BOOK IN STOCK

1. Introduction

The workshop was held on the island of Djerba, Tunisia, from 22 to 25 November 2000. A total of 15 scientists (see list at end of volume) originating from nine countries attended the seminar at the invitation of CIESM. In opening the meeting, Dr Frederic Briand, Director General of CIESM, and Dr Jean Mascle, Workshop chairman and President of CIESM Marine Geoscience Committee, expressed their great pleasure to see eminent researchers from both shores gathered here – and in many cases meeting each other for the first time – to tackle in an original and intensive fashion a complex and challenging issue : the African continental margins of the Mediterranean Sea.

Although practical considerations did prevent, in the end, the attendance of a few invited specialists from further countries, the organizers felt gratified to see scientists from Algeria, Morocco and Tunisia, representing both academic research and national petroleum organizations, able to exchange their knowledge, hypotheses, and perspectives with specialists from Europe (France, United Kingdom, Germany, Belgium and Russia), and among themselves as well. As pointed out by several participants, the workshop permitted an intensive scientific dialogue, not only across the north-south divide of the Mediterranean Sea, but also, and for the first time, among top geoscientists from north Africa.

1.1. Background and objectives

The north African and Levant continental margins are quite contrasted, representing either segments of an old Mesozoic passive margin, or more recent and tectonically active continental borderlands. In spite of their scale and importance, they remain surprisingly poorly known, at least from an academic point of view, and poorly understood. These continental margins have never been studied by academic drilling and no real synthesis exists concerning their morphology, history, structure and sedimentation. Marine surveys remain sparse, and information from hydrocarbon exploration remains selective (*e.g.*, Nile delta).

In brief, from the Levantine region to central Tunisia, the continental margin is considered to be an old passive-type margin that has resulted from a long Mesozoic rifting history. It was later subjected to important sedimentary accumulations (such as the Nile deep sea fan construction) and might be locally starting to collide with the southern border of Europe (particularly south of Cyprus and south of Crete). From western Tunisia to Gibraltar, the north African margins result from complex interactions between the much more recent opening of the Western Mediterranean and Alboran basins, and from active tectonic evolution related to the surrounding Alpine collisional setting which has created the north African mountain ranges. As a consequence this sector can be considered as a tectonically active area, still subjected to damaging earthquakes and shaped by fairly active sedimentary processes.

Currently many questions are a matter of debate. For instance what is the timing and process of continental break-up and spreading of the Tethys ocean ? Is the north African margin of volcanic- or non-volcanic-rifted type, or both ? When did ocean spreading begin? Is the Levant margin segment of transform, or orthogonally rifted type ? To what extents are parts of the margin now affected by collisional tectonics ?

BOOK IN STOCK

During the course of their four-day meeting, the participants extensively covered such ground, reviewing current knowledge of the structural and geodynamic evolution of the region. Based on the various discussions and presentations, Dr Alastair Robertson summarizes below the geodynamic evolution of the north African and Levant margin segments, integrating the many threads provided by the participants.

In addition, two workshop sessions were set apart to evaluate the best ways (and the best tools) to further our understanding of the structural and sedimentary processes at work in this zone, and to identify promising paths for future research projects at regional and/or thematic scales. These are presented in sections 3 and 4 below.

2. Outline of geodynamic evolution (summary by A. Robertson)

2.1. Pre-Late Permian

During the Paleozoic the present north African/Levant passive margin formed part of Gondwana bordering the Paleotethys ocean to the north (see references in Robertson, this volume). A major question is whether the south-Gondwana margin was also affected by the Hercynian orogeny (*e.g.* Sengör *et al*., l984). Recent work has tentatively correlated the Betics (S Spain) with Gondwanian crust, which would favour such an hypothesis (Michard *et al*., l997; Chalouan and Michard, this volume).

2.2. Late Permian-Mesozoic rifting

Hercynian deformation was followed by extension during Late Permian-early Mesozoic time (Guiraud and Bosworth, l999). For example, in Tunisia, a deformed Paleozoic succession is unconformably overlain by Late Permian rift-related sediments (Dridi, this volume). Rifting was there active until Early Cretaceous time and facies belts are oriented E-W, while sub-surface successions thicken northwards (see references in Bedir, this volume). The Tunisian margin is dissected by N-S faults which can be interpreted as reactivated Pan-African basement structures, and the southward continuation of the Malta Escarpment is seen as one of the major N-S transverse structures.

Local Late Triassic fault blocks in Tunisia, associated with rift magmatism, are sealed by Lower Jurassic sediments (Bedir, in press; Bedir, this volume), and the Late Jurassic was marked by a regression and clastic sedimentation that could correspond to flexural uplift of the margin, coeval with opening of an oceanic connection between the Central Atlantic and the Western Mediterranean. In general, individual fault blocks in Tunisia exhibit contrasting movement histories controlled by regional extension, wrench movements and salt tectonics (Bedir, in press). A comparable rift history is seen Morocco, Libya and Egypt (Guiraud and Bosworth, l999).

2.3. Two-phase continental break-up

It is generally agreed that the north African/Levant passive margin owes its origins to two fundamentally different spreading episodes (Sengör *et al*., l984; Robertson and Dixon, l984; Dercourt *et al*., l986, l992; Ziegler, l990; Ziegler *et al*., in press). The first is the opening of the eastern Neotethys, westwards from Oman, along the Arabian margin into the Eastern Mediterranean. Evidence from the Levant margin (Ben-Avraham, this volume), from Cyprus, and from southern Turkey (*e.g.* Antalya) is consistent with initial spreading in Late Triassic-Early Jurassic time, but this early oceanic basin probably remained narrow (<500 km) (Robertson, l998a).

The second spreading event relates to the opening of the North Atlantic in the Late Jurassic. This is well documented by the existence of ophiolites (*e.g.* Liguria, E Alps) and deep-sea sediments in the Atlantic and Western Mediterranean region (Bernoulli and Jenkyns, l974; Ogg *et al*., l980). The oceanic crust extended thus through the Betics/Rif area of S Spain/N African into the western Neotethyan ocean, and this, in turn, extended through the Eastern Alps into the Pannonian Basin.

An important question is the nature of this oceanic connection. The linkage is marked by fragmentary ophiolitic rocks of the Betic and Rif regions (Puga *et al*., l989, 1999). Recent work indicates that the ophiolite-related volcanics are of MORB type, supporting an oceanic connection,

BOOK IN STOCK

rather than a rift basin. In recent models the southern Betics are seen as a rifted microcontinent that was separated from the Iberian plate to the north, both by a westward extension of the late Mesozoic-early Tertiary “Pennine” ocean in the north, and by another oceanic strand to the south.

How the eastern (Triassic-Early Jurassic) and western (Late Jurassic-Early Cretaceous) spreading systems interacted remains controversial. In one view the two stopped short of each other, leaving an Apulian promontory as a small N-S barrier. Alternatively, the two ocean stands converged (or overlapped), opening an E-W oceanic basin along the entire length of the north African margin by Late Jurassic time. It is likely that by Early Cretaceous time the north Africa/ Levant margin existed as a continuous rifted continental margin, separated from a number of rifted continental fragments by Neotethyan oceanic crust. It is not surprising, therefore, that the Late Permian/Early Mesozoic rift history of north Africa was multi-pulsed and locally variable, as this region was affected by continental break-up of both the eastern and western Neotethys.

2.4. Passive margin history

After formation of a single north-facing passive margin, by Late Jurassic time the north African/ Levant margin underwent passive margin subsidence during Cretaceous-Early Tertiary time. The history of the Mediterranean deep southern margin and adjacent basin during this time remain poorly known. As an exception, the structure and stratigraphy of the passive margin is relatively well documented in the Levant, based on seismic and well data (Garfunkel, l998). Locally, the Early Cretaceous and younger history of a rifted continental fragment, the Eratosthenes Seamount, in the Eastern Mediterranean was recorded by drilling during ODP Leg 160 (Robertson, l998b; Mart and Robertson, l998). Additional information on the mid-Cretaceous (Aptian-Albian) and younger sedimentary history is provided by sedimentary rock clasts recovered from mud volcanoes in a number of areas of the Mediterranean, notably south of Crete (see references in Akhmanov *et al*., this volume).

During its “passive margin phase” the north African/Arabian margin was also influenced by far-field tectonic events, including pulses of compression, strike-slip and extension in specific areas (Guiraud and Bosworth, 1999). Most notably, in the Arabia-Egypt region the margin was affected by stratigraphic inversion along the Syrian arc (Garfunkel, l998), probably related to ophiolite emplacement along the northern margin of Arabia (Robertson, l998).

2.5. Neogene collisional deformation in the west

During the Tertiary, especially in Eocene time, many parts of the north African margin experienced deformation and thrusting related to the convergence of the African and Eurasian plates (Dewey *et al*., 1989; Robertson and Grasso, 1995).

Along the southern Mediterranean margins, collisional deformation was manifested first in the west (west of the Malta Escarpment), as documented by Early Miocene thrusting onto north Africa. As a result, former passive margins units (platform/slope) were detached and thrust southwards as a stack of thin-skinned thrust sheets. Thrusting was accompanied by widespread genesis of Burdigalian olistostromes and related foreland basin flysch sediments, as in Sicily, the Maghrebides and Kabilides) (Saad and Caby, l994; Braik, this volume). Recent marine studies of the Sardinia channel area now allow detailed correlations between Tunisia, Sardinia and Sicily (Bouillin, this volume; Torelli *et al*., this volume).

The favoured explanation of the Miocene collisional deformation is that oceanic crust of the western Neotethys was subducted generally northwards beneath the Iberian margin. With time, the subducting slab migrated southwards (i.e. “rolled back”) until it collided with the north African continental margin. This margin was, however, irregular in shape, such that collision in Sicily and Tunisia and Algeria occurred in Early Miocene time, whereas the eastern Neotethys further east remained largely unaffected. Collision with the northwest African margin was thus diachronous, becoming younger westwards towards the Rift/Betics region (Chalouan and Michard, this volume).

2.6. Orogenic collapse

The Neogene compressional deformation was accompanied by pervasive extension of hinterland areas, dating from latest Oligocene/Early Miocene (Aquitanian-Burdigalian) time. Rifting took

BOOK IN STOCK

place extensively, as documented in the Alboran Sea (Watts *et al*., l993). Extension also gave rise to core complexes in some areas (*e.g.* Grande Kabylie in Algeria; Saad and Caby, l996). In addition, Neogene extensional basins are well developed in some onshore areas of north Africa (*e.g.* in Algeria) and include records of Miocene extensional volcanism (Kheidri *et al*,. this volume).

Rifting proceeded to oceanic spreading in some hinterland areas, including the North Balearic (Provencal) Basin, the north Algerian Basin and the Tyrrhenian Sea. Extension was coupled with the detachment of Corsica and Sardinia from southern Europe (see papers in Robertson and Comas, l998).

Rifting of the present-day north African continental margin was accompanied by left-lateral strike-slip, as documented by onshore and offshore studies (Comas *et al*., 1992; Braïk, this volume). Some margin areas remained tectonically active into Pleistocene-Holocene time (Kheidri *et al*, this volume; El Moumni *et al*., this volume). In Tunisia, for example, Plio-Quaternary tectonics included fault inversion, mud diapirism and rapidly subsiding basins in a strike slip dominated setting (Bedir, this volume).

One apparent problem is that the northern (European) Neotethyan northern margin collided with north Africa during the Early Miocene, yet further NE the Tyrrhenian Sea did not open until Late Miocene onwards (Kastens *et al*., l990). To accommodate this, Gueguen *et al*. (l998) postulate a major NW-SE trending dextral strike-slip zone between the Algerian and Provencal basins, which thereby allowed the Ionian Sea oceanic plate to continue to “roll-back” southwards, accommodating further back-arc extension in the Tyrrhenian Sea.

2.7. Incipient collision in the Eastern Mediterranean

East of the Malta Escarpment, the Cyrenaica region of Libya is interpreted as a large promontory of the north African continental margin during its rift/passive margin development; this promontory would be the first area to collide with the Mediterranean Ridge accretionary wedge as it migrated southwards in Neogene-Recent time (Camerlenghi *et al*., l995).The timing of initial collision remains controversial, but was arguably since Late Miocene (Chaumillon and Mascle, l995; Mascle *et al*, this volume), but more probably since Late Pliocene time. An intriguing aspect is that processing and interpretations of seismic refraction data suggest that an undeformed continent-ocean transition zone may still exist between Cyrenaica and Crete (Brönner and Makris, this volume). By contrast, interpretations based on shallow seismic reflection data favour a collision-related setting, implying that the pre-existing margin should be partly disrupted in this region (Mascle *et al*, this volume).

Further east again the Egypt and Arabian margin segments remain in a “passive margin” state. However, these areas were affected by Eocene-Oligocene collision of the regional Arabian promontory with Eurasia, as documented by the thrust belts in southern Turkey and Iran (Fourcade *et al*., l991; Yilmaz, l993; Robertson, l998a) and this, in turn reactivated the Syrian Arc in the Levant and northern Egypt. More recently, during the Plio-Quaternary, the Eratosthenes Seamount, interpreted as a rifted north African/Arabian continental fragment (Woodside, l977; Ben-Avraham *et al*., l986; Kempler and Ben-Avraham, l987; Robertson *et al*., l995; Mart and Robertson, l998) began to collide with the Cyprus active margin to the north. However, the Nile cone area to the south has so far remained largely unaffected by these collisional effects (Bellaiche *et al*., l999). Instead, salt tectonics has played a critical role in subsurface structuration and sediment dispersal (Gaullier *et al*. 2000; Loncke, this volume). In addition, the Late Oligocene and subsequent rifting of the Red Sea/Gulf of Suez/Gulf of Aqaba systems might have also affected the eastern Nile Cone area (Mascle *et al*., 2000).

2.8. Summary

In summary, the north African-Arabian margin was affected by the evolution of the Tethys ocean to the north from early Paleozoic times onwards. The eastern Mediterranean basin experienced Late Permian rifting, followed by early Mesozoic oceanic spreading and has survived relatively unscathed until present time. On the other hand, the Western Mediterranean basin also experienced Late Permian-early Mesozoic rifting, but Tethyan oceanic spreading was delayed until Late Jurassic time. Collisions in orogenic areas to the north reflect the amalgamation of continental

BOOK IN STOCK

fragments, mainly in Late Cretaceous-Paleogene time. Owing to the narrowness of the western Neotethys, relative to further east, and the Africa-Eurasia convergence history, this western region adjacent to north Africa experienced collisional deformation first (Early Miocene), creating the Maghrebides-Kabilides-Rif chain and counterparts in the Betics. By contrast, the north African margin in the Eastern Mediterranean has experienced only incipient collision in some areas. Back-arc marginal basins in the Eastern Mediterranean basin were restricted to rifts (*e.g.* N Aegean), but widened into small ocean basins in the Western Mediterranean (*e.g.* Tyrrhenian Sea).

The probable ultimate fate of the Mediterranean region is a thrust belt similar to the Zagros Mountains of Iran.

3. Research needed on structural framework and geodynamic evolution (session summarized by Alastair Robertson)

With regard to fundamental scientific objectives, available techniques and potential achievements, the group identified two main topics: 1) Neogene-Recent tectonic evolution of the Western Mediterranean basins and margins; 2) Rifting/passive margin history of the Mesozoic Tethys ocean.

Potential study on the Neogene-Recent tectonic evolution of the Western Mediterranean basin (from Tunisia to Morocco) requires both new data and synthesis of existing data. The main techniques to use should be swath bathymetry, coring and other marine studies of the offshore areas, coupled with onshore studies, mainly of neotectonic processes (*e.g.* field mapping, kinematic analysis of faults, tectonic geomorphology). The end product would be an improved understanding of the geodynamic evolution of Neogene deep-sea basins and margins in the Western Mediterranean region.

Research on the Rifting/passive margin history of the Mesozoic Tethys ocean (Libya, Egypt, Levant, Syria) should be mainly based on existing, largely industrial data, that could be used for collaborative studies with academia, supplemented, where possible, by acquisition of new data. Aspects could include improved interpretation of deep geophysical data and land-based studies. The end-product would be a better understanding of the Mesozoic history of rifting and sea-floor spreading in the central and eastern Mediterranean regions.

Both topics appear to have important social relevance. For example topic 1 (Western Mediterranean basins/margins evolution) relates to seismic hazard (earthquakes and tsunamis), and topic 2 (Mesozoic rifting/spreading) clearly relates to hydrocarbons, especially exploration of deep-water and ultra deep water continental margins.

3.1. Fundamental scientific objectives

The following topics were assessed for possible future research.

3.1.1. Mesozoic rifting of Tethys

This concerns the opening of the Mesozoic Tethys ocean (Neotethys), as documented in north Africa/Arabia. Aspects that require future investigation include :

- timing of rifting and spreading;
- location and nature of the continent-ocean transition zones;
- role of extension, or strike-slip in basin evolution;
- role of magmatism in continental break-up;
- sedimentary facies and depositional processes related to rifting.

3.1.2. Tethyan passive margin history

Aspects that require future investigation include :

- onshore passive margin subsidence history;
- tectonically emplaced passive margin units (i.e. N Syria and parts of the NW African flysch nappes);
- deep-sea drilling (ODP);
- sampling of clasts from mud volcanoes.

BOOK IN STOCK

3.1.3. Collision and emplacement history

Future investigations should include :

- late Cretaceous emplacement of margin/oceanic units onto the Arabian margin (N Syria);
- oligocene-Miocene emplacement of margin/oceanic units onto the NW African margin (Tunisia, Algeria, Morocco);
- Miocene?-Recent collision of the Mediterranean accretionary wedge with the Cyrenaica Peninsula (Libya).

3.1.4. Neotectonic rift /spreading processes

This concerns the opening of the Neogene-Recent deep water/oceanic basins of the Western Mediterranean Sea. Expected investigations include :

- timing of rifting and spreading;
- location and nature of the continent-ocean boundary in different basins;
- sedimentary facies and depositional processes related to rifting;
- role of extension, or strike-slip in basin evolution.

3.1.5. North Mediterranean margins

It was stressed that the conjugate margins of the rifted Mesozoic Tethyan margins (*e.g.* in Greece, Cyprus, Turkey) and also of the Western Mediterranean Neogene basins (S Spain, S France, Italy) must be taken into account in any synthesis.

3.2. Research techniques

Participants discussed the techniques that could be applied to the studies listed in the above section, as follows :

3.2.1. Standard marine geophysical surveys

This includes classical marine survey techniques, including magnetic and gravity surveys. Most of the necessary data already exist, but may merit re-interpretation *e.g.* to examine the possible role of magmatism in rifting.

3.2.2. Deep refraction seismic investigations

Data now exist (*e.g.* Libya-Crete), but could be re-analysed in terms of E-W, in addition to N-S trends. There is also a need for refraction surveys of the N Levant margin (Syria) and some other areas.

3.2.3. High-resolution swath bathymetry

This is an efficient and very effective technique that recently has shed much light on the topography and recent deformational sedimentary processes of the Mediterranean basins (*e.g.* Mediterranean Ridge). A future aim is to complete the coverage of the Mediterranean Sea. Surveys should include inshore areas as far as possible. For the Western Mediterranean there is a need to collate data from surveys by different countries to achieve an overview.

3.2.4. Deep-sea drilling (ODP)

This is an effective, but very expensive, approach that yields invaluable evidence of the subsurface stratigraphy of the Mediterranean basins (*e.g.* recent drilling in the Alboran Sea and on the Eratosthenes Seamount). However, no future drilling is likely in the Mediterranean within the foreseeable future (next 3-5 years) despite good and relatively mature proposals (*e.g.* Rhone fan and Mediterranean Ridge.

3.2.5. Sampling of clasts from mud volcanoes

This is a relatively cheap and effective way of shedding light on the facies and age of rock types beneath the Mediterranean seafloor. It is, however, limited by the localised development of mud volcanoes and does not allow the surface stratigraphy to be reliably reconstructed.

3.2.6. Onshore fieldwork

This approach is of limited value in determining the Neotethyan rift/passive margin history, as coastal areas are widely buried by younger sediments. Exposures further inland (*e.g.* Sirte Basin, Libya) commonly reflect north African intraplate deformation as a whole and not simply Mediterranean basinal tectonics. However, onshore fieldwork is invaluable where Tethyan mar-

BOOK IN STOCK

gin/basin units have been tectonically emplaced. Principally this is seen in north Syria (Baer-Bassit), where Triassic-Late Cretaceous margin units were tectonically emplaced in latest Cretaceous time. Tethyan units in NW Africa were emplaced in Oligo-Miocene time (*e.g.* Maghrebides, Cabilides, Rif) as basal levels of "flysch nappes". However, these do not document the pre-Cretaceous history. In addition, onshore fieldwork can shed much light on the Miocene-Recent history of the Western Mediterranean Neogene basins, as coeval sedimentary succession, are exposed in Tunisia, Algeria and Morocco.

3.2.7. Industry well and seismic data analysis

Extensive industry seismic reflection and well data sets exist for the Mesozoic north African basins, including the Suez Rift and Sirte Basin. However, the deep-water continental margin areas have, to date, been largely ignored, with the exception of the Nile Cone. For some areas (*e.g.* Sirte Basin) industry data sets are already partly in the public domain. Opportunities may exist for industry/academic collaboration to utilise existing seismic and well data sets. The north African deep-water margins are one of the world's major frontier areas that are likely to be explored further in coming years.

3.3. Regional aspects

Bearing in mind the scientific objectives and the potentially available techniques, the following specific research targets were identified for individual continental margin segments. Working clockwise around the southern Mediterranean these are as follows :

3.3.1. North Levant margin

This is the least well known segment of north African/Arabian Tethyan margin, requiring all types of data to progress with regional understanding.

3.3.2. South Levant Margin

This is currently the best documented area, but much industry data are still not in the public domain.

3.3.3. Nile Cone

Areas in the south (Gulf of Suez rift) are well documented, but the data are mainly not in the public domain. Recent swath bathymetry has shed much light on the northward prolongation of the Gulf of Suez structures beneath the Nile Cone (distributed extension zone, or microplate boundary). The deep structure and stratigraphy of the deep-water basins beneath the inner and outer Nile Cone are under industry study, including reflection seismics and planned drilling. However, little of these data are likely to be available for academic study in the near future. The surface morphology of the Nile Cone is revealed most effectively by high-resolution swath bathymetry and a more complete coverage is expected to be obtained during future cruises. In addition, ultra-long piston cores will be acquired from the Nile Cone within several years.

3.3.4. Cyrenaica Peninsula

A recent deep-seismic refraction survey from offshore Libya to Crete has revealed the deep structure of the central north African margin. This apparently includes a transition from subsided continental crust to oceanic crust. Two along-strike survey lines reveal apparently different continental margin structure and further processing of data is needed to determine if this is a real effect (*e.g.* caused by segmentation of the continental margin), or an artefact of processing. Shallow seismic reflection and swath bathymetric surveys extend from the Mediterranean ridge as far south as the Cyrenaica Promontory. Interpretation suggests that the accreted sediments of the Mediterranean Ridge are in the process of being thrust southwards over the Cyrenaica margin segment, which is interpreted as a promontory of the north African margin. However, the deep structure of this inferred collision zone remains largely unknown and the timing of any initial collision is unclear.

The inference from seismic refraction data of a still extant continent-ocean transition zone located basinwards of the Cyreniaca Promontory appears to conflict with the interpretation of this region as a continental collision zone, as hypothesised from shallow seismic data. Processing of seismic reflection and refraction data to evaluate E-W, in addition to N-S structural trends may help resolve the problem. Onshore fieldwork may contribute, as the coastal Cyrenaica hills (sev-

BOOK IN STOCK

eral hundred metres high) may reflect collisional processes. Their timing of uplift might be determinable, *e.g.* by fission track dating. Also, analysis of onshore industry seismic data may reveal tilting or other structuration referable to continental collision.

3.3.5. Tunisian margin

This is a critical segment of the north African margin as it illustrates the combined effects of 1) E-W extension; 2) N-S extension/strike slip; 3) mid-Tertiary collision in the north (Maghrebides); 4) Plio-Quaternary localised extension/strike slip. To the east, the Tunisia margin segment is bounded by the southward extension of the Malta Escarpment, a N-S lineament that was probable active in segmenting the north Africa margin during Mesozoic rifting of Neotethys. Additional evidence of N-S tectonics is revealed by the N-S Axis, further west. This extends northwards from the Saharan basement, through surface outcrops, then beneath basinal units, as revealed by industry seismic reflection data. The Tunisian margin segment underwent overall oblique extension during Late Permian to Jurassic rifting. Basement structures were reactivated to create a mosaic of rift blocks, with variable extensional and trans-tensional effects between. Salt tectonics have played an important role during rifting and subsequently. Recent studies link the tectonics of Tunisia across the Sardinia Channel to Sardinia. Correlations have also been established with Sicily.

Although quite well known from combined industry and academic studies, several aspects of the Tunisian margin segment are worthy of further study, as follows :

- role of Pan-African basement fault lineaments in Neotethyan rifting;
- relationship of the N-S Axis to rift processes and later Mesozoic-Tertiary history;
- nature and geometry of rift-related blocks;
- effect of Oligo-Miocene collision on margin structure;
- nature of Plio-Quaternary tectonics including exceptionally rapid local basin subsidence;
- role of salt tectonics.

3.3.6. Algerian margin

Data exist from industry, but only few are available from academia for onshore areas. However, there is a need for some updating and improved correlation with adjoining regions (Tunisia and Morocco). Some relevant aspects are as follows :

- the Maghrebides/Kabilides formed by Oligo-Miocene collision; this requires additional study to link sub-surface structure with outcrop geology in detail;
- Neogene history of post-collisional extensional tectonics and related (dextral?) strike slip in the north (coastal areas); fission track studies could elucidate the timing of exhumation (as determined for Sicily and elsewhere);
- study of onshore Plio-Quaternary sedimentary basins could shed more light on the rift/ extensional history;
- neotectonic studies (i.e. of fault planes) are needed to determine kinematic history and the role of strike-slip *versus* extension;
- the structure of the continent-ocean transition zone in deep water off Algeria remains poorly understood.

3.3.7. Moroccan margin

Much data from industry and academia for onshore areas have already been published, and synthesis is well advanced. Few major problems appear to remain. A more extensive collisional orogenic history is present in the Betics of southern Spain and detailed comparisons between the two areas will be useful. Aspects for potential studies include:

- comparison of the Hercynian orogenesis of the Atlas, the Moroccan margin area and the Betics;
- comparison of the processes and timing of Neotethyan rifting within adjacent areas (*e.g.* Tunisia, Libya, N Egypt);.
- comparison of the timing and processes of Neogene collision, particularly to test models suggesting diachroneity of collision from east to west;
- more detailed geochemical studies of magmatic units, including the Rif peridotites and extrusive igneous rocks.

BOOK IN STOCK

- neotectonic and sedimentary studies of the northern coastal areas to compare with marine studies of the rifting/subsidence history of the Alboran Sea;
- detailed comparative studies of the Betics to decide whether the Rif nappes record a Tethyan rift, narrow (transform dominated ?) oceanic connection between Tethys and the Atlantic, or a wider ocean basin;
- finally, the alternative models for the origin of the Alboran Sea (i.e. delamination of the dense lower crustal root or slab "roll-back" need to be tested by all possible means (*e.g.* seismic tomography).

4. SEDIMENTARY PROCESSES ON THE NORTH AFRICAN MARGIN (session summarized by N. Kenyon)

4.1. Mapping

The mapping of bathymetry and reflectivity (backscattering) using the latest swath mapping techniques is proving to be of considerable value as a basis for many marine scientific and commercial activities. This will be increasingly so over the next decades with the growth, for instance, of deep sea hydrocarbon exploitation, the submarine cable industry and deep sea fishing. Such data exist for most of the deeper waters of the northern Mediterranean and for the eastern north African margins. Mapping should be extended to the remainder of the north African margins and will provide part of the infrastructure for the development of the submarine margins of north Africa. The shelf and coastal zones should clearly be included: although their mapping will be more costly in ship time, they are of great importance to the health and wealth of the coastal population.

Such mapping should include acquisition of high quality, high resolution seismic profiles. Added value from such mapping would be to place the many existing bottom samples in context and to better evaluate the recent tectonic activity. Data banks currently being developed in the EU countries for cores, seismic profiles and sidescan sonar, etc., necessarily include all of the Mediterranean Sea. They should be supported by both the provision of data and by the use of such data for research and commercial purposes.

The meeting participants considered of particular value to develop studies on slope instability and sedimentary processes, as well as on mud volcanoes which are relatively common features in the deep basins of the Mediterranean sea

4.2. Slope instability and sedimentary processes

The failure of sediments is particularly significant along the north African margins and was noted early, *e.g.* in the study of the turbidity current following the Orleansville-El Asnam earthquake of 1954 (Heezen and Ewing, 1955).

The types of slope failure need to be identified and their ages determined. The north African margin appears as a particularly good place for studies aimed at understanding failure triggered by tectonic or rapid sedimentation. The effects of tsunamis, caused by some combination of earthquake displacement of the sea bed, major sediment failure or volcanic eruption, need to be checked by studies of characteristic tsunami deposits on low lying coasts of north Africa.

The very thick and extensive turbidites found in Mediterranean abyssal plains are also believed to be associated with major tsunamis, such as the one inferred to be associated with the eruption of Thera, or with sediment failures, such as a slide identified on the Libyan margin. The origins and age of these deposits need to be established and this approach, together with modelling studies, should lead to improved hazard prediction.

Extreme end members of the types of turbidite system are present on the north African margin. The Nile Cone is an example of a very large system, fed by a major river input and having many sinuous channel-levee systems that lead down to the abyssal plain. The detailed history of channel migration and switching has yet to be determined for any such system. For this, very detailed high resolution seismic profiles and dated samples on a portion of a single sinuous channel are required.

BOOK IN STOCK

Steep, tributary canyon fed systems are common along most of the north African margin, west of the Nile cone, particularly off Algeria. Such systems are poorly studied. The frequency of flows needs to be determined at times of both high and low sea level, as does the role played by debris flow processes as compared to turbidity current processes. The fate of fine grained materials, both sediment, organic matter and pollutants that are fed into the sea, mainly during peak river flow should be studied. Canyons are expected to play a major role in this on the north African margin from western Tunisia to Morocco, where many of the shelves are particularly narrow.

Strong near surface currents are known on the north African margin, for instance the gyres in the Alboran Sea and the eastward flowing Algerian current. Erosion and deposition from such currents will be significant but are as yet unstudied. The discovery of long term currents, both at great depths and on the shelf, can be readily discovered by geophysical and sedimentological techniques, to the mutual benefit of interdisciplinary studies. Once the main current patterns are established, their high frequency variability can be investigated by measuring the effect of, for instance, benthic storms.

4.3. Mud volcanism

A major addition to the knowledge of deep stratigraphy in the deep Mediterranean marine basins, at a relatively low cost compared with drilling, can derive from the analysis of rock clasts brought to the surface by deep seated mud volcanoes.

Further work is required, particularly along the western arm of the Mediterranean Ridge and in the Alboran Sea. Improved provenance for clasts should come from detailed comparisons with rocks recovered from boreholes and outcrops.

Data bases for Mediterranean-wide rocks should contain improved descriptions, without which provenance studies are less reliable. A study of the veining in clasts will lead to a better understanding of fluid flow mechanisms as should more measurements of heat flow. The study of specific life colonies, associated with such environments, is in its infancy. Careful biological sampling in relation to more detailed geological mapping is strongly recommended

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Evolution of the Mediterranean continental margin of northern Morocco

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INTRODUCTION

The present-day north Morocco continental margin bounds to the south the westernmost Mediterranean basin, or Alboran basin, and the Gibraltar Strait. However, this margin corresponds to a recent, mostly Neogene feature, post-dating the Alpine orogeny of the Betic-Maghrebide fold belts. In northern Morocco, the orogeny formed the Rif belt through the collision of an exotic terrane, defined hereafter as the Alboran Terrane, with the previous, Mesozoic continental margin of north Africa. In the following, we summarize the Rif and Alboran basin structure and geological records, then present a new scenario for their tectonic evolution.

STRUCTURAL OUTLINE

The Rif belt includes three major geological domains or nappe complexes, trending from ENE in the Eastern Rif to NNE at the tip of the Western Rif. The External Rif consists of a fold-and-thrust belt detached along the Upper Triassic evaporitic beds from the attenuated crust of the north African margin. The belt is usually divided into three parautochthonous-allochthonous zones, from SW to NE the Prerif, weakly displaced with respect to the Atlasian-Mesetian foreland; the Mesorif, more displaced southwestward, and the Intrarif, thrust over the Mesorif windows (Suter, 1980a,b; Wildi, 1983; Ben Yaich, 1991). In the eastern Intrarif, Upper Jurassic-Lower Cretaceous sediments are transgressive onto serpentinized lherzolites (Beni Malek massif), likely originating from a serpentinite ridge at the tip of the African margin during the Neo-Tethyan opening (Michard *et al.*, 1992).

The Maghrebian Flysch nappes overlie the External Rif, with the exception of some inliers on top of the Internal Zones (Jebel Zemzem, Riffiene). It is now widely accepted that these unrooted nappes originate from an ocean- or transitional crust-floored Maghrebian Trough extending during the Mesozoic-Paleogene in between the north African margin and the Internal Zones domain (Suter, 1980a,b; Wildi, 1983; Durand-Delga, 2000).

The Internal Rif, i.e. the Moroccan part of the Alboran Terrane, includes three nappe complexes, from top to bottom, the Ghomaride and Sebtide nappes (equivalent to the Malaguide and Alpujarride nappes of the Betic Cordilleras, respectively), and the Dorsale calcaire in front of them. The Dorsale calcaire consists of stacked carbonate slabs, made of Upper Triassic-Liassic platform to passive margin deposits with a thin cover of Jurassic-Late Cretaceous pelagic and

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Tertiary detrital sediments (Wildi, 1983; Nold *et al.*, 1981; El Hatimi *et al.*, 1991; El Kadiri *et al.*, 1992). The Ghomaride complex, which overlies either the Sebtide or Dorsale units, includes four nappes of low-grade Paleozoic metasediments covered by thick, unconformable Triassic red beds, and much thinner Liassic and Paleocene-Eocene limestones (Chalouan and Michard, 1990; Maaté, 1996). The Sebtide complex is a stack of metamorphic units with dominant shallow-dipping foliation, deformed by the Beni Bousera and Beni Mezala late-metamorphic antiforms (Kornprobst, 1974; Michard *et al.*, 1983, 1997). The highest Sebtide units (Federico units) mainly show Permian-Triassic phyllites and quartzites, and Triassic carbonates. The underlying Filali unit consists of schists and gneisses, and overlies the Beni Bousera granulite (kinzigite) and peridotite unit (Kornprobst, 1974; Reuber *et al.*, 1982; Saddiqi *et al.*, 1988).

The Alboran Basin shows a rugged subsea topography and is divided in two parts by the northeast-trending Alboran ridge. The Alboran basin basement consists of Ghomaride-Malaguide, and mostly Sebtide-Alpujarride rocks (Comas *et al.*, 1992; Watts *et al.*, 1993; Platt *et al.*, 1998; Sanchez-Gomez *et al.*, 1999) and external rifan units. The thickness of the basin crust is extremely variable, <15 km in the western sub-basin, <10 km in the eastern one, and in several sectors the underlying, abnormally hot mantle might be very near the sediments (Polyack *et al.*, 1996; Seber *et al.*, 1996; Morales *et al.*, 1999). The Alboran Basin has very thick Neogene sediment accumulations, anoxic subbasins (potential sources), evaporites (traps, seals), many stratigraphic traps, and moderate to high heat flow (up to 1,6°C/30m in Andalucia A-1 well). The thickness of the Middle-Miocene-Quaternary sediments explored by our profiles reaches approximately 9 km in the depocenters.

THE GEOLOGICAL RECORD

Sedimentary, metamorphic and volcanic records in the Rif belt

Several tectonic events are recorded by the rifan rocks and by the Alboran deposits. The earliest sedimentary events recording potential orogenic movements in the Rif-Maghrebide domain itself date from the Late Cretaceous, and can be essentially observed in the Dorsale units. In the Rifian Dorsale, the Late Cretaceous hemipelagic deposits overlie disconformably or even unconformably the Tithonian-Berriasian limestones (Wildi, 1983; Ben Yaich, 1981; El Kadiri *et al.*, 1992). In the External Rif, one can only notice a sudden crisis of Triassic rock resedimentation in the Upper Cretaceous marls, likely related to an increase in diapiric deformation of the thick Triassic evaporites.

In the Internal Rif, an Alpine metamorphic event is recorded by the Sebtide nappes. The Alpine P-T conditions in the upper Sebtide Federico nappes were estimated through the study of their Permian-Triassic metapelites (Bouybaouene *et al.*, 1995). In the Beni Mezala antiform, the critical metamorphic assemblages abruptly change from one unit to the other, the peak P-T conditions increasing downward from greenschist to HP-greenschist, to blueschist, and finally eclogite-facies in the lower Beni Mezala unit. In the latter unit, the retrograde P-T path includes a roughly isothermal, early unloading evolution from 15-20 kbar down to 8-6 kbar, while further unloading occurred under decreasing T. After this syn-metamorphic evolution, the final exhumation of the Sebtide antiforms occurred by excision of their Ghomaride overburden through brittle normal faults, as illustrated in the Beni Bousera region by the Zaouia low-angle, and Ras Araben high-angle normal faults (Chalouan *et al.*, 1995a, b). This extensional faulting represent a post-metamorphic stage, which can be correlated with the Alboran Sea rifting itself (Chalouan *et al.*, 1997). Such an evolution strictly compares with that of the equivalent units in western Betics, as described by Balanyá *et al.* (1997) and Azañón *et al.* (1997). The salient feature of our results from the Beni Mezala antiform is the relative convergence of the ages yielded by white micas which are comprised between 20.9 ± 0.8 and 27.4 ± 0.6 Ma. These ages of 23 Ma correspond to the time when the Beni Mezala units crossed the 350°C isotherm, assuming a minimum closure temperature of 350°C for white micas (Purdy and Jäger, 1976). Hence, exhumation of the whole Sebtide units up to upper crustal levels, and their cooling below ca. 350°C occurred at about 21 Ma.

In the External Zones, metamorphic recrystallization is lacking in the western part of the belt, while it may be conspicuous in the eastern part. The Jurassic-Cretaceous rocks of the Ketama and

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northern Temsamane units display externally verging, tight reclined folds with axial-planar foliation and stretching lineation mostly parallel to fold axis (Andrieux *et al.*, 1971; Frizon de Lamotte, 1985; Michard *et al.*, 1992). Coeval greenschist-facies recrystallization was dated in northern Temsamane at 27.8 ± 0.3 Ma by $^{40}Ar/^{39}Ar$ on phengite (Monié *et al.*, 1984).

Pre- to synorogenic volcanic outpours have been identified in the eastern Prerif nappe, where the Sidi Maatoug analcime-bearing basanites are dated from the Paleocene (Hernandez *et al.*, 1976). Undersaturated lamprophyre dykes of similar age (60-57 Ma) are known from the same area (Leblanc, 1979), and more to the east in the Atlasian foreland.

The earliest trace of late-orogenic volcanism is found in western Betics (Malaga), where a basaltic-andesitic dyke swarm with arc-tholeite affinities yielded K/Ar ages of 23-20 Ma (Torres-Roldan *et al.*, 1986), and $^{40}Ar/^{39}Ar$ ages of 30-18 Ma (Turner *et al.*, 1999). The Alboran Island andesites yielded K/Ar ages of 18-7 Ma (Aparicio *et al.*, 1991), but most of the trans-Alboran calc-alkaline volcanism (andesites, shoshonitic trachy-andesites, rhyolites) took place during the Middle-Late Miocene (Hernandez *et al.*, 1987).

Sedimentary and tectonic Alboran basin record

In the Alboran Basin, two main tectonic events are recognized:

- Early-Middle Miocene to Tortonian rifting. The great thickness of the Neogene accumulation in the Alboran Basin results from a dramatic extension initiated during the Aquitanian-Burdigalian (Comas *et al.*, 1992; Watts *et al.*, 1993). Due to the insufficient depth resolution of our profiles, the earliest extensional structures shown here are middle Miocene in age. Hence, we can see a half graben, with divergent middle Miocene reflectors, gently dipping northeastward, similar to the top of the underlying basement, and striking against a southwest-dipping growth fault (Bally, 1981). Some shallow-dipping normal fault, observed in the western Alboran Basin, is likely to extend into the onshore Araben fault zone (Suter, 1980b; Chalouan *et al.*, 1995a). The dip of the Araben fault zone, which cuts across the Beni Bousera peridotites and the overlying nappes, is about 50-60° to the northeast; therefore, the whole normal fault system exhibits a listric geometry. Many of the middle Miocene growth faults remained active during part of the Tortonian.
- Post-Messinian compressional evolution. During the late Miocene, compressional structures developed in the southwestern Alboran Basin. An early shortening pulse occurred before the latest Tortonian-Messinian sedimentation. A further pulse occurred toward the end of the Messinian. The pre- and post-Messinian unconformities spread over similar areas; that is, the west and north borders of the Xaouen Bank (southern tip of the Alboran ridge) and its western approaches. Most post-Messinian structures correspond to the reverse faults with a dominant northward (seaward) vergence. These N110°E- to N80°E-trending structures display an echelon patterns with respect to the N60°E trend of the Alboran ridge. We suggest that they developed in response to sinistral strike-slip movement along the northeastern prolongation of the Jebha fault, a post-Aquitanian, sinistral wrench fault observed onshore (Olivier, 1981).

A third contractional pulse developed toward the end of the Pliocene. Whereas the two earlier pulses were chronologically close (being separated by less than 1 my.), the Pliocene-Quaternary compression occurred after a longer period (approximately 3 my.) of subsidence and relatively quiet sedimentation. The new compression reactivated the previous faulted folds, particularly in the Alboran ridge, and also affected a zone that had previously escaped deformation. This zone, located south of the folded ridge, was converted into a southverging, N90°E to N60°E trending synclinorium, while the ridge itself was pushed upward onto the inverted north Bokkoya fault.

A TECTONIC SCENARIO IN CROSS-SECTION

We present hereafter a tectonic interpretation of the Rif mountain building in four cross-sections (Figure 1a-d). The initial, pre-orogenic setting of the Rif-Betic structural units, as assumed in this paper, is schematically shown in the Figure 2.

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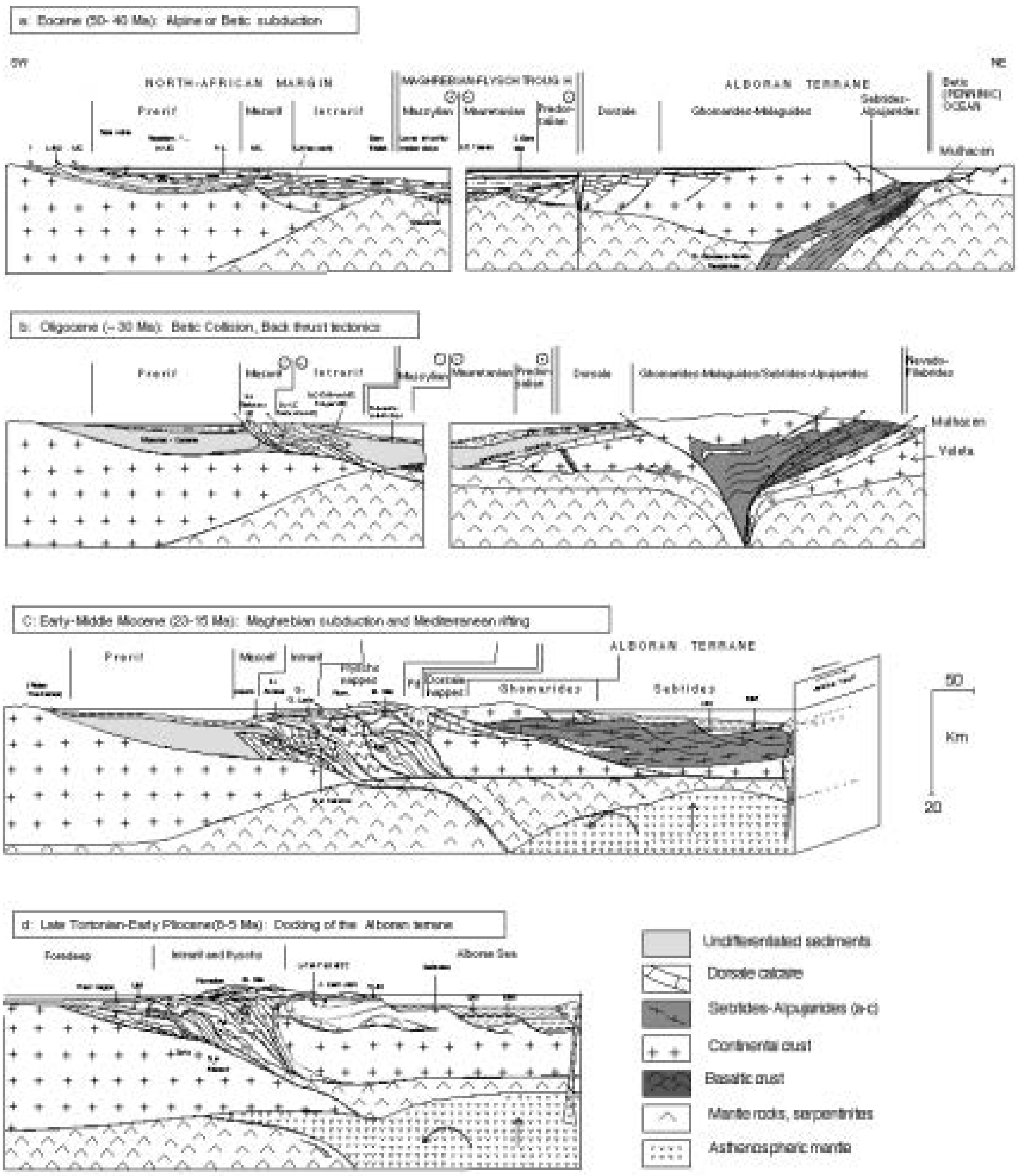

Fig.1. A tectonic scenario for the Rif Mountain building: interpretative, geologic-crustal scale cross-sections corresponding to four critical stages. The northern part of each cross-section except **(d)** is shifted to the east with respect to the southern one. **(a)** The Alpine southeast-dipping subduction is operating along the northern edge of the Alboran Terrane. **(b)** Onset of the Apennine-Maghrebide subduction along the southern edge of the Alboran orogenic arc. **(c)** Rifting Alboran Terrane (operates since the Oligocene-Early Miocene), suturing of the Maghrebian Flysch Trough, incipient docking of the Alboran Terrane, formation of the Proto-Rif belt. **(d)** Collapse of the external accretionary prism (thickened since stage c), continuation of the African margin subduction, asthenosphere uplift, and calc-alkaline volcanism.

Symbols, with **P**: Paleocene; **(M)E**: (Middle)Eocene; **O-M**: Oligo-Miocene; **LM**: Lower Miocene; **S1**: slaty cleavage.

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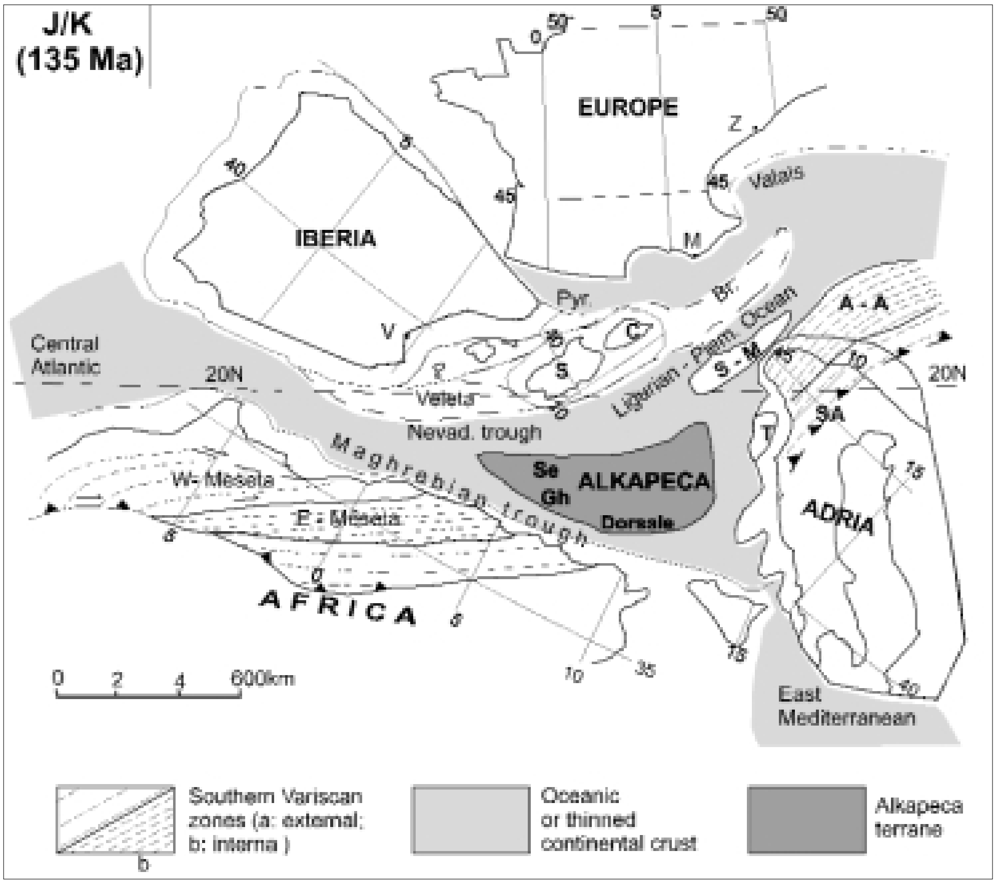

Fig. 2. Origin of the Alboran Terrane and associated blocks in the Western Mediterranean–Alpine area. **A-A**: Austroalpine; **C**: Corsica, **Gh**: Ghomarides-Malaguides; **M**: Marseilles ; **SA** : south-Alpine ; **Se :** Sebtides-Alpujarrides **; S-M :** Sesia-Margna **; T :** Tuscany ; **Z** : Zurich.

Late Cretaceous-Eocene

At the end of the Late Cretaceous-Eocene, the north African passive margin is still essentially undisturbed, except the large Triassic diapirs emplaced within the Cretaceous marly sediments. The northern boundary of the continental margin is marked by the Beni Malek serpentinites (Michard *et al.*, 1992). The Maghrebian Flysch Trough extends onto an oceanic or transitional crust, now almost entirely subducted. The Maghrebian Flysch Trough includes a transform fault zone allowing the Alboran Terrane to drift along north Africa. Moreover, there is good reasons to assume that the Sebtide-Alpujarride margin was bounded to the NW by an oceanic seaway (Guerrera *et al.*, 1993). Relics of this domain, initially located between Iberia and the Alboran Terrane, can be found in the Nevado-Filabride units, which underlie the Alpujarride nappes in the Central-Eastern Betics. The serpentinites, metagabbros and meta-pillow basalts of the Mulhacen units (upper Nevado-Filabride) may be regarded essentially as ophiolitic remnants (Puga *et al.*, 1989, 1999), although the occurrence of transitional crust must not be excluded (Gómez-Pugnaire and Muñoz, 1991). SE-dipping subduction zone is shown in the marginal Alboran blocs. This subduction is recorded by HP-LT metamorphic imprints in the Sebtides-Alpujarrides units, and it would then represent the southwest projection of the Western Alps-Corsica subduction

Oligocene

The Africa-Europe convergence is now accommodated by two zones of contractional deformation, located on both sides of the Maghrebian Flysch Trough (Fig. 1b). South of the trough, the African margin deformation is mainly documented in the Eastern Rif. In this area, the transpressional shortening of the Ketama-northern Temsamane units is associated with low-grade recrystallizations, dated at about 28 Ma (Monié *et al.*, 1984).

As a matter of fact, the Ghomaride-Malaguide domain is emerged after the Middle-Late Eocene, and likely changes during the Late Eocene-Oligocene into an elevated mountain belt.

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The uplift of the Ghomaride-Malaguide mountains likely resulted from the collision of the Alboran orogenic arc with the Iberian margin. The arc collision would have also caused contractional deformation of the previously thinned Alpujarride-Sebtide nappes beneath their Ghomaride-Malaguide overburden.

Late Oligocene-Middle Miocene

The topographic profile of the Rifian transect dramatically changed by the end of the Early Miocene (Fig. 1c). The inner Alboran Mountains collapse, being converted into a marine basin through an intense rifting process during the Late Oligocene-Early Miocene (Mauffret *et al.*, 1992; García-Dueñas *et al.*, 1992; Watts *et al.*, 1993; Chalouan *et al.*, 1997). Onshore, the extensional tectonics is responsible for late- to post-metamorphic low-angle normal faults, such as the Zaouia fault in between the Ghomaride and Sebtide units, and equivalent structures in Spain (García-Dueñas *et al.*, 1992; Balanyá *et al.*, 1997). Exhumation of the Sebtide-Alpujarride metamorphic units is achieved by the late Aquitanian, with Alpujarride rocks and HP-LT minerals being reworked in Burdigalian sediments on the Betic side of the orogen (Durand-Delga *et al.*, 1993).

On the external side of the rifted area, the Alboran Terrane units are now thrust above the adjacent Flysch Trough. The oceanic trough infill is converted into an accretionary prism of flysch units, which indicates that the underlying lithosphere is subducted, together with that of the Predorsalian zone, beneath the Alboran domain. The uplift of the accretionary prism would have also allowed the detachment and inward sliding of flysch masses onto the Ghomaride-Sebtide post-nappe cover, giving birth to the Jebel Zemzem and Riffiene inliers (Fig. 1d; Feinberg *et al.*, 1990; Chalouan *et al.*, 1995b).

Along the external boundary of the suturing Maghrebian Trough, the contraction now affects the whole Intrarif-Mesorif zones, the sediments of which are incorporated into the growing accretionary prism.

Late Miocene-Pliocene

From the Middle Miocene stage (Fig. 1c) to the Late Miocene one (Figure 1d), the main changes concern the External Zones: they have been affected by a strong contraction, before being partly overlain by late Tortonian-Messinian molasses. In the Eastern Rif, folding of the Lower-Middle Miocene ante-nappe deposits is accompanied by axial-plane cleavage development and very low-grade recrystallization, dated at 7-8 Ma (Tamda, Kouine and south Temsamane windows; Monié *et al.*, 1984; Favre, 1992).

The Upper Miocene-Lower Pliocene post-nappe sediments onlapped widely onto the External Rif. After this period of mild surface extension, contraction occurred again in the Central Rif with a dominant NNE compressional trend, responsible for the folding of the post-nappe syncline area during the early Pliocene (Michard *et al.*, 1996; Samaka *et al.*, 1997; Samaka, 1999).

CONCLUSION

The Rif belt appears as the southern part of an orogen formed by a three-fold collage, involving two continental plates, Iberia and Africa, and an exotic Alboran Terrane, together with remnants of the intervening oceanic areas. The Western Mediterranean Alpine belt was successively bounded by i) a SE-dipping, Late Cretaceous-early Oligocene Alpine subduction zone, bringing the Sebtide-Alpujarride and adjacent oceanic/transitional crust beneath the Ghomaride-Malaguide lithosphere; and ii) a NW-dipping, late Oligocene-Miocene Apenninic-Maghrebide subduction zone, born in the southward projection of the Western Alps back-thrust zone (Doglioni *et al.*, 1998).

The coeval extension (Alboran Sea) and contraction (Rif belt) can be interpreted in the frame of a subduction zone roll back model (Lonergan and White, 1997). The sinking slab detachment (slab break-off) hypothesis may explain the onset of the Alboran calc-alkaline magmatism (Zeck *et al.*, 1999).

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BOOK IN STOCK

Mécanismes sédimentaires et structuration récente de la marge méditerranéenne marocaine (partie occidentale)

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ABSTRACT

This study is based on the analysis of drilling core samples on the Moroccan Mediterranean margin (western part). The distribution of the different sedimentary facies is controled by hydrodynamic (Atlantic-Mediterranean exchange) and glacio-eustatic factors, morphostructural inheritance and lithology of the outcropping geological formations. It attempts to hierarchize these differents factors, in order to provide a type sequence of the upper Pleistocene-Holocene. Mineralogical and textural studies allow to distinguish between a marine and a continental origin for sedimentary components and to better understand late Quaternary and modern sedimentation in this key area of the Mediterranean. In particular, the fine sequential analysis details the characteristics of the recent Quaternary sedimentary filling. Sedimentological, mineralogical and micropaleontological records show a good correlation with climato-eustatic changes which are well established in the deep Western Mediterranean On this basis, we establish a biostratigraphic model, estimate sedimentation rates and reconstruct the different paleoclimatic, paleoeustatic and paleohydrologic episodes for the Alboran Sea since the last glacial (isotopic stage 2) until the present (isotopic stage 1).

INTRODUCTION

Notre présentation est basée sur une étude pluridisciplinaire de l’environnement littoral et marin de la marge méditerranéenne marocaine (Ammar, 1987; El Moumni, 1987; Tesson *et al.*, 1987; El Moumni et Gensous, 1991; El Moumni et Monaco, 1992; El Moumni, 1994; El Moumni *et al.*, 1995; Hassouni *et al.*, 1998; El Moumni *et al.*, 1999; El Hmaidi, 1999; El Khanchoufi *et al.*, 2000). Ce travail porte sur l’extrême sud-ouest de la Méditerranée occidentale (Fig. 1). Il s’appuie sur l’analyse de profils sismiques multitraces, hautes résolution et sur l’étude sédimentologique, géochimique et micropaléontologique d’échantillons de surface et de dépôts, prélevés par carottage, durant plusieurs campagnes (ALBOMAR 1985, ALBOSED 1986 et STRAKHOV 1994). Il se fixe comme objectif l’étude des mécanismes hydrosédimentaires dans une zone de forts échanges : échanges hydrologiques à travers le détroit de Gibraltar, échanges continent-océan à travers la marge SW-Alboran. Ce qui est recherché à travers l’étude de l’accumulation sédimentaire, c’est l’évolution de ce contexte au cours des variations glacio-eustatiques et climatiques récentes.

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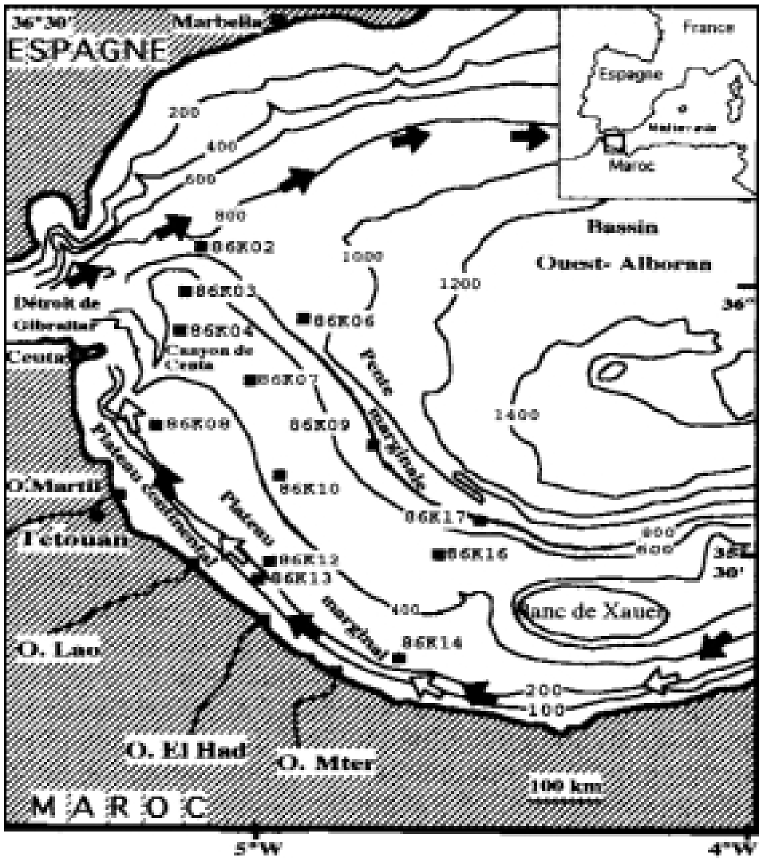

Fig. 1. Bathymétrie de la zone d'étude et positionnement des prélévements par carottage Kullenberg. Les flèches pleines et vides indiquent respectivement la circulation superficielle des masses d'eaux d'origine atlantique (Arnone *et al.*, 1990) et la circulation des masses d'eaux méditerranéennes profondes (Preller, 1986; Arnone *et al.*,. 1990).

Evolution structurale récente

L'évolution structurale récente est marquée par:

- la compression nord-sud, largement démontrée dans le Rif et la chaîne bétique, entre lesquelles la mer d'Alboran est enclavée (Maldonado and Comas, 1992; Chalouan and Michard, ce volume); la structuration majeure de ce secteur est marquée par la ride d'Alboran. Cette compression remobilise parfois des accidents anciens tel que le décrochement sénestre de Jebha (passant par le flanc méridional de la ride) qui rejoue en faille inverse (Fig. 2);
- les manifestations diapiriques qui ont joué un rôle important aussi bien dans l'organisation actuelle de la couverture sédimentaire qu'au niveau de la morphologie; il s'agit d'un diapirisme argileux qui s'enracine au Miocène inférieur et qui parfois recoupe l'ensemble de la couverture sédimentaire pour imprimer son empreinte au niveau de la morphologie actuelle; il faut toutefois souligner que le diapirisme n'intéresse que le bassin ouest-Alboran dans lequel il est cantonné.

Matériel et méthodes

Les prélèvements (Fig. 1, Tab. I) ont été réalisés à l'aide de carottiers de type "Kullenberg" et "Box Core" à grosse section, lors de la campagne océanographique ALBOSED II - 1986 à bord du N/O *Catherine Laurence* (CNRS - France), complétée ensuite par la campagne STRAKHOV à bord du N/O *N. Strakhov* en 1994. Après analyse par gammadensimétrie et radiographie aux rayons X., les carottes sont ouvertes et un log lithologique détaillé est établi. Les taux de carbonates ont été déterminés par calcimétrie sur l'échantillon total. Après séparation sur tamis de 40 µm, les constituants de la fraction sableuse ont été observés sous la loupe binoculaire et la fraction argileuse (< 2 µm) analysée par diffractométrie de rayons X à partir de dépôts orientés sur lames, sur diffractomètre Philips PW 1729 à anticathode de cuivre.

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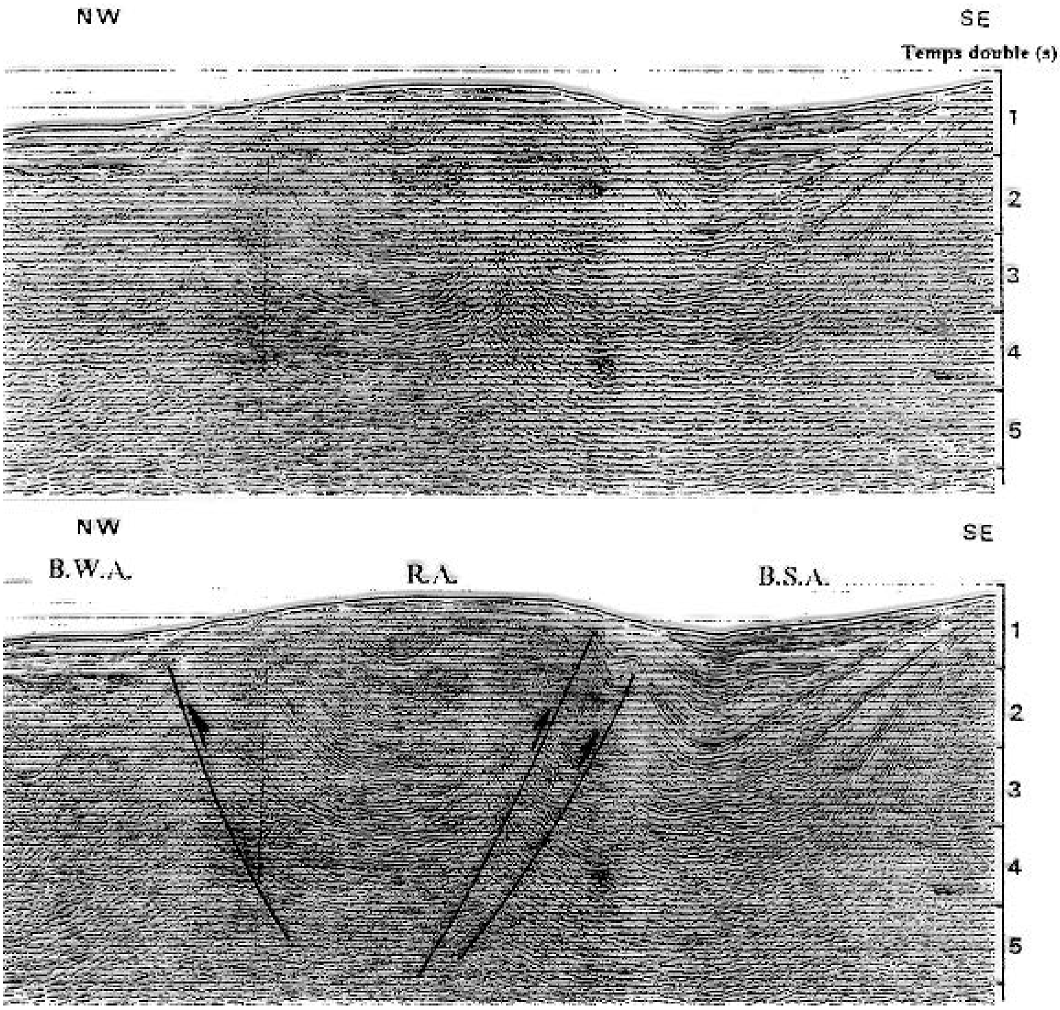

Fig. 2. Profil de sismique reflexion multitrace montrant le jeu en faille inverse et la montée actuelle de la ride d'Alboran (profil situé au niveau du banc de Xauen).
BWA = Bassin west d'Alboran; **R.A** = Ride d'Alboran; **BSA** = Bassin Sud d'Alboran.

Au sein de la fraction sableuse, la détermination et le comptage des espèces de foraminifères planctoniques permettent, par comparaison avec les études réalisées sur les secteurs voisins (Devaux, 1985; Pujol et Vergnaud-Grazzini, 1989; Vergnaud-Grazzini *et al.*, 1989; Vergnaud-Grazzini et Pierre, 1991), de tracer l'évolution paléoclimatique et de fournir un cadre chronostratigraphique. Les données obtenues ont été corroborées par datations absolues.

RESULTATS

Quatre faciès sédimentaires (vases beiges, vases grises, vases grises à monosulfures et sables vaseux, sont distingués sur la base des caractéristiques lithologiques, texturales, minéralogiques et micropaléontologiques (Fig. 3, 4, 6 et 7). Du sommet vers la base, on distingue :

Vases beiges

Elles constituent la partie sommitale des carottes situées au-delà du plateau continental (à partir de l'isobathe 300 m). Il s'agit de vases beiges homogènes, plastiques (hémipélagiques à pélagiques). Leur épaisseur est d'environ 25 cm sur la pente continentale, de 0 à 60 cm sur le plateau marginal et peut atteindre jusqu'à 120 cm au niveau de la pente marginale. Elles sont limitées à la base par un contact franc qui les sépare des vases grises sous-jacentes. La texture très fine (valeur moyenne de la médiane de 1 µm) diminue globalement (de 2 à 0,5 µm) de la base vers le sommet de la couche et du plateau continental vers le large. Le taux moyen de carbonates est de 15% du poids sec du sédiment avec souvent une légère diminution (de 20 à 10%) de la base vers

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Tab. 1. Coordonnées géographiques des prélévements étudiés.

N° carottes	Latitude (N)	Longitude (W)	Profondeur (m)	Longueur (m)
K02	36° 05,00	05° 04,00	830	4,5
K03	35° 57,80	05° 07,00	520	4,8
K04	35° 55,30	05° 08,70	520	4,9
K06	35° 51,50	04° 55,10	670	4,7
k07	35° 48,40	04° 59,80	440	4,3
k08	35° 43,40	05° 08,80	313	2,5
k09	35° 42,71	04° 44,85	785	7,5
K10	35° 37,45	04° 55,80	440	9,5
K12	35° 28,70	04° 59,00	214	4,5
K13	35° 27,50	05° 01,00	140	0,5
K14	35° 19,20	04° 43,00	300	4,8
K16	35° 29,30	04° 35,10	440	4,8
K17	35° 33,32	04° 32,68	755	4,8

le sommet. La fraction sableuse, en moyenne de 8% par rapport au poids sec du sédiment, organogène, est essentiellement constituée de foraminifères planctoniques souvent glauconitisés.

Dans la fraction argileuse (<2 µm), le cortège minéralogique des argiles est caractérisé par la prédominance de l'illite (45%) et de la chlorite (25%) au niveau des embouchures des oueds et de la pente continentale contre 30% seulement pour l'ensemble smectite (15%) et kaolinite (15%). Vers le plateau marginal et au niveau de la pente qui lui est adjacente, c'est plutôt la smectite (30%) et la kaolinite (22%) qui prédominent légèrement sur l'ensemble illite (40%) et chlorite (8%). A ce niveau, la smectite et la kaolinite montrent également un gradient décroissant d'ouest vers l'est jusqu'à des teneurs voisines de 14% chacune. Verticalement et dans la plupart des carottes, la phase argileuse montre en outre un enrichissement vers le sommet des teneurs en illite de 40 à 50% et une diminution des teneurs en smectite de 40 à 25% (Fig. 3).

Vases grises

Ce sont des vases de couleur sombre (hémipélagiques), homogènes argileuses (médiane <1,4 µm). Elles sont absentes au niveau du plateau continental externe (86k-13) et de la pente continentale (86k-08, 86k-12) sauf à l'est de l'embouchure de l'oued Mter (86K-14) où sont présentes sur une épaisseur de 30 cm. Elles atteignent 150 à 200 cm au niveau du plateau marginal (86K-07 et 86K-16) et 100 à 80 cm sur la pente marginale (86K-06, 86K-17). Elles sont limitées à la base par un contact franc ou par un niveau coquillier qui les séparent des vases grises à monosulfures sous-jacentes. Les teneurs en carbonates par rapport au poids sec du sédiment diminuent de la base (23%) vers le sommet (15%), parallèlement à l'augmentation de la médiane granulométrique de 1 à 2 µm.

Dans la fraction sableuse (<10% du sédiment brut), les composants sont principalement organogènes (débris de coquilles, foraminifères benthiques et planctoniques) à la base, et terrigènes vers le sommet. Dans la fraction argileuse (<2 µm), la minéralogie montre toujours une abondance d'illite (45%) et chlorite (25%) par rapport à la smectite (<10%) et kaolinite (17%) ; à partir du plateau marginal, les taux de smectite augmentent jusqu'à 30%. A ce niveau, la smectite et la kaolinite montrent encore un gradient légèrement décroissant d'ouest vers l'est jusqu'à des teneurs voisines de 14% chacune.

Vases grises à monosulfures

Elles sont absentes au niveau du plateau et de la pente continentale, où elles n'ont été recoupées que par la carotte Alb 86K-14 vers 300 m de profondeur au large de l'oued Mter. Il s'agit de vases grises (hémipélagiques), relativement compactes (teneur en eau d'environ 40%), parsemées de taches noires de monosulfures et de fragments charbonneux et pyritisés, nombreux à la base et diminuant vers le sommet. Elles sont parfois homogènes et recoupées momentanément de passées sablo-coquillières d'épaisseur décimétrique à centimétrique. Le contact avec les vases grises sus-jacentes est franc et il est parfois souligné par un niveau coquillier. Au niveau de la pente marginale, on note la présence de galets mous de vases; les radiographies aux R.X. mon-

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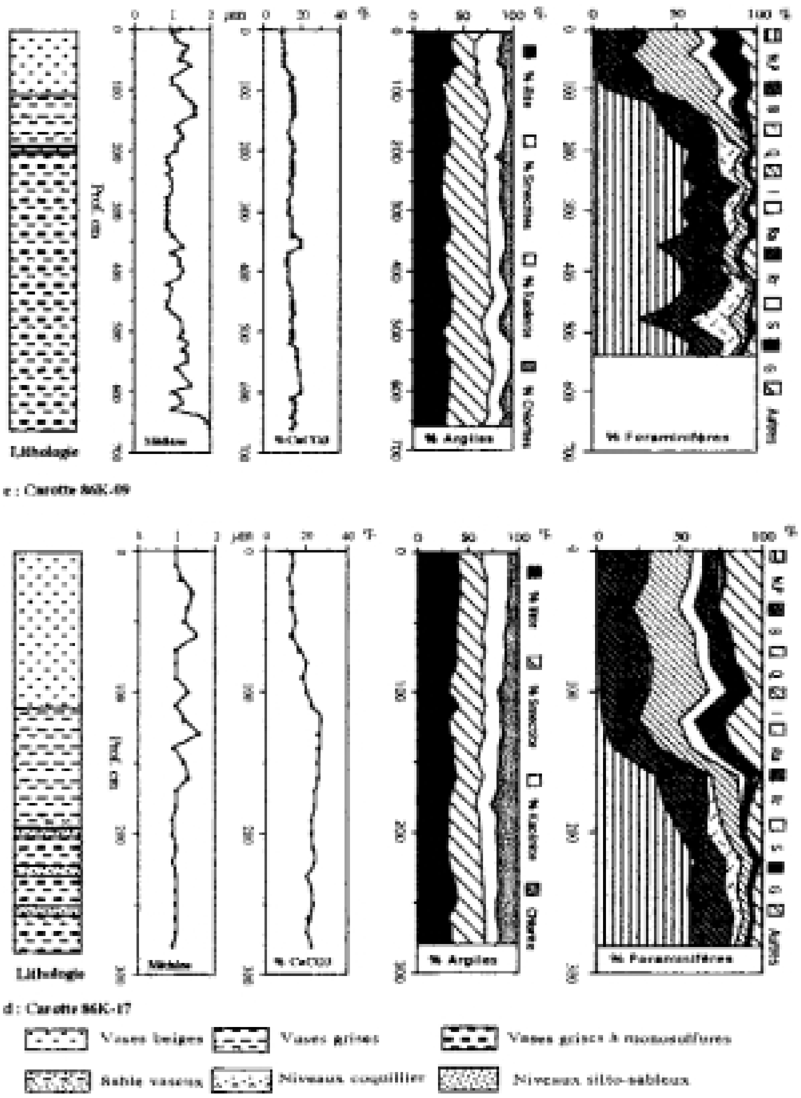

Fig. 3. Données sédimentologiques, minéralogiques et micropaléontologiques des 86K09 et 86K17 de la pente marginale.
N.P : *N. pachyderma* (forme senestre et dextre), **B** : *G. bulloides*, **Q** : *G. quinqueloba*, **I** : *G. inflata*, **Ra** : *G. ruber alba*, **Rr** : *G. ruber rosea*, **S** : *G. scitula*, **G** : *G. glutinata*, **Autres** : Autres espèces.

trent la présence de bioturbations (filaments et traces d'Annélides, débris de coquilles) abondantes à la base et dont la fréquence décroît progressivement vers le haut. La médiane diminue (de 2 à 1 µm) de bas en haut.

Les teneurs en carbonates, comprises entre 20 et 30% par rapport au sédiment brut sur la pente continentale et le plateau marginal, diminuent jusqu'à 15% au niveau de la pente marginale. Ces teneurs sont plus élevées, voisines de 40% dans les niveaux coquilliers.

La fraction sableuse représente en moyenne 7% du sédiment total. Sur le plateau marginal, la phase organogène est constituée de débris de coquilles, de foraminifères benthiques et planctoniques généralement usés. Vers le sommet du faciès, la phase détritique (minéraux en grains, micas, minéraux lourds et débris de roches) devient prédominante. Au niveau de la pente marginale, la fraction sableuse est constituée presque entièrement de foraminifères planctoniques.

Dans la fraction argileuse (<2 µm), la minéralogie montre la prédominance de l'illite (50%) et de la chlorite (25%) au niveau de la pente continentale, qui s'atténue vers le plateau et la pente marginale au profit de la smectite (jusqu'à 40%) et de la kaolinite (jusqu'à 30%).

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Sables vaseux

Ils ont été recoupés sur une épaisseur d'environ 4,5 m sans que leur base ne soit atteinte. Il s'agit de sables vaseux massifs, gris-clair, entrecoupés par des niveaux organogènes (lamellibranches, bryozoaires, ...) d'épaisseur centimétrique à décimétrique. Ils sont recouverts, localement, de vases beiges (86K-08) et sont caractérisés par la présence de quelques galets mous de vases dispersés de façon sporadique dans tout le faciès.

La fraction sableuse représente 60 à 90% du sédiment total. L'évolution granulométrique est granodécroissante (de 230 à 150 µm). La radiographie aux rayons X montre des traces de bioturbations dans les 2/3 inférieurs du faciès et des laminations vers le sommet. Les teneurs en carbonates totaux sont de 30% à la base et de 15% au sommet. La phase détritique est composée de fragments de roches (micaschistes, gneiss) et de minéraux (quartz, grenat et pyroxène). La phase organogène est constituée de débris de Lamellibranches, d'échinodermes, de spongiaires et de foraminifères, surtout benthiques. La majorité des constituants sableux est glauconitisée et présente des traces d'usure et de remaniement. Dans la fraction argileuse, la minéralogie est caractérisée par l'abondance de l'illite (45%) et de la chlorite (30%).

Micropaléontologie

Trois carottes (86k-06, 86k-09 et 86k-17), situées sur la pente marginale et recoupant la majorité des faciès fins, ont fait l'objet d'une étude micropaléontologique (Fig. 3 et 4). 15 espèces de foraminifères planctoniques ont été reconnues et regroupées en associations caractérisant différents types de climat (Cossement *et al.*, 1984; Moullade, 1983; Blanc-Vernet, 1984; Devaux, 1985; Venec-Peyre *et al.*, 1991). Les espèces Globigerinoïdes ruber (formes alba et rosea) caractérisent un climat chaud; l'espèce *Globorotalia inflata* indique un climat transitionnel à tendance chaude; l'espèce *Globigerina bulloides* indique un climat transitionnel plutôt froid et les espèces *Neogloboquadrina pachyderma* et *Globorotalia quinqueloba* caractérisent un climat froid.

Les vases beiges sont caractérisées par une association d'espèces "d'eaux chaudes" ou encore sub-tropicales, comme *G. ruber* (formes alba et rosea) avec 15 à 27% (Fig. 3 et 4) et *G. inflata* (20 à 40%). Les espèces *N. pachyderma* et *G. quinqueloba*, considérées "d'eaux froides" ou subarctiques, ne dépassent pas 10% de l'ensemble faunique. Une datation au ^{14}C effectuée au sein de ce faciès nous ont donné un âge d'environ 4504 ± 38 ans B.P. Ces caractéristiques et la position de ce faciès au sommet de la séquence de remplissage conduisent à dater sa mise en place durant la période chaude Holocène. Plus au large, dans le bassin occidental d'Alboran, Devaux (1985), Pujol et Vergnaud-Grazzini (1989), Vergnaud-Grazzini et Pierre (1991) confirment, sur la base de datations absolues et d'isotopes stables, la mise en place de cet assemblage faunique durant la période chaude Holocène terminal entre 7 000 ans B.P. et l'actuel.

Les vases grises, situées sous les vases beiges, se caractérisent par l'augmentation progressive du pourcentage d'espèces subarctiques. Les espèces *N. pachyderma* et *G. quinqueloba* passent de 36% à 120 cm à 76% à 200 cm vers la limite vases grises-vases grises à monosulfures. L'espèce *G. bulloides* varie entre 20 et 28%. A l'inverse, l'espèce *G. inflata*, à comportement chaud ou transitionnel, montre des pourcentages relativement élevés à 120 cm (18%) et disparaît vers la base. Il en est de même des espèces sub-tropicales (*G. ruber alba* et *rosea*) dont les pourcentages diminuent de 15% au sommet à 1 à 2% à la base (Fig. 3 et 4). Une datation au ^{14}C effectuée au sein de ce faciès nous a donné un âge d'environ 8653 ± 55 ans B.P. Les vases grises pourraient être attribuées à la période Holocène basal entre 10 000 et 7 000 ans B.P. (Devaux, 1985; Pujol et Vergnaud-Grazzini, 1989; Vergnaud-Grazzini et Pierre, 1991).

Les vases grises à taches noires de monosulfures sont caractérisées par la prédominance des espèces "d'eaux froides" ou subarctiques: l'espèce *N. pachyderma* à elle seule atteint environ 60%; les espèces *G. quinqueloba* (subarctique) et *G. bulloides* (à tendance froide) atteignent près de 40%. L'espèce *G. inflata* est présente en faibles proportions (<à 10%) (Fig. 3 et 4). Ces résultats permettent d'attribuer aux vases grises à monosulfures une mise en place durant la période de déglaciation entre 18 000 et 10 000 ans B.P.

A la base de ces données chronologiques et micropaléontologiques, les phases majeures de la transgression Postglaciaire-Holocène sont relativement bien identifiées. Ainsi, le passage

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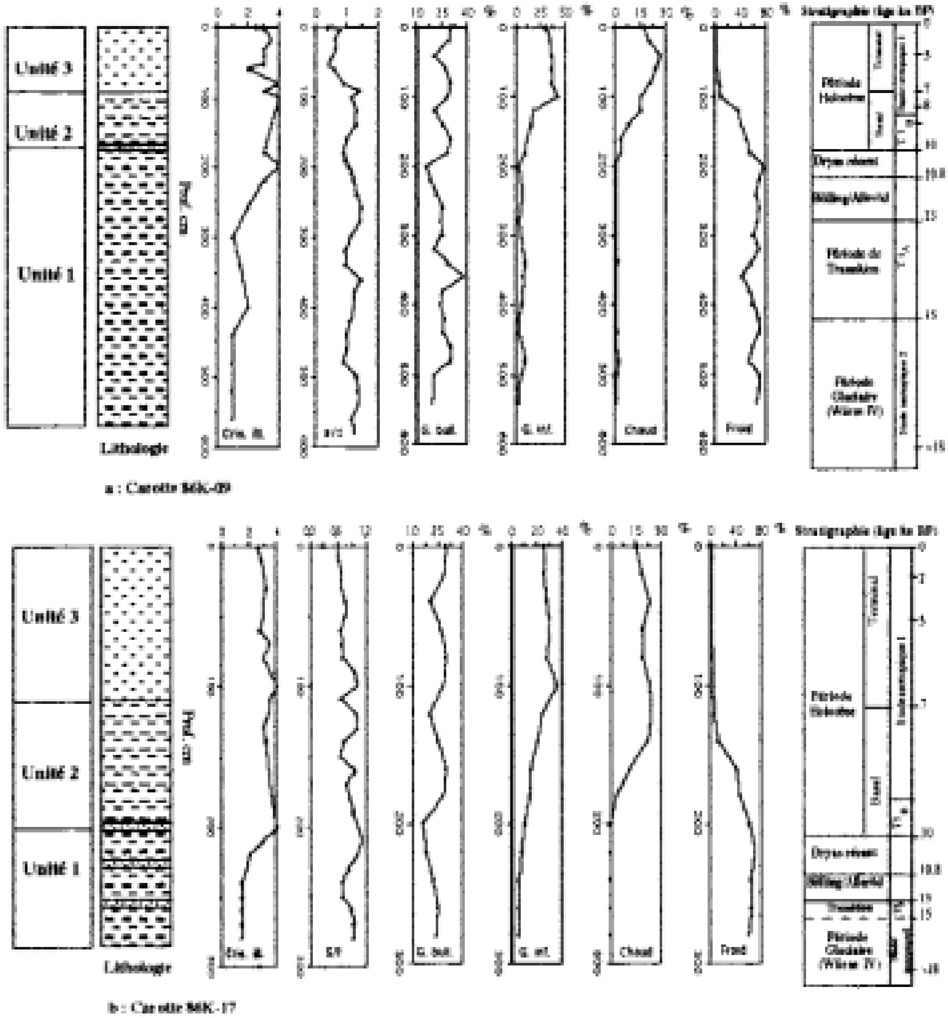

Fig. 4. Stratigraphie des dépôts et corrélation avec les épisodes paléoclimatiques depuis le dernier glaciaire sur la marge méditerranéenne marocaine à travers les carottes de la pente marginale 86K09 et 86K17.
Cris. ill. : cristallinité de l'illite, **S/I** : smectite / illite, **G. bull**. : *G. bulloides*, **G. inf.** : *G. inflata*, **Chaud** : espèces d'eaux chaudes = *G. ruber* (alba et rosea), **Froid** : espèces d'eaux froides = *N. pachyderma* et *G. quinqueloba*.

Pléistocène-Holocène peut être fixé à la limite vases grises à monosulfures-vases grises; la transition Holocène basal-Holocène terminal est fixé au passage vases grises-vases beiges (Fig. 4, 5 et 6).

Taux de sédimentation

D'après le découpage stratigraphique proposé pour la pente marginale (Fig. 4, 5 et 6), la séquence Holocène, épaisse de 180 à 200 cm, aurait un taux de sédimentation moyen de 18 à 20 cm/1000 ans environ. Dans le détail, l'Holocène terminal entre 7 000 ans B.P. et l'actuel, épais d'environ 60 à 120 cm, aurait un taux de sédimentation d'environ 17 à 8 cm/1000 ans ; l'Holocène basal, entre 10.000 et 7.000 ans B.P., épais d'environ 80 à 120 cm, aurait un taux de sédimentation d'environ 40 à 27 cm/1000 ans. La période de déglaciation, entre 18 000 et 10 000 ans B.P., épaisse de plus de 500 cm, aurait un taux de sédimentation supérieur à 60 cm/1000 ans. Des datations, obtenues récemment sur des carottes prélevées dans le bassin occidental (en cours de publication avec l'équipe de l'Institut des Sciences de la mer de Barcelone), d'autres en utilisant les foraminifères benthiques (El Khanchoufi *et al.*, 2000), corroborent ce découpage stratigraphique avec des taux de 20 cm/1000 ans pour la période Holocène et de 37 cm / 1000 ans pour la période anté-Holocène (entre 15 000 et 11 000 ans B.P).

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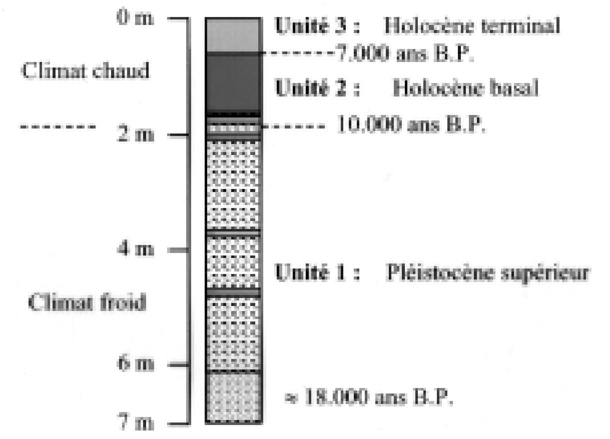

Fig. 5. Séquence sédimentaire type au Quaternaire terminal au niveau de la marge méditerranéenne occidentale marocaine.

DISCUSSION

Séquence type et facteurs climato-eustatiques de la sédimentation

Dans notre zone d'étude, et bien que les âges exacts des dépôts ne soient pas définitivement établis, une tentative de corrélation entre les intervalles de la courbe eustatique (Aloisi *et al.*, 1978; Bard *et al.*, 1993) et les différentes unités de dépôt de la séquence type, est proposée (Fig. 4 et 5). Cette séquence a été établie sur la base des critères lithologiques, sédimentologiques, minéralogiques, micropaléontologiques et des corrélations inter-prélèvements. Il s'agit d'une séquence type transgressive, paléoclimatique, globalement granodécroissante. L'unité 1, absente au niveau du plateau continental et de la pente supérieure, qui étaient alors émergés, est caractérisée par des vases grises à monosulfures et à microfaune de climat froid à sub-arctique. Elle correspond au bas niveau marin du dernier glaciaire et à la première étape de la trangression post-glaciaire. L'unité 2, sus-jacente, à vases grises hémipélagiques à pélagiques et à microfaune de climat chaud à caractère transitionnel, serait d'âge Holocène basal et mise en place lors de la deuxième étape de la transgression post-glaciaire. L'unité 3, représentée par des vases beiges pélagiques et à microfaune de climat chaud à sub-tropical, se serait déposée à l'Holocène terminal lors de la période du haut niveau marin installé depuis environ 7 000 à 6 000 ans B.P.

Les sables vaseux du plateau externe rappellent, par leurs caractéristiques, les sables reliques du large, bien connus sur le rebord de l'ensemble des plateaux continentaux méditerranéens et datés d'environ 18 à 20 000 ans B.P. (Aloisi, 1986; Ercilla, 1992; El Hmaidi, 1993; El Moumni, 1994). Ce faciès représenterait donc des dépôts côtiers et serait déposé lors du dernier maximum régressif et pendant les premiers stades de la transgression post-glaciaire (unité 1). Les sables vaseux de la pente continentale pourraient être de même âge et seraient, d'après leurs caractéristiques, des sables turbiditiques déposés par les différents oueds débouchant directement sur le rebord de la pente continentale lors du dernier maximum glaciaire.

Évolution paléoenvironnementale et mécanismes sédimentaires

Mécanismes sédimentaires

Sur le plateau continental et le haut de pente (jusqu'à -200 m), où l'action hydrodynamique des eaux atlantiques superficielles (est-ouest) est forte, seuls sont rencontrés les sables moyens à fins et les sables vaseux. Ces faciès légèrement grossiers pourraient résulter de l'action de ces courants superficiels par vannage des dépôts et évacuation des argiles et des silts fins vers l'Ouest et les zones plus profondes (plateau marginal et bassin) et en laissant affleurer des dépôts sableux grossiers remaniés sur le plateau continental et la pente adjacente.

Au delà du bas de pente continentale et jusqu'au bassin (de -300 à -900 m), où l'action hydrodynamique des eaux méditerranéennes profondes (est-ouest) est importante, la séquence sédimentaire devient plus complète, diversifiée et bien différenciée; ainsi, les vases grises à monosulfures à la base sont surmontées par les vases grises, puis par les vases beiges qui occupent surtout la pente marginale.

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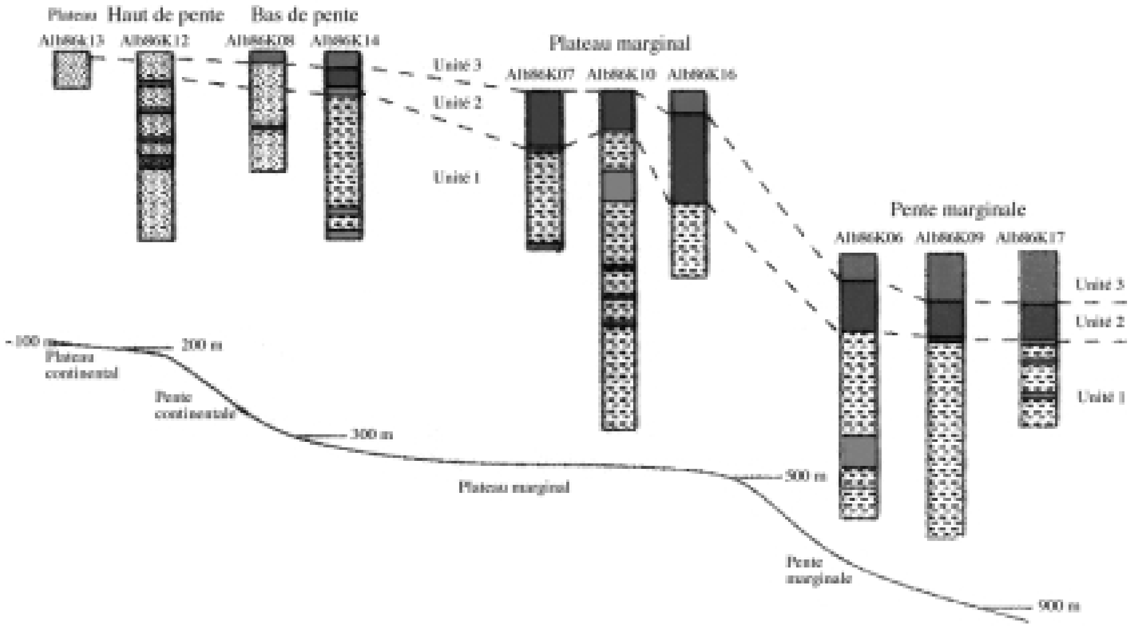

Fig. 6 : Découpage stratigraphique et corrélations inter-prélèvements dans la colonne sédimentaire depuis le rebord du plateau continental jusqu'au bassin (pour la légende voire Fig. 3 c et d).

Dans les secteurs de bas de talus et bassins, en dehors des zones de glissements gravitaires, les dépôts fins sont essentiellement hémipélagiques et/ou pélagiques.

Dans les sables vaseux de la pente, la présence de galets mous de vases témoigne des remaniements et des reprises sédimentaires d'origine hydrodynamique et gravitaire. Ainsi, la phase organogène, usée et riche en foraminifères benthiques, et la phase minérale détritique abondante proviendraient par reprise sédimentaire des sables remaniés situés sur la bordure du plateau continental. L'intercalation de niveaux silto-sableux coquilliers dans les vases grises à monosulfures permet d'interpréter les dépôts fins du plateau marginal comme des contourites (El Moumni, 1994) et ceux de la pente continentale comme des turbidites liées à des courants de turbidité de faible intensité (Stanley, 1972). Ces dépôts résultent probablement des reprises sédimentaires par des courants de fond, circulant d'est en ouest en longeant la pente méditerranéenne marocaine (Arnone *et al.*, 1990; Preller, 1986). Des faciès similaires ont été mis en évidence dans les dépôts post-glaciaires de la marge sud-portugaise au niveau de la ride de Faro le long du trajet des eaux méditerranéennes sortantes (Faugeres *et al.*, 1986).

La chenalisation des apports est peu importante du fait de la rareté des vallées sous marines à l'exception du canyon de Ceuta. De ce fait, les séquences de turbidites typiques de sables granoclassés sont rares par comparaison avec celles de la pente continentale située en face d'Al Hoceima et de la baie de Betoya (El Moumni, 1994) et avec celles de la pente espagnole (Ercilla *et al.*, 1994) où le domaine de pente est entaillée par de nombreux canyons sous marins.

Évolution des paléoenvironnements lors de la période glaciaire et de la transition entre 18 000 et 13.000 ans B.P.

Pendant la période froide glaciaire (Würm IV, stade isotopique 2) et la transition post-glaciaire, l'abondance des carbonates (30%) et les traces de bioturbation traduisent l'importance de la productivité organique de surface et de l'activité benthique à cette époque. La conservation exceptionnelle des tests carbonatés fragiles permet de classer le milieu de dépôt de l'unité 1 dans les milieux confinés à pH élevé (Mollineaux et Lohman, 1981), favorables à la préservation des carbonates. Le caractère réducteur du milieu de dépôt (présence de monosulfures et pyrite) a été accentué probablement par les forts taux de sédimentation (60 cm/1000 ans), l'abondance de la matière organique et une faible circulation. A cette époque, la mer d'Alboran aurait connu une éventuelle stagnation de l'eau profonde et une réduction de la circulation hydrologique comme c'était le cas dans le Golfe du Lion et dans toute la Méditerranée (Bourdillon, 1982; Nachite, 1984; Béthoux, 1984; Zahn *et al.*, 1987; Murat, 1991; El Hmaidi, 1993; El Hmaidi *et al.*, 1998; El Hmaidi, 1999).

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L'abondance de la phase détritique dans l'unité 1 reflète le fait que durant la période glaciaire et de transition post-glaciaire, le milieu était soumis à l'influence des zones continentales et côtières. Le mélange de foraminifères benthiques et planctoniques témoigne de l'étroitesse de la marge et par conséquent de l'interpénétration des étages bathymétriques à cette époque. L'état usé de la phase organogène montre l'importance des phénomènes de remaniements à partir du plateau externe. La bonne cristallinité de l'illite (indice faible) montre que le climat était froid. La période de transition T IA (entre 260 et 420 cm de la surface), caractérisée par un léger réchauffement climatique, est marquée par une petite augmentation de l'indice de cristallinité de l'illite (carottes 86k-09 et 86k-17) et du rapport smectite / illite (S/I) (Fig. 4a et b); de même, les foraminifères planctoniques montrent une légère augmentation des formes transitionnelles (*G. bulloïdes* et *G. inflata*) et une légère diminution des formes froides subarctiques.

Évolution des paléoenvironnements lors de la période de transition Pléistocène-Holocène

Dans les assemblages de foraminifères planctoniques, le maximum de formes subarctiques, *N. pachyderma* et *G. quinqueloba*, vers 200 cm de la surface pourrait indiquer le retour bref au froid lors de l'épisode du Dryas récent (11 000 à 10 000 ans BP). La transition Pléistocène-Holocène est marquée globalement par la diminution des minéralisations sulfurées et par conséquent du caractère réducteur. Cette évolution refléterait les premiers indices d'une circulation océanique qui s'amplifiera progressivement avec le temps. Aussi, le réchauffement climatique indiquant le passage à l'Holocène se marque par l'apparition des formes sub-tropicales *G. ruber*, l'épanouissement des formes transitionnelles *G. inflata* et la diminution progressive des formes subarctiques. Dans le même sens, la mauvaise cristallinité de l'illite (indice élevé-carottes 86k-09 et 86k-17) et l'augmentation du rapport S/I confirment le réchauffement climatique (Fig. 4a et b). Cette évolution minéralogique traduit généralement le réchauffement post-glaciaire avant les faunes planctoniques avec augmentation précoce des teneurs en kaolinite et en smectite et dégradation précoce de la cristallinité de l'illite (Cossement *et al.*, 1984; Grousset *et al.*, 1988).

Les teneurs en carbonates diminuent de façon presque générale au passage Pléistocène-Holocène (Fig. 3). L'augmentation des teneurs en CO2 des eaux marines profondes pourrait expliquer cette diminution, associée à une intensification de la formation des eaux marines froides dans le Golfe du Lion (Cacho *et al.*, 1994); une baisse dans la productivité organique de surface pourrait également être en cause.

Les teneurs en smectite, qui serait d'origine atlantique, ne montrent aucune fluctuation majeure sur l'ensemble de la colonne sédimentaire (Fig. 3). Ceci pourrait indiquer l'absence de changement dans le régime de la circulation des masses d'eaux entre l'Atlantique et la Méditerranée depuis la dernière glaciation. Cette conclusion confirme les travaux antérieurs de Cossement *et al.* (1984), Pujol et Vergnaud-Grazzini (1989), Vergnaud-Grazzini *et al.* (1989) et Vergnaud-Grazzini et Pierre (1991) et contredit l'hypothèse d'une inversion vers 10 000 ans B.P. au passage Pléistocène-Holocène dans le régime hydrologique en mer d'Alboran comme cela avait été proposé par Huang et Stanley (1974).

Évolution des paléoenvironnements lors de la période de transition Holocène basal-Holocène terminal

Cette transition T IB (vers 100 cm de la surface) est marquée par le passage de faciès essentiellement détritiques à caractère relativement réducteur dans l'unité 2, vers des faciès, de plus en plus fins, oxydés (couleur beige) et riches en foraminifères essentiellement planctoniques glauconitisés dans l'unité 3; ce qui indique le passage dans l'unité 3 vers un milieu de plus en plus profond et ouvert à une ventilation croissante des masses d'eaux. Parallèlement, pour les foraminifères planctoniques, les formes subarctiques s'éteignent et les formes transitionnelles à subtropicales s'épanouissent (Fig. 4). On note également dans les argiles des valeurs relativement élevées du rapport S/I ainsi qu'une faible cristallinité de l'illite lors du passage à l'Holocène terminal. Ceci indique probablement le réchauffement climatique de la période atlantique (vers 7 500 ans BP) également bien connue en Méditerranée (Aloisi *et al.*, 1978; Cossement *et al.*, 1984; Blanc-Vernet, 1984; Pujol et Vergnaud-Grazzini, 1989; Vergnaud-Grazzini *et al.*, 1989; Vergnaud-Grazzini et Pierre, 1991; Venec-Peyre *et al.*, 1991; Murat, 1991; Cacho *et al.*, 1994). Cette période de réchauffement serait à l'origine d'une forte stratification des eaux de surface et d'une stagnation des eaux profondes au niveau du bassin d'Alboran, et pourrait être contemporaine de celle du dépôt du dernier sapropèle S1 de Méditerranée orientale (Murat, 1991; Cacho *et al.*, 1994) également connu dans le bassin d'Alboran (Comas *et al.*, 1996).

Évolution des paléoenvironnements lors de la période Holocène terminal à Actuel

Lors de l'Holocène terminal (unité 3), un autre maximum de réchauffement climatique apparaît (vers 40 cm de la surface) avec une diminution des pourcentages de *G. bulloïde*s, un maximum d'espèces sub-tropicales *G. ruber* et une diminution du rapport S/I et de la cristallinité de l'illte (Fig. 4). Cette fluctuation positive du climat pourrait correspondre au réchauffement de la période sub-boréale à sub-Atlantique (entre 4 500 et 2 000 ans BP) également bien connue en Méditerranée (Aloisi *et al.*, 1978; Cossement *et al.*, 1984; Vénec-Peyré *et al.*, 1991). Après cet événement, la diminution de *G. ruber* et l'augmentation de *G. bulloides* reflètent un léger refroidissement dans les eaux de surface et l'installation d'un climat semblable à l'actuel.

CONCLUSIONS

L'ensemble des travaux réalisés permet de proposer une chronologie relative des dépôts et de préciser les mécanismes et les conditions de leur mise en place depuis le dernier Glaciaire. Les enregistrements sédimentologiques, minéralogiques et micropaléontologiques se corrèlent relativement bien avec les changements climato-eustatiques bien établis dans les zones profondes de la Méditerranée occidentale. Cette corrélation permet d'établir un modèle biostratigraphique, d'estimer les taux de sédimentation et de reconstruire les différents épisodes paléoclimatiques, paléoeustatiques et paléohydrologiques pour la mer d'Alboran depuis le dernier glaciaire (stade isotopique 2) jusqu'à l'actuel (stade isotopique 1).

• La dernière période glaciaire et la transition post-glaciaire connaissaient une productivité organique de surface et une activité benthique relativement importante; cependant, la circulation était faible et le milieu de dépôt était réducteur.

• L'optimum climatique de l'Holocène se caractérise par la stratification des eaux de surface et la stagnation des eaux profondes au niveau du bassin occidental de la Méditerranée et serait l'équivalent du sapropèle S1 en Méditerranée orientale.

• L'action des courants atlantiques de surface se manifeste au niveau du plateau externe et le haut de pente par les remaniements à l'interface et le vannage de la fraction fine vaseuse vers l'Ouest et les zones profondes. Ces courants superficiels sont également responsables de l'alimentation en smectite et en kaolinite à partir de l'Atlantique. A ce propos, la minéralogie des argiles n'indique aucune inversion des courants de surface à travers le détroit de Gibraltar au passage Pléistocène-Holocène et par conséquent la constance du régime hydrologique depuis la dernière période glaciaire jusqu'à l'actuel.

• L'action des courants méditerranéens de fond est responsable de la mise en place des contourites au niveau du plateau marginal. Les courants de turbidité contrôlent la mise en place des dépôts turbiditiques sur la pente continentale et la pente marginale. Les faciès hémipélagiques à pélagiques, rencontrés au delà de la pente continentale, seraient déposés par décantation.

• Les séquences de turbidites typiques sont peu nombreuses du fait de la rareté des canyons sous marins. Les taux de sédimentation, élevés lors de la dernière période glaciaire, diminuent progressivement jusqu'à l'actuel, du fait du transfert des sources d'apports vers le continent par la transgression post-glaciaire.

• Le développement d'une sédimentation terrigène, par un phénomène de continentalisation à l'Holocène terminal, semble être généralisé à l'ensemble de la Méditerranée occidentale et serait lié à une période d'aridification, favorisant l'érosion sur le continent.

• La mise en place des dépôts est donc contrôlée par les facteurs hydrodynamiques (échange Atlantique-Méditerranée) et glacio-eustatiques auxquels s'ajoute l'héritage morphostructural et la nature lithologique de l'arrière pays métamorphique.

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Synthèse des connaissances géologiques et géodynamiques de la Méditerranée occidentale et de la marge algérienne en particulier

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CADRE GÉOLOGIQUE ET STRUCTURAL

Le bassin méditerranéen est ceinturé par des chaînes de montagnes appartenant à l'édifice alpin, conséquence de la fermeture des espaces océaniques téthysien et mésogéen.

Caractères généraux du domaine méditerranéen méridional

L'analyse structurale du domaine interne nord-algérien montrent des rapports de cause à effet entre certaines de ses structures et la distension qui accompagne l'ouverture du bassin méditerranéen occidental. Pour exemple, les failles transformantes méditerranéennes se prolongent avec des accidents dans la chaîne des Maghrébides. Elles épousent la trajectoire de l'Accident Sud Kabyle (A.S.K.) polyphasé, et le décalent.

Les premières constatations qu'il est possible de faire au sujet des marges périméditerranéennes, et en particulier ouest-algériennes, sont :

- une remobilisation tectonique récente "post-rift" qui affecte les séries du Miocène terminal à Plio-Quaternaire;
- réactivation d'anciens accidents sans apparition évidente de failles inverses; elle concerne une zone de dislocation majeure de la pente et de la marge qualifiée comme un "Boundary fault"; l'essentiel de la déformation est enregistré à terre;
- une subsidence importante aux pieds des marges associée à la subsidence générale du bassin;
- une pente continentale remarquablement abrupte, accompagnée d'une marge étroite.

Au niveau de la marge algérienne pratiquement tous ces caractères se retrouvent avec en plus certaines particularités:

- absence de prisme tectonique impliquant les séries mio-plio-quaternaires marines;
- décalage de l'ordre de 2500 m entre la base du Pliocène sur la marge et dans le bassin profond;
- absence de failles inverses sur la marge méditerranéenne, plutôt un redressement des failles normales structurant la pente;
- développement d'une dépression de bas de pente et fluage du salifère vers un domaine distal;
- surrection des massifs côtiers et des paléorivages quaternaires;

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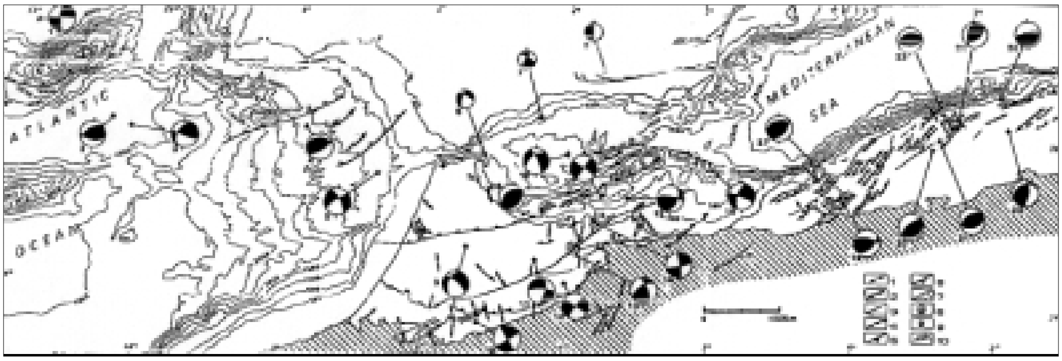

Carte néotectonique et sismotectonique du domaine Gorringe-Alboran-Rif-Tell (GALTEL). Les déformations compressives plio-quaternaires sont mises en évidence le long de la limite des plaques Afrique-Europe à l'Est de la transformante Açores-Gibraltar. Le domaine GALTEL apparaît comme une zone de cisaillement E-W marquée par un système transpressif.
1. direction de raccourcissement quaternaire; **2**. axe anticlinal; **3.** faille indifférenciée; **4**. faille normale; **5**. faille inverse active; **6**. faille inverse quaternaire; **7**. limite nord de l'avant-pays africain; **8**. mécanismes au foyer des séismes superficiels; **9.** mécanismes au foyer des séismes intermédiaires et profonds (taille réduite des cercles); **10.** bathymétrie en mètres (Meghraoui *et al.*, 1996).

- développement d'un système "plis-failles inverses" dans le Tell septentrional et une tendance à la réduction des espaces occupés par les bassins néogènes;
- caractéristiques essentielles de la sismicité et des mécanismes au foyer des séismes traduisant un style tectonique transpressif et décrochevauchant.

En intégrant un maximum de données marines et terrestres, le modèle de déformation exposé implique la bande "Gorringe-Alboran-Tell" (GALTEL) qui enregistre l'essentiel de la déformation intercontinentale.

LE REMPLISSAGE SÉDIMENTAIRE DU BASSIN MÉDITERRANÉEN

Les unités sédimentaires

Le remplissage sédimentaire évalué est de l'ordre de 7000 à 8000 m au niveau du bassin algéro-provençal où 3 unités principales ont été synthétisées :

- une unité supérieure d'âge "Plio-quaternaire" composée essentiellement par des dépôts marneux et turbiditiques;
- une unités moyenne constituée par un matériel évaporitique d'âge "Messinien";
- une unité inférieure "Infrasalifère" composée par des sédiments détritiques et marneux d'âge Miocène moyen à inférieur "Burdigalien précoce" qui d'ailleurs n'a pas été entièrement forée.

Dans les golfes de Valence et du Lion, des forages pétroliers ont traversé des séries peu épaisses d'âge "Aquitanien". Au Nord de l'Algérie, la série typique méditerranéenne a été reconnue par sismique réflexion.

CONCLUSION

L'ouverture du bassin est consécutive à l'orogène alpin qui s'accompagne d'une dislocation et dérive des différents éléments constituant les domaines internes et le bloc corso-sarde.

L'évolution géodynamique retenue fait intervenir l'ouverture d'un bassin marginal d'arrière-arc qui admet une distension oligo-miocène. Ce rift à croûte océanique, de moins en moins développée et de plus en plus jeune en allant vers le sud-ouest, concerne la partie centrale des bassins algéro-provençal et algéro-baléare.

Phase d'initiation du rifting (30 et 21 Ma.)

L'extrapolation du taux de sédimentation élevé aux premiers dépôts dans le bassin permet d'envisager un âge "Oligocène terminal" à "Aquitanien" (Miocène inférieur).

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Phase de rifting (21 et 18 Ma.)

L'expansion océanique et la distension active accélèrent l'ouverture du bassin. La sismique-réflexion montre une tectonique synsédimentaire en horsts et grabbens avec maintient d'un taux de sédimentation élevé: les sédiments du terme basal de "l'Unité inférieure" seraient synrift. L'âge attribué à cette étape est "Burdigalien inférieur".

Phase post-rifting

Elle est caractérisée par les sédiments issus des différentes transgressions miocènes. Les horizons sédimentaires post-rift, d'âge "Burdigalien terminal" (terme sommital), "Langhien", "Serravallien" et "Tortonien précoce" ne sont plus contrôlés par la tectonique. Ils drapent les différentes structures. La sédimentation post-tortonienne montre de nouveau un contrôle tectonique jusqu'à l'Actuel.

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Evolution structurale de la partie orientale du bassin d'Alboran au cours du Néogène (offshore occidental, Algérie)

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Résumé

L'histoire géologique de la partie orientale du bassin d'Alboran débute au Serravalien terminal-Tortonien et se continue jusqu'à l'heure actuelle. Elle est caractérisée par une suite d'événements géologiques ayant contrôlé l'organisation des dépôts sédimentaires. L'évolution géodynamique de la région montre un contexte général compressif lié à la convergence des plaques Europe-Afrique et au blocage du bloc d'Alboran et sa suture définitive à l'Afrique et à l'Ibérie. Dans ce cadre tectonique, les bassins tortoniens s'ouvrent en transtension.

Introduction

Notre zone d'étude s'intègre dans le domaine situé au Sud-Est du bassin d'Alboran oriental. Elle est comprise entre 35°10' et 35°55' de latitude Nord et 1° et 2°10' de longitude Ouest (Fig. 1). Elle appartient au plateau marginal algéro-marocain qui constitue la zone de transition entre le bassin algéro-baléare et le domaine d'Alboran au sens large.

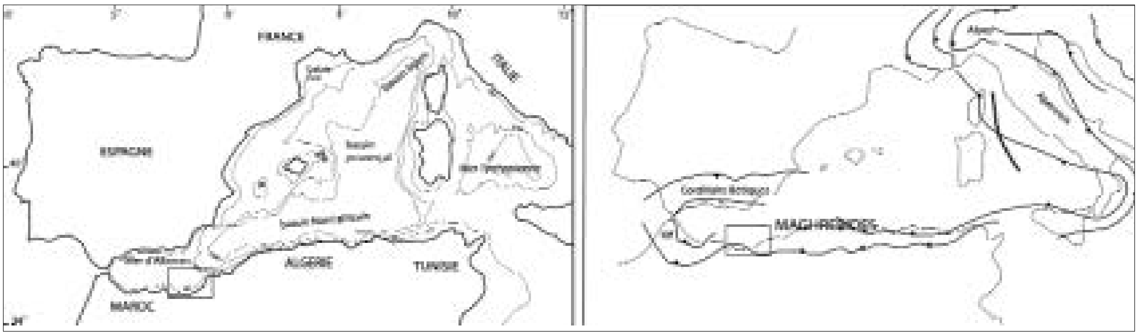

Fig. 1. Situation de la partie orientale de la mer d'Alboran.

Stratigraphie

Etablie sur un substratum de nature métamorphique composé de schistes verts, la série sédimentaire comprend de bas en haut (in Benabdelmoumen, 1997):

• le Serravalien-Tortonien inférieur. Il atteint une épaisseur de 843 mètres et est caractérisé par une série monotone d'argiles carbonatées et de marnes silteuses avec la présence de deux bancs gréseux de 34 mètres d'épaisseur chacun.

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• le Tortonien supérieur. Il est épais de 325 mètres. On observe la répétition de la série inférieure sauf que cette dernière est plus riche en matériel gréseux.

• le Messinien. Il est épais de 1271 mètres. Sa base est marquée par l'apparition de *Globorotalia mediterranea* (Iaccarino, 1984). Il peut être subdivisé en deux ensembles distincts:

- un Messinien inférieur, constitué d'une série marneuse avec des intercalations gréseuses. Son épaisseur est de 184 mètres.
- un Messinien supérieur, composé d'une série argilo-marneuse à fines passées de grès fins, épaisse de 1087 mètres. L'épisode évaporitique est marqué par la présence de quelques rosettes de gypse. La base du Messinien supérieur correspond au niveau de tufs volcaniques qui marquerait le début du confinement du bassin.

• le Pliocène. Il est épais de 574 mètres. Constitué à sa base d'une série argilo-carbonatée comportant de fines passées de grès fins avec intercalation d'un banc de grès fin de 20 mètres d'épaisseur. Cette série atteint 170 mètres. Au-dessus, on observe une série de 404 mètres d'épaisseur identique à la première avec toutefois une absence de niveaux gréseux. La série Pliocène et Quaternaire n'a pu être observée.

ÉVOLUTION STRUCTURALE

Serravalien à Tortonien-Messinien inférieur

L'ouverture du bassin se réalise au Serravalien terminal à Tortonien. Ce fait est corroboré par les datations radiochronologiques effectuées sur des échantillons de socle (Bellon, 1976) récupérés lors du forage "Habibas 1"(x: 01°56'31"13W; y: 35°37'58"51N; z: 4496,5 mètres; tranche d'eau: 923 mètres). Cet âge est beaucoup plus récent comparativement aux bassins les plus occidentaux du domaine d'Alboran qui se sont structurés au Burdigalien (Olivet *et al.*, 1973; Chalouan *et al.*, 1997).

La structuration du bassin se fait vers l'ouest. Celle-ci est guidée par le jeu de deux familles d'accidents. Des failles de direction NE-SW structurent le bassin vers le sud-ouest en demi-graben à regard nord-ouest, avec développement du bassin vers l'ouest. Les failles E-W sont beaucoup plus impliquées dans la portion est et nord du bassin; elles sont synsédimentaires et ont un rôle primordial dans la subsidence du bassin.

Au Tortonien supérieur, une transgression apparaît comme l'événement le plus marquant dans la paléogéographie Néogène de la région (Montenat, 1977). Elle est marquée par la discordance intra-tortonienne. Pendant cette période, l'ouverture du bassin se fait dans un contexte général compressif lié au rapprochement des plaques Eurasie-Afrique ainsi qu'au blocage du bloc d'Alboran qui finit par être solidaire au nord-ouest de l'Afrique et au sud-ouest de l'Ibérie. Il constitue un poinçon entre ces deux grandes plaques (Thomas, 1985). Dans le Tell méridional oranais et au Maroc oriental (Bassin de Melilla), une compression a été observée, se manifestant par l'existence de plis et d'accidents inverses de direction NE-SW (Philip et Thomas, 1977), alors que dans notre zone d'étude, on assiste à une ouverture de bassins dans un régime de type transtensif. Ce dispositif structural s'intègre de manière cohérente dans les schémas et modèles cinématiques associant des déformations en extension dans un contexte compressif (Platt et Vissers, 1989 et Frizon de Lamotte *et al.*, 1991 *in*: Meghraoui, 1996).

Au Messinien inférieur se poursuit le remplissage du bassin avec une accentuation de la subsidence qui a débuté au Tortonien.

Au Messinien supérieur

Des indices de confinement du bassin commencent à apparaître avec une microfaune planctonique mono-spécifique exprimant des conditions de vie hostiles (*in* Benabdelmoumen, 1997). Cette période est caractérisée par le dépôt d'une série monotone argilo-marneuse. Ce sont des dépôts à faunes pélagiques suggérant des apports d'eau de l'océan Atlantique. Ainsi, le bassin ne serait pas entièrement isolé. D'autre part, l'étude nouvelle de la microfaune du forage "Joides 121" fait apparaître un Messinien marin totalement dépourvu d'évaporites. Ceci est également le cas des bassins néogènes Bétiques et du bassin central de Vera (Almeria, Espagne) où le Messinien est marneux à foraminifères planctoniques (Montenat *et al.*, 1975).

La tectonique continue par le biais des accidents E-W uniquement. De ce fait, notre zone d'étude montre un remplissage sédimentaire du bassin vers l'ouest. Par contre à l'est, l'activité de ces failles est observable avec le ralentissement de la subsidence.

Au Plio-Quaternaire

La limite Miocène supérieure-Plio-quaternaire est caractérisée par le passage d'un épisode régressif vers un épisode transgressif. Ce passage est matérialisé par la discordance messinienne et l'évolution vers le milieu marin ouvert au Pliocène.

A ce moment, les communications avec l'Atlantique s'effectuent uniquement au travers du détroit de Gibraltar (Montenat, 1977). Notre bassin étant demeuré plus ou moins à l'écart de la crise de salinité messinienne, il est le siège d'une sédimentation mio-pliocène ininterrompue.

Les séries stratigraphiques pliocènes succèdent normalement aux séries miocènes supérieures comme cela est le cas pour d'autres bassins qui n'ont pas été le théâtre de la crise de salinité messinienne, tel que le bassin du Rharb (Maroc) et le bassin de Guadalquivir (Espagne) (Feiberg et Lorenz, 1977 et Viguier, 1974 *in* Montenat, 1977). Il est probable que la période messinienne correspondrait plutôt à un milieu à faible tranche d'eau qu'à un milieu émergé.

La tectonique continue à la faveur des accidents normaux E-W avec toujours des rejeux verticaux et une croissance verticale des bassins.

Au Pliocène supérieur, tout le pourtour de la mer d'Alboran enregistre une phase compressive qui apparaît également dans notre zone d'étude.

Relation entre les accidents Est-Ouest et le volcanisme

Les cartes en isochrones montrent des axes structuraux positifs (volcaniques) plio-quaternaires alignés selon une direction E-W en bordure du réseau d'accidents de Habibas. La nature volcanique de ces structures positives a été établie grâce à une parfaite corrélation avec des structures magnétiques ayant des valeurs élevées organisées selon la direction E-W.

La sismique montre que les horizons plio-quaternaires sont tronqués par ces remontées magmatiques qui sont parfois affleurantes (Benabdelmoumen, 1997). De plus, l'étude du volcanisme à terre, en Oranie Nord-occidentale (Algérie), a montré que les complexes volcaniques de Ain Temouchent et de Béni Saf correspondent à des alignements E-W d'appareils volcaniques d'âge plio-quaternaires (Megartsi, 1985; Abad, 1993). Cela corrobore l'âge plio-quaternaire proposé pour le volcanisme en mer.

Conclusions

La zone d'étude représente le domaine de transition entre le bassin d'Alboran et le bassin algéro-baléare. Elle s'inscrit au sein d'un plateau marginal qui constitue la continuité de la marge Nord Orientale marocaine.

Au Tortonien supérieur, les mouvements d'ouverture s'accélèrent, engendrant la structuration et l'installation des différents bassins tortoniens du domaine d'Alboran. Cette structuration s'effectue selon les accidents NE-SW décrochants senestres. Ce processus d'ouverture semble avoir fonctionné en un système de type "pull apart". Au niveau de ce bassin losangique, une subsidence tectonique active est assurée par les failles E-W dont le rôle majeur se retrouve à l'échelle du domaine d'Alboran, et cela jusqu'au Messinien inférieur. Cette phase transtensive est accompagnée par une transgression généralisée.

Au Messinien supérieur, une relaxation tectonique est observable; seuls les accidents E-W rejouent. Elle est accompagnée par un comblement sédimentaire du bassin. De plus, la sédimentation argileuse se maintient bien qu'une baisse eustatique ait eu lieu.

La transgression pliocène s'observe par le retour d'une sédimentation pélagique. Au Pliocène supérieur (Quaternaire ancien) s'installe un épisode compressif qui induit une inversion et un jeu décrochant dextre des accidents de direction E-W. C'est à ce moment également que s'est formé le bassin losangique de Yusuf.

BOOK IN STOCK

L'évolution géodynamique de la région au Tortonien supérieur montre un contexte général compressif lié à la convergence des plaques Europe-Afrique et au blocage du bloc d'Alboran et sa suture définitive à l'Afrique et à l'Ibérie. Dans cet environnement tectonique, les bassins tortoniens s'ouvrent en transtension.

Cette tectonique se poursuit jusqu'au Plio-Quaternaire où les accidents E-W décrochants dextres, entre autres l'accident Yusuf-Habibas, représentent des structures néotectoniques majeures dans cette région. L'accident régional Yusuf-Habibas a contribué à la mise en place du volcanisme plio-quaternaire observé sur les cartes structurales, aligné selon la direction E-W. Cette direction d'accidents se prolonge vers l'Est jusque dans le Tell (Cheliff) et à l'Ouest jusqu'au banc de Gorringe (Açores). Ce réseau fait partie d'une bande de déformation (GALTEL: Gorringe-Alboran-Rif-Tell) (Meghraoui *et al.*, 1996) qui redistribue la contrainte tectonique dans cette région.

BOOK IN STOCK

Paleozoic to Cenozoic crustal and basin structuring of the Tunisian north African margin

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The geological and geophysical data from the different regions of the Tunisian Atlassic and Mediterranean zones allow us to reconstruct the structuring and evolution of the crustal margin as well as its tectonic and sedimentary cover mechanisms. Reflection and refraction seismic investigations and gravimetric Bouguer and residual anomalies (Midassi, 1982; Abbès, 1983; Buness *et al.*, 1992; Bédir, 1995; Jellouli and Kevin, 2000) have highlighted the deep tectonic framework of the Tunisian margin. This domain is characterized by deep structural discontinuities trending north-south, east-west, northeast-southwest and northwest-southeast corresponding to deep-seated flower fault corridors (Bobier *et al.*, 1991; Ben Ferjani *et al.*, 1990; Bédir, 1995; Bédir, 1999). These discontinuities bound the main tectonic blocks of the margin. The Mohorovitchi discontinuity depth reaches from the south to the north and to the north-east from forty to twenty kilometres. The upper levels are respectively to the north at the Galite granitic Island latitude and to the north-east along the Tunisian-Sicilian Pantellaria and Linosa volcanic islands strait (Buness *et al.*, 1992). This rising of the Moho corresponds to the important magmatic activity in these zones since the Triassic until the Neogene times and the high geothermal rates encountered in the petroleum wells of these areas. From the south to the north the margin is subdivided in three main geotectonic blocks (Fig. 1): the Saharan platform bloc, the Atlassic domain with the meridional, the central and the northern Atlas and the eastern-northeastern Sahel-Pelagian block. From the Paleozoic to the Cenozoic times, these blocks evolved progressively from the southern Saharan continental domain to the northern Tunisian basinal trough area across a mosaic of tilted platforms and grabens.

The Saharan domain has a granitic and metamorphic basement (Laaridhi-Ouazaa, 1994) which sustains the Paleozoic and Mesozoic platforms (Busson, 1972). These platforms are cut by dominately E-W faults accompagned by magmatic basaltic rocks. Sedimentary deposits are mainly composed of continental sandstones, shales and carbonates with a Permian reef along the E-W Talemzane arch which bounds the Saharan platform from the northern Atlassic domain. This area is affected by the Caledonian and the Hercynian unconformities (Busson, 1972; Bouaziz, 1995).

The Atlassic domain is characterized by thick Mesozoic cover and thinning and lack of Tertiary deposits around the Upper Cretaceous and Paleogene “Kasserine Island“. The Triassic basal series are composed of shallow marine continental sandstones, shales, dolomites and thick evaporite and salt deposits which are engaged in halokinetic structures along the Meso-Cenozoic

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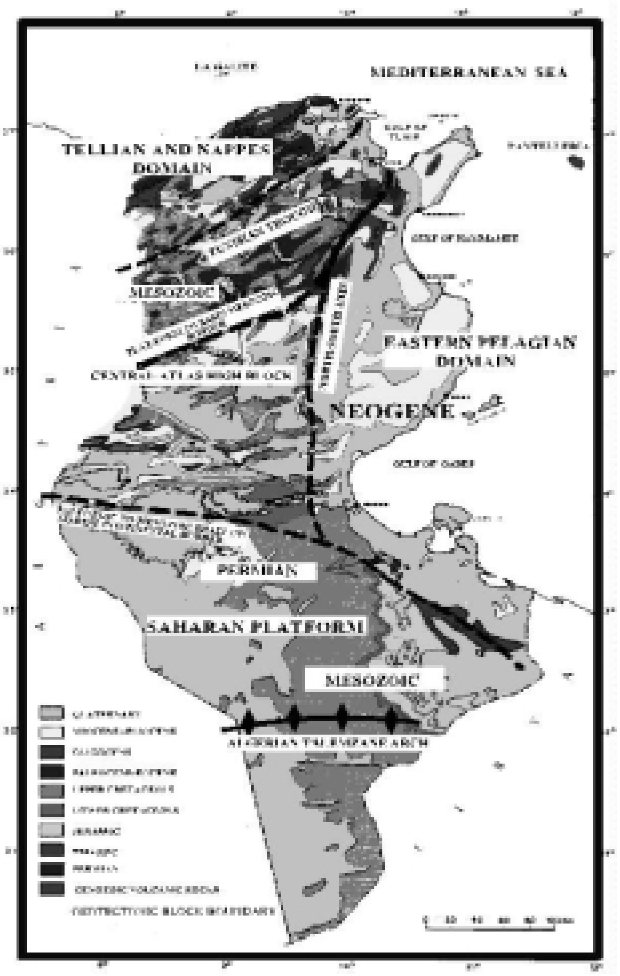

Fig. 1. Geotectonic Tunisian margin domains

cover. The structures are affected by the Tethyan rift along deep-seated E-W, N-S and NE-SW strike-slip faults inducing the formation of Triassic, Jurassic and Cretaceous graben and rim syncline basins bounded by rhombic platforms (Bédir, 1995; Soussi, 2000). These deposits consist of marine to shallow marine carbonates, shales, continental sandstones and evaporites. Since the Middle Cretaceous compression the basin-platform borders are transpressed and folded preparing the Atlassic fold domain. The northern Atlassic area comprised the Tunisian trough, which is the basin ward domain since the Jurassic according the meridional and central Atlas. This area is bounded to the north by the “Triassic Diapir zone” trending NE-SW. Mesozoic series became more thicken with deep marine facies of turbidites and reefs (Ben Ferjani *et al.*, 1990). The Thrust imbrication zone is composed of Upper Cretaceous to Paleogene and Miocene overthrusted units (Rouvier, 1977) with a 15 to 20 kilometers displacement toward the south-east. To the extreme north the Tellian zone is the most basinal area overlapped by the Oligo-Miocene Numedian nappes (Rouvier, 1977).

The eastern sahelian and pelagian area is affected by an important tectonic activity of transtensional and transpressional deep-seated faults that create a mosaic of Mesozoic and Cenozoic grabens, horsts, platforms and folds (Bédir, 1995, 1999). During the Mesozoic times, a marine to deep marine carbonate basin was form with platform deposits intruded by magmatic volcanic basaltic rocks A great subsidence occurs in the Tertiary where the horizons are composed by thick deposits of Paleocene marls, Eocene platform carbonates, deep-marine marls and black shales. The Oligocene consists on deltaic and fluvial sandstones deposits. Paleocene to Eocene series are interspaced by volcanic alcaline explosive rocks. Miocene deposits comprise deep-marine to thick deltaic series marked by an alternations of marls, platform carbonates and clay packages with sandstone turbidites. Burdigalian to Langhian series are interbedded in some offshore wells by basaltic rocks. To the offshore basins Pliocene deposits are formed by a thick deep marine marls and sandstone turbidites (Bédir *et al.*, 1996). These Neogene deposits are engaged in a claykinesis and clay intrusives along the strike-slip fault corridors (Bédir *et al.*, 1998). Quaternary Pleistocene and Holocene sandstones and Continental red beds and caliches overly the ancient series.

BOOK IN STOCK

Le canal de Sardaigne, à la croisée des bassins algéro-provençal et tyrrhénien de la Méditerranée et des segments kabylo-tunisien et siculo-calabrais de la chaîne des Maghrébides

Jean-Pierre Bouillin
et les participants aux campagnes SARCYA et SARTUCYA

Laboratoire de Géodynamique des Chaînes Alpines, UMR 5025 CNRS/ Université Joseph Fourier, Grenoble, France.

De Gibraltar jusqu'à Bizerte, la chaîne des Maghrébides se développe parallèlement à la côte africaine de la Méditerranée. L'élément le plus oriental de la chaîne, isolé, forme l'intérieur de l'arc Calabro-péloritain (Fig. 1). En position intermédiaire, un tronçon de la chaîne a été profondément submergé dans le canal de Sardaigne.

La morphologie du canal de Sardaigne (Fig. 1) présente des pentes abruptes qui ont permis d'y réaliser des dragages puis des observations et des prélèvements au moyen du submersible Cyana. Les roches récoltées sont analogues à celles des zones internes de la chaîne des Maghrébides dont les affleurements les plus proches sont ceux de Petite Kabylie, des Monts Péloritains en Sicile, et de Calabre (Compagnoni *et al.*, 1989). A terre, on observe un socle de roches métamorphiques et de granites hercyniens recouvert par des grés et conglomérats discordants d'âge Oligocène supérieur à Burdigalien inférieur. Cet "Oligo-Miocène kabylo-péloritain" est surmonté par des olistostromes remaniant des flyschs crétacés-paléogènes et par des panneaux de grès numidiens. Des grès et marnes d'âge langhien reposent sur cet ensemble. Les observations en plongée et la répartition des échantillons récoltés permettent de reconstituer une disposition analogue sur les pentes sud du canal de Sardaigne et sur la Ride Médiane (Fig. 1).

Les observations en plongée (équipes des campagnes SARCYA et SARTUCYA, sous la responsabilité de G. Mascle, P. Tricart et L. Torelli) associées aux analyses pétrographiques (F. Rolfo), géochimiques (H. Lapierre) et radiochronologiques (P. Monié, G. Poupeau, H. Bellon) des échantillons ont permis de préciser l'évolution géodynamique du canal de Sardaigne.

On constate que le socle y est sensiblement moins déformé que celui des massifs de Petite Kabylie, des Monts Péloritains et de Calabre. Par ailleurs les mesures $^{40}Ar/^{39}Ar$ montrent que seuls les minéraux des échantillons situés vers le SE, le long de l'escarpement Sud-Cornaglia, ont subi des réouvertures partielles à l'Eocène supérieur-Oligocène (Bouillin *et al.*, 1999). Cela signifie que le socle du canal de Sardaigne occupait une position haute dans l'édifice de la chaîne ou/et qu'il était situé en arrière de la zone la plus déformée. La transition progressive entre le socle de la chaîne des Maghrébides et le socle sarde s'accorde avec l'interprétation paléogéographique qui replace, avant l'orogenèse alpine, les zones internes maghrébides sur la bordure européenne du bassin téthysien maghrébin.

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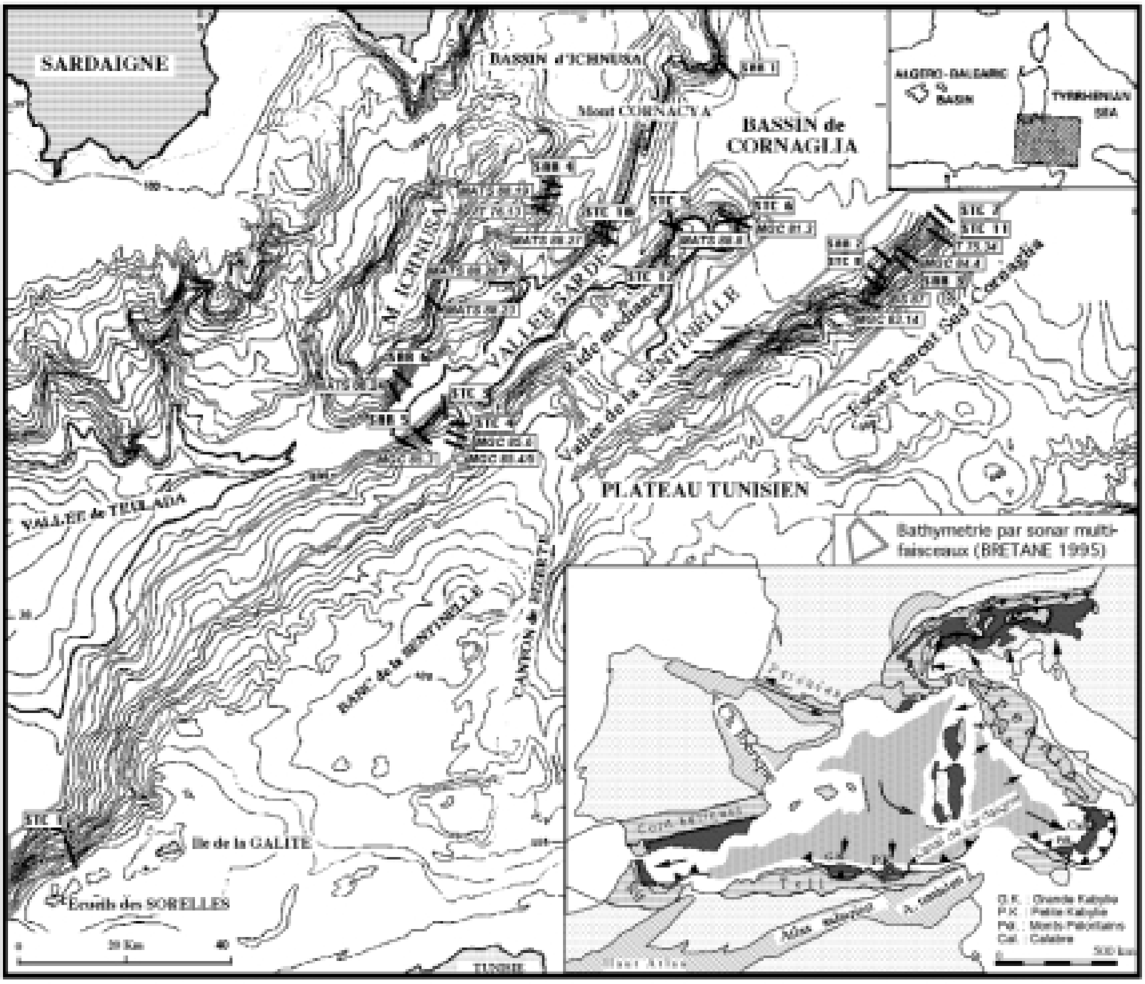

LOCALISATION DES SITES DRAGUEES ET DES PLONGEES dans le Canal de Sardaigne

SAR : plongées SARCYA (1994)
STC : plongées SARTUCYA (1995)

T.76, MGC.85, BS.87, MATS.88 : sites dragués (d'après Compagnoni et al., 1989).

Fig. 1.

A l'Oligocène, du fait de la subduction de la Téthys sous la marge européenne, puis de la collision avec l'Afrique, le bloc formé par les Kabylies, les Monts Péloritains, la Calabre et leur arrière-pays sarde a chevauché le bassin maghrébin. Le socle chevauchant a lui-même été découpé par des surfaces à faible pendage, que l'on peut observer à terre. Les réflecteurs sismiques, faiblement pentés vers le NW, visibles dans le socle du canal de Sardaigne ont été interprétés comme de tels contacts (Fig. 2; Tricart *et al.*, 1994).

A la fin de l'Oligocène, la bordure méridionale des massifs de socle a émergé tandis que les zones plus internes, jusqu'alors exondées et profondément érodées, ont été envahies par les dépôts détritiques de l'Oligo-Miocène kabylo-péloritain. Des failles en extension, déterminant des blocs basculés, sont associées à cet effondrement (Kézirian *et al.*, 1994). La dénudation du socle calabro-péloritain au cours de l'Oligocène et du Miocène inférieur a été datée en Calabre et en Sicile par la méthode des traces de fission (Thomson, 1994). Un échantillon du versant sud de la vallée sarde, daté de 22,8 Ma. par la même méthode, montre que le socle du canal de Sardaigne a subi la même dénudation. Les mylonites en extension qui affectent le socle de l'Aspromonte vers 25-30 Ma (Platt et Compagnoni, 1990) sont probablement l'expression profonde de cet épisode de distension. Certaines des surfaces observées sur les profils sismiques du canal de Sardaigne (Fig. 2) peuvent être interprétées comme de tels détachements. A partir de l'Oligocène terminal, les massifs kabylo-calabro-péloritains se seraient donc détachés de la Sardaigne. Ils en auraient été séparés par un bassin marin, probablement étroit, mais qui s'ouvrait plus largement vers l'Ouest. On peut penser que l'essentiel de la structure du canal de Sardaigne a été acquise dès cette étape.

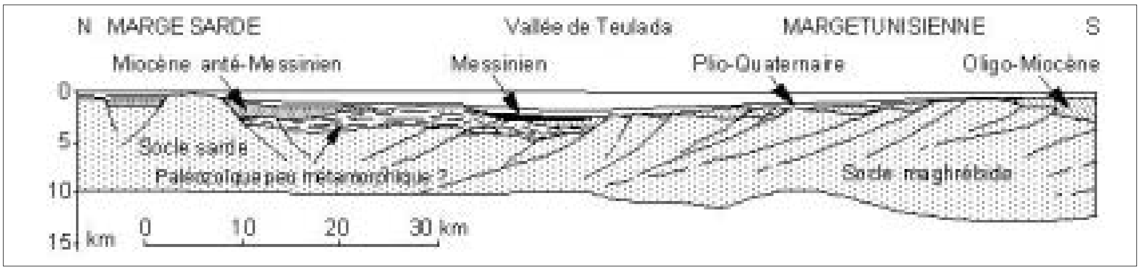

Fig. 2. Coupe schématique du canal de Sardaigne d'après les données sismiques, d'après Tricard *et al*., 1994.

Le bassin Oligo-Miocène s'est ensuite approfondi jusqu'au Burdigalien inférieur, époque à laquelle il a recueilli des olistostromes et des nappes gravitaires à matériel de flysch provenant des zones plus externes de l'orogène vers lesquelles la compression s'est déplacée.

Le volcanisme du Mont Cornacya (Fig. 1), découvert lors des plongées SARCYA, a marqué le début d'une nouvelle étape de la structuration du canal de Sardaigne et de ses abords. Ce sont des andésites, datées de 12,6 Ma par la méthode $^{40}Ar/^{39}Ar$ et apparentées à une suite shoshonitique. Elles incluent des xénolithes lamprophyriques. Cette association indique une mise en place en contexte post-collisionnel (Mascle *et al.*, 2001). On connaît à Sisco, dans le Cap Corse, un volcanisme de même nature géochimique et de même âge, situé dans une position analogue en bordure orientale de la mer Tyrrhénienne. Ce volcanisme marque le début de la formation du bassin Tyrrhénien qui s'est ouvert de l'Ouest vers l'Est, en arrière arc de la subduction apenninique et calabro-péloritaine.

Cette ouverture a disloqué la chaîne des Maghrébides en entraînant loin de la Sardaigne et des Kabylies, les massifs péloritain et calabrais. Pour sa part, le socle du canal est resté attaché à la Sardaigne. Il a subi néanmoins une nouvelle extension. Les traces de fission dans l'apatite montrent que le socle des versants de la vallée sarde et de l'escarpement Sud-Cornaglia a été exhumé entre 10,3 et 8,3 Ma, probablement par le jeu de failles normales (Bouillin *et al.*, 1998).

La bathymétrie par sonar multifaisceau (campagne BRETANE, 1995) et les observations en plongée montrent, surimposée à des escarpements correspondant à des failles, une morphologie de détail modelée par une forte érosion fluviatile et par des mouvements gravitaires. Cette érosion est attribuable à la crise messinienne.

Enfin les observations en plongée ont montré des indices en faveur d'une reprise en compression récente des structures, bien que l'activité sismique actuelle soit très faible (Brocard, 1998).

En conclusion, le canal de Sardaigne a enregistré deux histoires distensives successives. Il s'est formé à l'Oligocène supérieur-Miocène inférieur, en constituant la partie orientale du bassin nord-algérien, en arrière de la chaîne des Maghrébides. Dans un second temps, à partir du Tortonien, il a été impliqué dans l'ouverture du bassin Tyrrhénien, en arrière de l'arc calabro-péloritain.

BOOK IN STOCK

Cenozoic collisional and extensional structures among Sardinia, Sicily and Tunisia (Central Mediterranean): examples and constraints from seismic reflection profiles

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Introduction

Within the Central Mediterranean the geology of the south-western Tyrrhenian Sea and of the western end of the Strait of Sicily is still relatively poorly known. However, this marine area presents several aspects that may be relevant to the comprehension of the geologic evolution of the entire Central Mediterranean.

Among the aspects worthy of interest are the following: i) the relationship between the Tunisian and the Sicilian segments of the Maghrebian fold-and-thrust belt, ii) the distribution of the Kabylo-Calabrian terranes, iii) the collision of the Sardinia-Corsica block, during its Neogene rotation, with the African margin, and iv) the nature and evolution of extensional tectonics in the Tyrrhenian basin, in the Strait of Sicily and in the Sardinia Channel, where the previous two areas merge.

In order to understand the geology of the south-western Tyrrhenian Sea/western Strait of Sicily, the acquisition of deep seismic profiles and a reprocessing of existing seismic data were undertaken. In addition sampling and Well data available along the seismic profiles have been taken in account. The results are here summarized as a geologic cross section derived from one of these crustal scale seismic profiles (Fig. 1).

Geologic setting

The study area (inset of Fig 1) presents a complex arrangements of tectonic units derived from both the European (internal massifs) and African domains. Its evolution was accomplished mostly during the Tertiary in relation to the the collision between Africa and Europe.

In this area, as in the whole of the Mediterranean basin, contraction and extension often occurred at the same time. In particular, the opening of the Balearic basin, from Oligocene to Early Miocene, and the ensuing rotation of Corsica-Sardinia, gave rise to a southeastward propagating fold-and-thrust belt that involved internal terranes (Kabylo-Calabrian units) and part of the African continental margin. Apparently, the shortening continued after the rotation of Corsica-Sardinia stopped.

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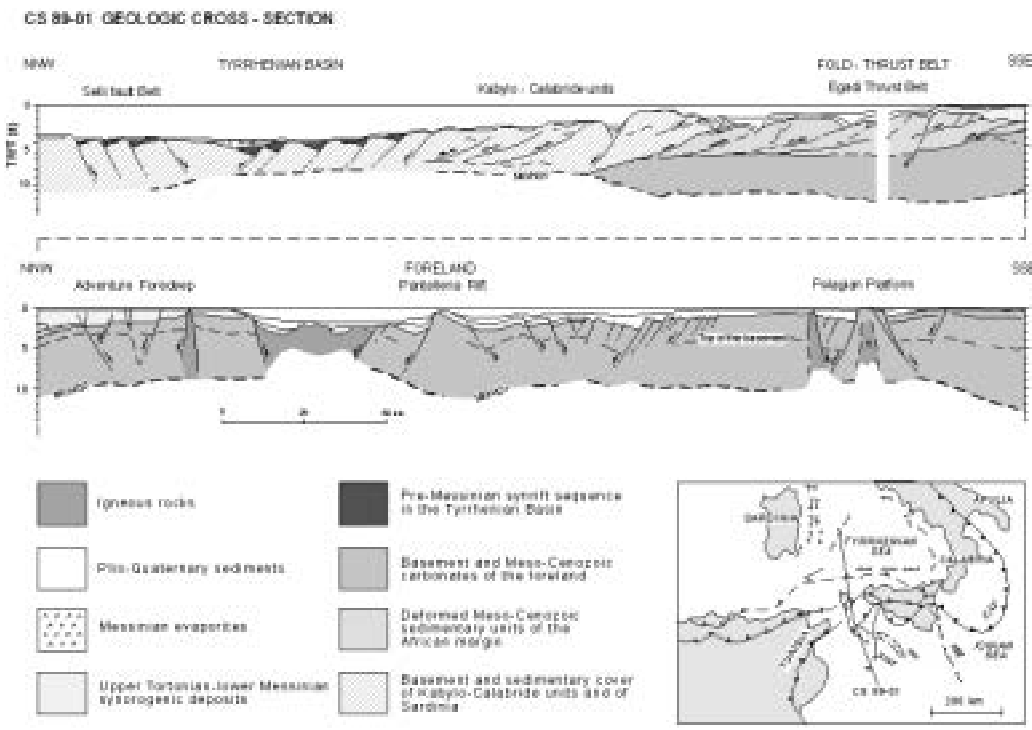

Fig.1. Geologic cross-section from the foreland to the hinterland in the Central Mediterranean

The Adventure basin, filled by Tortonian-lower Messinian sediments, is the youngest foredeep in the western Strait of Sicily and, although it was deformed contractionally in the Messinian, it represents the last major episode of shortening occurred in the area. However, shortening continued to operate farther to the east until Early Pleistocene, giving rise to the Plio-Quaternary Gela foredeep. The sense of thrusting in the fold-and-thrust belt has been shown to rotate in time from southeast in the Late Miocene to south in the Pliocene.

The onset of Tyrrhenian extension occurred in the Tortonian; such extension propagated southeastward in time until the Early Pleistocene and affected also, to some extent, the internal part of the fold-and-thrust belt.

The Strait of SiciIy rift zone can also be related to this episode of extension and can be seen as a foreland response to the opening of the Tyrrhenian backarc basin.

RESULTS

The seismic profile here illustrated (Fig. 1) crosses all the domains present in the area; namely, the foreland, that in this case is affected by extensional tectonics (Pantelleria rift zone), the fold-and-thrust belt with associated foredeep (Egadi fold-thrust belt and Adventure foredeep), and the extensional margin of the Tyrrhenian backarc basin. For each of these domains a crustal-scale image is obtained.

A geologic cross-section has been constructed integrating the deep seismic data with data from conventional seismic, welI logs and dredge hauls. Three major domains are represented in the cross section: i) African foreland, ii) fold-and-thrust belt, and iii) Tyrrhenian backarc basin.

The African foreland is affected by extensional tectonics of Pliocene age that shows up as fault systems the most prominent of which is the Pantelleria rift system. The master faults are rather steep and bound slightly rotated blocks indicating that extension was Iimited. Along some of the faults magmatic bodies were intruded. The maximum amount of volcanics, as interpreted from seismic facies, appears to occur within the Pantelleria trough. The package of reflections attributed to the Moho decreases in TWT underneath the Pantelleria trough where the maximum crustal thinning is also indlcated by refraction data.

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The domain of the fold-and-thrust belt can be further subdivided into three sub-domains: i) the Adventure foredeep, ii) the Egadi fold-and-thrust belt, and iii) the Kabylo-Calabrian units.

The foredeep basin is filled with Serravallian-lower Messinian sediments that were later shortened and partially detached from their substrate during the Messinian. Towards the northern shoulder of Pantelleria trough some extensional faults, likely related to the above mentioned rifting, cut the foredeep sediments and along a few of them volcanic bodies were emplaced.

In the Egadi fold-and-thrust belt a few shallow-dipping south-facing reflections can be observed. They are Interpreted as thrust faults bounding the units of the African continental margin piled up southward during the Miocene.

Farther to the north, the terranes piled up belong to the Kabylo-Calabrian units. Following the contraction, some of these thrust faults were reutilised in extension and gave rise to basins fiIled with Plio-Quaternary sediments. Roughly at the boundary between the African and the Kabylo-Calbrian units one of these Plio-Quaternary basin shows contractional structures. Such structures are thought to be indicative of a recent episode of inversion tectonics.

In the area of the Kabylo-Calabrian units the reflectivity is very poor; however, short south-dipping reflections can be observed. These reflections are interpreted as defining rotated fault blocks originated during the opening of the Tyrrhenlan basin.

This structural style becomes more evident moving towards the Tyrrhenian basin domain where fault blocks and also syn-rift sediments can be recognised along the southern Tyrrhenian slope. In the Tyrrhenian plain it is not possible to define a particular structural style but a depocentre of Messinian evaporites shows up quite weIl. At the very end of the cross section some faulted blocks, tilted northward, are bounded to the north by the Selli Fault. lt is worthy of note that in the Tyrrhenian domain the PIio-Quaternary sediments, and in the Tyrrhenian plain even the Messinian evaporites, are not affected by extensional faulting. Therefore, in this part of the Tyrrhenian basin, the stretching occurred mostly in the early phase of opening. The PIio-Quaternary extensional basins superposed onto the fold-and-thrust belt prove that extension migrated southward in time (In some way keeping up with the southward migration of the thrust front).

Using this cross-section and other deep regional seismic images we try to reconstruct the Neogene kinematics of this sector of the Central Mediterranean, comparing the main deformative events of the area with the tectonic and sedimentary evolution of adjacent regions, namely the Kabylo-Calabrian and Maghrebian units in Sicily and Tunisia.

BOOK IN STOCK

Overview on the stratigraphy, paleogeography and structural evolution in southern Tunisia, during the Mesozoic, as part of the African margin

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Etudes explorations, Tunis, Tunisie

During the Mesozoic, the Jeffara and the Chotts trough formed a linking between the stable Saharan platform and the active alpine domain. So, the lithostratigraphic columns have recorded the main paleogeographic and structural events. The sedimentation patterns were under control of local and regional tectonic events and the sea level changes. Thanks to its economic interests and the outcrop's good exposure, the area is one of the most studied. Renewed studies start in the beginning of this century.

In the Sahara, several Paleozoic and Mesozoic targets are explored for hydrocarbon as Ordovician, Silurian, Devonian, Triassic and Jurassic and where non neglect amount of oil and gas is discovered and produced, until now. Into the Chotts, the phosphates are intensely produced from Paleocene and Eocene series.

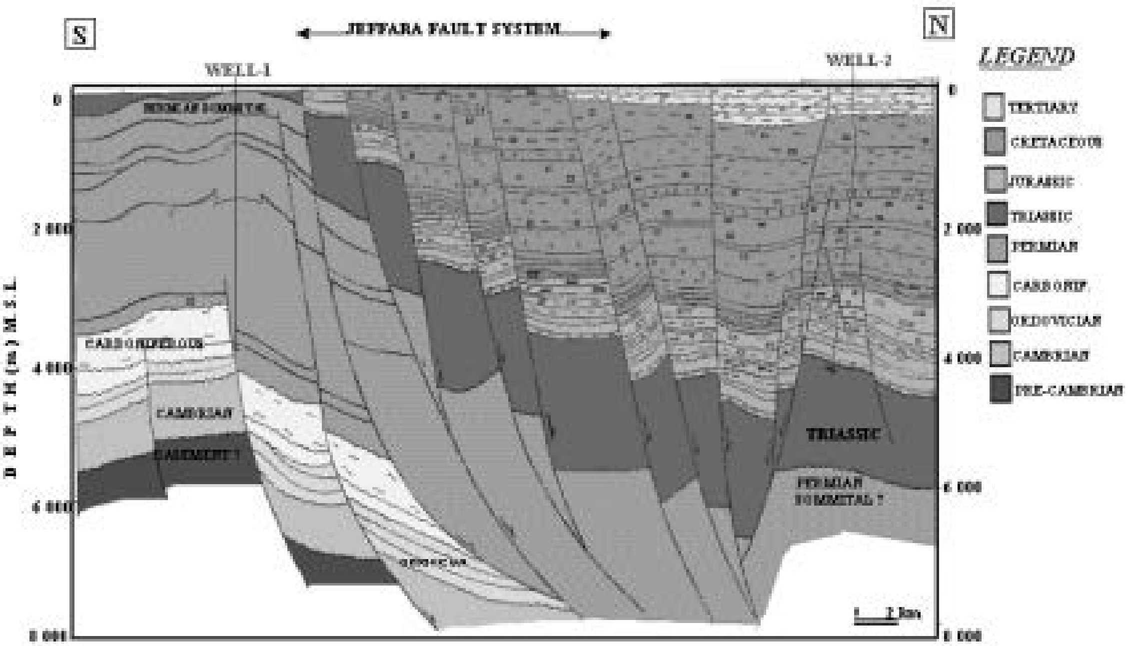

Fig. 1. Geoseismic Cross-section illustrating the active subsidence in the Jeffara area during the Mesozoic time in contrast with the Saharan platform resistant domain.

BOOK IN STOCK

In the following, we will summarise an overview of the geologic history of the area during the Mesozoic. Particularly, the Triassic dynamic deposition, in the Jeffara and Tripolitain basins will be treated.

The Hercynian Orogeny affected severely the area. During Late Carboniferous to Early Triassic, southern Tunisia was entirely uplifted and sharply eroded, except the eastern part of the Jeffara and partially the Chotts area, where virtual continued Triassic series are encountered.

The Lower and Middle Triassic marine sediments, predominantly siliciclastic were restricted to the Jeffara province and partially into the Chotts. Firstly, the Triassic basin matched the inherited Permian basin.

The Upper Triassic siliciclastic, carbonates and evaporites series extended progressively and widely into the Saharan platform and the Chotts basin.

Jurassic tectonism, by reactivation of Paleozoic normal and transtensional faults led to the rifting of the Chotts. The Chotts trough was formed and experienced as a pronounced subsidence, particularly in the Middle and Upper Jurassic. Southward the Chotts trough, extended a resistant domain.

During Liassic and Middle Jurassic PP, evaporitic conditions were established into the Saharan platform, while in the Chotts, the marine limestone and marls were deposited .

The Pliensbachian transient transgression interrupted momentary the evaporitic conditions and permitted the deposition of 15m of marine oolithic carbonates over the Saharan Platform.

During Upper Jurassic, occurred the main transgression. The resistant domain was covered by marine deposits :fossiliferous limestone (locally reef mound), sandy or oolithic carbonates, associated with sandstone and shales extending widely.

The major Cretaceous regional tectonic event was the effect of the Austrian Orogeny expressed by obvious unconformity. The Austrian unconformity was developed through the region. In fact, the Hiatus of Lower Albian is noted largely; the Upper Albian sets, unconformably, on the Aptian carbonate unit. This unconformity, combined to locally uplifts, led to the erosion and hiatus of large series as in the Melab Paleohigh where Cretaceous is overlying the Paleozoic.

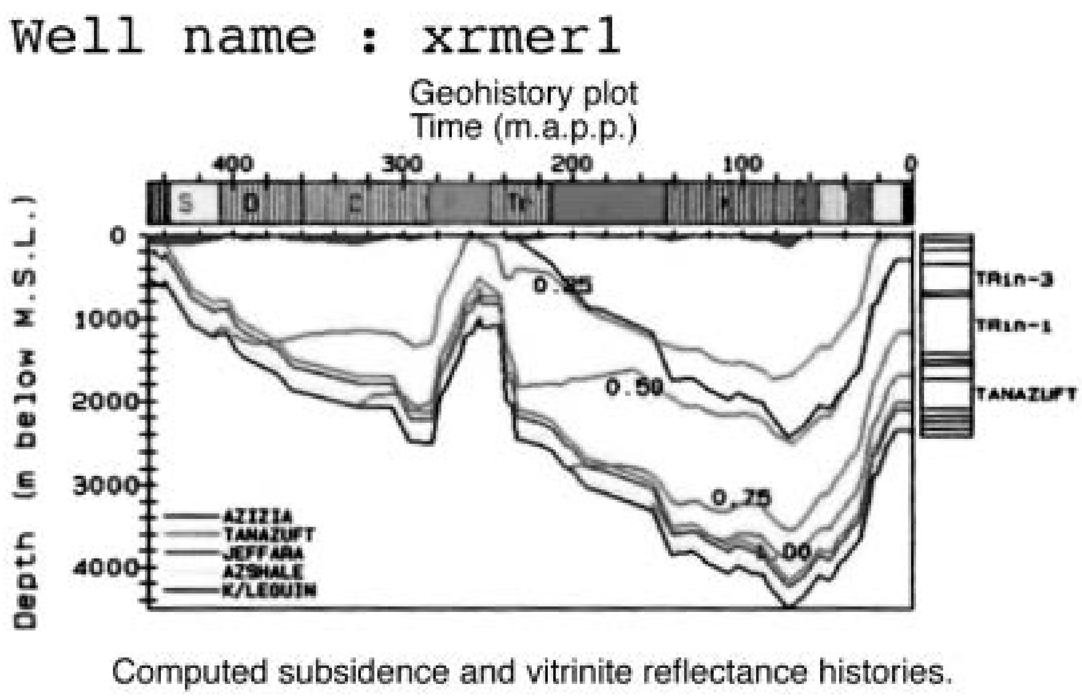

Fig. 2. Geohistory Plot of MER-1 well showing Hercynian Orogeny, computed subsidence and hydrocarbon generation.

BOOK IN STOCK

During Neocomian time, siliciclastics and evaporites transgressed over all the domains. The Chotts trough was compensated and siliciclastics crossed further north. The troughs transferred further north.

The Aptian was a transgressive period; it is represented by distinguished carbonates extending widely into the Saharan platform.

In the Late Albian and Late Cretaceous, marine transgressions succeeded alternating with relatively short regressions. The marine conditions prevailed upon into the Saharan platform.

Much of the southern Tunisia was remained emerged since the Paleocene, although, there was mild subsidence and deposition of Neogene sediments in the Chotts area and Jeffara maritime. Into the Gafsa basin were deposited a thick series of shales, intercalated with phosphates, evaporites and fossiliferous limestone.

Alpine, Late Miocene and Pliocene compression caused substantial uplift and folding in the Chotts basin with strike-slip movements. Extensional faults developed in the Jeffara at this time and are interpreted as transtensional features.

BOOK IN STOCK

Crustal structure of the Libyan margin

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INTRODUCTION

The Eastern Mediterranean sea is characterized by high seismicity and a complex tectonization which is not fully understood yet. Several geodynamic models have been developed to explain the processes in this region (Makris, 1976; Le Pichon *et al*., 1982; Mc Kenzie, 1970 and 1972). Still the subduction process south of Crete, the crustal structure below the Mediterranean Ridge and the deep structure of the north African passive continental margin remain poorly understood. The geophysical data – especially for the African margin – are limited to potential fields and some industrial reflection seismic lines, while deep seismic soundings are either very old or completely missing.

From reflection seismic experiments (Chaumillon *et al.*, 1996) at the southern edge of the Mediterranean Ridge (MR) a backthrusting tectonic structure in the sediments was identified in a southward direction. This confirms the MR as an accretionary complex (Le Pichon *et al.*, 1982; Ryan *et al.*, 1982; Mascle *et al.*, 1995), but it is not yet clear if and how far this backthrusting reaches onto the continental margin. Furthermore the question arises about how far the extant compression affects a tectonization of the African margin and of the crust itself.

To investigate questions and better understand the geodynamic processes in the Eastern Mediterranean region two active seismic studies south of Crete and the Mediterranean Ridge were carried out. We used densely spaced Ocean Bottom Seismographs (OBS) with distances of 3.5 to 5 km and shot a 48 lt. airgun array at 120m intervals. Two profiles extended to the Libyan margin of Africa and mapped its crustal structure. The locations of both lines are indicated in figure 1, where the bathymetric-topographic features demonstrate the complexity and lateral variability of the area.

PRELIMINARY RESULTS

The data were combined to common receiver point gathers (CRP) and corrections derived from navigation and bathymetric data were taken into account. For both lines about 130 CRP sections could be used for evaluation. The data were evaluated with forward modelling by using two point raytracing of traveltimes and amplitudes (Cerveny and Psencik, 1983) For each line we generated a 2D P-wave velocity depth model which resolved the crustal structures along these lines. We identified by different velocities at least four sedimentary layers, upper and lower continental crust and oceanic crust.

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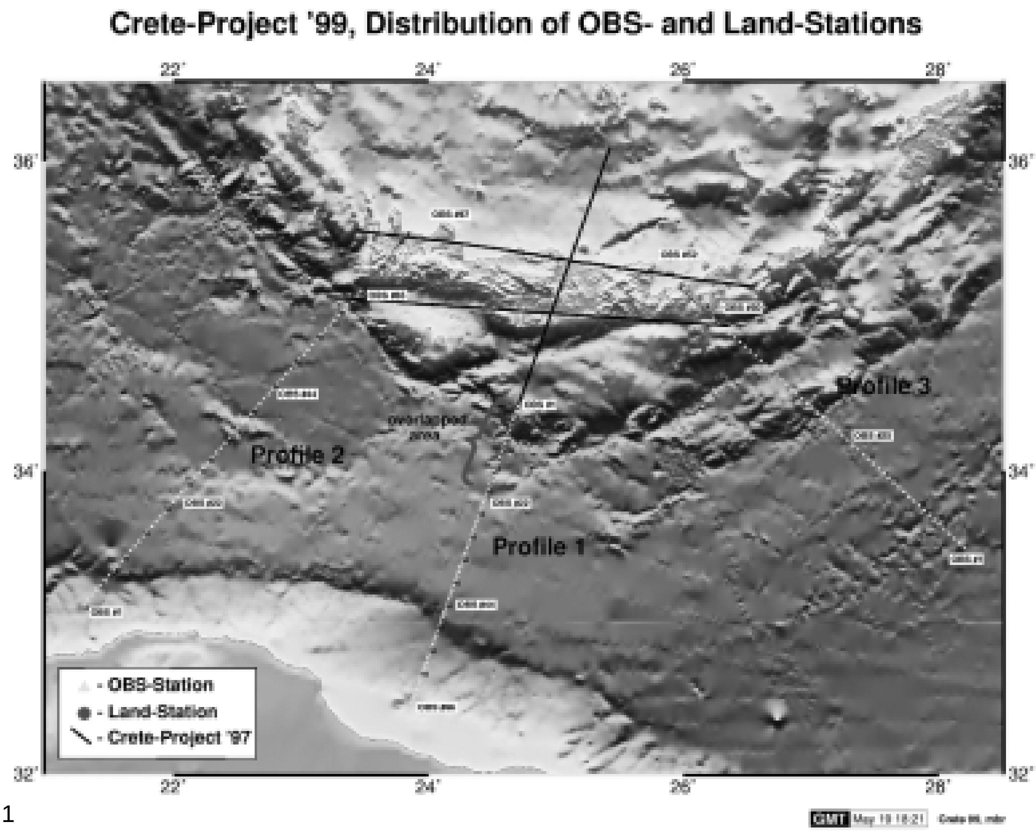

Fig. 1

Profile 1

The first line is N-S oriented and extends from the Cretan continental margin to Tobruk in Libya. The sea bottom varies on this profile from approx. 900 m on the African margin to more then 3400 m depth at the Hellenic Arc. In the middle of the profile – at the Mediterranean Ridge – an almost constant depth of 1800 m is observed. Along the line 66 OBS were deployed with a spacing of 3.5 km, and 2258 airgun shots were fired. For evaluation and computation of a 2D velocity depth model along this line (Fig. 2), 53 CRP sections were applied.

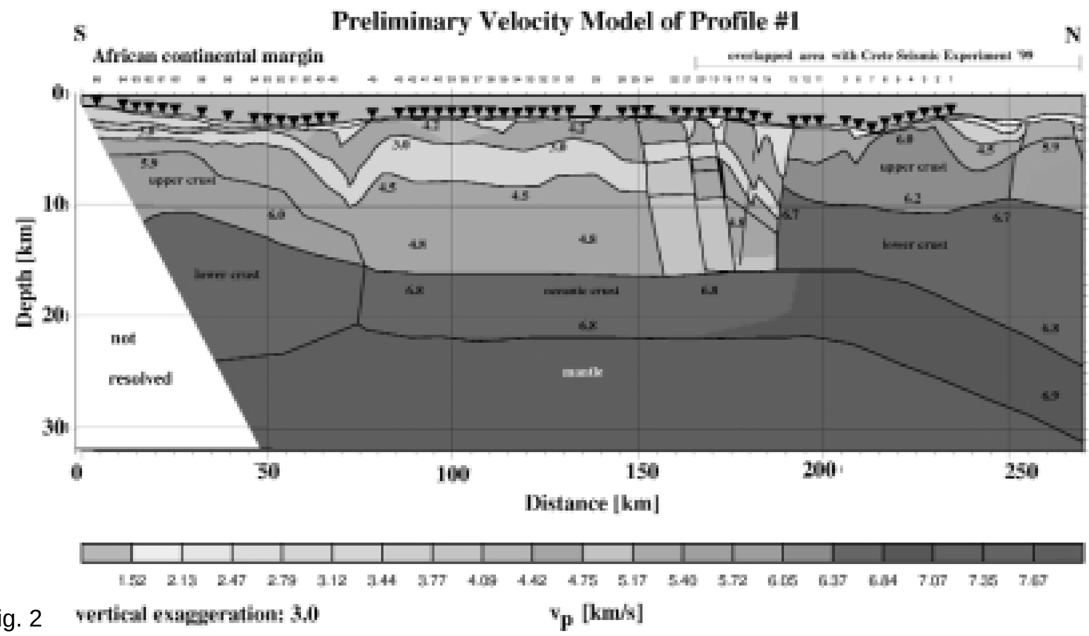

Fig. 2

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The African passive continental margin extends to nearly 90 km offshore the coastal line and has an abrupt transition to an oceanic crust buried under 12 to 14 km of Mesozoic and Neogene sediments. The sediments on the Libyan margin along this profile are 4 to 5 km thick and they have at least one major velocity inversion zone. They are depressed to the continent-ocean margin where their thickness is nearly doubled and where they show backthrusting to the south. The depth of water here increases to some 3000 m. The sediments and the crust at this part of the margin are not severely deformed, until well off the transitional area. The continental crust is 23 km thick and thickens towards the coast to a value of approx. 28 km.

Profile 2

The second line of NE-SW orientation extends from western Crete towards Bengazi and the Sirte basin. Along this seismic profile strong variations in bathymetry and topography are observed. Onshore Crete the altitude ranges between approx. 20 m and 870 m, south of Crete the depth of water increases rapidly to approx. 3700 m in the Hellenic Arc. Towards the south the seafloor depth ranges between 2500 m and 2000 m and decreases to approx. 1400m towards the African margin. Sixty seven OBS and 17 landstations were deployed. The spacing between the OBS was approx. 4.2 km, the distance between landstations was about 2 km, and 2852 offshore shots were fired. For evaluation and computation of a 2D velocity depth model along this line (Fig. 3), 80 CRP sections were used.

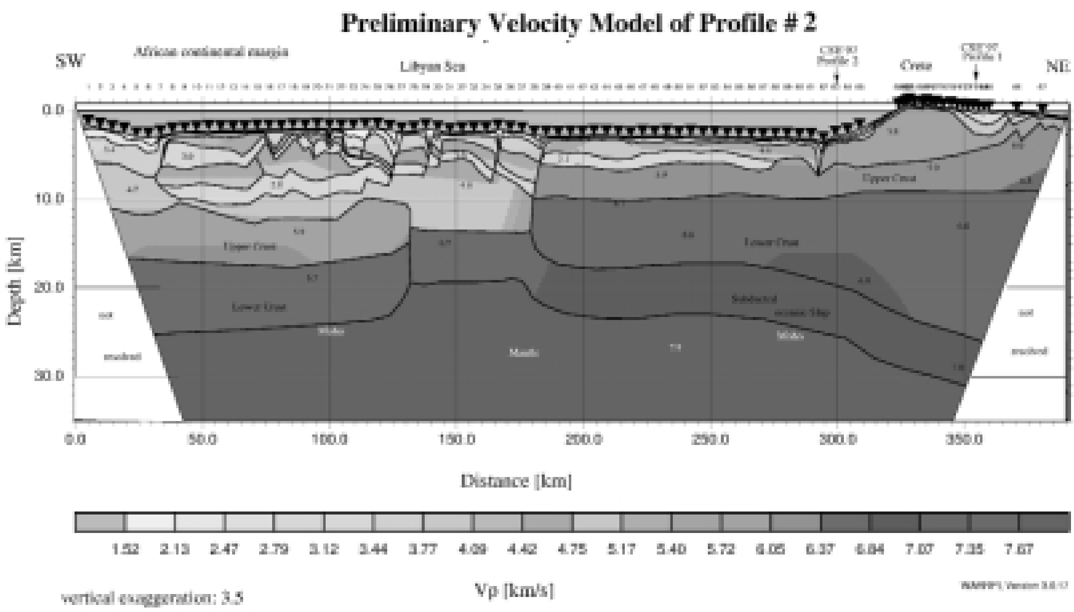

Fig. 3

The continental structure of this area extends at least to 160 km off the Libyan coast and is severely tectonized. Towards the coast a well developed basin was identified following the bathymetric depression. It is limited to the northeast by a major fault. The crust is 22 to 25 km thick including 6 to 7 km sediments under a 2 to 2,5 km waterdepth. Particularly intense is the tectonization of the structures at the continent-ocean margin. Two inversions of the v_p-velocities were identified, demonstrating the complexity and extended history of this passive margin. The tectonic structures in the sediments are dominated by large and inordinate sequences of overthrust tectonics towards the south, that becomes chaotic at the transition. The following oceanic structure is of limited extend, not more than 50 km wide and covered by 12 to 13 km of strongly tectonized sediments. Subsequently it is in contact to the European continental margin southwest of Crete. The identification of the collision of the African-European continental margins is most advanced in this area, where crustal shortening should produce a continent-continent collision in the near geological future. The complexity and lateral variability of the Mediterranean margin of north Africa are demonstrated in a simplified way for the Libyan coastal areas.

CONCLUSIONS

The crustal structure of the African margin as it is identified along profile 1 and profile 2 is extremely different. While the passive continental margin at profile 2 shows thick and strong tectonized units of sediments lying on thin continental crust, the African margin – basement and sediments – along profile 1 is barely affected by faulting. This changes abrupt at the transition where huge packages of sediments start to be backthrusted towards the African margin. Furthermore at profile 2 we identified two layers with velocity inversion in contrast to only one at profile 1. The upper one can be explained by weak sediments covered by Messinian salt. It can be observed on both profiles up to the European continental margin south of Crete. An explanation for the second, deeper low velocity layer at profile 2, might be given by tectonic deformations and broad overthrusting which is a consequence of continuous crustal shortening. Therefore thicker and stronger tectonized sediments along profile 2 may indicate a more advanced compressional process between Africa and Europe on the western profile, which would also explain the smaller relic of oceanic lithosphere between the African and European continental margins.

Although the crustal structure of the Libyan margin is mapped quite well along these two profiles, a correlation and interpretation of the evolution of the African margin remain highly speculative until more deep seismic studies are performed by active on-offshore experiments.

So far P-wave traveltimes were taken into account to generate present models. A refinement in resolution and a further confirmation of the models will be accomplished by evaluation of converted waves and gravity data.

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Evidences for tectonic reactivation along the African continental margins from Egypt to Libya

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ABSTRACT

Several areas of the Mediterranean continental margins of Africa, from Egypt to Libya, have recorded, and are still recording, regional tectonic events, which may have strongly disrupted their original structures, created in Mesozoic times during one, or two, rifting episods.

In this paper we focus on three regions, off Egypt and Libya, which have been recently reactivated. These are respectively : (1) the Erathostenes Seamount, a continental block, previously rifted away from Africa mainland and now entering in collision with the Cyprus arc; (2) the Eastern Nile Deep Sea Fan, beneath the thick recent cover of which a northern, may be still active, segment of the Suez Rift system is inferred; finally, (3) most of the continental slope north of Cyrenaica (Libya), where various structural effects of its incipient collision with the Mediterranean Ridge, and potentially with the southern border of the Anatolia-Aegean microplate, are initiating. Most of this presentation is based on a set of marine geophysical data recorded in 1998, aboard the R.V. *l'Atalante*, during PRISMED II survey.

INTRODUCTION

We present and discuss here evidences of tectonic reactivations that have affected several areas of the African continental margin within the Eastern Mediterranean Sea, at different times.

Most of the data, on which this presentation is based, have recently been gathered during a marine geophysical survey, PRISMED II, conducted over large areas of the Nile Deep Sea Fan (Nile DSF) and of the Mediterranean Ridge (MR), between Crete and Libya. The concerned regions are respectively : (a) the Erathostenes Seamount (ES), a continental block, formerly part of Africa, now half-way between northern Egypt and Cyprus, (b) an elongated, N-W trending, and thickly sedimented, region of the Eastern Nile DSF, strongly affected by both thin-skinned salt tectonic and thick-skinned tectonics and, finally, (c) along the Libyan continental margin, a wide segment of continental slope (off Cyrenaica promontory), which faces the overlapping southern border of the MR.

Along southern Eastern Mediterranean, the African passive margins are thought to have been created in Mesozoic times, in response to rifting processes whose exact timing remains a matter of debate, but that have led to the creation of an oceanic space located between southern Eurasia and Africa , the Eastern Mediterranean Sea (Guiraud and Bosworth, 1999). For some authors the oceanic crust of the deep Eastern Mediterranean basins, as well as of the Ionian Sea, may be as

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old as Triassic, according to onshore geologic data which support a first rifting episod, sometimes between Permian and early Trias (Guiraud and Bosworth, 1999). For others, the main rifting phase has occurred chiefly in Jurassic times (Robertson, 1998a), leading to the creation of an oceanic space whose remains are now consumed, as a consequence of the subduction of the African plate beneath southern Europe, recently combined with a south-eastward-directed motion of the Aegean-Anatolia microplate.

Whatever the exact rifting periods and timing during which continental break up took place, and dispite sparse geophysical supports, it is well admitted, that large areas of the deep Ionian sea as well of the Levantine basin, are floored by old oceanic crust remains, whose bordering, northern active and southern Mesozoic passive, margins are progressively entering collision south of Crete (Mascle *et al.*, 1999).

Off Africa, some of these margins segments have, however, been reactivated at different periods, and following different processes.

The Erathostenes Seamount (Fig. 1) is, for example, experiencing, since post-Miocene, the results of its progressive collision with Cyprus arc (Robertson, 1998b). The now thickly sedimented Eastern Nile DSF (Fig. 1) cover a probable submerged northward extension of the Suez-Red Sea system, which initiated some times in early Miocene (Mascle *et al.*, 2000) and, finally, the continental margin that lies north of Cyrenaica (Fig. 2) is submitted to deformations likely related to, both, the in progress southward thrusting of the MR, and a probable indenting of the Cyrenaica promontory against the Crete continental margin.

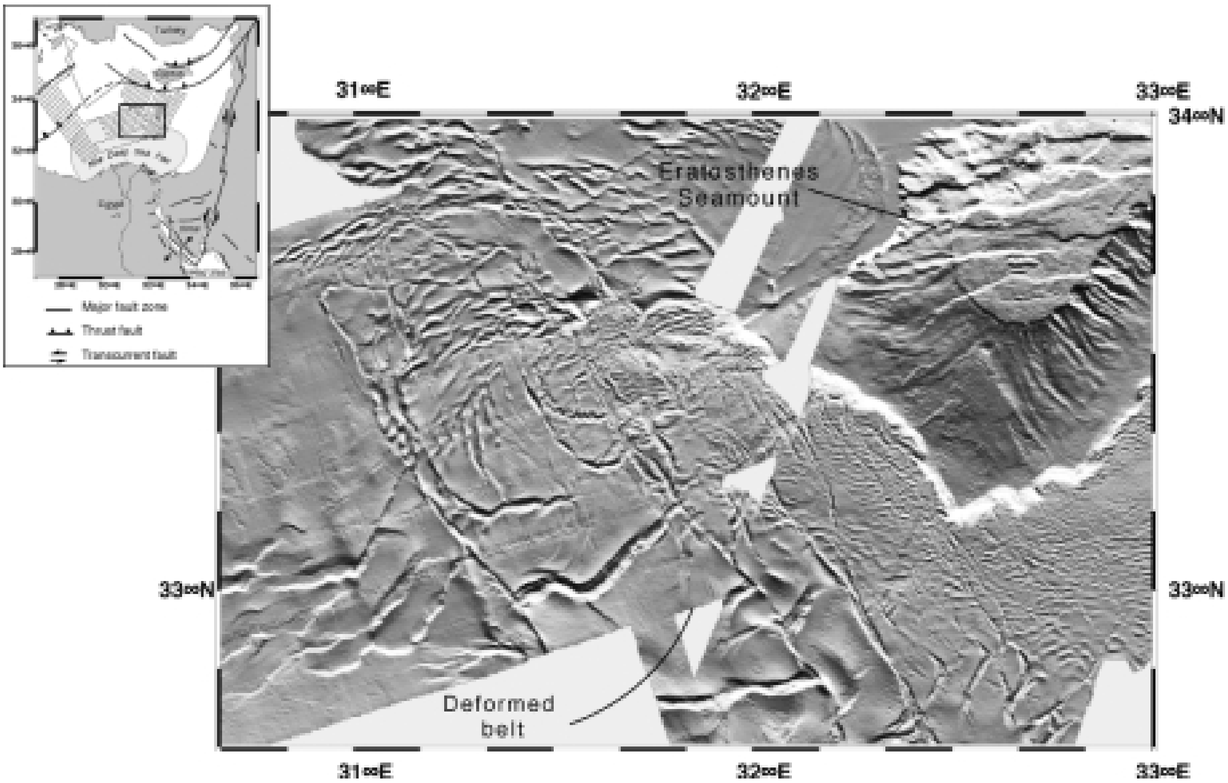

Fig. 1. Shaded bathymetry (from PRISMED II swath data) showing Eratosthene Seamount (right side corner), and the tectonized belt (central area) that runs across the Eastern Nile deep sea fan. The morphology illustrates he present day fracturing of Eratosthene seamount and the tectonics across the deep sea fan.

ORIGIN OF DATA

We have used for this paper a few published data almost only available for Erathostene Seamount (*e.g.* Ben Avraham *et al.*, 1976; Robertson *et al.*, 1995; Mart and Robertson, 1998; Robertson, 1998). For the two others study areas, only very scarce data were available, at least in academic files: Kenyon *et al.*, 1975; Ben Avraham *et al.*, 1987. We therefore based our interpretation chiefly on PRISMED II data (Mascle *et al.*, 1999; Bellaiche *et al.*, 1999; Gaullier *et al.*,

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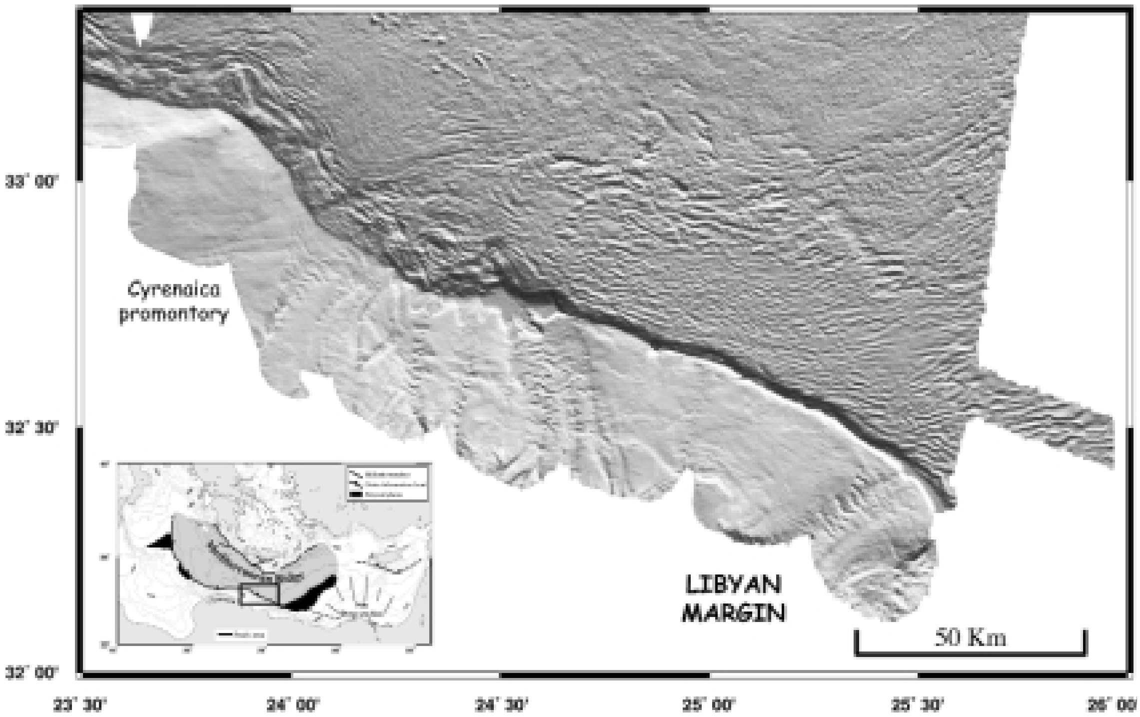

Fig. 2. Shaded bathymetry (from PRISMED II swath data) depicting the contact between the Libyan continental slope (to the South) and the southern border of the Mediterranean Ridge. Note the contrasted morphology of the Libyan slope facing different structural domains of the Mediterranean Ridge.

2000). This survey, run aboard the RV *l'Atalante*, allowed to continuously record, at 10 knots, swath data (bathymetry and sea floor backscatter images), 3.5 kHz profiling, seismic reflection, gravity and magnetic data. Only the results from swath bathymetry, imagery and seismic reflection profiling are taken into account in this preliminary study.

ERATHOSTENES SEAMOUNT (Fig. 1)

ES is a prominent, continental crust-rooted, feature, about half way between Egypt and Cyprus. It as been demonstrated, from previous studies, that the seamount, which shows as a flat-topped, about 80 km long, and relatively shallow (900 meters) relief, is progressively breaking as a result from its bending due to collision with Cyprus arc. Our data illustrate how the seamount, made of a thick pile of Mesozoic and partly Cenozoic sediments, is cut by a dense set of normal east-west trending, faults (with some strike slip components) and tilting towards north. Facing the though that bounds the Cyprus margin to the south, and as already proposed by Robertson (1998), some of the normal faults seem to be progressively re-activated into southward verging thrusts, bounding previous horsts, now on the process to be incorporated to Cyprus collision zone. The progressive tilting towards north of the seamount is probably at the origin of numerous slumps well seen along its southern slope (Gaullier, personnal communication)

EASTERN NILE DEEP SEA FAN (Fig. 1)

Large areas of the Nile DSF have been mapped during PRISMED II survey (see Loncke *et al.* this volume). We only focus here on the Eastern province which show very contrasting differences when compared to the Central and Western ones. These rely to a NW-SE directed, and more than 150 km long, fault belt where are probably expressed the combined effects of salt tectonics and of deep crustal extension, both controlling sediment deposition and complex topography. We have interpreted the area as a possible northern submerged extension of the gulf of Suez Rift zone (Said, 1962), delineating the north-western boundary of a Sinai microplate (Joffe *et al.*, 1987; Salomon *et al.*, 1996) whose others boundaries are relatively well constrain (Badawy and

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Horbath, 1999; Mascle *et al.*, 2000). If such an hypothesis is correct, it would indicate that parts of the Mesozoic margin of Egypt has been re-activated, during early to middle Miocene, as a consequence of the Gulf of Suez rifting episod and following an east-west directed extension. In the study area, the resulting structural pattern is enhanced by underlying thick Messinian evaporites that are likely progressively gliding towards north, i.e. towards the base of the deep sea fan (Gaullier *et al.*, 2000) and generating either specific salt deformations, and/or salt reactivations, in an active transtensive tectonic belt. In most of the Nile DSF the presence of underlying salt is at the base of widespread and important sedimentary collapses associated to numerous growth faults rooting in the evaporitic layers (Gaullier *et al.*, 2000).

LIBYAN CONTINENTAL MARGIN (Fig. 2)

Very little is known on this segment of the Mesozoic African margin which appears, surprisingly narrow on average. The data collected during PRISMED II concern mainly an area of the Libyan continental slope off Cyrenaica, and its contact with the southern border of the Mediterranean accretionary wedge, which is itself directly overthrusting above parts of the Libyan lower continental slope (Mascle *et al.*, 1999; Huguen *et al.*, in prep.). Fourteen seimic reflection profiles, full coverage swath bathymetry and acoustic imagery records make possible, however, to demonstrate (1) that the continental slope shows there in fact different segments, (2) that it has locally been submitted to some southward directed tilting, may be in Miocene times, and (3) is now partly affected by apparently transcurrent faulting (Mascle and Huguen, in prep.). We do not know yet the reason for its tilting prior to Pliocene times. It may be a consequence of reactivation of the nearby onshore Cyrenaica domain, or related to some precollision events between the Cyrenaica indenter and the MR to the north. Our data illustrate also the strong variability that exists at the contact between the MR and the base of the continental slope, which is locally seen to be directly overthrusted by the deformed sedimentary wedge. Finally, acoustic imagery shows that the upper to middle slope is cross-cut by linear recent faults (detected over more than 30 km) indicating thus that this region of the Libyan continental margin is also starting to be affected by tectonics likely related to its incipient collision with the Mediterranean ridge complex.

CONCLUSIONS

Different segments of the African Mesozoic continental margins have been and/or are still submitted to tectonic reactivations related either to their actual specific pre-collision setting, or to distensive events, probably linked to the rifting and subsequent opening of the Red Sea, in Miocene times. From our data set, at least three of these segments can be distinguished between eastern Egypt and Central Libya : the Erathostenes Seamount area, not directly belonging to the present margin but once part of the African Mesozoic margin, and the Libyan margin (off Cyrenaica) are both involved in collision to pre-collision processes with the southern borders of the Aegean-Anatolian microplate. This results, in both areas, in general bending, recent fracturing, and/or structural inversion at various degrees. A Mesozoic margin segment off Egypt, now covered by the thick piles of the Nile DSF, and particularly its eastern domain, has been re-activated sometimes in Miocene by distensive events that may have created an up to now unknown northward prolongation of the Suez Rift system. Messinian salt-rich sediments are probably enhancing the effects of this event, which may be still slightly active as it seems to be the case for the Suez Rift.

BOOK IN STOCK

Recent depositional pattern of the Nile deep-sea fan from echo-character mapping : interactions between turbidity currents, mass-wasting processes and tectonics

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Abstract

The Nile deep-sea fan has recently been surveyed using swath multibeam bathymetry, backscatter imagery, and seismic profiling. Three main morpho-structural and sedimentary provinces have been recognized: (1) a western domain mainly characterized by growth faults and a dense network of sinuous distributary channels; (2) a central province showing segments of sinuous channels, large growth faults, and salt ridges separating ponded basins; (3) an eastern province displaying impressive NW/SE-oriented tectonic structures interpreted as a potential northward extension of the Suez rift system. These physiographic characteristics seem to result from the combination of thin-skinned tectonics driven by thick Messinian layers and thick-skinned crustal-scaled tectonics linked to the inferred existence of a transtensive plate boundary under the Nile deep-sea fan. Recent sediment pattern, from echo-character mapping, shows that gravity-induced sedimentary deposits are prevailing on this huge fan. They are expressed either through mass wasting deposits or turbidites. This study highlights the importance of both salt-related and deep-seated active tectonics that shape the Nile cone seafloor, and therefore control its sedimentary distribution.

Introduction

The Nile deep-sea fan is the largest sedimentary clastic accumulation within the Eastern Mediterranean (Fig. 1). This system is relatively old since it was active at Oligocene times, southwest of the current Nile Delta. The modern Nile valley has been created during late Miocene, simultaneously with the Mediterranean desiccation. During Messinian times, seawater evaporation led to deposition of thick evaporites (Ryan *et al.*, 1973). Once the Mediterranean Sea was reconnected to the global ocean system through the Strait of Gibraltar, Messinian salt deposits were covered by a large deep sea fan from Pliocene to present day. Seismic surveys across this cone (Ross and Uchupi, 1977; Bellaiche *et al.*, 1999; Gaullier *et al.*, 2000, Mascle *et al.*, 2000) have shown that unconsolidated plio-quaternary sediments have an average thickness of 2000 kilometres, exceeding in place 3000 meters. This cover is frequently disturbed by salt diapirs, as in many other places in the Mediterranean Sea (Ryan *et al.*, 1970; Ross and Uchupi, 1977; Ryan, 1978, Ben-Avraham and Mart, 1981; Pautot *et al.*, 1984; Gaullier and Bellaiche 1996; Gaullier *et al.*, 2000).

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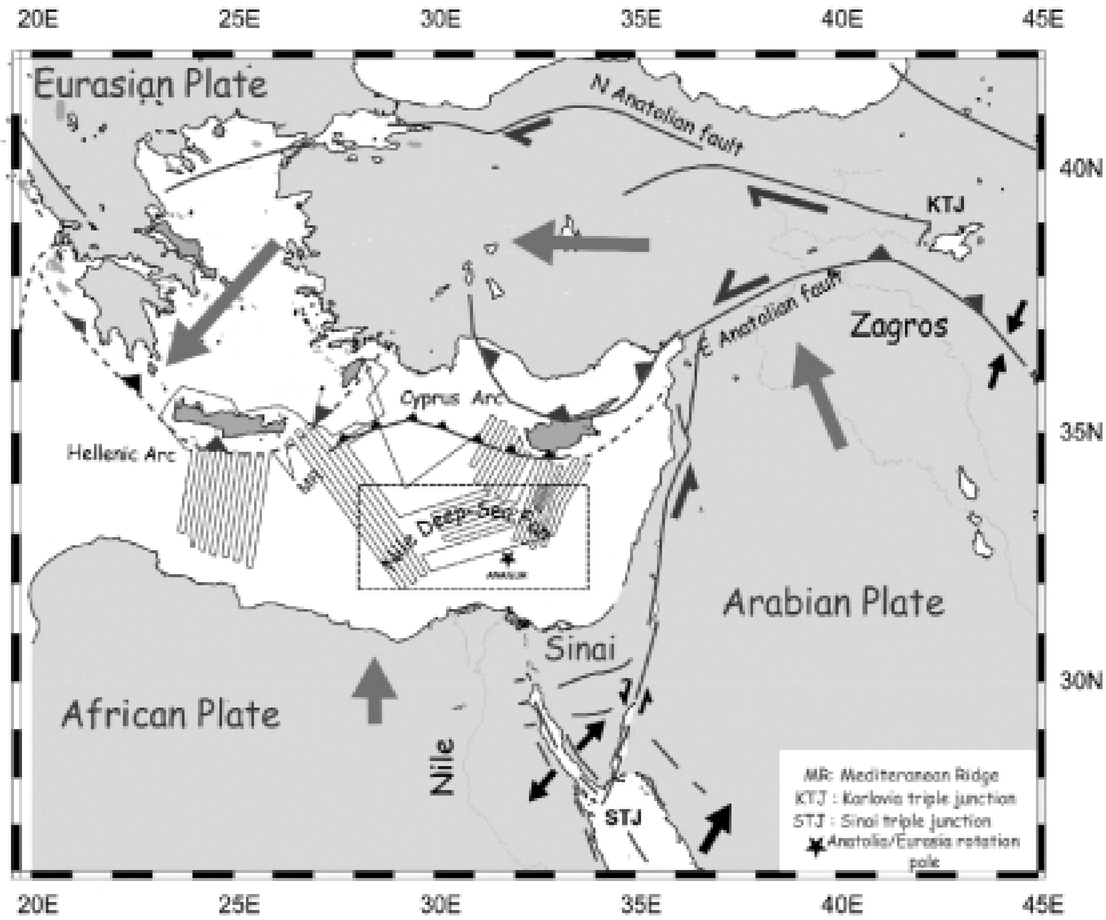

Fig. 1. Geodynamic sketch of the Eastern Mediterranean. Modified after Le Pichon *et al.*, 1988, 1995; Reillinger *et al.*, 1997 and Mc Clusky *et al.*, 2000. PRISMED II tracklines are indicated. The dotted box shows the study area. Heavy arrows indicate general plate motions.

The Nile deep-sea fan is constructed in a complex geodynamic setting (Fig. 1). Eastward, it is bounded by the Dead Sea shear zone, and to the north, by the Cyprus convergent zone and the Mediterranean Ridge. Furthermore, a Sinai microplate, locked between Arabia and Africa has been suspected either from plate-tectonics considerations (*e.g.*, McKenzie, 1970; Le Pichon and Francheteau, 1978, Courtillot *et al.*, 1987), and earthquake distribution (Salamon *et al.*, 1996; Badawy and Horváth, 1999). The western boundary of this microplate is believed to prolong the northwest-trending Suez-Cairo-Alexandria fault zone offshore Egypt and should therefore be covered by the eastern Nile deep-sea fan. This hypothesis is in good agreement with observations collected during PRISMED II cruise (1998) that evidenced a 150 kilometres long NNW-SSW narrow fault system, bisecting the eastern part of the surveyed area. (Bellaiche *et al.*, 1999; Mascle *et al.*, 2000, Gaullier *et al.*, 2000). The surface expression of these probably deep-rooted tectonic structures is complicated by salt tectonics (Neev *et al.*, 1976; Kenyon *et al.*, 1975, Woodside, 1977; Mart *et al.*, 1978; Ryan, 1978; Garfunkel and Almagor, 1987; Bellaiche *et al.*, 1999; Gaullier *et al.*, 2000; Mascle *et al.*, 2000).

The aim of the present study is to recognize the recent sedimentary pattern and to evaluate interactions between sedimentary processes, deep-seated tectonics and salt tectonics.

DATA SET AND METHODS

The PRISMED II survey, carried out on board the R.V. *l'Atalante* during early 1998, allowed to refine the physiography of this area using a Simrad EM12 Dual multibeam sounding system. This system allowed to survey along track strips 5-6 times wider than the water depth. The survey also recorded back-scatter images of the seafloor. Simultaneously, near-surface sedimentary structures were recorded up to 50-100m of penetration using a 3.5 kHz profiler. Deeper structures, locally

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up to 3 seconds two way travel time, were also imaged using a six-channel streamer and a single 75 ci Sodera GI gun. About 3500 km of continuous seismic and 3-5 kHz profiles have been recorded over the Nile deep-sea fan. This data set allows us to differentiate the three major morphostructural domains briefly described in the next section (Bellaiche *et al.*, 1999) (Fig. 2). To assess near-bottom sedimentary processes along the Nile Deep Sea Fan, we used mainly 3-5 kHz profiles, combined with bathymetry and imagery, following a method developed by Damuth (1980, 1994) that consists in constructing an echo-character map.

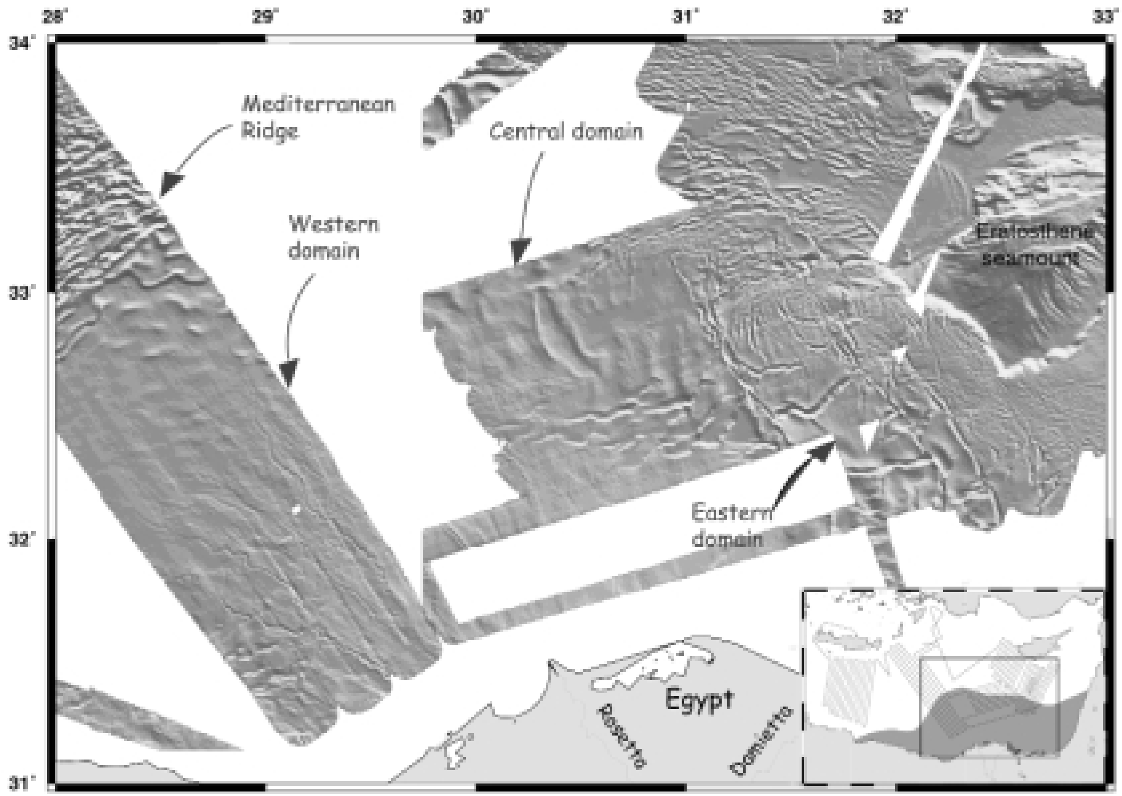

Fig. 2. Shaded bathymetry of the Nile Deep-sea fan, acquired during the PRISMED II cruise. This fan has been divided in 3 main morphostructural provinces: a western, a central and an eastern province.
The onshore Damietta and Rosetta trends, constituting the main distributary branchs of the Nile river, are indicated on this map.

RESULTS

The first step was to inventory the different echo-types observed in the study area. This was accomplished using guidelines established by Damuth in 1980 and refined by others (*e.g.*, Coutellier *et al.*, 1984; Le Cann, 1987; Gaullier and Bellaiche, 1998). We differentiated 11 echo-types gathered into 5 main families. Once this classification was established, we attempted to understand echo-type distribution in terms of sedimentary processes. Previous works allowed to allocate specific processes to some echo-types.

Western province

Physiographic characteristics:

This province is chiefly characterized by the N145-trending, more than 200km long, northern submarine extension of the onshore Rosetta branch of the Nile Delta. This prolongation corresponds to a meandering and ramified channel network that ends approaching the Mediterranean Ridge. Spoon-shaped normal growth faults, rooted on the Messinian salt layers, affect the upper slope in this area. To the north, the channel system and their lobe deposits are relayed by folded topographies that delineate the accretionary Mediterranean Ridge front. There, active compressive tectonics leads to a series of folds, reverse faulting, and piggy-back basins. These folds are generally salt-cored.

BOOK IN STOCK

Echo-character mapping

Two different grains are expressed: a proximal N145 grain, corresponding to the channel network, and a distal N70 grain, corresponding to the Mediterranean Ridge folds. Channels generally return Rough or Hyperbolic echoes on their axis, and are flanked by Bedded or Bedded-undulated Levees. These levees frequently display destabilizations expressed through Bedded-transparent or Hyperbolic echoes. The Mediterranean Ridge shows an acoustic domain made of a series of Transparent parallel ridges, separated by elongated basins filled with Bedded or Transparent-bedded sediments.

In summary, this province shows a well-developed drainage system that controls turbiditic flows which, subsequently, overflow levees. The middle cone channel's end by lower fan large lobes, probably of sandy nature. Some turbidites are likely directly incorporated in the Mediterranean Ridge. Proximal mass wasting deposits may have led to channel avulsions and migrations. In this area, tectonic processes, except salt tectonics, are poorly expressed.

Central province

Physiographic characteristics

Compared to the western area, the channel network appears far less developed. Small abandoned channels, less than 15 km in length, are detected in intraslope basins. These downslope channels appear to be isolated features. The main morphological characteristics of this province is the existence of numerous proximal spoon-shaped growth faults, roughly perpendicular to the slope. These are connected to each other in a 200 km long belt and appear to be the response to the sediments spreading above Messinian salt layers. Downslope, N145-trending topographic ridges affect the abyssal plain. They underline, as observed on seismic lines, elongated salt anticlines.

Echo-character mapping

On acoustic grounds, this province can clearly be divided into 2 domains :

• a first domain, in the upslope area, forms a broad allochtonous tongue, exhibiting Transparent, Transparent-bedded, Transparent-perturbated and Chaotic echo-types. All these echo-types form a continuum and are considered as typical of mass wasting processes.

• a second more distal domain, is characterized by Bedded echo-types located in depressions. They might indicate either distal turbidites or hemipelagites. The bathymetric ridges show, in contrast, Transparent to Rough echoes. In this area, the sediment traps, catching turbidites, seem to be strongly influenced by salt-related topographies (especially by the interaction of E-W and NW-SE trending salt pillows that form intraslope basins).

Eastern province

Physiographic characteristics

There is a significant contrast between the two previous provinces and the eastern one, marked by a 150 km-long NNW-SSE trending tectonic belt. This belt consists in a series of linear N145 fault zones bounding SW-NE sigmoidal grabens. Seismic reflection data indicate that these two types of structures are commonly cored by salt diapirs and that the SW-NE depressions correspond to crestal-grabens. We also notice in this region, deep-sea nilotic distributary channels. Some of these channels may have been enhanced by tectonics and their well-preserved sinuous pattern could be observed on nearly 100 km. Most channels appear however disrupted by faulting. Around the Eratosthenes Seamount, the deep-sea fan is bounded by a more than 500m-high arcuate escarpment, quite similar to the well-known Sigsbee Scarp in the Gulf of Mexico. We infer that Messinian salt layers were expelled and inflated by combined loading and seaward translation of the prograding sedimentary cover. Finally, northeast of the Eratosthenes Seamount, we evidenced another deep-sea channel system, oriented WNW-ESE, which is probably fed by canyons of the Cyprus or Levantine continental margins.

Echo-character mapping

This province exhibits, at least, two main morpho-acoustic domains :

• a westernmost domain corresponds to the NW-SE tectonic belt. It is mostly characterized by Transparent, or Transparent-Bedded, corresponding to destabilizated deposits. A former drainage system, feeding this area with turbiditic sediments, seems to be disrupted by faulting.

BOOK IN STOCK

• the Levantine platform shows surficial small folds, 50 to 100 meters in wavelength, probably due to a general progradation of the thin sedimentary cover above Messinian salt layers; they directly control the emplacement of further bedded sedimentary deposits that appear trapped in small depressions.
• the Eratosthenes Seamount is itself characterized by Bedded or Hyperbolic echoes. Hyperbolic echoes are generated by steep slopes and normal WSW-ENE faults cutting across the seamount, whereas Bedded echoes are attributed to hemipelagites.

Globally, the sedimentation seems to be strongly influenced by the NW-SE trending tectonic belt. Evidences of recent tectonic activity are numerous: channels are frequently interrupted, and inferred destabilizated sediments are systematically associated with large sigmoidal crestal grabens.

CONCLUSIONS

Our echo-character mapping allows to define the different recent sedimentary processes that shape the Nile deep-sea fan and to evaluate their regional influence (Fig. 3). The data examined in this study show firstly that the terrigenous sediment deposition is clearly controlled by two main gravity-induced processes: (1) turbidity currents and related gravity-controlled density flows; and (2) mass-transport processes including slumps, slides and debris flows. We assume that broad tongues of mass flow deposits were generated by slope destabilizations, which can be triggered by seisms, salt-related faults activity, eustatic fluctuations, gas seeps, or mud volcano activity; the data show secondly that the sedimentary distribution of the Nile deep-sea fan is strongly influenced by its own physiography which is, itself, controlled by thin-skinned and thick-skinned tectonics. A fairly recent activity of the resulting structures is systematically attested by associated inferred destabilizated sediments.

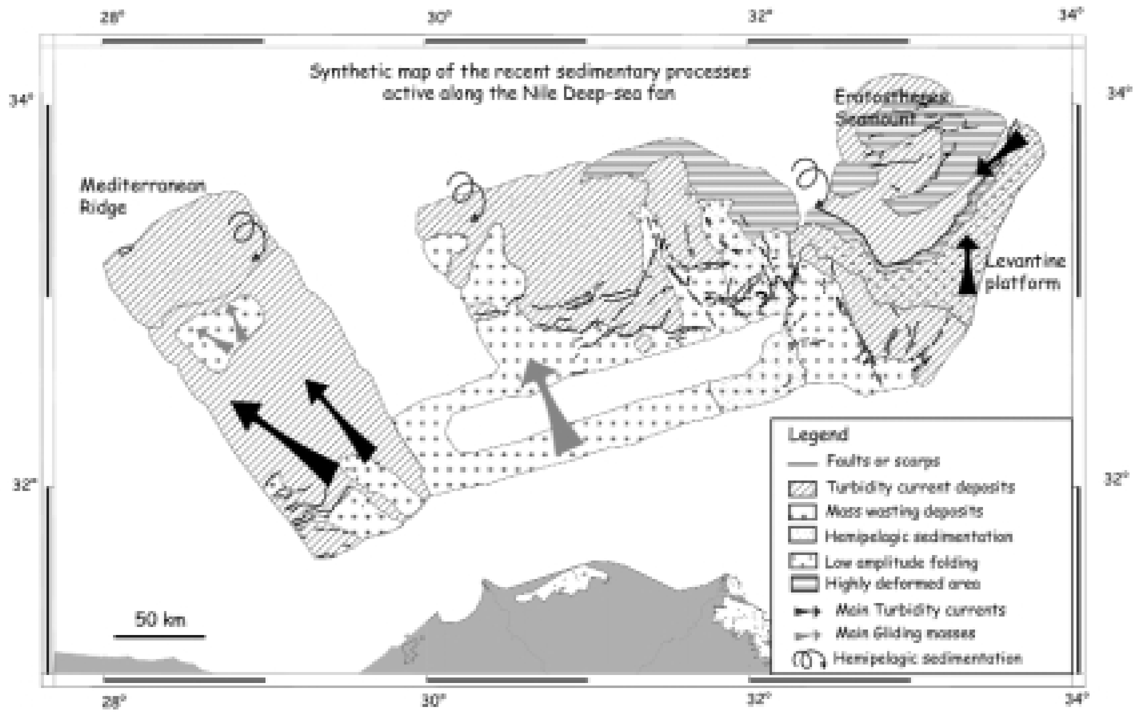

Fig. 3. Synthetic map of the recent sedimentary processes active over the Nile deep-sea fan, determined by echo-character mapping. Is overlain on this map a simplified morphostructural sketch.

BOOK IN STOCK

Evolution of sedimentation patterns on the north African margins: an update from studies of recent mud volcanic deposits

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Since 1991 mud volcanism of the Mediterranean and adjacent regions has been a focus for scientific exploration carried out within the framework of the UNESCO "Training-through-Research" (Floating University) programme. Intensive investigation of this phenomenon has resulted in the discoveries of new mud volcanoes in the Eastern Mediterranean, the Alboran Sea, and the Gulf of Cadiz and a large database has been compiled on lithology of mud volcanic deposits.

Mud volcanic deposits (mud breccia) sharply differ from the surrounding pelagic sediment and are composed of silty sandy clay (mud breccia matrix), very poorly sorted, with significant gravel admixture (mud breccia clasts). Clay mineral compositions of the matrix and clast lithologies vary for different mud volcanic areas. Large mud breccia clasts are represented by a wide variety of fragments of deep-seated sedimentary rocks of different lithology, age, and genesis. They provide a unique opportunity of direct geological observation of samples from deeply buried strata, through which mud volcanoes erupted and which are still hard-to-reach for drilling.

Up to now the most studied mud volcanic fields are the Cobblestone 3 Area, the Olimpi/Prometheus 2 Area, and the UN Rise Area, located on the Mediterranean Ridge (Premoli *et al.*, 1996; Robertson and Kopf, 1998). Providing the most significant information about the composition of the deep-seated series (Akhmanov and Woodside, 1998), lithology of mud breccia clasts from these areas was analysed in detail. This was followed by paleontological investigation performed in the University of Milan (Italy) by Prof. I. Premoli and Dr. E. Erba and a representative collection of the clasts was dated. This study has resulted in reconstruction of deep-seated sedimentary series of the Mediterranean Ridge (Fig. 1).

The results of the mud breccia clast study suggest that a deep-sea basin has been present in the Eastern Mediterranean at least from the Aptian. The Aptian-Albian claystones, found as rock clasts in mud breccia, are very fine, well-sorted, and formed in marine conditions with high input of terrigenous clayey material. This interpretation accords perfectly with the conception of an Early Cretaceous global sea level fall (Haq *et al.*, 1987) and with conclusions that at this time a wide deltaic front was being formed on the north African continent (Sestini, 1984) and thick tur-

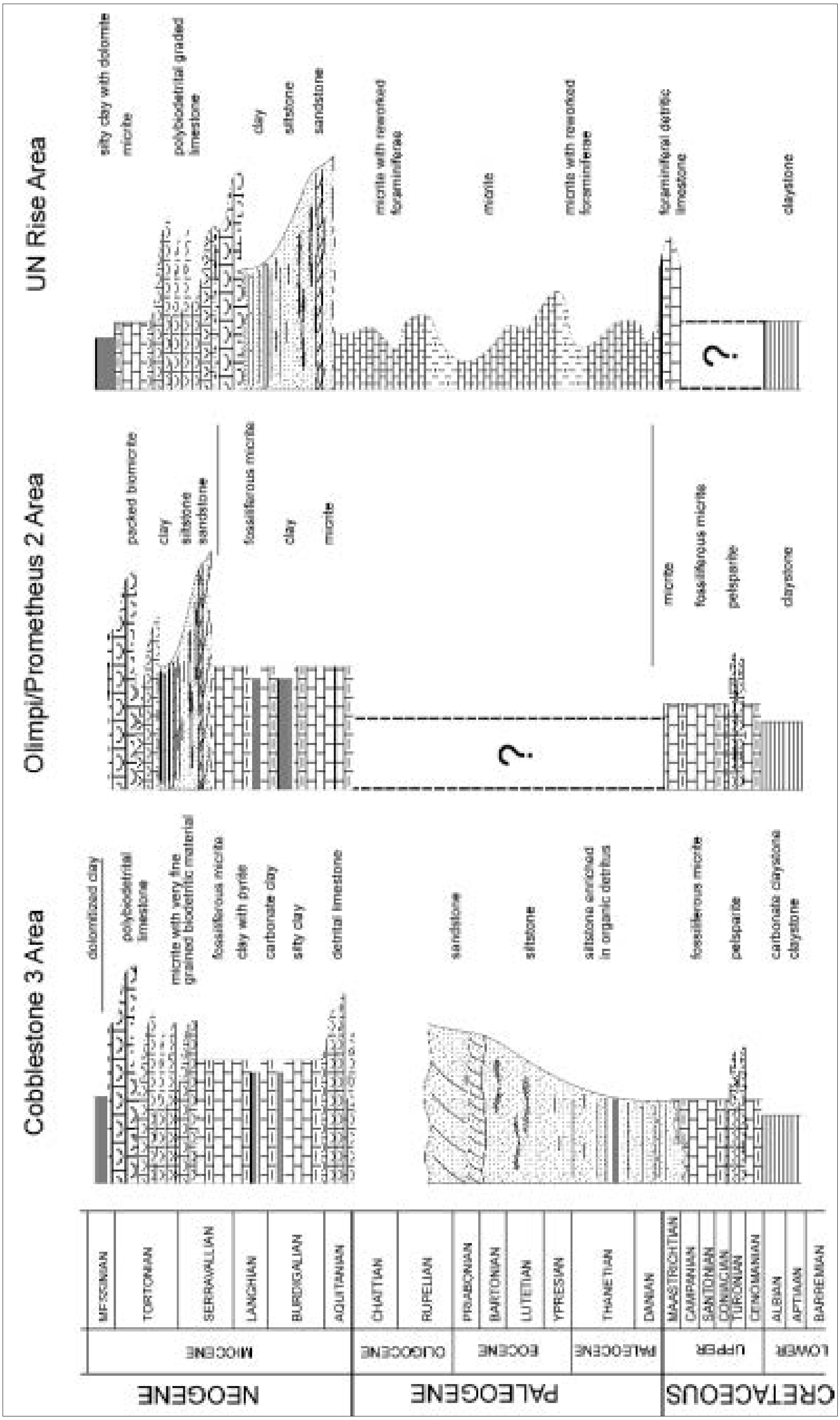

Fig. 1. Sedimentary series of the Mediterranean Ridge. Reconstruction is made on the base of a study on recent mud volcanic deposits.

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bidite series were being deposited at the south Pelagonian continental rise (Hall *et al.*, 1984). A large amount of terrigenous material was being supplied into the basin and the finest particles could reach its deep-sea parts. A relatively narrow deep-water gulf of the southern Neotethyan basin existed in the western sector of the Eastern Mediterranean and served as a settling basin for terrigenous clayey material at the end of the Early Cretaceous. A thick clayey series accumulated in the basin. northern Africa was undergoing denudation and its margin was an area of active sediment transport.

Upper Cretaceous rock fragments from the Cobblestone and Olimpi/Prometheus mud breccias are represented by micrites and fossiliferous micrites. Foraminiferal assemblage and micritic composition of these rocks imply a pelagic origin, suggesting that in the Late Cretaceous the Eastern Mediterranean basin enlarged, terrigenous input into deep part of the basin decreased, and carbonate pelagic sedimentation became predominant.

The uppermost Cretaceous and Paleogene rocks, found as clasts in the mud breccia, differ very much for different areas of the Eastern Mediterranean. The Paleogene rocks of Cobblestone 3 Area are represented by cross-laminated quartzose sandstones and compositionally similar siltstones, enriched significantly in terrestrial organic fragments. Good sorting, compositional maturity, cross-lamination and abundance of terrestrial organic matter in these rocks imply a relationship with a fluvial system. The series could be formed by rapid sedimentation in an area of deltaic or prodeltaic progradation, resulting from a falling sea-level. Clasts of rocks, which can be dated as Paleogene, were not found in mud volcanic deposits of the Olimpi/Prometheus 2 Area. In contrast, clasts of Uppermost Cretaceous - Paleogene rocks are widely represented in the mud breccia of the UN Rise mud volcanoes. Among them there is well-sorted, clast-supported biodetrital, mainly foraminiferal limestone. Its formation can be attributed to bottom current activity, strong enough to washout the finest material and to condense biodetritic material. Bottom current activity probably also affects the deposition of other Paleogene rocks from the UN Rise Area, represented by fossiliferous micrites with reworked foraminifera.

The results of mud breccia clast study allow us to conclude that the Paleogene was characterised by a decrease of the Eastern Mediterranean basin areas and by a probable increase in their separation. Within dissociated basins the specific depositional environments were set up in accordance to local geological and hydrological conditions. The regression in the western part of the basin led to deposition of thick terrigenous siliciclastic series of deltaic-prodeltaic sediments. The eastern part of the basin was characterised by predominant carbonate pelagic sedimentation and bottom-current activity. This interpretation agrees with the hypothesis that the regional tectonic re-arrangement of the Neotethyan basins occurred at the end of the Late Cretaceous (Robertson and Dixon, 1984; Robertson *et al.*, 1991; Robertson, 1998a).

The Miocene rock fragments are the most abundant and accordingly the most diverse in the mud volcanic deposits of all the studied areas. The Miocene sequence in the Cobblestone 3 Area starts with the Low Miocene detritic limestones with large amount of silisiclastic sandy admixture. These rocks are poorly to medium sorted and show gradual lamination, suggesting a turbiditic origin. The Middle Miocene series are represented by pelagic fossiliferous micrites and well-sorted claystones with pyrite nodules. This series are followed by very fine grained, graded micritic limestones with foraminifera, interpreted as intrabasinal turbites redepositing pelagic material within the basin. The Upper Miocene is composed of coarse-grained, graded, polybiodetrital limestones of turbiditic origin and by dolomitized claystones.

Among the Miocene rock fragments from mud breccia of the Olimpi/Prometheus 2 Area micrites and fossiliferous micrites predominate, both of pelagic origin. Other rocks presented in the mud breccia from this area are graded sub-arkose sandstones and packed detrital biomicrites formed as turbidites.

The Miocene sequence of the UN Rise area is represented by sub-arkose sandstones. They are medium to poorly sorted, with convolute and graded lamination, medium to fine grained and often in contact with siltstones (of similar grain composition) enriched in organic matter. These siliciclastic rocks are characterised by genetic indicators allowing us to consider them as turbidite deposits. Rocks of other turbiditic series of the Miocene are also widespread, among the mud

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breccia clasts of the UN Rise area. They are represented by coarse to medium grained, graded, rather poorly sorted polybiodetrital limestones. Turbiditic series are followed by pelagic Upper Miocene fossiliferous micrite, very abundant among mud breccia clasts.

Mud breccia clast study suggests that in the Miocene several turbidite systems were widely developing in the Eastern Mediterranean region, together with the predominant carbonate pelagic sedimentation. Turbiditic sedimentation in the western sector of the basin was characterised by a supply of carbonate detritus mixed with siliciclastic terrigenous material. In the central part of the basin thick clastic terrigenous and detrital carbonate series have been deposited separately. Siliciclastic material supplied into the central part of the basin was carried out mainly by the palaeo-Nile, shifting gradually to the east, while the Eocene carbonate series of north Africa were the main source for carbonate turbidite deposition. Besides, the Miocene tectonic activation led to episodic re-deposition of pelagic material within the basin, forming very fine-grained mostly carbonate basinal turbitides.

In general, the configuration of the Eastern Mediterranean basin has changed significantly during its geological evolution since the Cretaceous, gradually approaching to the present contours. Beside global sea-level fluctuations, the style of sedimentation has been affected by local tectonic movements which change the provenances within the basin. The north African margin was the main supplier of siliciclastic and carbonate detritic material into the basin.

Ideas about evolution of sedimentary patterns on the western sector of the north African margin (*e.g.* the Moroccan margin) also benefit from study of mud volcanic deposits.

Signs of very complex sedimentary history are found in rock fragments from the mud breccia of mud volcanoes, discovered in 1999-2000 on the Moroccan margin in the Gulf of Cadiz. These rock fragments were dated with Lower/Middle Eocene and Miocene ages and were represented by a large variety of lithologies. They are mainly different sandstones and claystones of turbiditic origin, suggesting very complex evolution of the depositional environment on the continental rise. Many hemipelagic claystones and clayey limestones of Miocene age were also described among mud breccia clasts, implying a predominance of deep-sea basin conditions. Some shallowing of the basin can be assumed for the Middle Miocene to the beginning of the Late Miocene, when characteristic foraminiferal and biodetrital limestone were deposited. Mud breccia clasts represented by various carbonate rocks, such as micrites, marlstones and bituminous limestones, characterise the Upper Miocene depositional environment in the area.

Recently discovered and sampled mud volcanoes on the Moroccan margin of the Alboran Sea also might provide some additional information about the north African margin (Kenyon *et al.*, 2000). The mud breccia from the Granada mud volcano (West Alboran Sea) contains numerous fragments of Upper Cretaceous-Neogene rocks, implying the location of mud volcano roots within the Early Miocene olistostrome unit overlying the basement. Some insights into pre-Miocene history of the region can be expected from further analysis of the mud volcanic deposits.

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Tectonic setting of the Levant margin

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The Levant continental margin off Israel, Lebanon and Syria is a passive margin that was formed during early Mesozoic by continental rifting. The morphology of the margin changes significantly from south to north. The width of the continental shelf is about 25 km in the south, about 8 km offshore northern Israel and about 3 km offshore Lebanon. Off Haifa Bay there is a local widening of the continental shelf to 15 km. The continental slope which is gentle in the south, about 2 degrees, becomes much steeper, about 8 degrees offshore southern Lebanon. The continental slope off northern Israel and southern Lebanon is cut by numerous submarine canyons which form disturbed bathymetry.

The morphology partially reflects the deep crustal structure of the continental margin and the tectonic processes which are taking place over there. As seen by the morphology, the continental margin is divided into two distinct provinces separated by the Carmel structure which extends northwestward from land across the continental margin into the Levant basin. Seismic refraction and gravity data indicate that in the southern province the transition from continental to oceanic crust is gradual and takes place tens of kilometers offshore, at the base of the continental slope. On the other hand, in the northern province the transition is rather sharp and takes place close to the coast (Fig. 1). Magnetic anomalies are also of different patterns in the two provinces. It is possible therefore that the continental margin south and north of the Carmel structure were formed in different ways and maybe even at different times.

The transition from the southern province to the northern province is sharp across the Carmel structure. This structure therefore is an important element. On its northern boundary the Carmel structure is bounded by the Yagur fault, which forms the northwest edge of a fault system which branches from the Dead Sea rift. This fault system forms a zone of seismic activity. The northern province of the Levant margin is also seismically active. In addition, continuous seismic profiles indicate the existence of active faults in this region. Part of the faults trend east-west and continue from land to sea. However, at the base of the continental slope a wrench fault system parallel to the trend of the margin was identified. The southern province is more stable.

In order to explain the tectonic activity in the northern province, a model which relates the activity in this region to the movement along the Dead Sea transform was developed. The model, which is based on a mathematic simulation, shows that part of the movement between the Arabian plate and the Sinai plate is transferred to the northern province of the continental margin through faults in northern Israel and southern Lebanon, which connect the Dead Sea rift with the continental margin. Other part of the movement between the plates continue northeasterly along the Yammuneh Fault in Lebanon and other faults. According to the model, the Dead Sea

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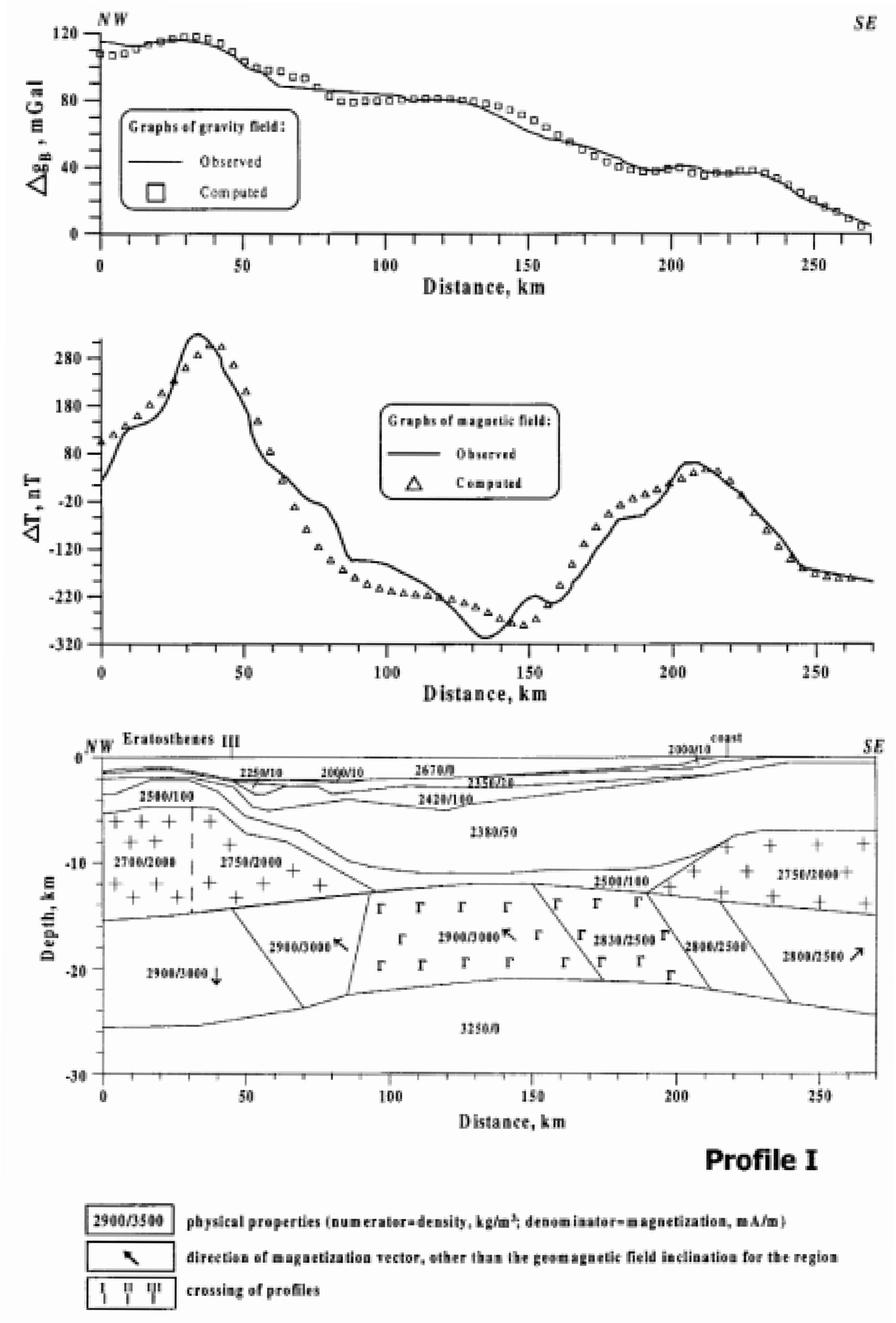

Fig. 1. A combined gravity and magnetic interpretation of a seismic refraction/wide angle reflection profile across the northern province of the Levant continental margin. The interpreted oceanic crustal layer ( ) and upper continental crust of the Eratosthenes Seamount and the Levant (+) are marked.

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rift was formed by propagation of rifting simultaneously from the Red Sea in the south and from Taurus Mountains collisional zone in the north. With time, faulting is also developed along the southern province of the continental margin, but this is not a wrench faulting. Understanding of the nature of faulting along the Levant margin, thus, is important for the understanding of tectonic process in the entire Middle East.

The results of various geophysical studies strongly suggest the existence of oceanic crust under portions of the Levant basin and continental crust under the Eratosthenes Seamount. The geophysical data indicate a large sedimentary sequence, 10-14 km thick, in the Levant Basin and below the Levant continental margin. Assuming the crust is of Cretaceous age, this gives a fairly high sedimentation rate. The sequence can be divided into several units. A prominent unit is the 4.2 km/s layer, which is probably composed of the Messinian evaporites. Overlying the evaporitic layer are layers composed of Plio-Pleistocene sediments, whose velocity is 2.0 km/s. The results also indicate that the Levant Basin is composed of distinct crustal units.

BOOK IN STOCK

Tectonic setting of the north African/Arabian continental margin in relation to the bordering Tethys ocean

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Given its scale and importance the north African/Levant continental margin (Fig. 1) remains surprisingly poorly known. Exceptionally amongst the major continental margin segments it has never been studied by academic drilling; marine surveys remain sparse, and information from hydrocarbon exploration remains selective (*e.g.* Nile delta). Currently, the timing and process of continental break-up and initial spreading of the bordering Tethys ocean remain controversial. To name a few questions: Is the north African margin of “volcanic”-or “non-volcanic”-rifted type, or both ? When did initial ocean spreading take place ? Is the Levant margin segment of transform, or orthogonally rifted type ? To what extent are different segments of the margin now affected by collisional tectonics ?

During the Palaeozoic (Silurian-Carboniferous) the present north African/Levant passive margin formed part of Gondwana bordering the Palaeotethys ocean to the north. This was followed by a switch to episodic extensional pulses of Late Permian to Jurassic age (see Guiraud and Bosworth, l999). One key question is which of these rift pulses actually corresponds to continental break-up and initial spreading of a southerly Neotethys ocean.

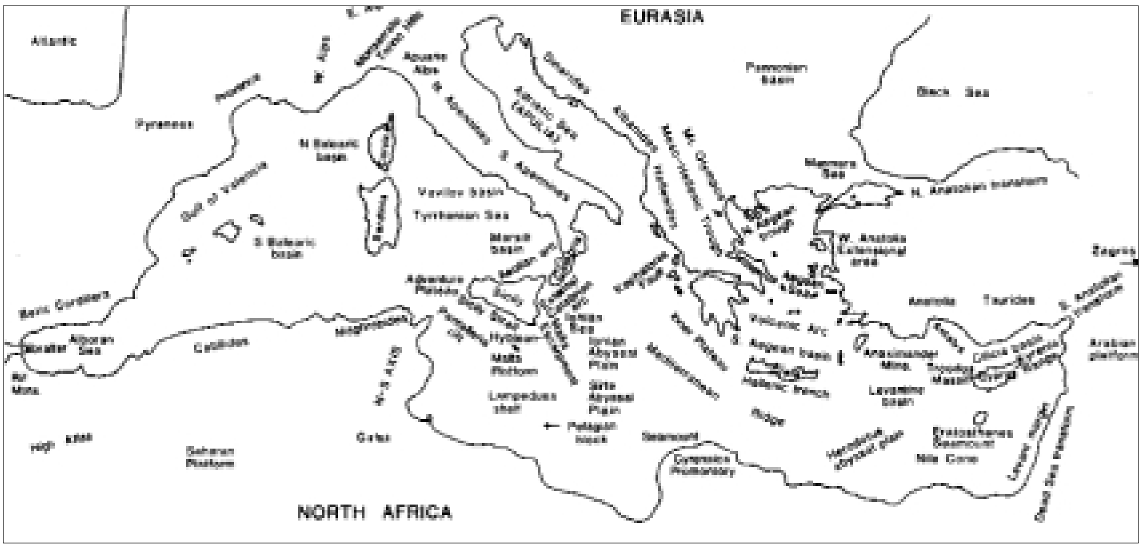

Fig. 1. Location of features in the Mediterranean region, as discussed in this summary. From Robertson and Grasso, 1995.

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It is generally agreed that the north African/Levant passive margin owes its origins to two fundamentally different spreading systems (see papers in Ziegler *et al.*, in press). First, the opening of the Neotethys, westwards from Oman, along the Arabian margin into the Eastern Mediterranean. There is evidence of a major rifting event in the Late Permian from Crete, S-W Turkey and Italy. According to different authors the continental break-up of the eastern Tethyan ocean basin (southern Neotethys) took place in Late Permian (Stampfli *et al.*, 1998, in press), Late Triassic-Early Jurassic (Garfunkel and Derin, 1984; Garfunkel, 1998; Robertson *et al.*, 1991), or not until mid-Late Cretaceous time (Dilek and Moores, 1990). However, evidence from the Levant margin, from Cyprus, and from southern Turkey (Antalya) is indicative of initial spreading in Late Triassic-Early Jurassic time. Additional evidence from north Africa (Tunisia/Morocco) documents pulsed rift events during Late Permian, Triassic and Jurassic time, but does not provide unambiguous evidence for the timing of initial seafloor spreading within Neotethys to the north. Recent geochemical data from Triassic igneous rocks in Greece, Turkey and Cyprus (Pe-Piper, 1998; Dixon and Robertson, 1998) indicate that voluminous alkaline- to transitional MORB-type volcanics show an apparent plume-related influence, suggesting comparison with volcanic-rifted margins, or an intermediate sate between ideal “volcanic” and “non volcanic”-type rifted margins.

The second spreading event relates to the opening of the North Atlantic in the Late Jurassic systems (see papers in Ziegler *et al.*, in press). The spreading basin extended through the Betics/Rif area of south Spain/north Africa, through Liguria (north Italy), and the Eastern Alps into the Pannonian basin region. Evidence of units such as the Ronda peridotite in south Spain can be compared with the asthenosphere exposed near the continent ocean boundary on the Iberian margin, in turn, suggesting a non-volcanic margin type of rifting in this western basin, as also supported by evidence of ophiolite-related units in the Eastern Alps and the Liguria.

How the eastern and western spreading systems interacted remains controversial. In one view the two stopped short of each other leaving an Apulian promontory as a N-S barrier. Alternatively, the two ocean stands converged (or overlapped) opening an E-W ocean basin all along the north African margin by Late Jurassic-Early Cretaceous time. Several lines of evidence support the latter hypothesis: 1) faunal provinces were linked even by Late Permian time, suggesting free exchange of ocean waters (Kozur, 1993); 2) palaeomagnetic data indicate that Apulia has evolved as a microplate independent of Africa and Eurasia at least during Late Mesozoic/Early Tertiary time (E. Marton, pers. com. 2000); 3) geophysical evidence suggests that the Ionian Sea is likely to be underlain by oceanic or transitional crust that is now being subducted northwestwards beneath the Tyrrhenian Sea (G. Rehault *et al.*, 1985; Gueguen *et al.*, 1998); thus, crust south of Apulia is not merely continental.

After formation of a single N-facing passive margin by Late Jurassic time, the north African/Levant margin continued to undergo passive margin subsidence during Late Mesozoic-Early Tertiary time. However, it experienced echoes of far-field tectonic events, including pulses

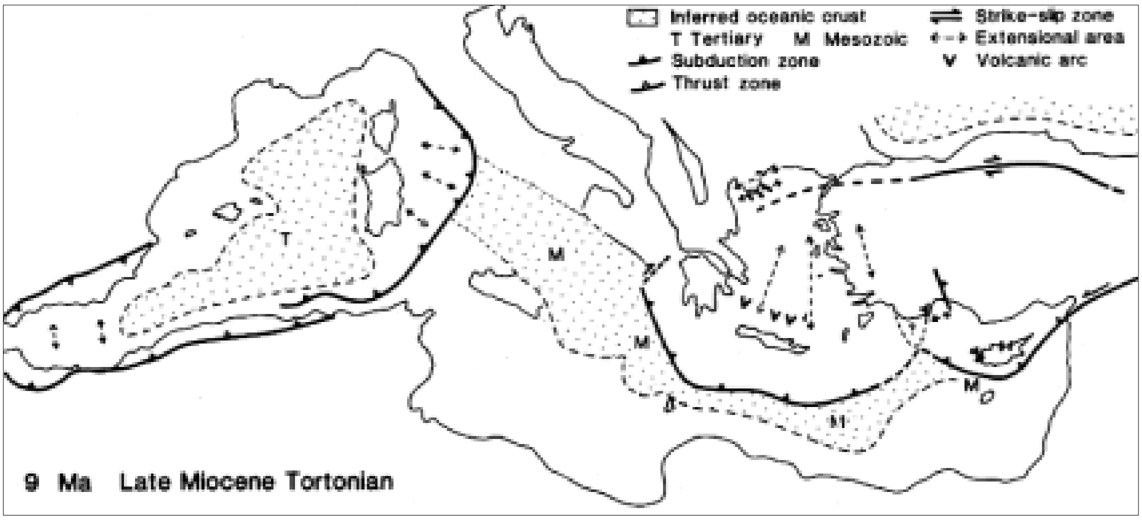

Fig. 2. Reconstruction of the tectonic setting of the north African margin in the Late Miocene. .

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of compression, strike-slip and extension in specific areas (Guiraud and Bosworth, 1999). Eventually the margin began to be affected by the collision of Africa and Eurasia (see Robertson and Grasso, l995; Fig. 2). This was manifested first in the west with Early Miocene thrusting of nappes onto north Africa and Sicily (*e.g.* Maghrebides). The driving mechanisms was possibly "roll-back" of surviving Neotethyan oceanic crust, or detachment of mantle lithosphere from the thickened root of the collision zone (Platt and Vissers, 1989 *in* Meghraoui, 1996). Further east the Tunisian Peninsula (N-S axis) collided with the Apulian microplate, reactiving N-S strike-slip lineaments. Eastwards again, the Cyrenaica Peninsula (Libya) began to collide with the Mediterranean Ridge accretionary wedge, arguably since Late Miocene (Chaumillon and Mascle, 1995; Mascle *et al.*, this volume) and more probably since Late Pliocene time. The Egypt segment further east remains in a passive margin state, except for the Eratosthenes Seamount (Woodside, 1977), a rifted fragment of north Africa/Levant continental crust, which is now undergoing collision with, and underthrusting beneath, the Africa-Eurasia plate boundary represented by the Cyprus arc (Robertson, 1998). The deep water Nile cone further south is apparently affected by northward propagation of rifting from the Gulf of Suez region (Mascle *et al.*, 2000). In addition, the Levant margin and Egypt segments were affected by earlier collisional events related to the northward migration of the Arabian promontory towards Eurasia, notably involving the Late Mesozoic and mid-Tertiary stages of compression-related inversion of the Syrian Arc. The overall present-day setting is shown in Figure 3.

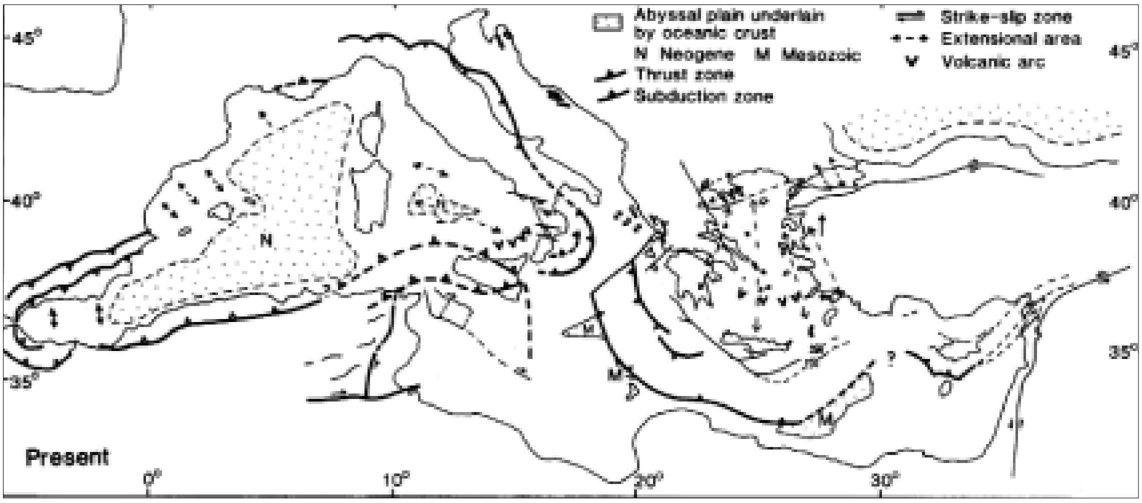

Fig. 3. Simplified present-day plate tectonic setting of the Mediterranean. From Robertson and Grasso, 1995.

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Channelised deep sea depositional systems in the Mediterranean : the value of sonar backscatter mapping

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Both long range GLORIA and OKEAN sidescan sonars, long range swath bathymetry and its accompanying backscatter information, and higher resolution deep towed sidescan sonars have been used to map deep sea "turbidite systems" in the Mediterranean over the past 30 years. The main gap in knowledge is detailed information on the distribution of the sands in the systems. This is because of the difficulties of obtaining good cores through sands and because of the lack of high resolution, deep towed seismic data. The main gap in the coverage is along the margins of north Africa, with the exception of the Nile Cone which is now relatively well known (Bellaiche *et al.*, 1999).

GLORIA and other data from along about 30,000 miles of low latitude (i.e. non glacial) continental margin, worldwide, provides sufficient data for a new classification of deep sea "turbidite systems" (Fig.1). The scheme is based on slopes with a fall of 3000m or more and does not hold for shelf edge deltas. The plan view emphasis is a valuable complement to sequence stratigraphic schemes that are derived mainly from a study of seismic profiles and rock outcrops. The simplest classification is into a spectrum of types (Fig.1), distinguished by their channel and lobe characteristics, in which the main control is the long term average rate of sediment input. Long

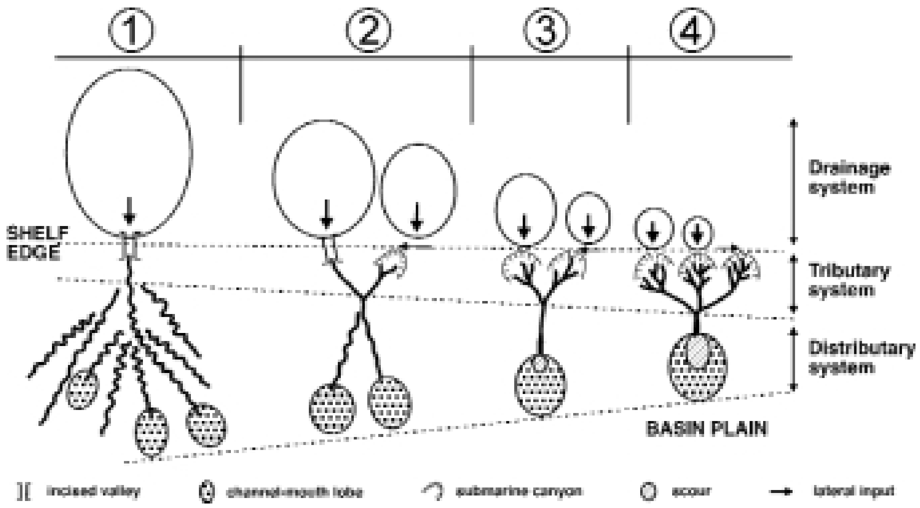

Fig. 1. Simplified classification of channelised depositional systems on low latitude margins.

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term rate of input will be affected by drainage basin size and gradient, and by climate, to a greater extent than by sea level. The model is tested for the Mediterranean using both published information, and unpublished data from West of Corsica and Sardinia and from the Tyrrhenian and Alboran seas.

TYPE 1. The model requires the mature, highest input types to have a point source, a large radial distributary channel system with sinuous channels and low fan gradients. Sinuosity should be greater than about 1.6 in the middle reaches of channels and maximum channel gradients should be less than about 1 in 100. Width to depth ratios of channels are usually less than 50. Sandy lobes are expected to be attached to the ends of channels, i.e. without extensive erosion upstream of the sand sheets. Flows are frequent, probably in excess of one per year at times of low sea level. The Nile Cone is the only system of this type in the Mediterranean, having a drainage basin of over 2 million km^2. It differs from the perceived ideal for a high input fan (*e.g.* the Indus or Amazon fans), in having slightly higher overall gradients. However the channels can be highly sinuous, up to 2 or more (Bellaiche *et al.*, 1999). Tectonics affects the fan surface and may be the cause of the higher gradients.

TYPE 2. The Rhone Fan is close to the norm for a medium to high input type. The drainage basin is about 100,000km^2 (Fig. 2) and it has a distributary pattern of channels with at least one avulsion (*e.g.* Torres *et al.*, 1997). Maximum sinuosity is greater than 2 for the abandoned channel but is only about 1.4 for the newly avulsed and entrenched channel. With further growth in the new fan lobe, the sinuosity should increase.

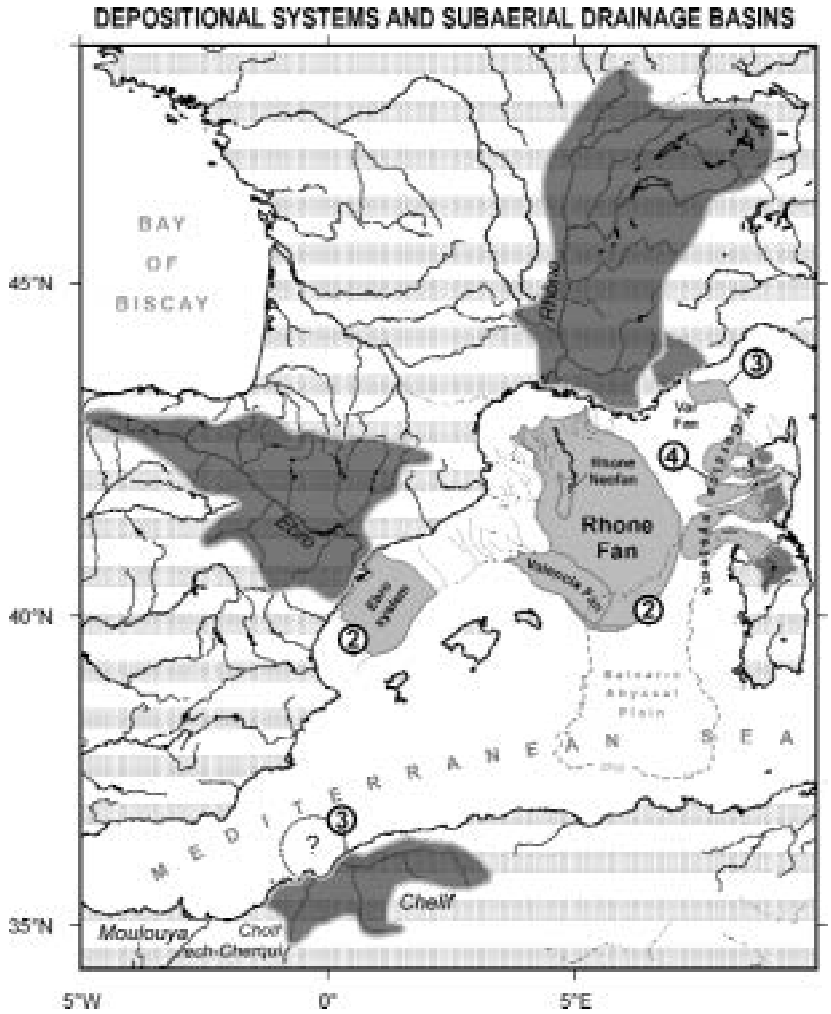

Fig. 2. Some subaerial drainage basins and their submarine depositional systems in the Western Mediterranean. The numbers refer to the types in Fig.1.

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TYPE 3. The Var fan and the Ebro are considered to fall within the medium to low input type. There is a drainage basin in north Africa (The Chelif, Fig. 2) that should be large enough to produce a type 3 system, but little is known of the submarine margin in this area. The Var drainage basin has an area of about 4,000km^2. It is fed by very steep tributary canyons with a maximum gradient of about 1 in 5. The channel is relatively straight and wide (up to 7km) and it has a well developed levee along its right bank. Downslope from the leveed section of the system the channel continues with a distributary pattern developing where it becomes very shallow (about 2m deep). A lobe can be identified from the backscatter pattern (Fig. 3). The lobe at the mouth of Ebro system has well developed scours and also constructional, probably sandy, bedforms (*e.g.* Morris *et al.*, 1998).

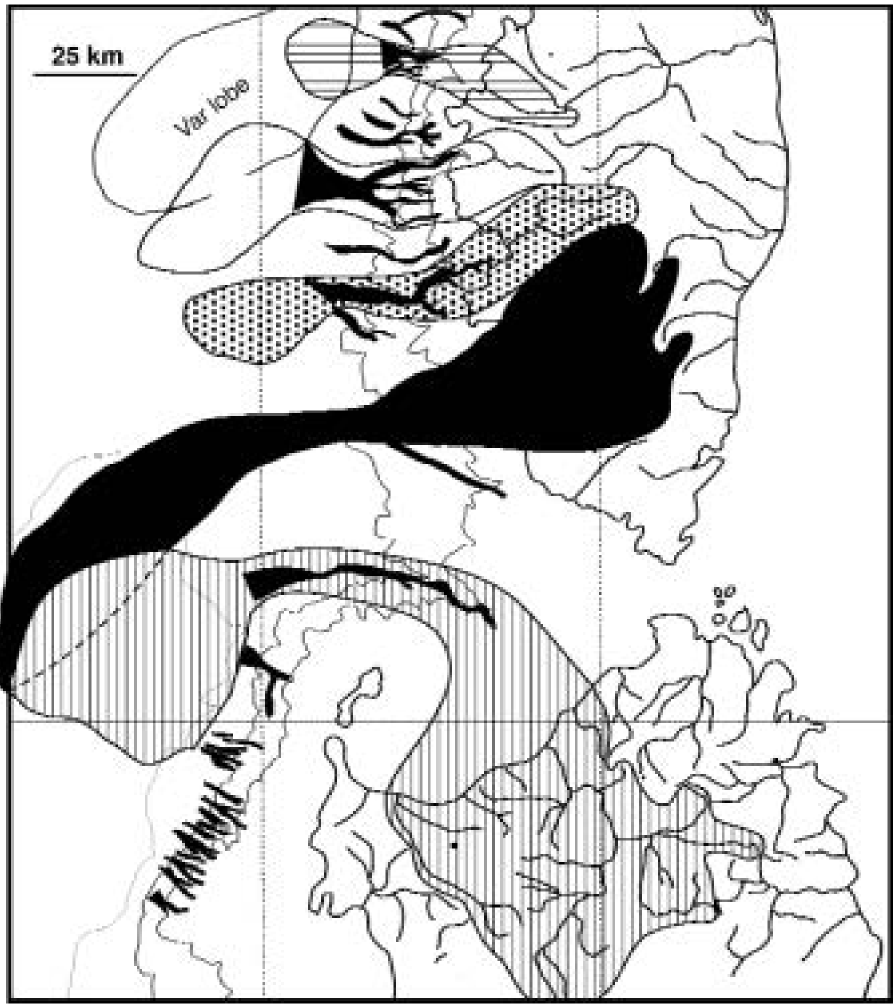

Fig. 3. Subaerial drainage basins and depositional systems west of Corsica and Sardinia (Kenyon *et al.*, in press).

TYPE 4. Mature lowest input types have tributary “canyon like” feeder systems and a lesser development of, or no, channel-levee systems. Channels are relatively straight, maximum channel gradients are generally over 1 in 70 and width to depth ratios are over 150 near the ends of channels. Channel-mouth sand lobes are common and extensive and usually detached from their channel by a zone of erosion. Flows are less frequent even at times of lower sea level, possibly fewer than 1 per 1000 years. There are many such systems in the Mediterranean, typified by the tributary canyon systems west of Corsica and northern Sardinia (Kenyon *et al.*, in press). These have subaerial drainage basins of less than 1,000 km^2 and straight, wide canyons. Short leveed channels, with a width to depth ratio of greater than 100, terminate at or just beyond the base of the steep slope. Maximum channel gradients are greater than 1 in 10. The floors of these canyons

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are acoustically rough, as first observed for similar canyon systems off Algeria in the first recorded sidescan sonar work in the deep Mediterranean Sea (Belderson *et al.*, 1970). The passage of fast moving turbidity currents leave scour holes and mobile gravel deposits in the western Corsican canyon axes. The rough canyon floors off Algeria were also attributed to the passage of turbidity currents such as that caused by the Orleansville earthquake, where speeds of up to 80km per hour are reported (Heezen and Ewing, 1955). Beyond canyon mouths there are lobe shaped areas (Fig. 3) where braid like strong backscattering patterns are found together with weak backscattering patterns. These backscattering patterns correspond to sand sheets that have been cored in places and found to be up to 3m thick (Kenyon *et al.*, in press) (Fig. 4).

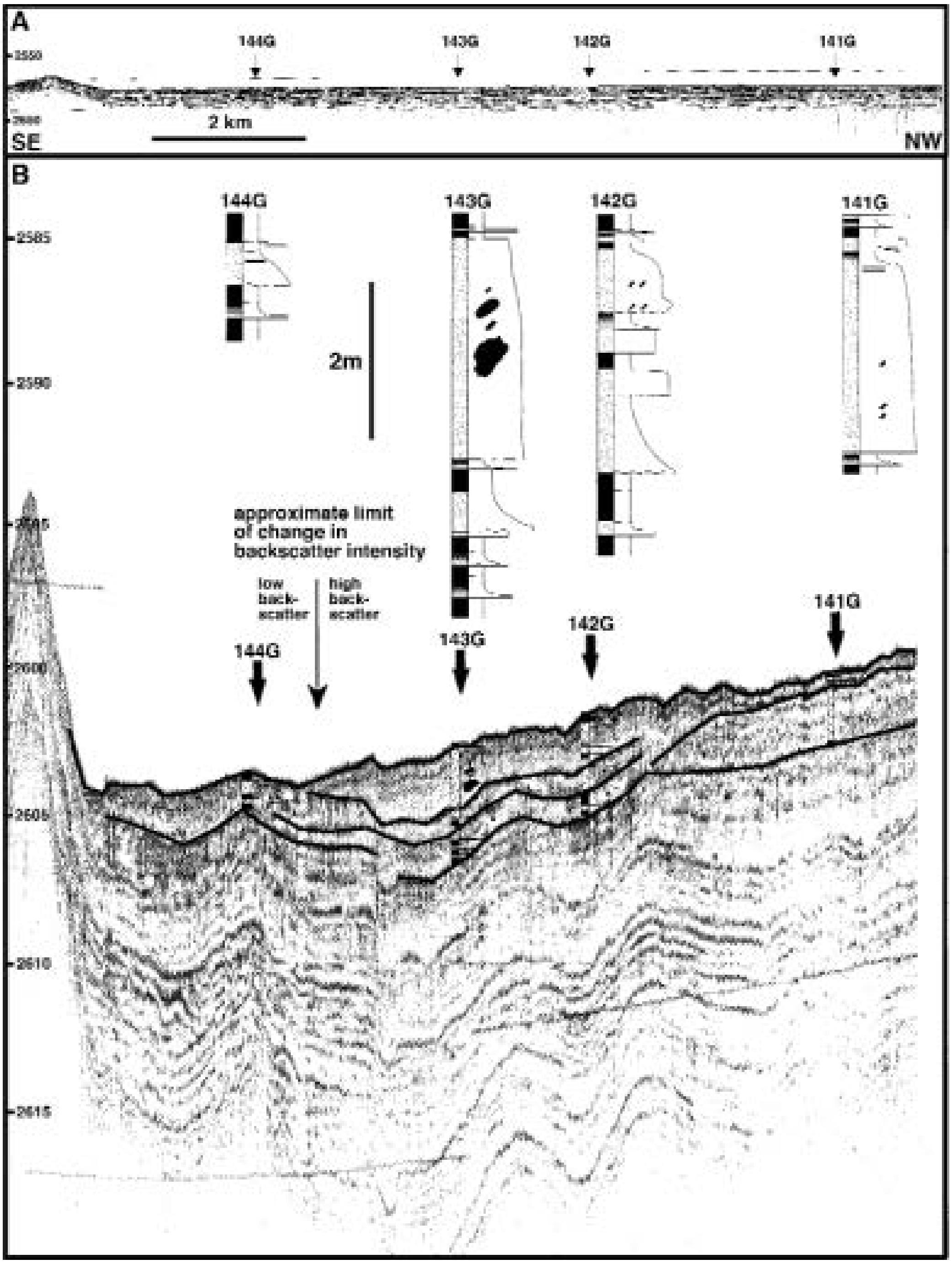

Fig. 4. 3.5kHz deep towed profile from a lobe west of Corsica. By increasing the vertical exaggeration extensive lenses can be identified. Cores of up to 3m of fining upward coarse sand are correlated with the lenses. TTR cruise 4 (Limonov *et al.*, 1995).