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Late Neogene evolution of the Taza–Guercif Basin (Rifian Corridor, Morocco) and implications for the Messinian salinity crisis

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Abstract

Magnetostratigraphic and biostratigraphic results are presented from Neogene deposits in the Taza–Guercif Basin, located at the southern margin of the Rifian Corridor in Morocco. This corridor was the main marine passageway which connected the Mediterranean with the Atlantic during Messinian times. Correlation of the biostratigraphy and polarity sequence of the Taza–Guercif composite section to the astronomical time scale, allows an accurate dating of three subsequent events in the Rifian Corridor. (1) The oldest marine sediments marking the opening of the Rifian Corridor were deposited at 8 Ma. At this age, a deep (600 m) marine basin developed in the Taza–Guercif area, marked by deposition of precession-controlled turbidite–marl cycles. (2) Paleodepth reconstructions indicate that a rapid (5 m/ka) shallowing of the marine corridor took place at the Tortonian/Messinian boundary, at an age of 7.2 Ma. This shallowing phase is primarily related to active tectonics, although a small glacio-eustatic sea level lowering also took place. (3) The Taza–Guercif Basin was emergent at an age of 6.0 Ma and, subsequently, continental sedimentation continued well into the Early Pliocene. We suggest that shallowing and restricting the marine passageway through the Rifian Corridor actually initiated the Messinian salinity crisis, well before the deposition of the Messinian evaporites in the Mediterranean.

Keywords: Messinian; Morocco; calcareous plankton; biostratigraphy; magnetostratigraphy; geodynamics

1. Introduction

Messinian sequences in the Mediterranean start with open-marine marls, followed by diatomites (Tripoli Formation), limestones (Calcare di Base), thick evaporites (Lower Evaporites) and terminate with brackish/marine to continental deposits (Lago Mare/Upper Evaporites). This succession of facies types reflects the progressive closure of the Mediterranean gateways, which first resulted in a salinity increase and, later, in the complete isolation of the Mediterranean during the final stage of the so-called 'Messinian salinity crisis'. Accurate time control on this Messinian succession is still remarkably poor,

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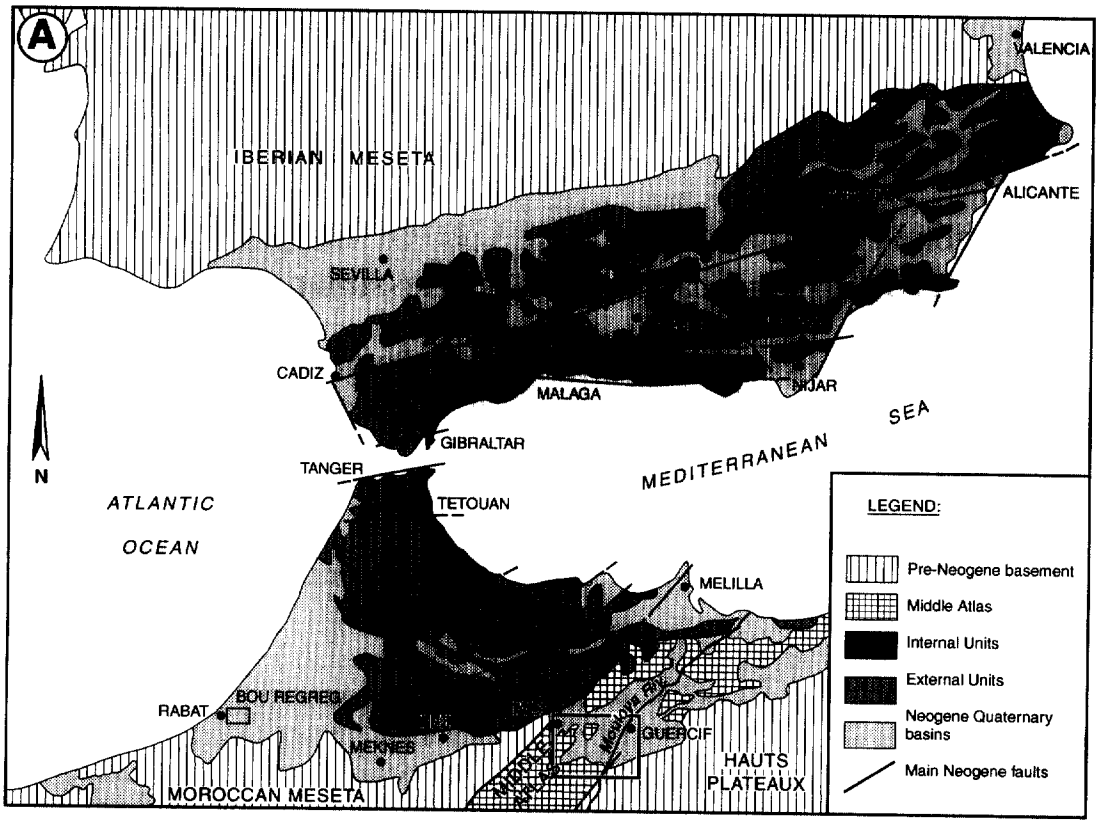
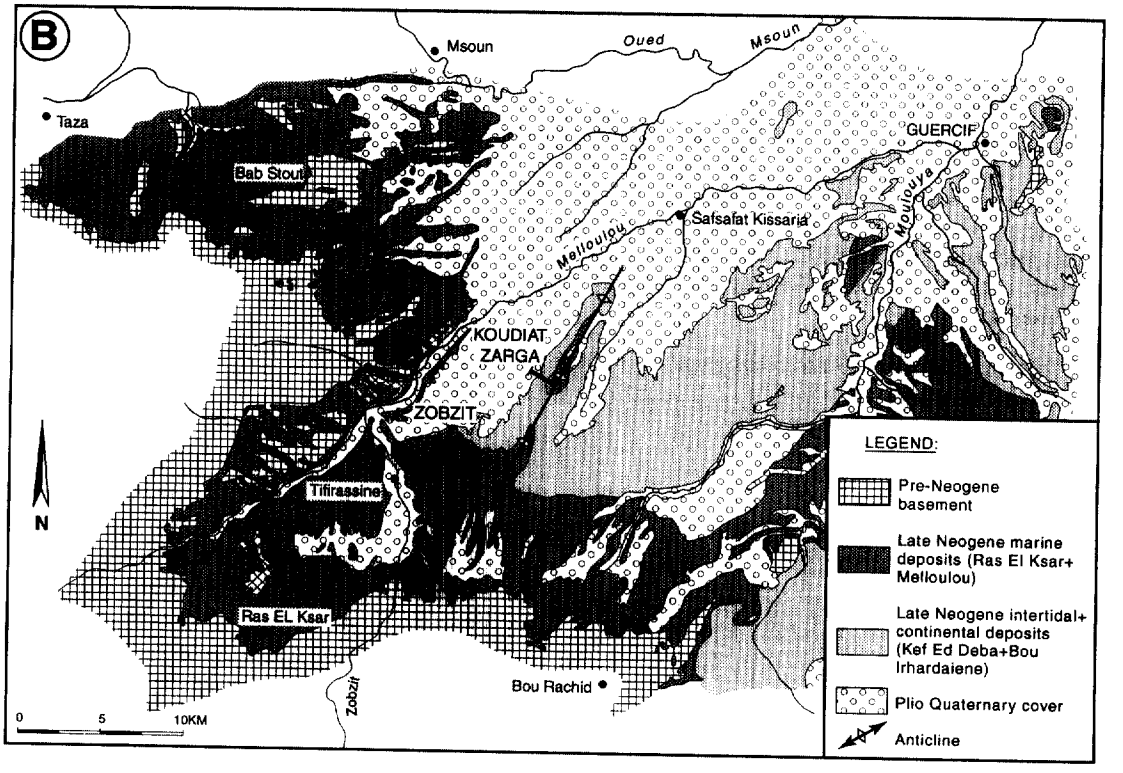

A
IBERIAN MESETA
VALENCIA
ALICANTE
SEVILLA
CADIZ
MALAGA
ORJAR
GIBRALTAR
TANGER
TETOUAN
MEDITERRANEAN SEA
ATLANTIC OCEAN
N
MELILLA
RABAT
BOU REGREG
MEKNES
GUERCIF
HAUTS PLATEAUX
MOROCCAN MESETA
LEGEND:
Pre-Neogene basement
Middle Atlas
Internal Units
External Units
Neogene Quaternary basins
Main Neogene faults


B
Msoun
Oued Msoun
Taza
Bab Stout
GUERCIF
Safsafat Kissaria
Melloulou
Moulouya
KOUDIAT
ZARGA
ZOBZIT
Tifirassine
Ras EL Ksar
Zobzit
Bou Rachid
N
0 5 10KM
LEGEND:
Pre-Neogene basement
Late Neogene marine deposits (Ras El Ksar+ Melloulou)
Late Neogene intertidal+ continental deposits (Kef Ed Deba+Bou Irhardaiene)
Plio Quaternary cover
Anticline

resulting in widely different and controversial models for this salinity crisis (e.g. Müller and Hsü, 1987; Benson et al., 1991; Hilgen et al., 1995; Clauzon et al., 1996; Vai, 1997).

The Rifian Corridor in Morocco is regarded as the most important passageway connecting the Messinian Mediterranean to the Atlantic. The present-day connection through the Strait of Gibraltar did not exist (Hsü et al., 1973) and no evidence is available of a Messinian passageway through the Betic Corridor at Messinian times (F.J. Sierro, pers. commun., 1997). Furthermore, marine connections with the Indian Ocean, through the Persian Gulf or the Red Sea, are poorly constrained and considered to have closed between 13 and 11 Ma (Yilmaz, 1993; Jacobs et al., 1996). In the Rifian Corridor, marine sediments of Late Miocene age extend from the Atlantic margin near Rabat (Bou Regreg area), via a series of interconnected basins near Meknes, Fez, Taza and Guercif, towards the Mediterranean near Melilla (Fig. 1A). Progressive closure of the Rifian Corridor during the Messinian was most likely caused by a complex interplay of both tectonic and glacio-eustatic processes (Benson et al., 1991; Hodell et al., 1989; 1994). The age of the final closure, however, is poorly constrained, because large-scale submarine sliding of olistolith complexes in the region between Fez and Taza cover the youngest marine sediments in this area (Benson and Rakic-El Bied, 1991).

To study the history of the marine passageway through Morocco, we selected the Taza–Guercif Basin, located at the southern margin of the Rifian Corridor (Fig. 1A). This basin contains a stratigraphic sequence of late Tortonian marine marls to continental deposits of unknown age (Bernini et al., 1992, 1994). The sequence reveals information on the opening and closure of the Rifian Corridor, although it is not located in the central part of this passageway. In this paper, we discuss the magnetostratigraphic and biostratigraphic data from the succession in the Taza–Guercif Basin. Correlation with the well-dated biochronology of the Mediterranean and with the astronomical polarity time scale (APTS) reveals when marine sedimentation started and ended in the Taza–Guercif Basin. Furthermore, the results will be discussed in the light of the Messinian salinity crisis in the Mediterranean.

2. The Taza-Guercif composite section

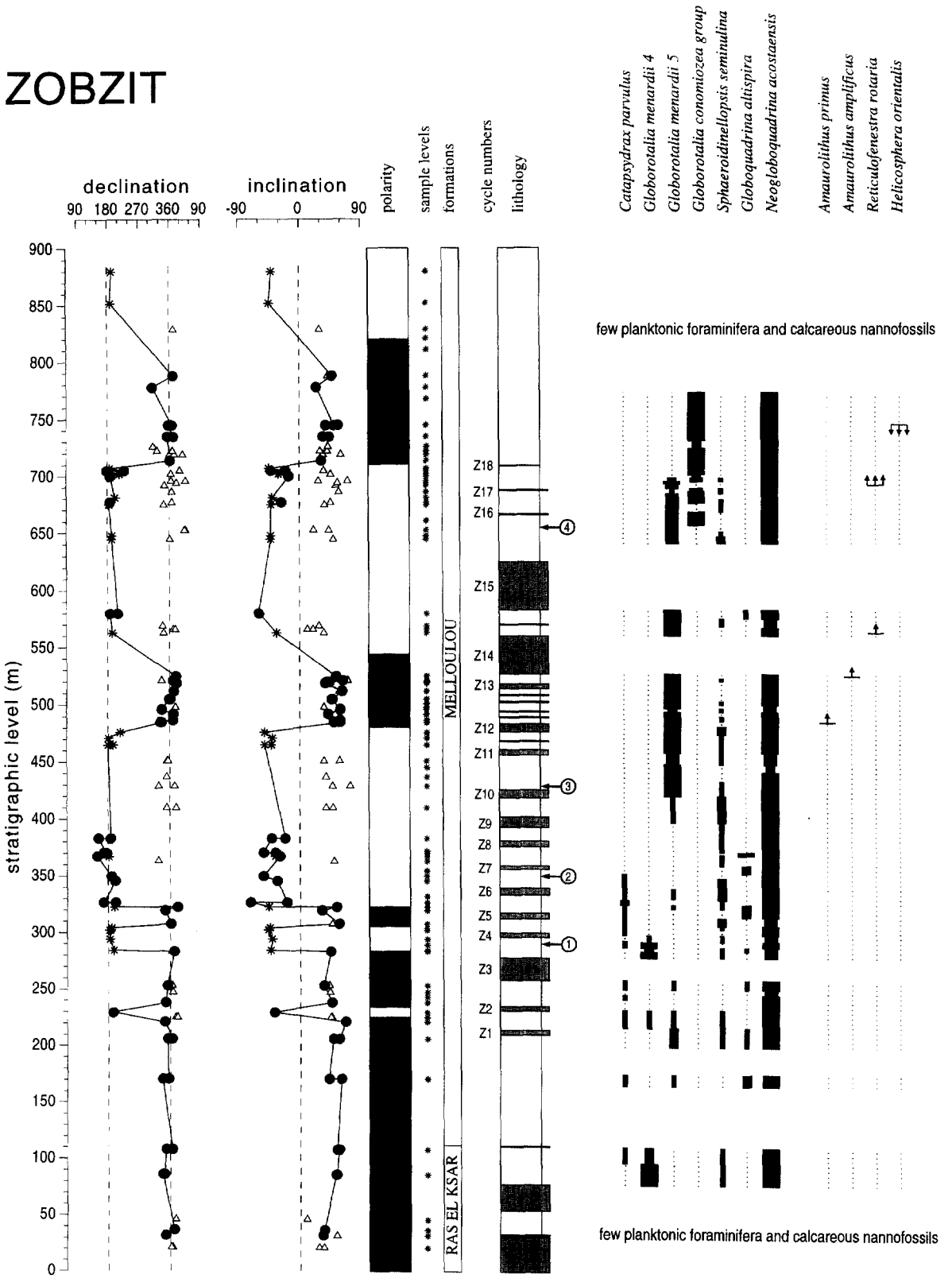
The Rifian Corridor is regarded as a residual foreland basin that developed, south of the external Rifian thrust front, on the central Moroccan Meseta and Middle Atlas. The Taza–Guercif Basin is located at the southern margin of the Rifian Corridor, on a previously emerged part of the Middle Atlas (Fig. 1A). It shows a more complex tectono-sedimentary history because of its position close to the Moulouya Belt shear zone and the arcuate Rifian front (Boccaletti et al., 1990). The Neogene deposits of the Taza–Guercif Basin consist of a thick sequence of deep to shallow marine and continental deposits. The succession was first described by Benzaquen (1965), followed by Coletta (1977) and Wernli (1988). Partly based on these results, Bernini et al. (1992, 1994) made a new stratigraphic subdivision in five lithostratigraphic units.

For our study, we sampled two sections including four of the five units distinguished by Bernini et al. (1992, 1994). The Zobzit section consists of the marine Ras El Ksar and Melloulou Formations and is exposed along the east-bank of the Zobzit River (Fig. 1B). The Koudiat Zarga section consists of the near-shore to intertidal Kef Ed Deba Formation and the continental Bou Irhardaiene Formation and is located in the west-flank of the Safsafat anticline (Fig. 1B). We did not sample the oldest unit (Draa Sidi Saada Formation) of Bernini et al. (1994), which consists of continental conglomerates and breccias.

2.1. The Zobzit section

The base of the Zobzit section consists of shallow marine sandstones and siltstones of the Ras El Ksar Formation (0–100 m; Fig. 2), which on-

Fig. 1. (A) Simplified structural sketch-map of the Gibraltar Arc (modified after Boccaletti et al., 1990) showing the locations of the Taza–Guercif Basin, the Bou Regreg area, and the Fortuna, Sorbas and Nijar Basins. (B) Stratigraphic map of the Taza–Guercif Basin (modified after Bernini et al., 1994) indicating the sample trajectories of the Zobzit and Koudiat Zarga section.


ZOBZIT
declination
90 180 270 360 90
inclination
-90 0 90
polarity
sample levels
formations
cycle numbers
lithology
Catapsydrax parvulus
Globorotalia menardii 4
Globorotalia menardii 5
Globorotalia conomiozea group
Sphaeroidinellopsis seminulina
Globoquadrina altispira
Neogloboquadrina acostaensis
Amaurolithus primus
Amaurolithus amplificus
Reticulofenestra rotaria
Helicosphera orientalis
stratigraphic level (m)
900
850
800
750
700
650
600
550
500
450
400
350
300
250
200
150
100
50
0
MELLOULOU
RAS EL KSAR
Z18
Z17
Z16
Z15
Z14
Z13
Z12
Z11
Z10
Z9
Z8
Z7
Z6
Z5
Z4
Z3
Z2
Z1
few planktonic foraminifera and calcareous nannofossils
few planktonic foraminifera and calcareous nannofossils

lap over the Jurassic basement of the Middle Atlas. These deposits mark the opening of a marine passageway through Morocco. The Ras El Ksar Formation gradually passes into the blue marls of the Melloulou Formation (100–1200 m). These classical 'blue marls' of Morocco can be traced through the entire Rifian Corridor. The Melloulou Formation shows a cyclic alternation of blue marls and sandy turbidites (Fig. 2). The turbidites are organised in lenticular sandy bodies ranging in thickness from less than 1 to more than 20 m and their provenance is from the south. These turbidite bodies extend laterally up to some hundred meters passing into yellow sandy marl layers. In our composite stratigraphic column, we counted a total number of fifteen main turbidite bodies (numbered Z1–Z15, Fig. 2). Some minor turbidites, usually at irregular positions, are not included in the numbering. Higher up in the Melloulou Formation (700 m), the turbidites become thinner (Z16–Z18) and finally disappear, although yellow sandy marls still remain being cyclically intercalated in the blue marls. At approximately 700 m, the blue marls start to contain secondary gypsum crystals. These gypsiferous marls pass gradually, via a number of oyster beds, into the near-shore to intertidal sediments of the Kef Ed Deba Formation. The contact between the two formations is, unfortunately, poorly exposed in the area. The presence of similar oyster beds at both sides of the contact suggests stratigraphic continuity, even though elsewhere a minor unconformity has been observed.

2.2. The Koudiat Zarga section

The Koudiat Zarga section is located in the west flank of the Safsafat anticline and consists of the Kef Ed Deba and Bou Irhardaiene Formations. The sediments of the Kef Ed Deba Formation consist of cyclic alternations of yellowish to reddish sands and marls (0–100 m; Fig. 3) with a reddish oyster-bearing conglomerate at the top. Facies analysis of the Kef Ed Deba Formation suggests a near-shore environment which passes upwards into an alluvial tidal flat or an inner delta front environment (Bernini et al., 1994). An erosional unconformity (at 100 m; Fig. 3) separates this succession from the overlying Bou Irhardaiene Formation. The latter formation (100–600 m) consists of reddish fluvial conglomerates, sands, marls, and whitish lacustrine limestones, indicative of a continental environment.

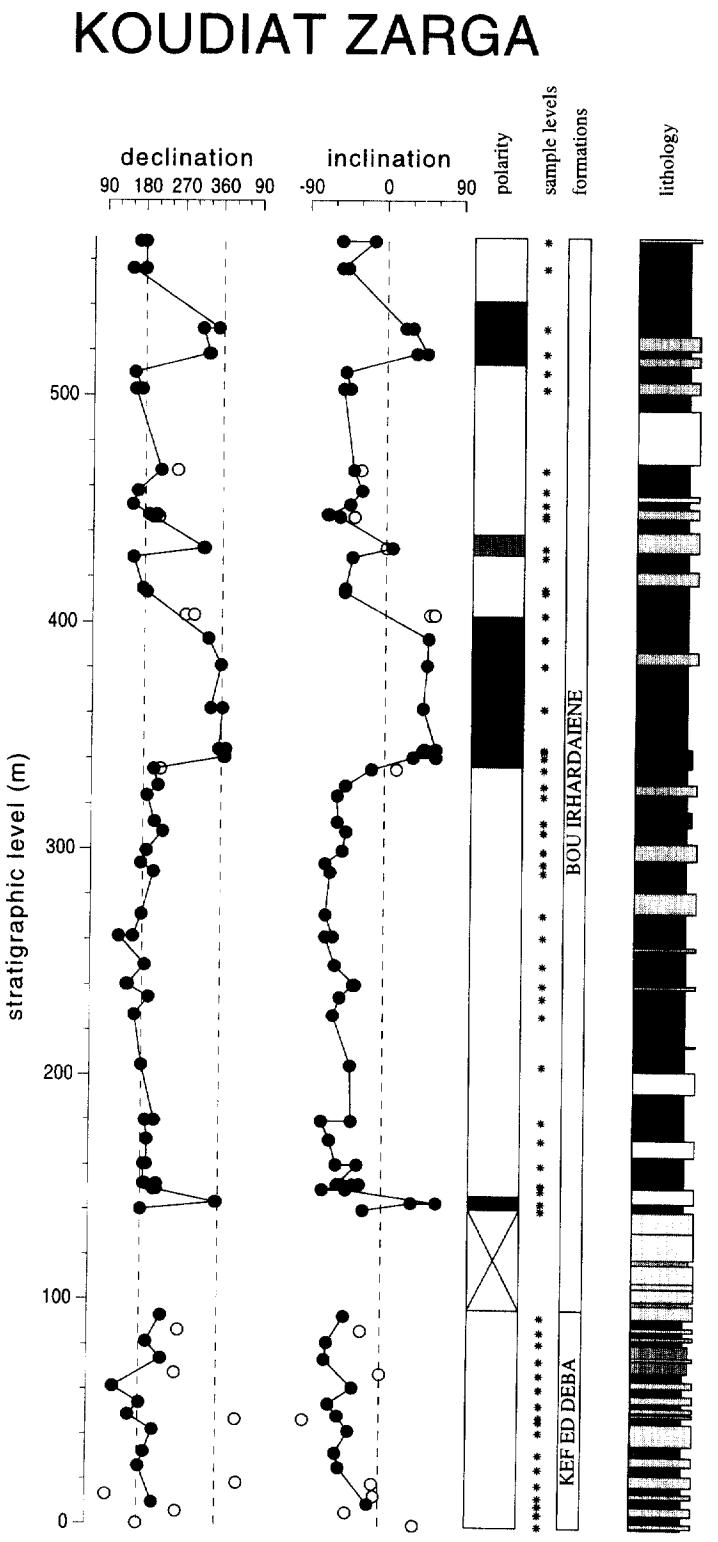
2.3. Sampling

Standard paleomagnetic cores were taken with an electric drill and a generator as power supply. As a routine procedure, we removed the weathered surface to drill in sediments as fresh as possible. In the Melloulou Formation, additional oriented handsamples were taken which were drilled with compressed air at the paleomagnetic laboratory Fort Hoofddijk. In the exposures along the Zobzit River, we sampled 6 levels in the Ras El Ksar Formation and 93 levels in the Melloulou Formation; we sampled at least 3 levels in the marls between each of the main turbidite bodies. We did not sample the turbidites. At the Safsafat anticline, we took 20 levels in the Kef Ed Deba Formation and 43 in the Bou Irhadaiene Formation, mainly in the marls and clays.

3. Biostratigraphy

The biostratigraphic analysis of the Zobzit section is based on planktonic foraminifera from the larger than 125 μm fraction and on calcareous nanno-

Fig. 2. Magnetostratigraphic and biostratigraphic data, lithology, sample positions and formations (after Bernini et al., 1994) in the Zobzit section. In the polarity column black (white) denotes normal (reversed) polarity interval. Dots represent reliable directions showing a linear decay to the origin, asterisks represent directions obtained by applying the great circle method (McFadden and McElhinny, 1988), triangles represent secondary directions showing a clustering. Lithology consists of marine marls (white) and sandy (turbiditic) layers (grey) numbered Z1 to Z18. The Ras El Ksar Formation consists of shallow marine sandstones and siltstones. Distribution of selected planktonic foraminiferal taxa is given in four semi-quantitative categories (indicated by increasing bar thickness: trace (<3 specimens per 9 fields of picking tray), rare (3–10), common (10–30) and frequent (>30). Planktonic foraminifera bioevents are: *1* = Last Common Occurrence (LCO) of *G. menardii 4*; *2* = Last Occurrence of *C. parvulus*,; *3* = First occurrence (FO) of *G. menardii 5*; *4* = First Regular Occurrence (FRO) of the *G. conomiozea* group. Five bioevents are recognised in the nannofossil record: FO of *Amaurolithus primus*, FO of *Amaurolithus amplificus*, FO of *Reticulofenestra rotaria*, FCO of *Reticulofenestra rotaria*, and LCO of *Helicosphaera orientalis*.


KOUDIAT ZARGA
declination
90 180 270 360 90
inclination
-90 0 90
polarity
sample levels
formations
lithology
stratigraphic level (m)
0
100
200
300
400
500
BOU IRHARDAIENE
KEF ED DEBA

plankton. Diverse assemblages of these calcareous microfossils occur up to 770 m. Above this level, these microfossils are rare and not useful for biostratigraphic analysis. The planktonic foraminiferal record of the Zobzit section contains four bioevents which are widely used for intra-Mediterranean correlations: (1) the last common occurrence (LCO) of *Globorotalia menardii* 4 at 290 m; (2) the last occurrence (LO) of *Catapsydrax parvulus* at 350 m; (3) the first occurrence (FO) of *Globorotalia menardii* 5 at 430 m; and (4) the first regular occurrence (FRO) of the *Globorotalia conomiozea* group at 650 m (Fig. 3).

Globorotalia menardii 4 and 5 are labels used for two Late Miocene groups of Mediterranean menardine globorotaliids. These two groups differ in (1) chamber outline (being less elongate in group 5), (2) the angle formed by the proximal part of the intercameral and spiral suture (being larger in group 5), (3) the degree of convexity of the umbilical side (being smaller in group 5), and (4) the coiling (predominantly left in group 4 and right in group 5) (see also Zachariasse, 1979). The *G. conomiozea* group incorporates a planoconvex group of keeled globorotaliids with a crescent-shaped chamber outline. This group contains two end-member morphologies, which are labelled *Globorotalia miotumida* (flat) and *Globorotalia conomiozea* (conical). The *G. conomiozea* group in the Zobzit section consists of the flat *G. miotumida* type. *Catapsydrax parvulus* is a small-sized species, of which the taxonomic status with respect to *Globorotaloides falconarae* is not entirely solved.

In the Mediterranean, *G. menardii* 4 (LCO is at 7.512 Ma, Hilgen et al., 1995) is succeeded by *G. menardii* 5 (FO is at 7.355 Ma, Hilgen et al., 1995). *Globorotalia menardii* 5 is subsequently replaced by the *G. conomiozea* group at a level which is equated with the Tortonian/Messinian boundary (age 7.240 Ma, Hilgen et al., 1995). In the Zobzit section, *G. menardii* 5 shows stratigraphic overlap with both *G. menardii* 4 and the *G. conomiozea* group (Fig. 2). *Globorotalia menardii* 5 is considered to be a warm-water element because its morphology is close to the low-latitude *Globorotalia menardii*. The somewhat expanded stratigraphic range of *G. menardii* 5 compared to that in the Mediterranean seems therefore to be related to the more southern latitudinal position and the proximity to the open ocean of the Taza–Guercif Basin during the Late Miocene. A similar explanation may hold for the warm-water species *Sphaeroidinellopsis seminulina* and *Globoquadrina altispira* (Fig. 2), which occur more frequently in the Zobzit section than in time-equivalent sections in the Mediterranean.

The LCO of *G. menardii* 4, FO of *G. menardii* 5, and FRO of the *G. conomiozea* group, however, are well-distinguishable on the basis of their frequency distribution (Fig. 2). Together with the LO of *C. parvulus*, these four planktonic foraminiferal bioevents can be used to correlate the Zobzit section to Late Miocene sections in Greece and Italy (Krijgsman et al., 1995, 1997). Noteworthy is furthermore the observation that *Neogloboquadrina acostaensis* is predominantly (>90%) left coiled in the Zobzit section, which is in agreement with its coiling in time-equivalent sections in the Mediterranean (Krijgsman et al., 1995).

In addition to the four planktonic foraminiferal bioevents, we distinguished five bioevents in the calcareous nannofossil record. These five bioevents are: (1) FO of *Amaurolithus primus* at 485 m, (2) FO of *Amaurolithus* cf. *amplificus* at 530 m, (3) FO of *Reticulofenestra rotaria* at 550 m, (4) FCO of *Reticulofenestra rotaria* at 690 m, and (5) LCO of *Helicosphaera orientalis* at 750 m.

4. Magnetostratigraphy

Paleomagnetic analyses were performed by stepwise thermal demagnetisation with small temperature increments of 30–50°C. Rock magnetic experiments show that both magnetite and hematite contribute to the natural remanent magnetisation

Fig. 3. Magnetostratigraphy, sample positions, formations, and lithology of the Koudiat Zarga section. Black (white) denotes normal (reversed) polarity interval. Dots represent reliable directions showing a linear decay to the origin; circles represent unreliable directions of low-intensity samples which only show a scatter around the origin. Lithology consists of lacustrine limestones (white), sandy layers (light grey), reddish/brownish clays and marls (dark grey).

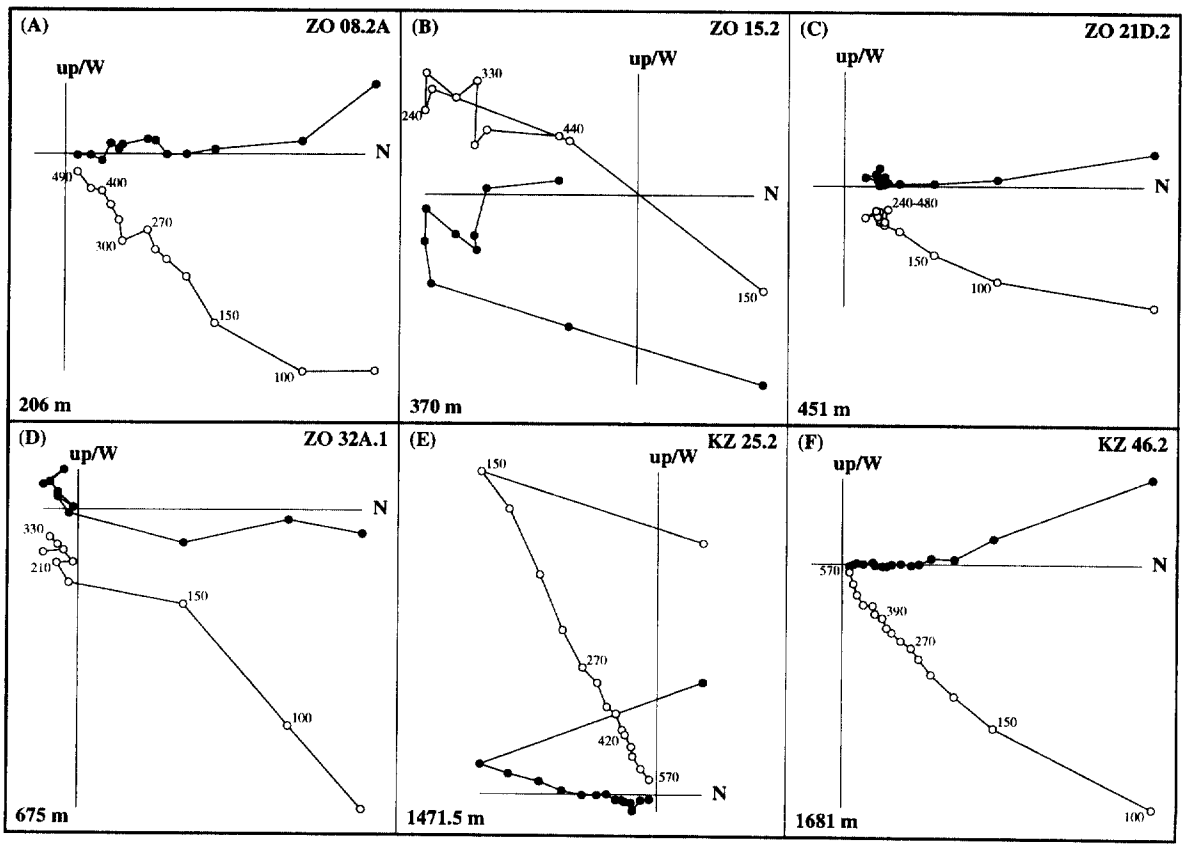


Fig. 4. Orthogonal projections of stepwise thermal demagnetisation of selected samples (*ZO* = marine sediments; *KZ* = continental sediments). Dots (circles) represent the projection of the NRM vector end-point on the horizontal (vertical) plane. Values represent temperatures in degrees Celcius; stratigraphic levels are in the lower left corner.

(NRM). Demagnetisation diagrams of the samples from the Zobzit section are of mixed quality. For samples with relatively high NRM intensity (0.1–1 mA/m) directions and polarities can usually be reliably determined (Fig. 4A,B). Low intensity (<0.1 mA/m) samples often show a clustering, with temperatures ranging from 200 to 500°C (Fig. 4C). We have interpreted these clusters as secondary, weathering-induced, components. Some samples do not show a linear decay to the origin but tend to 'move' to the reversed quadrant of the demagnetisation diagram (Fig. 4D). Plotted on an equal-area diagram, remanence vectors progress along a great circle toward a southerly upward direction, indicating removal of a normal phase from a reversed primary component. In these cases, a best-fitting great circle plane is determined according to the method of McFadden and McElhinny (1988). The NRM directions and polarity zones (Fig. 2) show that nine polarity reversals are recorded in the Zobzit section.

The reddish intertidal and continental clays and marls of the Koudiat Zarga section are characterised by higher NRM intensities (1–10 mA/m). Demagnetisation diagrams generally show a linear decay to the origin, after removal of a secondary present-day direction below temperatures of 200°C (Fig. 4E,F). The NRM is totally removed below 600°C which indicates that magnetite is the dominant carrier of the magnetisation. The Kef Ed Deba Formation only reveals reversed polarities, the Bou Irhardaiene Formation reveals three intervals of normal polarity, although the lowermost interval only consists of one single level (Fig. 3). The relatively large dip of the section (40–60°) was useful to discriminate between

primary and secondary components. One level (440 m) in the upper part of the section shows an anomalous result. Declinations suggest a normal polarity but inclinations, after tilt correction, appear to be too shallow for a primary normal component.

5. Correlation to the APTS

5.1. The Zobzit section

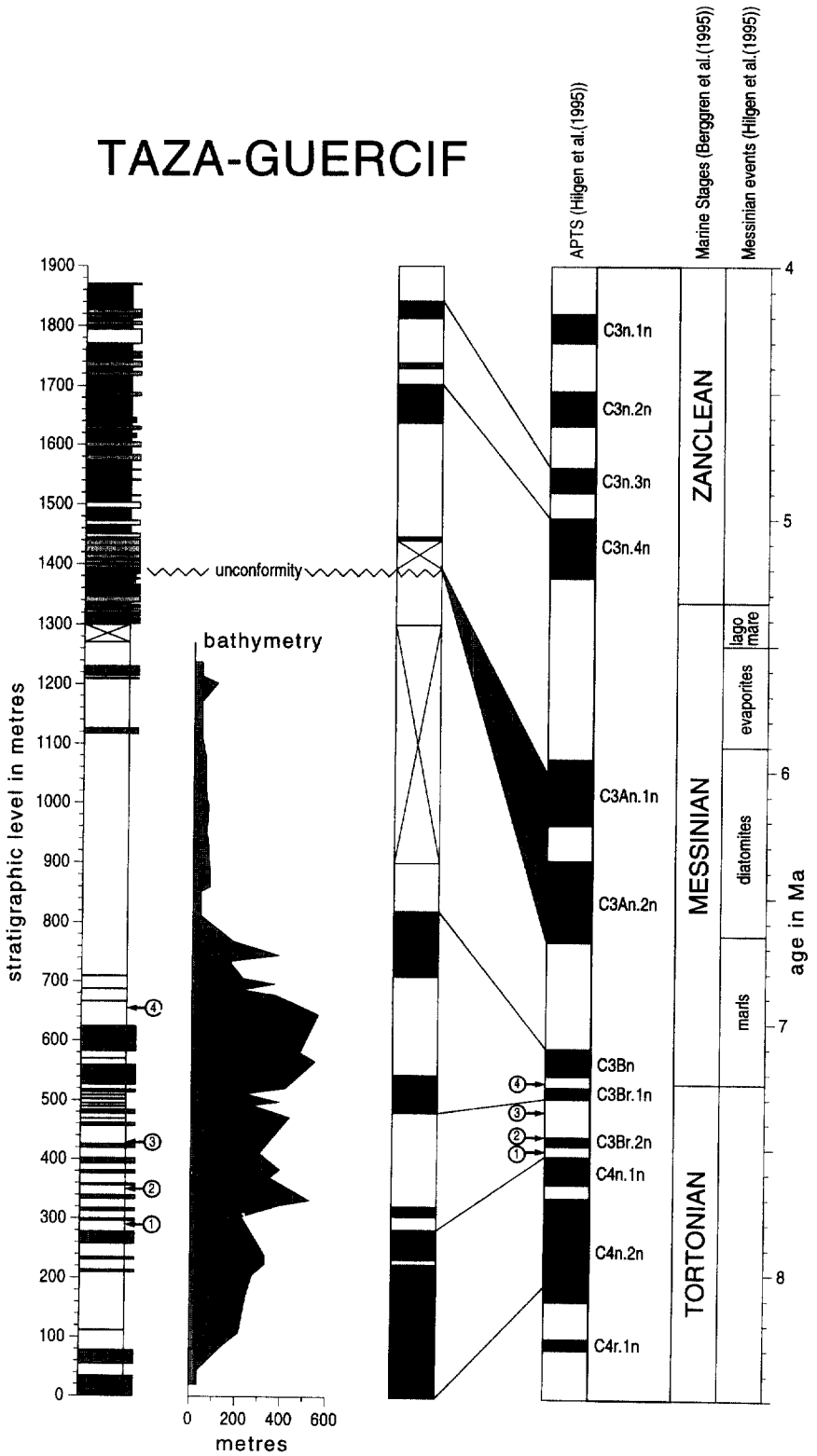
The planktonic foraminiferal bioevents can be used to correlate the polarity sequence of the Zobzit section to the astronomically tuned magnetobiostratigraphies of Late Miocene key sections in Greece and Italy (Krijgsman et al., 1995, 1997; Hilgen et al., 1995) and hence to the astronomical polarity time scale of Hilgen et al. (1995) (Fig. 5). The reversed interval with the LCO of *Globorotalia menardii* 4 corresponds to chron C3Br.3r. Consequently, the normal zone in the basal part of the Zobzit section is correlated to chron C4n.2n. The reversed interval with the LO of *Catapsydrax parvulus* and the FO of *Globorotalia menardii* 5 is identified as chron C3Br.2r, whereas the reversed interval with the FRO of the *Globorotalia conomiozea* group corresponds to chron C3Br.1r (Fig. 5). These results are generally in agreement with the magnetobiostratigraphy of the Bou Regreg area at the Atlantic margin of Morocco (Sale drill-hole; Hodell et al., 1994).

The magnetostratigraphic correlation of the Zobzit section to the APTS permits an age estimate for the calcareous nannofossil bioevents in this section. These ages are: 7.30 Ma for the FO of *Amaurolithus primus*; 7.27 Ma for the FO of *Amaurolithus* cf. *amplificus*; 7.25 Ma for the FO of *Reticulofenestra rotaria*; 7.23 Ma for the FCO of *Reticulofenestra rotaria*, and 7.15 Ma for the LCO of *Helicosphaera orientalis*. Work in progress on the Upper Miocene sections Faneromeni (Crete) and Monte del Casino (northern Italy) suggests slightly older ages for the FO's of *Amaurolithus primus* and cf. *amplificus*, and *Reticulofenestra rotaria*. This age discrepancy is probably caused by the lower stratigraphic resolution and the (much) higher sedimentation rate in the Zobzit section.

Calculating the periodicity of the main turbidite bodies, using the astronomical ages for the LO of *G. menardii* 4 (7.512 Ma) and FRO of the *G. conomiozea* group (7.240 Ma) (Hilgen et al., 1995; Krijgsman et al., 1997), reveals an average periodicity of 22.6 Ma. This suggests that the turbidite deposition is controlled by climatic variations related to the precessional component of orbital forcing. With respect to the phase relationship, we suggest that the main turbidite bodies in the Zobzit section are in phase with the sapropels in the Mediterranean, which are related to maxima in northern summer insolation and thus to warmer periods and increased runoff (Lourens et al., 1996). Although the number of precession cycles between the planktonic foraminiferal bioevents is in good agreement with that obtained from time-equivalent sections in Crete and Italy (Krijgsman et al., 1995, 1997), there are (minor) discrepancies per individual polarity zone. This is probably caused by uncertainties in the true position of the magnetic reversals, related to the low stratigraphic resolution and the low intensity paleomagnetic signal.

5.2. The Koudiat Zarga section

The nearshore to intertidal deposits of the Kef Ed Deba Formation are entirely of reversed polarity. Stratigraphic continuity between the Kef Ed Deba Formation and the underlying shallow marine marls of the Melloulou Formation suggests that these polarities correlate with the reversed chron C3Ar. Hence, the top of the Kef Ed Deba Formation, i.e. the base of the unconformity, is not younger than 6.7 Ma (Fig. 5). The correlation of the polarity zones in the Bou Irhardaiene Formation (above the unconformity) is less unambiguous. The absence of additional age constraints in the Bou Irhardaiene Formation (no biostratigraphy, no clear cyclostratigraphy) allows only a tentative magnetostratigraphic correlation, assuming a constant sedimentation rate and no major hiatuses. The most conspicuous characteristic of the observed polarity column is the relatively long reversed interval in the lower part of the measured sequence, which is almost three times longer than the next (younger) normal interval. A first option in which the three normal zones correlate with C3An.2n, C3An.1n, and C3n.4n would imply unrealistic changes in sedimentation rate. The second option, in which the upper two normal zones correlate with C3An.1n, C3n.4n and C3n.4n agrees


TAZA-GUERCIF
stratigraphic level in metres
1900
1800
1700
1600
1500
1400
1300
1200
1100
1000
900
800
700
600
500
400
300
200
100
0
unconformity
bathymetry
0 200 400 600
metres
APTS (Hilgen et al.(1995))
C3n.1n
C3n.2n
C3n.3n
C3n.4n
C3An.1n
C3An.2n
C3Bn
C3Br.1n
C3Br.2n
C4n.1n
C4n.2n
C4r.1n
Marine Stages (Berggren et al.(1995))
ZANCLEAN
MESSINIAN
TORTONIAN
Messinian events (Hilgen et al.(1995))
lago mare
evaporites
diatomites
marls
age in Ma
4
5
6
7
8

reasonably well with the lengths of the polarity zones above the unconformity. This option is preferred because any other correlation to a younger part of the APTS is less convincing. The preferred option implies that the unconformity at ca. 1400 m represents a hiatus of some 700 ka.

6. Discussion and conclusions

6.1. Opening of the Rifian Corridor

The Rifian orogeny started when the westward relative movement of the Alboran microplate collided with the African plate (Dercourt et al., 1986). This phase was accompanied by metamorphism, with its apex being radiometrically ($^{40}Ar/^{39}Ar$) dated at 8 Ma (Monie et al., 1984). The Rifian Corridor represents a remnant part of the foredeep basin of the Rif Orogen, which is related to the southward shifting of the chain–foredeep system (Boccaletti et al., 1990). The Taza–Guercif Basin developed on the southern margin of the Rifian Corridor and is strongly influenced by transpressive movements along the Moulouya Belt shear zone (Boccaletti et al., 1990). The oldest marine sediments in this basin onlap over Jurassic substratum or Miocene fluvial conglomerates and have an extrapolated age of 8 Ma (Fig. 5). Because the evolution of the Taza–Guercif Basin is closely related to that of the Rifian Corridor, we assume that the opening of this corridor has an age close to 8 Ma. This age is in agreement with the major tectonic deformation phase in the Rifian Orogen.

Also in other areas, tectonic activity is recorded at more or less the same time. A major tectonic rotation phase of Calabria, inferred from paleomagnetic data, occurred between 8.6 and 7.6 Ma and is suggested to correspond with the opening of the Tyrrhenian Basin (Duermeijer et al., 1998). This is in agreement with an age of 7.8 Ma (recalibrated to Hilgen et al., 1995) for the oldest sediments in the Tyrrhenian Basin, east of Sardinia (Kastens et al., 1987). In the Northern Apennines, turbidite deposition (Formazione Marnoso–Arenacea) suddenly ceased at an age of 8 Ma, followed by deposition of open marine marls (Krijgsman et al., 1997). Furthermore, it has been suggested that Tibet was uplifted substantially (±1000 m) and abruptly at about 8 Ma, coinciding with folding of the Indo-Australian plate (Molnar et al., 1993).

6.2. Shallowing of the Taza–Guercif Basin

The paleodepth reconstruction in Fig. 5 is based on counting the ratio planktonic to benthic foraminifera and using the paleodepth equation of Van der Zwaan et al. (1990). This reconstruction indicates a rapid deepening to outer shelf–upper slope depths in the basal part of the section, and a rapid shallowing to near-shore depths between 7.2 and 7.1 Ma (i.e. during the earliest Messinian). This near-shore facies passes into the intertidal deposits of the Kef Ed Deba Formation, and finally into the continental deposits of the Bou Irhardaiene Formation. The rapid shallowing by at least 400 m of the Taza–Guercif Basin most likely reflects tectonic activity, whereby the rapid infilling probably is related to further advancement of the thrust front. A glacio-eustatic sea level lowering during the earliest Messinian may have added to this shallowing because the oxygen isotope record at the Atlantic side of the Rifian Corridor (Bou Regreg area) shows a distinct increase across the Tortonian/Messinian (T/M) boundary which is interpreted in terms of global ice volume increase (Hodell et al., 1994). If the enrichment (0.4‰) in the benthic foraminiferal $\delta^{18}O$ record at Bou Regreg is interpreted solely in terms of global ice volume increase, then it would correspond with a glacio-eustatic sea level fall of 40 m (using the calibration of Fairbanks and Matthews, 1978). The reconstructed relative sea level fall in the Taza–Guercif Basin between 7.2 and 7.1 Ma of at

Fig. 5. Lithology, bathymetry and polarity pattern of the Taza–Guercif composite section and correlation to the APTS (Hilgen et al., 1995), marine stages (Berggren et al., 1995) and the chronology of Messinian facies types (Hilgen et al., 1995). Chron nomenclature after Cande and Kent (1992). Bathymetry is based on the P/B ratio using the paleodepth equation of Van der Zwaan et al. (1990). Encircled numbers refer to planktonic foraminiferal bioevents (see caption to Fig. 2). Their position along the APTS follows from tuning Late Miocene sections to the astronomical solutions and summer insolation at 65°N (Hilgen et al., 1995).

least 400 m, thus contains a glacio-eustatic component of 40 m. The remaining shallowing of 360 m requires (1) uplift of the Taza–Guercif Basin or (2) an increase in the sedimentation rate to 360 cm/ka under continuous tectonic subsidence. Since the sedimentation rate between 7.2 and 7.1 Ma is ca. 115 cm/ka, uplift must have occurred during the earliest Messinian, possibly in concert with increased sedimentation rates as inferred from the increase in cycle thickness in the Bou Regreg area (Oued Akrech section; field data, 1996).

Hence, the rapid shallowing of the Taza–Guercif Basin during the earliest Messinian must be primarily driven by tectonic activity and seems to be related to uplift in the Rifian Corridor and associated deposition of olistromes. This process began in the late Tortonian or early Messinian and finally closed the corridor somewhere in the late Messinian (Benson et al., 1991). There is also evidence for tectonic activity in the Betic Orogen at this time. Santisteban and Taberner (1983) and Müller and Hsü (1987) report a shallowing of the Fortuna Basin during the T/M boundary interval. At the same time, however, the Sorbas and Nijar Basins show a rapid deepening because late Tortonian shallow marine calcarenites are overlain by early Messinian deep marine marls (Ott d'Estevou and Montenat, 1990; Sierro et al., 1997).

We believe that the rapid shallowing of the Rifian Corridor between 7.2 and 7.1 Ma actually initiated the Messinian Salinity Crisis, i.e. well before the deposition of evaporites. Shallowing or restricting the connection between the Mediterranean and Atlantic should have resulted in an increase in the residence time of the Mediterranean leading a.o. to an increase in salinity, providing that the excess of evaporation over fresh-water inflow remained equal. The residual of 2.5‰ between the early Messinian benthic and planktonic foraminiferal $\delta^{18}O$ record from Crete (Van der Zwaan and Thomas, 1980) and the time-equivalent benthic foraminiferal $\delta^{18}O$ record from Atlantic Morocco (Hodell et al., 1994), suggests an increase in salinity of the Mediterranean during the early Messinian (although we are aware of the fact that both records are based on different benthic foraminiferal species). The early Messinian $\delta^{13}C$ decrease of ca. 1‰ in the record of Atlantic Morocco is interpreted by Hodell et al. (1994) to represent the Late Miocene carbon shift. However, this early Messinian decrease in $\delta^{13}C$ is less than the time-equivalent $\delta^{13}C$ decrease in the Cretan record. A similar excess $\delta^{13}C$ decrease in the Mediterranean has been inferred by Van der Zwaan and Gudjonsson (1986) by comparing planktonic foraminiferal $\delta^{13}C$ records from the Mediterranean and the open ocean. This early Messinian excess $\delta^{13}C$ decrease might reflect an increase in the average age of the Mediterranean waters caused by an increase in the residence time. Underpinning of this model in which the residence time of the Mediterranean begins to increase at about the T/M boundary requires more high-resolution stable isotope and benthic foraminiferal records from in and outside the Mediterranean and their placing in a rigorous and undisputable time frame.

6.3. Closure of the Rifian Corridor

It is a commonly accepted scenario that the final closure of the Atlantic passageways isolated the Mediterranean from the open ocean during the latest Messinian. Correlation of the upper two normal polarity zones in the Koudiat Zarga section to chron C3n.4n and C3n.3n implies that the unconformity at ca. 1400 m represents a hiatus which at least spans the interval between the base of chron C3An.2n and the uppermost part of chron C3An.1n, i.e. roughly the interval between 6.7 and 6.0 Ma (Fig. 5). This correlation further implies that the Taza–Guercif Basin was emergent at about 6.0 Ma, which is well before the time the Mediterranean became isolated at the end of the Messinian (between 5.5 and 5.3 Ma according to Hilgen et al., 1995). Whether the entire Rifian Corridor became emerged at this time is still uncertain, and repeated influxes of Atlantic waters through a very shallow and narrow Corridor can not be excluded. It is, however, worth noting that the first reported mammal exchange between Africa and Europe took place at 6.1 Ma (Benammi et al., 1996) indicating that a continental passageway existed in the western Mediterranean well before the evaporite deposition.

If the entire Rifian Corridor was closed at about 6.0 Ma, then there must have existed another marine connection because the deposition of the Lower Evaporites, which is suggested by Hilgen et al., 1995 to have started at about 5.9 Ma, has taken place in a marine environment. This marine connection

could have been the Strait of Gibraltar or the Betic Corridor but data are not conclusive. Alternatively, a marine passageway to the Indian Ocean during the late Messinian cannot be completely excluded since fossil fishes of Indian Ocean–Red Sea affinity are found in the Lower Evaporites of northern Italy (Vai, 1997). Stratigraphic data from the eastern Mediterranean margin, however, are still extremely scarce.

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