonic margins of the WAC, but not in the WAC itself.

Thence, the Alpine Orogeny of the Maghrebide Belt corresponds to a new phase of continental accretion, due to the collage of exotic terranes. The observed continental growth remained very limited, as the collage of Iberia against Africa was hampered by the efficient roll-back process which carried the AlKaPeCa terranes toward Africa, then toward the Gibraltar Arc, and opened contemporaneously the Mediterranean Sea between the converging continents.

10.2 Continental Deformation and Topography

Topography is the surface expression of the 2 Ga-long continental evolution reported above. At first glance, the topography of NW Africa offers two strongly contrasting domains (Fig. 10.3). The southern and largest domain consists dominantly of low lands with smooth topography except in the Hoggar (Ahaggar) and Tibesti swells. In contrast, to the north, the Maghreb strip shows a rugged topography with high mountain ranges including lozenge-shaped plateaus and basins.

It is clearly tempting, and indeed largely correct to correlate this contrasted topography with the age of the orogens, which form the corresponding crustal domains. On the one hand, in the main continental domains, the crust consists of Precambrian orogens, almost totally erased during the Paleozoic. Both the Variscan and Alpine deformations have been negligible there, relative to the northern domain. Only the Anti-Atlas and Ougarta belts show significant deformations as foreland

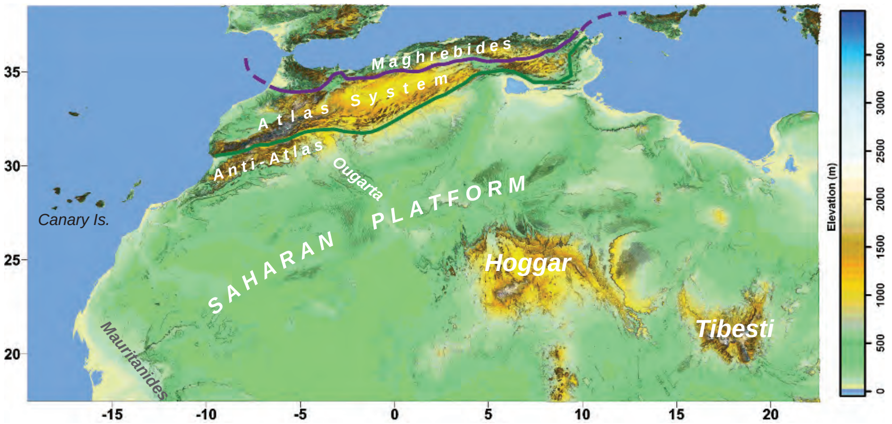

Fig. 10.3 Topography of NW Africa (GTOPO30 database) showing that the main current relief is concentrated in the NW fringe of the continent, i.e. in the Maghreb (Atlas Belts) but that major swells (Hoggar, Tibesti) also occur within the continent itself, in the Pan-African and metacratonic regions of the Saharan platform (compare with Fig. 10.2)

fold belts during the Variscan Orogeny. On the other hand, the Maghreb rugged topography correspond to the continental rim which suffered the superimposed Variscan and Alpine orogenic cycles, each one involving successively rifting and collisional events. The youngest and still ongoing collision affected a weak, deeply fractured continental crust, resulting in the observed mountainous topography.

However, this scheme does not account for all the topographic particularities in these regions of NW Africa, neither qualitatively or quantitatively. One of the issues in debate concerns the origin of the Hoggar and Tibesti swells in the main, cratonic-metacratonic domain. As they coincide with the distribution of Cenozoic (Late Eocene-Pleistocene) volcanic centres (Fig. 10.2; Dautria et al., 2005), they have been classically considered to be the products of mantle plumes (e.g. Aït Hamou et al., 2000, with references therein). However, Liégeois et al. (2005) argue that at present, several observations and data including geological, heat flow and tomographic results do not support the mantle plume model (cf. Chap. 1, Fig. 1.21). They emphasize the volcanic centres close relationship with fractures zones in the Pan-African basement. Thence, they suggest that intraplate stress induced by the Africa-Europe convergence reactivated the Pan-African mega-shear zones inducing linear lithospheric delamination, asthenospheric upwelling and melting due to pressure release. This debate certainly deserves new studies, in particular more accurate tomographic images and geophysical models.

Another issue concerns the asymmetry of the Maghreb topography from west to east. Why do the highest mountains occur in Morocco (Marrakech High Atlas)? Why the only mountain range south of the South Atlas Fault (SAF) is the Moroccan Anti-Atlas? An intuitive cause should be the presence there of the West African craton that induced a larger rheological contrast that has been used already during the Jurassic rifting. Rift weakening and craton rigidity could have concentrated the shortening effects. However, it is known for long that the crustal root created by the Alpine shortening beneath the Atlas Mountains is not thick enough to isostatically support the topography. Different authors modelled the lithospheric structure taking into account gravity, geoid, heat flow and topography. They showed that the lithosphere is thinned to 60 km below the Anti-Atlas and High-Atlas (see Chap. 1, Fig. 1.19). Modelling the effect of such lithospheric thinning shows that the whole topography of the Anti-Atlas and up to half of the relief of central High Atlas is due to lithospheric thinning, not only to crustal shortening (cf. Chap. 4, Fig. 4.46). Moreover, the NE-trending thinned lithosphere strip coincides with the distribution of Miocene to recent alkaline volcanism. This lithospheric structure is oblique on the structural grain of the crust as it crosscuts not only the SAF, but also the Jurassic Atlantic margin up to the Canary Islands to the south, and the Oligocene-Miocene Alboran Basin up to eastern Spain and possibly the French Central Massif to the north ("Morocco Hot Line" hypothesis; Fig. 4.47). Again, this issue deserves further geophysical and geological studies. In this general framework, the asymmetry of the Maghreb topography from west to east is mainly due to the existence of the Cenozoic "Moroccan Hot Line", whose development was probably enhanced by the lithosphere tearing along the Maghrebian margin (Frizon de Lamotte et al., 2008).

10.3 Further Studies

Morocco is certainly one of the African countries whose geology is best known. As shown in the Chap. 9, this results from a long history but above all from an exemplary coordination between the Geological Survey of Morocco ("Service Géologique du Maroc") and the academic research handled in national and foreign universities. By the beginning of the seventies, oil and gas exploration, under the supervision of ONAREP ("Office National pour le Recherche Pétrolière") now included in the ONHYM ("Office National des Hydrocarbures et des Mines"), led to a better knowledge of sedimentary basins, borders of orogenic belts and both Atlantic and Mediterranean margins. The exploration remains very active, in particular the deep offshore targets, which bear the main hopes of discovery. Concurrently, mining exploration and related academic research restarts as a direct consequence of the price of raw materials worldwide.

Beyond these economic stakes, societal stakes are manifold. They firstly concern the question of water resources and environmental problems. Among them, we emphasise the evaluation of seismic hazard. Indeed Morocco is globally subject to a moderate risk. However the recent Ms 6.0 earthquake of Al Hoceima (2004 February 24th) recalls that the Earth remains murderous and that we have to increase our vigilance.

From an academic point of view, Morocco is a wonderful natural laboratory attracting researchers from many countries. This territory offers numerous opportunities for detailed studies. For a better understanding of its geological evolution, we are waiting for more and more data: more geochemical data, more geochronological and thermochronological data, more stratigraphic data (in particular in continental series) etc… One of the main frontiers is the knowledge of the deep crustal and lithospheric structure of the country. Morocco provides a relatively restricted region where a great variety of geophysical, geodynamic and geochemical processes associated with extension, subduction and continental collision can be addressed. To improve our understanding of these processes, we suggest promoting collaborative, international multi-disciplinary projects. The first general objectives could be:

- to determine the three-dimensional structure of the crust and lithosphere through high resolution tomographic images and seismic profile acquisition across the orogenic systems (Rif, Atlas) and the margins (Atlantic and Alboran margins);
- to monitor the motions of crustal blocks in time and space in order to constrain earthquake models as well as mountain building processes;
- to analyse the feedback relationships between tectonics, topography and climate.

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Index[1]

[1]: Indexes both geographic and geological (subject) names, together with some fossil names (mostly Vertebrata). Stages, epochs and eras are not indexed, except those of the Quaternary. The only author's names indexed are those of the pioneers of the Moroccan geology.