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Palaeogeography, Palaeoclimatology, Palaeoecology 158 (2000) 353–370
PALAEO
www.elsevier.nl/locate/palaeo

Precession-related sapropelites of the Messinian Sorbas Basin (South Spain): paleoenvironmental significance

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Abstract

The Sorbas Basin (Southern Spain) is a narrow and elongated basin located in the Betic Corridor that formed the northern connection between the Atlantic and the Paleomediterranean during the late Miocene. In the centre of the basin, about 120 m of well-preserved cyclic marine sediments crop out, ranging in age from 7.2 to 6 Myr. These cycles make up the Abad Member and are an excellent example of precession control over sedimentation. In the upper part of the sequence (Upper Abad Member), the cycles include organic-rich laminated brown marls (sapropelites) alternating with homogeneous marls displaying diatomite layers. The OM-rich laminated marls are the westernmost sapropelites of the Mediterranean. This paper is a geochemical, mineralogical, sedimentological, and micropaleontological analysis of three cycles with sapropelites.

The data enable us to distinguish two main climatic scenarios. The climate, which dominated each of these scenarios, regulated the sedimentology and the physico-chemistry of the water column, resulting in the sapropelite/homogeneous marl cyclic alternations. The sapropelites developed during a climate characterised by a long period of temperate and humid conditions that occur at times of precession minima. These climatic conditions changed progressively to a subarid and drier weather, during which homogeneous sediments were deposited. The changes between these scenarios were gradual, as expected from precession dynamics. Our data suggest that the change from the subarid to the humid climate (transition from homogeneous marls to laminated sapropelites) took place at a faster rate or, at least, was recorded more rapidly by the sediments than the transition from the humid interval to the sub-arid period.

Keywords: cycles; Messinian; paleoceanography; paleoclimatology; sapropels; Sorbas Basin; Western Mediterranean

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0031-0182/00/$ - see front matter
PII: S0031-0182(00)00058-4

1. Introduction

In the centre of the Sorbas Basin (Southern Spain), well-preserved cyclic marine deposits are exposed. The cycles form the Abad Member (Ruegg, 1964) and are an excellent example of precession control over sedimentation. The overall thickness is about 120 m. The upper cycles have organic-rich laminated brown marls that constitute the westernmost sapropelites of the Mediterranean (Sierro et al. 1995, 1999; Zamarreño et al., 1995).

During the last decades, Mediterranean sapropels and sapropelites have been intensively studied in Holocene and late Pleistocene piston cores recovered in the Eastern Mediterranean, and in Pliocene and early Pleistocene land sections from Italy (Olausson, 1961; Rossignol-Strick et al., 1982; Calvert, 1983; De Lange and Ten Haven, 1983; Howell et al., 1990; Cita, 1991; Rohling and Hilgen, 1991; Rohling, 1994; Van Os et al., 1994; Rio et al., 1997; among others). Most of these studies propose an astronomical control over the formation of the sapropels by enhancing the run-off of the Nile and the North Mediterranean rivers at times of precession minima, when the perihelion occurs during the Northern Hemisphere summer. Kutzbach and Webb (1993) supported this interpretation by means of a numerical climatic model that indicates an increase in summer rains in the Northern Hemisphere during precession minima.

By contrast, the paleoceanography and paleoclimatology of the Miocene sapropelite have been much less studied, and the interpretations show some disagreement. Earlier studies (Cita and Grignani, 1982) interpreted Miocene sapropels as being linked to a general warm climate and a sluggish bottom water circulation. Van der Zwaan and Gudjonsson (1986) suggest that climate-induced salinity stratification associated with short-term eustatic sea-level changes formed the Sicilian Miocene sapropels, whereas the most recent studies (Nijenhuis et al., 1996; Ten Veen and Postma, 1996) suggest that the Miocene cyclic sediments of Crete are controlled by insolation. At that time, precession must have occurred, as has been proposed for the Plio-Pleistocene, increasing seasonality and fresh water imput during summer insolation maxima in the Northern Hemisphere. Such variations triggered high export production and oxygen consumption rates and concomitant decreases in deep-water formation, allowing the formation of sapropelites.

This paper addresses the formation of late Miocene sapropelites in the Western Mediterranean and their comparison with equivalent sediments in the Central and Eastern Mediterranean. We selected three sapropelites in the 'Los Molinos del Rio de Aguas' section (Sorbas Basin in South Spain) for an integrated geochemical, mineralogical, sedimentological and micropaleontological study.

2. Geological setting

The Sorbas Basin forms part of the Betic Corridor, which was the northern connection between the Atlantic and the paleo-Mediterranean during the late Miocene (Fig. 1A). The basin is narrow and elongated (10×25 km), being surrounded by metamorphic and carbonatic Betic Ranges: the Sierra de los Filabres forms the northern boundary of the Sorbas Basin, whereas the Sierra de Alhamilla and the Sierra de Cabrera constitute the southern margin (Fig. 1B). The sedimentation in the basin started during the middle Miocene and extended into the Quaternary. Völk and Rondeel (1964), Völk (1967), Ott d'Estevou (1980), Riding et al. (1991), Braga and Martín (1992) and Martín and Braga (1994) give detailed descriptions of the general stratigraphic record of this basin. During the pre-evaporitic Messinian, the sedimentation in the central part of the basin was typically hemipelagic, the sediments being a variable mixture of pelagic and detritic imputs reflecting the balance between productivity and dilution.

Homogeneous and laminated marls alternating cyclically with Si-rich deposits characterise the Abad Member. This member was deposited in the centre of the basin between 7.2 and 6 Myr (Sierro et al., 1997, 1999). The cycles overlie Tortonian calcarenites and are capped by the late Messinian evaporites. Coeval with the Abad Member deposition, reef growth progressed along the northern and southern margins. Fourier analyses of the

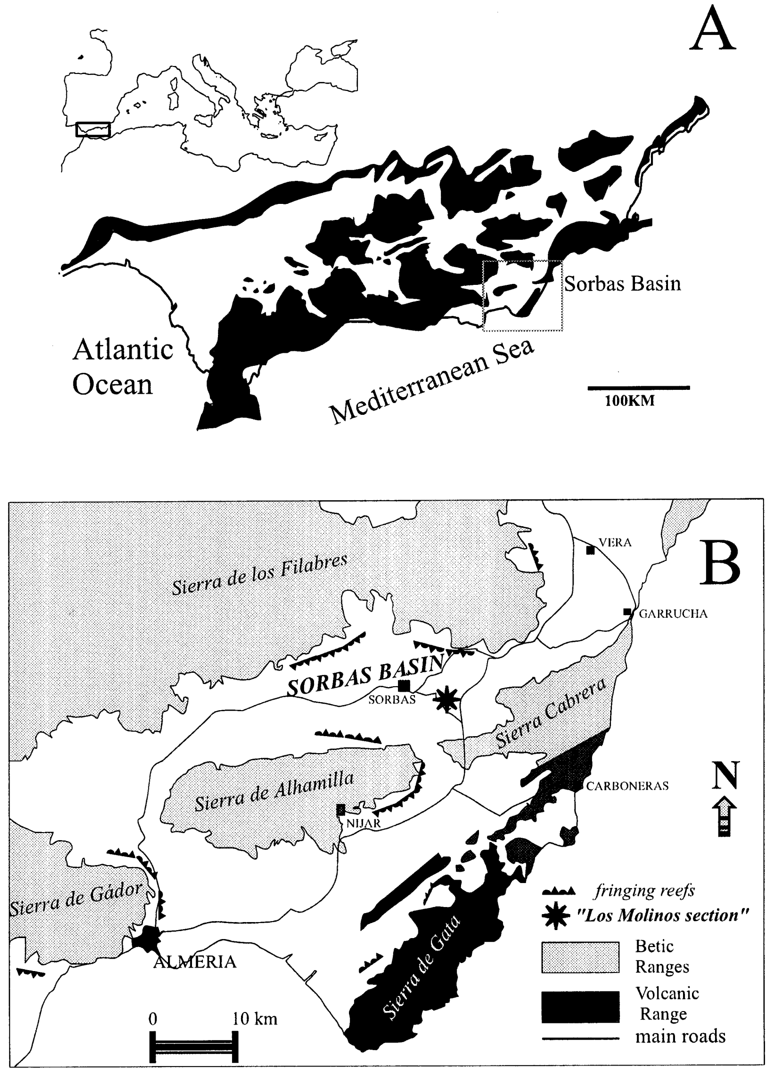

Fig. 1. (A) Messinian paleogeography of the Betic Corridor in south Spain (modified after Serrano, 1975). Black colour indicates emerged domains. (B) Sketch of the present-day geology of the Sorbas Basin showing the small size and elongated geometry. The 'Molinos del Rio de Aguas' section is located in the centre of the basin. The Messinian reef distribution is based on Dabrio et al. (1981).

cycles clearly reveal the precession control over the cyclicity. The Abad Member is divided into Lower and Upper Abad (Sierro et al., 1997). The cycles of the Lower Abad are grey homogeneous marls (2–4 m thick) alternating with CT-opal rich layers (30–50 cm thick). CT-opal is due to the diagenesis of diatomites (Zamarreño et al., 1995). Homogeneous marls display the characteristic polygenetic hemipelagic composition. The Upper Abad cycles are differentiated by (1) a decrease in the cycle thickness, (2) a good preservation of the diatomites, and (3) the existence of brown laminated, organic-rich hemipelagic marls intercalated in the homogeneous marls (Fig. 2). The differences between the lower and Upper Abad are related to the shallowing of the basin during upper Abad times, previously reported by Dronkert (1977) and Troelstra et al. (1980). Dronkert (1977) suggested

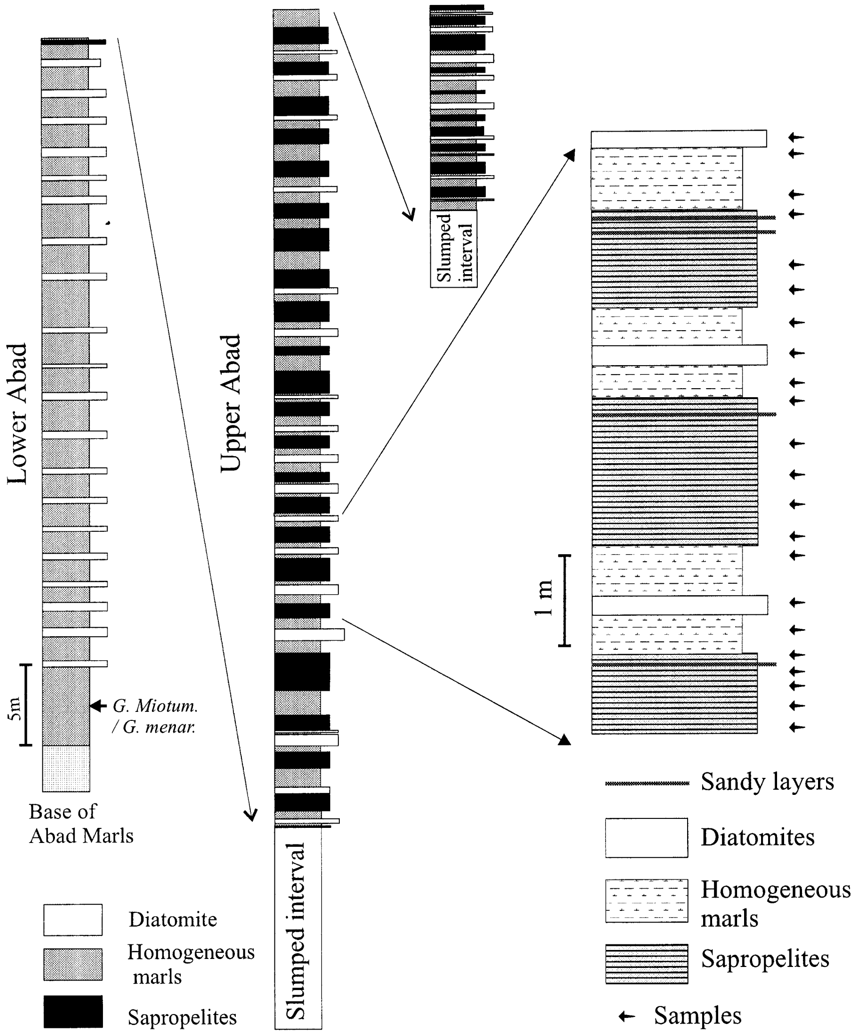

Fig. 2. Composite section of the Abad Member at the 'Los molinos del Rio de Aguas' location. The main differences between Upper and Lower Abad Precessional Cycles are evident. The alternation between sapropelite and homogeneous marls with diatomites characterises the precession cycles of the upper Abad. In the cycles studied, sandy layers occur in the uppermost parts of the sapropelites. The transitions between lithologies are gradual with microbioturbations and diffuse laminations.

a maximal depositional depth of 400 m for the lowermost Abad marls. However, Troelstra and co-workers, from foraminiferal data of the central part of the basin (Los Perales section, which is nearby to Los Molinos del Rio de Aguas section), indicate a decreasing water depth of 200–300 m for the lower Abad to around 150 m in the upper Abad. These authors indicate that the shallowing produced a restriction in the basin favoured by the existence of an eastern sill between the Sorbas basin and the open Mediterranean. Van de Poel (1992) also indicate a similar shallowing and water restriction during the Upper Abad in the Nijar basin, which was still connected to the Sorbas Basin in Abad marls times (Roep et al., 1998). Finally, from abundance, diversity and oxygen isotope of planktic foraminifers, Sierro et al. (1999) suggest normal Mediterranean waters for the lower Abad marls deposition, whereas the strong reduction in diversity and abundance and the heavier oxygen isotopes of *G. bulloides* indicate that the water exchange with the open Mediterranean was strongly reduced during the Upper Abad deposition.

3. Materials and methods

3.1. Sediments

We chose three precession cycles of the Upper Abad Marls, which crop out in the Los Molinos del Rio de Aguas section (Sorbas Basin). A interval of brown laminated marls, rich in organic matter (sapropelites), and grey homogeneous marls with white diatomites form each cycle (Fig. 2). Contacts between the different lithologies are gradual. Diffuse laminations and microbioturbations exist in the transitions between sapropelites and homogeneous marls (Vázquez et al., 1997). Several sandy millimetre-thick layers exist in the upper part of the sapropelites (Fig. 2). The three cycles were deposited between 6613 and 6560 (Krijgsman et al., 1999). The samples of this study were collected every 20–25 cm after removing the weathered surface.

3.2. Mineralogical analyses

Mineralogical analyses were performed with a Siemens D-500 X-ray diffractometer using CuKα radiation. The percentages of quartz, calcite (LMC), magnesian calcite (HMC), and dolomite in bulk powdered samples were obtained following the method of Chung (1974). The Mg/Ca ratios in LMC, HMC and dolomite were calculated by measuring the position of the d_{104} peak compared with a standard (Lumsden, 1979).

The clay mineralogy of the <2 μm size fraction was also obtained by X-ray diffraction. Samples were pretreated with 0.2 N HCl to remove carbonates. After centrifugation and microhomogeneisation, the <2 μm fractions were mounted on glass slides. Each sample was scanned three times: from 2 to 40° 2θ on unexpanded and unheated samples; from 2 to 20° 2θ on glycolated and from 2 to 20° 2θ on samples heated for 2 h at 550°C. Percentage evaluations are based on peak heights, corrected by factors depending on the crystallinity of minerals (Holtzapffel, 1985).

SEM and EDAX observations and analysis on selected samples allowed particles and facies characterisations.

3.3. Grain-size analysis

Particle size distributions were performed on samples after digestion of carbonates and organic matter contents by a Malvern Mastersizer/E laser particle analyser. The mean diameter and the percentages of terrigenous sand, mud, and clay fractions were calculated.

3.4. Mass accumulation rates

Measuring the volume of the samples by a Micromeritics Helium Picnometer (Dadey et al., 1992), we calculated the dry bulk density (DBD). Afterwards, the mass accumulation rates (MAR) were obtained by the formula: $MAR_{COMP.} = LSR \times DBD \times (\%COMP/100)$ g cm^{-2} kyr^{-1}, where $MAR_{COMP.}$ = accumulation rate of each component; LSR = linear sedimentation rate (cm kyr^{-1}) in each cycle; and DBD = dry bulk density in g cm^{-3}.

3.5. Geochemical analyses

Total organic carbon (TOC), total carbon (TCC), and nitrogen contents were determined using a C E Instruments NA 1500 elemental analyser. TCC and TOC were measured by flash combustion, using 'reference lake sediments lksd-2' from The Canadian Certified Reference Materials Project. TOC was determined after attack with 2 M HCl (twofold in excess). The carbonate carbon was calculated from the difference between these two measurements.

3.6. Micropalaeontological analyses

The samples for foraminiferal analyses were washed and sieved through 150 and 62 μm, but only the >150 fraction was counted. The sieved samples were split to obtain a fraction of around

250 specimens, that were then identified and counted.

Slides for the calcareous nannofossil analysis were prepared following the procedure of Flores and Sierro (1997). Around 500 nannoliths from different species and/or morphotypes were counted to estimate their relative abundance.

Samples for siliceous microfossil analyses were subjected to hot treatment with HCl and H_2O_2 and to successive washes in demineralised water. Absolute numbers were determined from slides with randomly distributed microfossils. The recommendations of Schrader and Gersonde (1978) were used to count diatom valves.

4. Results and discussion

4.1. Total organic carbon and nitrogen

One of the essential parameters in sapropel identification is the total organic carbon content (TOC). The mass accumulation rates of the total organic carbon (TOC_{MAR}) in Sorbas sediments are lower than the values obtained in most of the Quaternary sapropels. The TOC_{MAR} maximum of the laminated sediments reaches $0.5\ g\ cm^{-2}\ ka^{-1}$ (Fig. 3). This accumulation rate is equivalent to a TOC of 1.8%, indicating that the Sorbas laminated marls should in fact be regarded as sapropelites. Similar amounts of TOC have been described in the Tortonian laminites from Gavdos, Crete (Nijenhuis et al., 1996), and in the Plio-Pleistocene laminites from South Italy and Sicily (Howell et al., 1988, 1990; Van der Weijden, 1993; Van Os et al., 1994). In contrast to sapropelites, the TOC_{MAR} values of the homogeneous sediments are around $0.15\ g\ cm^{-2}\ ka^{-1}$ ($\sim$0.6%), corresponding to common homogeneous hemipelagic oozes. Generally, the increases in TOC are well marked and located in the centre of the laminated sediments. Although this peak is considerably lower in the upper cycle, there are several differences with the TOC of homogeneous marls.

The carbon/nitrogen (C/N) ratio (Fig. 3) displays a distribution similar to TOC. The increases in the C/N ratio reflect appreciable increases in Terrigenous Organic Matter (TOM) (Müller, 1977). In the Upper Abad cycles, these C/N (TOM) increases coincide with abundant well-preserved land plant remains, suggesting the importance of the vegetal cover on the organic carbon of Sorbas sapropelites. Likewise, Van de Poel (1991) indicates the existence of abundant floating plant material in the upper Abad marls in the nearby Nijar basin. Howell et al. (1988, 1990) in Vrica laminites, and Brosse and Herbing (1990) report similar TOM increases in Plio-Pleistocene sapropels from the ODP leg 107 in the Tyrrhenian. Moreover, in the Eastern Mediterranean sapropels S_7 and S_1, Ten Haven et al. (1987a,b) describe increases in TOM via nutrient-rich river discharges. Furthermore, Shaw and Evans (1984) find increments in total pollen content that are clearly related to river inputs in sapropel S_1.

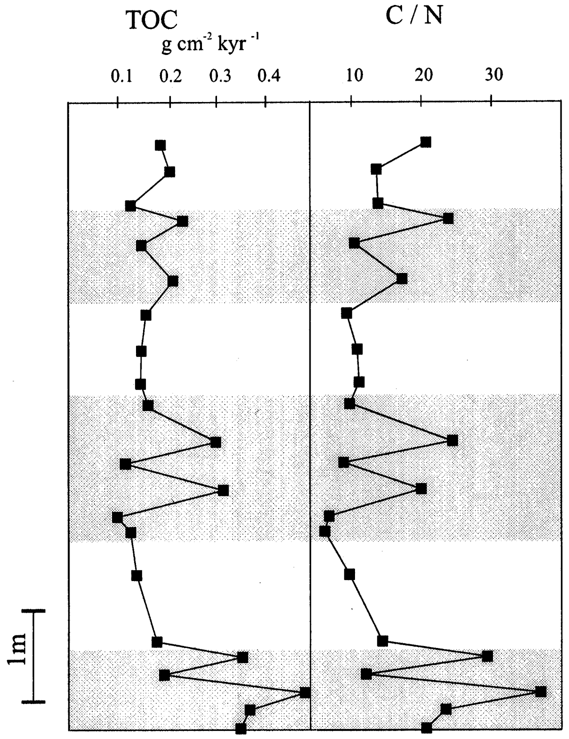

Fig. 3. Mass accumulation rate of total organic carbon (TOC) and ratio between TOC and nitrogen (C/N). The maxima TOCMAR values of Upper Abad cycles correspond to the laminated marls ($0.5\ g\ cm^{-2}\ kyr^{-1}$). These values are lower than 2% and indicate that the laminated marls are sapropelites. C/N peaks correlate with TOC maxima and suggest notable inputs of terrestrial organic matter during sapropel formation.

4.2. Grain-size analysis

It is well known that the amount of terrigenous particles and their grain-size variations provide evidence of detritic transport of particles to the Basin. The sediments of the Abad Member settled in the centre of the basin, and the non-carbonate particles are smaller than 63 μm. Thicker diameters have only been detected in some sandy layers occurring in the upper part of the sapropelites (Fig. 2). These layers suggest sporadic stormy floods, and we did not plot their grain-size values.

The terrigenous grain-size distributions oscillate in the cycles (Fig. 4). There are marked differences between sapropelites and homogeneous marls. In the sapropelites, the amount of terrigenous particles is low, the clay fraction is prevalent, and the mean diameter is smaller. In the homogeneous marls, the MAR of terrigenous is higher, and the coarse fractions prevail. The changes between both facies are gradual.

These data suggest changes from humid to sub-arid climates forced by precession cyclicity. In a more humid climate, at times of sapropelite deposition, the amount of terrigenous particles is low, and the fluvial transport of clays dominates. During the sub-arid periods, sporadic stormy floods and eolian inputs prevail, resulting in an increase in terrigenous abundance and in the mean diameter of terrigenous particles. This sub-arid climate resembles the conditions of the present-day Western Mediterranean in many areas of Spain and North Africa. The maximum terrigenous contents in the upper part of homogeneous intervals may represent maximum aridity, which is in agreement with microfaunal inferences (Sierro et al., 1997).

Despite the scant data on the terrigenous grain-size variations in sapropels, De Visser et al. (1989) and Van Os et al. (1994) report smaller sizes for decalcified particles in the sapropelitic sediments in the Trubi Formation of Sicily. Foucault and Mélières (1995) suggest a decrease in terrigenous minerals in the sapropels of the Narbona Formation of Sicily. These authors suggest a diminution in the wind transport efficiency from the Sahara during sapropel deposition. Finally, Aksu et al. (1995) describe increases in terrestrial particles in the sapropel S_1 from Aegean cores linked to a higher degree of precipitation and run-off from rivers flowing into the Aegean Sea. By contrast, no variations in grain size are found by Feldhausen and Stanley (1980) in Holocene and Upper Pleistocene sapropels from the Hellenic Trench, by Cramp and Collins (1988) in sapropel S_1 of the Sphorades Basin (North Aegean Trough), or by Nijenhuis et al. (1996) in the Miocene laminites from Gavdos (Crete).

The different patterns of terrigenous proxies suggest the control of the paleogeography of each basin over terrigenous accumulation. One possible explanation could be that the wind transport efficiency exerts a strong influence over the terrigenous inputs in the basins that are removed from major rivers and close to the Sahara (i.e. Sicilian basins). By contrast, precipitation controls the detritic input, and winds become important only during the drier periods in marginal basins such as Sorbas and some Aegean basins. Finally, the sediments do not record terrigenous variations at other sites, i.e. some basins of the Hellenic trench.

4.3. Carbonate mineralogy

The carbonate particles are made up of four different minerals: low magnesian calcite (LMC), Ca-rich dolomite, stoichiometric dolomite and high magnesian calcite (HMC). Each carbonate has its own origin, and its incorporations into the sediment have different meanings. Sedimentological and X-ray data, with binocular and SEM observations, indicate that LMC corresponds to the pelagic rain (foraminifera and coccoliths), whereas the HMC particles derive from the carbonates of the continental shelves surrounding the basin (e.g. red algae, molluscs). A similar scenario of carbonate depositions exists today in the nearby Southeastern Balearic Basin (Vázquez et al., 1991; Vázquez and Zamarreño, 1993). The X-ray diffraction also reveals two types of dolomites: (a) Ca-rich dolomite with a composition of $(Mg_{0.44}Ca_{0.56})CO_3$, and (b) Stoichiometric dolomite with a composition of $(Mg_{0.49}Ca_{0.51})CO_3$.

The accumulation rates of the four carbonates change cyclically (Fig. 5). LMC_{MAR} has the highest values, reaching maxima above the transition from

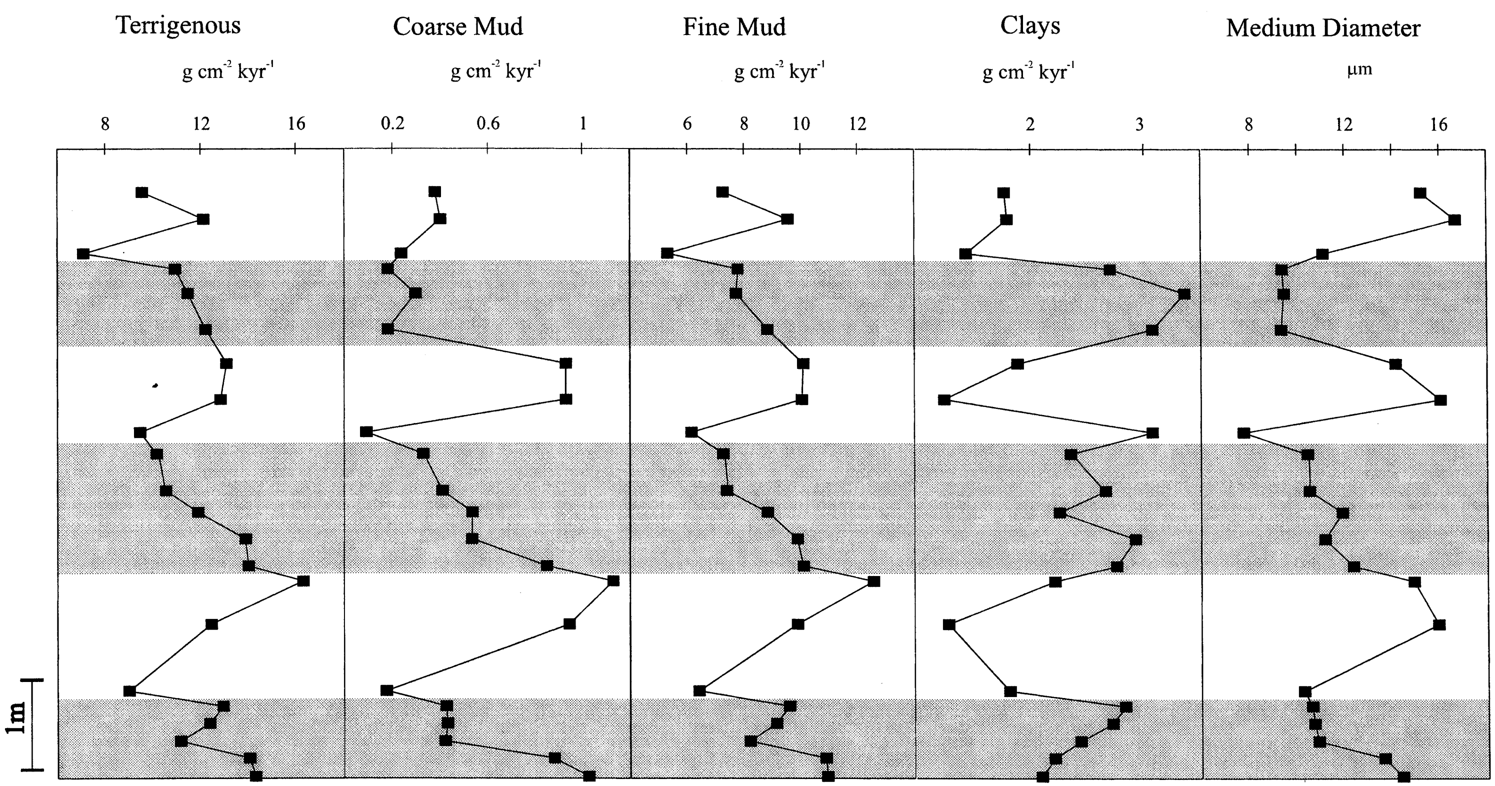

Fig. 4. Textural variation of the non-carbonate particles in the Upper Abad cycles. The oscillations of the mass accumulation rates of terrigenous particles and mean diameter reflect changes in the model of the detritic transport to the Basin. Coarse mud is the fraction between 50 and 20 µm, the fine mud comprises particles from 20 to 4 µm in size, and clay is <4 µm. The shaded area represents sapropelites.

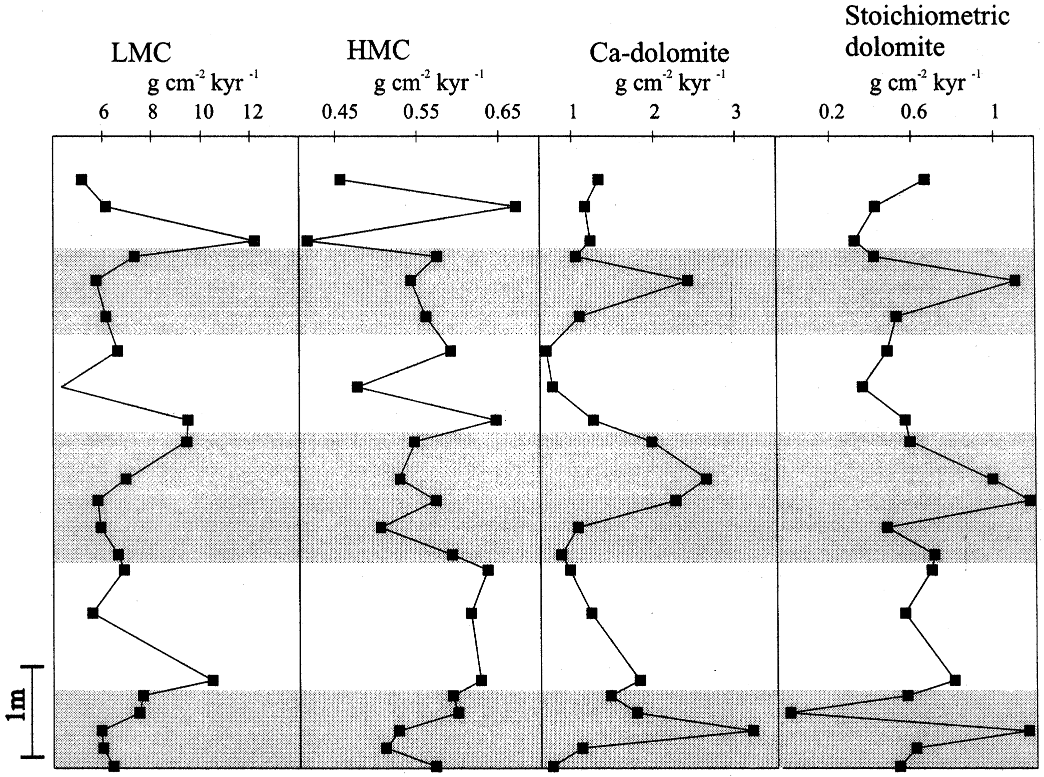

Fig. 5. Mass accumulation rates of low magnesium calcite (LMC), high magnesian calcite (HMC), Ca-rich dolomite and stoichiometric dolomite. See text for explanations. Shaded areas indicate sapropelites.

sapropel to homogeneous sediments. These maxima coincide with the maximum diversity and abundance of foraminifers (Sierro et al., 1997, 1999), together with a great abundance and diversity of coccoliths revealed by SEM observations. Ca-dolomite is the second mineral in abundance. Its MAR is at a maximum in the middle of the sapropelites. In the other samples, the Ca-dolomite$_{MAR}$ is lower. Stoichiometric dolomite$_{MAR}$ values display a similar distribution, but their values are three times lower than those of the Ca-dolomite$_{MAR}$. Finally, the HMC$_{MAR}$ values are higher in the homogeneous interval and decrease in sapropelites. The repetition of the mineral distribution pattern should be noted throughout the cycles, in particular the maximum LMC$_{MAR}$ at the beginning of the homogenous marls and the maxima of the dolomite in the middle of the sapropelites.

Despite the scant data on carbonate minerals in sapropels in other Mediterranean Basins, Sigl et al. (1978) documented the presence of LMC, HMC and dolomite in the Plio-Pleistocene to Miocene sapropels recovered in DSDP sites 374, 375, 376, and 378. In such sapropels, LMC is the most abundant carbonate, its origin being related to coccolith and planktonic foraminifer abundance. The authors linked the HMC to inputs from shelf carbonates, the dolomite being detritic or diagenetic. There are LMC and dolomite in the Pliocene Narbona Formation. LMC is made up of 80–90% of coccoliths and 10–20% of foraminifers, whereas the dolomite is detritic (Hilgen, 1987; Foucault and Mélières, 1995). LMC has a distribution pattern similar to that of cycles in Sorbas, and the maximum values are located above the sapropel–homogeneous transition. These oscillations are attributed to increases in the clay content in sapropels changing the productivity–dilution balance. The dolomite in the Narbona Formation displays a different distribution, being maximal in the homogeneous sediments. There are more data for total carbonate in the sapropelitic sequences of the Mediterranean. Most studies show a decrease in total carbonate in sapropels, which is frequently interpreted as a dilution effect given the increase in terrigenous particles (Nesteroff, 1973; Calvert et al., 1987; Howell et al., 1988, 1990; Emeis et al., 1991; Pruysers et al., 1991; Van der Weijden, 1993; Van Os et al., 1994; Aksu et al., 1995; Nijenhuis et al. 1996).

In order to improve our understanding of the relationships between the dilution and productivity in the Upper Abad cycles of the Sorbas Basin, we plotted the accumulation rates of LMC and terrigenous components (LMC_{MAR}, and $terrigenous_{MAR}$ in Fig. 6). LMC_{MAR} is related to the calcareous planktonic productivity, whereas the MAR of terrigenous is linked to the fluvial and eolic inputs. Fig. 6 shows the high negative covariance between the sets of data values, indicating their inverse relationship. Furthermore, we calculated the accumulation rates of LMC, assuming the terrigenous sedimentation to be constant [$LMC_{(t)}$], and the MAR of terrigenous particles, assuming the sedimentation of LMC components to be constant [$t_{(LMC)}$]. We used the equations of Van Os et al. (1994) and plotted the results in Fig. 6. The $t_{(LMC)}$ is higher than $LMC_{(t)}$, and the maximum values of $LMC_{(t)}$ coincide with the decrease in $t_{(LMC)}$.

Although the productivity/dilution balance undergoes cyclical changes, the variations do not match the laminated/non-laminated facies changes. The LMC_{MAR} is maximal after each new onset of the well-mixed water column, following sapropelite deposition. After the LMC_{MAR} maxima, the diatomite blooms (Fig. 2), and after these

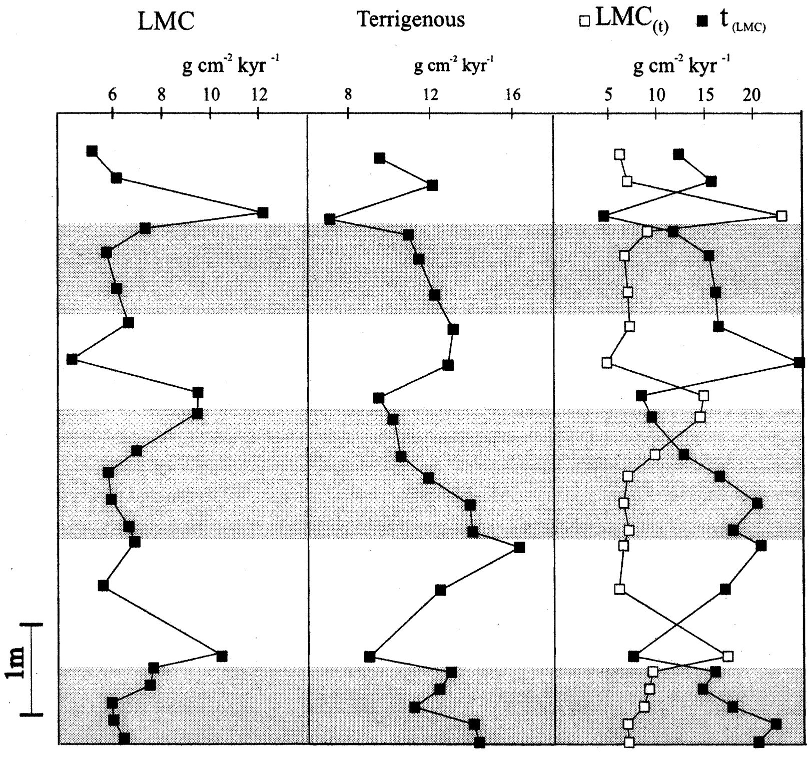

Fig. 6. Mass accumulation rates of low magnesium calcite (LMC) and terrigenous fraction. The third column displays the MAR of LMC assuming the terrigenous input to be constant (LMC(t), open squares) and the MAR of the terrigenous components assuming the flux of LMC to be constant [t(LMC), solid squares]. Calculations after Van Os et al. (1994). See text for explanations.

blooms, the terrigenous fluxes suddenly increase and remain high up to the middle of the sapropelite.

4.4. Clay mineralogy

The Upper Abad cycles have the following clay minerals, in descending order of abundance: smectite, illite, paragonite, paligorskite, kaolinite and clorite (Fig. 7). This association is similar to that of the Messinian sediments of Vera (Galán et al., 1985) and Fortuna (Chamley and Müller, 1991) Basins in South Spain. Azzaro et al. (1988) also describe a similar clay association in the marls of the Tripoli Formation in Sicily. The authors found no indication of secondary alteration and they assumed that the clay mineral oscillations reflect changes in the nature and intensity of the pedogenesis and erosion of the hinterland.

The smectites of the Neogene Betic basins, are Al–Fe beidellites (Galán et al., 1985; Chamley and Müller, 1991). This mineral occurs in regions characterised by short humid seasons and poor drainage, which is typical of subarid climates (Chamley and Robert, 1980; Chamley, 1989). Kaolinite results from chemical weathering (Millot, 1964, 1970) in relatively well-drained soils in warm and humid climates (Chamley, 1989; De Visser and Chamley, 1990). Illite, clorite and paragonite, are classic detritic clays formed by physical weathering of metamorphic and igneous rocks (Millot, 1970; Gaucher, 1981; Chamley et al., 1986). Galán et al. (1985) suggests an anchimetamorphic source for these detrital clays in the Vera basin. Paligorskite is reworked and transported by the Sahara winds from North Africa to diverse areas of the Mediterranean (Chamley and Millot, 1975; Chamley, 1975a,b; Tomadin et al., 1984; Robert et al., 1984; Coudé-Gaussen, 1988).

The mass accumulation rates of the clay minerals (Fig. 7) oscillate throughout the cycles recording changes in the climatic conditions and in the mechanisms of detritic transport. Smectite$_{MAR}$ and paligorskite$_{MAR}$ display irregular trends. The accumulation rates of kaolinite and detritic clays have a marked correlation. These minerals have peaks in the middle of the sapropelites and are less abundant in the rest of the cycle.

To clarify clay oscillations, we plotted the kaolinite/smectite and paligorskite/kaolinite ratios and the MAR of detrital clays (Illite, chlorite and paragonite) in Fig. 8. Given the different formation of smectite and kaolinite, the kaolinite/smectite ratio (k/s) could reflect continental climatic variations. The higher the k/s, the higher the humidity (Chamley, 1989; Chamley and Müller, 1991; Chamley et al., 1993). The maximum k/s values are higher in the lower half of the sapropelite. Upwards, they show a decrease, reaching minimum values in the transition between sapropelites and homogeneous marls. However, the higher the k/s, the higher the detrital clays (Fig. 8). Both patterns outline a scenario with a gradual climatic change from a more humid climate to a subarid weather, which has a short stormy precipitation. In the more humid environment, kaolinite is formed, run-off is increased, and the transport of clays is more abundant. During the subarid weather conditions, precipitation and river discharges are sporadic, the smectite is prevalent and the clays are less abundant.

The paligorskite/kaolinite ratio (p/k) has a strong negative correlation with the k/s ratios and detritic clays (Fig. 8). Maximum values of p/k correspond to the samples with lower k/s and detritic clays. Thus, paligorskite becomes more important when aridity increases, owing to the decrease in the run-off transport of clays and to the increase in the eolian inputs of paligorskite from the Sahara. The winds coming from the Sahara and the 'red rains' in South Spain are associated with strong cyclonic depressions, which are more frequent in drier periods (Querol et al., 1998). Several similar climatic clay distributions exist in the Tripoli Formation (Chamley et al., 1986; Suc et al., 1995) and in the sapropel S_1 from the Aegean Sea (Aksu et al., 1995). By contrast, the clay variations are different in the Pliocene Trubi and Narbona formations. The smectite is higher in sapropelitic levels, whereas kaolinite and detrital clays increase in non-sapropelitic intervals in the Trubi (De Visser et al., 1989) and in the Narbona cycles (Foucault and Mélières, 1995). De Visser et al. (1989) report the contrasting climatic inferences between clay mineralogy and the other proxies in the Pliocene of Catalnissetta basin.

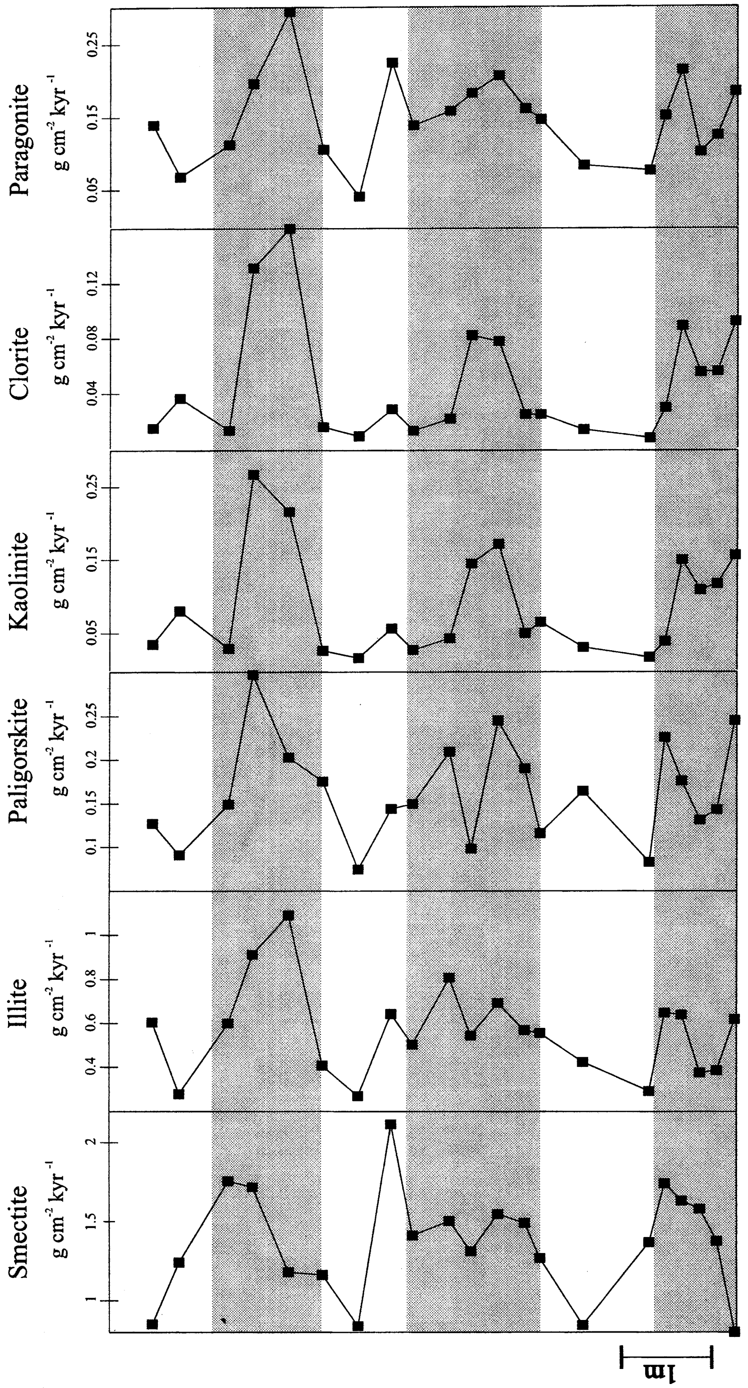

Fig. 7. Mass accumulation rates of the clay minerals identified in the Upper Abad cycles. Shaded areas denote sapropelite. See text for discussion.

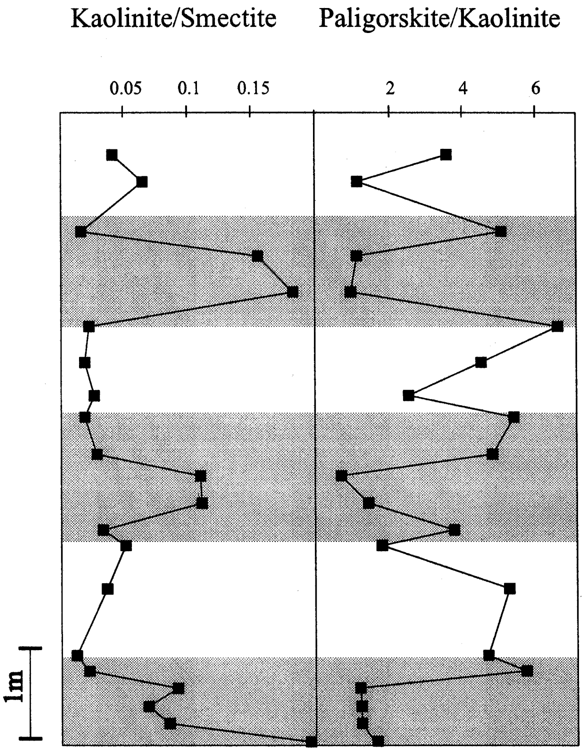

Fig. 8. Clay mineral indicators of paleoclimatic changes. Note the inverse correlation between kaolinite/smectite ratio and the paligorskite/kaolinite ratio. Shaded areas indicate sapropelite.

These authors suggest that kaolinite and paligorskite in these cycles were reworked from older soils and sediments that were wind-supplied from North Africa.

Summarising, clay data from the Miocene Upper Abad support the fact that the climate regulates the type of clay mineral and the mechanisms of transport. Consequently, the gradual change from a more humid climate to a subarid weather in which the precipitation is sporadic exerts a control over the kaolinite/smectite ratio and the eroded clay minerals. The cyclonic winds transport paligorskite from the Sahara to the Sorbas Basin, especially during subarid episodes.

4.5. Microfaunal data

In common with other proxies, foraminifers and coccoliths also show well-marked precession cyclicity and provide valuable insights into the climatic and oceanographic variations (Sierro et al., 1997, 1999). *Globigerinoides obliquus*, *Globigerinoides sacculifer*, *Globigerina apertura* and *Orbulina universa* are prevalent in the sapropelite. These are forms living, or are equivalent to forms currently proliferating, in warm oligotrophic waters (Be, 1977; Luz and Reiss, 1983; Hemleben et al., 1988; Pujol and Vergnaud-Grazzini, 1995). By contrast, these species become rare in the homogeneous sediments where they are replaced by species such as *Globigerina bulloides*, *Turborotalita quinqueloba* and Neogloboquadrinids that are typical of cold and nutrient-rich waters (Be, 1977; Luz and Reiss, 1983; Hemleben et al., 1988; Pujol and Vergnaud-Grazzini, 1995). Thus, the ratio between such assemblages (warm/cold PF index, in Fig. 9) shows a marked cyclicity. This pattern is also displayed by the calcareous nanoplankton assemblages. Peaks of *Coccolithus pelagicus* and other cold-water species such as *Dictyococcites antarcticus* characterise the homogeneous intervals. By contrast, sapropels register considerable increases in Discoasterids especially in the uppermost part. Cachão (1995) observed a good correlation between the abundance of *C. pelagicus* and nutrient-rich waters of the present-day coastal upwelling of Portugal. However, Chepstow-Lusty et al. (1989) and Flores et al. (1995) observed a higher discoaster abundance in Pliocene deep-sea sediments under oligotrophic conditions.

Diatoms, silicoflagelates and sponge spicules are only present in the indurate, opal-rich layers in the middle of the homogeneous intervals, whereas the sapropels and the rest of the homogeneous are devoid of siliceous microfossils. This suggests that silica, one of the main biolimiting nutrients in the sea, was relatively abundant in surface waters favouring the opal preservation during the deposition of the indurate layer. The diatom assemblage is clearly dominated by *Thalassionema nitschioides*, which is a species characteristic of the present-day low-intensity upwellings in the Western Mediterranean (Abrantes, 1988; Bárcena and Abrantes, 1998). The benthic foraminifera (Sierro et al., 1999) are abundant and diverse in the homogeneous marls, being absent in the sapropelites (Fig. 9).

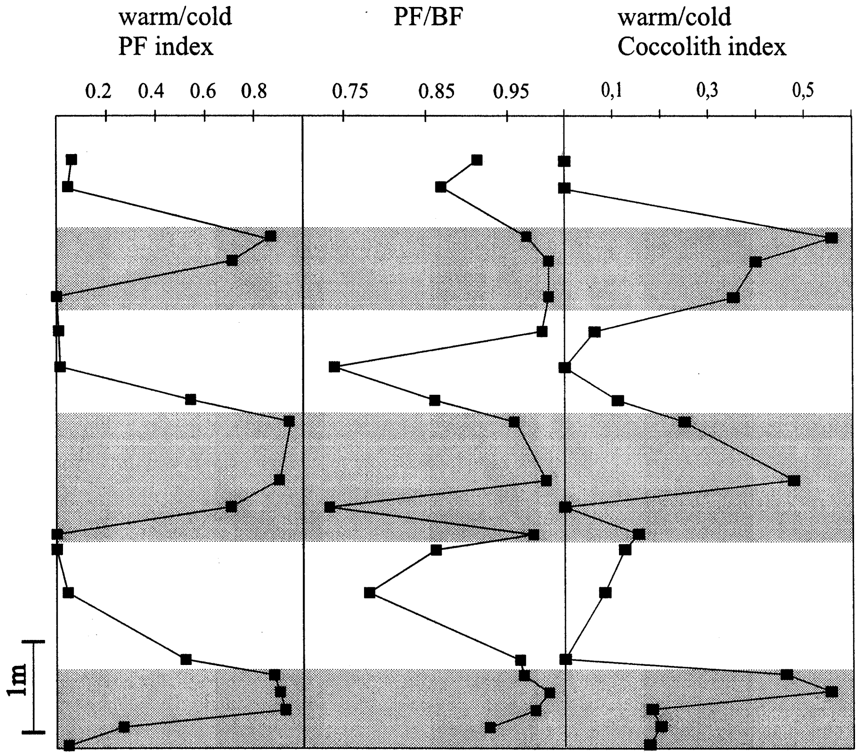

Fig. 9. Microfaunal ratios provide a good definition of the water-column changes. The warm/cold PF and coccolith indexes indicate changes between a warm-temperate oligotrophic scenario and a cold eutrophic regime. PF/BF is the ratio between planktonic and benthic foraminifera and gives indications of the oxygen content in the bottom water and therefore of water-column stratification. Both ratios match the sapropelites (shaded area) and homogeneous marls.

These faunal relationships yield two distinct scenarios in the Upper Abad cycles of the Sorbas Basin: a scenario controlled by a warm/oligotrophic water column corresponding to the sapropelites, and a scenario with cold/eutrophic conditions and homogeneous hemipelagic marl deposition. The warm oligotrophic interval developed in a humid climate with a permanently stratified water column. The surface layer is less saline and nutrient-poor, and the oxygenation of deep, nutrient-rich waters is considerably reduced, allowing sapropelitic laminae deposition. The cold eutrophic scenario is brought about when the climate becomes drier. The stratification of the water column becomes unstable, and the gradient of vertical density is less pronounced. This change favours bottom oxygenation and re-colonisation of the sea floor by benthic species. Concomitantly, the nutrient levels of the photic layer increase, favouring eutrophic fauna and the deposition of homogeneous marls.

5. Conclusions

The Upper Abad Member of the Sorbas Basin (early Messinian, South Spain) consists of precession cycles made up of sapropelites and homogeneous marls. These cycles have highly amplified records of the paleoceanographic and paleoclimatic changes that occurred in the westernmost Mediterranean Sea before the deposition of the evaporites. Our data support the hypothesis of climatic control over the development of the sapropelitic–homogeneous cyclic alternations. The data outline two main climatic scenarios. Each scenario climate regulates the sedimentology and physicochemistry of the water column, resulting in different sediment facies. Thus, at times of precession minima, when the perihelion coincides with the Northern Hemisphere summer, the sapropelites developed in a temperate climate with a long humid period. Rains intensified the river discharges of clays and terrestrial organic matter, and the water column became stratified with a less saline upper layer characterised by oligotrophic fauna. This condition changed to subarid, drier weather with stormy sporadic rains, during which homogeneous sediments settled. Aridity affects the precipitation/evaporation balance, and as soon as the upper layer becomes more saline, the density gradient is reduced, and the water stratification breaks up, favouring the existence of eutrophic fauna in the photic zone. Sea-floor oxygenation encourages benthic organisms at the sea bottom. Storms become more frequent, and sporadic highly energetic river floods generate the sandy layers observed in the cycles.

In conclusion, our data suggest that the climatic changes between these scenarios occur gradually. Despite the sharp transitions between sapropelitic and non-sapropelitic sediments, detailed observations reveal the existence of centimetre-scale intervals between the two main stratigraphic facies. Diffuse laminations and microbioturbations exist in these thin levels. Moreover, the variations in most of the textural, mineralogical, faunal and geochemical paleoceanographic proxies indicate that the changes in climate took place gradually. Furthermore, in the light of our data, it seems that the change from the subarid to the humid climate (transition from homogeneous marls to laminated sapropelites) proceeded at a faster rate or, at least, was recorded more rapidly in the sediments than the change from the humid to the subarid climate.

Acknowledgements

We thank F. Hilgen for the discussion on an earlier draft. Helpful comments of A.R. Fortuin and an anonymous reviewer are also very much appreciated. This study was supported by research grants of the Fundación Ramón Areces and DGCYT Projects PB95-0927-CO2, PB 98-0288-CO2. Special thanks are given to Jose Elvira and Jesus Roncero for technical assistance in sample processing.

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