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Tectonic evolution of the Gibraltar Arc

Flinch, Joan Francesca, Ph.D.
Rice University, 1994
RICE UNIVERSITY

TECTONIC EVOLUTION OF THE GIBRALTAR ARC

by

JOAN F. FLINCH.

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APPROVED, THESIS COMMITTEE:

D. L. Bally, Chairman
Harry Carothers Wiess Professor

Manik Talwani
Schlumberger Professor

John Anderson
Professor of Geology

Ricardo Yamal
Professor of Spanish

Houston, Texas
August, 1993
ABSTRACT
TECTONIC EVOLUTION OF THE GIBRALTAR ARC
Joan Flinch

The Betic Cordillera of Spain and the Moroccan Rif constitute the northern and southern branches of the Gibraltar Arc, which is the western limit of the Alpine-Mediterranean system. The frontal units (i.e., the Guadalquivir Allochton of the Betic Cordillera and the Prerifaine Nappe of Morocco) have in the past been interpreted as olistostromes. Seismic data from the frontal part of the Gibraltar Arc suggest an accretionary complex migrating towards the west from the Western Mediterranean Basin. Seismic data in the Gulf of Cádiz, in the northwestern Atlantic margin of Morocco between Rabat and Tanger and in the Rharb Basin of northern Morocco have been interpreted and compared with field examples from the external Western Rif and the Guadalquivir region of the Betic Cordillera. The structure of the accretionary wedge consists mainly of imbricated thrusts and low-angle extensional detachments. The structures and internal deformation observed are similar to present-day accretionary wedges. Extensional and compressional structures are coeval with the foredeep development. The emplacement and collapse of the wedge were very rapid and occurred during Tortonian and Messinian time. The internal structure of the accretionary wedge is difficult to map since it originated from a deep-water passive margin succession with allochthonous Triassic evaporites and turbiditic wedges (flysch domain). The Prerifaine Nappe and the Guadalquivir Allochthon record several stages of accretion and westward motion of the Alborán domain, providing important constraints for the evolution of the Gibraltar Arc. Frontal accretion is coeval with uplift in the internal domain and back-arc extension in the Alborán region. A detailed sequence stratigraphic analysis of the Supra-Nappe succession has provided insights into the geodynamic evolution of the region and the effect of Pleistocene glacio-eustatic fluctuations.
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1. INTRODUCTION

1.1 OBJECTIVES

The Betic Cordillera of Spain and the Moroccan Rif constitute the northern and southern branches of the Gibraltar Arc, which is the western limit of the Alpine-Mediterranean system. The frontal units (i.e. the Guadalquivir Allochton of the Betic Cordillera and the Prerifaine Nappe of Morocco) have been interpreted asolistostromes. Seismic data from the front of the Gibraltar Arc suggest an accretionary complex migrating towards the west from the Western Mediterranean Basin. Defining details of this accretionary wedge and a comparison of the seismic expression of the wedge with outcrops in Spain and in Morocco are the basic objectives of this study. These data will be integrated to construct a three-dimensional model of the wedge.

The evolution of the accretionary complex, the continuity of the Gibraltar Arc and particularly its foreland provide important constraints on the plate displacements between Iberia and North Africa. The main problems that will be addressed are:

- The relationship of the frontal accretionary complex emplacement to the back-arc extension of the Alborán Sea.

- Constraints of the progression of accretion of the Alboran block to Africa and Iberia.

- The accretionary wedge emplacement and the development of the foredeep.

- The 3D structure of the accretionary wedge and the integration of field observations with seismic data.
1.2 SOURCES OF DATA AND METHODOLOGY.

1.2.1. INTRODUCTION

Offshore seismic data of the Gulf of Cádiz and the northwestern Atlantic continental margin of Morocco, together with additional data from the Rharb Basin in northern Morocco, constitute the main data set for this study. Surface data from a number of geologic maps of the Bético and Rif Cordilleras and additional data collected in the field have been integrated with seismic profiles and well logs (Fig. 1).

To integrate many sources of data over a large area, a tectonic map of the Gibraltar Arc and structural sections across the arc have been compiled. Seismic data from offshore areas and foreland regions also have been integrated with surface and subsurface data in the folded belt. Several seismic lines have been converted to depth and used to construct regional cross-sections.

1.2.2 SUBSURFACE DATA

Seismic data were provided by ONAREP-PETROCANADA for the Moroccan side and by REPSOL for the Spanish side of the Gibraltar Arc. SECEGSA provided high resolution data from the Straits of Gibraltar. The seismic data set used in this project is not entirely comparable. Seismic parameters vary depending on the area and year of acquisition (Table 1.1 describes the seismic data used for this study).
Fig. 1. Seismic data presented in this work. Dashed line shows the shelf break.
<table>
<thead>
<tr>
<th>Area</th>
<th>source</th>
<th>migration</th>
<th>penetration (sec)</th>
<th>fold</th>
<th>year</th>
</tr>
</thead>
<tbody>
<tr>
<td>Offshore Rabat</td>
<td>Air gun</td>
<td>migrated</td>
<td>6 sec</td>
<td>48</td>
<td>1987</td>
</tr>
<tr>
<td>Offshore</td>
<td>hydroseis</td>
<td>non-migrated</td>
<td>4 sec</td>
<td>24</td>
<td>1970</td>
</tr>
<tr>
<td>Larache</td>
<td>airgun</td>
<td>non-migrated</td>
<td>5sec</td>
<td>48</td>
<td>1974</td>
</tr>
<tr>
<td>Lalla Yto</td>
<td>vibrator</td>
<td>migrated and</td>
<td>4.2sec</td>
<td>24</td>
<td>1984</td>
</tr>
<tr>
<td></td>
<td></td>
<td>non-migrated</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nouriat</td>
<td>vibrator</td>
<td>migrated and</td>
<td>5 sec</td>
<td>24</td>
<td>1981</td>
</tr>
<tr>
<td></td>
<td></td>
<td>non-migrated</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sebou</td>
<td>dynamite</td>
<td>non-migrated</td>
<td>5 sec</td>
<td>24</td>
<td>1972</td>
</tr>
<tr>
<td></td>
<td>dynamite</td>
<td>non-migrated</td>
<td>4 sec</td>
<td>24</td>
<td>1974</td>
</tr>
<tr>
<td></td>
<td>vibrator</td>
<td>migrated</td>
<td>3 sec</td>
<td>24</td>
<td>1980</td>
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<tr>
<td>Prerif</td>
<td>dynamite</td>
<td>non-migrated</td>
<td>8 sec</td>
<td>24</td>
<td>1977</td>
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<tr>
<td>Gulf of Cádiz</td>
<td>Maxipulse</td>
<td>non-migrated</td>
<td>4 sec</td>
<td>48</td>
<td>1974</td>
</tr>
<tr>
<td></td>
<td>Vaporchock</td>
<td>migrated</td>
<td>4 sec</td>
<td>48</td>
<td>1978</td>
</tr>
<tr>
<td></td>
<td>Vaporchock</td>
<td>non-migrated</td>
<td>4 sec</td>
<td>48</td>
<td>1979</td>
</tr>
<tr>
<td></td>
<td>HP air gun</td>
<td>migrated</td>
<td>6 sec</td>
<td>48</td>
<td>1982</td>
</tr>
<tr>
<td>Gibraltar Strait</td>
<td>Flexichoc</td>
<td>migrated</td>
<td>1 sec</td>
<td>24</td>
<td>1987</td>
</tr>
<tr>
<td>Alborán</td>
<td>Vaporchock</td>
<td>non-migrated</td>
<td>4 sec</td>
<td>24</td>
<td>1975</td>
</tr>
</tbody>
</table>

Table 1.1. Main parameters of the seismic data used for this study.

Dense well-log coverage in the Rharb Basin of Morocco permits a good correlation in the onshore area. But for the offshore area of Morocco, only one well is available to constrain the timing of the seismic units. However, the lack of synthetic
seismograms for most of the wells does not allow an accurate tie with the seismic information. The deep well BB1 Société Chérifienne des Pétroles (SCP 1957) of the Prerifaine zone has been used to constrain the top of the Jurassic platform underlying the outcropping thrust sheets. In the Gulf of Cádiz area, a few selected wells were used to calibrate the main seismic reflectors. Published well data in the Betic Cordillera (IGME 1987) were used to construct geological cross-sections through the Guadalquivir Basin and the External Betic Cordillera. Velocity information was compiled from sonic logs of several onshore and offshore wells and used for depth conversion. The most typical interval velocities for the study are outlined in Table 1.2. Key seismic lines have been reproduced as line drawings (see Panels 1, 2, 3 and 4). Planktonic foraminiferal biostratigraphy from wells was tied into seismic data to constrain the timing and evolution of the accretionary complex.

Table 1.2 lists typical velocities for the frontal part of the Gibraltar Arc.

<table>
<thead>
<tr>
<th>Zone</th>
<th>Acoustic basement</th>
<th>Lower Imbricates</th>
<th>Accretionary Complex</th>
<th>Supra-complex units Pliocene</th>
<th>Supra-Complex units Pleistocene</th>
</tr>
</thead>
<tbody>
<tr>
<td>velocity m/s</td>
<td>6000</td>
<td>4500</td>
<td>3250</td>
<td>2600</td>
<td>2200</td>
</tr>
</tbody>
</table>

Table 1.2. Typical velocities for the Offshore Rabat-Western Rharb Basin.

1.2.3 SURFACE DATA

Field work was conducted in selected key areas of the most external units of the Rif and Betic Cordilleras. In the Betic Cordillera, field-work was done in the Guadalquivir, i. e. the provinces of Jaén, Sevilla, Córdoba and Cádiz. In the Moroccan Rif, field observations were made in the following areas:

- Rharb Basin, regions of Souk el Arba, Arbaoua and Lalla Zhara.
- Northwestern Moroccan Atlantic Coast in Tanger, Cap Spartel and Asilah-Larache.
• Central Rif in Ouezzane, Beni Amar, Karia ba Mohamed, Taounate and Fès.
• Internal Rif in Oued Laou Valley.

Surface exposures in the most external regions of the Gibraltar Arc are rare and of poor quality. For this reason, high-quality seismic data were used to constrain the geometry of the structures. A comparison between seismic and field data was undertaken to verify the results of seismic interpretation. Some seismic lines have been constrained, when possible, by surface regional cross-sections. The hypothesis derived from seismic observation, that the frontal units of the Gibraltar Arc represent an accretionary complex, has been explored in some detail in the field.

1.3 PREVIOUS KNOWLEDGE ABOUT THE BETIC AND RIF CORDILLERAS

1.3.1 BETIC CORDILLERA

Mallada's (1895-1911) explanatory note of the "Mapa Geológico de España" was the first serious contribution to the knowledge of the geology of the Betic Cordillera following many isolated observations made by scientists or engineers during the XIX Century. After the big Granada earthquake of 1884 the Academy of Science of Paris sent the "Mission d'Andalousie". The report by French geologists (Mission d'Andalousie 1888) contributed much to the knowledge of the Betic Cordillera. These authors were the first to contrast the difference between the Internal and External zones and to recognize the presence of Alpine facies, as in other Alpine domains.

Suess (1885-1909) was the first to point out the connection between the Rif Cordillera, the Maghrebian Mountains and the Balearic Islands. At the beginning of the XX century, Gavala (1916) made important contributions to the geology of the Cádiz region. Orueta (1917) made interesting petrographic observations in the Ronda area. The allochthony of the Betic units, defined by the presence of thrust sheets, was first shown by
Nickles (1904). Gentil (1918) emphasized the allochthony of the most frontal units of the Cordillera.

During 1925-1936, knowledge of the Betic Cordillera greatly increased thanks to the works of Fallot, Blumenthal and Brouwer, all authors with experience in the Alps and other areas of the Alpine system. Staub (1934) described the structure of the Betic Cordillera as a pile of thrust sheets comparable with the Eastern Alps. Blumenthal (1927) focused his work in the Betic Cordillera west of Granada. Fallot established the first correlation of the Mesozoic series (Fallot 1931-34) and focused on the Subbetic Zone (Fallot 1948). Fontboté and Vera (1983) outlined the geology of the Betic Cordillera in the context of the Iberian Alpine folded belts, presenting a synthesis of previous work as well as their own view of the main problems.

The "Mapa geológico y minero de Andalucía" (Junta de Andalucía 1985) represents a synthesis with a new interpretation of the Betic Cordillera, emphasizing the role of extensional tectonics. SECEGSA (1990) presented a detailed map of the Gibraltar Straits area. The most recent works by the school of Granada (Galindo-Zaldivar et al. 1989, Balanyá and García-Dueñas 1991, García-Dueñas et al. 1992, Comas et al. 1992) emphasize the role of extensional detachments in the structure of most of the Betic Cordillera and its relationship with the extension in the Alborán region.

1.3.2 RIF CORDILLERA

Coinciding with the XIX International Geological Congress, Choubert and Marçais (1952) presented a basic condensed synthesis of Moroccan geology titled: "Géologie du Maroc". A geologic map of the Rharb and Western Preifir at scale 1:200,000 was put together by the Société Chérifienne des Pétroles (S.C.P.) for the occasion of the International Congress (Bruderer and Lévy 1954). Durand-Delga and others (1960-62) presented a geological synthesis of the Rif. The tectonic map of Africa "Carte tectonique
internationale de l'Afrique au 1/50000,000" (1968) brought new contributions to the knowledge of Moroccan tectonics.

Faure-Muret and Choubert (1971) described the Moroccan folded belts in the context of African tectonics. Andrieux (1971) studied the Central Rif, especially the Intrarifean Ketama unit, tracing the Alpine schistosity front.

A special meeting of the Geological Society of France, "Réunion extraordinaire de la société Géologique de France" (1972) organized by Faure-Muret, was dedicated to the Gibraltar Arc. Contributions to the geology of the Betic and Rif Cordilleras and the connections between both folded belts were the main topics of the conference.

Michard (1976) outlined the geology of Morocco, dedicating a chapter to the Rif; his work is a summary and updated version of previous works.

All mapping done by the Société Chérifienne des Pétroles (S. C. P.) (Tilloy 1955 a, b, c, d) and a compilation of new data were put together in the "Carte géologique et Carte structurale de la chaîne Rifaine 1:50,000" (Geological and Structural map of the Rif Cordillera) by Suter (1980a,b).

The structural map of Morocco "Schema structural du Maroc" (Saadi 1982) outlines the main structural domains of Morocco.

Wildi (1983) presents a stratigraphic and structural synthesis of the Rif Cordillera, emphasizing its connection with the Tell and other Maghrebian folded belts. Wernli (1988) established the Neogene biostratigraphy of northern Morocco that will be used in this study. A new contribution to the neotectonics of the Rif Cordillera was presented by Morel (1988).
2. THE GIBRALTAR ARC

2.1. GEOGRAPHIC AND GEOLOGIC SETTING.

The Gibraltar Arc is the western limit of the Alpine-Mediterranean system. The Betic Cordillera in southern Spain and the Rif Cordillera in northern Morocco constitute the northern and southern part of the arc (Fig. 2.1). The Betic Cordillera extends from the Straits of Gibraltar in the west to the Jucar Valley (Valencia) in the east. The NE-SW structural trend of the Central and Western Betics becomes a NNE-SSW trend in the Eastern Betics. This change occurs east of Sierra de Cazorla, coinciding with the Tiscar fault. The Guadalquivir Valley is located in front of the mountain belt and is drained by the Guadalquivir River. The Guadalquivir Basin represents the foreland basin of the Betic Cordillera. Highly deformed metamorphic Paleozoic and Triassic rocks are exposed in the Sierra Nevada, the Sierra de los Filabres and the Montes de Málaga Mountains. The highest elevation in the Betic Cordillera is the Mulhacen peak in Sierra Nevada (3,478 meters). The Balearic Islands in the Western Mediterranean are part of the Eastern Betics, but the opening of the Balearic Basin separated them from mainland Iberia. The frontal allochthonous units of the Betic Cordillera (i.e., the Guadalquivir allochthonous units) extend to the Gulf of Cádiz and farther west to the so-called Horseshoe "Fer du Cheval", east from Gorringe Bank and the Seine abyssal plain in the Central Atlantic.

The Rif Cordillera in northern Morocco extends from the Tanger Peninsula in the north to the Melilla Bay in the east. The highest mountains are located close to the Alborán Sea, reaching almost 2000 meters above sea level. The Martil and Laou Rivers flow into the Mediterranean. The relief rapidly decreases from the Mediterranean to the Atlantic, passing into a hilly country of mild topography ("collines prérifaines"). The Rharb Basin is a plain located south of the mountains, drained by the Sebou, Loukkos, Dra and Bou Regreg Rivers. In the Eastern Rif the basins of Guercif and Saïs separate the Rif from the Middle Atlas Mountains.
Fig. 2.1. Major tectonic units of the Western Mediterranean region. After Dewey and others (1989).
2.2. THE WESTERN MEDITERRANEAN

2.2.1. INTRODUCTION

The Western Mediterranean region consists of a collage of several blocks located between the Euro-Asiatic and the African plates. These intermediate blocks or microplates (i.e. Iberia, Alboran, Corsica-Sardinia and Apulia) interacted between each other and with Eurasia and Africa, defining the geodynamics of the region.

Dewey and others (1989) noted:

"The western Mediterranean basins evolved between the converging European and African plates. Previous kinematic models have attempted to relate these large plate motions directly to the tectonics of the region. However, Africa’s motion with respect to Europe is not always the factor that determines the geological evolution of this region. Rather, as shown by vector triangles, the motion of individual blocks with respect to each other often outweighs the importance of the Africa-Europe motion in creating the geology we observe today".

The present-day Western Mediterranean postdates the cessation, during Late Eocene-Oligocene, of several important orogenic events, such as the Penninic thrusting in the Alps, the Pyrenean orogeny and the collision in Corsica (Dewey et al. 1989). The configuration of the Alborán Sea is the result of rifting and oceanic spreading superimposed into a collage of Eo-Alpine orogenic belts.

The evolution of the Western Mediterranean can be subdivided into a number of stages:

2.2.2 BREAKUP OF PANGEA

During Permian and Triassic time the European and North African Hercynian folded belt was affected by extensional tectonics. In East Greenland and the Pyrenees rifting started during Middle Triassic (Ziegler 1987). Half-grabens filled with red-beds and shallow-water limestones and evaporites developed in the European and northern African
domains with facies similar to the German Basin. These facies are known as Germanic-Type Triassic, versus the Alpine-Type Triassic consisting of platform carbonates that occupied the Tethyan region.

In northern Africa, the breakup of Pangea took place on the Atlantic margin and in the Atlasic region. The Atlantic rift is represented by the Essaouira-Agadir coastal basins, and the Atlasic rift is composed of the Central High Atlas and the Middle Atlas grabens. Upper Triassic basic volcanism was associated with Atlasic rifting (Fig. 2.2). According to some authors the South-Atlasic fault was connected and facing Newfoundland transforms (Laville and Piqué 1991).

2.2.3 TETHYAN CARBONATE PLATFORM

Several carbonate platforms developed in the Tethys domain during Lower-Jurassic time: The Algarve-Gulf of Cádiz and the Betic platform in the southern Iberian continental margin (Vera 1983, García-Hernández et al. 1987, Martín-Algarra 1987), and the platform carbonates of the Rides Prerifaines and the Middle and High Atlas systems in northern Africa (Fig. 2.3). The carbonate platform of the northern African rifted margin extended into the Apenninic domain into the Apulia platform (Dewey et al. 1987, Ziegler 1987) (Fig. 2.3).

During Lower Pliensbachian (190 Ma) the basement underlying the Betic platform broke up (Vera 1983, Martín-Algarra 1987). Jurassic mafic volcanism and pillow-lavas in the Subbetic Zone of the External Betic Cordillera are associated with this rifting event (Puga et al. 1989). During Late Callovian time the second rifting stage occurred. Oceanic spreading, active since Bathonian time in the Central Atlantic, was transferred to the Tethys (Andrieux et al. 1989). NE-SW trending shoals and troughs characterized by normal faults and tholeitic to intermediate volcanism in the Panormide-Imerese-Trapanese-Sicano area represent rifting in the Apenninic region (Catalano and d'Argenio 1978).
Fig. 2.2. Plate tectonic reconstruction of the African and Iberian plates during Hettangian time. Modified after Andrieux et al. (1989), Favre and Stampfli (1992) and Ziegler (1987).
Fig. 2.3. Plate tectonic reconstruction of the Western Tethys and Northern Atlantic regions during Pliensbachian time. Modified from Ziegler (1987), Andrieux et al. (1989) and Favre and Stampfli (1992). The figure shows the location of the main carbonate platforms.
2.2.4 OCEANIC SPREADING.

Oceanic spreading during the Callovian (165 Ma) is recorded by the oldest ophiolitic remnants in the western Mediterranean (Puga et al. 1989). Upper Jurassic pelagic sedimentation, coeval with spreading, is thought to be associated with a hypothetical north African transform zone. This transform zone was probably located in the present-day Alborán domain (Andrieux et al. 1989).

A large input of terrigenous turbidites (deep flysch basin) occurred in the Western Atlas and Algeria, extending to the Rif and the Oran region during Callovo-Oxfordian time (Wildi 1983). During Aptian-Albian time renewed terrigenous influx from the Sahara covered the whole north African margin (Wildi 1983). The north-African margin evolved as a terrigenous siliciclastic margin, while the south Iberian margin remained as a carbonate dominated margin, a situation that resembles the North American and Caribbean margins of the present-day Gulf of Mexico. The separation of Iberia from North America was initiated during the Hauterivian (118 Ma) by seafloor spreading north of the Azores-Gibraltar fracture zone (Ziegler 1988).

2.2.5 EARLY-ALPINE COMPRESSION AND SUBDUCTION

From 118 Ma to 34 Ma (Middle Cretaceous to Uppermost Eocene) the motion between Africa and Eurasia changed from east-west transtension to a northeasterly directed compression. This change in motion could be reflected by the 92 Ma (Cenomanian-Turonian boundary) old High Pressure (HP) eclogitic metamorphism in the Alps (Dewey et al. 1989). HP/HT metamorphism related to subduction took place 115-85 Ma ago in the Betic Cordillera (de Jong 1991). Widespread turbiditic ("Flysch") sedimentation extending from the Western Alps to the Gibraltar Arc (Dercourt et al. 1986) was coeval with this initial HP/LT metamorphism. After anomaly 30 (earliest Paleocene) (66.7 Ma) the convergence of Africa and Europe slowed dramatically and was associated with a change in orientation of the Charlie-Gibbs fracture zone (Dewey et al. 1989).

2.2.6 PYRENEAN OROGENY.

During Late Eocene time (40 Ma) the main Pyrenean compression and the structural inversions of the North Sea occurred (Ziegler 1988). Iberia moved as an independent plate and compression developed in its contact with the Euroasiatic (Pyrenees) and the African plate (Betic and Rif) (Roest and Srivastava 1991). Strike-slip along the Gloria fault zone of the Azores-Gibraltar fracture passed laterally into compressional ridges in Gorringe Bank.

2.2.7 OLIGOCENE RIFTING.

During Oligocene time, rifting developed in several regions of western Europe, i.e. the Rhine Graben, the Balearic Basin, the southern Rhône Valley, the Gulf of Lion, Sardinia and the Valencia Trough (Dercourt et al. 1986, Ziegler 1988) (Fig. 2.4). During this time the structure of the Pyrenees was completed and the Ebro and Aquitanian foredeeps were filled by fluvial sediments (Fig. 2.4). Turbiditic "Flysch" sedimentation took place in the foredeep of the Western Alps (Dercourt et al. 1986).

2.2.8 ALPINE STRUCTURAL EVOLUTION

From Lower Oligocene to Middle Miocene time (33 to 13 Ma), volcanic calc-alkaline magmatism occurred in several regions of the Western Mediterranean, e.g. Sardinia, the Western Alps (Rehault et al. 1985), and the trans-Alborán corridor (Hernández et al. 1987).

The continued structural development of the Alpine orogenic belts and the onset of subduction in the Bay of Biscay started during Late Oligocene time (Ziegler 1988)
Oceanic spreading followed rifting and was contemporaneous with oceanic crust generation in the Balearic Basin (Rehault et al. 1985).

The Apenninic deformation started during Late Oligocene (27 Ma). By that time much of the older oceanic crust of the Tethys was subducted, and an accretionary wedge with NE transport direction developed (Knott 1987). Extension occurred in the internal part of the Kabylie-Calabrian belt. Subduction was accompanied by abundant calc-alkaline volcanism in western Sardinia and southeastern France. In the Gulf of Valencia area, Mallorca and Menorca rotated away and detached from Iberia (Dewey et al. 1987) (Fig. 2.5).

During Aquitanian time (23 Ma), extensional tectonics occurred throughout the Mediterranean region. The north Balearic Basin opened along transform faults, and Sardinia began to drift from mainland France as rifting developed in the Catalonian coastal ranges (Dercourt et al. 1986) (Fig. 2.5). Numidian flysch was deposited in a trough developed in front of the Apeninnes and the Maghrebian folded belts (Catalano and d’Argenio 1978) (Fig. 2.5). Sedimentary filling of the Swiss molasse basin in the Western Alps was initiated during Aquitanian time. Oceanic spreading continued in the Balearic Basin. Forearc compression in the western Mediterranean orogenic belts was coeval with back-arc extension (Horvath and Berkhemer 1982). During Burdigalian time (18 Ma), Corsica and Sardinia had already completed their rotation. In the central Western Mediterranean, only the southernmost part of the South Balearic Basin remained to be formed. The Kabylies and Calabria migrated southeastward towards the Algerian-Tunisian margin (Bouillin 1984) (Fig. 2.5). The Liguride accretionary wedge developed during that time (Dewey et al. 1989). During Tortonian time, compression continued in the Apeninnes, Western Alps, Carpathians, Betic and Rif Cordilleras, while the foredeep basins were filled with marine sediments (Dercourt et al. 1986) (Fig. 2.6). Numidian flysch was reworked and deposited in the Apenninic foredeep. Subduction in the Ionian Sea was
Fig. 2.5. Early Miocene plate tectonic reconstruction of the Alpine-Mediterranean system. Modified after Descout et al. (1986) and Dewey et al. (1989).
Fig. 26. Tortonian plate tectonic reconstruction of the Alpine-Mediterranean system. Modified after Dercourt et al. (1986) and Dewey et al. (1989).
coeval with accretion in western Sicily (Dewey et al. 1989) (Fig. 2.6). At anomaly 5 (Late Miocene, 8.9 Ma), Africa began moving to the NW. This change in direction of the African plate enhanced extension and strike-slip in the Tyrrenian Basin (Dewey et al. 1989). At this time, Calabria and NE Sicily displaced towards the southeast. During Late Miocene time (from 7 Ma to 5 Ma), a westward-northwestward directed movement of Africa with respect to Eurasia resulted in tightening of the Calabrian (Dercourt et al. 1986, Dewey et al. 1989) and the Gibraltar Arc (Dercourt et al. 1986, Aît Brahim and Chočín 1984, 1989, Aît Brahim 1985, Morel 1988, this thesis). During Messinian time, a special event took place throughout the Mediterranean region, the so-called Messinian salinity crisis (e.g. Hsü et al. 1973). The Mediterranean was disconnected from the Atlantic, resulting in thick evaporite deposition. Back-arc extension continued throughout this event, while intra-montane grabens developed in the Calabrian and Gibraltar Arcs and the southern Apennines. The locus of Tyrrenian extension shifted to the SE (Dewey et al. 1989). A large accretionary wedge developed in front of the Calabrian Arc (Roure et al. 1991) and the Gibraltar Arc (this study). During Middle Pliocene-Pleistocene time, deformation in the southern Apennines and the Calabrian and Gibraltar Arcs was controlled by strike-slip faulting. Subduction in the Aegean Arc took place in this scenario (Dercourt et al. 1986) (Fig. 2.7). The Straits of Gibraltar as we know today were formed during Pliocene time (Fig. 2.7). Late uplift culminates the evolution of the Alpine folded belt in the Balearic Islands (Roca 1992) and in the Betic and Rif Cordilleras (Cadet et al. 1977, 1978).
Fig. 2.7. Present-day structural sketch of the Alpine-Mediterranean system. Modified after Dercourt et al. (1986) and Dewey et al. (1989).
2.3 MODELS OF THE GIBRALTAR ARC

Several models have been proposed to explain the arcuate geometry and evolution of the Gibraltar Arc. These models and the authors who support them have been summarized in Table 2.

Following the earliest geological surveys in the Betic and Rif Cordilleras, the presence of an Internal domain with a completely different stratigraphy and tectono-metamorphic history than the External domain became evident. The Internal domain, strange to Iberia and Africa, was assigned to the Alborán microplate or block which was initially located (i. e. before collision) in the eastern Tethys (Andrieux et al. 1971).

2.3.1 STRIKE-SLIP/CURVED INDENTER MODEL.

This model is based on the plastic indentation of an arcuate block or plate (the Alborán block) with the Iberian and African blocks (Andrieux et al. 1971, Andrieux and Mattauer 1973) (Fig. 2.8). According to most authors this indentation is accompanied by strike-slip faulting on the southern and northern borders of the Alborán block. Geological studies in the Internal domain of the Betic and Rif Cordilleras emphasized the important role of strike-slip faults, like the Jebha and Nekor faults of the Rif and the Carboneras, Palomares or Alhama de Murcia faults of the Betic Cordillera.

These observations led to the proposition of strike-slip models. According to these models, the emplacement of the Alborán block took place along two megashear zones (Kampschuur and Rondeel 1975) or else four major strike-slip faults (Leblanc and Olivier 1984). These models were modified by Bouillin et al. (1986) introducing the ALKAPECA block concept (Alborán-Kabylies-Calabria). Olivier (1981-82, 1990) emphasized the role of Miocene strike-slip tectonics in the curvature of the Gibraltar Arc. Sanz de Galdeano (1990) presented a detailed paleogeographic model in which strike-slip deformation has a significant role in the westward emplacement of the Gibraltar Arc.
### Geodynamic Models of the Gibraltar Arc

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Table 2. Proposed geodynamic models for the Gibraltar Arc.
Fig. 2.8. Emplacement of the Alborán domain according to Andrieux et al. (1971). a Present day situation of the Alborán domain. b Location during Early Miocene. c Emplacement model.
2.3.2 EXTENSIONAL COLLAPSE MODEL.

A better understanding of the Alpine-Mediterranean system and the Cordilleran system in North America attested to the importance of extensional structures in orogenic belts. Horvath and Berckhemer (1982) considered the Pannonian, the Tyrrenhian, and the Alborán Basins as back-arc basins of the respective Carpathian, Calabrian and Gibraltar Arcs.

Extensional collapse of a thickened continental crust has been proposed as a valid mechanism for the formation of the Alborán Sea (Dewey 1988). In Dewey's model the present-day Alborán Sea represents an intermediate evolutionary stage of collapse. According to Platt and Vissers (1989) the compression in the Gibraltar Arc and the coeval extension in the Alborán Sea can be explained by the extensional collapse of an east-west oriented collisional ridge. Extensional collapse was associated with the elimination of the lithospheric root (Fig. 2.9). Platzman and Lowrie (1992) and Platzman (1992) support this model on the basis of paleomagnetic data from the Betic and Rif Cordilleras (Fig. 2.9). According to Doblas and Oyarzun (1989) the structure of the Gibraltar Arc is the result of the emplacement of Neogene core complexes.

2.3.3 OROCLINAL BENDING MODEL

Microstructural data from the ultramafic massifs and crustal rocks of the Internal Betic and Rif domains are consistent with oroclinal bending of the arc (Tubia and Cuevas 1987). Micro-structural studies in the Internal domain of the Betic (Balanyá et al. 1987) and Rif Cordillera (Chalouan and Michard 1990) indicate superposition of Alpine thrusting. Older transport direction of thrusts has rotated counterclockwise with respect to the younger directions in the Rif Cordillera and clockwise in the Betic Cordillera, suggesting oroclinal bending (Fig. 2.10).
Figure 6. Tectonic evolution of Alboran region. a: Mid-Oligocene Collisional ridge with thickened lithospheric root formed by Late Cretaceous-Paleogene convergence. Map (above) shows coastlines around Gibraltar and Tangiers for reference, and plate motion vectors for Africa relative to Europe from 27 Ma to present after Dewey et al. (1989). Basins were probably underlain by thin continental crust. b: Burdigalian. Convective removal of lithospheric root has caused uplift, increase in potential energy of collisional ridge, and extension. Extension is accommodated by crustal shortening around margins. High-T peridotite emplaced at base of crust. c: Extending Alboran domain has been emplaced onto surrounding continental margins; center subsides as lithosphere thickens by cooling and continued slow convergence. Note: Crustal volume is conserved in these figures, but cross-sectional area is not because of radial pattern of motion.

Fig. 2.9. Tectonic evolution of the Alborán region. After Platt and Vissers (1989).
Fig. 2.10. Kinematic indicators in the Gibraltar Arc region. After Frizon de Lamotte et al. (1991).
2.3.4 ROLL-BACK MODEL

According to the roll-back model (Dewey 1980), compression and coeval extension are due to the roll-back of a subduction zone. This model has been applied to several folded belts in the Mediterranean region (Royden 1993) (Fig. 2.11). Based on micro-structural and other kinematic indicators, Frizon de Lamotte et al. (1991) emphasize the westward motion of the Gibraltar Arc. The deformation observed in the Betic and Rif Cordilleras can be explained according to these authors by an E-W migrating subduction zone. The westward migration of this east-dipping subduction zone would explain deformation within the arc. García-Dueñas et al. (1992) support this model to explain frontal compression and coeval extension in the westward migrating Gibraltar Arc.

Field and seismic data provided in this thesis significantly constrain some of the models presented in this chapter.
Fig. 2.11. The retreating subduction zone of the Gibraltar Arc according to the roll-back model. After Royden (1993).
3. STRUCTURAL FRAMEWORK

3.1 INTRODUCTION

The Gibraltar Arc overthrusts and successfully covers the Iberian and Moroccan Mesetas and the passive margins of the African and Iberian plates. The Gibraltar Arc is conveniently split into External and Internal domains. The External domain involves sedimentary allochthonous units derived from the south-Iberian and north-African passive margins. The Internal domain involves various metamorphic units (Andrieux 1971) that are mostly derived from an independent Alborán platform that was separate from both Europe and Africa (Fig. 3.1).

The tectonic map of the Gibraltar Arc (see attached) integrates several maps of other authors (see list of maps), with some subsurface and field data collected for this thesis. The northern part of the Gibraltar Arc, i.e. the Betic Cordillera, is based mainly on the geologic map of Andalusia, "Mapa geológico de Andalucía" (Junta de Andalucia 1985), and the tectonic map of the Gibraltar Arc, "Mapa tectónico del Arco de Gibraltar" (SECEGSA 1990). The southern part of the arc, i.e. the Rif Cordillera, is based on the structural map of the Rif Cordillera, "Carte Structurale de la Chaîne Rifaine" (Suter 1980b). Detailed geological maps published by the "Instituto Geológico y Minero de España" (IGME) and by the Geological Survey of Morocco, as well as several publications cited in the reference list, have also been used to construct this tectonic map. The map incorporates new data in the front portion of the Gibraltar Arc. The legend includes the correlation of tectonic units of the Betic and Rif Cordilleras (see map 1).

Note that Neogene extension associated with the collapse of the Alborán Sea and late inversion and transpressional tectonics severely modify the overall compressional development of the Gibraltar Arc. Equivalent structural units of the Betics and the Rif are presently separated by the Alborán Sea (see Internal domain).
Fig. 3.1. Major structural domains of the Gibraltar Arc.
The following description of the structural units of the Gibraltar Arc proceeds from the simple foreland to the progressively more complex External and Internal domains.

3.2 FORELAND

Because the main topic of this thesis is the relationship of the foreland and the evolution of the adjacent accretionary wedges, this part of the introduction is kept to a minimum. A specific chapter will be dedicated to the Rif and Betic foredeeps.

The Gibraltar Arc is superposed on basement rocks of the Late Paleozoic Hercynian folded belt. The Hercynian orogenic system constituted a continuous folded belt that extended along Pangea (or Pangean suture) from the Marathon-Ouachitas and Appalachians in North America to the Urals in Eastern Europe (Ziegler 1985). The Hercynian folded belt of the Iberian Peninsula is connected to the north with the Hercynian of Bretagne and Montaigne Noire in France and the Bohemian Massif in central Europe. To the south of the Anti-Atlas, the Hercynian structures cross-cut the Mauritanide system (Laville and Pique 1991).

3.2.1 IBERIAN MESETA

Major structures of the Betic Cordillera trend nearly perpendicular to the structures of the Hercynian folded belt (Fig. 3.2). These Paleozoic folds underlie the Guadalquivir Basin (i.e. from the foreland to the hinterland: the South-Portuguese, the Ossa-Morena and the Central-Iberic Zones). The South-Portuguese Zone, which is the front part of the orogen, consists of Devonian and Carboniferous turbidites, the "Culm facies". A pyritic belt running parallel to the main structures is the most characteristic feature of this zone. The Ossa-Morena Zone is characterized by Precambrian metamorphic and granitic rock and Carboniferous siliciclastics. The "Los Pedroches" granitic batholite separates the Ossa-Morena from the Central-Iberic Zone, the most internal zone of the Hercynian orogen. The Central-Iberic Zone is characterized by high-grade metamorphism and abundant granite intrusions (Julivert 1983).
Fig. 3.2. Structural units of the Iberian Hercynian Meseta, north of the Guadalquivir Basin. After Apalategui et al. (1990).
3.2.2 MOROCCAN MESETA

The northern Moroccan Meseta occupies the foreland region of the Western Rif. It consists of Paleozoic rocks deformed during the Hercynian orogeny (Fig. 3.3). Ordovician quartzite or Cambro-Ordovician schist and quartzite constitute the core of NE-SW trending anticlines, which are flanked by Lower and Middle Devonian limestone and Upper-Devonian-Carboniferous synorogenic turbidites (Dinantian and Lower Carboniferous flysch). Post-orogenic Permian sandstones and conglomerates fill small basins that trend parallel to the direction of the Hercynian structures. West of Rabat in the Tiflet area, the direction of the structures in the Hercynian basement is parallel to the WNW-ESE trending Rif Cordillera. This anomalous orientation is due to post-Messinian basement-involved tectonics discordantly superimposed on the Hercynian structure. The growth of the Tiflet Anticline is recorded by Messinian sedimentary wedges (Tilloy 1955b).

The Paleozoic basement is overlain by a half-graben system filled by Triassic red-beds, consisting of sandstone, shale and conglomerate (Fig. 3.4). Lower Cretaceous horizontal marl and sandstone beds unconformably overlie both Triassic and Paleozoic rocks. These overlying deposits might be associated with Neocomian transgression (Gigout 1951, Michard 1976). Cenomanian marly-limestone covers Hercynian paleohighs. Turonian limestone, Senonian marl, Maastrichtian to Lower Eocene phosphates and Lutetian limestone constitute the upper platform succession of the Hercynian Meseta (Fig. 3.4). Locally Neogene basaltic flows overlie this sedimentary succession.

3.3 FOREDEEP

A thick section of Neogene sediments overlies the Hercynian basement of both the Iberian and Moroccan Mesetas. These Neogene basins deepen towards the orogenic belt and constitute the foredeep basins of the Gibraltar Arc. The Guadalquivir Basin in the
Fig. 3.3. Structural map of the Moroccan Meseta. After Pique et al. 1991.
Fig. 3.4. Sedimentary cover of the Moroccan meseta. Modified from Gigout (1951) and Michard (1976).
Betic Cordillera is the northern foredeep. The southern foredeep, linked to the Rif Cordillera, is represented by the southern Rharb, the Mamora and the Zaërs Basins. The Paleozoic basement and its Mesozoic cover dip gently towards the Rif folded belt and extend below the frontal thrust sheets. Unconformably overlying these rocks are Miocene transgressive sandstones derived from the Paleozoic basement. This widespread sandstone unit, present in both Betic and Rif foredeeps, grades upward into a marly section that records the progressive deepening of the basin. Neogene Betic and Rifean foredeep sediments may be subdivided into two major transgressive-regressive facies cycles bounded by major flooding events: an Upper Tortonian-Lower Messinian cycle and an Upper Messinian-Lower Pliocene cycle. These cycles are characterized by deltaic prograding systems which grade into deep-water pelagic and/or turbiditic deposits.

The accretionary wedge loads and partially fills the Betic and Rif foredeeps. Emplacement of this accretionary complex took place mostly during Tortonian time. The upper foredeep succession is coeval with the extensional collapse of the wedge, creating separate accommodation space for the Supra-Nappe units. Some normal faults, common in the distal part of the foredeep, involve the lower foredeep succession and the underlying basement.

The Betic and Rif foredeeps and their relation to the accretionary complex are discussed in greater detail in Chapter 4.

3.4 FRONTAL TECTONO-SEDIMENTARY UNITS.

The frontal part of the Gibraltar Arc is occupied by tectono-sedimentary complexes, consisting of a chaotic mixture of Triassic, Cretaceous, Paleogene and Neogene sediments, overlying the thin sedimentary cover of the Iberian and Moroccan Mesetas. The Prerifaine Nappe of the Rif and the Guadalquivir Allochthon of the Betic Cordillera represent respectively the southern and northern parts of this frontal tectono-
sedimentary complex. The frontal tectono-sedimentary complex of the Gibraltar Arc is an accretionary complex related to the interaction between the Internal and External zones of the Betic and Rif Cordilleras. The Accretionary Zone can be subdivided into an Inner and an Outer accretionary complex. The Inner unit is represented by some of the classical flysch units, and the Outer unit occupies the frontal part of the accretionary zone, being represented by the Guadalquivir allochthonous units in the Betic Cordillera and the Prerifaine Nappe in the Rif Cordillera.

3.5 EXTERNAL DOMAIN

Classically, the external domain of an orogenic belt is represented by non-metamorphic sedimentary successions characterized by thin-skinned tectonics. The External domain of the Betic and Rif Cordilleras is separated into a number of structural units or thrust sheets. Some thrust sheets consist primarily of platform carbonates and other sheets consist mostly of siliciclastic sediments. The different types exhibit widely different styles of decollement tectonics. The External domain of the Betic and Rif Cordillera represents respectively the south-Iberian and north-African passive margin successions incorporated in the folded belt.

Overall, the south-Iberian and north-African passive margins experienced a similar tectonic evolution, although some differences between them need to be emphasized. A number of stages, common to most of the margins of the Tethys, can be distinguished in the Betic and Rif Cordilleras (Figs. 3.5, 3.6 and 3.7).

3.5.1 TRIASSIC RIFTING

Permian and Triassic half-grabens overlie the Hercynian basement of the Iberian and Moroccan Mesetas. They developed in the southeastern margin of Iberia and north and northwestern margins of Africa. These half-grabens were filled with continental red-beds (Permian and Lower Triassic) and shallow-water evaporite and limestone (Middle
Fig. 3.5. Legend of the stratigraphic sections and diagrams used in this chapter
and Upper Triassic). In the Atlas system red-beds also fill branches of the Triassic rifting system (Favre and Stampfli 1992).

In the Betic Cordillera, thin Triassic red-bed successions overlie the Hercynian Meseta to the north of the chain. These red-beds contrast with the thicker evaporitic section of the originally southern part of the External domain. Diabase-type subvolcanic rocks are associated with thick shale-evaporitic sections (Pérez-López and López-Chicano 1989). The facies distribution reveals a southward deepening and thickening. In the Rif the authochthonous Triassic mainly consists of continental red-beds (Salvan 1974), while the allochthonous Triassic is richer in evaporites and has a thicker section. Seismic and well data suggest Triassic deposition in a half-graben setting.

3.5.2 CARBONATE PLATFORM

During Lower Jurassic time a widespread carbonate platform dominated by dolomites was established in the Betic Cordillera. In the Rif, tidal dolomites occur in the Rides Prerifaines (the southernmost exposed Jurassic). Shallow-water algal limestones were deposited in the Ketama region north of the Rif. During Middle Jurassic time (Pliensbachian), the Betic and Rif platforms broke up and drowned (see Fig. 3.2). Pelagic sedimentation, characterized by deep-water limestones and fault-breccias, covered previously deposited shallow-water carbonates (Vera 1983, Favre et al. 1991, Favre and Stampfli 1992). In the Rif, the carbonate platform persisted in the Prerifaine Rides until Bajocian time. This platform represents the prolongation of the Middle Atlastic platform. This carbonate platform probably extended toward the Mesorif, grading to the north into talus deposits and basinal calci-turbidites (Wildi 1983). In the Betic Cordillera, platform carbonates were deposited in two regions: in the north attached to the south-Iberian continental margin (Prebetic platform), and in the south (Internal Subbetic and Penibetic platforms). Both platforms were separated by a deep subsiding basin (García-Hernández et al. 1988). In the Western Betics a northwestern platform attached to the Iberian basement,
the Algarve-Gulf of Cádiz platform persisted until Lower Cretaceous time and a southwestern platform, the Penibetic was probably the prolongation of the Subbetic of the Central region (Martín-Algarra 1987). The carbonate platform units of northern Africa and southern Iberia were both covered during the Early Cretaceous by a thick detritic passive margin sedimentary succession.

3.5.3 PASSIVE MARGIN

In the Rif Cordillera, Toarcian-Bajocian deltaic systems (in the Prerifaine Rides) grade to the north into calci-turbidites and basinal carbonates (Favre et al. 1991). Thick turbiditic deposits, the so-called "ferry-flysch", represent deep sea fans derived from the African Craton (Wildi 1981). Talus breccias and conglomerates (carbonate debris flows) associated with fault scarps may suggest rifting along the north African margin during Pliensbachian-Toarcian time. Deep-water pelagic and hemi-pelagic shales overlie carbonate units. Since Bathonian time, emergence of the Prerifaine Rides caused a sedimentary hiatus and detritic siliciclastics (mainly deltaic) were transported northward to form deep-sea fan sediments (Favre et al. 1991). These deposits are covered by widespread pelagic nodular limestones, referred to in the Western Mediterranean as "ammonitico rosso".

In the Betic Cordillera, basaltic volcanic rocks and pillow lavas are interbedded with radiolarites, radiolarian marl and nodular marly-limestone, typical "ammonitico rosso" facies. The "ammonitico rosso" was deposited in structural highs, while calci-turbidites were sedimented in extensional troughs. This tectonic alignment of blocks suggests the break-up of the platform (Upper Jurassic). Only in the northern platform region (Prebetic zone) did lagoonal shallow-water carbonate deposition persist during Upper Jurassic time.

The Lower Cretaceous (Neocomian) is characterized by its uniform facies in both the Betic and Rif Cordilleras. In the Betic Cordillera, widespread Neocomian pelagic
sedimentation took place throughout the External domain, except in the northern platform (Prebetic), where shallow-water deposition persisted. The most characteristic deposits of this age are rhythmic marl-limestone couplets and thin-bedded greenish-dark shales of Aptian-Albian age. Siliciclastic turbidites interfingerling with pelagic marls were deposited in a growth fault setting (Banks and Warburton 1991). In the Rif there is widespread deposition of deep-water marls and limestones and siliciclastic turbidites that are often referred to as "Apto-Albian flysch".

Deep-water organic rich sediments, represented by bituminous shales, are interbedded with deep-sea fan deposits in the Cenomanian-Turonian boundary in both the Betic and the Rif Cordilleras (Herbin et al. 1986). In the Rides Prerifaines, Turonian dolomitic limestone with pelagic macrofaunas (i.e. Inoceramus, ammonites) occurs in facies similar to those of the Moroccan Meseta and the Atlasic system.

In the Western Rif (Loukkos zone) Cenomanian-Turonian deposits consist of limestone, marl and chert rich in planktonic fauna, silicified calci-turbidites with radiolaria and pyrite, as well as hemipelagic black shales.

In the Betic Cordillera the Upper Cretaceous is characterized by tidal-lagoonal limestone and shallow-water platform carbonates in the north. These grade into Globigerina-bearing salmon-red and white pelagic marl and marly-limestone "scaglia" facies (Vera 1983).

3.5.4 ALLOCHTHONOUS EVAPORITES

Upper Cretaceous pelagic sediments contain allochthonous Triassic blocks of several sizes and sheets up to 2000 meters thick, consisting of gypsum and salt. I think most of the Triassic of the Betic Cordillera and the Rif Cordillera was emplaced as allochthonous evaporites during the passive margin stage in the Cretaceous in a similar way to the allochthonous salt of the Gulf of Mexico (Worrall and Snelson, 1989; Wu et al.
1990). The most striking evidence for this is that the Triassic is systematically embedded within Cretaceous sediments and the Jurassic is always absent.

Extensional basins filled by marls and marly-limestone seem to be related to the emplacement of the allochthonous Triassic in a manner similar to the extensional basins riding piggy-back on the allochthonous salt of the Gulf of Mexico (Wu et al. 1990). Abundant olistostromes of Triassic evaporites are common within the Senonian of the Central Rif (Asebriy 1983, Asebriy et al. 1987).

During Coniacian to Lower Campanian time, sub-CCD pelitic slope sedimentation took place in the southern part of the Rif Cordillera. The Upper Campanian to Lower Maastrichtian section is characterized by abundant slumps, olistostromes, pebbly mudstone, debris flows and calci-turbidites containing reworked Turonian microfossils (Thurow and Kuhnt 1986), suggesting compressional setting. In the Rif, pelagic limestone with gradually increasing siliciclastic content comprises slope provenance sediments.

In the Betic Cordillera, during the Paleogene there was a change in the polarity of the basin. Paleogene turbidites have a high content of platform foraminifera and come from the north, that is from the incipient folded belt, in contrast with the Jurassic and Cretaceous turbidites that were sourced from the cratonic area in a passive margin scenario. The emplacement of the initial allochthonous thrust sheets resulted in increasing subsidence evidenced by the deposition of thick marl successions.

3.6 INTERNAL DOMAIN

The Internal or Alborán domain, which includes the Internal zones of the Betic and Rif Cordilleras (Fig. 3.1), differs stratigraphically and structurally from the External zones. The most notable characteristics that distinguish the Internal from the External zones
include: Alpine-type Triassic carbonates, Early Alpine (Cretaceous-Paleogene) polyphase compressional deformation and HP/LT metamorphism.

The Alborán domain is often visualized as a microplate, initially located between the Eurasian and African plates, which collided with Iberia and Africa (Andrieux et al. 1971). The tectono-metamorphic evolution from HP/LT to LP/HT experienced by the Internal domain records subduction, extension and inversion (de Jong 1991). The collision of this plate with Iberia and Africa in the Late Aquitanian to Burdigalian was accompanied by west-vergent thrusting along the Gibraltar Crustal Thrust (G. C. T.) (Balanyá and García-Dueñas 1987, 1988). Following collision, the allochthonous Alborán domain was thinned and dismembered by extension (García-Dueñas and Martínez-Martínez 1988, García-Dueñas et al. 1992). Today, this allochthonous “terrane” forms part of the Alborán back-arc basin basement. The Internal domain is composed of the Nevado-Filabrides, Alpujarrides-Setebides, Malaguídes-Ghomarides, Dorsale and the Pre-dorsale complexes. The Nevado-Filabrides and Alpujarrides-Setebides are the lowermost structural units, represented mainly by HP/LT metamorphic rocks. The upper, low-grade, structural units are represented by the landward Ghomarides-Malaguídes complex and the more distal Dorsale and Pre-dorsale complexes of the Alborán passive margin.

3.7 STRUCTURE OF THE BETIC AND RIF CORDILLERA

3.7.1. INTRODUCTION

Two generalized regional geological cross-sections were constructed to compare the structure of the Betic and the Rif Cordilleras (Fig. 3.8, 3.9). The sections are based on seismic, well-log and mainly surface data. They provide a broad approximation of the structure of the Gibraltar Arc and help to define the main structural problems.
The cross-sections show great similarities i.e.:

- Foreland-vergent thin-skinned thrusting involving the passive margin succession of the north-African or south-Iberian domains.
- Piggy-back thrusting.
- A main Triassic decollement
- The Frontal Tectono-sedimentary unit
- Extensional (Negative inversion) structures affecting the Internal domain.

The main difference between these folded belts is that the Betic is a carbonate dominated margin, while the Rif is a detritic dominated margin. The volume of allochthonous Triassic evaporites is much larger in the Betic Cordillera than in the Rif. The principal problems of the area concern the role of extension related to the opening of the Alborán Sea, the role of strike-slip faults, the way to accommodate the shortening in the lower part of the passive margin succession, the area of origin of the frontal allochthonous tectono-sedimentary complexes and its relationship with the allochthonous Triassic, and the sequence of thrust emplacement.

3.7.2. RIF CORDILLERA

The description of the cross-section will proceed from foreland to hinterland. A thin Cretaceous section overlies the Paleozoic basement of the Moroccan Meseta. The overlying Neogene succession represents the foredeep of the Rif Cordillera. The Prerifaine Nappe represents the most frontal allochthonous unit of the Rif thrusted onto the foredeep basin. The internal structure of this unit is characterized by thrusts and superimposed normal faults. The Prerifaine Nappe overlies the uppermost part of an antiformal stack formed by Cretaceous-Paleogene foreland-vergent imbricates. A regional decollement separates them from underlying Triassic-Jurassic imbricates characterized by broader anticlines. These crop out in the region classically called Mesorif. Thus the lower
Fig. 3.8. Main structural units of the Gibraltar Arc. Location of the cross-sections of figure 3.9. GCT Gibraltar Crustal Thrust.
Fig. 3.9. Cross-sections through the Betic and Rif Cordillera. Based on data from Suter (1980), García-Hernández et al. 1980, Blankenship (1992) and García-Dueñas et al. (1992).
imbricates are overlain by foreland-vergent, closely-spaced imbricates composed mainly of Cretaceous and, occasionally, Paleogene sediments.

Proceeding northward, Cretaceous imbricates are overlain by Triassic and Jurassic limestones of Alpine-type facies that differ from those encountered in the previously described imbricates. These and all the other units of the Internal domain are derived from the Alborán microplate (Andrieux et al. 1971). These Jurassic limestones constitute a structural unit known as Dorsale calcaire. The structure of the Dorsale consists of steep northward-dipping thrusts reactivated as normal faults. Several structural units consisting of Paleozoic rocks and a thin Permo-Triassic cover are thrusted onto the Alpine Jurassic of the Dorsale. These Internal units have a completely different stratigraphy and tectono-metamorphic history than the units of the External domain. The lowermost unit (The Sebtides) contains metamorphic rocks and mantle peridotites (Beni-Bousera ultramafics). The upper unit (Ghomarides) overlies both the Sebtides and the Dorsale unit. The structure of the Internal domain is characterized by low and high angle extensional detachments that cut or invert ancient thrusts. These extensional structures extend farther north in the Alborán Sea, which is characterized by Neogene half-grabens.

3.7.3. BETIC CORDILLERA

This description will proceed from the Iberian Meseta to the Alborán Sea, that is, from foreland to hinterland.

The Paleozoic of the Iberian Meseta is overlain by a thin Triassic cover and the thick Neogene succession of the Guadalquivir Basin which represents the foredeep of the Betic Cordillera. The Guadalquivir Allochthon is thrust onto the Neogene of the Guadalquivir Basin and represents the frontal-most structural unit in the chain. The Guadalquivir is underlain by Cretaceous and Jurassic domes. Proceeding northward, the structure consists of foreland-vergent imbricates consisting of Jurassic and Cretaceous platform and basinal sediments detached from Triassic shales and evaporites, which
provide the main decollement level. Locally, backthrusting occurs close to the contact with the Internal zones. In the same area, the Jurassic imbricates are cut by the extensional basin of Granada. Alpine-type carbonates of the Dorsale unit and metamorphic rocks of the Malaguides overlie the Jurassic of the External domain. The External-Internal domain boundary is a dextral-transtensional contact. To accommodate and fill the space located below the contact with the Internal zone a duplex involving Triassic and Jurassic sediments is hypothesized. The detached Cretaceous-Paleogene of these hypothetical duplexes could be represented by the Guadalquivir Allochthon. The structure of the Internal domain is characterized by low- and high-angle extensional detachments, superposed on an antiform that consists of three stacked thrust sheets, i.e. from bottom to top: Nevado-Filabrides, Alpujarrides and Ghomarides. High pressure and low temperature metamorphism characterizes this antiform. Omissional contacts with the absence of stratigraphic units are common in this region due to extensional delamination.

For additional details on the stratigraphy and structure of the Betic and Rif Cordillera, see Appendix.
4. FOREDEEP

4.1 INTRODUCTION

Foredeep basins develop in front of fold and thrust belts as a response to loading caused by the emplacement of thrust sheets. The rheology of the underlying basement, the rates of sediment supply versus subsidence, and the role of eustacy are the main factors controlling foredeep development. Traditionally, the filling of the foredeep was represented by two "tecto-facies", a deep-water turbiditic "flysch" sedimentation stage, followed by shallow-water marine and fluvial deposition "molasse". The foredeep succession in the study area was subdivided into several stages that characterize the evolution of the foreland region adjacent to the folded belt (Fig. 4.1). This scheme, simplified and modified from Bally (1989), facilitates the comparison between the two foredeep basins of the Gibraltar Arc (i.e. the Betic and the Rif foredeep). A lower unit includes rift and platform stages that underlie a complex unconformity on top of the platform sequence, the so-called "basal foredeep unconformity"; it marks the inception of the foredeep. The overlying foredeep succession includes three stages: deep-water foredeep, shallow-water foredeep and alluvial foredeep. The initial stage of the foredeep is characterized by longitudinally transported turbidites and/or deep-water pelagic marls (deep-water foredeep). Prograding deltaic complexes from both margins tend to fill the depression (shallow-water foredeep). The final filling of the foredeep is represented by fluvial and alluvial sediments (alluvial foredeep).

Two foredeep basins are in front of the Gibraltar Arc, one linked with the Betic Cordillera and the other with the Rif Cordillera (Fig. 4.2). Table 4.1. outlines the main tectono-stratigraphic events that occur in these foredeep basins.
Fig. 4.1 Foredeep subdivision used in this study. Modified from Bally (1989). Arrows show sediment input direction.
Fig. 4.2. Tectonic map of the Gibraltar Arc. Location of the Betic and Rif foredeeps.
Table 4.1 Major Tectono-stratigraphic events in the Foredeep basins of the Gibraltar Arc.
4.2 BETIC FOREDEEP


4.2.1 STRATIGRAPHY

4.2.1.1. Platform section.

The Platform section includes sediments located underneath the basal foredeep unconformity. These sediments represent the sedimentary cover of the Late Paleozoic Hercynian folded belt and involve a lower rifting event and an upper platform stage (Fig. 4.3).

Rifting stage

In the Central Guadalquivir Basin the Hercynian substratum is mainly composed of Carboniferous shale and slate and granites (IGME 1987). The basal Triassic section consists of a paleosol interval overlain by quartzitic red conglomerate and sandstone grading into red mudstone followed by sandstone, gypsiferous marl and gypsum (IGME 1977). This unit has a typical thickness of about 150 meters.

Platform stage.

In the western and eastern Guadalquivir Basin, Jurassic and Cretaceous sediments directly overlie the Paleozoic basement and its thin Triassic cover. In the onshore Gulf of Cádiz (Western Guadalquivir), the foredeep section overlies a thick Jurassic and Late Triassic evaporite and carbonate section related to the development of the Atlantic passive
Fig. 4.3. Betic Foredeep subdivision with indication of local stratigraphic units.
margin. In the Eastern Guadalquivir Basin, the Jurassic carbonates are continuous with the Iberian Meseta cover. Overlying the Jurassic carbonates are Upper Cretaceous sandy-marl and calcarenite, attributable to the widespread "Utrillas facies". Both the Jurassic and the Cretaceous sections are exposed in the Sierra de Cazorla frontal imbricates and thicken toward the east (IGME 1985).

A widespread Upper Serravallian-Lower Tortonian basal sandstone unit "Arenas basales" covers the basement of the Guadalquivir Basin (Fig. 4.3). This unit, referred to as the "Atlántida Group" (Martínez del Olmo et al. 1984, Suárez-Alba et al. 1989), consists of poorly sorted, sub-angular medium-grained (0.5-0.25 mm) sandstone with quartz and minor metamorphic rock fragments, benthic foraminifera, Globigerinoides, bryozoa, echinoderms and diverse carbonate litho-clasts (Palomares 1990). Detailed studies suggest that the provenance of this sandstone is from the Hercynian basement of the Iberian Meseta. The top of this sandstone unit represents the basal foredeep unconformity and records the initial emplacement of the Betic thrust sheets and the inception of the foredeep.

4.2.1.2 Foredeep stage

The south-dipping basal transgressive sandstone unit is overlain by turbidites. Bidirectional onlap characterizes this foredeep; the onlap is on one hand on the platform sediments and on the other on the accretionary wedge (Fig. 4.3). The initial filling of the foredeep consists of deep-water deposits "deep-water foredeep" that correspond to the Bética and Andalucía Groups (Martínez del Olmo et al. 1984, Suárez-Alba et al. 1989) Bética Group

The lowermost deposits of the foredeep consist of Lower Tortonian-Lower Messinian (N17-N15 Blow) progradational depositional systems separated by an axial turbiditic system, referred to as the Bética Group (Martínez del Olmo et al. 1984, Suárez-
Alba et al. 1989). The progradational deposits are located in the northern and southern margins of the Guadalquivir Basin (Fig.4.3).

Northern prograding system

In the northern margin of the Guadalquivir Basin, backstepping Lower Tortonian deltaic sediments interfinger with marine marl wedges. Deltaic deposits are represented by lower channelized conglomerates fining upwards into sandstones, calcarenites and mudstones with evidence of subaerial exposure. Coquinoid intervals with large oysters are present in the upper part of the section (Santos-García et al. 1991a). In the onshore Gulf of Cádiz area, the "Calcarenitas de Niebla" calcarenitic formation (Civis et al. 1987) unconformably overlies the substratum. This calcarenitic unit, representing backshore-foreshore deposits, consists of basal quartzitic conglomerates with oyster fragments, overlain by marly-sandstones and cross-bedded glauconitic sandstones. The most common fossils present in this unit are brachiopods, ostracods, echinoids and foraminifera, which yield an Upper Tortonian age (Civis et al. 1987). A glauconitic sandstone interval lies in the contact between calcarenites and deep marine shoreface-offshore marls (Castelló et al. 1975, Martínez del Olmo et al. 1984, Sierro 1984, Civis et al. 1987, Galán et al. 1989 a,b).

Southern prograding system

Tortonian blue marls, referred to as "Arcillas de Gibraleón Formation" in the western Guadalquivir Basin (Civis et al. 1987), overlie Early Miocene siliceous pelagic marl ("Albarizas and Moronitas") of the Guadalquivir Allochthon. Along the southern margin of the basin, this marly unit is discontinuously covered by sandstone and calcarenite represented by the "Calcarenitas de Porcuna" and "Arenas de Écija" Formations. These sandstones consist of carbonate-cemented, well-sorted, fine-grained bioclastic and quartzitic sandstone with a micritic matrix. The sandstone composition suggests an important change in paleocurrent direction and provenance (Palomares 1990). These sediments were shed from the orogenic belt, with the carbonate abundance
suggesting an external Betic provenance. To the north, these southward prograding sandstone units interfinger with marls. Notice that these deposits are not related to the flexural process linked to the Guadalquivir Basin but to the evolution of the Guadalquivir Allochthon and therefore they should be considered as "Supra-Nappe" deposits.

Central turbiditic system

The southern and northern marginal depositional systems grade into an axial marly section with sandstone intervals, the so-called "Arenas del Guadalquivir". These consist of well-sorted, fine to very fine-grained polymictic sandstone with fragments of metamorphic rocks, limestone, bioclasts, lignite fragments, glauconite and quartz (Castelló et al. 1975). Parallel bedding, flute casts and groove casts, dish structures, flame structures and mud rip-ups are common. Massive or amalgamated beds are organized in an overall thickening upwards section composed of Bouma Ta-b sequences. The thickness of this sandstone unit is highly variable, ranging between 3-100 meters, reflecting basin paleorelief (Suárez-Alba et al. 1989). Six turbiditic fan systems with channel-overbank, lobe and levee facies have been recognized in this turbiditic deposit and the turbiditic infill migrates from east to west and slightly to the north (Suárez-Alba et al. 1990). Longitudinal turbiditic systems are characteristic of early foredeep stages (Bally 1989). The turbiditic sandstone composition represents an important change with respect to the basal sandstone unit "arenas basales" (Palomares 1990).

Andalucia Group

The Andalucia Group consists of Upper Tortonian-Lower Pliocene (N18-N17 Blow zone) prograding sandstone and carbonate deltaic sediments. The deltaic system progrades from the Guadalquivir Basin margins, interfinger with marine wedges in the central portion of the basin (Fig. 4.3). These deposits overlie Tortonian marls of the "Arcillas de Gibraleón" Formation (Civis et al. 1987). Upper Tortonian-Lower Pliocene
shoreface-offshore marly facies with *Globorotalia humerosa, G. miotumida* and *G. margaritae* cover these deltaic deposits (González-Regalado 1989).

**Northern prograding units**

In the northern Guadalquivir Basin margin, glauconitic sandstones unconformably overlie Tortonian blue marls (Viguier 1974, Sierro 1984, Mayoral 1986 and Galán et al. 1989). Towards the west, this sandstone unit is represented by the Arenas de Huelva (Civis et al. 1987) and Arenas de Bonares Formations (Mayoral and Pendón 1986-87). The lower part of the Arenas de Huelva Formation belongs to the *Globorotalia margaritae* biozone and the middle and upper portions of the section belong to the *Globorotalia puncticulata* biozone (Sierro 1984, Flores 1987). This formation constitutes an upward coarsening deposit, where cross-bedding, parallel lamination, bioturbation, erosional surfaces and storm beds are common. The Arenas de Huelva Formation has been interpreted as shoreface sandstones with platform bars, deposited in an anoxic restricted bay environment. Paleontological studies (González-Regalado 1989, González-Regalado and Ruiz-Muñoz 1989) indicate that the lower part of the unit was deposited under open marine conditions (150-200 meters water depth). The upper part of the section was deposited in a shallow-water (50-100 meters) restricted environment with anoxic conditions. The Arenas de Bonares Formation overlies and partially interfingers with the Arenas de Huelva Formation (Fig. 4.3). It consists of Middle-Upper Pliocene foreshore sandstones characterized by low-angle cross-bedding and abundant ichnofacies (Mayoral and Pendón 1986-87).

**Southern prograding units**

In the southern margin of the Guadalquivir Basin, mixed carbonate-siliciclastic shallow-water deposits unconformably overlie the edge of the Guadalquivir Allochthon and the Supra-Nappe Tortonian marls. The shallow-water succession consists of grainstones with serpulids, bivalva, benthic foraminifera, bryozoa and algae. They
represent a coastal environment in a high-energy ramp setting (Clauss 1991). These mixed deposits are exposed in the following localities: Alcalá de Guadaira, Arcos de la Frontera, Bornos, Osuna, Estepa and Carmona "Calcarenitas de Carmona" Formation (Martínez del Olmo et al. 1984). In the Carmona-Alcalá de Guadaira area mollusca and benthic foraminifera dominate, while in the Arcos de la Frontera-Bornos area bryozoa and echinoderma dominate. In Osuna this formation is represented by Messinian bioclastic sandstone with metric-scale herringbone structures, suggesting tidal-delta deposition. Northwest from Estepa, in the Cerro de los Canterones, this Messinian bioclastic sandstone contains abundant fragments of Triassic shales. The overall stratigraphic unit represents Upper Tortonian-Messinian transgressive deposits constituting an upward deepening succession. Basal deltaic deposits interfingerling with marine wedges are overlain by backshore-foreshore deposits that grade into deeper marine shoreface-offshore facies. The "Arenas de Guadarcazar" is a very mature sandstone that resulted from previously reworked sediments in a beach environment (Palomares 1990).

Messinian progradational units are overlain by widespread Lower Pliocene marl suggesting deepening of the basin. These deposits correspond to the basal part of the Marismas Group (Upper Pliocene-Quaternary N21-N18 Blow of Martínez del Olmo et al. 1984, Suárez-Alba et al. 1989). During the sedimentation of this group, progradational depositional systems occurred only in the northern margin of the basin (Fig.4.3).

Shallow-water foredeep

Most of the deposits of the Marismas Group represent the shallow-water foredeep stage. The Middle Pliocene (G. crassaformis) (Benkhelif 1976) to Pleistocene succession of the Cádiz Bay area consists of shoreface sandstone and lagoonal marls with interbedded washover fan sequences, bounded by soils, root marks and subaerial exposure surfaces (Zazo et al. 1983). Upper Pliocene coquinoïd conglomerates, known as facies "ostionera", are related to Pliocene transgressive episodes.
Alluvial foredeep

In the northern and central sectors of the Guadalquivir Basin, Pleistocene fluvial braided sediments characterized by debris flows, channel and flood plain deposits, unconformably overlie the "Arenas de Huelva" and "Arenas de Bonares" Formations (Romero-Segura and Pendón 1991). These deposits are referred to locally as "Alto nivel aluvial" (Fig. 4.3) (Pendón and Rodríguez-Vidal 1986). The present-day foredeep is dominated by the Guadalquivir longitudinal drainage system. This meandering river with well developed point bars runs parallel to the basin axis. Several terrace levels record incision of the Guadalquivir River (Díaz del Olmo et al. 1986). Transversal secondary rivers and alluvial fans join the main axial system developed along the border of the Hercynian Meseta. Longitudinal drainage systems are characteristic of underfilled foredeeps (Flemings and Jordan 1990, Burbank and Beck 1991).

4.2.2 STRUCTURE AND MECHANICS

Surface data from published maps and wells (IGME 1987) were used to construct two dip sections and one strike section across the Guadalquivir Basin. These cross sections display the relationship between the dip of the basement, the Guadalquivir Allochthon and the overlying Neogene sedimentary succession. The strike section shows the stratigraphy and geometry of the pre-foredeep unit (Fig. 4.4). This strike section displays an arching of the Hercynian basement in the Central Guadalquivir, sinking to the west and east below the Mesozoic succession. In the western Guadalquivir Basin, the basal foredeep unconformity overlies thick Triassic and Jurassic evaporites and carbonates of the Algarve platform. Normal faults offset the basement and part of the Neogene succession in the central portion of the section. Proceeding eastward, the Jurassic and Triassic cover of the Iberian Meseta is involved in NW-vergent imbricates, cropping out in the Sierra de Cazorla area. This imbricated zone coincides with the eastern termination of the foredeep.
Fig. 4.4. Strike section along the Guadalquivir Basin showing...
Basin showing the pre-Neogene units distribution.
Fig. 4.5. Two cross-sections of the Betic foredeep based on well data. Vertical exaggeration 5:1. Well information from IGME (1987).
Dip sections illustrate the relationship between basement flexure, the Guadalquivir Allochthon and the Neogene foredeep (Fig. 4.5). The Hercynian basement, overlain by the Arenas Basales Formation, dips underneath the base of the Guadalquivir Allochthon (Suárez-Alba et al. 1989). Normal faults cut the basement of the Guadalquivir Basin and the overlying Neogene succession (IGME 1977, 1987). Section A-A' images the Triassic-Jurassic carbonate platform underneath the Neogene of the Guadalquivir Basin. This platform is the eastward prolongation of the Algarve-Gulf of Cádiz platform. Section B-B' (Carmona section) is located farther east. In this area, the Jurassic is absent and the basement is shallower and directly overlain by the Neogene "Arenas basales" transgressive sandstones (Fig. 4.5).

The flexural response to loading of the Betic thrusts reveals a low effective elastic thickness of the underlying lithosphere. This low value is due to rifting and thinning of the margin before collision according to Van der Beek and Cloetingh (1992). Changes in the rheology of the crust would explain the eastern termination of the Guadalquivir foredeep. The crust is weaker in the Eastern Betics due to late Oligocene-Early Miocene rifting events that also affected the Valencia Trough and the Alborán region (Cloething et al. 1992). According to these authors, the rigidity of the crust is controlled by extensional events that took place before the inception of the foredeep.

4.3. THE RIF FOREDEEP

Exploration work conducted by the Société Chérifienne des Pétroles (SCP) in the 1950's revealed that the Hercynian basement of the Moroccan Meseta dips gently to the north, underneath the Mamora and Rharb regions, attaining depths in excess of 2400 meters (Fig. 4.6). The Rif foredeep is represented by the Mamora-Zaër-Southern Rharb Basin in the Western Rif and by the so-called "coulors sud-Rifain" in the Central and Eastern Rif. The "coulors sud-Rifain" is an E-W trending depression located north of the
Correlated surface sections

Fig. 4.7

Fig. 4.6. Cross-section of the Betic foredeep. Box shows correlated surface data taken from Wernli (1988). Subsurface data from Lorenz (1972) in Suter (1980).
Fig. 4.7. Stratigraphic section of the foredeep succession in the Mamora area. Modified from Wernli (1988). For location of the section see fig. 4.6.
Middle Atlas, that includes the Guercif and Saïs Basins. South of the Eastern Rif the foredeep sequence directly overlies Mesozoic carbonates of the Middle Atlas. In the easternmost Rif, the Atlas system is directly in contact with the frontal thrust sheets of the folded belt, because the accretionary complex is absent (a situation similar to that of the Eastern Betic Cordillera) (see Fig. 4.2). The foredeep partially onlaps the Nappe Prerifaine, which was also encountered by exploration wells (Lorenz 1972 in Suter 1980a).

4.3.1 STRATIGRAPHY

The stratigraphy of the Rif foredeep is subdivided in the same manner as the Betic as follows:

4.3.1.1 Pre-foredeep sequence:

Rift stage

According to subsurface data in the Rharb Basin (SCP 1955 and Lorenz 1972 in Suter 1980a) and field data in the northern margin of the Meseta (Michard 1976), Triassic shales and sandstones unconformably overlie Paleozoic granite, quartzite and mica-schist of the Hercynian basement. They fill half-grabens related to Triassic rifting (Fig. 4.8).

Platform stage

On the Moroccan Meseta and below the Rharb Basin platform sediments are limited to Upper Cretaceous sediments. Note however that in a westward direction the platform also includes Lower Cretaceous and Jurassic sediments of the Atlantic passive margin. Farther east in the Prerif and in the Saïs Basin, Jurassic carbonates and evaporites overlie the same Paleozoic basement corresponding to the Jurassic platform-halfgraben system of the Middle Atlas and the Rides Prerifaines.

Well data from the Rharb Basin and the Mamora area (Société chérifienne des Pétroles 1951 in ONAREP internal reports) indicate that a widespread Upper Cretaceous
detritic unit overlies the basement or its Triassic cover. The Cretaceous section starts with Cenomanian-Turonian quartz-rich conglomerate, coarse sandstone and sandy-marl. Locally, breccias with Paleozoic elements and red matrix overlie the basement. This basal section is overlain by marls and limestones with white dolomitic and oolitic limestones, green marls and marly limestones with lignite remnants. Turonian marls are exposed in Oued Mellah, southeast of Rabat (Gigout 1951, Michard 1976). The upper part of the section consists of Upper Maastrichtian (?) sandy-limestone and marl with phosphate nodules, black silex and brachiopod debris (Tilloy 1955b).

North of the Rides Prerifaines, the platform sequence also includes Aquitanian and Burdigalian sandy-limestone and marl, unconformably overlying Jurassic and Cretaceous sediments (Michard 1976). The top of the platform succession is characterized by sandy-limestone bearing planktonic foraminifera neritic facies.

Messinian basal coquinoid conglomerate and bioclastic yellow sandstone, sandy-limestone and marl, referred to as "Molasse de base" (Fig. 4.7, 4.8) (Feinberg and Lorenz 1970), unconformably overlie Paleozoic, or locally Triassic, rocks of the northern flank of the Moroccan Meseta (Wernli 1977). In the Mamora region, this unit has a constant thickness throughout the area. The sandstone is rich in Pecten, bryozoa, echinoderms (Clypeaster), ostracods, molluscs and *Heterostegina sp.* The presence of mammal teeth indicates continental proximity (Feinberg 1986). The upper part of the section is highly bioturbated. The sandstone unit is topped by a coral-bearing unit referred to as "banc a coraux" (Chevalier 1962). Pecten, brachiopods, gastropods, brachiopods, solitary corals and pebbles have a phosphatic coating and comprise a hard-ground. In the Khemisset area, unconformably on top of the Triassic there are sandy matrix conglomerates with Paleozoic pebbles. These grade upwards into marl with occasionally interbedded conglomerate. This stratigraphic unit, known as the Taguelmane Formation, is unconformably
Tajemout" unconformity) overlain by Pliocene sandy and conglomeratic limestone and marl (Morel et al. 1983).

4.3.1.2 Foreddeep succession

The fordeep succession of the Rif consists of a deep-water pelagic facies overlain by shallow-water (littoral) deposits and alluvial-fluvial deposits (ONAREP int. reports 1991).

Deep-water stage

The basal fordeep unconformity coincides with the top of the condensed interval at the top of the Coral unit. The "Banc a Coraux" is overlain by Messinian to Lower Pliocene pelagic blue marls "Marnes de Salé" with *G. dutertrei, G. margaritae and G. crassaformis* (Wernli 1977,1979) (Fig. 4.7, 4.8). The base of the unit close to the contact with the Coral level has a higher glauconite content. The planktonic/benthic foraminiferal ratio is 4/1 to 6/1 (Wernli 1977). Throughout the region, there is a continuous transition from Miocene to Pliocene pelagic sediments. In the Mamora and Zaër (El-Hanech area), the basal marl is Upper Tortonian in age. These pelagic deposits record deepening of the basin related to the inception of the fordeep and represent the "deep-water fordeep".

Shallow-water stage

In the Fès and Saïs area (Central Rif) the Pliocene section commences with cross-beded yellow sandstone ("sables fauves") reworked from Messinian sandstone. This siliciclastic section is succeeded by 50 meters of lacustrine white limestone marl and lime conglomerate malacologically dated as Lower Pliocene (Jodot 1955). Lacustrine deposits are overlain by travertines (Tilloy 1955b).

Upper Pliocene glauconitic grey sandy-marl overlies the pelagic marly section. A conglomeratic and bioclastic carbonate interval, the so-called "dalle moghrebiene", defines the base of the "Moghrebian transgression" and the Moghrebian stage of Choubert (1965). The "dalle moghrebiene" consists of well-rounded biotritic and coquinoïd glauconite rich coarse sandstone with cross-bedding. The "dalle moghrebiene" contains fauna of
*Lithotamnium*, and benthic and planktonic foraminifera belonging to the *G. crassaformis* biozone (Wernli 1979) (Fig. 4.7). In the Zemours area (northern border of the Moroccan Meseta) Upper Pliocene and Pleistocene sandstones, silts and limestones suggest a coastal barrier island / lagoon system developed in the margin of the Moroccan Meseta (Cirac 1978). These deposits represent the "shallow-water foredeep" stage.

Alluvial foredeep

Brown quartzitic sandstones and cross-bedded conglomerates overlying the "dalle moghrebiene" represent the initial fluvial filling of the foredeep in the Zaër-Mamora area (Fig. 4.8). Widespread deposits of Villafranchian fluvial and alluvial conglomerate and sandstone crop out on the contact with the uplifted Meseta in the south and with the Prerifaine Nappe in the east and north. Villafranchian deposits are covered by aeolian sandstones associated with the "Fouaratan transgression", manifested by the landward shift of the nearshore dunes belt. The coquinoide and sandy-limestone ("grès de Rabat" or "Grande dune") represent dunes backstepping onto coastal plain red siltstone, sandstone and alluvial conglomerate (Biberson 1970, Stearns 1978) (Fig. 4.8). Shallow-water marsh-type pisolithic sandstone developed on the landward side of dune ridges, filling the depressions. In the Fouararat area, consolidated dunes are overlain by red shales and sandstones " formation rouge de la Mamora" or " marnes de Témara" (Tilloy 1955b). Recent aeolian sandstone "sables dunaires recents" constitute a belt of dunes parallel to the shoreline "grès de Mamora". Due to the barrier effect caused by the aeolian ridges, shallow-water lakes or merjas, separating the Rharb Basin from the coastal area, developed on the landward side of the dunes.

In the Mamora area, downcutting erosional surfaces affect Plio-Pleistocene sediments. The drainage system of the rivers in the southern border of the Rharb Basin shows a straight pattern with flow direction from south to north, that is, from the uplifted
foreland into the subsiding foredeep. Some of these fluvial systems capture the drainage system of the Sebou and the Ouerha Rivers.

4.3.3 STRUCTURE

The typically northeastwardly-dipping Paleozoic basement and the overlying Neogene of the Rif foredeep are disrupted by normal and occasionally by strike-slip faults. Basement-involved thrusting in the northern border of the Moroccan Meseta occurred after the inception and initial development of the foredeep. E-W normal faults affect the Hercynian basement and the Neogene sequence, except the Villafranchian in the northern border of the Moroccan Meseta (Wernli 1977). Normal faults (Wernli 1988, Feinberg 1986) affecting the basement of the foredeep are often interpreted as flexural extension (Bradley and Kidd 1991), common to the distal part of the foredeep.

On the southeastern margin of the Rharb Basin, i.e. the region of Sidi Fili, the Neogene foredeep is cut by the Rides Prerifaines (Fig. 4.9). The inversion of the underlying Mesozoic system, characterized by a thick Jurassic section, postdates the inception and early development of the foredeep (Zizi, pers. comm.).

Farther south in the region of Tiflet, a tight E-W trending anticline cored by Paleozoic rocks is overlain by Messinian and Pliocene sediments that record anticlinal growth. In the hinge zone, Villafranchian conglomerates directly overlie the Hercynian basement, post-dating the growth of the structure. The Taguelmane Formation is coeval with the suerrection of the Meseta (Morel et al. 1983).

4.4. CONTRASTING THE BETIC AND RIF FOREDEEPS.

The Betic and Rif foredeeps present a similar structural evolution. Table 4.2 compares the Betic and Rif foredeeps. The inception of the foredeep during Tortonian time is coeval with the emplacement of the frontal accretionary complex of the Gibraltar Arc. The eastern end of the foredeep, in the Betics and the Rif, coincides with the
termination of the accretionary complex. The foredeep developed mainly during Tortonian-Messinian time. A widespread basal transgressive sandstone unit, onlapped by deep-water facies, marks the inception of the foredeep in the Betic and Rif. In the Rif Cordillera, the initial deep-water stage is represented by pelagic marls (marnes de Salé), while the Betic Cordillera is characterized by turbidites. Another major difference is that the Neogene section is much thicker in the Betic than in the Rif foredeep.

Fig. 4.9. The Rif foredeep disrupted by the Rides Prerifaines.
<table>
<thead>
<tr>
<th>Foredeep characteristics</th>
<th>Betic Foredeep</th>
<th>Rif Foredeep</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maximum thickness</td>
<td>3800 m</td>
<td>1900 m</td>
</tr>
<tr>
<td>Drainage system</td>
<td>Longitudinal</td>
<td>Transversal</td>
</tr>
<tr>
<td>Alluvial stage</td>
<td>&quot;Alto Nivel Aluvial&quot; (Pleistocene)</td>
<td>Sables de Rabat &amp; Sables de Mamora (Pleistocene)</td>
</tr>
<tr>
<td>Shallow-water stage</td>
<td>&quot;Marismas group&quot; (Middle-Upper Pliocene)</td>
<td>Sables glauconieux (Upper Pliocene)</td>
</tr>
<tr>
<td>Deep-water stage</td>
<td>Turbidites (Arenas del Guadalquivir) (Tortonian-Messinian)</td>
<td>Pelagic marls &quot;Marnes de Sale&quot; (Tortonian-Middle Pliocene)</td>
</tr>
<tr>
<td>Basal Foredeep Unconformity</td>
<td>Tortonian</td>
<td>Tortonian</td>
</tr>
<tr>
<td>Platform stage</td>
<td>Arenas basales (Upper Serravallian-Lower Tortonian)</td>
<td>&quot;Molasse de base&quot; (Tortonian) Cretaceous</td>
</tr>
<tr>
<td>Rift stage</td>
<td>Triassic red-beds</td>
<td>Triassic red-beds</td>
</tr>
</tbody>
</table>

Table 4.2. Similarities and differences between the Betic and Rif foredeeps.
5. THE ACCRETIONARY ZONE

The frontal units of the Gibraltar Arc form an accretionary zone related to the westward migration of the Alborán domain. According to the time of accretion and position, this zone can be split into an outer and an inner accretionary complex. The boundary between both units is not well defined but approximately coincides with the external limit of the classical flysch units (Fig. 5.1). The inner accretionary complex is not the topic of this study and therefore will be only discussed later in more general terms. It is convenient to describe these zones separately since traditionally they have been treated as separate zones.

5.1. THE PRERIFAINE NAPPE

The frontal units of the Betic and Rif Cordilleras (i.e. Frontal Tectono-sedimentary Complex) consist of highly deformed Triassic, Cretaceous, Paleogene and Neogene strata, that were detached from their original base and thrust over the Mesozoic to Lower Miocene of the foreland. The unit is known as the "Guadalquivir allochthonous units" in the Betic Cordillera and is equivalent to the Prerifaine Nappe of the Moroccan Rif (Fig. 5.1) (Table 5.1). Both the Guadalquivir Allochthon and the Prerifaine Nappe constitute a wide belt, extending to the "Horseshoe" or "Fer du Cheval" east from Gorringe Bank in the Atlantic Ocean (Lajat et al. 1975) (Fig. 5.2). The Guadalquivir Allochthon incorporates voluminous Triassic shales and evaporites, as well as marls, carbonates and siliciclastics. In contrast, the Prerifaine Nappe includes only minor amounts of Triassic evaporites, but mainly younger detritic turbidites or shales.

The overall type of deformation suggests an accretionary prism involving deep-water sediments, related to the westward motion of the Alborán domain over the attenuated Iberian and African passive margins. The timing of deformation and the age of
Fig. 5.1. Tectonic map of the Gibraltar Arc. Location of the Frontal Accretionary Wedge.
Table 5.1. Major Tectono-stratigraphic events
The text refers to stratigraphic events in the Frontal part of the Gibraltar Arc.
Fig. 5.2 Prolongation of the allochthonous units of the Betic and Rif Cordillera into the Atlantic Ocean. After Lajat et al. (1975).
the sediments involved suggest an accretionary progression towards the external portion of the arc. The outer accretionary complex is the most external unit of the Gibraltar Arc (i.e. the Guadalquivir Allochthon of the Betic Cordillera and its equivalent of the Rif). Deformation within the complex was previously explained by several phases of deformation and by gravitational tectonics (Vidal 1977, Feinberg 1986). In contrast, a continuous accretion model, similar to current models of more conventional accretionary complexes (e.g. Dickinson and Seely 1979, von Huene 1986), is presented here. On the basis of field data and seismic data, a block diagram of the accretionary complex has been drawn (Fig. 5.3). Note that the accretionary wedge is presumably underlain by a normal to transitional continental crust that is dipping towards the Mediterranean (A-type subduction), unlike the more conventional accretionary wedges that are related to subducting oceanic lithosphere (B-type subduction).

5.1.1. THE PRERIFAINNE NAPPE: PREVIOUS PUBLICATIONS.

5.1.1.1 Initial exploration

The frontal part of the Rif Cordillera consists of allochthonous units involving a mixture of ductile sediments. This structural unit was initially recognized by Daguin (1927) as the "Nappe Trias-Nummulitique". Later, the unit was called "Nappes prérifaines" by Termier (1936). During the 1950s, following exploration work done by the Société Chérifiene des Pétroles (SCP) (Tilloy 1955a,b,c,d), the structure of the Prerifaine Nappe was found to consist of marls inter-mixed with Triassic shales and evaporites known as "complexe salifère". Cretaceous shale is the dominant lithology of the Nappe. The SCP interpreted the Prerifaine Nappe as a "Complexe crétacée" or "Nappe crétacée" due to the presence of Cretaceous foraminifera in shales sandwiched within Miocene sediments. The first available seismic reflection profiles suggested the presence of a detachment at the base of the Prerifaine Nappe. The allochthony of these units led to the
name "Prerifaine Nappe" (Bruderer and Lévy 1954). For the first time, the age of emplacement of the Nappe was defined as Upper Vindobonian (Tortonian). The leading edge of the Nappe was traced by Bruderer and Lévy (1954). Several wells (SCP in ONAREP internal reports) found intermixed Triassic lithologies within the Nappe. Because of its heterogeneous lithology, the term Complexe Prerifain (Leblanc 1977) or "Tecto-sedimentary complex" (Feinberg 1986) was suggested for this geological unit.

Feinberg (1986) p. 74 states:

"L'emploi du terme Complexe Prerifain est préféré ici à celui d'olistostrome à cause de l'ampleur exceptionnelle et de la variabilité des phénomènes sédimentaires synchro-nappes reconnus dans le Prerif". (The term Prerifaine Complex is preferred to the term olistostrome because of the huge dimensions of the unit and diversity of sedimentary processes that take place during the emplacement of the Nappe).

5.1.1.2 Models of the Prerifaine Nappe

Several models have been proposed for the structure and emplacement of the Prerifaine Nappe. The allochthony of the Prerifaine Nappe was initially proposed by Bruderer and Lévy (1954) and Durand-Delga in Suter (1965). Vidal (1977) suggested an authochthonous origin for the Prerifaine Nappe, involving a "melange of authochthonous olistoliths" which he called the "Mélange Prerifain". According to Leblanc (1977), the Prerifaine Nappe was emplaced by gravity thrusts. Feinberg (1986) supports a model based on thrusting followed by gravitational emplacement (Fig. 5.4). According to this model, the emplacement of the Nappe Prerifaine starts in the Middle-Upper Serravallian (N14 of Blow) and ends during the Lower Tortonian (early Globorotalia acostaensis zone, base of the zone N16 of Blow). Thus, planktonic faunas belonging to biozones of G. acostaensis and G. menardi are present within the Prerifaine Nappe. A Pliocene episode of gravity gliding occurs afterwards, in the southwestern Rharb Basin. This phase of
Fig. 5.4. Polphasic model of emplacement of the Prerifaine Nappe. After Feinberg (1986).
gravity gliding "phase de glissement" was identified only in the southwestern portion of the Rharb Basin resulting in the "Unite Rharbienne" (Feinberg 1986). Newly available seismic data in the region add valuable information to support a more refined model of the Prerifaine Nappe.

5.1.2. RHARB BASIN AND NEIGHBORING AREAS.

5.1.2.1 Introduction

The topographic Rharb Basin overlaps the western front of the Rif Cordillera and its foreland. It is bounded to the east and north by the frontal ranges of the Rif Cordillera and to the west by the Atlantic coast. The southern limit is the Paleozoic Moroccan Meseta (Fig. 5.5). The surface expression of the Rharb Basin is a fluvial-alluvial coastal plain drained by the Sebou River. The Rharb may be considered a peneplain in the sense of England and Molnar (1990), that is, a plain close to sea level that has reached geodynamic equilibrium. Subsurface data presented in this work display the structure of the Neogene sedimentary succession of the Rharb Basin and the underlying Prerifaine Nappe. In the following section I will illustrate the general stratigraphy and the main structural features of the Rharb Basin based on wells and seismic sections.

5.1.2.2 Stratigraphy.

In this section I will describe the stratigraphy of the Prerifaine Nappe and discuss the subsurface stratigraphy of the Rharb Basin. The stratigraphy of the Nouriat, Souk el Arba, Rhabet Jebila-Lalla Zhara and Arbaoua areas will be discussed separately. The stratigraphy of the Rharb Basin is subdivided into: Infra-Nappe, Nappe, Supra-Nappe. The Infra-Nappe succession was already discussed in the Rif foredeep chapter (section 4.3).
Fig. 5.5. Rharb Basin location map.
Prerifaine Nappe stratigraphy

The stratigraphy for the Prerifaine Nappe is obscured by complex compressional and superposed gravity tectonics. Resedimentation and reworked faunas are common and make stratigraphic identification difficult. The presence of reworked Cretaceous faunas (i.e. ammonites) together with Triassic and Miocene sediments leads to difficult biostratigraphic problems (Feinberg 1986). The Prerifaine Nappe consists of Triassic to Miocene sediments that are bounded by stratigraphic and/or tectonic contacts. Thickness estimates are not easily made. The stratigraphy of the Prerifaine Nappe is based on very few outcrops located in the areas surrounding the Rharb Basin and on exploration wells located in that basin. The following description of the stratigraphy is based on surface data obtained by the SCP (Tilloy 1955a, b, c, d) (Feinberg 1986) and information from exploration wells that penetrated the Nappe (Fig. 5.6 and 5.7). The paleontology is derived from Feinberg (1986) and Wernli (1988). When possible the Neogene planktonic biozonation of Wernli (1988) was used. Even though there is not a discernable stratigraphic order because of imbrication and gravitational tectonics, the sediments involved in the Nappe proceed from older to younger. Table 5.2 describes the lithology and fossil content of the Prerifaine Nappe.

Supra-Nappe stratigraphy based on wells

The Supra-Nappe subsurface stratigraphy of the Rharb Basin has been described from selected wells located on panels 1 and 2. There is no agreement on the Neogene biostratigraphy of the Rharb. Table 5.3 shows the suggested or preferred correlation between the biostratigraphy of Feinberg (1986) and Wernli (1988) mainly based on surface exposures and subsurface data by SCP (ONAREP internal reports).

The Neogene biostratigraphy of the Rharb and Rif areas, according to Wernli (1988), is outlined in Fig. 5.8. The stratigraphy of the Rharb Basin based on selected wells, (Fig. 5.9) is shown in Fig. 5.10.
<table>
<thead>
<tr>
<th><strong>CARBONATES</strong></th>
<th><strong>SILICICLASTICS</strong></th>
</tr>
</thead>
<tbody>
<tr>
<td>Limestone</td>
<td>Marl</td>
</tr>
<tr>
<td>Dolomite</td>
<td>Sandy-marl</td>
</tr>
<tr>
<td>Sandy-limestone</td>
<td>Cross-bedded Sandstone</td>
</tr>
<tr>
<td>Shaly-limestone</td>
<td>Siltstone</td>
</tr>
<tr>
<td>Shelly-limestone</td>
<td>Shale</td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td>Turbidites (Flysch)</td>
<td>Marly-limestone</td>
</tr>
<tr>
<td></td>
<td>Conglomerate</td>
</tr>
<tr>
<td></td>
<td>Breccias</td>
</tr>
<tr>
<td></td>
<td>Coarse Sandstone</td>
</tr>
<tr>
<td></td>
<td>Sandstone</td>
</tr>
<tr>
<td>Organic matter/ Coal</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Evaporites</td>
</tr>
<tr>
<td></td>
<td>Salt</td>
</tr>
<tr>
<td></td>
<td>Volcanic Ophites</td>
</tr>
<tr>
<td></td>
<td>Metamorphics</td>
</tr>
<tr>
<td></td>
<td>Granite</td>
</tr>
<tr>
<td></td>
<td>Pillow lavas</td>
</tr>
</tbody>
</table>

Fig. 5.6. Legend of stratigraphic sections and diagrams used in this chapter.
Fig. 5.7. Synthetic stratigraphy of the Prerifaine Nappe. Thicknesses are approximate restored values of intensely deformed layers. Based on the descriptions of Tilloy (1955) and Feinberg (1986)
<table>
<thead>
<tr>
<th>AGE</th>
<th>LITHOLOGY</th>
<th>FORAMINIFERA</th>
<th>THICKNESS</th>
<th>BOUNDARIES</th>
<th>COMMENTS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Middle Miocene</td>
<td>Interbedded sandy-marl and sandstone with occasional limestone beds.</td>
<td><em>Globorotalia mioz ea nilensis</em> Orbulina suturalis</td>
<td>200-300 m</td>
<td>Transitional</td>
<td></td>
</tr>
<tr>
<td>Lower Miocene</td>
<td>Interbedded sandstone and marl with occasional marly-limestone levels.</td>
<td><em>G. klugeri</em>, <em>G. primordius</em>, <em>G. trilobus</em></td>
<td>100-300 m</td>
<td>Transitional</td>
<td></td>
</tr>
<tr>
<td>Upper Oligocene</td>
<td>Interbedded marl and turbiditic sandstone with occasional breccia</td>
<td><em>G. angulatetrialis</em></td>
<td>150-200 m</td>
<td>Transitional</td>
<td>Paleontologic problems to define the Oligo-Miocene boundary.</td>
</tr>
<tr>
<td>Lower Oligocene</td>
<td>intervals and limestone beds.</td>
<td><em>Lepidocyclina sp.</em>, <em>Amphistegina sp</em></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper Eocene</td>
<td>Interbedded limestone and sandstone</td>
<td><em>G. seminivoluta</em>, <em>G. coocanensis</em></td>
<td>400 m</td>
<td>Transitional</td>
<td></td>
</tr>
<tr>
<td>Middle Eocene</td>
<td>Interbedded marly-siltstone and sandstone with conglomerates</td>
<td><em>Hankenina aragonensis</em></td>
<td>450-500 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>and olistostromes</td>
<td><em>Globigerinatheka subconglobata subconglobata</em></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td><em>Globorotalia lehneri</em></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower Eocene</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper Paleocene</td>
<td>White marl and grey marly-limestone with nodular siltex.</td>
<td><em>G. caucasica</em>, <em>G. palmerae</em>, <em>G. aragonensis</em></td>
<td>200-300 m</td>
<td>Transitional</td>
<td>These facies are referred to locally as &quot;marines blanches a silex&quot; or &quot;le Numulite&quot; due to the abundance of Numulites</td>
</tr>
<tr>
<td>Lower Paleocene</td>
<td></td>
<td><em>G. pseudomemardii</em>, <em>G. velascoensis</em></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper Cretaceous</td>
<td>Magnesium-bearing green marly-shale with planktonic foraminifera</td>
<td><em>G. edita</em>, <em>G. pseudobulloides</em>, <em>G. compressa</em></td>
<td>300-300 m</td>
<td></td>
<td>Triassic evaporites with blocks of sedimentary, volcanic and</td>
</tr>
<tr>
<td></td>
<td></td>
<td><em>G. elongata</em>, <em>G. triloculinoides</em>, <em>G. daubieregensi</em> b*</td>
<td></td>
<td></td>
<td>metamorphic rocks appear to be</td>
</tr>
<tr>
<td></td>
<td></td>
<td><em>G. trinitatis</em></td>
<td></td>
<td></td>
<td>interbedded with Cretaceous marls</td>
</tr>
<tr>
<td>Lower Cretaceous</td>
<td>Albion calcic-turbidites consisting of interbedded marl and marly-limestone</td>
<td><em>Trinitata scotti</em>, <em>Plummerina hankeninoides</em></td>
<td>300-300 m</td>
<td></td>
<td>Tectonic or diapiric</td>
</tr>
<tr>
<td></td>
<td>Ammonite-bearing Neocomian gray marl and white marly limestone</td>
<td><em>Globotruncana sp.</em></td>
<td></td>
<td></td>
<td>Locally blocks of Paleozoic gneiss micaschist or organic rich black</td>
</tr>
<tr>
<td>Triassic</td>
<td>Varicolored and often purple marl with interbedded salt and gypsum.</td>
<td></td>
<td></td>
<td></td>
<td>schist are intermixed with Triassic</td>
</tr>
<tr>
<td></td>
<td>Occasional diabase and pillow-lavas</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 5.2. Stratigraphy of the Prerifaine Nappe. After Tilloy (1955) and Feinberg (1986).
Fig. 5.8. Planktonic biostratigraphy of the Neogene post-Nappe of northern Morocco. After Wernli (1988).
Fig. 5.9. Seismic and well data from the Rharb Basin presented in this work.
Fig. 5.10. Stratigraphic correlation of some selected wells of the Rharb Basin. Data from Wernli (1988), Feinberg (1986) and unpublished ONAREP reports. For location of the wells see Fig. 5.9. * Messinian Infra-Nappe.
The presence of restricted anoxic environments that are poor in planktonic faunas, resedimentation and tectonic complications hinder the establishment of a generalized biostratigraphy. The littoral faunas of the peripheral regions of the Rharb Basin are difficult to correlate with the pelagic faunas of the center of the basin. A detailed description of well CGD-5 (Caïd el Gueddari) (SCP 1969) located in the Central Rharb is presented as an example of biostratigraphic zonation (Fig. 5.11).

SCP (Tchret des Tombes) Feinberg 1986 and Wernli 1988

1967)

Zone VI Littoral fauna *G. puncticulata, crassaformis and inflata.*
Zone V G. primitiva+marg. *G. margaritae zone*
Zone IV G. dalii *G. dutertrei zone*
Zone III orbulina *G. menardi zone*
Zone II Menardi-
  Saphoe-Nephentes.

Zone I Littoral fauna *G. menardi zone*

Table 5.3. Correlation between the biostratigraphic zonation of the SCP 1967 (ONAREP internal reports) and the more recent of Feinberg (1986) and Wernli (1988).

5.1.2.3 Structure of the Supra-Nappe Rharb Basin

The Rharb Basin is a composite of extensional basins, referred to as satellite basins. The orientation of the main faults varies from one area to another. In the southern part of the Rharb Basin compressional structures associated with the front of the Prerifaine Nappe occur (Fig. 5.12). The Sebou coastal area consists of NE-SW trending anticlines and synclines related to SW-vergent imbricates of the Prerifaine Nappe (see
CGD-5 Well Caïd el Guedari

PLEISTOCENE

Pliocene

Upper

G. margaritae

Lower

G. dumerilii

Messinian

Upper Miocene

G. mutard

Tortonian

Vulvulina pannatula, Gleboceras menardi
Cyrtammina cancellata, Orbulina sp.
G. newtoni and G. menardi-saphoe

Pyrita-cemented glauconitic sandstone
Epistilobium sp., Ammonia sp. and Neovolcon incisum
Glauconite, conules and ostracoda
Dorothy gibbosae
Sidaucia deboris, pyrite and glauconite
Testudinaria depressa, Tritaxia sp.
G. margaritae, G. premargaritae, G. dali
Bolvina Scalprata var. mionconica
Martiniellia communis, Test. depressa
Rheophax sp., Saccammina sp.

Anomalina trinnotatus, Testudinaria sp.
Evanbergina dinapoli, Bathysiphon sp.,
Haplophragmoides sp., Martiniellia communis
Repoxa-Saccammina, Martiniellia communis
Bathysiphon sp.

Verneuilina sp., Rizammina sp.
Gyrtexia longispica var., Gyrtexia bavirgata
Uvigerina subpepergrina, Sphaerodinia dahiscens
Pyrita remnants

500 m

Fig. 5.11. Stratigraphic section of well CGD-5 Caïd el Guedari
(SCP 1969 in ONAREP internal reports).
Fig. 5.12. Structural map of the Rharb Basin. Modified from ONAREP (1991).
Panels 1 and 2). The southern area is occupied by E-W and N-S trending troughs associated with extensional faults superimposed on top of the Nappe. The frontal part of the Nappe is occupied by NE-SW oriented anticlines and synclines that represent the SW-vergent frontal imbricates of the Prerifaine Nappe. Three E-W, N-S and NE-SW trending troughs are related to extensional basins bounded by low-angle listric normal faults. Depocenters in excess of 4500 meters occur southwest from Mechra bel Ksiri. (ONAREP internal report 1991, Dakki 1992). Table 5.3 shows the major tectono-stratigraphic events in the Rharb Basin.

5.1.2.4 Rharb Basin regional sections

The structure of the Rharb Basin is best shown by a series of composite regional sections. The best seismic profiles were selected and spliced together to construct line drawings. Panel 1 shows four regional dip sections and Panel 2 shows eight strike sections. Dip sections are roughly oriented NE-SW, trending nearly perpendicular to the leading edge of the Prerifaine Nappe. Strike sections are oriented perpendicular to the transport direction of the Nappe and parallel to its leading edge. Panels 1 and 2 display five northeastward trending sections in the Lalla Yto area. Three E-W oriented sections, extending from the Nouriat region in the east to the Sebou region in the west; four NE-SW trending sections; two in Lalla Yto in the East, one through the central Rharb extending from Rhabet Jebila in the north to Lalla Yto in the south and one along the Sebou coastal area. On the basis of seismic character and structural significance, the seismic units present in the transects can be subdivided into a number of structural units, from bottom to top of section:

2. Infra-Nappe = Mesozoic and Lower Miocene sedimentary cover of the basement, here also referred to as Pre-foredeep platform stage.
Table 5.4. Major Tectono-stratigraphic events in the Rharb Basin.
4. Supra-Nappe = Upper Miocene and Plio-Pleistocene siliciclastics.

In the following these units will be referred to in the text as: Basement, Infra-Nappe, Nappe and Supra-Nappe.

**Structural features**

To simplify the description of the regional sections, figure 5.13 illustrates a general cross section across the Prerifaine Nappe with the most common structural features and the relationship to each other.

**Dip sections-General remarks (see Panel 1).**

The most characteristic features shown on the dip sections are:

1. The wedge-like character of the Prerifaine Nappe.
2. The northward dip of the basement and the Infra-Nappe succession underneath the Prerifaine Nappe.
4. Extensional faults offsetting the top of the Nappe.
5. Supra-Nappe clinoformal patterns related to southwestward progradation.

**Regional section R1.**

This N-S and NE-SW regional section extends from Rhabet Jebila to Lalla Yto. In the northern portion of the transect a series of southward-dipping growth faults offset the top of the Prerifaine Nappe. The region of Rhabet Jebila-Lalla Zhara will be discussed in detail in section 5.1.4.6. The basal decollement is located at 2 seconds recording time. Thrusts connected with this basal detachment are present within the Nappe. The maximum thickness (2.5 sec) (1800 meters) of the Supra-Nappe Neogene is attained in the central part of the section. The southern part of the section (Lalla Yto region) is represented by lateral ramps merging into the basal decollement of the Prerifaine Nappe. Antithetic and synthetic normal faults also offset the Mio-Pliocene boundary. Normal faulting is younger
Fig. 5.13 Regional transect connecting onshore and offshore seismic data. Planktonic biostratigraphy is based on onshore wells.
Terminology of Structural Features

LOCATION MAP

The most common structural features and the terminology used are indicated.
in this region than in the northern Rharb, where the Mio-Pliocene boundary is not offset.

**Regional section R3.**

This northeastward trending section follows the Sebou River coastal plain and shows the wedge-shaped Prerifaine Nappe. In the southeastern portion of the section, gently northeastward-dipping Infra-Nappe reflectors unconformably overlie the Hercynian basement. The Supra-Nappe prograding units onlap directly the basal foredeep unconformity. Well EM-3, located in the foredeep region just in front of the Nappe, penetrated the whole sedimentary succession, encountering Cretaceous and Tortonian-Messinian Infra-Nappe and Plio-Pleistocene Supra-Nappe. North of the leading edge of the wedge, the Infra-Nappe unit is characterized by a basal zone of imbricated layered reflectors. The structure of the Prerifaine Nappe itself is characterized by northeastward-dipping thrust faults. Extensional faults are superimposed on the Nappe. Supra-Nappe units show a southward prograding clinoformal pattern. Well MO-1 penetrated the Nappe and Infra Nappe units, reaching the Paleozoic basement.

**Regional section R5**

This NE-SW oriented section located in Lalla Yto illustrates essentially the same features as on section R3. Again, the geometry of the wedge, the basal imbricates of the Infra-Nappe unit, and the normal faults that cut the Supra-Nappe succession are the most interesting features shown on this profile. On the SE margin of the section, south of the leading edge of the Prerifaine Nappe i. e. in the foredeep, the Supra-Nappe units onlap directly on the northeastward-dipping Infra-Nappe reflectors, thus defining the basal foredeep unconformity. Well KC-1, located in front of the Nappe, reaches the basement after penetrating Infra-Nappe Triassic, Cretaceous and Middle Miocene.

**Regional section R7**

This northeastward trending section is located in the region of Lalla Yto, west of section R5. The section shows the same features as section R7. In the frontal part of the
Nappe, extensional and compressional structures are detached at the same décollement level, thus defining toe-thrusts.

**Strike sections-General remarks (see Panel 2)**

The most conspicuous features shown by these strike sections are:

1. The thickness change of the Infra-Nappe unit.
3. Lateral ramps of the extensional system (the oblique orientation of the section reveals lateral ramps).

**Regional section R2**

This section is oriented E-W, extending from the region of Nouriat to the Atlantic Coast. The line shows the subsurface expression of the contact between the Rharb Basin and the frontal ranges of the Rif. The eastern end of the section shows westward-dipping listric normal faults, responsible for the westward-thickening of the Supra-Nappe succession. Most normal faults merge into a low-angle extensional detachment. Rotated blocks develop in the hangingwall of the extensional system. Extension in the Nouriat area occurs mostly during Messinian time. Proceeding westward, the structure of the central Rharb Basin consists of a series of westward-dipping listric normal faults. Lateral ramps, suggesting a transport direction oblique to the plane of the section, are common in the central portion of the section. In the western part of the section (Sebou region), the base of the Prerifaine Nappe is at 3 sec recording time. Infra-Nappe layered reflectors are well imaged. Occasional thrusts are present within the Nappe. The Prerifaine Nappe is not significantly affected by extensional faults in this western area.

**Regional section R4**

This E-W trending regional section is parallel to R2 and extends from Nouriat to the Atlantic. Most conspicuous on this section is the westward deepening of the top-of-
the-Nappe extensional detachment. Ramps and flats define the geometry of the basal
extensional detachment. The extensional décollement deepens down to 2.5 sec in the
central part of the section. Supra-Nappe sediments fill the downthrown depressions of the
extensional system. Only on the easternmost portion of the section is the Mio-Pliocene
boundary offset by normal faults. In the Sebou region the basal decollement of the
Prerifaine Nappe is imaged on the profile at 3 sec (TWT). Layered Infra-Nappe reflectors
underlie the basal decollement.

**Regional section R6**

This section shows listric normal faults superimposed on the Prerifaine Nappe.
Significant growth and rotation along faults occur in the Supra-Nappe succession. These
growth faults are lateral ramps of the extensional system and suggest a SW-transport
direction oblique to the plane of the section. The base of the Nappe is located
approximately at 2.5 sec recording time. Thrust faults within the Nappe merge into this
basal decollement. The basal detachment of the Prerifaine Nappe shallows westward and
offsets younger sediments in the westward direction. This suggests westward thrusting of
the Prerifaine Nappe onto the Infra-Nappe succession. The timing of the units is
constrained by three wells (MA-101, MO-1 and CGD-5). Wells MA-101 and MO-1
reached the basement.

**Regional sections R8, R10, R12, R14.**

These sections located in the Lalla Yto area are closely spaced (2 to 3 Km). They
are oriented NW-SE, providing more examples of strike sections across the Rharb Basin.
Wells MA-101 and MO-1 were tied into the seismic. The basal décollement steps down
from 1.8-2 to 3-3.2 sec recording time. Thickness changes of the Infra-Nappe succession
coincide with the location of nearly vertical faults that offset the basement. The top of the
basement and the overlying Infra-Nappe succession dip to the northwest. The Prerifaine
Nappe is thrusted onto Cretaceous and Middle Miocene Infra-Nappe sediments that
overlie the Hercynian basement of the Moroccan Meseta. Lateral ramps of listric normal faults offset the top of the Nappe and the overlying Supra-Nappe succession.

**Regional section R16**

This section across the central Lalla Yto area is roughly oriented WNW-ESE. The section is located near the leading edge of the Prerifaine Nappe. A complete Infra-Nappe succession penetrated by the KC-1 well consists of a Triassic half-graben and is unconformably overlain (Post-Rift unconformity) by parallel-bedded horizontal Cretaceous and Miocene sediments. Plio-Pleistocene units show a downlap pattern that suggests west-northwestward progradation.

**Selected areas of the Rharb Basin**

Some areas located in the peripheral part of the basin, where surface and subsurface data are abundant, were selected to show the detailed structure of the accretionary complex and the overlying sediments. These areas will be discussed separately, and are shown on figure 5.14.

**5.1.2.5 Nouriat Region**

The region of Nouriat is located near the central-eastern border of the Rharb Basin, east of Mechra bel Ksiri (see Fig.5.15). During the 1950s SCP made several gas discoveries in this area. This area of the Rharb represents the transition between the exposed Prerifaine Nappe (External Prerif in the sense of Suter 1980b) and the Rharb Basin. Most of the wells drilled in this region just reached the top of the Nappe. Thus, only Supra-Nappe information is available.

The following stratigraphy is based on wells NRT-1, NRT-2, NRT-3, NRT-4 and NRT-5 (Figs. 5.16 and 5.17).
Fig. 5.14. Location map of some selected areas of the Rharb Basin that will be discussed in detail.
Fig. 5.15. Location map of the Nouriat area (Rharb Basin).
Fig. 5.16. Structural contour map of top of the Nappe in the Nouriat Region (Eastern Rharb). Location of seismic and well data also shown. Contour interval is 100 milliseconds.
Fig. 5.17. Stratigraphic correlation of the Nouriat wells. After ONAREP (1983) (Internal reports). White arrows indicate low-angle extensional detachment.
Plio-Quaternary.

Lithology:

Fine-grained, carbonate-cemented sandstones with interbedded clay and occasional intervals of microconglomerates and coquinas.

Thickness range:

300-50 meters

Messinian

Lithology:

Grey silty and sandy-marl with occasionally interbedded siltstone, carbonate-cemented sandstone, shale or coquina intervals.

Paleontology:

Zone a *G. miotumida dalii*.

Thickness range:

900-1750 meters.

Comments:

Sandstone intervals contain methane gas.

Structure

A structural contour map of the top of the Nappe was constructed using all available seismic data of the region (Fig. 5.16). Six seismic profiles crossing the area have been selected to show the structure of this area (Fig. 5.18-5.23). The structure of Nouriat is characterized by closely-spaced westward-dipping listric normal faults indicating a westward transport direction over the eastern half of the area. In map view these faults show an anastomosing pattern. East-west trending doubly-vergent lateral ramps branch into north striking listric normal faults. Structural highs of the top of the Prerifaine Nappe, referred to as ridges, occur in the central part of the Nouriat area and were penetrated by
Fig. 5.18, Section NT-1, Nouriat region. The contact between the Nappe and the overlying sediments is a low-angle extensional detachment with ramps and flats. The Supra-Nappe sediments located on the hangingwall show significant rotation. Normal faults developed in the Supra-Nappe succession coinciding with ramps of the extensional system. Three wells (NRT1, NRT5 and NRT2) are located on this section. Wells NRT5 and NRT2 reached the top of the Prerifaine Nappe.
Fig. 5.19. Section NT-2 images another aspect of the basal extensional detachment of the Rharb Basin. Supra-Nappe sediments are strongly rotated in the hangingwall of the detachment system. The basal extensional decollement offsets thrusts within the Prerifaine Nappe. This cross-cutting relationship is also observed in other areas (see Souk el Arba area).
Fig. 5.20. Section NT-3 illustrates a westward-dipping normal fault merging into a west-vergent thrust (toe-thrust). Displacement due to extension is accommodated by compression. Extensional and compressional structures are coeval on time.
Fig. 5.1. Section NT-8 is a strike line along the Naurait extensional system. Northeastward and southward dipping low-angle extensional detachment faults form westward-directed extensional ramp sequences and a NE-trending fault zone. A culmination of the Naurait and a NE-thrust are present in the central part of the section.
Fig. 5.22. Section NT-10 shows another aspect of the lateral ramps of the extensional system superimposed on thrusts within the Nappe.
Fig. 5.23. Section NT-16 images lateral ramps of the extensional system described in the preceding sections. They constitute very low-angle extensional detachments located on top of the Nappe. Extensional features affecting only the Supra-Nappe succession are present in the central part of the section.
exploration wells. These ridges are similar to those described from the central portion of the offshore Larache area. They are due to lateral flow of shales and evaporites in response to sediment loading and extension. In the southern half of the area, the basal extensional system merges into an eastward-dipping N-S oriented toe-thrust (this feature is similar to the one described in regional section R7 of Panel 1). The arrows on the structural map indicate the apparent sense of transport associated with the extensional system. E-W and N-S trending seismic sections show different aspects of the Nouriat region (Figs. 5.18-5.23).

The E-W trending sections display:
1. Thrusts within the Nappe.
2. An overall westward deepening of the top of the Nappe.
3. A ramp-flat geometry of the Supra-Nappe extensional detachment.
4. The relationship between ramps of the extensional detachment and normal faults of the Supra-Nappe succession.
5. A toe-thrust associated with the low-angle extensional system.

The N-S oriented sections show:
1. Ridges in the central portion of the extensional system.
2. Lateral ramps of the extensional system.

For description of each line see figure captions.

5.1.2.6. Rhabet-Jebila Lalla Zhara area.

The region of Rhabet Jebila is located near the northern margin of the Rharb Basin, east of Moulay bou Selham (Fig. 5.24). This area is a Quaternary alluvial plain that overlies the Pliocene of Dhar el Ahrech. The hills of Lalla Zhara are located on the northern border of the Rharb Basin, which is one of the few areas where the Miocene is exposed on the surface. The structures trend E-W, an atypical orientation for the Rharb Basin where the structures normally trend NW-SE.
Fig. 5.24. Location map of the Rhabet Jebila -Lalla Zhara area (Northern Rharb).
Stratigraphy

The stratigraphy of this area is known from oil wells drilled by the SCP in the late 1960s (Fig. 5.25) and by field work in the Lalla Zhara hills (Morel 1988, Wernli 1988, and own observations). The stratigraphy of the Lalla Zhara region is outlined on Fig. 5.26. A major sequence boundary separates the transgressive coquinaid sandstone of the G. crassaformis biozone and bioclastic sandstone with G. inflata. An important erosional surface cutting down into the Miocene affects previous sediments and predates the deposition of Villafranchian deposits (pre-Villafranchian sequence boundary). Channeled coarse conglomerate overlies tabular parallel-bedded conglomerate and sandstone. This succession suggests a highstand progradation.

The subsurface stratigraphy of the Rhabet Jebila area is based on three wells, i.e. well RJB-1 (SCP 1967), well RJB-3 (SCP 1969) and well LZ-6 (SCP 1950) (Fig. 5.25,5.27). A complete description including paleontology of well RJB-3 (Tachet des Tombes 1967) shows in greater detail the stratigraphy of the Supra-Nappe succession of this area (Fig. 5.28). Biostratigraphy is based on the Paleontology zones of Tachet des Tombes (1967).

Well RJB 3 (After SCP 1969).

Top of the Nappe is located at 1062 meters depth.

Zone III = G. menardi zone

Depths

1062-625 meters

Lithology

Pyritic grey marl with glauconite and lignite. The lower part of the section (1030-980 m.) is rich in organic matter and plant remnants. The interval between 1020-1062 m. consists of silts and rounded microconglomerates with silex.
Fig. 5.25 Structural contour map of the top of the Nappe in the Rhabet Jebila area. Contour interval 100 milliseconds. Subsurface data after ONAREP (1991) Internal reports. Surface data from Wernli (1988) and Morel (1988).
Fig. 5.26. Stratigraphic section of the Lalla Zhara hills. After Wermli (1988) and Morel (1988).
Fig. 5.27. Stratigraphic sections of wells LZ-6, RJB-1 and RJB-3 of the Rabet Jebila-Lalla Zhara area. After SCP (1950, 1967 and 1969) in ONAREP internal reports.
RJB 3 well (Rhabet Jebila)

Fig. 5.28. Biostratigraphic log of well RJB-3. After SCP (1969) in ONAREP internal reports.
Paleontology and Environment

Directly on top of the Nappe, littoral conditions are suggested by large size bryozaonans and shell debris, abundant small size ostracodes, *Ammonia sp.* and *Elphidium sp.*. These forms decrease upwards in number and size. The pelagic forms increase upwards and the ostracoda decrease in number and size upwards. *Elphidium sp.*, *Bolivina sp.*, *Uvigerina sp.*, *Cibicides sp.*, *Ammonia sp.* and ostracoda occur from 1020-800 meters. *Ammonia sp.* and *Uvigerina sp.* appear at 800m, *Eponides praecinctus* and *Buliminella sp.* appear at 200m and *Anomalina trinitatensis* at 650m. Down to depths of 800 m the succession is represented by a littoral shallow-water environment.

Zone IV = G. dutertrei

Depths

625-325 m.

Lithology

Grey marl with glauconitic and pyritic sandstone intervals. Glauconite is present as microconglomeratic grains. Asphalt remnants, lignite, bipyramidal quartz and pyrite are common.

Paleontology

625-500 m. Large size Orbulines and abundant reworked Cretaceous and Oligocene fauna. Abundant pelagic fauna .

500-325 m. *Uvigerina cf. flinti*, large *Rubulus sp.*, *Cassidulina sp.*, *Planulina aris tizensis*, *Cibicides ungerianus*, *Cibicides pseudoungerianus*, they diminish upwards passing to an assemblage of:

*Anomalina trinitatensis*, *Anomalina helicina*, *Cibicides sp.*, *Textularia sp.*, *Plectofrondicularia pentecostata*, *Plectofrondicularia gemina*, *Ceratobulimina pacifica*, *Nodosaria longiscata*. 
Zone V= G. margaritae zone

Depths
325-50m

Lithology
Fine-grained pyritic sandstone, and sandy grey glauconitic marl, red and rose quartz, often bipyramidal, lignite, asphalt.

Paleontology and environment
Faunal assemblage: small size Robulus sp., small size Ammonia sp., Uvigerina subperegrina, Cibicides sp., Sphaeroïdina bulboides, Bolivina Scalprata v. miocenica.
Reworked faunas from lower zones in the Miocene are common.

At 250-200 m. Abundant Elphidium sp. and Nonion incisum indicating a more littoral condition of deposition.

180-80 m. Ammonia sp, small size ostracoda and debris of equinodermata.

Zone VI= G. puncticulata, G. crassaformis and G.inflata.
Absent.

Plio-Quaternary=Villafranchian+Pleistocene to recent.
Sandy clays and red and yellow sandstones and conglomerate.

Structure
The Rhabet Jebila / Lalla Zhara region is subdivided into a northern extensional satellite basin and a southern ridge (Fig. 5.24). The top of the Prerifaine Nappe is offset by listric normal faults. Deep depocenters in excess of 2.1 sec (TWT) are associated with eastward trending southward-dipping normal faults (ONAREP internal report 1991). The Lalla Zhara hills coincide with an E-W trending structural high on top of the Prerifaine Nappe (Ridge of Lalla Zhara). The structure of the Lalla Zhara Ridge is characterized by E-W and WNW- ESE trending anticlines and synclines offset by ENE-WSW trending faults. In the central sector of Lalla Zhara north-vergent thrusting has been observed

According to Morel (1988), the compressional structures of the region of Lalla Zhara are the result of strike-slip movement along a N70-80 deep-seated strike-slip fault, a continuation of the Jebha fault. This hypothesis is not consistent with the seismic evidence. In this area the data available consist of three seismic sections and the contour map of the top of the Nappe (ONAREP 1991 internal report). N-S trending sections (Fig. 5.29), extending from Rhabet-Jebila to the Lalla Zhara area (see also section R1 of Panel 1), illustrate the internal structure of the Prerifaine Nappe and the extensional system superimposed on the wedge. The genetic relationship between the extensional detachment and the compressional ridge of Lalla Zhara is clearly displayed by the sections. The ridge of Lalla Zhara is explained as a toe-thrust connected with northward-dipping extensional faults. North-directed thrusting, reported in the field (Morel 1988, Wernli 1988), can be connected with a deep-seated toe-thrust, constituting a triangle zone or a wedge of conjugate thrust faults. Compression in Lalla Zhara coincides with a hiatus of the biozone *G. puncticulata* and is marked by a major angular unconformity between Pliocene marls and Villafranchian alluvial / fluvial deposits.

**Section RJ-1**

This is a N-S view of the Rhabet Jebila-Lalla Zhara area. A series of southward-dipping normal faults offsets the top of the Nappe. The extensional faults sole out in a basal décollement located at 2.5-2.7 sec (TWT). The region of Lalla Zhara constitutes a structural high caused by a rotated tilted block. The extensional faults do not affect the Plio-Pleistocene boundary. Well RJB-3 is located on this section (for description of well see fig. 5.30). This section coincides with the northern part of section R1 of Panel 1.

**Section RJ2**

This is another N-S oriented section through the Rhabet Jebila / Lalla Zhara area. Southward and northward-dipping listric normal faults cut the top of the Prerifaine Nappe.
Fig. 5.29. Sections through the Rhabet-Jebila Lalla Zhara area. Sections show the extensional system superimposed on the Prerifaine Nappe.
These faults affect Upper Miocene sediments but do not affect the overlying Plio-
Pleistocene succession. The southern portion of the section is occupied by the Lalla Zhara
structural high. This high is on a rotated block associated with the basal extensional
detachment of the Prerifaine Nappe. Well LZ-6 is located in the southern portion of the
section.

Section RJ3

This northeastward trending section illustrates southward-dipping normal faults
offsetting the top of the Nappe. A listric normal fault in the northern part of the section
shows considerable growth of the Supra-Nappe unit. The section also images thrust
structures within the Nappe. Well RJB-1 is located in this transect.

5.1.2.7 Souk el Arba area.

The region of Souk el Arba is located in the northeastern corner of the Rharb
Basin (see Fig. 5.30). The SCP (Société Chérifienne des Pétroles) explored this region in
the 1940s. Wells in the Fokra area show thrusting within the Prerifaine Nappe. The Nappe
is overlain by a thick Miocene Supra-Nappe succession (Bruderer 1942, in ONAREP
internal reports). The abandoned oil well of Ain Hamra is located in the western part
of this region. Gas was discovered in Neogene Supra-Nappe sandstone intervals (Fig. 5.31).

Stratigraphy

The stratigraphy of the area is constrained by several wells. The Nappe succession
is exposed north and northwest of the village of Souk el Arba and in the northern Fokra
area. The description of wells Fokra FO-1 (SCP 1942) (Fig. 5.32) and SEA-2 (SCP 1942)
(Fig. 5.33) gives an indication of the Neogene stratigraphy of the area.
Fig. 5.30. Location map of the Souk el Arba area (Rharb Basin).
Fig. 5.31. Structural contour map Top of the Nappe. Contour interval 200 milliseconds. Modified from ONAREP internal reports (1991).
Fig. 5.32. Stratigraphic section of well Fo-1 (Fokra) NW of Souk el Arba. After the SCP (1942). Black arrow indicates thrust fault and white arrow low-angle extensional detachment.
Fig. 5.33. Stratigraphic section of well SEA-2 with indication of lithology, biozones and sedimentary environment. After SCP(1966) (in ONAREP internal reports).
Well Fokra FO-1 (SCP 1942)

Prerifaine Nappe.

Lithology

The Nappe consists of grey marls with reworked Cretaceous fauna and Triassic salt and diabases or dolerites.

Supra-Nappe.

Thickness

353 meters

Lithology

A 15 m thick sandstone interval at the base of the Tortonian directly overlies the top of the Nappe. Tortonian blue marl with three sandstone intervals contain gas; two of them can be correlated with the Aïn Hamra well.

Paleontology and age.

The paleontology of this well was done by the SCP (Bruderer 1942 in ONAREP internal reports). The Supra-Nappe succession was at that time reported as Tortonian.

Comments

Most of the Supra-Nappe succession of the peripheral regions of the Rharb were reported initially (40s and 50s) as Tortonian (Bruderer 1942 in ONAREP internal reports). More recent paleontologic studies in neighboring areas (ONAREP 1982 internal report) indicate that the Miocene Supra-Nappe succession is mainly Messinian.

Well SEA 2 (After SCP ) (Fig. 5.33)(see Fig. 5.31 for location).

Top Nappe at 870 m.

Zone II = G. menardii-saphoe

Depths

600-870 m.
Lithology

Passage through a basal sandstone located on top of the Nappe. Marls with siltstone intervals with abundant pyrite and lignite-carbon remnants and continuous glauconite and oxidized surfaces.

Paleontology

Abundant reworked fauna, small size specimens (general: small size Ammonia sp., Nonion incisum, Uvigerina bononiensis compressa, Virgulina schreibersiana, small sized Bolivina sp. and Haplophragmoides sp.).

Passing into an assemblage of Nonionella sp. Virgulinella sp. and bolivinids.

Passing into an assemblage of: Virgulina schreibersiana, Uvigerina bononiensis compressa and Butonia sp. (ostracoda).

The upper part of this zone is defined by the association of Rohulus cf. calcar., Nonion incisum, Haplophragmoides sp.

Environment

Shallow-water littoral mud dominated.

Zone III = G. menardi zone

Depths

600-300 m.

Lithology

Silty marls with occasional sandstone intervals grade into pyritic marly-limestones with organic detritus. Glaucanite is common in sandy and silty beds. Ferruginous incrustations.

Paleontology

Microfaunal assemblage of: Cibicides sp. and Cibicides mexicanus, Cassidulina laevigata, Uvigerina subperegrina, Uvigerina cf. rutila. Association of arenaceous foraminifera: Saccammina sp., Haplophragmoides sp. and Bathysiphon sp. Below 550m.
appearance of *Nonion incisus*.

**Environment**

Littoral to sub-littoral. Upward increase in benthic fauna indicating shallowing upwards conditions.

**Zone IV** = *Orbulina* = *G. puncticulata, G. crassaformis inflata*.

**Depths**

300-15 m.

**Lithology**

Pyritized grey marl with locally interbedded siltstone and sandy-marl beds, containing glauconite, lignite remnants, pyritized foraminifera, gastropods, brachiopods, annelid tubes, and echinoderms.

**Paleontology**

20-30 % pelagic fauna. It is dominated by *Orbulina sp.*, occasional *G. nephentes*.

Pelagic fauna: *Uligerina cf. rutila, Bulimina pyrula, Bulimina pupoides* and *Spiroplectammina carinata*. Rare: *Ehrenbergina dinapolii, Textularia sp.*, *Lingulina costata, Ramulina globulifera, Robulus spinulosus, Epinodes praeinctus, Bairdia sp.* and pyritized lamelibranchia.

**Environment**

Sub-littoral conditions.

**Plio-Quaternary**.

**Thickness**

15 m

**Lithology**

Sandstones, conglomerates and clays.
Structure

Souk el Arba is a key area for the understanding of the contact between the Rharb Basin and the Rif front. The Prerifaine Nappe imbricates, consisting of closely-spaced steeply northeastward-dipping thrusts, are intersected by a southwestward-dipping low-angle extensional detachment (Bruderer 1942) (Fig. 5.31 and 5.34). The structure of Souk el Arba is characterized by a northern extensional basin and a southern compressional ridge. The Prerifaine Nappe, exposed on the surface in the northwestern portion of the area, is offset by northwest trending southwest-dipping normal faults, forming a fault-bounded extensional basin. Proceeding southward the structure consists of a NW-SE oriented compressional ridge cropping out in the village of Souk el Arba (Souk el Arba Ridge), referred to as the "bombement de Souk el Arba". The internal structure of the Souk el Arba Ridge consists of southwest-vergent thrusts (Tilloy 1955a, Suter 1980). The Prerifaine Nappe reaches a maximum depth of 1.5 sec (TWT) in the depocenter of the basin (ONAREP internal report 1991). A lateral culmination of the Nappe approximately trending N-S occurs also in Aïn Hamra (Aïn Hamra Ridge). The seismic data suggest that this structural high may be associated with the basal extensional detachment of the Nappe. Three seismic profiles (Figs. 5.35, 5.36 and 5.37) illustrate the extensional-compressional setting of this region. The northern part of the sections is represented by southwestward-dipping growth faults, and the southern and western part is formed by the Souk el Arba and Aïn Hamra Ridges. The genetic relationship between extension and compression is not obvious from the local sections, but it becomes clear in the regional context (see model Fig. 5.34).

Section S1 (Fig. 5.35)

This NE-SW oriented dip section illustrates the structure of the extensional basin of Souk el Arba. Southwestward-dipping normal faults merge into an extensional décollement that flattens out at 1.5 sec of the recording time. The Supra-Nappe sediments
Fig. 5.34. Depth-converted section integrating field data from SCP (Souk el Arba Geol. map) and Subsurface data from the SCP, Bruderer (1942) in ONAREP internal reports.
Fig. 5.35. Section S1. Uninterpreted and interpreted dip sections along the Sask El Arba Basin and Ridge.
show significant growth and rotation. The southern portion of the section is occupied by the Souk el Arba Ridge. This structural high is due to the rotation of the extensional system.

**Section S2 (Fig. 5.36)**

This NW-SE oriented strike section is oblique to S1. It offers another perspective of the extensional system. The northwestern margin of the section is occupied by a culmination of the Prerifaine Nappe.

**Section S3 (Fig. 5.37)**

This is another oblique section that displays a view of the extensional system. The seismic line is oriented WNW-ESE and shows the relationship between the axial culmination of the Prerifaine Nappe and the extensional system. The seismic character of the Nappe reflectors suggests compressional structures connected with the basal extensional decollement. The extensional detachment is located at 1.6-1.7 sec of the recording time of this profile.

**5.1.2.8 Arbaoua region**

The region of Arbaoua is located south of Ksar el Kbir, north of Rhabet Jebila. It represents the frontalmost NW-SE trending structures of the Rif (see Fig. 5.38). There are no seismic data for this region; only surface data are available. The Arbaoua region is characterized by broad undulations of large wavelength, the "vastes ondulations" of Morel (1988) affecting the Prerifaine Nappe and its thin Neogene cover.

**Stratigraphy**

**Prerifaine Nappe.**

Consists of Tortonian white marl and Triassic shales and evaporites.

**Neogene Supra-Nappe.**

A thin Supra-Nappe succession overlying the Prerifaine Nappe characterizes this area. A thin Pliocene yellow and red siltstone interval (Middle Pliocene aeolian
Fig. 5.38. Location map of the Arbaoua area (Northern Rharb).
sandstones), overlain by red, rounded Villafranchian conglomerates with sandy matrix and ferruginous coating, represents the Supra-Nappe succession (Morel 1988) (Fig. 5.39).

**Structure**

Supra-Nappe Villafranchian conglomerates and sandstones that involve steeply to gently-dipping, onlapping units are bounded by angular unconformities with similar geometry to those found in diapiric flanks or on growing anticlines (Fig. 5.40). These angular unconformities indicate that anticlinal growth is syn-Villafranchian (Lower Pliocene). Because of the lack of seismic data, it is not possible to know if these structures are related to diapiric anticlinal growth or to ridges associated with extensional detachments (toe-thrusts), and to trace the trailing edge of the extensional system that affects the Rharb Basin. The structure of this region has been interpreted as due to late stage compression (Villafranchian) related to strike-slip tectonics (Morel 1988).

**5.1.3. OFFSHORE NORTHWESTERN MOROCCO.**

A large number of seismic profiles covering the northwestern Atlantic margin of Morocco were used to map the structure of the Prerifaine Nappe in the offshore region.

**5.1.3.1. Stratigraphy**

Since only one well was drilled in the northwestern Moroccan offshore, stratigraphic information is largely based on correlation with onshore data. The only offshore well drilled (LAR-1 well) is located near Larache (Fig. 5.41). This well provides the sole stratigraphic control for the offshore region (Fig. 5.42). The transition to the Pleistocene is not well defined due to lost circulation and consequent lack of samples. The paleontology of this well was done by the BRPM (1975) and in part follows Wernli (1988). The Mio-Pliocene boundary is defined in both cases by the base of the biozone *Globorotalia margaritae*. 
Fig. 5.39. Stratigraphic section of Aioun Bsal (Arbaoua region). After Morel (1988).
Fig. 5.40. Progressive and angular unconformities in Villafranchian conglomerates associated with a culmination of the Prerifaine Nappe. Arbaoua (Northern Rharb).
A. Unconformities in Villafranchian alluvial conglomerates. South of Kherarkal.
Fig. 5.41. Structural map Offshore Northwestern Morocco. Location of well LAR-1.
Fig. 5.42. Stratigraphy and paleontology of LAR-1 well (Offshore Larache). After BRPM (Bureau de Recherches et de Participations Minieres1975) and Wernli (1988).
The biostratigraphy of the LAR-1 well has been tied into two seismic sections. Figures 5.43 and 5.44 show the line drawings of the seismic sections with the location of the biostratigraphic zones.

5.1.3.2. Structural map.

Seismic data provided by ONAREP and PETROCANADA were used to make a structural map of the Atlantic margin offshore northwestern Morocco. The map covers an area of more than 10,500 Km², extending from Tanger in the north to Rabat in the south, and from the coastline to 1700 meters water depth. The structural map shows the offshore prolongation of the Rif frontal thrusts and the structures of the Rharb Basin (Fig. 5.45). The region can be subdivided into four structural domains (Fig. 5.46) (see Fold-out 1) as follows:

1. Offshore Tanger-Asilah / Fold and Thrust belt

   The northern part of the Moroccan Atlantic margin consists of WSW-vergent folds and thrusts. Fold axes and thrusts strike NNW-SSE, following the general trend of the Rif Cordillera. They represent the continuation of the structures of the Tanger and Habt units. The quality of the seismic data in this area is mediocre, and there are some regions with virtually no useful data.

2. Offshore Larache / Extensional zone.

   Proceeding southward, folds and thrusts are cut by NW-SE trending SW-dipping normal faults. The structure of this region is defined by troughs and ridges. In the eastern portion of the area, normal faults anastomose, merging into each other to define a complex extensional system that is superimposed on the accretionary wedge. Structural culminations roughly oriented in a NW-SE direction occur in the central and western parts of this extensional zone. This zone is the prolongation of Prerifaine Nappe.
Fig. 5.43. Structural map offshore Larache. Location of lines A and B and LAR-1 well.
Fig. 5.44. Line drawings of a dip and a strike section in the offshore Larache region. Sections show biostratigraphic zones tied with LAR-1 well.
Fig. 5.45. Structural map of the frontal Western Rif based on seismic data. The structural map of the Rharb Basin is partially taken from ONAREP internal reports (1991).
Fig. 5.46. Structural map Offshore Northwestern Morocco with indication of major structural domains.
3. **Offshore Rharb / Frontal Imbricates-Extensional-Compressional Zone.**

South of Moulay Bou Selham, a combination of compressional and extensional elements can be identified. In this complex area, NW-SE trending normal faults and occasional NE-SW-oriented normal faults cut NW-SE trending SW-vergent folds and thrusts. To the east of the area, normal faults share the same décollement level as thrust faults, defining toe-thrusts that accommodate the normal fault displacement.

The frontal part of the wedge is characterized by closely spaced NW-SE trending, NE-dipping thrust faults that define a zone of frontal imbricates. Lateral and oblique ramps related to these frontal thrusts are common in the central and western portions of the area. The front of the accretionary complex has a NW-SE orientation. This zone will be extensively analyzed and illustrated with seismic data in section 5.2.1.2

4. **Offshore Rabat / Foredeep**

South of the leading edge of the accretionary complex, the structures consist of NE-dipping Infra-Nappe units that dip underneath the accretionary wedge. Basement-involving normal faults disrupt the otherwise continuous Infra-Nappe succession. An angular unconformity with its onlap defines a "basal foredeep unconformity". This region is the offshore prolongation of the Rif foredeep that was described in section 4.3.

5.1.3.3. **Regional Transects offshore Northwestern Morocco.**

Five NE-SW offshore regional sections, extending from Asilah to Rabat, and four NW-SE oriented sections in the offshore Larache area have been selected to show the structure of the northwestern Moroccan Atlantic margin (Panel 3). The transects display a variety of structural styles. According to the seismic character and structural significance, a number of structural units are differentiated, from bottom to top:

2. Infra-Nappe : Mesozoic and Lower Miocene cover of the basement.


In the following these units will be referred to as: Basement, Infra-Nappe, Nappe and Supra-Nappe.

Dip sections

The NE-SW oriented sections traverse the main structural units of the accretionary complex shown on the structural map. The regional sections will be described proceeding from the southern foreland basin to the northern frontal folded belt. The southern part of the transects shows northward-dipping layered reflectors of the foreland that project under the frontal imbricates of the accretionary complex. These reflectors are occasionally detached from the acoustic basement to form imbricates. The frontal imbricates are characterized by thrust planes that dip steeply to the north. Thrust sheets emanate from a gently northward-dipping basal décollement that separates them from the underlying autochthon. Proceeding northward, the complex is overprinted by northward-dipping normal faults and associated extensional basins with no significant growth. Farther north, extensional faults step down to form thick extensional basins. They constitute the southern part of an extensional system running nearly perpendicular to the plane of the sections. Ridges cored with accretionary complex sediments occur in the central region of the extensional system. The southern branch of the extensional system is characterized by northward-dipping normal faults. These faults are connected with conjugate southward-dipping listric normal faults that often constitute the lateral ramps of the extensional system. The central portion of the extensional basin, with its high ridges and deep troughs, is detached from the basal extensional contact. This portion of the margin provides exceptional sections across the extensional system. Proceeding northward, the northern branch of the extensional system cuts thrusts and folds of the underlying accretionary complex. The northern portion of the sections is characterized by north-dipping thrust
planes and related ramp anticlines. For a more detailed description of the main structural features shown by these sections, see Panel 3.

Offshore Larache Strike sections

The structure of the offshore Larache region is characterized by down-to-the-basin listric normal faults and associated rotated blocks (Fig. 5.47 and 5.48). The basal detachment of this system steps down from 0.5 sec in the east to 3.5 sec in the west. The regional sections offer a strike perspective of the extensional system. Four E-W oriented dip sections, shown in the lower-right corner of panel 3, illustrate the extensional system superimposed on the accretionary complex. The offshore Larache structure consists of strongly rotated blocks on the hangingwall of a low-angle extensional detachment. In the eastern portion of these sections, growth-faulting results in large Supra-Nappe expansion controlled by westward-dipping normal faults that sole out into the basal detachment. In map view these faults show an anastomosing pattern, resulting in extensional horses. Sediments above the extensional system show significant growth. In the central portion of the area, the top of the Nappe attains depths of 3.7 sec (TWT). Nearly E-W oriented ridges, are present in the central part of the extensional system.

5.1.3.4. Offshore Rharb area.

This area is located west of the southern Rharb Basin, between the confluence of the Sebou River and the village of Moulay Bou Selham. The seismic data consist of NE-SW and NW-SE trending seismic lines, covering an area of 4500 square kilometers (Fig. 5.49). The seismic data of this region show the details of the frontal part of the accretionary complex and the northernmost portion of the Rif foredeep. The structure of this area is illustrated by several selected high quality seismic sections (Fold-outs 1-26, see fig. 5.3 for location of the lines). For convenience, the seismic grid is subdivided into dip and strike lines. Dip lines are those trending perpendicular to the leading edge of the
Fig. 5.47. Structural map Offshore Larache. Location of dip sections.
Fig. 5.48. Line drawings of dip sections, offshore Larache. Dashed line indicates top Miocene.
Fig. 5.49. Location map of the Offshore Rharb seismic data presented in this work. (see Plates 1-26).
accretionary complex, that is NE-SW. Strike lines are those trending roughly parallel to the front of the wedge, that is NW-SE.

**Dip sections**

The most conspicuous features shown by these sections are:

1. The wedge-like geometry of the Nappe.
2. Imbrications within the Infra-Nappe authochthonous succession.
3. Frontal thrusts of the Accretionary Complex (i.e. Nappe).
4. Extensional faults crosscutting the wedge.
5. A northwestward-dipping basement underneath the Nappe.
6. The northernmost portion of the foredeep.
7. Frontal slumps

For details of the "Dip sections" see captions and plates 1, 2, 4, 6, 8, 10, 12, 14, 16, 18, 20, 22, 24 and 26.

**Strike sections**

The strike sections show the following structural features:

1. Steeply dipping basement-involved faults offsetting the base of the Infra-Nappe units.
2. Eastward thinning of the Prerifaine-Nappe.
3. Lateral ramps of the frontal imbricates.
4. Westward progradation of the Supra-Nappe succession.
5. Frontal slumps and detached units.

Notice that the apparent gentle westward deepening of the wedge is a velocity effect due to the water column.

For explanation of each single "strike" line see figure captions in fold-outs 1, 3, 5, 7, 9, 11.
General comments on the seismic character of the units.

Supra-Nappe
Well-bedded reflectors that often show clinoformal pattern. Some transparent zones.

Nappe
Chaotic reflectors with some organized reflectors. A strong high-amplitude basal reflector indicates the basal décollement.

Infra-Nappe
Well-bedded nearly horizontal reflectors.

Acoustic basement
Lack of significant reflectors.

5.1.3.5 Comments on the Infra-Nappe succession

The geology of the west-African Atlantic margin in Morocco is well known from the regions of Aaiun, Tarfaya and Essaouira. The structure is characterized by a shelf margin with Jurassic and Lower Cretaceous carbonates overlying blocks of Triassic red-beds bounded by steeply seaward-dipping normal faults (Jansa and Wiedmann 1982). A Triassic to Lower Jurassic salt province extends from Tarfaya to offshore Rabat (Heyman 1989). This province is characterized by large diapirs and northward thinning of the evaporites. In the northernmost portion of the Mesozoic Atlantic margin, structures are overlain by the frontal accretionary wedge of the Rif Cordillera. In the Mazagan Plateau a thin deep-water Jurassic and Cretaceous pelagic succession contrasts with the Tarfaya and Essaouira margins, which consists of platform carbonates (Hinz et al. 1982 a,b). The Infra-Nappe succession of the offshore Rharb area has several angular unconformities and thins towards the west. The structural style, thickness and distribution of the sediments are similar to the Mazagan Plateau,
even though there are no wells that indicate the nature of these sediments. Tentative correlation with onshore wells suggests that part of the Infra-Nappe reflectors of the offshore area are Cretaceous (see Fig. 5.13). According to seismic and well data (Heyman 1989) two main unconformities are present in the Cretaceous succession of Tarfaya and Essaouira and in the exposed cover of the Moroccan Meseta (Gigout 1951, Michard 1976). One is located at the base of the Neocomian and the other at the base of the Cenomanian. They could be equivalent to some of the unconformities seen on the seismic sections. Normal faults and related tilted blocks suggest Triassic half-grabens superimposed on the Hercynian basement (Fig. 5.50). The Infra-Nappe unit is offset by NE-SW trending basement faults that coincide with thickness changes. These faults are part of a fault system that is responsible for a spectacular thickness change within the Mesozoic sedimentary cover that overlies the Moroccan Hercynian basement (Fig. 5.51). The Mesozoic Infra-Nappe thins towards the east in the Rharb Basin, where the Cretaceous is absent, and thickens dramatically towards the east across the Sidi-Fili fault. East of Sidi-Fili there is a thick Jurassic succession (more than 2000 meters) overlying the Paleozoic basement. The lack of Mesozoic in the Rharb Basin is perhaps due to erosion and exhumation of a shoulder uplift connected with the Atlantic rifting (Favre and Stampfli 1992). The fault system of northwestern Morocco is part of a complex mosaic of blocks located between the ancient Tethys and Atlantic rift margins. The connection between these systems and its prolongation in the south-Iberian margin (Betic Cordillera) is not well understood and a source of debate. Figure 5.52 shows a possible reconstruction during Early Jurassic time of the African and Iberian plates.
Fig. 5.50. Block diagram of the Infra-Nappe succession of the offshore Rharb area reconstructed before compression.
Fig. 5.51 Structure of the Infra-Nappe succession in the Western Rif. Reconstructed for the Top Cretaceous.
Fig. 5.52. Plate tectonic reconstruction of the Western Tethys and Northern Atlantic regions during Pleinsbachian time. Modified from Dercourt et al. (986), Ziegler (1987), Andrieux et al. (1989) and Favre and Stampfli (1992). The figure show the location of the main carbonate platforms.
5.1.3.6 Sequence Stratigraphy of the Offshore Rharb Supra-Nappe succession.

Planktonic biostratigraphy from wells and outcrops in the Rharb Basin has been tied into offshore seismic lines to estimate the age of the sedimentary units (see Fig. 5.13). The seismic expression of the Supra-Nappe succession suggests a prograding mud-dominated shelf margin contrasting with conventional deltaic progradations (Berryhill and Suter 1986). The lack of sandstone prevents the deposition of coarse clastic basin floor fans. Sequences are represented by thick lowstand prograding complexes dominated by slumps and mass flow deposits, well-developed transgressive backstepping units and thin or deeply eroded highstand deposits (Fig. 5.53). Lowstand deposits consist of prograding complexes overlain by stacked slope fans. Prograding complexes are characterized by clinoforms and chaotic reflectors at the toe, that represent slump or mass flow deposits. These mass transport deposits are characteristic of late lowstand deposition (P. Vail pers. comm. 1992). Substantial erosion, associated with canyons and incised valleys, can attain more than 150 meters. The sequence stratigraphic analysis of the seismic data reveals nine Pleistocene sequences. According to their estimated time span (~ 0.1 m.y.), these sequences correspond to fourth-order cyclicity (Vail et al. 1991), of about the same number as the Pleistocene sequences recognized for the Gulf of Mexico (Wornardt and Vail 1991). Sequences are probably related to glacio-eustatic cycles, which are characterized by smaller magnitude but higher frequency than tectonically induced transgressive-regressive facies cycles (Vail et al. 1991). Isotopic, geomorphologic and geologic data suggest high frequency sea-level fluctuations during the Pleistocene (Shackleton and Opdyke 1973, Bartek et al. 1990, Vail et al. 1991). Changes over the last 0.8-0.9 Ma have a periodicity of 0.1 Ma and an average amplitude of over 1.5 per mil (a sea level equivalent of > 130 m) (Williams 1988). Several stages recognized by Stearns (1978) based on outcrops located along the Atlantic Coast of Morocco have a similar order of magnitude.
Fig. 5.53. Fourth-order glacio-eustatic sequences in the Offshore Rharb Pleistocene succession.
Outcrop expression of Plio-Pleistocene sequences

The Atlantic Coast of Morocco is an exceptional area to study sea-level oscillations, since Pliocene to Quaternary continuous uplift has exposed several strand lines, dune ridges and sediments that can record sea-level fluctuations during the past 5 Ma of the earth's history (Stearns 1978). These deposits are particularly well exposed in specific areas.

Mio-Pliocene sequences

A major Pliocene transgression is characterized by quartz-rich sandstone with a basal lag deposit suggesting transgressive ravinement. Pliocene sandstone overlies Messinian marls through an important erosional surface in Charf el Akab (north of Asilah) associated with a hiatus (Wernli 1988). The Messinian-Pliocene contact is a major sequence boundary followed by transgressive ravinement in the northern Rif (end-Pontian surface). Lower Pliocene sediments (*G. margaritae* zone) onlap onto substratum consisting of imbricates of the Nappe prerifaine and the Flysch units in the Asilah area (Wernli 1988).

In the Grottes d'Hercules (South of Cap Spartel, western Tanger peninsula), there is a continuous marine section from the Pliocene to the Pleistocene (the only one in Morocco). This marine section is represented by marls with planktonic microfauna of *Globorotalia truncatulinoides*, *Globorotalia menardii*, *Globorotalia crassaformis* and *Globorotalia inflata*, a characteristic faunistic association of Pleistocene age (Feinberg 1986). Lower Pliocene sandy marls (*G. margaritae* zone) grade into Upper Pliocene marls with *G. inflata* (Messaoudian transgression 1.3 my). Higher in the section marls with *G. inflata* pass into marls with *G. truncatulinoides* that represent the base of the Pleistocene (Calabrian stage).
Pleistocene Sequences

The offshore sequences recognized on seismic lines on the Atlantic margin have an expression onshore in the coastal area. Table 5.7 shows a compilation of marine and continental stages for the Pliocene of Morocco based on several authors (Biberson 1965, Stearns 1978, Texier et al. 1985, Raynal et al. 1986 and Weisrock and Fontugne 1991), and the estimated correlation with the offshore seismic fourth order sequences.

Widespread deposits of aggradational Villafranchian fluvial-alluvial conglomerates and red-beds overlying the Rharb Basin and surrounding areas suggest highstand progradation related to a high sea-level stand (Fig. 5.54).

An End-Villafranchian erosional surface affecting Villafranchian yellow sandstone is well exposed in the Khemis du Sahel area (foret du Sahel area, SW of Asilah). This surface can be traced along the top of the sandstone unit (Carte Géologique du Rif 1:50,000, Asilah). A superimposed fluvial system incises the flat erosional surface.

The Fouaratan transgression or Moghrebian transgression is evidenced in the coastal Rharb region by landward shift of the near shore dunes belt ("grès de Mamora") and shallow-water merja deposits ("grès des merjas") and pisolithic sandstone backstepping onto coastal plain red siltstones and alluvial conglomerates and sandstone (Tilloy 1955b) (Fig. 5.55, 5.56). A coarse marine conglomerate level in the Mamora region represents the transgressive deposits associated with the Moghrebian transgression (Choubert 1965). This transgression accounts for an important landward shift of the shelf break and the end of the transgression may be equivalent to the important offlap break shift recognized in offshore seismic sections (sequence K in Table 5.7).

Late Pleistocene-Holocene sea-level fluctuations are represented in the coastal Rharb area by shifting of dune ridges (Fig. 5.57). Shallow brackish-water lakes, referred to locally as Merjas, developed on the landward side of the dune ridges due to an older sand dunes belt ("grès de Rabat") related to higher former sea level (Fig. 5.58). The Flandrian
Table 5.5. Chronology of the Moroccan Plio-Pleistocene and tentative correlation with Seismic sequences. Compiled from Biberson (1965), Stearns (1978), Texier at al. (1985), (Raynal et al. (1986), Wengler and Vernet (1992). (For Offshore seismic sequences see fig. 5.53).
Fig. 5.54. Sequence Stratigraphic interpretation of the Villafranchian conglomerates. SB Sequence Boundary HST Highstand Systems Tracts.
Fig. 5.55. Chrono-stratigraphic diagram of the Coastal Rharb area. Tentative Sequence Stratigraphic interpretation. SB Sequence Boundary, MFS Maximum Flooding Surface, TST Transgressive Systems Tracts, HST Highstand Systems Tracts. Based on data from Tilloy (1955).
Fig. 5.6. Recent to Present sedimentary environments of the Rharb coastal area. Transgressions and regressions are evidenced by facies shifts. Inspired by data from Tilley (1955b).
Fig. 5.57. Dune ridges in the Atlantic coastal area of northern Morocco, 5 Km south of Rabat. HST Highstand Systems Tracts, TST Transgressive Systems Tracts. SB Sequence Boundary. Modified of Beaudet et al. 1969.
Fig. 5.58. Surface expression of Pleistocene sea level oscillations. Marie Feuillet Hospital (Rabat). Modified after Choubert (1965) and Stearns (1978). HST Highstand Systems Tracts, TST Transgressive Systems Tracts SB Sequence Boundary MFS Maximum flooding surface.
(Mellahian) transgression is represented by blue marine marls on the Tanger Bay and conglomerates on the beaches located along the shoreline of the Straits of Gibraltar (El Gharbaoui 1977).

According to Raynal et al. (1986) transgressions and regressions are indicated by the bio-rhexistase, that is stages of humid and dry climate. Some of these cold and warm cycles defined onshore coincide with major sequence boundaries in the offshore Rharb region. Analysis of the molluscs of coastal strand sediments reveals cold and warm stages. These stages can be radiometrically dated using the C14 method. High marine terraces reveal a sea level rise with respect to those that occupy a lower topographic position. Arid climate is associated with caliches and red-soils, common on top of aeolian dune-type deposits. The Maarifian stage, for example, is characterized by a conglomerate with large blocks that erode the Cretaceous. These sediments have a cold water fauna. They grade into sandy-limestone that represents aeolian dune deposits.

The Amirian II is coeval with the Lower Anfatian, and is represented by two marine levels G0 and G1 recognized by Biberson (1970). These levels contain cold water malacofaunas. The Middle Anfatian, represented by a coarse conglomerate, is equivalent to the transgressive maximum of this marine cycle. The malacofauna indicates a warm climate episode that can be correlated with the Middle Sicilian (0.6-0.47 Ma) (stages 15 to 13 in the isotope curve). The Upper Anfatian corresponds to the regressive part of the Anfatian cycle and the beginning of the Ouljian transgression (0.3 to 0.13 Ma). It is characterized by an erosional surface with caliches and corresponds to an important sea level drop during arid climate (rhexistatic stage) (inter-Tensiftian I-Tensiftian II=Pre-Soltanian). Littoral deposits are associated with Harounian-Agadirian transgression dated as 0.26 Ma. The Tensiftian II stage could be equivalent to the high marine level of Harounian-Rabatian in the area of Rabat dated as 145000 years BP. The marine Middle Ouljian is equivalent to the continental inter-Tensiftian-Soltanian dated as 0.13-0.10 Ma. It is represented by a
to 8 m high marine terrace. A cold water strand line malacofauna of is associated with this terrace level. This stage can be correlated with the last inter-glacier period and with the stage 5 of the isotopic curve. At 125000±10000 B.P., 120000±10000 B.P. and 110000±6000 B.P. soil levels with ferruginized surfaces (red paleosols were formed). The marine malacofauna indicates warm water conditions. The Upper Ouljian regressive interval is characterized by deposition of dune ridges topped by arid soils. These aeolian deposits are referred to as "sables beiges" and indicate an arid climate.

During Holocene time sea level reached a maximum at about 15000 years B.P. indicating the beginning of the Mellahian cycle. After 6500 years BP there is a maximum sea level high. This highstand situation characterizes the Middle Mellahian. Minor transgressive pulses occurred around 3500 and 2000 years BP. Three Holocene dune systems were recognized by Weisrock and Fontugne (1991).

In conclusion Plio-Pleistocene sea level changes have been observed or postulated by various authors. The precise correlation with the cycles observed on the seismic with the surface is unclear. Table 5.7 represents a first generic attempt to correlate land observation with seismic data. Cycles A-K are roughly compatible with data from the Gulf of Mexico (Wornardt and Vail 1991).
5.2. THE GUADALQUIVIR ALLOCHTHON.

The Prerifaine Nappe of Morocco is equivalent to the so-called "Guadalquivir allochthonous units" in Spain. The Guadalquivir allochthonous units occur in the Central and Western Betic Cordillera, extending westward into the offshore Gulf of Cádiz area. Classically, the Guadalquivir allochthon has been interpreted, like the Prerifaine Nappe, as an olistostrome (Perconig 1960-62) or a "melange" (Vidal 1977). Seismic and well data from the Gulf of Cádiz, supported by field observations from the External Betic Cordillera, show features and a geodynamic evolution that correspond to the Prerifaine Nappe, suggesting that both are part of a frontal accretionary wedge of the Gibraltar Arc.

5.2.1 GULF OF CADIZ

The Gulf of Cádiz is located offshore the Portugese and Spanish southwestern Atlantic margin and west of the Gibraltar Straits (Fig. 5.59, Table 5.8.). In the Gulf of Cádiz, old oceanic crust related to the initial stages of the opening of the Atlantic and very young Neogene structures related to the Gibraltar Arc co-exist (Maldonado and Snelson 1988). This region provides a uniquely complete section of the south-Iberian passive margin that underlies the accretionary wedge, showing several stages of the Mesozoic passive margin evolution and the superposed Neogene accretion related to the westward escape of the Alborán block. The structure of the Gulf of Cádiz was initially studied in the 1960s by seismic reflection, gravimetric and bathymetric profiles. Roberts (1970) recognized the main NE-SW structural trends (i.e. ridges and valleys) of the Gulf of Cádiz, relating them to halokinesis. A more detailed picture of the Gulf of Cádiz structure, based on seismic reflection profiles, was presented by Malod and Didon (1975). Baldy et al. (1977) and Malod and Mougenot (1979) mapped the main diapiric structures, folds and faults, and recognized flysch sediments in the eastern Gulf of Cádiz. According to these authors the structure of the region is the result of Late Miocene and Pliocene extensional
Fig. 5.59. Tectonic map of the Gibraltar Arc. Location of the Gulf of Cádiz area.
<table>
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<th>REGION</th>
<th>TIME</th>
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<th>OFFSHORE RHARB</th>
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Table 5.6. Major Tectono-stratigraphic events in the Gulf of Cádiz or the External Betic Cordillera
phases followed by a Pleistocene compressional phase. Overall knowledge about the structure of the Gulf of Cádiz has increased since gas discoveries were made in the area (Delaplanché et al. 1982).

5.2.1.1 Stratigraphy.

The stratigraphy for this study is based entirely on published wells (IGME 1987). The stratigraphic units of the Gulf of Cádiz can be subdivided, following the same nomenclature used for the Moroccan side, into: Infra-Nappe, Nappe and Supra-Nappe (Fig. 5.60).

Infra-Nappe.

In the Gulf of Cádiz area, the Infra-Nappe unit corresponds to the passive margin succession of southern Iberia. The stratigraphy and structure are similar to other Atlantic-type margins. Proceeding from bottom to top, a thick Triassic salt succession covered by shale, marl, anhydrite and gypsum in diapiric structures was penetrated by exploration wells and overlies the Paleozoic basement. Jurassic-Lower Cretaceous sediments are filling tilted blocks. Jurassic-Lower Cretaceous sediments consist of shallow-water limestones and dolomites and are overlain by Lower Cretaceous shales with interbedded chalk-limestones, marls and dolomites.

The Jurassic-Lower Cretaceous succession represents the prolongation of the Algarve-Huelva carbonate platform. The rotated blocks probably do not involve the basement and may be better compared to the rollers and growth faults reported from the Gulf of Mexico and other areas of the world (Bally 1981, Jackson and Talbot 1991). An unconformity with downlap separates these Jurassic to Middle Cretaceous rotated blocks from Upper Cretaceous to Middle Miocene prograding shelf margin prograding units.

The Upper Cretaceous consists of shale, siltstone and marl. Paleocene limestone-conglomerates and shaly and marly-limestone interbedded with limestones unconformably overlie the Upper Cretaceous succession. Eocene-Oligocene limestone, Lower Miocene
Fig. 5.60. Synoptic stratigraphic diagram of the Offshore Gulf of Cádiz area based on published exploration wells (IGME 1987).
(Aquitanian) sandstone and Middle Miocene interbedded marl and limestone topped by Langhian-Serravallian limestone represent the top of the Infra-Nappe succession.

The Upper Cretaceous to Middle Miocene shelf margin occurs below the basal foredeep unconformity and is referred to as a platform sequence. **Nappe. "Guadalquivir allochthonous units".**

The Guadalquivir Allochthon is mainly dominated by Triassic evaporites and redbeds while the Prerifaine Nappe is mainly composed of marls. Most of the wells of the Gulf of Cádiz area pierced strongly deformed Lower Miocene marls and allochthonous Triassic of the Guadalquivir unit. **Supra-Nappe and Foredeep units.**

**Foredeep**

A major erosional surface with canyons separates Lower Miocene shales from Middle-Upper Miocene detritic limestone and sandstone overlain by limestone, sandstone and siltstone. Tortonian-Messinian is represented by pyritic and glauconitic marl with interbedded micro-conglomerates and fine-grained sandstones. The Upper Miocene turbiditic sandstones of the Gulf of Cádiz are the prolongation of the Guadalquivir Basin turbiditic trough. The most characteristic facies are channel-fill and channel-overbank deposits (Delaplanché et al. 1982). Pliocene is represented by shale and siltstone with sandstone lenses and occasionally interbedded limestone.

The eastern part of the Gulf of Cádiz consists of structurally imbricated Cretaceous and Paleogene siliciclastic turbidites, limey-sandstones and marl of the flysch domain. Satellite basins filled with well-cemented Miocene conglomerates discordantly overlie Upper Eocene-Oligocene dark-brown pelagic marls (Maldonado 1983). Plio-Quaternary bioclastic sandstone and breccias with algae, mollusca and corals cover the entire region (Maldonado 1983).
5.2.1.2 Structure

Structurally the Gulf of Cádiz can be subdivided into a northwestern and a southeastern area separated by the leading edge of the frontal accretionary complex.

Northwestern Gulf of Cádiz

The structure of the northwestern region is characterized by southwestward-dipping normal faults offsetting Lower Cretaceous and Jurassic carbonates, showing significant rotation (Fig. 5.61). As mentioned earlier these structures probably do not involve the basement; they appear to be related to gravitational tectonics on a passive margin. Upper Cretaceous through Middle Miocene prograding shelf margin sediments represent the post-rift succession superimposed on the tilted syn-rift units. Locally Triassic diapirs intrude the passive margin succession. A widespread erosional surface with pronounced canyon incisions constitutes the basal foredeep unconformity in this region. Upper Miocene turbiditic sandstone, related to the initial deep-water stage of foredeep development, onlaps onto this unconformity. The offshore prolongation of the Jurassic Algarve platform and the main structural features of the overlying passive margin succession of this northwestern portion of the Gulf of Cádiz are shown in Fig. 5.61.

Southeastern Gulf of Cádiz

The southeastern area is occupied by the frontal accretionary wedge of the Gibraltar Arc (i.e. the Guadalquivir Allochthon). The internal structure of this unit is characterized by NW-vergent imbricates merging into a basal décollement. Toe-thrusts connected with extensional detachments and extensional basins ride on top of the Nappe. Supra-Nappe Upper Miocene and Pleistocene sediments onlap onto the accretionary wedge.

The main tectonic features of the Gulf of Cádiz can be depicted along several transects (see Panel 4).
Fig. 5.61. Cross-section of the South-Iberian passive margin in the Gulf of Cadiz. J-LK Jurassic-Lower Cretaceous. UK-UO Upper Cretaceous-Upper Oligocene LM Lower Miocene, MM Middle Miocene, UM Upper Miocene PL-PL Plio-Pleistocene. BFU Basal foredeep unconformity.
Section A

This is an E-W trending strike section along the northern part of the Gulf of Cádiz. The section displays a major unconformity characterized by canyons that dissect Lower-Middle Miocene and Cretaceous sediments. This erosional surface is perhaps associated with the basal foredeep unconformity and is overlain by Upper Miocene to Pleistocene deep-water foredeep units.

Section B

This N-S trending section displays rotated Jurassic to Lower Cretaceous blocks overlain by prograding shelf margin units. Southward-dipping listric normal faults form the boundary of rotated Lower Cretaceous blocks which are covered by a thick Upper Cretaceous to Middle Miocene post-rift succession characterized by clinoforms that suggest a southward progradation. The top of the Middle Miocene is a major unconformity and represents the basal foredeep unconformity in the shelf margin area. The onlapping Upper Miocene and Plio-Pleistocene constitute the initial filling of the foredeep.

Section C

This northward trending section shows the front structure of the wedge. A set of imbricates with southward-dipping thrusts constitutes the main structural feature of the complex. The wedge is thrust onto the distal platform succession of the southern Iberia margin. Supra-Nappe sediments directly overlie folded Infra-Nappe reflectors.

Section D

This N-S oriented section illustrates the accretionary wedge overlying southward-dipping Lower-Middle Miocene sediments. The infra-Nappe succession includes strongly rotated Jurassic and Lower Cretaceous blocks. The accretionary wedge consists of southward-dipping thrust faults merging into a basal décollement. Upper Miocene sediments onlap onto the top of the accretionary wedge and the basal foredeep
unconformity.

Section E

This E-W oriented section shows the geometry of the south-Iberian passive margin units. Lower Cretaceous rotated blocks are covered by Upper Cretaceous to Lower Miocene shelf margin sediments. The angular unconformity between eastward-dipping Middle Miocene strata and the onlapping horizontal Middle Miocene to Pleistocene corresponds to the basal foredeep unconformity.

Section F

This section is composed of a N-S and a E-W portion. The most conspicuous feature shown on this section is the contact between the frontal part of the wedge and a Triassic salt diapir. The N-S portion of the section displays southward-dipping listric normal faults and a significant rotation of the Lower Cretaceous unit. The overlying Upper Cretaceous to Middle Miocene post-rift succession is pierced by a Triassic salt diapir. The structure of the accretionary wedge consists of steeply eastward-dipping thrust faults merging into a gentle eastward-dipping basal décollement. Occasional normal faults and related extensional satellite basins are superimposed on the upper part of the wedge.

5.2.2. EXTERNAL BETIC CORDILLERA

The "Guadalquivir allochthonous units" are exposed in the southern margin of the Guadalquivir Basin and in the frontal ranges of the Western and Central Betic Cordillera, terminating with the Tiscar fault, west of Sierra de Cazorla (see Fig. 5.59). The eastern termination of the Guadalquivir Basin coincides with the eastern edge of the accretionary wedge. The northernmost part of the accretionary complex is not exposed, because it is covered by the Neogene of the Guadalquivir Basin. Perconig (1960-1962) initially named this unit for the region of Sevilla as "Nappe de glissement" or "olistostrome". The
Guadalquivir allochthon is characterized by a mixture of strongly deformed sediments overlain by Neogene satellite basins. The frontalmost part of the allochthon is covered by Neogene sediments of the Guadalquivir Basin, here referred to as Supra-Nappe. They are equivalent to the upper part of the foredeep succession and were extensively described in Chapter 4.2.

5.2.2.1 Stratigraphy

The lithology and age of the sediments involved in the Guadalquivir allochthonous units are highly variable, and it is difficult to establish a general stratigraphy. Strong deformation, synsedimentary tectonics and lateral facies changes throughout the region prevent a good stratigraphic correlation. As in the Prerifaine Nappe, the stratigraphy is subdivided into: Infra-Nappe, Nappe and Supra-Nappe (Fig. 5.62).

Infra-Nappe.

The Infra-Nappe succession is locally exposed in tectonic windows associated with post-Nappe domes or anticlines. Jurassic through Cretaceous platform and basinal facies of the south-Iberian margin constitute the Infra-Nappe succession in the northern part of the Guadalquivir Basin. In the southern portion of the basin, the only information on the Infra-Nappe is provided by exploration wells. The Lower Tortonian-Serravallian sandstone of the "Arenas basales" unit underlies the base of the allochthon (see Betic foredeep Chapter 4). This basal sandstone also directly overlies the basal Late Paleozoic unconformity (i.e. Hercynian unconformity) in the northern Guadalquivir Basin.

Nappe.

The Guadalquivir Allochthon consists of a mixture of Triassic, Cretaceous, Paleogene and Neogene sediments, often referred to as a melange or olistostrome. The Guadalquivir allochthon involves a relatively much larger volume of Triassic shales and evaporites than the Prerifaine Nappe. Even though the sediments involved in the chaotic
Fig. 5.62. Chronostratigraphic diagram of the Guadalquivir Allochthon in the Central External Betic Cordillera. Data from geological maps of IGME and own data.
structure of the Nappe do not follow any stratigraphic order, they will be described from older to younger.

**General distribution and thickness**

The rocks of the Guadalquivir allochthon are chaotic and therefore do not show a simple distribution pattern. However in general the sediments become younger in the northward direction (Fig. 5.63). The original thickness of these deposits is unknown, as it is difficult to find a complete section. Only partial sections bounded by decollement levels are described from the Guadalquivir region (Baena and Jerez 1983).

**Triassic**

The oldest sediments present in the Guadalquivir allochthonous units consist of well-bedded micritic limestone and dolomite of the Muschelkalk facies (Germanic-type Triassic). Iron mineralizations, referred to locally as "Ocres rojos de Jaén", are associated with this limestone. Triassic shales, evaporites and marls, containing ophites i.e. diabase are interpreted to be of the Germanic Keuper facies. The Triassic sediments occupy widespread cultivated areas and good exposures are rare. Dolerites and Muschelkalk carbonates constitute small isolated massifs or blocks surrounded by shales and evaporites. Triassic sediments are frequently in contact with strongly deformed Cretaceous to Eocene pelagic marls and limestones.

**Lower Cretaceous**

Interbedded pelagic white limestone and marly-limestone. Biomicritic limestone and marly-limestone with abundant ammonites. Bourgois (1978) recognized Valanginian, Hauterivian and Barremian faunas in these sediments.

**Upper Cretaceous- Paleocene**

A discontinuous level of ammonite-bearing Neocomian light grey marl overlies Triassic sediments. Conformably overlying the Neocomian marl are Barremian-Aptian
Fig. 5.63. Idealized block diagram illustrating the age distribution of sediments in the frontal part of the Guadalquivir Allochthon. Ages of allochthonous units are the youngest beds involved in the Pre Supra-Nappe deformation.
dark red sandy-limestone, sandstone and marl with *Orbitolina sp.*. The Senonian and Paleocene consists of well-bedded reddish marly-limestone and marl of the pelagic "Capas Rojas" facies.

**Eocene-Oligocene.**

The Eocene-Oligocene succession is represented by alternating green marls and marly limestone, green or reddish shales and marls with interbedded bioclastic sandstones of turbiditic character ("flysch"). These facies are similar to the Paterna Formation of Chauve (1968) and to the Paleogene succession of the flysch domain in the Campo de Gibraltar area.

**Miocene**

The Miocene succession is very uniform and widespread throughout the Guadalquivir Basin. Radiolaria and diatom-bearing marl and marly-limestone with occasionally interbedded sandstone and bioclastic limestone constitute the dominant lithology. The age of these sediments ranges from Aquitanian to Tortonian. It is difficult to distinguish the age of the sediments in the field without paleontologic analysis. These facies were initially known locally as "Albarizas" and later as "Moronitas" after the village of Morón de la Frontera (Calderón and Paul 1886).

**Aquitianian-Lower Burdigalian**

- **Lithology:**
  
  White marl and marly-limestone with occasionally interbedded limestone consisting of silicified wackestones and packstones, and sandstone.

- **Paleontology:**
  
  **Middle Aquitanian.**
  
  *Globigerinoides primordius, Turborotalia semivera, Globquadrina dehiscens.*

  **Upper Aquitanian.**
  
  *Globigerinoides gr. trilobus, Globigerina brazieri*
Lower Burdigalian.

Globigerinoides altiaperurus

G. subquadratus and Globigerina euapertura upper part of Lower Burdigalian.

Upper Burdigalian-Serravalian

- Lithology:
  Limestone and bioclastic limestone.

- Paleontology:
  Upper Burdigalian
  Globigerinoides sicanus, Globorotalia praescitula, Globorotalia acrostoma.

  Langhian
  Praeorbulina glomerosa, Globigerinoides sicanus, G. bulloides, Praeorbulina transitoria, Globorotalia praemanardii, Orbulina universa.

  Lower Serravalian
  Lack of Globorotalia menardii

  Middle Serravalian
  Globorotalia menardii and Globigerinoides subquadratus.

  Upper Serravalian
  Presence of Turborotalia siakensis lack of Globigerinoides subquadratus.

Tortonian:

- Lithology
  White-grey marl with sandstone intervals and white marl referred to as Albarizas s.s.

- Paleontology
  T. acostaensis, T. humerosa, T. pachyderma, Globorotalia plesiotumida, Globorotalia dalii.
Environment

Curtó and Matías (1986-87) distinguished five nannofacies within the Miocene sediments of the Guadalquivir Allochthon: aciculate diatoms, biogenic, coccolithic-detritic, spiculithic-coccolithic-detritic and detritic. The deposits form a rhythmically alternating diatomitic and biogenic laminae that developed in an anoxic depositional environment due to local upwelling.

Supra-Nappe units.

The Supra-Nappe stratigraphy was already described in the foredeep section (Chapter 4). Only a few points need to be re-emphasized here.

Upper Tortonian-Messinian

The Upper Tortonian-Messinian succession is characterized by a coarsening upward sequence of channelized conglomerates interfingering with sandstone, siltstone and yellow marl, representing a prograding fan-delta sequence. Conglomerates are composed of fragments or grains of Jurassic and Cretaceous provenances. This succession is topped by Messinian burrowed bioclastic cross-bedded quartzy sandstone with abundant lamellibranchs. This sandstone is rich in Triassic shale fragments.

Pliocene

The Pliocene is lithologically very variable, consisting of blue marl, sandy-marl, sandstone, siltstones, laminated limestone and poligenic conglomerates rich in quartz, potassium feldspar and limestone fragments. Some of these deposits represent lacustrine sediments grading into marginal coarse breccias.

5.2.2.2. The transition with the flysch domain: The Paterna unit.

Chauve (1962) defined the Paterna unit as a transitional unit between the Subbetic units and the so-called flysch units. This widespread unit of the Guadalquivir area is
equivalent to other units described by several authors (Table 5.9).

<table>
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<td>Chauve (1962)</td>
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<tr>
<td>Didon (1969)</td>
<td>Arcillas de Jimena</td>
</tr>
<tr>
<td>Didon (1969)</td>
<td>Benaiza Formation</td>
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<tr>
<td>Cruz SanJulian (1974)</td>
<td>Guadalteba Formation</td>
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<tr>
<td>Bourgois (1978)</td>
<td>Arcillas con bloques</td>
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Table 5.7. Nomenclature of Paterna unit facies.

The Paterna unit is exposed in tectonic windows underlying allochthonous Triassic of Subbetic affinity or as part of imbricates with these Triassic sediments (Fig. 5.64). Part of these sediments is incorporated in the Guadalquivir Basin. The Paterna unit consists of Cretaceous shales and marls with blocks of mixed lithologies and has a chaotic and very complex structure. The shaly matrix, referred to locally as "Arcillas variegadas", consists of variegated marls and shales with Tobotomaculum (Martín-Algarra 1987). These reddish clays and shales are characterized by scaly cleavage, flattening of fossils, resedimented fauna (i.e. mixture of fossils from Upper Cretaceous to Oligocene) and the lack of carbonates. In the shales of the Paterna unit (i.e. variegated shales) foraminifera and nannoplankton are sometimes dissolved, suggesting deposition below the carbonate compensation depth (CCD). Allochthonous blocks of several lithologies detached from the other units are present within these scaly clays. These clays appear to be similar to the "Argille Scagliose" of the Apennines (Martín-Algarra 1987). Blocks of several sizes and lithologies are found in the complex. Triassic marls and marly-limestone, sandstone and breccias, "numidoid" shales with interbedded lenses, dismembered beds or blocks of
Fig. 5.64. Stratigraphic diagram of the Aljibe and Paterna units. Based on IGME geologic maps. Tk Triassic Keuper, UK Upper Cretaceous, LM Lower Miocene and N Numidian Sandstone.
Numidian-type sandstone are the most common blocks present within the scaly clay matrix.

The sediments of the Paterna unit constitute the base of the Aljibe unit, which consists of Aljibe sandstone, and it is very well exposed in the Campo de Gibraltar area (Gibraltar Peninsula). The Numidian Aljibe-type sandstone partially onlaps and also is incorporated as blocks of several sizes, suggesting that part of the complex is Aquitanian in age.

5.2.2.3 Structure

Few studies have been done on the detailed structure of the Guadalquivir allochthon. However, some geological maps at scale 1:500,000 done by the "Instituto Geológico y Minero de España" (IGME) provide useful information on the structure of the region. This frontalmost part of the Betic Cordillera has been classically interpreted as olistostrome or melange, based on the chaotic nature of the region. In the Western Betic Cordillera Bourgois (1978) defines this region as a tectono-sedimentary complex "Complexe du Guadalquivir" formed by three geologic units:

1. Triassic breccia unit referred to as "brèches polygéniques à ciment gypseux du Trias germano-andalou" (polygenic breccias with gypsum cement of Germanic-Andalusian type Triassic).

2. The Subbetic unit.

3. Scaly clays with blocks, referred to as "Arcillas a bloques".

Because these units can be statistically found in all possible combinations, the Guadalquivir Complex is considered by Bourgois (1978) as a perfect "Mélange".

The geologic maps of the region, combined with field work for this study, reveal some structural features within this chaotic assemblage of rocks. The accretionary complex includes part of the classical Subbetic and Circumbetic units of Baena and Jerez (1982) and the classical Guadalquivir allochthonous units. The Guadalquivir Allochthon is
dominated by allochthonous Triassic shales and evaporites (Fig. 5.65). The structure of this frontal region of the Betic Cordillera is characterized by multiple décollement tectonics defined by three main detachment levels: Triassic, Upper Aptian-Lower Cenomanian and Middle Eocene (Baena and Jerez 1982). Imbricated thrusts, ridges, normal faults, low-angle extensional detachments, slumps and olistostromes are common features within the nappe (i.e. Guadalquivir Allochthon). Neogene satellite basins characterized by pelagic sediments ride piggy-back on the nappe. In the Central Betic Cordillera (region of Porcuna) Roldan-García and Rodríguez-Fernández (1991) recognized thrust structures within the frontal part of the Guadalquivir unit, overlain by piggy-back basins in the sense of Ori et al. (1986).

Ages of emplacement

Thrusting at Middle Serravallian-Early Tortonian time is coeval with the inception of the foredeep. In the Western Betic Cordillera the age of emplacement of the allochthon is intra-Tortonian (Perconig 1960-62, Chauve 1967). Bourgois (1978) recognized the diachronic emplacement of the Guadalquivir Allochton, ranging from Middle Miocene, (referred to in the past as Helvetian) in the eastern Guadalquivir area to Upper Miocene Tortonian in the Western Guadalquivir.

5.2.2.4. Selected field examples

Field work was done in selected key areas of the Guadalquivir Basin where the Guadalquivir Allochthon is well exposed (Fig. 5.66). These local field observations show similar to those described from seismic sections, supporting the accretionary wedge model for the Guadalquivir allochthonous unit. The field examples from the Western and Central external Betic Cordillera show the following: 1. Internal deformation within the nappe, 2. Thrusts within the nappe, 3. low- and high-angle extensional detachments, 4. Ridges, 5. Olistostromes and 6. the Supra-nappe / nappe contact.
Fig. 5.65. Cross-section through the Western Betic Cordillera based on well log data extracted from IGME (1987). The section displays the superposition of Allochthonous Triassic over the autochthonous Triassic and Cretaceous of the south-Iberian margin.
Fig. 5.66. Location of the main regions of field observations along the Guadalquivir Allochthon presented in this study.
1. Internal deformation within the nappe

The most common macro-structures affecting the sediments of the nappe are scaly cleavage, s-c shear structures, folds and duplexes. These structures account for the internal deformation and thickening of the wedge (see Fig. 5.67).

2. Thrusts within the nappe.

Thrust faults, frequently defining imbricates, were observed in several spots throughout the Guadalquivir area. The internal structure of the Guadalquivir Allochthon consists mainly of northwest vergent imbricates emerging from several décollement levels. Slickensides, scaly cleavage and shear zones with related s-c fabrics are common structures on fault planes. These steeply-dipping imbricated thrust systems represent the frontal imbricates of the accretionary wedge. Some examples of Triassic-Cretaceous imbricates are shown in figures 5.68 to 5.72.

3. Low and high angle extensional detachments

Several tectonic contacts are characterized by stratigraphic omissions that suggest extensional faults. Normally there are no microstructures, indicating a sense of displacement associated with these contacts. The sediments on the hangingwall of these low-angle extensional detachments are strongly rotated and show roll-over geometries. The most common low-angle extensional detachments observed in the External Betic Cordillera separate Cretaceous marl and marly-limestone from strongly deformed Triassic evaporites (Figs. 5.73 to 5.75).

4. Ridges

East of Jerez de la Frontera (external Western Betic Cordillera), Triassic shales and evaporites are overlain by steeply deeping Neogene sediments and are offset by normal faults (Fig.5.76). This Triassic ridge or culmination represents the onshore prolongation of the Gulf of Cádiz diapirs.
Fig. 5.67. Internal structure of the Guadalquivir Allochthon in the region of Huelma (Central Betic Cordillera). A Triassic blocks deformed within a matrix of Cretaceous marls. B Folded level of Triassic blocks. C Scaly cleavage and shear planes affecting Cretaceous pelagic marls. D. s-c shear fabrics and scaly cleavage affecting Cretaceous marl.
Fig. 5.68. Triassic and Cretaceous Imbrications. 2 Km west of Huelma (Jaén) (Central Betic Cordillera).
Fig. 5.69. Foreland-vergent thrust system in the Guadalquivir Allochthon. Eastern Jodar region (Jaén) (Central Betic Cordillera).
Fig. 5.70. Triassic and Cretaceous Imbricates. Torrecera (Cádiz, Western Betic Cordillera).

Fig. 5.71. Imbricates of Triassic Muschelkalk and Cretaceous sediments in the region of Paterna (Cádiz, Western Betic Cordillera).
Fig. 5.72. Imbrications of Triassic and Cretaceous sediments of the Paterna unit in the Cortijo de las Piletas (Cadiz, Western Betic Cordillera).
Fig. 5.73. Low-angle extensional detachment between Triassic Keuper (Tk) and Cretaceous (K) in Paterna de Rivera (Cádiz, Western Betic Cordillera).

Fig. 5.74. Cross-section Jéduła-Arcos de la Frontera (Cádiz, Western Betic Cordillera). The cross section shows low- and high-angle extensional faults and a Triassic ridge.

Mu Upper Miocene sandstones MI Lower Miocene marl K Cretaceous marl Tk Triassic shales and evaporites
Fig. 5.75. Low-angle extensional detachment between Triassic and Cretaceous sediments. (Arcos de la Frontera, Western Betic Cordillera). Tk Triassic Keuper (shales and evaporites), Tm Triassic Muschelkalk (limestones) and Ku Upper Cretaceous (pelagic marl).

Fig. 5.76. Extensional fault offsetting Cretaceous shales with blocks of Numidian Sandstone and Triassic evaporites of the Paterna unit (Medina Sidonia, Cádiz) (Western Betic Cordillera). Tk Triassic Keuper (shales and evaporites) K Cretaceous marls and shales with blocks of Numidian Sandstone.
Fig. 5.77. Map and cross-section of a Triassic Ridge. East of Jerez de la Frontera (Cádiz, Western Betic Cordillera). Modified from IGME Mapa geológico 1:50,000 Hoja de Jerez de la Frontera. Tk Triassic Keuper, N Neogene.
5. Olistostromes and slumps

Throughout the most frontal parts of the External Betic Cordillera, chaotic mélange or marls ranging in age from Cretaceous to Miocene are often difficult to separate cartographically. Blocks of Triassic shales and evaporites, Jurassic oolitic or algal limestones, Cretaceous pelagic marls and marly-limestones and Tertiary sandstones and marls are the main type of blocks present within the nappe. Fig. 5.78 shows an example in the region of Osuna of Jurassic olistolithes embedded in a marly matrix overlying allochthonous Triassic sediments.

6. Contact Nappe-Supra-Nappe

The contact between the nappe and the Supra-nappe units is well exposed only in a few areas of the Guadalquivir Basin. One is the Puente-Genil area (Córdoba), located in the External Central Betic Cordillera, southwest from Córdoba (see fig.5.64 for location). The contact between the accretionary wedge and the Supra-Nappe sediments is very well exposed near the village of Puente-Genil. In this region the accretionary complex (i.e. the Guadalquivir allochthonous units) consists of Triassic shales with interbedded evaporite and micritic limestone and Aquitanian-Middle Tortonian white pelagic marl (Fig. 5.79). The Miocene succession is strongly deformed and detached from the underlying Triassic (Fig. 5.80). The Supra-Nappe succession consists of Upper Tortonian sandy-marl and Messinian bioclastic sandstone. The Supra-nappe sediments onlap onto the underlying deformed units of the accretionary complex. Normal faults offset the top of the accretionary complex and the overlying sediments (Fig. 5.80).

Since the outcrops in the Guadalquivir Basin are rare and poor only few examples were selected. They illustrate partial aspects of the structure and style of deformation of the accretionary wedge. These sketches are presented after the seismic data, so the reader can compare the surface and subsurface expression of the wedge and also can integrate
Fig. 5.78. Surface expression of olistostromes in the Guadalquivir Allochthon. Hacienda del Soldado. (Osuna, province of Sevilla, Central-Western Betic Cordillera). After IGME Mapa geológico 1:50,000 Hoja de Osuna.
Fig. 5.79. Surface expression of the Nappe/Supra-Nappe contact in the area of Puente Genil (Cordoba, Central-Western Betic Cordillera). Modified from IGME (1985) Mapa Geologico de Espana 1:50,000 Hoja de Puente-Genil.
Fig. 5.80 Cross-sections of the Nappe / Supra-Nappe contact in Puente Genil. See Fig. 5.79 for location.
them into the big picture of the accretionary wedge provided by the seismic data. Most of the structures are common in many ancient and present day accretionary wedges.
CHAPTER 6
DISCUSSION

In this chapter I will address several aspects related to the structure and evolution of the frontal accretionary wedge of the Gibraltar Arc. The frontal part of the Gibraltar Arc is an accretionary complex related to the westward escape of the Alborán block. The inner part of the wedge includes turbidites (Flysch Zone), but the most frontal part also involves a passive margin succession of southern Iberia and northern Africa. Southern Iberia is a carbonate margin and northern Africa is a siliciclastic margin. A large volume of Triassic evaporites, more abundant in the Betic Cordillera than in the Rif, is present within the accretionary complex (see Fig. 5.63).

The relationship between the accretionary wedge and the underlying units, the evolution of the wedge, and a comparison with other accretionary wedges will be discussed in this chapter.

6.1. THREE-DIMENSIONAL DIAGRAM OF THE ACCRETIONARY WEDGE.

Seismic, well and surface data in the frontal region of the Gibraltar Arc were integrated to construct a three-dimensional diagram of the accretionary wedge (Fig. 6.1). The accretionary zone was subdivided into the Infra-Nappe, the Nappe and the Supra-Nappe. The Infra-Nappe includes sediments underlying the accretionary prism. The structure of these units is characterized by foreland vergent duplex structures (Infra-Nappe Imbrications) detached from the acoustic basement. Locally steep reverse faults offset these units and suggest a thickness change of the Infra-Nappe sediments. In some areas inverted Triassic half-grabens were postulated. The seismic data suggest that the Infra-Nappe units
Fig. 6.1. 3D block-diagram of the accretionary wedge.
were emplaced after and underneath the accretionary wedge; this aspect will be discussed in section 6.2.

The wedge s.s. or the nappe itself has a complex structure (Fig. 6.1). The most characteristic structural features are: Extensional detachments, Ridges, Frontal Imbricates, Slumps.

**Slumps** are located at the toe of the frontal slope of the wedge. These submarine slides are bounded by a shallow low-angle extensional detachment that can locally merge into small-scale thrusts. The distribution of the slumps is controlled by the slope of the wedge (for examples see plates 10, 12, 14, 16, 18, 20, 22, 24 and 26 and section E of panel 3).

**Frontal Imbricates** consist of foreland-vergent closely-spaced steeply-dipping imbricates emerging from the basal detachment of the nappe. They are underlain by the Infra-Nappe Imbricates and locally overlain by slumps (for example see plates 2,4 and 10 and panels 1, 3 and 4).

**Toe-thrusts** are compressional-extensional structures consisting of a normal fault that merges into a thrust. These features are parallel to the main structural trends and indicate that extensional and compressional structures are coeval (for example see plates 3, 20, 22, 24 and 26, sections R3, R5 and R7 of panel 1, sections D and E of panel 3, section F of panel 4).

**Ridges** are elongated culminations of the wedge due to the plasticity of sediments involved in the accretionary prism sediments. Ridges trend nearly parallel to main thrusts or extensional detachments. Some ridges are caused by shale diapirism (for example see plates 24 and 26 and sections B, D, E and H of panel 3).

**Extensional detachments** often offset the uppermost part of the wedge in the rear of the accretionary wedge (for examples see panels 2 and 3).

The Supra-Nappe units overlie the accretionary wedge and mainly consist of satellite basins and prograding units.
Satellite Basins are small basins that ride piggy-back on the accretionary wedge and are bounded by extensional and/or compressional faults. Extensional satellite basins are deeper and do not show a systematic orientation. They can trend parallel or perpendicular to the main thrusts. Some satellite basins are controlled by toe-thrusts and therefore are considered "mixed" extensional-compressional basins. Other satellite basins are related to thrusts and can be considered as classical "piggy-back basins". The significance and implication of these basins for the geology of the folded belt will be discussed in section 6.3.

Prograding units consist of Plio-Pleistocene shelf margin siliciclastic wedges that prograde and downlap onto the accretionary prism and the satellite basins. These progradations as mentioned in section 5 reflect glacio-eustatic sea level oscillations as well as the late geodynamic evolution of the orogen, as will be discussed in section 6.7.

The structure of the accretionary wedge of the Gibraltar Arc will also be compared (see section 6.5) with present-day accretionary wedges to show the similarity of megascopic and regional structures, as well as deformation processes.

6.2. RELATIONSHIP BETWEEN THE ACCRETIONARY WEDGE AND THE UNDERLYING UNITS.

A look at the tectonic map of the Gibraltar Arc shows that the accretionary wedge (i.e. Guadalquivir Allochthon and Prerifaine Nappe) overlies structural units of the external domain. The Betic and the Rif foredeeps are partially filled by this frontal accretionary complex. The classic explanation was in terms of olistostromes of more internal units emplaced by gravity gliding into the frontal part of the Arc. However, seismic and field data in the Betic and Rif Cordilleras suggest that an accretionary wedge
is deformed from underneath due to late thrusting that involves the underlying passive margin successions of northern Africa and southern Iberia.

6.2.1. CONTACT OF THE ACCRETIONARY COMPLEX WITH THE MAGHREBIAN DOMAIN.

Seismic, well and field data were integrated to characterize the structure of the Maghrebian domain and its contact with the overlying accretionary complex. The most spectacular example is offered by the Rides Prerifaines, where Jurassic thrust sheets are emplaced and displace the overlying accretionary complex. (Fig. 6.2)

In the Central Rif (Had-Kourt-Teroual region) (Fig. 6.3) two main structural units can be recognized:

**Lower Imbricates.** They are characterized by foreland-vergent thrust sheets made up of Triassic and Jurassic carbonates and a thin Lower Cretaceous section, equivalent to the Mesorif and Prerifaines Rides.

**Upper Imbricates.** They consist of Cretaceous and Eocene marl and siliciclastics characterized by multiple decollement tectonics, resulting in close-spaced imbricates.

Fig. 6.3 shows the structural map of the Had-Kourt-Teroual area to the east of the Rharb Basin. This tectonic map was modified from the map sheets of Had-Kourt-Khenisset and Terwall-Oulad Aïssa (Serv. Geol. du Maroc 1984, 1990) and shows the surface expression of the Lower and Upper Imbricates and the overlying Prerifaine Nappe, which constitutes the uppermost structural unit of the region. Piggy-back basins filled with Upper Cretaceous to Miocene sediments (the Ouezzane Unit) or else only Tortonian to Pliocene sediments overlie the main thrust sheets. Lower Miocene Zoumi sandstones fill additional satellite basins that overlie both the upper imbricates as well as the Ouezzane unit. Three interpretations of seismic sections and one surface section show the structure of the Had-Kourt / Teroual region and suggest a sequence of thrust emplacement.
Fig. 6.2. The Rif foredeep and the Prerifaine Nappe disrupted by the Prerifaine Rides
Fig. 6.3. Structural map of the Had Kourt-Teroual area. Modified from Serv. Geol. du Maroc (1984) and (1990). 1:50,000 scale maps of Had-Kourt and Teroual.
SECTION HT 1 (Fig. 6.4)

High amplitude reflectors at 3 sec represent the top of the Jurassic carbonates that dip towards the NE and can be followed from the outcropping Rides Prerifaines to the southern border of the Mesorif. According to data from well BB1 (Société Chérifienne des Pétroles 1952) located in the Ouerrha Valley (see Fig. 6.3 for location), Cretaceous and Eocene imbricates with a basal Triassic breccia overlie Jurassic carbonates. This allochthonous sheet of mixed Triassic, Cretaceous and Eocene sediments ("melange") constitutes the Prerifaine Nappe. The nappe directly overlies Jurassic carbonates in the north and Lower Miocene sediments in the south (Prerifaine Ridges area). Tortonian and Messinian sediments unconformably overlie the Prerifaine Nappe and are furthermore folded by the underlying compressional structures. On top of these sediments lies the "Ouezzane Nappe". In my view this unit is not a nappe but a set of synformal satellite basins filled with Eocene, Paleocene and Lower-Middle Miocene materials. On the northwestern end of the section, the Jurassic imbricates reach the surface to define the southern boundary of the Mesorif unit of Suter (1965).

HT2 (Fig. 6.5)

Section HT2 is based only on surface data from the map of Terwall-Oulad Aïssa (1990) Service Geol du Maroc, 1:50,000. This section shows the superposition of décollement levels. The Lower Imbricates, made up of Jurassic carbonates, are overlain by Cretaceous Upper Imbricates. The Upper Imbricates are overlain by the Prerifaine Nappe which constitute a melange with Triassic matrix. The structure of the area suggests piggy-back thrust propagation.

Section HT3 (Fig. 6.6)

Section HT3 shows the following structural units from bottom to top: Lower Imbricates, Upper Imbricates and Prerifaine Nappe. Mainly Eocene and Miocene sediments constitute basins on top of the Prerifaine Nappe (Ouezzane Unit). In the central
Fig. 6.4. Line drawing from seismic section HT-1. See Fig. 6.3 for location.
Fig. 6.5. Sketch cross-section HT-2 based only on surface data modified from (Serv. Geol. du Maroc 1990, geologic map 1,50,000 Terwall-Oulad Aissa).
Fig. 6.6. Line drawing of seismic section HT-3. See Fig. 6.2 for location.
part of the section a Triangle Zone defines the contact between Neogene and Cretaceous sediments.

Section HT4 (Fig. 6.7)

Section HT4 is a strike line along the region that shows the imbrications of the Prerifaine Nappe and its relationship with the underlying units, suggesting again a piggy-back emplacement sequence. The section also shows Neogene basins and the Ouezzane Unit overlying the Prerifaine Nappe.

6.2.2 CONTACT OF THE ACCRETIONARY COMPLEX WITH THE SOUTH IBERIAN DOMAIN.

According to seismic and field observations, the Guadalquivir Allochthon was deformed by the underlying southern Iberian platform units. Three areas have been selected to illustrate the relationship between the accretionary complex and the underlying units:

6.2.2.1 Jodar Region (Central Betic Cordillera).

6.2.2.2 Sierra de Estepa-Osuna (Western Betic Cordillera).

6.2.2.3 Sierra de Esparteros (Western Betic Cordillera).

6.2.2.1. Jodar Region (Central Betic Cordillera)

The Prebetic structure in the Jodar-Mancha Real region consist of domes and NE-SW oriented folds and thrusts (Figs. 6.8 and 6.9). The overlying Guadalquivir allochthonous units and the Subbetic thrust sheet are deformed by the underlying Prebetic domes, suggesting that the accretionary complex was emplaced prior to the underlying doming (Fig. 6.9). Field data support a piggy-back sequence of emplacement of the Guadalquivir Allochthon, the Subbetic, and finally the underlying Prebetic. Steeply dipping shear zones indicate south-vergent thrusting of the Guadalquivir allochthonous unit (i.e. accretionary complex) onto the Prebetic unit (Fig. 6.9).
Fig. 6.7. Line drawing of seismic section HT-4. See Fig. 6.2 for location.
Fig. 6.8. Structural map of the Jodar area. Central Betic Cordillera.

**Guadalquivir Alloclithon**: Tr Triassic, K-M Cretaceous to Miocene

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*SI*, South Iberian domain
Fig. 6.9. Cross-section of the Sierra de Jodar. A General cross-section. B Shear zone related to the contact between the Guadalquivir Allochthon and the Infra-Nappe succession. C. Structural reconstruction sketch.
6.2.2.2. Sierras de Estepa-Osuna (Western Betic Cordillera).

The Sierra de Estepa is located in the Western Betic Cordillera, south of the Guadalquivir Valley (Fig. 6.10). The structure of this region is characterized by a Subbetic anticlinorium overlain by the Guadalquivir allochthonous Triassic and Miocene. In map view the Jurassic and Cretaceous Subbetic units are completely surrounded by allochthonous Triassic. Triassic shales and evaporites and Miocene marl constitute the main lithologies of the Guadalquivir Allochthon. The structure of this area is coherent with an early emplacement of the accretionary complex, followed by emplacement of the underlying Subbetic.

6.2.2.3. Sierra de Esparrerios area (Western Betic Cordillera).

The Sierra de Esparrerios is located in the Western Betic Cordillera, southwest of Morón de la Frontera (Fig.6. 11). The Sierra de Esparrerios is an NE-SW trending Jurassic anticline characterized by a northern steeply-dipping thrust. Jurassic is thrust onto Pliocene lacustrine breccias, revealing post-Pliocene tectonics. In map view, Jurassic is surrounded by Miocene marl of the accretionary complex.

6.3. SUPRA-NAPPE SATELLITE BASINS: THE OUEZZANE AND RELATED UNITS.

The three-dimensional diagram presented here is inspired by the seismic data and has some important implications for the geology of the Betic and Rif Cordilleras. The data presented in this thesis suggest that several seismic units classically referred to as "unites flottantes" or "Nappes rifaines superieures" (upper thrust sheets) (Wildi 1983) are in fact satellite basins overlying the accretionary wedge. The Ouezzane Unit of the external Western Rif is the most obvious example of a complex of satellite basins (Fig. 6.12 and Fig. 6.1). In the so-called flysch domain, the Numidian Unit represents the highest structural unit located above the underlying imbricates and may also represent satellite
Fig. 6.10. Contact between the allochthonous Triassic of the Guadalquivir Allochthon and the underlying passive margin units. Osuna-Estepa region (Sevilla) (Central-Western Betic Cordillera).
Fig. 6.11. Geologic map of the Sierra de Esparteros. Modified from IGME Mapa geológico 1:50,000 Hoja de Morón de la Frontera. Detail of the Jurassic-Pliocene contact.
Fig. 6.12. Structural map of the Western Rif integrating seismic data and surface data modified from Suter (1980b). Location of the Ouezzane-type satellite basins.
basins. The Numidian Unit consists of a set of turbiditic basins riding piggy-back on the accretionary wedge. These basins were deformed later.

In the Betic Cordillera similar types of units overlie the Guadalquivir Allochthon. They have been described in the Central Guadalquivir Basin area and are referred to locally as the Castro del Rio Unit (Roldán-García and Rodríguez-Fernández 1991). These units overlie the wedge and are considered piggy-back basins in the sense of Ori and Friend (1984). In this section I will focus on the Rif Cordillera since more information is available.

6.3.1 THE OUEZZANE UNIT

The Ouezzane Unit was initially known as the "grands synclinaux nummulitiques" or the "Ouezzane Nappe" (Hottinger and Suter in Durand-Delga 1960-1962). Later the term "unités flottantes" or "nappes de glissement" was used (Ben Yaïch 1991). The Ouezzane Unit is exposed in broad synclinoriums overlying the Prerifaine Nappe, constituting the highest structural unit (Fig. 6.12). This unit was interpreted as a thrust sheet or "nappe" located on top of the Prerifaine nappes (Suter 1965, 1980b). According to Michard (1976) the Ouezzane Nappe was the cover of the Tanger Unit.

The stratigraphic section of the Ouezzane Unit is highly variable and consists of Senonian to Miocene marls and siliciclastic sediments (Fig. 6.13).

I postulate that these units may have origins analogous to the Neogene satellite basins which were observed in the offshore and onshore Rharb area. If this hypothesis is correct a re-study of the outcrops should reveal both normal faults and thrusts faults underlying the Ouezzane Unit in addition to the conventional unconformities.

6.3.2. THE NUMIDIAN

The Numidian Unit is the uppermost structural unit of the Flysch domain (Fig. 6.14). It consists of Oligo-Miocene well-rounded coarse quartzy sandstones, with occasionally interbedded micaceous pelites, attaining an average thickness of 500 meters
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Fig. 6.13. Idealized stratigraphic section of the Ouezzane unit. After Serv. Geol. du Maroc (1990).
Fig. 6.14. Structural map of the Western Rif integrating seismic data and surface data modified from Suter (1980b). Location of the Numidian Unit.
in the Rif. The Numidian Unit and its equivalent in the Betic Cordillera, the "Arenisca del Aljibe" (Aljibe Sandstone), represent Upper Aquitanian to Burdigalian turbiditic sandstones. They consist of massive or amalgamated deep-water sandstones interbedded with deep-water hemipelagic shales. The Numidian Sandstone crops out in the hills located between Tanger and Chaouen (Northern Rif) (see Fig. 6.14). In 1958 Mattauer and Durand-Delga correlated these facies with the previously defined Numidian of Algeria. The Numidian Sandstone commonly overlies the Senonian of the Tanger Unit, represented by interbedded m-scale shale with Tubotomaculum and black shale. But they may also overlie the cover of the Tanger Unit, i.e. the Oligo-Miocene Larache Sandstone ("grès de Larache"), which may represent earlier satellite basins. According to zircon provenance studies (Lancelot et al. 1977), the source area for the Numidian sand is presumed to be in Africa, perhaps in eastern Tunisia. It is a polyphasic sandstone representing several sedimentary cycles. The sediments located below the Numidian were paleontologically dated as Lower Oligocene in the Eastern Rif (Leblanc and Feinberg 1982). The Numidian Flysch occupied the southern part of the Massylian-flysch domain and the northernmost portion of the External domain (Wildi 1983). In the light of this hypothesis the age relation between the Numidian and the Larache Sandstone needs to be re-examined.

In the Betic Cordillera the Aljibe Sandstone is the equivalent of the Numidian of northern Africa. This sandstone is extensively outcropping in the Campo de Gibraltar area. The Aljibe Sandstone overlies Aquitanian limestones (Didon 1969, 1973, Peyre 1974). They constitute the highest structural thrust sheet, known as the Aljibe Unit. The Aljibe Sandstone consists of > 75% well-rounded and polished quartz grains and < 10% matrix. Sandstone beds are amalgamated and channelized with dish and flame structures. These deposits have been interpreted as contourites (Wezel 1970, Pendón 1978) or as fluxo-turbidites for Lespinasse (1990). There is some diachronism in the timing of the Numidian
sandstone around the Western Mediterranean, between Sicily and Gibraltar (Martín-Algarra 1987) even though the main sedimentary influx ranges between Upper Aquitanian and the base of the Burdigalian. The Numidian sandstone was deposited over imbricates involving previously deposited turbiditic deposits of the Tanger and Ketama Units. I suggest that the Numidian Sandstone represents turbiditic satellite basins overlying the Oligocene accretionary wedge. These satellite basins may be detached from the underlying units by thrust faults or normal faults simulating an independent thrust sheet, with sediments that are younger than the formations located underneath (Fig. 6.15).

6.4 EVOLUTION OF THE FRONTAL ACCRETIONARY WEDGE

Seismic and field data presented in this work provide important constraints on the evolution of the frontalmost part of the Gibraltar Arc. The evolution of the frontal accretionary wedge may be conveniently split into the following stages:

6.4.1 Compressional stage

6.4.2. Extensional stage.

6.4.3. Foredeep stage.

6.4.1. COMPRESSIONAL STAGE (CRETACEOUS-MIDDLE MIOCENE).

This stage is characterized by foreland-vergent thrusting. First generation thrusts are observed on seismic sections from the offshore Asilah area (see Fig. 5.46 and sections B and F of panel 3), onshore in the Fokra area (north of Souk el Arba) (see Fig. 5.34 and Fig. 5.35) or in the field in the External Rif or Guadalquivir region (Fig.6.16).

6.4.2 EXTENSIONAL STAGE (TORTONIAN-MESSINIAN-PLIOCENE).

Extensional structures cut previous imbricates and thrusts within the Nappe (Fig. 6.17) (see Figs. 5.34, 5.35, 5.47, 5.48 and Panel 3). Extensional collapse advances from hinterland to foreland. Part of the extension is compensated by compression. Compressional structures are the direct compensation of extension (Lalla Zhara, Arbaoua and Souk el Arba).
Fig. 6.15. Deposition and deformation of the Numidian unit according to my hypothesis. Arrow shows sedimentary input direction.
Fig. 6.16. Compressional structure folds and thrusts in the offshore Asilah region.
Fig. 6.17. Extensional structures in the Offshore Larache region superimposed on compressional structures.
Fast extensional collapse is evidenced by:

1. Drastic superposition of deep-water pelagic facies on top of shallow-water littoral facies as indicated by wells.
2. The lack of growth and the parallel filling pattern of the supra-complex units also suggesting rapid subsidence of the extensional basins (see Panels 2 and 3).
3. Olistostromes, reworked sediments.
4. Thick deposition in short intervals also indicating fast subsidence.
5. Paleogeographic evolution.

6.4.3 FOREDEEP STAGE (TORTONIAN-PLIOCENE).

During Lower Miocene the Prerif was partially emerged; the basin deepened towards the Mesorif. Tortonian sediments unconformably overlap Aquitanian or Middle Miocene sediments (upper platform sequence). The Rif foredeep developed mainly during Tortonian-Messinian time. The initial stage of foredeep development was coeval with uplift in the Mesorif and penetrative deformation in the Intrarif (Michard 1976).

Loading by thrust sheets appears to be responsible for the formation of the Betic and Rif foredeeps. A mountainward-dipping monocline of the basement and the overlying Mesozoic-Middle Miocene platform sediments are topped by a basal foredeep unconformity which merges with the leading edge of the accretionary wedge. The overlying Neogene sediments onlap onto the external platform and on the accretionary wedge itself (for example see sections R3, R5 and R7 of panel 1, sections B, C, D and E of panel 3 and section C of panel 4). Thus the foredeep is in part filled by the accretionary wedge and in part by the post-nappe infill of both the foredeep and the supra-nappe satellite basins. The Rif foredeep developed during Tortonian-Messinian time when most of the extensional collapse occurred. Continuing subsidence coeval with extensional collapse of the wedge and loading subsidence created accommodation space for a thick Supra-Nappe sedimentation starting at Tortonian time. The emplacement of the Prerifaine
Nappe is coeval with extensional activity in the wedge. The development and filling of the foredeep takes place in less than 9 Ma.

6.4.4. PALEOGEOGRAPHIC MAPS

Several paleogeographic maps were constructed using data from several authors: Feinberg 1986, Morel 1988, Wernli 1988, ONAREP internal report 1991, geologic maps of the Serv. Geol. du Maroc and own field and seismic data.

During Upper Tortonian time, thrusting in the accretionary wedge was coeval with extensional collapse, resulting in extensional basins (Fig. 6.19). Deep-water pelagic facies developed in these regions while shallow-water sediments were deposited in the peripheral region of the present-day Rharb Basin. The frontal foredeep developed as the wedge moved southward.

During Messinian time the structure of highs and lows is controlled by extensional faults. There is an increase of paleowater depth recorded by the wells of the Rharb Basin (Fig. 6.20). Widespread pyrite and characteristics of the fauna, as well as the rates of sedimentation (Cirac and Peypouquet 1983), suggest anoxic conditions related to water circulation problems. The peripheral extensional system of the Rharb Basin controls the deposition of the pelagic facies. During Messinian time the Prerifaine Ridges were emplaced. On the border of the Meseta, uplift took place in the Tiflet region and a coastal sandstone belt was deposited overlying the Herynian basement.

During Lower Pliocene time the Prerifaine Rides, the Lalla Zhara area and the northwestern part of the Rharb Basin are subaerially exposed (Fig. 6.21). A barrier island system developed in the border of the uplifted meseta (Cirac 1978). Coastal sandstones on the border of the Rharb Basin grade into deep-water pelagic facies.

During Middle Pliocene time there is a widespread deposition of coquinas and shallow-water sandstones (Fig. 6.22) that indicate the end of subsidence and generalized uplift.
Fig. 6.18 Legend of paleogeographic maps
Fig. 6.19. Paleogeographic map of the frontal Western Rif during Upper Tortonian time.
Fig. 6.20. Paleogeographic map of the frontal Western Rif during Messinian time
Fig. 6.21. Paleogeographic map of the frontal Western Rif during Lower Pliocene time.
Fig. 6.22. Paleogeographic map of the frontal Western Rif during Middle Pliocene time.
During Upper Pliocene time the marine sedimentation in the Rharb Basin is restricted to the westernmost region. There is widespread deposition of Villafranchian conglomerates that grade into flood plain clays and sandstones (Fig. 6.23). Local uplift coeval with conglomerate deposition takes place in the Arbaoua area.

During Pleistocene time the facies distribution in the Rharb Basin is similar to the present-day situation. A coastal dune belt grades into marsh-type shallow-water lakes, referred to locally as merjas, interfinger with fluvial deposits (Fig. 6.24).

6. 5. COMPARISON WITH OTHER ACCRETIONARY WEDGES

6.5.1. INTRODUCTION

The Prerifaine Nappe, the Guadalquivir Allochthon and the flysch units represent an unconventional accretionary wedge. The regional structure, the age distribution of the materials involved, and the macrostructure and type of deformation are similar to present-day accretionary wedges. However, in contrast with most accretionary wedges the Prerifaine and Guadalquivir Nappes overlie a more or less attenuated continental crust.

6.5.2. REGIONAL STRUCTURE

The wedge geometry of the accretionary complex is well imaged in seismic profiles (Fold-outs 1-26). The internal structure of the wedge is characterized by seaward verging imbricates, bounded by steeply dipping thrust faults. The wedge is underlain by layered reflectors that represent imbricates detached from the acoustic basement. Overall the Prerifaine Nappe has some similarity with the Aleutian accretionary wedge (Moore et al. 1991) (Fig. 6.25). Seismic mapping indicates that listric normal faults pass laterally into thrust faults, constituting toe-thrusts. Mud diapirs parallel to the direction of thrust faults are common structures of this complex and of modern accretionary wedges (von Huene 1986, Vernet et al. 1992). In front of the wedge, large-scale slices have been slumped into the basin. Similar large-scale Plio-Pleistocene slumps have been described in front of the Gela Nappe of Sicily (Trincardi and Argiamenti 1990). According to seismic and field
Fig. 6.23. Paleogeographic map of the frontal Western Rif during Upper Pliocene time
PLEISTOCENE

Fig. 6.24. Paleogeographic map of the frontal Western Rif during Pleistocene time
Fig. 6.25. Cross-section along the Aleutian accretionary wedge (Moore et al. 1991).

Emplacement of layered reflectors by underplating in response to seaward growth of margin during early Tertiary time. A: Inferred geometry of prism evolution through time. B: Prism dimensions, cross-sectional areas, and rates of growth.

### Table

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**Possible Duplex Geometry**

- **Mesozoic Accretion**
- **Eocene-Oligocene Accretion**
- **Erosion**
- **Conglomerate Deposition**
- **Landward Vergent Thrusting**
- **Thick Incoming Sediments**

**Orogenic History:**

- **50 Ma (650 km²)**: Lateral Growth During Emplacement of Layered Reflectors
- **30 Ma (3800 km²)**: 10°
- **0 Ma (4500 km²)**: 8°

Scale: 0 - 10 km (50 km)
data from the Rharb Basin in Morocco and the Western Betic Cordillera, low-angle extensional detachments offset thrust-bounded imbricates of the accretionary wedge. Extensional collapse has been described in several present-day accretionary wedges (von Huene 1986). As outlined in section 6.3 satellite basins, occasionally associated with extensional structures, ride piggy-back on the wedge.

6.5.3. MACROSCOPIC STRUCTURES

Macroscopic structures observed in well-exposed areas of the Guadalquivir Allochthon, Prerifaine Nappe and flysch units are similar to structures recognized in deep-sea cores from modern accretionary wedges (Lundberg and Moore 1986, Moore et al. 1986) and in ancient accretionary complex melanges (Aalto 1982, Cowan 1985, Knipe and Needham 1986, Kimura and Mukai 1991). The most common structures observed in these units are: scaly cleavage, clastic dikes, breccias, vein structure, spaced foliation, fissility, thrusts, folds, mud diapirs and layer-parallel extensional features. These structures developed under conditions of high fluid (pore) pressure and dewatering and were associated with extensional fracturing. Gravity processes, common in accretionary wedges, are evidenced by the presence of abundant olistostromes and slumps interbedded in the section. Different types of melange have been recognized (Cowan 1985). Each has been related to a deformational process in the accretionary wedge (Fig. 6.26). Type I and II melanges are characterized by boudinage and disrupted beds that indicate layer-parallel extension. The role of gravity in the genesis of these deposits is discussed by Cowan (1982, 1985). Type III is characterized by block-in-matrix mudstone chaos. Some authors claim an olistostromic origin for this type of melange (Page 1978), but some of these deposits seem to be related to hydroplastic deformation. Finally, Type IV mélangé consists of mudstone-dominated brittle fault zones, characterized by elongated inclusions bounded by a system of sub-parallel faults exhibiting an anastomosing pattern. They are formed by shearing between thrust planes in the frontal part of the accretionary prism.
Fig. 6.26. Types of mélange of the Western North American Cordillera and location in the accretionary wedge (Cowan 1985).
(Cowan 1985) (Fig. 6.26). Knipe and Needham (1986) compile structures related to accretionary wedge deformation (Fig. 6.27) and, based on field observations, describe "spoon" structures related to downslope movement similar to the toe-thrust described in this study.

6.5.4. PROGRESSION OF ACCRETION

Seismic, well and field data suggest that the emplacement of the frontal accretionary wedge is rapid but not synchronous along the Gibraltar Arc. Bourgois (1978) had already recognized the time gap between the emplacement of the Guadalquivir Unit in the Western and Eastern Betic Cordilleras. Feinberg (1986) illustrated several stages of emplacement of the Prerifaine Nappe, a Late Tortonian and a Late Pliocene stage. The structure of accretionary wedges is characterized by several belts that become younger towards the front. They indicate the progression of accretion within the wedge (Dickinson and Seely 1979, Moore et al. 1992). Fig. 6.28 shows the record of accretion of the Gibraltar Arc, the distribution of the sediments, and the time of accretion.

6.6. EMPLACEMENT OF THE ALLOCHTHONOUS TRIASSIC

The Moroccan Atlantic margin is also characterized by extensive areas of Triassic and Lower Jurassic salt (Heyman 1989). Upper Triassic reconstruction of the north African and south Iberian regions indicates the presence of extensive evaporitic depocenters in the Atlantic and Western Tethys (Fig. 6.29). The Algerian Basin could be the homeland of the evaporites of the Betic Cordillera.

The Guadalquivir Allochthon involves deep-water pelagic facies "Capas Rojas" and deep-water pelagic marls with radiolaria, referred to as "calcaria maiolica". Triassic sediments are strongly deformed involving kilometric scale sheets to centimetric or decametric blocks interbedded within marls.
Fig. 6.27. Review of deformational features within an accretionary wedge (Knipe and Needham 1986).
Fig. 6.28. Several stages of accretion of the Gibraltar Arc. The frontal tectono-sedimentary units of the Betic and Rif Cordilleras record a progression of accretion towards the external part of the arc. Arrows show the displacement direction of the arc.
Fig. 6.29. Plate tectonic reconstruction of the African and Iberian plates during Hettangian time. Modified after Andrieux et al. (1989), Favre and Stampfli (1992) and Ziegler (1987).
In the Rif Cordillera, Triassic shales and evaporites (mainly gypsum) are intermixed with deep-water Cretaceous (Senonian) marl referred to locally as "marnes à gypsum". Locally these sediments have a breccia character including blocks of metamorphic and igneous rocks, "brèche polygénique à matrice gypseuse". These deposits are well exposed in the Central Rif (Asebriy 1983, Asebriy et al. 1987, Lespinasse 1990). Lower Eocene is unconformably overlying these evaporites, suggesting that the emplacement was already completed at that time.

I suggest that the passive margins of southern Iberia and northern Africa may have some allochthonous salt sheets similar to those reported in the Gulf of Mexico (Worrall and Snelson 1989, Wu et al. 1990, Wu 1993). The salt climbed up the stratigraphic section and was emplaced within younger sediments (Fig. 6.30). This hypothesis would help to explain the presence of large volumes of Triassic sediments within the Guadalquivir Allochthon as well as the absence of Jurassic strata. A non-deformed analog of the Mesozoic southern Iberian passive margin could also in part be similar to the Nova Scotia margin (Friedenreich 1987), which shows similar dimensions and facies distribution similar to the restored Betic margin (Fig. 6.31). Triassic half-grabens superimposed on the Hercynian basement of the Iberian Meseta represent the initial rifting of the margin. Rifting results in crustal thinning and is discordantly covered by evaporitic deposits that thickened towards the Tethys (SSE). A widespread Lower Jurassic carbonate platform occupied the south Iberian margin. The carbonate platform was broken up during Middle Jurassic time and the region became an Atlantic-type passive margin. Initiation of oceanic crust is indicated in Figure 6.32 by seaward-dipping reflectors. Allochthonous salt deposits may occupy distal positions of passive margins. A tentative hypothetic reconstruction of the External zones of the Central Betic Cordillera suggests northward displacement of the Guadalquivir Allochthon in the order of 200 Km (Fig. 6.32). In the Rif the allochthonous Triassic is not as abundant as in the Betic Cordillera, which suggests an asymmetry of the
Fig. 6.30. Development of allochthonous salt and related down-to-the-basin growth fault (Wu et al. 1990).
Fig. 6.1: Hypothetic reconstruction of the South-Iberian passive margin in the Central Betic Cordillera.

Fig. 6.32. This sketch shows the progression of deformation of the Betic passive margin. The allochthonous evaporites of the distal passive margin are transported about 200 Km north and occupy the present-day Guadalquivir allochthon. The deformation progresses from the south to the north and top to bottom 1,2,3 and 4.
north African and south Iberian passive margins rather than a Neogene tectonic feature related to the collision of Alborán.

6.7. GEODYNAMIC EVOLUTION OF THE GIBRALTAR ARC.

6.7.1 INTRODUCTION

The present-day Gibraltar Arc is the result of a complex geological evolution due to the escape and collision of an Alpine-type lithospheric domain (Alborán domain) with the passive margins of the Iberian, African and Atlantic plates. Post-collisional extensional collapse, strike-slip and Late Pliocene-Pleistocene uplift considerably modified the compressional structure of the Arc. The geodynamic evolution of the Gibraltar Arc can be summarized in several stages:

6.7.2 EARLY-ALPINE COMPRESSION

An early stage of Cretaceous-Paleogene accretion recorded by the Flysch units is related to the westward escape and subduction of the Alborán domain and reflects the plate kinematics of Iberia and Africa (see Dercourt et al. 1986 and Dewey et al. 1989). Subduction before 81 Ma resulted in HP/LT metamorphism in the Mulhacen Complex (De Jong 1991) (van Wees et al. 1992) (Fig. 6.33).

6.7.3 COLLISION

After subduction and ductile thrusting under metamorphic conditions within the Alboran domain, collision took place with the southern Iberian and northern African passive margins. The Gibraltar Crustal Thrust (G. C. T.) (Balanyá and García-Dueñas 1987) represents the suture between the External and Internal domains (Fig. 6.34). Thrusting under brittle conditions and accretion coeval with the sedimentation of thick turbiditic sandstone units took place during Aquitanian-Burdigalian time.
Fig. 6.33. Reconstruction of the Gibraltar Arc during Late Cretaceous time. Subduction is associated with HP/LT Blueschists metamorphism.
Fig. 6.34. Reconstruction of the Gibraltar Arc during Aquitanian-Burdigalian time. Collision takes place along the Gibraltar Crustal Thrust (GCT). Back-arc extension initiated in the Alborán region.
6.7.4 BACK-ARC EXTENSION AND FRONTAL ACCRETION

Generalized uplift and exhumation of deep crustal rocks in the Internal domain of the Gibraltar Arc during Miocene time (22-15 Ma), supported by fission track data (Zeck et al. 1992a,b, Michard et al. 1991), was coeval with extensional collapse in the back-arc basin of the Alborán sea. The extensional collapse I reported from the frontal accretionary wedge, took place at the same time as the uplift of the Internal domain and the back-arc extension (Fig. 6.35). Extension related to the elimination of a lithospheric slab associated with a previous collisional ridge was suggested by Platt and Vissers (1989) as a mechanism to create the Alborán back-arc basin.

6.7.5 NEO TECTONICS

Extension in the back-arc region is coeval with compression in the frontal part of the folded belt, where an accretionary complex with well-developed thrusting and imbrication was emplaced (Fig. 6.36). The emplacement of this active wedge was associated with the foredeep development. The Pliocene transgression covered large regions previously exposed. On the Mediterranean Coast, Pliocene marine sediments covered previous estuaries. Incision of the Late Miocene fluvial network resulted from Miocene uplift. In the region of Oued Laou, Pliocene sediments have been uplifted more than 500 meters above sea level. Differentially uplifted ancient beaches can be traced along the Atlantic and Mediterranean coasts of Morocco (Cadet et al. 1977). They show an arcuate pattern along the Atlantic coast. Uplift coincides with the Internal zones of the folded belt (Fig. 6.37).

Quaternary compression is suggested by strike-slip and related transpression along major faults in the Eastern Betic Cordillera (Bousquet and Philip 1976, Sanz de Galdeano 1983, 1990) and reactivation of the Alhama de Murcia fault (Martínez-Díaz and Hernández-Enrile 1991). N-S compression of Iberia-Africa is suggested by the tectonics in
Fig. 6.35. Reconstruction of the Gibraltar Arc during Tortonian time. Frontal accretion is coeval with back-arc extension.
Fig. 6.36. Neotectonics of the Gibraltar Arc. Uplift in the Internal domain of the arc is coeval with frontal extensional collapse and associated compression, foredeep subsidence and foreland uplift.
Fig. 6.37. Differential uplift along the shoreline of the Gibraltar Arc and Alborán Sea from Middle Pleistocene (Sicilian) to Holocene (Late Tyrrenian) time inferred from present day position of paleobeach levels. After Cadet et al. (1977).

Seismicity is also coherent with a generalized N-S compression (García-Dueñas et al. 1984, Medina and Cherkaoui 1992) that could be responsible for the late uplift.
CHAPTER 7
CONCLUSIONS

- The frontal tectono-sedimentary complexes of the Betic and Rif Cordilleras (i.e. the Guadalquivir Allochthon and the Prerifaine Nappe) constitute an accretionary wedge superposed on an attenuated passive margin.
- The structure of the accretionary wedge consists of frontal imbricates, ridges, toe-thrusts and low-angle extensional detachments.
- The Supra-Nappe sediments involve compressional-extensional and extensional satellite basins trending parallel and perpendicular to the arc.
- Satellite basins are not directly related to the opening of the Alborán Sea; instead they are due to oversteepening of the wedge (slope tectonics).
- An undisturbed prograding succession suggests that the wedge was stabilized during Pleistocene time.
- Particularly in Spain the accretionary wedge (Guadalquivir Allochthon) may involve overthrusting of allochthonous Triassic evaporites probably emplaced earlier on a passive margin in a manner similar to the allochthonous salt of the Gulf of Mexico. This hypothesis explains the lack of Jurassic strata and the abundant evaporites within the Guadalquivir allochthon and the less abundant evaporites of the Prerifaine Nappe.

- Very rapid extensional collapse affected the accretionary wedge during Tortonian and Messinian time. Several arguments support this:
  1. paleogeographic evolution,
  2. superposition of deep-water facies onto shallow-water sediments,
  3. lack of significant growth.
• The style of extension is characterized by low-angle listric normal faults similar to the Gulf of Mexico.

• Extensional displacement is compensated by frontal compression. Toe-thrusts are common structures in the frontal part of the Gibraltar Arc.

• Geodynamic evolution

  The geodynamic evolution of the Gibraltar Arc can be summarized in the following stages:

1. Subduction and westward accretion during Cretaceous time.

2. Collision and westward accretion during Aquitanian-Burdigalian time.

3. Frontal accretion coeval with hinterland uplift and back-arc extension. Foredeep development is coeval with frontal extensional collapse.

4. Late uplift, transpression and compression (and related inversions in the Alborán Sea). Deep-seated thrusting is responsible for the emplacement of lower thrust sheets involving passive margin sediments (envelopment of overlying units).

5. The Gibraltar Arc is the result of the westward escape of the Alborán block, resulting in: strike-slip, roll-back of a subduction plane, extensional collapse and oroclinal bending.
APPENDIX
CLASSICAL UNITS OF THE BETIC AND RIF CORDILLERAS

The following notes are intended to serve as additional background for the regional map. They provide useful references about stratigraphic, sedimentologic, petrologic and structural aspects of several regions of the Betic and Rif Cordilleras.

A.1 SOUTH-IBERIAN DOMAIN

The south-Iberian domain includes the External zones of the Betic Cordillera, which consists of thrust sheets involving Jurassic to Lower-Middle Miocene sediments. Triassic evaporites form the main décollement level. The Mesozoic sediments are the deformed Tethyan margin of the Iberian continent. Thrusting superposes basin over platform paleogeographic domains. Published well-log data and geologic maps in the western Betic Cordillera suggest repetition of carbonate platforms. The Penibetic-Subbetic platform (Martín-Algarra 1987) is thrust onto the Gulf of Cádiz-Algarve platform. In the Central Betic Cordillera seismic data indicate superposition of several paleogeographic domains (e.g. García-Hernández et al. 1980, Blankenship 1989, 1992).

The external zones of the Betic Cordillera have been subdivided into different units based on structural and stratigraphic criteria (Blumenthal 1927, Fallot 1948, Azema et al. 1979, García-Hernández et al. 1980, 1988). This study follows the most recent subdivision (García-Hernández et al. 1988), from north to south: External Prebetic, Internal Prebetic, Intermediate Units, External Subbetic, Middle Subbetic, and Internal Subbetic (Penibetic) (Fig. A-1).

A.1.1. PREBETIC.

The Prebetic zone of García-Hernández et al. (1988) is the northernmost paleogeographic unit of the External domain of the Betic Cordillera (Fig. A-1).
Fig. A.1. Classical structural units of the External domain of the Betic Cordillera. After Junta de Andalucia (1985).
The Triassic of the External domain of the Betic Cordillera is in the Germanic facies. In the southernmost regions of the mountain belt, the Triassic is represented by deep-water facies and a thicker section referred to locally as "Trias germano-andaluza" (Fig. A-2). The Upper Triassic (Keuper facies) consists of mudstones with interbedded sandstone, gypsum and carbonate. These facies have been interpreted as fluvial and coastal mud-flat and sabkha deposits (Pérez-López and López-Chicano 1989). The Triassic facies were deposited in rifts during the early development of the south-Iberian margin (Vera 1983). The detritic and evaporitic Triassic sediments are overlain by widespread shallow-water Jurassic carbonates, representing the northern Tethys carbonate platform attached to the south-Iberian continental margin. The Prebetic zone of the Eastern Betic Cordillera was continuous with the Mesozoic sediments that overlie the Hercynian Iberian Meseta. The section is composed of partially or completely dolomitized shallow-water carbonates, interrupted by numerous hiati and unconformities. The Jurassic of the Internal Prebetic zone consists exclusively of shallow-water facies, whereas the External Prebetic includes Lower and Middle Jurassic shallow platform carbonates as well as Upper Oxfordian-Lower Kimmeridgian pelagic deposits. Due to erosion, the Lower Cretaceous is absent over wide regions (see Fig. 3.2). The Upper Cretaceous section consists of limestone and dolomite deposited in a restricted lagoon (Azema et al. 1979). Paleogene sediments are present only in the southern part of the Prebetic domain. The Miocene cover thickens towards the south (Vera 1983). The Prebetic zone is not present in the Western Betic Cordillera, where the Subbetic zone is in direct contact with the Guadalquivir Allochthon.

A.1.2 INTERMEDIATE UNITS.

The Intermediate units are located between the classical Prebetic and Subbetic zones. The Triassic facies is similar to that of the Prebetic and Subbetic units. Lower Jurassic platform dolomites overlie Triassic evaporites. The Intermediate units are
characterized by the lack of post-Liassic Mesozoic platform carbonates (see Fig. 3.2). The Middle Jurassic section is made up of shallow-water oolitic limestone and radiolarian-bearing pelagic marl with interbedded calci-turbidites and breccias. The Lower Cretaceous section consists of a thick succession of detritic and/or calci-turbidites. Paleogene marl and marly-limestone, locally inter-fingered with sandstone, conglomerate and olistostrome levels, follow in the section (Vera 1983). In the Eastern Betic Cordillera there is a thick Paleogene and Neogene section. The Intermediate unit succession is much thicker than the Prebetic unit. Seismic data suggest that the I.U. were deposited in a synsedimentary extensional setting (Banks and Warburton 1991). These facies constitute the basinal facies deposited south of the Prebetic platform.

A.1.3. SUBBETIC

The Subbetic zone consists of three units: the External platform, the Internal platform unit, and an Intermediate unit characterized by basinal facies during Middle Upper Jurassic time.

A.1.3.1 External Subbetic

In the eastern Betic realm, the External Subbetic unit (García-Hernández et al. 1988) is located south of the Intermediate units (see Fig. 3.2). In the Western and Central Betics, the Subbetic directly underlies the Guadalquivir allochthonous units.

The Triassic and pre-Domerian Jurassic section, like in the Prebetic zone is made up of shallow-water carbonates. The Lower Jurassic is represented by shallow-water carbonates and dolomites and is very uniform throughout the Betic Cordillera.

The Domerian and Toarcian succession is represented by a thin section of ammonite-bearing marl and marly-limestone. The Middle Jurassic section consists of shallow oolitic limestone and interbedded micritic and nodular ammonitic limestone. A karstified unconformity marks the Middle Jurassic—Upper Jurassic boundary. Upper
Callovian-Lower Oxfordian pelagic sediments fill the karstic surfaces (Castro et al. 1990). The Upper Jurassic section is characterized by ammonite-bearing nodular limestones, Middle-Toarcian red nodular marly-limestone, known as "ammonitico rosso" facies, and coarse breccias (Vera 1983). The Jurassic sediments were deposited in a tilted block setting controlled by listric normal faults (Molina and Ruiz-Ortiz 1990, Ruiz-Ortiz et al. 1990)

A.1.3.2 Middle Subbetic

The Middle Subbetic consists of Mesozoic deep-water sediments and basic igneous rocks. Mesozoic igneous rocks are represented by small dolerite stocks and dykes intruded into Middle-Upper Triassic rocks, or as submarine basic volcanic flows (basaltic lavas and pillow lavas) and sills inter-layered with Jurassic pelagic sediments (see Fig.3.2). The climax of the magmatic activity occurred during Tithonian time. Geochemical studies of the basalts support a mantle origin and some continental crust contamination (Puga et al. 1989 b).

The Cretaceous succession consists of homogeneous pelagic marl and shale with inter-bedded turbidites. The Lower Cretaceous is represented by rhythmically alternating marly-limestone and marl containing ammonites and radiolaria (see Fig.3.2). The Upper Cretaceous consists of micritic marly-limestone and pink-white marl with coccolithic lime matrix and planktonic foraminifera. These facies are referred to locally as "Capas rojas" in the Betic Cordillera and are similar to the "scaglia" of Italy (Vera 1981). The Paleogene is represented by a discontinuous section of microcodium limestone. The Lower Eocene facies are similar to the "Capas rojas" pelagic marls. The Middle-Upper Miocene and the Oligocene consist of planktonic foraminifera-bearing marl with interbedded turbiditic sandstone and sandy-limestone. The Paleogene and Lower-Middle Miocene is represented
by marine limestone, bioclastic sandstone and planktonic foraminifera-bearing marl interbedded with turbidites (Vera 1983, Azema et al. 1979).

A.1.3.3. Internal Subbetic

The Internal Subbetic, also known as Penibetic west of Antequera (Vera 1983), is very similar to the External Subbetic. The stratigraphy of this zone is characterized by a thick succession of Middle-Upper Jurassic "amonitico rosso" facies. Thus the Jurassic section is entirely formed by carbonates, consisting of nodular and pelagic limestones (see Fig.3.2). Karstified unconformities separate the Jurassic succession from a Cretaceous and Tertiary section that is equivalent to the External and Intermediate Subbetic (García-Hernández et al. 1986).

A.1.4. STRUCTURE OF THE EXTERNAL BETIC CORDILLERA.

A.1.4.1 Introduction

The structure of the External Betic Cordillera is the result of Neogene compression of a passive margin succession. Earlier Mesozoic synsedimentary deformation was probably controlled by listric normal faults and locally by diapiric emplacement (Sanz de Galdeano 1973, Vera 1983, Martínez del Olmo et al. 1985). Neogene tectonics resulted in thin-skinned foreland-vergent thrust sheets. Locally, backthrusting occurred along the contact with the Internal domain (Balanyá 1984, Banks and Warburton 1991). The External Betic Cordillera is subdivided into three structural segments: Western, Central and Eastern Betic Cordilleras.

A.1.4.2 Western Betic Cordillera

The structure of the Western Betics is characterized by windows of the Subbetic units overlain by strongly deformed Triassic to Miocene marls and shales of the Guadalquivir Allochthon. Wells demonstrate the superposition of allochthonous Triassic
evaporites and Cretaceous deep-water sediments onto Triassic, Jurassic and Cretaceous platform sediments (see Fig. 5.63).

A.1.4.3 Central Betic Cordillera.

The structure of the Central Betic Cordillera consists of NW-vergent thrust sheets detached from Triassic evaporites. Thrust displacement increases southward, from 3 to 15 Km in the frontal Prebetic to 20-30 Km in the Subbetic (Vera 1983) (see Fig. 3.8, Fig. A-3). Seismic and well data suggest superposition of two northwestward-vergent thrust sheets detached within Triassic evaporites. These thrust sheets contain Jurassic and Cretaceous sediments and are characterized by footwall ramp-flat systems and hangingwall ramp anticlines. According to Blankenship (1992) 200 Km shortening was estimated. This shortening is contrasted with the 60-70% shortening of García-Hernández et al. (1980). Note that Blankenship's interpretation has been questioned by Sanz de Galdeano (in press).

In the Central and Eastern Betic Cordilleras diapirism seems to be a major factor controlling the structure of the Prebetic zone (Martínez del Olmo et al. 1986, Sanz de Galdeano 1973).

A.1.4.4. Eastern Betic Cordillera

A change in structural trends occurs at the eastern termination of the Guadalquivir Basin. ENE-WSW trending folds and thrusts of the Central Betics, shift to NNE-SSW oriented structures in the Eastern Betics (Sierra de Cazorla). This change coincides with the location of the Tiscar fault, which has been interpreted as a tear fault (Frizon de Lamotte et al. 1991) (Fig. A-4). East of the Tiscar fault, the structures consist of Jurassic rocks with thrust slices verging NW as well as doubly-vergent imbricates of Triassic, Jurassic and Cretaceous rocks referred to as "Alcaraz Imbricates". Thrust flats are located within Triassic and Early Cretaceous levels (Fig. A-5). The magnitude of thrusting of the
Fig. A.3. Cross-section through the Central Betic Cordillera. García-Hernández et al. (1988).
Fig. A.4. Structural map of the Sierra de Cazorla area. Notice the change of structural trend. After Frizon de Lamotte (1991).
Fig. A.5. Cross section through the Eastern Betic Cordillera Banks and Warburton (1991).
Subbetic onto the Intermediate units is estimated to be 40 Km (Banks and Warburton 1991).

A.2 THE MAGHREBIAN DOMAIN

A.2.1. INTRODUCTION

The Maghrebian domain includes the External domain of the Rif and Tell Cordilleras. The sedimentary succession involved in these folded belts represents the deformed northern passive margin of the African plate (Wildi 1983). On the basis of stratigraphic and structural criteria, the External domain of the Rif Cordillera was subdivided by Suter (1965) from hinterland to foreland into three main units: Intrarif, Mesorif and Prerif (Fig. A-6). The most internal zone, i.e. the Intrarif, consists of the Tanger and Ketama thrust sheets. The Intrarifean Nappes involve mainly Cretaceous deep-water epimetamorphic shales and turbidites (Andrieux 1971, 1973). The Mesorif consists of a thick non-metamorphic Mesozoic section that includes Lower-Middle Jurassic carbonates and Upper Jurassic, Cretaceous and Paleogene deep-water carbonates and siliciclastics (Fig. A-7). Jurassic and Lower Cretaceous carbonates are exposed in tectonic windows below Upper Cretaceous and Paleocene sediments characterized by foreland-vergent imbrication. Finally, the Prerif is the most external zone, consisting of allochthonous tectono-sedimentary complexes involving Triassic, Cretaceous, Paleogene and Neogene sediments (Fig. A-7). The main difference between the Prerif and the Mesorif-Intrarif is that the Prerif has a thin continuous Triassic to Cretaceous sedimentary succession, with conformable Tertiary. On the contrary, in the Intra and Mesorif, the Tertiary unconformably overlies the Mesozoic succession (Leblanc 1977). These classical domains (i.e. Intrarif, Mesorif and Prerif) can be split into several structural units or "Nappes" (Suter 1965, 1980 a,b, Leblanc 1977) (see Fig. A.6 and A.7). The following description will proceed from south to north.
Fig. A.6. Classical subdivision of the Rif Cordillera according to Suter (1965).
Fig. A.7. Cross-sections of the External domain of the Rif Cordillera. After Suter (1967) in Michard (1976). Legend: T Triassic, Li Lower Liassic, Lm Middle Liassic, Ls Upper Liassic, Jm1 Middle Jurassic, Jm2 Middle-Upper Jurassic, Jf Upper Jurassic (flysch), Ci Lower Cretaceous (limestones), Cf Lower Cretaceous flysch, Cm Middle Cretaceous (limestones), Cs Upper Cretaceous (marls), Ei Paleocene-Lower Eocene, Es Upper Eocene, O Oligocene, M1 Lower Miocene, M4a Middle Miocene, M4b Middle-Upper Miocene, M4c Upper Miocene, P-Q Pliocene-Quaternary.
A.2.2. THE RIDES PRERIFAINES.

The Rides Prerifaines are located in the southwestern Rif, east of Sidi Kacem and extending to the region of Fès. The stratigraphic succession consists of Mesozoic carbonates and Paleocene siliciclastic sediments, mainly carbonates (see Fig. 3.7). The stratigraphy of the Rides is like the stratigraphy of the Middle Atlas (Faugères 1978, 1981). The Rides Prerifaines involve a thick Jurassic carbonate succession overlain by a thin Cretaceous to Miocene cover. The Jurassic succession is from bottom to top as follows (Favre et al. 1991):

- Sinemurian and Carixian tidal limestones and dolomites rich in brachiopoda, topped by packstones and grainstones with corals.
- Domerian bivalve-bearing marly-limestones grading into interbedded marl and marly-limestone.
- Partially or totally dolomitized Toarcian marl with interbedded sandstone, limestone and algal mats (Faugères 1975). This marly section is overlain by sandy-limestone and deltaic sandstone topped by tidal packstones and grainstones.

Unconformably on top of the Jurassic are neritic Middle Cretaceous and Senonian marls and Lower Eocene phosphates. These in turn are overlain by Aquitanian continental red-beds and Upper Miocene (probably Tortonian) white marl of Beni-Amar which are the youngest sediments of this unit (Michard 1976).

A.2.3. PRERIF

The Prerifaine zone is the most frontal part of the Rif. It was subdivided into Internal and External Prerif (Marçais and Suter in Durand-Delga et al. 1962, Wildi 1983). The Internal Prerif involves autochthonous imbricates involving Jurassic carbonates, (Lias to Tithonian). The External Prerif is composed of a lower tectono-sedimentary complex (i.
e. Nappe Prerifaine) consisting of a marly-sandy matrix with blocks of different size, overlain by upper sedimentary units referred to classically as "Unités flottantes" or "Nappes de glissement" or "nappes rifaines supérieures" (Wildi 1983).

The stratigraphy of the Internal Prerif is similar to the Mesorif and consists of Middle-Upper Jurassic platform carbonates overlain by deep-water Tithonian limestone. The carbonates are exposed on the surface and form the so-called "Sofs" (see Fig. 3.7).

**A.2.3.1 Stratigraphy**

The Triassic succession is represented by marls, shales, gypsum and salt of thicknesses ranging between 100-2000 meters (Salvan 1974). The Jurassic section consists of Lower Jurassic limestone with silex and ammonitic marly-limestone, Callovo-Oxfordian interbedded turbiditic sand and shale referred to as Ferry-flysch (Wildi 1983) and ammonite rich Tithonian nodular limestone in "ammonitico rosso" facies (see Fig. 3.7). The Cretaceous succession is represented from bottom to top by: Albo-Aptian bioclastic limestones, Cenomanian-Turonian marl and Senonian marls and limestones with reworked Cenomanian and Triassic sediments. Upper Paleocene to Lower Eocene interbedded marl and sandstone with occasional conglomerates overlie Cretaceous marls. The Upper-Middle Eocene is quite variable in the Prerif, encompassing mainly sandstones, locally glauconitic, phosphatic limestone, Lutetian nummulitic limestone and white marl with silex. Oligocene marl with *Lepidocyclina sp.* and occasional sandstone grades into Lower Miocene sandy-marl and marl with *Globigerinoides bisphericus*. Middle Miocene marly-sandstone a *Orbulina sp.*.

Upper units.

The "unites flottantes" occupy the highest structural position of the Prerif. They are characterized by an Eocene-Oligocene sedimentary succession. There is no agreement on the origin of the upper units. The classical view suggests Intrarifean origin for all the
upper units of the Prerif (Suter 1980b). According to Wildi (1983) the Upper units can be subdivided into:

- "Nappes Rifaines Supérieures". Intrarifean affinity units represented by the Senhadja and Aknoul "nappes" consisting of epimetamorphic Lias to Upper Cretaceous sediments.

- "Nappes rifaines Inferieures" Mesorifean affinity units represented by the Bou Haddoud and Taïneste "nappes" (Leblanc 1979). They present a section similar to the Mesorif.

- "Ouezzane type Nappes" Prerifean affinity units including the Ouezzane, Habt and Tsoul "nappes".
  Each unit will be discussed in the zone of affinity.

**A.2.3.2 Ouezzanne unit.**

Introduction

The Ouezzane unit is equivalent to the "grands synclinaux nummulitiques" the "Ouezzane Nappe" (Hottinger and Suter in Durand-Delga et al.1960-62) the "unités flottantes or "nappes de glissement" (Ben Yaïch 1991). The Ouezzane unit is exposed in broad syncliniorms overlying the Prerifaine Nappe and constitutes the highest structural unit. This unit was interpreted as a thrust sheet or "nappe" located on top of the Prerifaine nappes (Suter 1965, 1980b). According to Michard (1976) the Ouezzane Nappe is the cover of the Tanger unit.

**Stratigraphy :**

The stratigraphic section of the Ouezzane unit consists of Senonian to Miocene marls and siliciclastic sediments (Fig. A-8).(Serv. Geol. Maroc 1990).
## OUEZZANE UNIT

<table>
<thead>
<tr>
<th>Thickness Range</th>
<th>Lithology</th>
<th>Biozones</th>
<th>Age</th>
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</thead>
<tbody>
<tr>
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<td>Orbulina sp.</td>
<td></td>
<td>Middle Miocene</td>
</tr>
<tr>
<td>500-200 m</td>
<td>G. primordius, G. sicanus, Globigerinoides sp, G. opima opima, G. increbescens, G. ciperoensis angulisuturalis</td>
<td></td>
<td>Lower Miocene</td>
</tr>
<tr>
<td>100-0 m</td>
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<td>Upper Eocene</td>
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<tr>
<td>500-0 m</td>
<td>G. pseudomenardii, G. velascoensis</td>
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<td>Middle Eocene</td>
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<td>500-0 m</td>
<td>G. marginodentata, G. lensiformis, G. palmerae, G. caucasica</td>
<td></td>
<td>Lower Eocene</td>
</tr>
<tr>
<td>400-0 m</td>
<td>G. edita, G. trinidensis, G. praecursoria, G. conicoiruncata</td>
<td></td>
<td>Middle Paleocene</td>
</tr>
<tr>
<td>200-0 m</td>
<td>Senonian</td>
<td></td>
<td>Lower Paleocene</td>
</tr>
</tbody>
</table>

Fig. A.8. Idealized stratigraphic section of the Ouezzane unit. After Serv. Geol. du Maroc (1990).
A.2.3.3 Habt unit

The Habt unit (Suter and Fiechter 1966) is exposed in the northwestern Rif and consists of Upper Cretaceous to Middle Miocene sediments. According to Suter (1965) the Habt unit is the detached cover of the Tanger unit. The Habt unit, known also as "Habt Nappe", overlies a basal Triassic breccia ("breche gypseuse") containing Paleozoic blocks, dolerites, gypsum and salt. Locally Cretaceous marl, but commonly Middle-Upper Eocene to Lower Oligocene clays and shales with gypsum, constitute the base of the Habt unit. This succession also includes limestone with *Orbitolina sp.*, turbiditic bioclastic sandstones and coquinas. An Upper Oligocene sandstone unit, referred to as "gres d'Asilah" (Asilah sandstone), consists of thick-bedded mainly channelized sandstone and interbedded marls and sandstones. These deposits have been interpreted as proximal turbiditic channel deposits that display paleocurrent directions towards the north-northwest (Morley 1987). The Paleogene succession contains deep-water arenaceous fauna. The upper part of the Habt succession consists of Lower Miocene pelagic marl with sandstone, intraformational conglomerates and conglomerates with limestones with *Orbulina sp.*.

The structure of the Habt unit is characterized by NNW-SSE oriented SW-vergent thrusts. The base of the Habt unit shows stratigraphic omission, suggesting extensional faults. Cross-cutting relationships and the overall internal structure of the Habt unit have been interpreted by out-of-sequence thrusting (Morley 1992).

A.2.4 MESORIF

The Mesorif (Suter 1965) consists of a set of tectonic windows that trend parallel to the main structures of the Rif Cordillera (see Figs. A-6, A-7). This zone was initially known, based on its outcrop pattern, as "zone des fenêtres" (Marçais 1937). The Mesorif section is overlain by the Flysch units and the Intrarif in the north and by the Prerifaine
Nappe in the south. The Mesorif represents part of the African passive margin succession involved in thrust sheets that were emplaced after collision with the Alborán domain.

**A.2.4.1 Stratigraphy**

The Mesorif is characterized by deeper water facies and a thicker sedimentary succession than the Prerif. The Triassic and Jurassic sections have Atlasic affinity (Michard 1976). The Triassic consists of red clays, shales and sandstones with dolomites, dolomitic breccias, dolerite and limestone (Salvan 1974). The Jurassic section consists of Callovo-Oxfordian deep-water turbidites, referred to locally as "Ferry-flysch" (Wildi 1981, 1983), overlain by widespread Tithonian Calpionella limestones and "ammonitico rosso" facies (see Fig.3.7). The Cretaceous succession consists of: Neocomian ammonite white marly-limestone, Albo-Aptian turbiditic shales, sandstones and conglomerates, referred to as Albo-Aptian "flysch", and Cenomanian-Senonian marls and nodular marly-limestones. The overlying Paleogene succession consists of Lower-Middle Paleocene marl and Upper Paleocene-Lower Eocene white marl with silex, topped by Oligocene siliciclastic turbidites (Lespinasse 1990) (Fig. A.9).

The lithology and thickness of the Neogene cover are variable but can attain a thickness of 1000 meters. It is composed of siliciclastic turbidites, marls and calcarenites. In the Central Rif there is a pronounced unconformity at the base of the Lower Miocene (Leblanc 1977). Conglomerates with abundant Cretaceous and Eocene reworked faunas are present at the base of the Lower Miocene section. The Aquitanian-Lower Burdigalian consists of Lepidocyclina-bearing white marl with reworked Nummulites and glauconitic sandstone.

The Upper Oligocene-Lower Miocene Zoumi Sandstone lies unconformably on top of Jurassic or Lower Cretaceous. The Zoumi Sandstone consists of approximately 800 meters of interbedded sandy-marl and graded sandstone, representing syntectonic turbiditic sandstone lobe deposits with NNW-SSE paleocurrent directions (Lespinasse
Fig. A.9. Stratigraphy of the Internal Prerif, Mesorif and Upper Intrarifean thrust sheets (Wildi 1983) For location see figure A.6.
1977, Morley 1987). Unconformably on top, the Dar el Oued Formation consists of white marls, limestones and shales with sandy beds and blocks of Eocene limestones with silex. Slumps are common in the sedimentary succession. This formation is supposed to represent a Lower Miocene olistostrome (Lespinasse 1990).

A.2.4.2 Structure

The structure of the Mesorif is defined by foreland-vergent thrust sheets involving Jurassic and Cretaceous sediments unconformably overlain by Oligo-Miocene siliciclastics (Suter 1980a,b, Wildi 1983, Favre et al. 1991) (see Fig. A-7). Eventually backthrusting occurs at the Mesorif/Intrarif contact (Favre et al. 1991, own observations). Weak regional metamorphism, affecting Mesorif rocks, was reported from the Eastern Rif (Andrieux 1971, 1973). West of the Loukkos River Valley, the Mesorif is overthrust by the Intrarifean Loukkos and Habit Nappes (Suter 1980b). The western termination of the Mesorif coincides with a NE-SW trending lateral ramp of the Habit unit (see Fig. A-6). This contact has been interpreted as a major cross-element (Morley 1987). The Mesorif zone is the surface expression of a Jurassic-Cretaceous duplex.

A.2.5. INTRARIF

The Intrarif unit (Suter 1965) is located between the Mesorif and the flysch units, and it is largely present in the Central Rif. This zone is considered to be para-autochthonous, representing the substratum of the detached flysch units (Michard 1976). The Intrarif section consists of a Lower Cretaceous succession that only occasionally preserves its Upper Cretaceous cover. These sediments represent thicker and deeper water facies than the Mesorif (see Fig. 3.7) (Suter 1965, Michard 1976, Favre et al. 1991). The Intrarif is made up of several structural units (Suter 1965, 1980b): Ketama, Tanger, Loukkos, and the "Nappes rifaines superieures" of the Aknoul / Senhaja units (Wildi 1983) (Fig. A-10).
Fig. A.10. Stratigraphy of the Tanger, Loukkos and Ketama units (Wildi 1983).
For location see figure A.6.
A.2.5.1 Ketama unit

The Ketama unit is widespread in the Central Rif (Andrieux 1971). The stratigraphic section consists of a thick Aptian-Albian shaly-sandy turbiditic succession, topped by Albian red and black ammonitic shale with lenses of quartzitic sandstone (Fig. A-10) (see also Fig. 3.7). These are overlain by Cenomanian-Turonian white marl and limestone with ammonites, green and black shales and marl with intraformational breccias (Lespinasse 1990) (see Fig. 3.7, A-10).


A.2.5.2 Tanger unit.

The Tanger unit is present only in the external Western Rif. It has been interpreted as the detached cover of the Ketama unit (Leblanc 1977). The stratigraphic section consists of Aptian to Lower Miocene marls. The Tanger unit was subdivided into "Tanger occidental" (External Tanger unit) characterized by a very thin section and "Tanger oriental" (Internal Tanger unit) represented by a thicker and more complete section (section of "Bab Taza") (Leblanc 1977, Suter 1980a).

The Cretaceous succession is represented from bottom to top (Fig. A-10), by:

- Cenomanian green and black shale
- Coniacian interbedded grey marl and limestone with Globotruncana.
- Santonian marl with interbedded decimetric intervals of microbreccias and reworked marl a Rotalipora.
-Campanian blue marl with *Globotruncanina sp.* with *Inoceramus sp.* and coquinooid intervals.

The Tanger unit is not affected by significant penetrative deformation (i.e. schistosity, crenulation cleavage or foliation).

A.2.5.3 Loukkos unit

The Loukkos unit is exposed only in the Western Rif and constitutes a thrust sheet located in front of the Tanger unit (Suter 1965, 1980a,b). The stratigraphic section consists of (Fig. A-10):

- Albian-Aptian black shales with quartzitic lenses.
- Cenomanian limey-shales with sandstone lenses, marl and ammonitic limestone with disperse gypsum.
- Turonian marly-limestone.
- Senonian grey marls with re-sedimented Cenomanian marly-limestones with *Globotruncanina sp.* and Triassic shales and evaporites.

The Upper Cretaceous-Paleogene contact is transitional.

- Eocene grey marl and limestone with Nummulites and fine-grained sandstone, conglomerate and microbreccia with *Globorotalia sp.* (Lespinasse 1990).

A.2.5.4 Aknoul unit

The Aknoul unit is considered as a "Nappe Rifaine supérieure" of Intra-Rif affinity (Wildi 1983). The Aknoul unit "Nappe d'Aknoul" represents the detached cover of the Ketama unit in the Eastern Rif (Leblanc 1977, Frizon de Lamotte and Leikine 1985), which, according to Frizon de Lamotte and Leikine (1985), was emplaced as a gravitational nappe. The stratigraphic section of the Aknoul unit is made up of Jurassic black shales and Cretaceous to Eocene marls, marly-limestones and limestones (Fig. A-9). Miocene low-grade metamorphism, characterized by chlorite-illite growth, affects the base of the Aknoul Nappe (i.e. the Lower Cretaceous section), resulting in a vertical
metamorphic gradient. Two phases of deformation were recognized in the Aknoul Nappe (Frizon de Lamotte and Leikine 1985):

1. Syn-schistosity phase characterized by sub-horizontal bedding-parallel schistosity affecting only the lower part of the section.
2. Reactivation of previous schistosity planes and generation of N-S oriented calcite veins.

A.3 FLYSCH UNITS

A.3.1 INTRODUCTION

A belt of deformed deep-water turbidites form the external part of the Gibraltar Arc s. s. . This zone separates the External and Internal domains of the Arc. The flysch domain "Nappes des Flysches" has always been interpreted as an elongated trough filled with turbiditic deposits; the "flysch trough" extends from the Gibraltar Arc to Sicily and the Adriatic. These flysch units are part of a continuous deformed belt present in the Apennines and in the Tell, Rif and Betic Cordilleras. The Flysch units of the north African alpine chain (Rif and Tell Cordilleras) have their equivalent in the Betic Cordillera. The Mauritanian units defined in the Chellata massif in the Gran Kabylie of Algeria and Tunisia are equivalent to the Beni-Ider and Tisirene units of the Moroccan Rif and the Algeciras and Nogales units of the Betic Cordillera (Didon et al. 1973). The Massilian units defined in the Petite Kabylie by Raoult (1969) are equivalent to the Mellousa and Chouamat units of Morocco and to the Facinas and Almarchal units. The Numidian units of northern Africa are equivalent to the Aljibe Sandstone of the Betic Cordillera (Didon et al. 1973, SECEGSA 1990). A complete correlation of the flysch units across the Straits of Gibraltar is presented in the "Mapa tectónico del Estrecho de Gibraltar" (SECEGSA 1990). González-Donoso and others (1987) define this zone as a Cretaceous to Lower Miocene tectono-sedimentary complex. In the Betic Cordillera, flysch unit sediments are
incorporated in the Guadalquivir allochthonous units (González-Donoso et al. 1987, Martín-Algarra 1987).

<table>
<thead>
<tr>
<th>Maghrebian flysch units</th>
<th>Betic Cordillera</th>
<th>Rif Cordillera</th>
</tr>
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<tr>
<td>Mauritanian units</td>
<td>Algeciras unit</td>
<td>Beni-Ider unit</td>
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<td>Nogales unit</td>
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<td>Massilian units</td>
<td>Facinas unit</td>
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<tr>
<td>Numidian units</td>
<td>Aljibe Sandstone</td>
<td>Numidian Sandstone</td>
</tr>
</tbody>
</table>

Table A-1. Equivalence between the flysch units on both sides of the Straits of Gibraltar. After Didon et al. (1973), SECEGSA (1990).

A.3.2. PROPOSED MODELS

Several models have been proposed to explain the structure and paleogeography of the flysch units. Initially autochthonist views were held (Gavala 1916). These were followed by allochthonist conceptions. Durand-Delga and Mattauer (1960) proposed the so-called "ultra" hypothesis. According to this conception, the provenance of flysch units was more internal than the Internal domain. This was based on the fact that the flysch units overlie not only the sediments of the External domain but also locally the sediments of the Internal domain. Following field observations in Sicily and Calabria, (Ogniben 1963, 1970) a new allochthonist conception known as the "intra" hypothesis was established. In contrast with the "ultra" hypothesis, the "intra" hypothesis supports flysch deposition between the Internal and External domains in a "eugeosyncline type" trough. In this model the Cretaceous to Paleocene paleogeographic location of the flysch was adjacent to the southern and western margins of the Alborán domain (Fontboté 1983, Wildi 1983,
Durand-Delga & Olivier 1988, Sanz de Galdeano 1990). According to Lespinasse (1990) the flysch domain was initially located between External Intra-Rif (Ketama-Tanger units) and the Dorsale. Several paleogeographic models of the flysch units have been outlined in Table A.2.

<table>
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<th>Autochthonist theories</th>
<th>Gavala (1929)</th>
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<td>Allochthonist &quot;ultra&quot;</td>
<td>Durand-Delga and Mattauer (1960-62)</td>
</tr>
<tr>
<td>Allochthonist &quot;intra&quot;</td>
<td>Obigny (1963)</td>
</tr>
<tr>
<td>External Intrarif-Dorsale</td>
<td>Lespinasse (1990)</td>
</tr>
</tbody>
</table>

Table A.2. Several models on the flysch units and the authors who support them.

A.3.3. STRATIGRAPHY

The stratigraphy of the flysch units is similar on both sides of the Gibraltar Straits. The Cretaceous flysch is very well developed on the African side (i.e. Rif Cordillera) and poorly represented on the Iberian margin (i.e. Betic Cordillera). The succession consists of Cretaceous to Lower Miocene deep-water siliciclastics and turbidites (Fig. A.11).

A.3.3.1 General stratigraphy

Middle Jurassic pelagic sandy-limestones and shales grading upwards to calcturbidites constitute the base of the flysch domain section (Wildi 1983). Middle Jurassic to Lower Cretaceous submarine volcanic series are known from the "Petite Kabylie" of Algeria. A thick terrigenous turbiditic formation was deposited from the Valanginian (in the Rif) or from the Barremian (in the Tell) to the Middle Aptian. The base of the Upper Cretaceous section consists of pelitic sediments containing radiolarite and bedded chert, grading upwards into marl and microbreccias. In the Kabylie region conglomerate pebble and breccia fragments shed from the Dorsale unit are common. Oligocene to Aquitanian
Fig. A.11. Stratigraphy of the Flysch units. After Durand-Delga and Olivier (1988).
For location see figure A.6.
micaceous flysch was discordantly deposited on folds and thrusts of the Kabylie unit and its contact with the Flysch units. In particular, Aquitanian to Early Burdigalian thick-bedded structureless quartzitic sandstone with argillaceous beds is known in the Western Mediterranean as "Numidian".

The structure of the flysch units is determined by the stratigraphy. In this context shale intervals act as efficient décollement zones at the base of thrust sheets or nappes. Major stratigraphic units coincide with major thrust sheets; each nappe therefore is characterized by its own stratigraphy.

A.3.3.2 Stratigraphy of the structural units.

- Units with argillo-quartzitic Lower Cretaceous, Cenomanian with phytanites and marly Senonian with microbreccias is represented by the thrust sheets of Mellousa and Chouammat (Durand-Delga 1965). According to Lespinasse (1990), the Mellousa unit of the northern Rif, west of Tetouan, (Durand-Delga 1965) is equivalent to the Chouammat unit in the Central Rif north of Ketama (Andrieux and Mattauer 1962). The Chouammat Nappe is equivalent to the Albo-Aptian flysch defined by Glangeaud (1962) in Algeria. The Tiserene unit which is made up only of Lower Cretaceous is located underneath the Chouammat unit (Michard 1976) or else in a lateral relationship with it (Lespinasse 1990).

- Units consisting of Upper Cretaceous and Paleogene turbidites and well developed Eocene carbonates or conglomerates and Oligocene micaceous sandstones are represented by the thrust sheet of Beni-Ider.

- The Numidian unit can be considered as the uppermost structural unit of the flysch domain. It consists of Oligo-Miocene well-rounded coarse quartz-rich sandstones, with occasionally interbedded micaceous pelites, attaining an average thickness of 500 meters in the Rif. The Numidian unit and its equivalent in the Betic Cordillera, the "Arenisca del Aljibe", represent Upper Aquitanian to Burdigalian onlapping turbiditic sandstones (some of these deposits could be basin floor fans? in the sense of Vail et al. 1991). They consist
of massive or amalgamated deep-water quartz rich sandstones interbedded with deep-water hemipelagic shales. The Numidian Sandstone outcrops in hills located between Tanger and Chaouen (Norther Rif). In 1958 Mattauer and Durand-Delga recognized these facies as the ones previously defined in Algeria. The Numidian sandstones commonly overlie the Senonian of the Tanger unit, represented by interbedded m-scale shale with Tubotomaculum and black shale. But they can also overlie the cover of the Tanger unit, that is, the Oligo-Miocene Larache Sandstone ("gres de Larache"). These facies overlie metric scale shale beds with Tubotomaculum interbedded with black shales. According to zircon provenance studies (Lancelot et al. 1977), the source area for these materials is presumed to be in Africa, probably in eastern Tunisia. It is a repeatedly reworked sandstone that corresponds to several sedimentary cycles. The sediments located below the Numidian were paleontologically dated as Lower Oligocene in the Eastern Rif (Leblanc and Feinberg 1982). The Numidian Flysch occupied the southern part of the Massylian-flysch domain and the northernmost portion of the External domain (Wildi 1983).

In the Betic Cordillera the Aljibe Sandstone is equivalent to the Numidian of northern Africa. This sandstone is extensively represented in the Campo de Gibraltar area. The Aljibe sandstone overlies Aquitanian limestones (Didon 1973, Peyre 1974). The Aljibe unit is the highest structural thrust sheet in the flysch zones of the Betic Cordillera. The Aljibe Sandstone consists of >75% well-rounded and polished quartz and < 10% of matrix. Sandstone beds are amalgamated and channelized, presenting dish and flame structures. These deposits have been interpreted as contourites (Wezel 1970, Pendón 1978) or as fluxo-turbidites for Lespinasse (1990). There is some diachronism in the timing of the Numidian Sandstone around the Western Mediterranean from Sicily to Gibraltar (Martín-Algarra 1987) even though the main sedimentary influx ranges between the Upper Aquitanian and the base of the Burdigalian. The Numidian Sandstone overlaid
imbricates of previously deposited turbiditic deposits of the Tanger and Ketama units.

The so-called "series externes rifaines" constituted by the Upper Cretaceous of the Tanger and Ketama units represent the least transported of the flysch allochthonous units (Michard 1976). They were probably in continuity with the passive margin sequence of the Maghrebian domain.

A.3.4 STRUCTURE

Many interpretations have been proposed to explain the complex structure of the flysch units (Fig. A.12). According to Didon (1969) the structure of the flysch domain is made up of a stack of thrust sheets. For Bourgois (1973) and Didon (1977) the structure of this region is the result of gravity-driven tectono-sedimentary processes (i.e. gravitational nappes). Esteras (1982) proposed a mixed model where compressional imbrication is followed by gravitational emplacement. Martín-Serrano (1985) interpreted this domain as compressional thrust sheets related to the westward movement of the Alborán block, emphasizing the important role of decollement levels.

The structure of the flysch units is defined by multiple decollement levels, coinciding with shale intervals that result in closely-spaced imbricates. Superposition of various decollement levels results in fold and thrust interference. The flysch units consist of foreland-vergent imbricates, but occasionally backthrusting occurs at the contact with the Internal domain, overturning the initial west-vergent pile of thrust sheets (Martín-Serrano 1985, Balanyá and García-Dueñas 1988, SECEGSA 1990, IGME 1990).

The flysch units partially overthrust the External zones of the Betics and Rif, with the exception of the Jbel Zem Zem region in the northern Rif, where the Numidian unit overlies Ghomarides rocks of the Internal Zones (Durand-Delga 1964).
Fig. A.12. Structure of the Flysch Domain in the Rif Cordillera. Three cross-sections in the frontal part of the Flysch domain, region of Chouamat and Targuist. After Lespinasse (1990).
A.4. INTERNAL DOMAIN

The Internal or Alborán domain, which includes the Internal zones of the Betic and Rif Cordilleras, differs stratigraphically and structurally from the External zones. The most notable characteristics which distinguish the Internal from the External zones include: Alpine-type Triassic carbonates, Early Alpine (Cretaceous-Paleogene) polyphase compressional deformation and HP/LT metamorphism.

The Alborán domain is often visualized as a microplate initially located between the Eurasian and African plates, which collided with Iberia and Africa (Andrieux et al. 1971). The tectono-metamorphic evolution from HP/LT to LP/HT experienced by the Internal domain records subduction, extension and inversion (de Jong 1991). The collision of this plate with Iberia and Africa during the Late Aquitanian to Burdigalian was accompanied by west-vergent thrusting along the Gibraltar Crustal Thrust (G. C. T.) (Balanyá and García-Dueñas 1987). Following collision, the allochthonous Alborán domain was thinned and dismembered by extension (García-Dueñas and Martínez-Martínez 1988, García-Dueñas et al. 1992). Today, this allochthonous "terrane" forms part of the Alborán back-arc basin basement. From bottom to top the Internal domain is composed of the Nevado-Filabrides, the Alpujarrides-Sebtides, the Malaguides-Ghomarides, the Dorsale and the Pre-dorsale complexes (see Fig. 3.7).

The Nevado-Filabrides and Alpujarrides-Sebtides complexes are the lowermost structural units, represented mainly by HP/LT metamorphic rocks. The upper, epimetamorphic structural units are represented by the landward Ghomarides-Malaguides complex and the more distal Dorsale and Pre-dorsale complexes portion of the Alborán passive margin.
A.4.1. THE NEVADO-FILABRIDES.

A.4.1.1. Introduction

The Nevado-Filabrides (Egeler and Simon 1969) occurs only in the Betic Cordillera and constitutes the lowermost structural unit of the Internal domain. HP metamorphic Paleozoic and Triassic rocks, exposed in the anticlinorium of the Sierra de los Filabres and Sierra Nevada, are characterized by both compressional and extensional structures.

A.4.1.2. Stratigraphy

The Nevado-Filabrides consists of two units: the lower Veleta Unit and the upper Mulhacen Unit, which in turn is composed of two thrust sheets, i.e. the Calar Alto thrust sheet and the Bedar-Macael thrust sheet (Fig. A.13). These units are bounded by the "Dos Picos" and "Marchal" thrusts (García-Dueñas et al. 1988). The Veleta Unit (also referred to as the Veleta Complex) is formed by alternating graphitic micaschists with garnet and chloritoid, and feldspar-rich quartzites. These rocks record successive stages of retrograde metamorphism (Fontboté 1983, González-Lodeiro et al. 1990, de Jong 1991). The Mulhacen Unit (also known as the Mulhacen Complex) was subdivided into four units (Fontboté 1983) from bottom to top:

1. Lower metapelitic unit. Polymetamorphosed micaschists, which also record pre-Alpine probably Hercynian metamorphism.
2. Upper metapelitic unit. Composed of micaschists with interbedded basic metavolcanics.
3. Volcanic and sedimentary unit. Consists of alternating marbles, metapelites gneisses and acid metavolcanics.
Fig. A.13. Location and stratigraphy of the Nevado-Filabrides. Stratigraphic section after García-Dueñas et al. (1987).
A.4.1.3. Ophiolites

A meta-ophiolitic association (Cobdar Formation) consisting of hundred meter thick sheets, or interrupted lenses, of harzburgites and rodingites, represents oceanic floor slices intercalated within the Mulhacen unit of the Nevado-Filabrides. The Cobdar Formation consists of volcanic, hypabyssal and plutonic ophiolitic metabasites and its sedimentary oceanic cover. Plutonic rocks consist of olivine-pyroxene gabbros, doleritic dykes and porphyritic basalts. The volcanic sequence is represented by aphanitic and amygdaloid basalts, amphibolite lenses and pillow-lavas. The geochemistry of the meta-ophiolitic association suggests a MORB ocean origin (Puga et al. 1989 a).

A.4.1.4. Structure

The Nevado-Filabrides is exposed in the core of large-scale, more than 20 Km wide, WSW-plunging antiforms (Metamorphic core complex). They consist of stacked thrust sheets bounded by syn- to post-metamorphic shear zones (García-Dueñas et al. 1988). W and WNW-directed thrusts are cut by low-angle extensional detachments with top-to-the SW transport direction (Fig. A.14) (García-Dueñas et al. 1988, García-Dueñas and Martínez-Martínez 1988, Galindo-Zaldívar et al. 1989, Soto et al. 1990). The contact between the Mulhacén and Veleta thrust sheets is a major compressional shear zone with 200 Km of top-to-the ENE horizontal displacement. This contact represents a major metamorphic boundary (Campos et al. 1986). The Nevado-Filabrides/Alpujarrides contact (Mecina fault) is an extensional detachment (Aldaya et al. 1984) with a dextral strike-slip component (Galindo-Zaldívar 1986). Extensional crenulation cleavage associated with this contact suggests top to the SW-SSW transport direction (Platt and Behrmann 1986, Galindo-Zaldívar 1986). Footwall deformation (i.e. Nevado-Filabrides) is characterized by mylonitic and brittle-ductile structures and hangingwall deformation (i.e. Alpujarrides) by brittle structures. Brittle deformation overprints ductile mylonitic fabrics as in core
Fig. A.14. Low-angle extensional detachments in the Nevado-Filabrides. After García-Dueñas et al. (1992).
complexes of the Basin and Range province of the USA (Galindo-Zaldívar et al. 1989). (Fig. A.15).

A.4.1.5. Tectono-metamorphic evolution

Pre-alpine and alpine age metamorphism occurs in the Nevado-Filabrides.

Pre-Alpine metamorphism

Pre-alpine (Hercynian ?) metamorphism is recorded in the Montenegro schists of the Mulhacen complex, by a LP paragenesis of andalusite, staurolite, chloritoid, biotite and garnet (Puga 1977, Puga et al. 1975), suggesting a pre-Alpine basement in the Mulhacen complex (Fontboté 1983).

Alpine metamorphism

Several stages of Alpine-age HP metamorphism have been distinguished (González-Lodeiro et al. 1990, de Jong 1991) in the Mulhacen and Veleta units. (Table A.3, A.4 and A.5). The Nevado-Filabrides is characterized by the superposition of the following metamorphic facies:

- Eclogitic facies.

  Early Alpine metamorphism is represented by Eclogitic facies with garnet, omphacite, kyanite calcic amphibole, paragonite, zoisite and rutile, indicating P of 12 Kb and T of 500°C.

- Blue schist facies.

  Retrograde metamorphism resulted in transformation of the initial eclogitic paragenesis into blueschists with garnet, zoisite, albite, chloritoid, paragonite and glaucophane, yielding an age of 48.4 ± 2.2 M.a. (Monié et al. 1992).

- Amphibolitic facies

  Represented by amphibolitic facies with garnet, epidote, albite, calcium-sodium amphibol, phengite and ilmenite. PT conditions of 6 Kb and 600°C are related to this episode.
Idealized detachment evolution diagram between the Alpujarra (AL) and Nevada-Filabride (NF).
1: Brittle deformation, 2: Ductile shear deformation.

Fig. A.15. Superposition of brittle-ductile structures in the Nevada-Filabrides. After Galindo-Zaldívar et al. (1989).
<table>
<thead>
<tr>
<th>Deformation phases</th>
<th>P-T conditions</th>
<th>Characteristics</th>
<th>Relationship to other nappes</th>
</tr>
</thead>
<tbody>
<tr>
<td>$D_{1}^{rd}$</td>
<td></td>
<td>micro fabric in porphyroblasts and microlithons</td>
<td></td>
</tr>
<tr>
<td>$D_{2}^{rd}$</td>
<td>$\geq 425^\circ C$; $\leq 0.6-0.7$ GPa</td>
<td>main tecto-metamorphic phase upwards increasing rotational component; ESE-WNW stretching; top-to-the-west shear in quartz mylonites in the top</td>
<td>W-ward overthrusting of the overlying Mulhacen Complex</td>
</tr>
<tr>
<td>$D_{3}^{rd}$</td>
<td>400-450$^\circ C$; 0.35-0.45 GPa</td>
<td>tight S-vergent folds</td>
<td>folding of the contact with the Mulhacen Complex</td>
</tr>
<tr>
<td>$D_{4}^{rd}$</td>
<td>retrogression</td>
<td>mylonites and ec'c's concentrated at the contact with the Mulhacen Complex</td>
<td>reactivation of the contact with Mulhacen Complex, coeval with $D_{5}^{\text{eq}}$, mylonitization in the top of the Mulhacen Complex during overthrusting of Alpujarride Complex</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Deformation phases</th>
<th>P-T conditions</th>
<th>Characteristics</th>
<th>Relationship to other nappes</th>
</tr>
</thead>
<tbody>
<tr>
<td>$D_1^{\text{mnh}}$</td>
<td>475-525°C; 0.9-1.1 GPa</td>
<td>in glaucophane schists and gneisses; ESE-WNW stretching</td>
<td></td>
</tr>
<tr>
<td>$D_2^{\text{mnh}}$</td>
<td>525-575°C; decompression: 1.1 to 0.7 GPa</td>
<td>main tectono-metamorphic phase transposition foliation ESE-WNW stretching; upwards increasing rotational component</td>
<td>W-ward thrusting over the underlying Veleta Complex</td>
</tr>
<tr>
<td>$D_3^{\text{mnh}}$</td>
<td>400-450°C; 0.35-0.45 GPa retrogression</td>
<td>local S-vergent folding associated with S-ward thrusting</td>
<td>folding of the contact between with the Veleta Complex</td>
</tr>
<tr>
<td>$D_4^{\text{mnh}}$</td>
<td>425-525°C; 0.2-0.3 GPa reheating climax</td>
<td>km-scale folds, local small scale structures; N-S compression</td>
<td>folds cut off by basal thrust of the overlying Alpujarride Complex</td>
</tr>
<tr>
<td>$D_5^{\text{mnh}}$</td>
<td>retrogression of 500 to 400°C</td>
<td>mylonitization at the contact with the Alpujarride Complex</td>
<td>N-ward overthrusting of Alpujarride Complex</td>
</tr>
<tr>
<td>$D_6^{\text{mnh}}$</td>
<td>≤ 400°C;</td>
<td>folds associated with and cut by brittle-ductile shear zones at contact with Alpujarride Complex and internal detachments</td>
<td>imbrication of the contact between the Mulhacen and Alpujarride Complex</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>$P-T-t$</th>
<th>$P$ (GPa)</th>
<th>$T$ (°C)</th>
<th>$t$ (Ma)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ar2</td>
<td></td>
<td>360 ± 50</td>
<td>21–22</td>
<td>Modelled (^{40})Ar/(^{39})Ar phengite cooling ages, inferred from (^{40})Ar/(^{39})Ar age plateaus</td>
</tr>
<tr>
<td>D4</td>
<td>0.25–0.45</td>
<td>450–525</td>
<td></td>
<td>Peak heating</td>
</tr>
<tr>
<td>D3</td>
<td>0.3–0.4</td>
<td>350–425</td>
<td>ca. 25</td>
<td>Low-grade metamorphism. Starting heating + partial thermal resetting of (^{40})Ar/(^{39})Ar cooling ages, inferred from phengite single grain</td>
</tr>
<tr>
<td>Ar1</td>
<td></td>
<td>360 ± 50</td>
<td>30–31</td>
<td>Oldest (^{40})Ar/(^{39})Ar cooling ages in core of phengite single grain</td>
</tr>
<tr>
<td>C</td>
<td>0.4–0.55</td>
<td>400–500</td>
<td></td>
<td>Cooling trajectory</td>
</tr>
<tr>
<td>Rb1</td>
<td>0.7–1.0</td>
<td>500 ± 50</td>
<td>65 ± 10</td>
<td>(^{87})Rb/(^{86})Sr phengite cooling age</td>
</tr>
<tr>
<td>D2</td>
<td>0.7–1.0</td>
<td>525–575</td>
<td></td>
<td>Medium-grade metamorphism, synkinematic decompression, main tectono-metamorphic phase</td>
</tr>
<tr>
<td>D1</td>
<td>0.9–1.1</td>
<td>475–525</td>
<td></td>
<td>Synkinematic eclogitization / glaucophane schist facies</td>
</tr>
<tr>
<td>B</td>
<td>1.0–1.1</td>
<td>375–425</td>
<td></td>
<td>Static eclogitization</td>
</tr>
<tr>
<td>A</td>
<td>1.0–1.1</td>
<td>300</td>
<td></td>
<td>Incipient eclogitization</td>
</tr>
</tbody>
</table>

• Green schist facies

Amphibolitic facies have been transformed into green schists facies with albite, chlorite and actinolite, as a result of retro metamorphism aged as 24.6 ± 3.6 M.a. (Monié et al. 1992).

• Hydrothermal alteration.

An important episode of volcanism between 15 and 17 Ma is responsible for ore deposition and hydrothermal alteration. Magmatism is related to crustal thinning in the Alborán domain. Heat flow anomalies are especially important in the Eastern Betics, where widespread Neogene volcanism took place (Albert-Bertrán 1979).

P-T-t paths

Several P-T-t paths have been proposed for the Nevada-Filabrides Complex. Gómez-Pugnaire and Fernández-Soler (1987) proposed a continuous P-T-t path. On the contrary Puga and Díaz de Federico (1978) and De Jong et al. (1991), consider a non-continuous metamorphic evolution, characterized by two stages related to subduction, interrupted by an episode of uplift and exhumation. Several phases of deformation associated with the tectono-metamorphic evolution of the Veleta and Mulhacen units have been distinguished (De Jong 1991, Van Wees et al. 1992) (see Tables A.3, A.4 and A.5).

A.4.2. THE ALPUJARRIDES-SEBTIDES.

A.4.2.1. Introduction

The Sebtide Complex is the lowermost structural unit of the Internal zones of the Rif. Its equivalent, the Alpujarride Complex of the Betic Cordillera (Van Bemmelen 1927), overlies the Nevada-Filabride Complex. The Alpujarrides-Sebtides are composed of two major lithostratigraphic groups: a lower group composed of Paleozoic basement
rocks and an upper group consisting of a thin Triassic cover. Post-Triassic rocks are not present in this complex (Fontboté 1983, Balanyá and García-Dueñas 1989) (Fig. A.16).

A.4.2.2. Stratigraphy


Two lithostratigraphic groups have been distinguished: a lower metapelitic group containing metapelites with metapsammitic interbeds and an upper carbonatic group, dated as Middle-Upper Triassic (Gonzalo and Tarín 1882, Fallot et al. 1954). Each unit presents from bottom to top the following lithologic intervals (Galindo-Zaldívar 1990).

Paleozoic Lower Group.

This basement unit consists of, from bottom to top:

1. High grade metamorphic orthogneiss and peridotite (the ultramafic massifs will be discussed separately).

2. Graphitic and quartzitic Paleozoic micaschists known as the Lower Metapelitic Formation in the Betic Cordillera (Fontboté 1983) and El Filali Micaschist in the Rif (Kornprobst 1974, 1976). This unit attains a variable thickness of about 3000 m.

   2.1 Graphitic micaschist, quartzitic schist and schist with quartz, biotite, muscovite, plagioclase, garnet, staurolite and andalucite. Quartzites and schists are characterized by cross-bedding, graded-bedding and parallel lamination.

   2.2 Quartzite and quartzitic schist with quartz, biotite, mica, chlorite, plagioclase, chloritoid, andalusite iron oxides and occasional clinopiroxene and epidote. Migmatites and paragneisses with sillimanite, kyanite and K feldspar occur only in the upper thrust sheets of the Alpujarrides-Sebtides.
3. Grey metapelite and quartzite. Metapelite with interbedded quartzite, calco-schist or gypsum, and occasionally mafic volcanic rocks. The metapelite of the Lower and Intermediate thrust sheets contains: mica, chlorite, quartz, albite, calcite and iron oxides. This metapelitic unit is referred to as "Unités de Federico" in Morocco (Milliard 1959 in Durand-Delga et al. 1960-62). The "Unités de Federico" are present in the Nappes of Beni-Mezaal and Tisgarin and in the upper metapelitic formation of the Betic Cordillera. This succession is believed to be Permo-Triassic because of its continuity with Triassic rocks (Fontboté 1983).

**Triassic Upper Group.**

A Triassic carbonate unit consisting of limestones, marbles and dolomites interbedded with calcareous slates and pelites top the Alpujarrides-Sebtides (Fontboté 1983, González-Lodeiro et al. 1990). Dasycladacea and Ostracoda of the lower and intermediate thrust sheets are Middle-Upper Triassic (Flügel et al. 1984). This lithologic assemblage is characteristic of Alpine-type Triassic facies.

**A.4.2.3. Tectono-metamorphic evolution.**

The Alpujarrides show a tectono-metamorphic evolution similar to the Mulhacen Complex of the Nevada-Filabrides, characterized by plurifacial metamorphism during polyphase deformation (Table A.6). There is a reverse relationship between structural position and metamorphic grade. The upper thrust sheets represent the highest metamorphic grade while the lower ones represent the lowest grade. However with an overall reverse metamorphic polarity, the polarity within each single unit is normal (González-Lodeiro et al. 1990). Four tectono-metamorphic phases have been distinguished in the Alpujarrides (de Jong 1991) (see Table A.6). The metamorphic evolution is characterized by the following succession of facies.
<table>
<thead>
<tr>
<th>Deformation phases</th>
<th>P-T conditions</th>
<th>Characteristics</th>
<th>Relationship to other nappes</th>
</tr>
</thead>
<tbody>
<tr>
<td>$D_1^{\text{Alpu}}$</td>
<td>350-400°C; 0.7 GPa</td>
<td>micro fabric in porphyroblasts and microlithons</td>
<td></td>
</tr>
<tr>
<td>$D_2^{\text{Alpu}}$</td>
<td>400-450°C; 0.6-0.7 GPa decompression</td>
<td>main tectono-metamorphic phase transposition foliation ESE-WNW stretching</td>
<td></td>
</tr>
<tr>
<td>$D_3^{\text{Alpu}}$</td>
<td>525-575°C; 0.15-0.35 GPa retrogression</td>
<td>tight N-vergent folds downward increasing intensity strongly curvi-linear folds in base of overthrust mass</td>
<td>N-ward overthrusting on the underlying Mulhacen Complex by gravity spreading with local lateral extension</td>
</tr>
<tr>
<td>$D_4^{\text{Alpu}}$</td>
<td>≤ 400°C; 0.15 GPa</td>
<td>folds associated with and cut by brittle-ductile shear zones at contact between tectonic units and other nappe complexes</td>
<td>imbrication of contacts with Mulhacen Complex and with the overlying Malaguide Complex; formation of contact with Almagride Complex</td>
</tr>
</tbody>
</table>

- Eclogite facies

According to Tubia and Gil Ibarguchi (1991) eclogites (partially retrograded into amphibolites) of the Ojen Nappe (Central Betic Cordillera) record subduction in the Alpujarride Complex.

- Blue schist facies

Blue schist facies represented by synkinematic growth of blue amphiboles: (i.e. glaucophane and croxite) constitute the first tectono-metamorphic episode. The presence of grossular and the absence of omphacite indicates temperatures below 400 °C (Bakker et al. 1989). Cm-scale tight to isoclinal folds with axial plane cleavage affecting quartz veins and bedding planes would be related to this metamorphic event.

- Green schists

Retrograde metamorphism resulted in transformation of blue schists into blue-green amphiboles. Increase in temperature was manifested by widespread kyanite growth. This metamorphic phase is associated with open to tight north-vergent folds and thrusts. Staurolite and andalusite growth suggests a drop in pressure and increase in temperature. Breakdown of staurolite into chlorite, oxy-chlorite and biotite points to a decrease in temperature. This metamorphic phase could be represented by open to tight folds and brittle-ductile shear zones associated with the contact between the Mulhacen (Nevado-Filabride) and Alpujarride complexes (de Jong 1991).

A.4.2.3. Structure

The Alpujarrides-Sebídides mainly comprises two sets of thrust sheets, a lower Los Reales thrust sheet and an upper Blanca thrust sheet. The Los Reales thrust sheet consists of three structural units or nappes (Table A.7). Major thrust sheets are punctuated by synmetamorphic mylonitic zones (Balanyá and García-Dueñas 1991b).
Betic Cordillera

Blanca thrust sheet group
  Ojen thrust sheet
  Guadaiza thrust sheet

Los Reales thrust sheet group
  Benarrabá imbricates
  Jubrique thrust sheet
  Bermeja thrust sheet

Rif Cordillera

Unites de Federico
El Filali unit
Beni-Bousera unit.


The rocks of the Alpujarride-Sebtide Complex record several stages of deformation. Synmetamorphic compressional deformation under ductile/brittle conditions related to crustal thickening are overprinted by extensional detachments, commonly developed under brittle conditions (Balanyá et al. 1987, Simancas and Campos 1988, González-Lodeiro et al. 1990).

Compressional structures.

Kornprobst (1974) and Olivier (1990) recognized several phases of deformation in the Sebtides of the Rif. The first phase is characterized by bedding-parallel centimeter-scale isoclinal folding with axial plane schistosity, coeval with sillimanite, biotite, muscovite, almandine and plagioclase growth. E-W trending folds, crenulation cleavage and kink-bands affect previous structures, constituting the second deformatinal phase.

A number of thrusting episodes have been distinguished in the Betic Alpujarrides according to the most recent publications.

- First episode: Synmetamorphic compressional structures under ductile conditions. (crustal thickening). Fabric analysis of synmetamorphic mylonitic foliation indicates
top-to-the ENE or NE sense of displacement (Balanyá et al. 1987, Galindo-Zaldivar 1990). The most significant structure of this episode is the "Los Reales" crustal thrust in the Internal-Western Betics (Balanyá et al. 1987).

- Second episode: Post-metamorphic episode. Takes place under ductile-brittle to brittle conditions. It is characterized by north-vergent folds (fold-nappes structures), resulting in s-c fabrics and cataclasis "fault gauges" (Simancas and Campos 1988) and imbrication (Estevez et al. 1985, Cuevas 1991). Displacement direction related to this episode is top-to-the NNW (Balanyá et al. 1987, Simancas and Campos 1988, Cuevas 1988). E-W trending folds are overprinted by N-S oriented folds, resulting in fold interference (Sanz de Galdeano 1989).

- Third episode N-NW-directed brittle thrusting consistent with the Gibraltar Crustal Thrust (G.C.T.) transport direction (Balanyá and García-Dueñas 1987).

**Extensional structures**


Thrusts and strike-slip faults disrupt the initial stacking sequence of the Alpujarrides imbricates in the Western and Central Betics (Galindo-Zaldivar 1986, Sanz de Galdeano 1989, Balanyá and García-Dueñas 1991).

**A.4.2.4. Ultramafic massifs**

Ultramafic massifs of alpine-age peridotites are present in the Alpujarride-Sebtiide Complex of the Betic and Rif Cordilleras. They represent upper mantle slabs thrusted onto upper crustal rocks (Kornprobst 1974, Tubia and Cuevas 1986). Extension following compression thinned and dismembered these slabs into a number of isolated massifs

**Betic Cordillera**

The ultramafic massifs of the Betic Cordillera are located in the upper thrust sheet of the Alpujarrides, referred to locally as "Los Reales Nappe". The peridotic succession consists of layered lherzolite, harzburgite and occasional gabbros. Hydrothermal alteration resulted in transformation of peridotite into serpentine, antigorite, crisotile and talc (Lundeen 1978, Tubia 1985, Balanyá and García-Dueñas 1991).

**Rif Cordillera**

In the Rif Cordillera, the ultramafic massif of Beni-Bousera is the largest exposure of Alpujarride-Sebta peridotites. The Beni-Bousera ultramafic massif is an antiform with regular zonation. The massif is cored by banded spinel and lherzolite with thin picroxene layers, surrounded by harzburgites, dunites and garnet-bearing dunites (Reuber et al. 1982). Kinzigites and migmatitic rocks (El Filali Gneiss) grading upward into micaschist with staurolite-kyanite-andalusite (El Filali Micaschist) represent the cover of the ultramafic rocks (Fallot 1937, Kornprobst 1974). Permo-Triassic rocks unconformably overlie El Filali Micaschists.

**Peridotitic Emplacement**

Several stages of deformation have resulted in the present-day structure of the ultramafic massifs of the Gibraltar Arc (Tubia 1985, Tubia and Cuevas 1986, Balanyá and García-Dueñas 1991). These stages can be summarized as follows:

1. Extension that delaminates the lower crust on top of the peridotites. This episode is manifested by stretching lineations within peridotitic mylonites (Tubia 1985).
2. Thrusting that results in the allochthony of the peridotite. The ultramafics of Sierra Bermeja (Lundeen 1980) and Sierra Alpujata (Tubia 1985) were emplaced by thrusting.
Peridotite was emplaced at high temperature, resulting in syntectonic migmatization and ductile thrusting (i.e., mylonitization) (Tubia and Cuevas 1986). Leucogranitic melt, radiometrically dated in Sierra Alpujata-Marbella area as 22.5 ±4.0 Ma (Zeck et al. 1992), HT garnets and migmatites developed in the footwall block, the so-called "Unidad de Blanca".

3. Extensional tectonics and related thinning. This stage is associated with the formation of the Alborán Basin. Boudinage of the peridotitic bodies is related to this episode (García-Dueñas et al. 1992). Geophysical data indicate that the peridotitic massifs are rootless lithospheric slabs (Casas and Carbó 1990, Torné et al. 1992).

4. Post-extensional structures. Low-angle extensional detachments are cut by N-S and E-W trending folds and high-angle normal faults in the peridotitic massifs of Sierra de las Aguas and Sierra de la Robla (Betic Cordillera) (Soto and Gervilla 1991). The basal decollement of the peridotites was folded during Late Miocene time (García-Dueñas and Balanyá 1991) probably in connection with deep-seated extensional faults (García-Dueñas et al. 1992).

A.4.3. THE GHOMARIDES-MALAGUIDES.

The Malaguide Complex of the Betic Cordillera (Blumenthal 1927) is the equivalent of the Ghomaride Complex of the Rif Cordillera (Kornprobst in Durand-Delga 1960-62). The Malaguide-Ghomaride Complex is the highest structural unit of the stack of thrusts of the Internal domain and the only one without HP Alpine-age metamorphism.

A.4.3.1. Stratigraphy

The Ghomaride-Malaguide Complex section consists of a metamorphic basement involving Paleozoic rocks that is overlain by a thin Mesozoic and Tertiary sedimentary cover (Fig. A.17).
Fig. A.17. Map distribution and stratigraphy of the Ghomarides-Malaguïdes. Stratigraphic sections after Durand-Delga and Olivier (1988).
Paleozoic basement succession

The basement succession commences with Lower Cambrian-Upper Ordovician grey and blue phyllite, referred to locally as "filitas de color de humo" or "schistes fumees", interbedded with occasional conglomerate beds with deformed pebbles of quartz and quartzite and lenses of carbonate. Upper Ordovician and Silurian interbedded sandstone and shale follow in the section. The Silurian-Devonian transition corresponds to spilites and cherts associated with conodont-bearing limestone. The Devonian is represented in the Betic Cordillera by conodont limestones with interbedded phyllite and quartzite (Fontboté 1983). In the Rif Cordillera the Devonian section varies from one thrust sheet to another. The Aakaili Nappe is characterized by distal calci-turbidites referred to locally as "calcareous flysch". The overlying Koudiat-Tizian Nappe contains proximal calci-turbidites. The higher, Beni-Hozmar Nappe comprises only a thin succession of tentaculites-bearing pelagic limestones with chert. The highest units of Talembote and Bokkoyas present Givetian massive reeval limestones (Chalouan and Michard 1990, Olivier 1990). In the Betic Cordillera the Devono-Dinantian section consists of spilitic lavas, volcano-detritics and calci-turbidites "carbonatic flysch". Calci-turbidites are characterized by rhythmically interbedded recrystallized dark limestone and shale ("calizas alabeadas") with occasional volcanic and radiolarite beds. This carbonate succession grades into lidites, which are overlain by rhythmically interbedded graywacke, shale and occasional conglomerate with Bouma sequences. This turbiditic facies is referred to regionally as "Culm" facies (i.e. the characteristic Carboniferous syn-orogenic facies of the Hercynian folded belt). In the Rif, these Carboniferous turbiditic sediments (Culm facies) include interbedded olistostromes, radiolarites and Upper Visean foraminiferal limestones. Carboniferous volcanism is manifested in the Rif Cordillera by doleritic sills, basaltic flows and rhyolitic-andesitic pyroclastic layers (Chalouan and Michard 1990).
Frequently polymictic coarse conglomerates "Marbella Conglomerates" top the basement succession (Blumenthal 1949).

Mesozoic-Tertiary Cover.

The Mesozoic-Tertiary cover of the Malaguide-Ghomaride Complex is thin and often incomplete. The cover succession commences with Permo-Triassic red-beds consisting of fine-grained sandstone, well-rounded quartz-rich conglomerate and dolomite. Interbedded basaltic rocks are present in the upper part of the Permo-Triassic succession (Chalouan 1985). The Jurassic consists of oolithic and nodular limestones. In the Rif Liassic dolomitic breccias and white massive algal limestones with debris of foraminifera unconformably overlie Paleozoic or Permo-Triassic rocks. This succession can reach a thickness of 500 m. A karst surface represents the boundary between the Jurassic and the Cretaceous. The Cretaceous section is composed of stratified limestone with sandy interbeds, glauconite and chert and attains only 100 meters of thickness. Where present, the Eocene consists of biogenic limestone, sandstone and marl overlain by Upper Eocene nummulitic limestone and conglomerate (Durand-Delga and Olivier 1988).

A.4.3.2. Structure

The structure of the Malaguides-Ghomarides is characterized by folds and thrusts associated with penetrative deformation. In the Betic Cordillera, the Malaguides consist of a large number of imbricates (Fontboté 1983). Stratigraphic omission suggests that the Malaguide basal contact with the Alpujarride is extensional (González-Lodeiro et al. 1990). The internal structure is characterized by axial plane cleavage associated with metric and decametric folds. Fold-interference patterns related to four phases of folding have been recognized in the Montes de Málaga area (Gáiez and Orozco 1980). Three generations of structures have been described in the Malaguide Complex of Sierra Espuña:
NNW-vergent folds are overprinted by SSE-vergent backthrusts. High-angle NNW
dipping reverse faults offset previous structures (Mäkel 1985).

In the Rif, the Ghomaride Complex is made up of four Paleozoic thrust sheets
(Chalouan and Michard 1990). From bottom to top: the Aakaili Nappe, the Koudiat-
Tiziane Nappe, the Beni-Hozmar Nappe and the Talembote Nappe. As previously
mentioned each of these nappes is characterized by a distinct Devonian section. Three
main phases of deformation, two of them probably related to the Hercynian orogenesis,
have been recognized in the Ghomaride Nappes (Chalouan and Michard 1990, Olivier
1990).

1. The first phase, D1, is represented by NNE-SSW trending minor folds with axial plane
cleavage. Pressure solution and stretching lineation was associated with this first
phase.

2. The second phase, D2, corresponds to NE trending metric to decametric-scale NW-
vergent folds. Shear planes, pressure solution and stretching lineations are related to
this episode.

3. The third phase, D3, does not affect the Triassic cover of the Ghomaride Nappe. It is
defined by NW-SE trending upright folds with associated crenulation cleavage. In the
Bokoyas these folds trend E-W. This last generation of structures could be Late
Hercynian or probably Alpine in age.

A.4.4. DORSALE AND PRE-DORSALE COMPLEXES.

The Dorsale unit, initially defined by Fallot (1937) in the Rif and Algerian Tell
(“Dorsale calcaire”), was also recognized in the Betic Cordillera and named “Dorsale
Betique” (Durand-Delga and Foucault 1967). The Dorsale Complex is exposed in the
narrow and discontinuous contact between the Internal and External domains (Fig. A.17)
Fig. A.18. Map distribution and stratigraphy of the Dorsale and Predorsale units. After Durand-Delga and Olivier (1988).
and is present only in the central and western part of the Gibraltar Arc. The Dorsale of the Rif and the Bokoya of Morocco extend farther east to the Kabylie massifs of the Algerian Tell (Wildi 1983). In the Betic Cordillera, the Dorsale has been subdivided into Internal and External Dorsale Units (Junta de Andalucia 1985). In the Rif Cordillera it has been subdivided into External, Intermediate and Internal Dorsale units (Wildi et al. 1977, Suter 1980b). The Internal Dorsale Unit is thought to be the cover of the Malaguide-Ghomaride Complex (Balanyá 1984), while the External Dorsale is the transition with the Pre-Dorsale Complex. The Pre-Dorsale Complex is a discontinuous strongly-deformed unit with similar but relatively deeper water deposits compared to the Dorsale units. The sedimentary succession of the Dorsale represents the rifted carbonate platform that rimmed the Alborán domain (Durand-Delga and Olivier 1989). Collision resulted in thrusting onto the south Iberian and north African margins and onto the Flysch units in the western part of the Gibraltar Arc.

A.4.4.1. Stratigraphy

As in the Malaguide-Ghomaride Complex, the Dorsale unit comprises Mesozoic limestone and dolomite and is unconformably overlain by Upper Eocene, Oligocene and Lower Miocene siliciclastics and marls. The Mesozoic succession commences with Alpine-type Triassic. It consists of Triassic massive dolomite and breccia, with recrystallized dolomitic-sparitic texture and Rhetian (Upper Triassic) interbedded limestone and dolomite. The Jurassic section consists of a 100 meter thick section of Lower Liassic massive grey limestone and 250 m. of Middle-Upper Liassic limestone with detritic silex and glauconite. The upper portion of the section consists of Eocene red marls unconformably overlain by Upper Oligocene-Lower Aquitanian quartzo-feldspatic calcite-cemented sandstone (Wildi et al. 1977, Wildi 1983, Olivier 1990).
A.4.4.2. Structure

The structure of the Dorsale Complex is the result of Mesozoic rifting, followed by Neogene compression linked to the building of the Gibraltar Arc and subsequent extension coeval with the opening of the Alborán Sea. Following is a description of the structures related to these major geodynamic events.

Structures related to rifting

The internal structure and Jurassic facies and thickness distribution suggest tilted block geometry associated with extension in the Dorsale of Haouz (Northern Rif) (El Hatimi et al. 1991). In the Bokoyas, mega-breccias and turbidites related to Liassic rifting have been described by Mouhssine et al. (1990). Lithology and thickness change in dip section, but lateral continuity along strike suggests elongated paleogeographic domains consistent with a rifted margin configuration.

Compressional structures

In the Central Betic Cordillera (Sierra Harana), the structure of the Dorsale is the result of several thrusting events. Northward-vergent imbricates are cross-cut by southward-directed thrusts. The thrust stack is offset by a high-angle reverse fault with dextral strike-slip component, that represents the boundary between the Internal and the External domain. Close to this boundary, the structure of the Dorsale Complex is characterized by backthrusting (Balanyá 1984, Balanyá and García-Dueñas 1986). In the Rif Cordillera backthrusting involving Lower Miocene sediments has also been reported along the Internal/External domain contact (Ben Yaïch et al. 1986).

Neogene extensional structures

Neogene extension connected with the opening of the Alborán Sea overprints previous structures. Low- and high-angle extensional detachments take advantage of previous thrust planes (negative inversion) (García-Dueñas et al. 1992), resulting in stratigraphic omission (Wildi et al. 1977, Suter 1980, SECEG 1990, own observations).
A.4.5. PRE-DORSALE COMPLEX.

The Pre-Dorsale Complex (Durand-Delga 1972) consists of tectonic slices with unusual stratigraphy, discontinuously located between the External Dorsale and the Tisirene or Beni-Ider flysch units. The Pre-Dorsale Complex may be thrust onto more external units or else back-thrust onto the Ghomarides-Malaguides. The paleogeographic position of this unit is not well known. According to Didon et al. (1973) the Pre-Dorsale succession may represent a pelagic high located between the External Dorsale and the flysch units. Lespinasse (1975) considers the Pre-Dorsale as the External Dorsale cover. For other authors like Wildi et al. (1977), the Pre-Dorsale does not represent an intermediate section between the flysch and the Internal units but rather a detached and more external unit than the Dorsale.

The Pre-Dorsale is characterized by a Liassic to Lower Miocene sedimentary succession (Olivier 1984, Durand-Delga and Olivier 1989). The most conspicuous stratigraphic features of this complex are the presence of Lower Cretaceous pelagic sediments, the absence of turbidites "flysch", and Numidian-type Aquitanian sandstones. The Jurassic section is similar to the External Dorsale unit and consists of Dogger-Malm silica with red radiolarites. Variegated marl, normally red, with planktonic and benthic foraminifera of Upper Eocene-Oligocene age overlie the radiolaritic Jurassic succession (Durand-Delga and Olivier 1989, Olivier 1990). Blocks of Triassic, Jurassic and Middle Cretaceous limestones ("klippes sedimentaires") (Olivier 1990, Ben Yaïch et al. 1988, Ben Yaïch 1991) and reworked fauna from the Dorsale are common in the section. Numidian type sandstones, referred to locally as "Numidoide" facies, overlie these deposits.
A.5 THE NEOGENE COVER OF THE INTERNAL DOMAIN

Neogene sediments fill widespread extensional or pull-apart intramountain basins that unconformably cover rocks of the Betic and Rif Internal domains. The Neogene succession records the extensional events that may or may not be related to the opening of the Alborán Sea and subsequent inversion of the early extensional structures and late uplift of the orogenic belt. These sediments, classically referred to as "post-orogenic", represent syn- and post-extensional deposits of the Alborán domain. The study of these sediments is important to constrain the late evolution of the orogenic belt. The distribution of the Neogene is quite irregular and variable throughout the Internal domain of the Gibraltar Arc. In the Betic Cordillera the Neogene is exposed on small outcrops as a tabular cover and in pull-apart basins mainly located in the Eastern Betics. In the Rif Cordillera the Neogene is exposed in small valleys (Oued Laou, Oued Martil of the Mediterranean slope and in the Tanger Peninsula); southwest of Sebta (Ceuta) Oligo-Miocene deposits overlie the rocks of the Internal zones.

A.5.1 LATE OLIGOCENE-LOWER AQUITANIAN

In the northern Tanger peninsula, southwest from Sebta, the Ghomaride basement is covered by basal conglomerates with reworked pebbles of Sebtide provenance, overlain by marls with Upper Oligocene fauna (Fnidek Formation). These Oligo-Miocene sediment seal Ghomaride thrusts, but are cut by the contact with the Dorsale and the Jbel Zem Zem region. They evidence an Oligo-Miocene transgression on the Internal Rif and represent the oldest sediments that cover the Internal domain in the Rif (Feinberg et al. 1990).
Table A.8. Late Oligocene-Lower Aquitanian sedimentary units that cover the Internal domain of the Betic Cordillera.

In the Betic Cordillera, shallow marine sands, clays and reddish conglomerates unconformably overlie the Malaguide Complex which constitutes the source area. These deposits post-date thrusts within the Malaguides and coincide with the first cooling event of the Nevado-Filabrides (Rodríguez-Fernández and Sanz de Galdeano 1992).

A.5.2 AQUITANIAN-LOWER BURDIGALIAN

In the Tanger Peninsula south of Oued Martil, Aquitanian-Lower Burdigalian deposits overlie the Ghomaride unit. This unit, referred to as the Sidi Ablesman Formation (Feinberg et al. 1990), consists from bottom to top of:

1. Bedded breccias with angular blocks of Paleozoic rocks interbedded with sandstones and clays.

2. Fine-grained calcarenites with interbedded marl intervals.

3. Grey, locally limy pelites topped by white silexites of Lower Burdigalian age.

The first transgressive deposit on the Internal zones of the Betic Cordillera in the Marbella-Estepona and Lecrín Valley regions includes breccias with angular blocks and intercalations of marly-limestone and turbidites in the upper part. Basal breccias unconformably overlie the Malaguide and Alpujarride basement, which is the source for these Lower Burdigalian to Lower Serravallian deposits (Martín-Pérez and Aguado 1990, Aguado et al. 1990).
Fuente Formation  Mac Gillavry et al. 1963, Soediono 1971
Alamo Formation  Volkand Rondeel 1964
Alamillos Formation  Rodríguez-Fernández 1982
Viñuela Formation  Boulin and others 1973, Rivière 1988
Las Millanas Formation  Bourgois and others 1972
San Pedro de Alcántara  Aguado and others 1990

Table A. 9. Lower Burdigalian-Lower Aquitanian sedimentary units that cover the Internal domain of the Betic Cordillera.

In the Western Betic Cordillera, large outcrops of Early Miocene deposits unconformably overlie the Alborán domain basement. This transgressive sequence of conglomeratic formations (Bourgois 1978) is referred to as the Alozaina sedimentary complex (Balanyá and García-Dueñas 1987, Balanyá and García-Dueñas 1991, García-Dueñas et al. 1992). This unit was initially included in the Pre-Dorsale Complex (Fontboté 1983). The stratigraphic succession of the Alozaina complex consists of Lower Burdigalian marine conglomerates overlain by clays, marls and turbidites with interbedded olistostromes and sedimentary klippes of flysch, referred to as Neonumidian (Bourgois 1978, Balanyá and García-Dueñas 1988).

These Aquitanian-Lower Burdigalian deposits are coeval with the denudation of the partially emerged reliefs of the Betic zone, coinciding with the second cooling peak of the Nevado-Filabrides (de Jong 1991, Rodríguez-Fernández and Sanz de Galdeano 1992).

A.5.3 UPPER BURDIGALIAN-LOWER LANGHIAN

Sediments of this age are scarce and poorly known. They consist mainly of grey calcareous marls, conglomerates and turbiditic sandstones with blocks or fragments of Alpujarride and Malaguide rocks. The deposition of these units coincides with the last

Espejos Formation Soedonjo 1971, Hermes 1985
Aguilas deposits Montenat and others 1978

Table A.10. Upper Burdigalian-Lower Langhian sedimentary units that cover the Internal domain of the Betic Cordillera.

A.5.4 UPPER LANGHIAN-SERRAVALLIAN

The Alpujarride basement is unconformably overlain by a conglomeratic formation made up by fragments and grains of Alpujarride and Malaguide provenance. Lower Serravallian transgressive deposits consisting of bioclastic platforms that changed laterally to yellow marl with interbedded turbidite sandstone levels overlie the basal conglomeratic unit. The Serravallian deposits consist of marginal clastic wedges of marine breccias and continental conglomerates, grading into central lacustrine formations of lutites with gypsum and limestones.

La Peza Formation Rodríguez-Fernández 1982
Umbría Formation Volk and Rondeel 1964

Table A.11. Upper Langhian-Serravallian sedimentary units that cover the Internal domain of the Betic Cordillera.

A.5.5 Tortonian

Lower Tortonian detritic sediments overlie the Alpujarrides and locally the Nevada-Filabrides. These deposits represent the first occurrence of detritus from the Nevada-Filabrides basement (Rodríguez-Fernández and Sanz de Galdeano 1992).
In the Ronda area (Western Betic Cordillera) Lower Tortonian detritic deposits unconformably overlie thinned South-Iberian and flysch units (García-Dueñas et al. 1992).

In the Central and Eastern Betic Cordillera E-W aligned Neogene polygonal basins were formed during Tortonian time (i.e. Ugijar-Sorbas, Guadix-Baza and Granada Basins). Extension and thinning of Alborán ceased during Tortonian time and strike-slip played the dominant role (Sanz de Galdeano 1990, Rodríguez-Fernández and Sanz de Galdeano 1992). The sedimentation in these basins commences during Tortonian time and is coeval with the Intra-mountain basins of the Central Rif.

The base of the Upper Tortonian is characterized by fan delta wedge deposition along the margins of the basins, that grade into prograding platforms and channelized breccias along the basin axis. These deposits record major uplift of the Nevado-Filabrides Complex.


A.5.6 MESSINIAN

In the Eastern Betics, Messinian deposition in the intra-mountain basins was characterized by marginal reefs grading into central evaporites, related to the "Messinian salinity crisis" (Saint Martin and Rouchy 1986, 1990, de Santisteban 1981). Pliocene-
Pleistocene alluvial fan deltas grading into shallow lacustrine deposits mark the beginning of the continentalization process.

A.5.7 INTRAMOUNTAIN BASINS OF THE CENTRAL RIF

E-W trending Neogene intra-mountain basins overlie allochthonous thrust sheets of the Central Rif (i.e. Middle Ouerha region, Dhar Souk, Taounate, Tafrant de Ouerha, and el Ksibat). These basins are syncloria on N-S sections but show half-grabens on E-W sections. (Morel 1988, this study). The substratum consists of allochthonous sediments of the Ketama, Aknoul and Prerifaine Nappes. The Neogene section consists of Lower Tortonian to Messinian strata, characterized by abundant facies changes, olistostromes, breccias and slumps. These features suggest synsedimentary tectonics. Pliocene is often not present. The stratigraphy of the intra-mountain basins can be summarized from bottom to top as follows (Morel 1988, Wernli 1989, own observations):

1. Rounded sandstones and conglomerates lie directly on top of the basement. Polygenic well rounded and cemented conglomerates also known as "poudinge" grade into red sandy-marl with interbedded conglomerates and levels of soils. Ferruginized pisolithes are common within this interval. An erosional surface separates the previous sediments from marine well-stratified, cross-bedded and channelized conglomerates with bioclasts of *Pecten sp.*, brachiopods, algae, bryozoa, echinoderms and benthic foraminifera, referred to locally as "molasse de base". Syn-sedimentary normal faults are common.

2. Grey-blue marine marls m5b of Wernli (1988) with interbedded well-stratified grey sandy-marls with large brachiopods and rich in terrigenous, bioclastic components and glauconite. The planktonic/benthonic ratio is 1/1 and is dominated by *G. dutertrei*, *G. humerosa* and *G. plesiotumida*. Pinch and swell structures, occasional interbedded nodular marly-limestone levels and well stratified sandstone in centimetric bancs with marly
interbanks and gypsum characterize this group (Wernli 1988). The succession is topped by conglomerates. A hiatus separates these facies from the overlying Villafranchian sandstones and conglomerates.
CHAPTER 8
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This NW-SE trending seismic line displays Intra-Nappe reflectors underneath the basal decollement of the accretionary wedge. In the southeastern portion of the section the internal structure of the Nappe consists of folds detached from a basal decollement located at 3.7 sec (TWT). Two sets of normal faults offset the top of the accretionary wedge. Low angle listric normal faults are cut by steeply dipping normal faults.
This seismic line displays the wedge-like character of the Prerifaine Nappe. The internal structure of the wedge itself, consists of several northeastward dipping thrust faults that define a zone of frontal imbrications. Thrusts emerge from a basal decollement located at 4 sec (TWT) between crossing 11 and 13. The Supra-Nappe succession is characterized by an angular unconformity that separates deformed beds of flat laying onlapping reflectors. The southeastern portion of the section is occupied by a "foredeep" located at the front of the accretionary wedge. The basal foredeep unconformity defines the base of the foredeep succession and merges into the basal decollement of the accretionary wedge. The Infra-Nappe succession suggest Triassic half-grabens overlying the Hercynian basement.
NW dipping basal decollement. The internal structure of the Nappe is characterized by SE dipping thrusts. This section illustrates normal faults detached at the same level than thrusts, defining toe-thrust structures. Normal faults offset the upper part of the Nappe and the Supra-Nappe reflectors.
This section displays a shallow horizontal extensional decollement associated with a slump structure that is superimposed on the Nappe. This section shows that the thickness changes of the accretionary complex coincides with nearly vertical faults that offset the basement and the Infra-Nappe reflectors. The supra-Nappe units show northwestward prograding clinoforms with slumps at their toe.
This section shows a much thinner wedge than the previous sections. Northeastward dipping normal faults offset the top of the basement and the overlying Infra-Nappe unit. An angular unconformity separates tilted Infra-Nappe sediments from onlapping Supra-Nappe sediments. The frontal portion of the nappe is topped by slumps. Between crossings 1 and 3, extensional basins bounded by northeastward dipping listric normal faults ride piggy-back on the Nappe resulting in satellite basins.
This section shows Infra-Nappe imbricates and the superposition of décollement levels. The section displays a dip view of the slumped frontal units, characterized by extensional faults in the rear, and thrust faults in the front, rooted into a shallow décollement level. Extensional faults overlie and crosscut the top of the accretionary wedge, resulting in satellite basins.
This seismic line displays the northeastward dipping Infra-Nappe succession underlying the accretionary wedge. To the southwest the Supra-Nappe sediments onlap onto these reflectors, defining the Basal foredeep unconformity. The slumped units are imaged between crossings 7 and 13. NE-dipping extensional detachments overprint the wedge in the NE sector of the line, resulting in strong Supra-Nappe rotation.
This section shows Intra-Nappe imbricates overlain by the accretionary wedge, which is characterized by internal southwest vergent thrusts. This section shows an interesting feature affecting the top of the accretionary wedge, normal fault merges into a toe-thrust. The northeastern portion of the section is occupied by a ridge.
This seismic section displays normal faults offsetting the Intra-Nappe succession. The upper part of the wedge is cut by a normal fault and its corresponding toe thrust. A typical frontal slump is present in the southwest end of the profile.
This profile displays well-imaged Infra-Nappe imbricates overlain by the Prerifaine Nappe basal decollement. In this section the Infra-nappe succession is very thin and involved in minor southwestward vergent imbricates. Extensional faults cutting the top of the Nappe occur in the northeastern end of the section (crossing 1). Normal faults passing into thrusts define toe-thrusts.
This section is characterized by a very thin Infra-Nappe succession. Southwestward dipping extensional normal faults overprint northeastward dipping thrusts of the frontal imbrications in the northeastern portion of the profile. The seismic images some folded internal reflectors within the Nappe. Normal faults turning into a thrust plane sharing the same décollement level represent the continuation of the toe-thrust feature described in the preceding sections. Note Pleistocene canyons.
This NW-SE trending seismic line displays Intra-Nappe reflectors underneath the basal decollement of the accretionary wedge. In the southeastern portion of the section, the internal structure of the Nappe consists of folds detached from a basal decollement located at 3.7 sec (TWT). Two sets of normal faults offset the top of the accretionary wedge. Low-angle listric normal faults are cut by steep-dipping normal faults.
This seismic line displays the wedge-like character of the Prentaine Nappe. The internal structure of the wedge itself is defined by several northeastward dipping thrust faults that define a zone of frontal imbricates. Thrusts emerge from a basal decollement at 4 sec (TWT) between crossing 11 and 13. The Supra-Nappe succession is characterized by an angular unconformity that deformed beds of flat lying onlapping reflectors. The southeastern portion of the section is occupied by a 'foredeep' located in front of the accretionary wedge. The basal foredeep unconformity defines the base of the foredeep succession and merges into a decollement of the accretionary wedge. The Intra-Nappe succession suggests Triassic half-grabens overlying the Hercynian base.
The internal structure of the wedge itself, consists of an overthrust system emerging from a basal decollement located at the base of the sequence. This is separated by an angular unconformity that separates an early diachronous section on the top of which is occupied by a foredeep located at the foredeep succession and merges into the basal half-grabens overlining the Hercynian basement.
The NW-dipping basal detachment of the Prerifaine Nappe and the Infra-Nappe reflectors appear offset near intersection 24. In the western portion of the line, the Infra-Nappe reflectors are not imaged by the seismic. In the eastern part of the section the base of the Nappe is located at 3.3 sec (TWT). The most conspicuous feature observed in this profile is a series of Supra-Nappe normal faults connected with thrust structures that show inversion. To the west of crossing 16, the structure is dominated by westward-dipping extensional faults offsetting the top of the Nappe.
Prerifaine Nappe

Near intersection 3. In the middle part of the section the base of Prerifaine Nappe normal faults are superposed by westward dipping
In this section the frontal imbricates of the wedge are underlain by Intra-Nappe imbricates detached from a basal decollement present between crossings 9 and 13. A ridge of the accretionary complex occurs near crossing 3. The trough located in front of the wedge in the southwestern portion of the profile represents the foredeep related to the emplacement of the Accretionary Wedge. The basal foredeep unconformity is defined by a strong reflector.
detached from a basal decollement

The trough located in front of the root of the Accretionary Wedge. The
NW dipping parasial decollement. The internal structure of the Nappe is characterized by SE dipping thrusts. This section illustrates normal faults detached at the same level than thrusts, defining toe-thrust structures. Normal faults offset the upper part of the Nappe and the Supra-Nappe reflectors.
This seismic line shows the frontal imbricates and the basal decollement of the Pre-Sicane Nappe. The top of the Nappe is defined by normal faults resulting in Supra-Nappe extensional satellite basins. The most interesting feature represented in this section is the angular unconformity separating folded northeastward-dipping Infra-Nappe reflectors from Supra-Nappe onlapping reflectors.
The Nappe is offset.

This section is the

2 Km
This section displays a shallow horizontal extensional decollement associated with a slump structure that is superimposed on the Nappe. This section shows that the thickness changes of the accretionary complex coincides with nearly vertical faults that cut the basement and the Intra-Nappe reflectors. The supra-Nappe units show northward prograding clinoforms with slumps at their toe.
This section shows a much thinner wedge than the previous sections. Northeastward dipping normal faults offset the top of the basement and the overlying Infra-Nappe unit. An angular unconformity separates tilled Infra-Nappe sediments from onlapping Supra-Nappe sediments. The frontal portion of the nappe is topped by slumps. Between crossings 1 and 3, extensional basins bounded by northeastward dipping listric normal faults ride up-gy on the Nappe resulting in satellite basins.
This section again shows an extensional detachment or slump that affects the uppermost part of the nappe. As in the previous sections, note that the nappe thickens dramatically in a westward direction. The northwestward dipping basement is offset by a series of high-angle dipping faults that coincide with thickness changes of the Intra-Nappe reflectors. Between crossings 8 and 10, the structure of the nappe is characterized by southwestern vergent imbrications that represent lateral ramps. The western part of the section, to the west of the crossing with section 8, shows northeastward dipping thrusts within the nappe. For an interpretation of Pleistocene sequences, see Figure 8.1.
This section shows intra-Nappe imbrications and the superposition of décollement levels. The section displays a dip view of slumped frontal units, characterized by extensional faults in the rear, and thrust faults in the front, rooted into a shallow décollement level. Extensional faults overlie and crosscut the top of the accretionary wedge, resulting in satellite basins.
The diagram displays a dip view of the geologic layers, indicating the presence of a canyon and the Hercynian Basement. The scale represents 2 Km along the SW direction, with TWT (two-way travel time) measurements increasing from 0 to 5 seconds.
In this strike section the basal decollement of the Accretionary Nappe is much shallower than in the previous sections so that here the wedge is much thinner than in the preceding sections. The same velocity effects described for the other sections here. West of crossing 6 the structure is dominated by thrusts within the Nappe. They are interpreted as lateral ramps of the imbricating system. Slumps overlying the top of the Nappe occupy the frontal slope of the wedge. The westward progradation of Supra-Nappe reflectors is well shown on this section.
LEGEND

Top of the Nappe

Decollements and faults

Hercynian Unconformity

Two sets of imbricates are well represented in this seismic line. Upper imbricates within the Nappe and lower imbricates. The slumped units are present between crossings 11 and 7. Extensional satellite basins can be seen northea.
This section is located close to the thin edge of the wedge. A shallow basement. In this line the basal decollement of the N. is located at 2 sec (TWT) in the southeastern portion of the section. Thickness changes of the autochthonous reflectors coincide with faults that are also displayed on this section. The thickness of Intra-Nappe reflectors increases dramatically toward the M. Angular unconformities are common within the autochthonous succession. Hypothetic inverted Triassic half-grabens are interpreted on the western side of the section.
This seismic line displays the northeastward dipping Supra-Nappe succession underlying the accretionary wedge. To the southwest the Supra-Nappe strata onlap onto these reflectors, defining the Basal foredeep unconformity. The slump units are imaged between crossings 7 and 13. NE-dipping extensional detachments overprint the wedge in the NE sector of the line, resulting in strong Supra-Nappe rotation.
This section shows infra-Nappe imbricates overlain by the accretionary wedge, which is characterized by intervergent thrusts. This section shows an interesting feature affecting the top of the accretionary wedge, normal faulting and thrust. The northeastern portion of the section is occupied by a ridge.
which is characterized by internal southwest
progression. A secondary wedge, normal fault merges into a toe-

The wedge is underlain by intra-Nappe imbrications. Southwestern dipping thrusts within the nappe characterize the eastern portion of the section. Normal faults offset the top of the nappe and a ridge in the northeastern portion of the section.
In this section the basal decollement of the Pretiiane Nappe can be seen along the whole line. Frontal imbricates are over this basal decollement, in the northwestern portion of the section northeastern and southwestern dipping extensional faults cut top of the Nappe. The frontal part of the wedge is overlain by a slump on top of the leading edge of the accretionary wedge.
PLATE 22B

LOCATION MAP

LEGEND

Top of the Nappe

Decollements and faults

Hercynian Unconformity

40 Km

This seismic section displays normal faults offsetting the Infra-Nappe succession. The upper part of the wedge is cut by a normal fault and its corresponding toe thrust. A typical frontal slump is present in the southwest end of the profile.
This profile displays well-imaged Intra-Nappe imbricates overlain by the Peri-Anne Nappe basal decollement. In this section the Intra-nappe succession is very thin and involved in minor southwestward vergent imbricates. Extensional faults cutting the top of the Nappe occur in the northeastern end of the section (crossing 12). Normal faults passing into thrusts define toe-thrusts.
This section is characterized by a very thin intra-Nappe succession. Southward-dipping extensional normal faults and northeastward-dipping thrusts of the frontal imbricacy in the northeastern portion of the profile. The seismic images some internal reflectors within the Nappe. Normal faults turning into a thrust plane sharing the same decollement level represent continuation of the toe-thrust feature described in the preceding sections. Note Pliocene canyons.
EXPLANATORY NOTE

The most characteristic features
dip sections are:

1. Wedge-like character of the Prerifai

2. Northward dip of the basement
Nappe succession underneath
Nappe.

4. Extensional faults offsetting the top of the Nappe.

5. Supra-Nappe clinoformal pattern related to southwestward progradation.
MAIN STRUCTURAL FEATURES

7. Toe-thrust
6. Extensional Lateral Ramp
5. Llistric normal fault
4. Low angle Extensional detachment
3. Thrusts within the Nappe
2. Prerifal Nappe basal detachment
1. Infra-Nappe Imbricates
PLEASE NOTE:

Oversize maps and charts are filmed in sections in the following manner:

LEFT TO RIGHT, TOP TO BOTTOM, WITH SMALL OVERLAPS

The following map or chart has been refilmed in its entirety at the end of this dissertation (not available on microfiche). A xerographic reproduction has been provided for paper copies and is inserted into the inside of the back cover.

Black and white photographic prints (17" x 23") are available for an additional charge.

UMI
EXPLANATORY NOTE

The most conspicuous features shown by these set of strike sections are:

1. Thickness change of the Infra-Nappe unit.


3. Oblique orientation of section reveals lateral ramps.

4. Thrusts and imbrications within the Nappe.

5. The main depocenters of the Basin.
MAIN STRUCTURAL FEATURES

7. Extensional Lateral Ramp
6. Listric Normal fault
5. Low angle Extensional detachment
4. Thrusts within the Nappe
3. Prerifaine Nappe basal detachment
2. Infra-Nappe Imbricates
1. Basement faults
ANEXE TO PANEL 2

R6
EXPLANATORY NOTE

1. Rotated blocks
2. Normal fault
3. Top synrift unconformity
4. Guadalquivir allochthonous basal detachment
5. Basal Fordeep unconformity
6. Thrusts within the Nappe
7. Toe-thrust

**LEGEND**
- Tops:
  - Upper Miocene
  - Middle Miocene
  - Upper Cretaceous
  - Lower Cretaceous
  - Jurassic
- Infra-Nappe
- Guadalquivir Allochthonous
- Supra-Nappe
- Post-Rift I
- Post-Rift II
- Syn-Rift
Panel 2: Rharb Basin

Regional Transects

Strike Sections

Gibraltar Arc

Atlantic Ocean

Location Map

Rharb Basin

Legend

Explanatory Note

The most conspicuous features shown by these sets of strike sections are:
1. Thickness change of the intra-nappe unit.
2. Nearly vertical faults affecting the intra-nappe succession.
3. Oblique orientation of section reveals lateral ramps.
4. Thrusts and imbrications within the nappes.
5. The main depocenters of the basin.
PANEL 3  REGIONAL TRANSECTS
OFFSHORE NORTH-WESTERN MOROCCO  ASILAH - RABAT

LEGEND
SUPRA-NAPPE
Basal Foreland Unconformity
Peak-Rift Unconformity
Philippian Unconformity
Nappe
Prerifaine
Intra-Nappe
Intrusion
Basement

SW
1. Hercynian Unconformity
2. Thrust within
3. Thrust within
4. Thrust within
5. Thrust within
6. Thrust within

50 Km

OFFSHORE NORTH WESTERN MOROCCO

SW
1. Hercynian Unconformity
2. Thrust within
3. Thrust within
4. Thrust within
5. Thrust within
6. Thrust within
RHARB BASIN

MAP 2

STRUCTURAL CONTOUR MAP
TOP OF THE PRERIFAINED NAPPE
CONTOUR INTERVAL 0.2 SEC
Modified from ONAREP (1991)

LEGEND

THRUST  
RIDGE  
NORMAL FAULT  
SYNCLINE  
ANTICLINE  
SLUMP SCARS  

J. F. FLINCH 1993

FLINCH, JOAN F.
9514173 c 1995