OCEANOLOGICA ACTA - VOL. 16 - N°2

Mediterranean sapropel formation, dynamic and climatic viewpoints

Mediterranean sapropels
Dynamics
Climate
Productivity

Sapropèles méditerranéens
Dynamique marine
Climat
Production planctonique

Jean-Pierre BETHOUX

Laboratoire de Physique et Chimie Marines, Université Paris VI, URA CNRS, B.P. 8, 06230 Villefranche-sur-Mer, France.

Received 15/12/92, in revised form 19/03/93, accepted 24/03/93.

ABSTRACT

Sapropels, *i. e.* discrete, black and organic-rich layers sedimented in the Eastern basin in anoxic conditions, constitute record, over the past 500,000 years, of eleven particular interactions between marine circulation and biology during glacial as well as temperate climates. Considering climatic and dynamic aspects, we argue that this formation is preconditioned by the stoppage of dense water formation in the Adriatic Sea, following sea-level and/or freshwater input changes. In this case, oxygenation of eastern deep water depends solely on the dense water formation in the South Aegean Sea, a process modified by episodic freshwater inputs. This scheme explains correlations between sapropel events and the strong African monsoons responsible for Nile river flooding over several thousand years (Rossignol-Strick, 1983). At the time of sapropel formation, Mediterranean general dynamics and planktonic productivity were not drastically changed. The deciphering of Mediterranean climatic records needs to take into account northern ice sheet changes (*via* sea-level, temperature and salinity in the adjacent Atlantic), as well as tropical climatic oscillations (*via* increased freshwater inputs and sapropel events).

Oceanologica Acta, 1993. **16**, 2, 127-133.

RÉSUMÉ

La formation des sapropèles en Méditerranée orientale : effets respectifs de la dynamique marine et du climat

Les sapropèles du bassin oriental de la Méditerranée, boues noires riches en carbone organique et sédimentées en conditions anoxiques, constituent onze couches individualisées dans le dépôt des derniers 500000 ans, en période glaciaire comme en climat tempéré. Considérant la dynamique et la climatologie de la Méditerranée, le dépôt de sapropèle est rendu possible par l'arrêt de la formation d'eau dense en mer Adriatique, suivant un changement du niveau de la mer ou des apports d'eau douce. Dans ce cas, l'oxygénation des eaux profondes orientales n'est plus assurée que par la formation d'eau dense dans le sud de la mer Égée, un processus qui peut être altéré par une augmentation des précipitations et des débits des fleuves. Ce schéma peut expliquer les corrélations entre les sapropèles et les moussons africaines responsables de fortes crues du Nil sur quelques milliers d'années (Rossignol-Strick, 1983). Lors des sapropèles, ni la dynamique de la Méditerranée ni sa productivité biologique ne sont fortement changées. L'utilisation des enregistrements climatiques méditerranéens nécessite de prendre en compte les évolutions des couvertures glaciaires de l'hémisphère nord (*via* le niveau de la mer, la température et la salinité dans l'Atlantique) ainsi que les oscillations climatiques tropicales (dont les sapropèles constituent un indice de fortes précipitations).

Oceanologica Acta, 1993. **16**, 2, 127-133.

INTRODUCTION

Sapropels were first discovered in the Eastern Mediterranean in core samples of the 1947-1948 Swedish Deep-Sea Expedition (Kullenberg, 1952). Since the 1960s, many studies have addressed some of the sapropelic horizons as relating to marine dynamics, biology and sedimentation, and to climatic variations. Despite the well defined paleochronology of sapropelic layers in the Ionian and Levantine basins, contradictions have appeared in hypotheses concerning the paleocirculation (more or less stagnant deep water, reversal of currents across the Sicily and Gibraltar straits, freshening of surface water...), the occurrence of enhanced planktonic production and the climatic changes (precipitation, temperature...) at the time of sapropel formation. Up to now, the lack of a hierarchy in the various proposed driving forces for sapropel formation has prevented most of the findings from being used in global change studies.

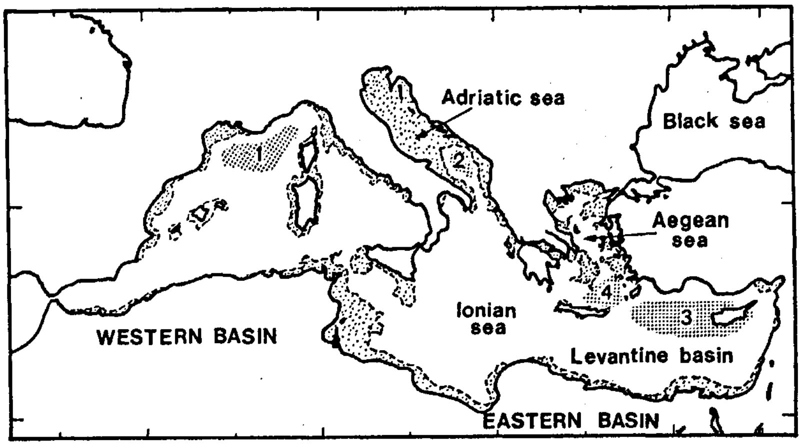
The dynamics of the semi-enclosed Mediterranean Sea is linked to heat, water and salt budgets, *i. e.* to exchanges with the Atlantic Ocean and the atmosphere. Heat advection from the Atlantic corresponds to a gain of 7 W m^{-2} which is added to solar radiation of 195 W m^{-2} absorbed by the sea. A constant mean annual temperature in the Mediterranean is maintained through the transfer to the atmosphere of those heat gains, which occurs by net infrared radiation (69 W m^{-2}) and the loss of sensible (13 W m^{-2}) and latent heat flux (120 W m^{-2}). The latter corresponds to a loss of evaporated water of 1.55 m a^{-1}, which ranges between 1.60 and 1.40 m a^{-1} in the eastern and western basins respectively. The greater evaporation in the eastern basin corresponds to a greater insolation of the most southerly part. These evaporation losses are much higher than the estimated freshwater gain P from precipitation and river runoff of 0.60 m a^{-1} (Bethoux, 1979). The water budget deficits (E-P = 0.95, 1.02 and 0.8 m a^{-1} for the Mediterranean and the eastern and western basins respectively) lead to concentration-basin dynamics, characterized by the formation of dense water and by superposed and reversed flows over the Gibraltar and Sicily sills. Even if the mid-Holocene climate was warmer and more humid, these present data and processes may refer to an almost optimal climatic age. During a glacial maximum, solar radiation was reduced by about 1 % and water temperature by 5°C, while humidity and cloudiness also decreased. As a result, net infrared radiation increased by about 5 % and water loss by evaporation decreased to about 1.25 m a^{-1} for the entire Mediterranean (Bethoux, 1984). On the watershed, palynological studies concur as to the arid climate and vegetation during glacial ages (Cheddadi *et al.*, 1991; Rossignol-Strick and Planchais, 1989). Precipitation and river runoff were certainly lower than the present day freshwater gain. Consequently, in a glacial period as well as in present Mediterranean conditions, evaporation losses linked to heat budget remained much greater than the freshwater gain, and the water budget probably remained more or less constant. According to this hypothesis, the mean freshwater input decreased to about 0.30 m a^{-1} and indeed indicates an arid climate trend, as it corresponds to about half of the present input. During the last glacial age (18 Ka BP), the global ocean salinity increase was about 1 (corresponding to the global disappearance of about 100-120 m of freshwater) and Mediterranean surface salinities calculated from oxygen isotope records were 2.7 and 1.2 higher than present values, in the eastern and western basins respectively (Thunnel and Williams, 1989), which again demonstrates concentration-basin dynamics. Analysis of benthic foraminifers in Atlantic sediment cores effectively indicates a continuous advection of nutrient-depleted Mediterranean outflow during the last 140 Ka BP (Zahn *et al.*, 1987). Consequently, reversal of flows over the Gibraltar sill and the associated hypothesis of nutrient-trap dynamics (*e. g.* Sarmiento *et al.*, 1988) cannot be retained to explain sapropel formation in the Eastern Mediterranean, either during glacial maximum or during a climatic optimum. Furthermore, sapropel formation by exclusively plankton production would require a four- to fivefold increase of the present new production (Bethoux, 1989; Mangini and Schlosser, 1986) *i. e.* an inconceivable change in the nutrient budget of the eastern basin.

SEA-LEVEL AND DEEP-WATER FORMATION CONSTRAINTS

Mediterranean sea-level variations follow changes in the Atlantic linked to global growth or the melting of ice-caps and glaciers. With a lowering of the sea level by 120 m (glacial age), the northern part of the Adriatic Sea and most of the Gulf of Qabes would emerge, as well as some 20 % of the Aegean Sea (Figure). The Black Sea, the Sea of Marmara and the Gulf of Corinth would be transformed into freshwater spillways. The whole eastern basin would be reduced by 3 x 10^{11} m^2 (about 19 %) and the resulting western basin by 0.5 x 10^{11} m^2 (about 6 %). Lowering of the sea level reduces the sea area involved in evaporation processes, extends the watershed and has significant effects on strait dynamics and deep-water formation in the eastern basin.

Strait dynamics

As reported in Ryan (1972), Bradley (1938) and Kullenberg (1952) have considered the possible effect of reduced sea-level on circulation and renewal of the eastern basin. In a concentration basin, heat, water, salt and energy budgets link deep outflow, strait geometry, water deficit and density difference between waters on either side of the sill (*e. g.* Armi and Farmer, 1985; Assaf and Hecht, 1974; Bryden and Stommel, 1984; Kullenberg, 1953; Whitehead *et al.*, 1974). The present Atlantic surface salinity (36.2), strait geometry and water deficit (7.5 x 10^4 m^3 s^{-1}) permit the association of a Mediterranean deep outflow Vs = 1.6 Sv (1 Sv = 10^6 m^3 s^{-1}) with a density difference $d\sigma_t$ = 1.6 kg m^{-3} between Mediterranean outflowing waters and Atlantic subsurface water (Bethoux, 1979). Lowering of the sea level by 120 m reduces the straits section, while Atlantic surface salinity increases by about 1 (Se = 37.2). The glacial water



Figure

The Mediterranean Sea. Broken line gives the glacial coastline when sea level is lowered by 100-120 m (adapted from J. Thiede:, 1978). Stippled surfaces concern dense water formation areas: 1) in the Gulf of Lions and Ligurian Sea; 2) in the Adriatic Sea; 3) in the Levantine basin; 4) in the South Aegean Sea or Cretan Sea.

La Mer Méditerranée. Le tireté représente la ligne de côte en période glaciaire pour un niveau de la mer abaissé de 100 à 120 m (d'après Thiede, 1978). Les surfaces en pointillés correspondent aux zones de formation d'eaux denses : 1) dans le golfe du Lion et en Mer Ligure ; 2) dans l'Adriatique ; 3) dans le bassin levantin ; 4) dans le sud de la Mer Égée ou Mer de Crête.

deficit, 6.4 x 10^{-2} Sv, permits calculation of a density difference $d\sigma_t$ reaching 2.2 kg m^{-3} and a Mediterranean deep outflow reduced to 1 Sv (Bethoux, 1984).

At the strait of Sicily, the eastern deep outflow follows two distinct channels, with sill depths of 430 and 350 m respectively, neither width nor depth being entirely occupied (the maximum depth of 300 m observed for the outflow seems to be governed by the Gibraltar sill depth). The mean outflow through the two channels amounted to 1.2 Sv as calculated from water deficit of the inner basin and from $d\sigma_t$ data (about 0.2 kg m^{-3}; Bethoux, 1979). Flux distribution between the Gibraltar and Sicily straits appears to be proportional to the water deficit of the inner basin, while the respective $d\sigma_t$ are quite different and partly linked to water deficit differences between adjacent basins (about 1 m a^{-1} between the Mediterranean and the Atlantic, and only 0.22 m a^{-1} between the eastern and western basins). At the strait of Sicily, sea-level change does not constitute a strong constraint for the outflow section, as the outflow does not occupy the whole sill section, and direct calculation of flux change induced by sea-level variation is not possible. In a first approach, it may be emphasized that outflow distribution between the Gibraltar and Sicily straits remains proportional to the water budgets of the inner basins. These amount to 6.4 x 10^{-2} Sv for the whole Mediterranean and to 4.4 x 10^{-2} Sv for the eastern basin, Vs through the Sicily strait being estimated at 0.7 Sv. If the outflow thickness Z is considered to be constrained by the change in water depth at Gibraltar, it varies from 220 to about 150 m, when the surface water inflow is situated in the 0-50 m layer (instead of the present 0-80 m). Vs outflow in the main channel, about 2/3 of the total Vs, may be linked to $d\sigma_t$ of about 0.26 kg m^{-3} (the present value being 0.23 kg m^{-3}).

Deep water rising to sill depth is linked to the outflow current (Bryden and Stommel, 1982; Stommel *et al.*, 1973) which may be written as being proportional to $(d\sigma_t\, Z)^{0.5}$, with Z representing the thickness of the outflow (Bethoux, 1984). Gibraltar cross-section and $d\sigma_t$ give a mean outflow current of 2 m s^{-1} in present as well as glacial times, and this current maintains the outflow of deepest western waters. At the Sicily strait, during the glacial age, the resulting outflow current is comparable to the present one and prevents stagnation of the deep eastern waters. With a high freshwater deficit, it appears that lowering of the sea level and reduced strait flows from present to glacial states (and conversely), do not constitute strong constraints in explaining sapropel formation, which mainly occurred during periods of rather high sea-level.

Dense-water formation

In the Mediterranean, four areas of dense water formation are known (Figure): the northwestern basin (Medoc Group, 1970), the Adriatic Sea, the Levantine basin and the Aegean Sea (Zore-Armanda, 1974; Miller, 1974). The eastern deep water originates in the mixing of Adriatic with Levantine dense waters and some dense water from the Cretan Sea, the southern part of the Aegean Sea. When flowing out over the Sicily sill, it constitutes the Levantine intermediate water of the western basin, which, after mixing with surface water, creates the deep western water and feeds the deep Mediterranean outflow across the Gibraltar straits. Upstream from the Sicily strait, the mean characteristics of eastern outflow are: T = 14.1°C, S = 38.8. According to a nomogram of T, S properties (El-Gindy and El-Din, 1986), this outflow corresponds to a mixing of 60 % Adriatic water, 30 % Levantine water and 10 % Cretan water. The characteristics of Adriatic dense water, the present main source of eastern deep water, are preconditioned in the shallow northern part of the sea, where the continental climate (cold Bora wind) cools the surface layer and somewhat increases the salinity previously lowered by freshwater inputs. The second main component of the deep eastern water is the Levantine intermediate water, created in the open sea by autumn and winter cooling of the highly saline surface layer, which sinks slightly to become an intermediate layer and moves westward towards the Otranto Strait and the Ionian Sea (Zore-Armanda, 1974). Its initial salinity is the highest of the eastern basin and results from the evaporation process in the surface layer distant from coastal freshwater effects and isolated from vertical mixing by the density gradient of the seasonal thermocline. The third (and minor) component of eastern deep water is produced by dense-water formation occurring in the South Aegean Sea and joining the Levantine basin through the Kasos and Karpathos straits (Miller, 1974).

The open-sea formation of Levantine dense waters is scarcely affected by climatic changes, while in the Adriatic Sea, dense-water formation and deep-layer oxygenation depend on climatic driving forces such as continental pressure gradient, freshwater budget or sea-level changes. Today, the water balance of the Adriatic as a whole is negative (rivers + precipitation - evaporation = - 0.13 m a^{-1}), but slightly positive in the shallow northern part (Bethoux, 1980). A 11 % increase in freshwater input (wet trend) or a lowering of the sea level by 20 m and a consequent reduction of the sea area exposed to evaporation processes, would lead to a balanced water budget of the whole Adriatic Sea and stop the salinity effect on dense-water formation. A 100 m lowering would dry up more than half of the sea and the resulting water budget would become positive (1.1 m a^{-1}). With freshwater inputs being halved and evaporation reduced by 20%, which corresponds to a glacial age hypothesis, the resulting budget would be positive and equal to 0.13 m a^{-1}. These two factors, sea-level change and/or wet trend, may arrest dense-water formation in the Adriatic Sea. The Levantine water increasingly constitutes the main component of eastern deep outflow and oxygenation of the deep eastern layer is realized by Cretan dense water, as hypothesized from oxygen isotopic records of planktonic foraminifera (Fontugne *et al.*, 1989).

With the hypothesis of Cretan deep water accounting for some 10 % of deep eastern water and therefore also of the outflow through the Sicily strait, its present formation amounts to 0.12 Sv. When deep-water formation is inhibited in the Adriatic Sea, Cretan dense water would progressively fill the eastern basin up to about 300 m depth and its oxygen content (about 235 x 10^{-6} mol kg^{-1} when it is formed) would be used for the remineralization of organic matter settling from the surface layer. With a Redfield molar ratio C/O_2 = 105/135 (Redfield *et al.*, 1963), this yearly input of oxygen would allow remineralization of 7 x 10^{11} mol C in the 300 m-bottom layer. Carbon flux reaching 300 m depth corresponds to 14 % of the surface primary production of carbon (Suess, 1980), *i. e.* about 42 % of the new production, assuming a ratio of 3:1 between primary and new production (Dugdale and Wilkerson, 1988). Consequently, the surface new production may amount to 17 x 10^{11} mol C a^{-1} or 12 g C m^{-2} a^{-1}, a value previously estimated from nutrient and oxygen budgets (Bethoux, 1989). In a glacial trend, the formation of Cretan dense water may be reduced as well as the flows through the straits, while its oxygen content will increase with decreasing water temperature. Despite the uncertainties concerning not only these respective changes but also the rate of vertical carbon flux and new production versus primary production ratio, it appears that Cretan dense-water formation alone can barely ensure oxygenation of most of the deep eastern water and prevent sapropel formation. This result agrees with most of the Mediterranean sedimentary history. If dense-water formation were stopped, oxygen content (about 190 x 10^{-6} mol kg^{-1} in deep water) and new production would not permit, under present conditions, sapropel formation deposit for about the next 500 years. In the case of reduced dense-water formation and oxygenation barely ensured at a concentration as low as 60 x 10^{-6} mol kg^{-1}, a criterion proposed by Rosenberg (1980) and marking a change in the benthic community, sapropel deposit may occur in about 150 years. Consequently, sapropel formation may accompany quasi-immediately the surface event: the arrival of freshwater stopping dense water formation.

FRESHWATER INPUTS AND NILE FLOODING

There are good time-correlations between sapropel occurrences and African summer monsoons (Rossignol-Strick, 1983; 1985), the obvious link being the Nile river flow, which at these times can increase about twofold its normal freshwater flow into the sea (Rossignol-Strick *et al.*, 1982). Present Nile river flow ranges between 4 and 8 x 10^{10} m^3 a^{-1} (Wadie, 1984) and would constitute, without the High Dam, a 2.4-5 cm a^{-1} freshwater layer over the eastern basin. Consequently, Nile flooding cannot transform the eastern concentration-basin (water deficit of about 1 m a^{-1}) into a dilution one and its nutrient load cannot drastically alter the new production of the basin. Cascadings of brackish water enriched with nutrient from the Black Sea (when isolated from Mediterranean inflow) would not have any greater effect over a thousand years.

The present inflow through the Sicily strait (equal to deep outflow + water deficit) corresponds to the yearly input of a 24 m thick layer over the eastern basin. The Nile river in spate (or an equal freshwater input) would result in an additional salinity decrease of that surface layer dS = S dh/h = 0.08 and in a density decrease of 0.06 kg m^{-3}. After some decades, density difference between eastern and western intermediate waters is reduced from 0.2 to about 0.14 kg m^{-3}. In a glacial trend, eastern inflow would be reduced and corresponds to a yearly input of about 16 m, and an equal additional input of freshwater would result in a salinity decrease of 0.13 and in a density decrease of 0.10 kg m^{-3}. In this case, the density difference between eastern and western intermediate waters would be just about halved, outflow current would be reduced and pumping of deep water slowed down. This flood effect (from the Nile river, the Black Sea or the Po river) cannot explain a complete oxygen starvation in deep water.

Coastal freshwater inputs amplify the general cyclonic circulation around the eastern basin and may concentrate freshwater in particular areas. According to the Mediterranean Sea atlas (Miller *et al.*, 1970), the Aegean water influenced by the Black Sea outflow follows the west Aegean coast, joins the Ionian cyclonic circulation through the Antihytera and Kythera straits and influences South Adriatic surface salinity. Freshwater from the Nile river flow follows the easternmost coast of the Levantine basin, joins the freshwater inputs from Lycian and Cilician Taurus mountains on the south-Turkish coast and influences surface salinity in the South Aegean Sea, *i. e.* in the Cretan dense-water formation area. Furthermore, during the summer monsoon in Africa, the rainy season lasts longer during wet years (Rossignol-Strick, 1983) and consequently the normal August-September flood from the Nile river may be extended by one or two months. Taking into account the transit time (about one to two months), freshwater from the Nile flood may directely influence the South Aegean Sea in November and December, at the time of preconditioning and formation of Cretan dense water. During Nile floods, the newly formed Cretan dense water may be less dense and spread out to lower depths, perhaps between 300 and 800 m, or to the Levantine intermediate layer depth (100-300 m), thereby isolating the deeper layer from oxygen input and allowing sapropel deposition from 1 000 m downwards.

Moreover, since the early 20th century, precipitations at Monaco (CSM, 1988) and over the northwestern basin (from analysis of the French meteorological data) have been decreasing. Similarly, the precipitations have also decreased over North Africa as well as between the 5°-35°N latitudes (Bradley *et al.*, 1987). As a conclusion, Mediterranean precipitations were uncoupled from the increasing trend over Europe and followed the decreasing trend of tropical precipitations. Thus, it may be assumed that during sapropel events, the African monsoon might have increased Mediterranean precipitations, perturbed the Adriatic freshwater budget and strengthened the Nile flood effect on Mediterranean hydrology.

FORMATION OF S_1 SAPROPEL, AN EXAMPLE

The formation of the more recent and best known S_1 sapropel may be re-examined according to our hypothesis. The last glacial maximum, occurring around 18-20 Ka BP, was followed by a slow deglaciation in an arid climate. In the Mediterranean area, the humid trend began at 12 Ka BP, reaching a maximum (summer precipitations) at 8 Ka and has been decreasing until now (Cheddadi *et al.*, 1991). S1 sapropel occured at 9 Ka BP (Fontugne *et al.*, 1993) simultaneously over the whole Eastern basin, when the rising sea level was 40 m below the present one (Fairbanks, 1989). This sapropel formation cannot be directly associated with freshwater discharge from the Black Sea, the maximum runoff of which occurred at about 13 Ka (Fontugne *et al.*, 1989,1993; Denton, 1981). But it coincided with a strong African monsoon (Rossignol-Strick, 1983) with effects on Nile discharge and probably on the entire Mediterranean precipitation and river runoff. These data are consistent with sea-level and water-budget driving forces. During deglaciation and the period of increasing humidity, low sea level, Po river and Black Sea runoffs prevented dense water formation in the Adriatic Sea. Oxygenation of deep water was ensured by the Cretan dense water until increasing river runoff from the Turkish coast and from the Nile river (effect of the strong African monsoon) prevented its formation. At around 7 Ka BP, probable decreasing monsoon intensity allowed reoxygenation of deep water only by Cretan dense water. The progressive oxygenation of Adriatic sediments, which ended only at about 2 Ka BP (Fontugne *et al.*, 1989), evidences the late revival of dense water formation in the Adriatic Sea, when the sea level was situated at about 5 m below the present one (Fairbanks, 1989), with freshwater inputs roughly greater than the present ones. Correlation between the monsoon index and each sapropel occurrence (Rossignol-Strick, 1983) shows that this scenario may apply to each of the eastern Mediterranean sapropels, which mainly occurred during interglacial but also twice during glacial periods (S6, S8). The present strong formation of dense water in the Adriatic Sea probably constitutes an anomaly in the Mediterranean Quaternary history, as does the present high sea level.

SAPROPEL FORMATION OR THE INEFFICIENCY OF OXYGEN VERTICAL TRANSPORT BY EDDY DIFFUSION

When dense water formation is arrested, the vertical oxygen transport depends on the eddy diffusion and may be written as:

$dO_2/dt = Kz\ dO_2/dz$

The first term represents the oxygen flux through a horizontal surface (mol O_2 m^{-2} s^{-1}) and is equal to the product of the eddy diffusion coefficient Kz (m^2 s^{-1}) and the oxygen concentration gradient (mol O_2 m^{-3} m^{-1}). In stratified conditions, $Kz = 0.25\ \varepsilon\ N^{-2}$, where N is the Brunt-Vaïsälä frequency (s^{-1}) and ε the rate of dissipation of turbulent kinetic energy (m^2 s^{-3}). In the eastern basin, the density difference between 300 and 1000 m (or 2000 m) is about 0.2 kg m^{-3} and the N^2 is equal to 2.7×10^{-6}. The corresponding ε value ranges between 0.5 and 1×10^{-9} (*e. g.* McDougall, 1988; Oakey,1988) and the resulting Kz amounts to 5-9 x 10^{-5}. When oxygen concentration decreases from about 190 x 10^{-6} mol kg^{-1} at 300 m to 0 at 1000 m (anoxic condition), the oxygen concentration gradient reaches a maximum value of 2.7×10^{-4} mol m^{-3} m^{-1}. Oxygen vertical transport by eddy diffusion is calculated to be between 1.4 and 2.4 x 10^{-8} mol m^{-2} s^{-1}. As previously shown, below 300 m depth the oxygen requirement for mineralization of organic matter originating from a new production of 12 g C m^{-2} a^{-1} amounts to 0.5 mol O_2 m^{-2} a^{-1} or 1.6×10^{-8} mol m^{-2} s^{-1}. Consequently, vertical eddy diffusion appears to be roughly sufficient to prevent anoxia with the present-day vertical density gradient.

If dense water formation is stopped by a change in the water budget through the sea surface, the densities of surface and intermediate layers are lowered and the initial density gradient between the new intermediate layer and the old deep layer is increased. For a 10 % decrease in evaporation (- 0.16 m a^{-1}) and a 10 % increase in precipitation (+ 0.06 m a^{-1}), the salinity of the surface layer (about 20 m thick) would decrease by 0.35, and the resulting intermediate layer density would decrease by 0.27 kg m^{-3}. The density gradient between 300 and 1000 m would increase from 0.2 to 0.47 kg m^{-3}. The resulting oxygen vertical transport through this density gradient would be 0.6-1 x 10^{-8} mol m^{-2} s^{-1}, which is lower than the required oxygen transport for remineralization of organic matter (1.6×10^{-8} mol m^{-2} s^{-1}). Consequently, despite a rather low density gradient, the vertical transport of oxygen by eddy diffusion cannot serve as a substitute for deep-water formation in the oxygenation of the deep layer. Nevertheless, it may explain the increasing trend with depth of organic carbon content in S1 sapropels, from about 1.5 % in core samples taken at 1000 m to 3 % in core samples taken at 3000 m depth (Murat *et al.*, 1990).

CONCLUSION

The present formation and renewal of dense eastern water is mainly preconditioned in the Adriatic Sea, where processes are strongly constrained by the European climate (Zore-Armanda, 1974), by the local water budget (in relation with tropical precipitations) and by the sea level (northern ice sheet changes). A long response time of the deep eastern basin when deprived from Adriatic dense water and a rather lengthy duration of African summer monsoon events (Rossignol-Strick, 1983), influencing the formation of Cretan dense water, may explain sapropel formationover several thousand years. In this case, Mediterranean dynamics (*e. g.* flows through the straits and negative freshwater budget) or planktonic productivity are not drastically changed.

Acknowledgements

We thank M.R. Fontugne, M. Rossignol-Strick and W.B.F. Ryan for helpful discussions or comments. This paper is a contribution to the MAST 2, MTP Geodyme sub-project.

REFERENCES

Armi L. and D.Farmer (1985). The internal hydraulics of the Strait of Gibraltar and associated sills and narrows. *Oceanologica Acta*, **8**, 1, 37-46.

Assaf G. and A. Hecht (1974). Sea straits: a dynamical model. *Deep-Sea Res.*, **21**, 947-958.

Bethoux J.-P. (1979). Budgets of the Mediterranean Sea. Their dependence on the local climate and on the characteristics of the Atlantic waters. *Oceanologica Acta*, **2**, 2, 157-163.

Bethoux J.-P. (1980). Mean water fluxes across sections in the Mediterranean Sea, evaluated on the basis of water and salt budgets and of observed salinities. *Oceanologica Acta*, **3**, 1, 79-88.

Bethoux J.-P. (1984). Paléo-hydrologie de la Méditerranée au cours des derniers 20000 ans. *Oceanologica Acta*, **7**, 1, 43-48.

Bethoux J.-P. (1989). Oxygen consumption, new production, vertical advection and environmental evolution in the Mediterranean Sea. *Deep-Sea Res.*, **36**, 769-781.

Bradley R.S., H.F. Diaz, J.K .Eischeid, P.D. Jones, P.M. Kelly and C.M. Goodess (1987). Precipitations fluctuations over northern hemisphere land areas since the mid-19th century. *Science*, **237**, 171-175.

Bradley W.H. (1938). Mediterranean sediments and Pleistocene sea levels. *Science*, **88**, 376-379.

Bryden H.L. and H.M. Stommel (1982). Origin of the Mediterranean outflow. *J. mar. Res.*, **40**, 55-71.

Bryden H.L. and H.M. Stommel (1984). Limiting processes that determine basic features of the circulation in the Mediterranean Sea. *Oceanologica Acta*, **7**, 3, 289-296.

Cheddadi R., M. Rossignol-Strick and M. Fontugne (1991). Eastern Mediterranean palaeoclimate from 26 to 5 Ka B.P. documented by pollen and isotopic analysis of a core in the anoxic Bannock basin. *Mar. Geol.*, **100**, 53-66.

CSM (1988). Un siècle d'observations météorologiques à Monaco. Centre Scientifique de Monaco, vol. 3, 79.

Denton G.H. (1981). *The Last Great Ice Sheets*, G.H. Denton and T.J. Hugues, editors. John Wiley and Sons, 484.

Dugdale R.C. and F.P. Wilkerson (1988). Nutrient sources and primary production in the Eastern Mediterranean. in: *Océanographie pélagique méditerranéenne, H.J. Minas and P. Nival, editors, Oceanologica Acta*, **vol. sp. No. 9**, 179-184.

El-Gindy A.H. and S.H. El-Din (1986). Water masses and circulation in the deep layer of the Eastern Mediterranean. *Oceanologica Acta*, **9**, 3, 239-248.

Fairbanks R.G. (1989). A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature*, **342**, 637-642.

Fontugne M.-R., M. Paterne, S.E. Calvert, A. Murat , F. Guichard and M. Arnold (1989). Adriatic deep water formation during the holocene: implication for the reoxygenation of the deep eastern Mediterranean Sea. *Paleoceanography*, **4**, 199-206.

Fontugne M., M. Arnold , L. Labeyrie, M. Paterne, S.E. Calvert and J.-C. Duplessy (1993). Paleoenvironment, sapropel chronology and Nile river discharge during the last 20,000 years as indicated by deep sea sediment records in the eastern Mediterranean. in: *Late Quaternary chronology and paleoclimate of the Eastern Mediterranean*, O. Bar-Yosef and R. Kra, editors (in press).

Kullenberg B. (1952). On the salinity of the water contained in marine sediments. Goteborgs Kungl. Vetenskaps. Vitt-Samhal. Handlingar, 6, 3-37.

Kullenberg B. (1953). Les échanges d'eau à travers le détroit de Gibraltar. Bull. Inf. Commn Centr. Océanogr. Ét. Côtes, 298-302.

McDougall T.J. (1988). Some implications of ocean mixing for ocean modelling. in: *Small-scale turbulence and mixing in the ocean*, J.C.J. Nihoul and B.M. Jamart, editors. Elsevier Oceanography Series, 46, 21-35.

Mangini A. and P. Schlosser (1986). The formation of eastern Mediterranean sapropels. *Mar. Geol.*, **72**, 115-124.

Medoc Group (1970). Observation of formation of deep water in the Mediterranean Sea, 1969. *Nature*, **227**, 1037-1040.

Miller A.R. (1974). Deep convection in the Aegean Sea. *Colloque international CNRS*, **215**, 155-163.

Miller A.R., P. Tchernia, H. Charnock and D.A. McGill (1970). Mediterranean Sea Atlas of temperature, salinity, oxygen profiles and data from cruises of R.V. *Atlantis* and R.V. *Chain* with distribution of nutrient chemical properties. *WHOI Atlas Series*, **3**, 190 pp.

Murat A., H. Got, G. Cauwet and R. Buscail (1990). Facteurs de la variabilité du taux de carbone organique du sapropèle holocène de la Méditerranée orientale. *Rapp. P.-v. Réun. Commn int. Explor. scient. Mer médit.*, **32**, 45.

Oakey N.S. (1988). Estimates of mixing inferred from temperature and velocity microstructure. in: *Small-scale turbulence and mixing in the ocean*, J.C.J. Nihoul and B.M. Jamart, editors. Elsevier Oceanography Series, 46, 239-247.

Redfield A.C., B.H. Ketchum and F.A. Richards (1963). The influence of organisms on the composition of sea water. in: *The Sea. 2*, M.N. Hill, editor. Interscience, New York, 26-77.

Rosenberg R. (1980). Effect of oxygen deficiency on benthic macrofauna in fjords. in: *Fjord oceanography*, H.J. Freeland, D.M Farmer. and C.D. Levings, editors. 499-514.

Rossignol-Strick M. (1983). African monsoons, an immediate climate response to orbital insolation. *Nature*, *304*, 46-49.

Rossignol-Strick M. (1985). Mediterranean Quaternary sapropels, an immediate response of the African monsoon to variation of insolation. *Palaeogeogr.Palaeoclimatol. Palaeoecol*, **49**, 237-263.

Rossignol-Strick M. and N. Planchais (1989). Climate patterns revealed by pollen and isotope records of a Tyrrhenian sea core. *Nature*, **342**, 413-416.

Rossignol-Strick M., W. Nesteroff, P. Olive and C. Vergnaud-Grazzini (1982). After the deluge: Mediterranean stagnation and sapropel formation. *Nature*, **295**, 105-110.

Ryan W.B.F. (1972) Stratigraphy of the late Quaternary sediments in the eastern Mediterranean, in: *The Mediterranean Sea*, D.J. Stanley, editor. Dowden, Hutchinson and Ross, Stroudsberg, Pa., 149-169.

Sarmiento J.L., T. Herbert and J.R. Toggweiler (1988). Mediterranean nutrient balance and episodes of anoxia. *Global biogeochem. Cycles*, **2**, 427-444 .

Stommel H., H. Bryden and P. Mangelsdorf (1973). Does the Mediterranean outflow come from great depth? *Pageophysics*, **105**, 879-889.

Suess E. (1980). Particulate organic carbon flux in the oceans-surface productivity and oxygen utilization. *Nature*, **288**, 260-263.

Thiede J. (1978). A glacial Mediterranean. *Nature*, **276**, 680-683.

Thunell R.C and D.F. Williams (1989). Glacial-Holocene salinity changes in the Mediterranean Sea: hydrographic and depositional effects. *Nature*, **338**, 493-496.

Wadie W.F. (1984). The effect of regulation of the Nile River discharge on the oceanographic conditions and productivity of the southeastern part of the Mediterranean Sea. *Acta adriat*, **25**, 29-43.

Whitehead J.A., A. Leetmaa and R.A. Knox (1974). Rotating hydraulics of strait and sill flows. *Geophys. Fluid Dynam.*, **6**, 101-125.

Zahn R., M. Sarnthein and H. Erlenkeuser (1987). Benthic isotope evidence for changes of the Mediterranean outflow during the late quaternary. *Paleoceanography*, **2**, 543-559.

Zore-Armanda M. (1974). Formation of Eastern Mediterranean deep water in the Adriatic. *Colloque international CNRS*, **215**, 127-133.