Review and new aspects concerning the formation of eastern Mediterranean sapropels

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Abstract

In this paper, it is proposed that the formation of eastern Mediterranean sapropels occurred in an anti-estuarine type of circulation, which was to some degree weakened relative to the present in response to reduction of the eastern Mediterranean excess of evaporation over freshwater input. This reduction of excess evaporation would have been imposed by intensifications of (1) the Indian Ocean summer (SW) monsoon, influencing the eastern Mediterranean via increased Nile discharge, and (2) the system of Mediterranean depressions (an element of the westerly Atlantic system) causing increased precipitation and decreased evaporation. Both the Indian Ocean summer monsoon, and the westerly Atlantic system, would be intensified in response to the occurrence of distinct minima in the cycle of precession.

It is demonstrated that reduced excess evaporation, whether or not coinciding with global phases of deglaciation, would lead to reduction of surface water salinities in the eastern Mediterranean, causing "isolation" of previously formed high salinity (cooler) deep water. Thus, mixing was severely reduced, possibly even restricted to eddy diffusion, with only occasional convective events that, because of the existing density gradient, hardly ever would reach the deepest parts of the basin. The consequently diminished oxygen advection down from a few hundred meters in the water column, favoured preservation of the sinking organic matter. This would, however, suffice only to enable the formation of sapropels with low organic carbon contents. The high organic carbon contents observed in various sapropels are argued to reflect superimposed increases of export production.

The described scenario accounts for previously reported (1) increases of organic carbon content with increasing depth of deposition within individual sapropels, and (2) asymmetrical sequences of sapropel deposition, characterized by gradual build-up and rather abrupt ending. It is, furthermore, in agreement with (3) isotopic and faunal reconstructions of the history of exchange transports through both the Strait of Sicily, and the Strait of Gibraltar, and (4) faunal and floral reflections of the presence or absence of a distinct Deep Chlorophyll Maximum with its associated increases in export production, indicating the presence or absence of a shallow pycnocline within the euphotic layer. Moreover, the described scenario (5) is in no way conflicting with the reports of sapropels in the western Mediterranean, and (6) seems to be endorsed by occasionally intercalated intervals suggestive of somewhat improved oxygenation, amidst anoxic (benthic desert) levels, a situation that has been observed in a few Quaternary and Pliocene sapropels.

1. Introduction

1.1. Eastern Mediterranean sapropels

In this paper, I review previous studies and introduce a new concept regarding the formation of sapropels in the eastern Mediterranean. Sapropels have been found in sedimentary sequences from the entire eastern Mediterranean. They are often well laminated, relatively dark brownish to black coloured, organic carbon-rich sediments, which contain more than average amounts of pyrite and benthic fauna indicative of low bottom water oxygen conditions (sometimes benthic fauna is completely absent) and notably high concentrations of elements like S and Ba. Sapropel occurrences and characteristics have been described and interpreted in terms of processes of formation and (syn-/postdepositional) alterations by numerous scientists, from a wide variety of disciplines (among many others, Kullenberg, 1952; Olausson, 1961; Van Straaten, 1966, 1972; Stanley et al., 1975; Cita et al., 1977; Vergnaud-Grazzini et al., 1977; Thunell et al., 1977; Stanley, 1978; Stanley and Maldonado, 1979; Mullineaux and Lohmann, 1981; Blechschmidt et al., 1982; Buckley et al., 1982; Cita and Grignani, 1982; Rossignol-Strick et al., 1982; Calvert, 1983; Thunell and Williams, 1983; Jenkins and Williams, 1984; Ross and Kennett, 1984; Shaw and Evans, 1984; Vergnaud-Grazzini, 1985; Mangini and Schlosser, 1986; Ganssen and Troelstra, 1987; Buckley and Johnson, 1988; Cramp et al., 1988; De Lange et al., 1989; Rohling and Gieskes, 1989; Nolet and Corliss, 1990; Murat, 1991; Pruysers et al., 1991, 1993; Van Os et al., 1991, 1994; Fontugne and Calvert, 1992; Perissoratis and Piper, 1992; Castradori, 1993; Rohling et al., 1993a,b). This study concentrates on sapropel formation in its relation to climate induced changes in hydrography and associated surface water productivity.

1.2. Present-day hydrography

The Mediterranean Sea is a marginal basin that is separated from the adjacent North Atlantic Ocean by the narrow Strait of Gibraltar which contains a 284 m deep sill (Bryden and Kinder,

1991). A topographically more complex sill-structure in the Strait of Sicily, which has an estimated average depth of 330 m (Wüst, 1961), but which may actually be as deep as 365 and 430 m in the two major channels (Garzoli and Maillard, 1979), separates the western and eastern Mediterranean subbasins. The exchange of water-masses through both straits is characterized by eastward surface flow and westward subsurface flow. This pattern of exchange results from net buoyancy loss in the basins on the easterly side of the sills. Surface water flowing in through the Strait of Gibraltar is traceable through the Strait of Sicily into the eastern Mediterranean, although its salinity increases steadily towards the east (Wüst, 1961; Malanotte-Rizzoli and Hecht, 1988; Malanotte-Rizzoli and Bergamasco, 1989).

Near Cyprus, in the eastern end of the eastern Mediterranean, surface water sinks to form Levantine Intermediate Water (LIW), which is the major constituent of the Mediterranean Intermediate Water (MIW) found throughout the eastern and western basins (Wüst, 1961; Malanotte-Rizzoli and Hecht, 1988; Malanotte-Rizzoli and Bergamasco, 1989). There is no equivalent formation of intermediate water in the western Mediterranean. The high salinity MIW accounts for a large part of the subsurface outflow across both the Sicilian sill and the sill at Gibraltar. In the eastern Mediterranean, MIW is found between about 150 and 600 m waterdepth (Wüst, 1960, 1961). The distinct halocline at about 150 m, which marks the transition between MIW and surface waters (Fig. 1), resides well below the euphotic layer (down to about 120 m). This situation inhibits substantial upward mixing of nutrients from the MIW into the euphotic layer, so that the MIW-surface water transition is associated with a nutricline between somewhat more nutrient-rich MIW and strongly depleted surface waters in most of today's eastern Mediterranean (McGill, 1961; Miller et al., 1970).

The above described vertical differentiation is complicated in the summer months, when the so-called "surface waters" overlying the permanent (halo-)pycnocline are subdivided by a very steep thermocline. In the Levantine basin near Crete and Rhodos [Fig. 2, after Wüst (1960)], for exam-
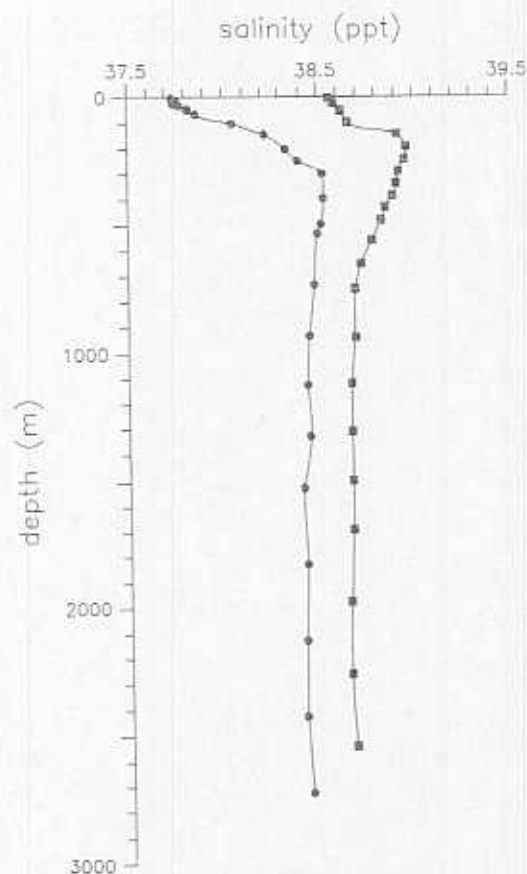
salinity (ppt)

Fig. 1. Typical salinity profiles for the western (dots) and eastern (filled squares) Mediterranean. Note the sharp decrease in salinity around 150 m, marking the top of Mediterranean Intermediate Water (after Rohling and Bryden, 1992).

ple, the very warm and also highly saline summer mixed layer above about 50 m is then separated from cooler waters below. These cooler waters below the summer thermocline still show the (permanent) halocline, separating relatively low salinity waters centered around 75 m from the high salinity intermediate water below about 150 m.

The eastern and western Mediterranean basins both have their own source of deep water, which settles below the MIW (resulting salinity profiles are shown in Fig. 1). Eastern Mediterranean Deep Water (EMDW) is formed in the Adriatic Sea. On the large north Adriatic shelf, surface waters of relatively low salinity (result of Po river dis-charge) are subject to cold and dry northeasterly "Bora" winds in the winter season. As a result, cold water with a somewhat increased salinity flows towards the deep south Adriatic Basin. The salinity of this watermass, however, remains lower than that of the MIW which penetrates the south Adriatic Basin across the Otranto sill between roughly 200 and 500 m. As the cold and relatively low saline watermass entering from the northern Adriatic mixes with MIW, which has a higher temperature and salinity, densification results in the formation of a water mass (EMDW) which, although of a somewhat lower salinity, is cooler and of higher density than MIW. As a consequence, EMDW settles below MIW, filling even the deepest parts of the eastern Mediterranean basin (Pollak, 1951; Wüst, 1961; Tchernia, 1980; Malanotte-Rizzoli and Hecht, 1988; Artegiani et al., 1989; POEM group, 1992).

Western Mediterranean Deep Water (WMDW) is formed in late winter in the northern sector of the western Mediterranean (so-called "MEDOC"-area), after a preconditioning phase during which the surface water density is increased by intense cooling and evaporation related to the northerly "Mistral" winds. As the density contrast between the surface water and the warm, high salinity MIW is eliminated, deep convective mixing of surface waters and MIW results in the formation of Western Mediterranean Deep Water (Wüst, 1961; MEDOC group, 1970; Stommel, 1972; Leaman and Schott, 1991).

EMDW and WMDW are effectively separated by the Sicilian sill, and WMDW is effectively separated from the Atlantic Ocean by the sill at Gibraltar. At present, strong net buoyancy loss from the basins, and resultant deep convective mixing in the Adriatic Basin and the "MEDOC" area, cause well oxygenated conditions down to the deepest parts of the eastern and western Mediterranean. Béthoux et al. (1990) estimated the flushing times of the western and eastern basins at 20 and 50 years, respectively. On the basis of chlorofluoromethane budget calculations, however, Schlitzer et al. (1991) argued that the renewal time for the entire deep eastern Mediterranean (>1500 m) should be more like 80 years.
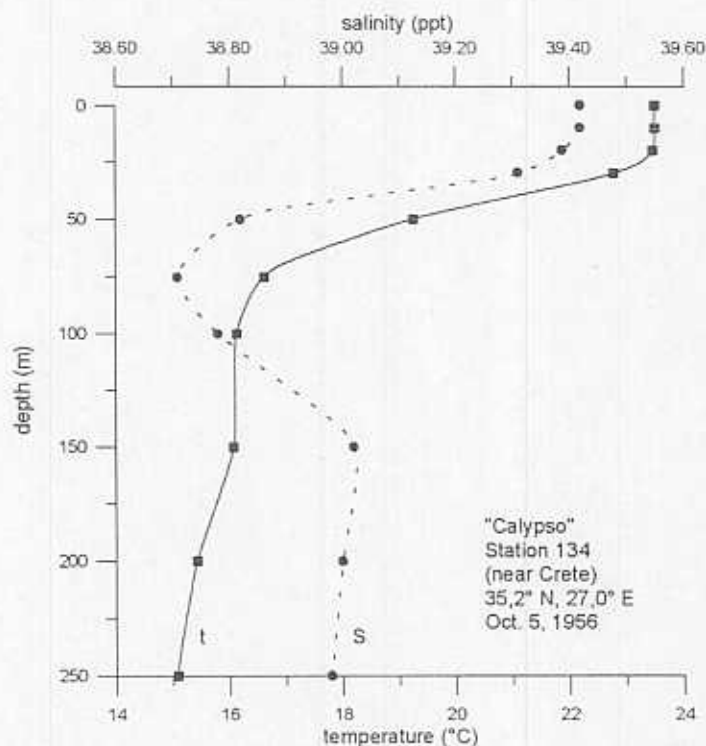
Fig. 2. Distribution of temperature and salinity in the upper 250 m near Crete. Note the very pronounced seasonal thermo-halocline at about 50 m depth, and the deeper (permanent) halocline around 125 m marking the top of MIW (after Wüst, 1960).

1.3. Timing and nature of past variations

Over the past 4 decades, a vast number of sediment cores has been recovered from the western and especially the eastern Mediterranean, most containing late Pleistocene sedimentary sequences. The most striking feature, superimposed on the glacial–interglacial cycles with a period of about 100,000 years, is the occurrence of often very distinct sapropels throughout the eastern Mediterranean. Sapropelic sediments are known from sediments as old as the Miocene (a.o. Cita and Grignani, 1982; Thunell et al., 1984; Howell et al., 1988; Hilgen, 1991). Although no equally well developed record of sapropels has been found in the western Mediterranean, the convincing discovery of sapropels in cores from the Tyrrhenian Sea (ODP Leg 107) demonstrated that sapropel formation was not restricted to the eastern Mediterranean (Thunell et al., 1990; Emeis et al., 1991). Earlier, Stanley (1978) reported radiocar-

bon dates of darker (olive-grey) layers containing pyritized burrows in cores from various basins in the western Mediterranean, evidencing regionally important reducing conditions. Stanley argued that these layers were formed at the same time as eastern Mediterranean sapropels. A few late Quaternary sapropels have been found in cores from the southern Balearic Basin (western Mediterranean), with radiocarbon ages very similar to those of eastern Mediterranean sapropels (Canals Artiguas, 1980).

The timing of sapropel deposition is thought to be astronomically controlled by minima in the cycle of precession, which occur about every 21,000 years (Rossignol-Strick et al., 1982; Rossignol-Strick, 1985; Prell and Kutzbach, 1987; Hilgen, 1991). The precession cycle reached its most recent minimum about 11,500 calendar years ago (Hilgen, 1991), corresponding to about 10,000 yr B.P. (^{14}C). According to modern AMS ^{14}C dates, without correction for reservoir age, the youngest

(Holocene) sapropel S_1 formed between about 8300 and 6300 yr B.P. (Jorissen et al., 1993), which is in reasonable agreement with previous datings (Stanley and Maldonado, 1979; Vergnaud-Grazzini, 1985; Lander Rasmussen, 1991; Troelstra et al., 1991; Perissoratis and Piper, 1992). High resolution studies, however, suggested that the processes leading to the formation of S_1 were initiated much earlier, namely at about 12,000 yr B.P. (Howell and Thunell, 1992), 12,700 yr B.P. (Rohling et al., 1993a), or even before 13,800 yr B.P. (Troelstra et al., 1991).

Tang and Stott (1993) showed that depletions in the $\delta^{18}O$ values of the planktic foraminifer *Globigerinoides ruber* (which thrives in the upper 25 m of the water column) started well before the formation of sapropel S_1. These authors did not perform AMS ^{14}C datings, but tentatively estimated an age of 13,000 yr B.P. for the early depletions in the $\delta^{18}O$ record of *G. ruber*. They furthermore found that it was not until (again estimated) 11,000 yr B.P. that a large change in surface water salinity occurred. Tang and Stott (1993) show the base of S_1 near 20 cm in their core TR171-27, and found early $\delta^{18}O$ depletion at about 26 to 28 cm. A large change of surface water salinity was noted at about 24 cm (11,000 yr B.P. according to Tang and Stott). Considering the aforementioned dates of Jorissen et al. (1993) for the base and top of S_1, found at about 20 and 14 cm, and assuming that the top of core TR171-27 corresponds roughly to 0 yr B.P., the average sedimentation rate would be between some 2.2 to 2.4 cm kyr^{-1}, which would be a rather typical rate for the Levantine basin. Applying the higher estimate (2.4 cm kyr^{-1}) to the interval below S_1, the early $\delta^{18}O$ depletion has an extrapolated age around 10,800–11,600, rather than 13,000 yr B.P. as proposed by Tang and Stott. The large change in surface water salinity noted at 24 cm would then be dated at about 10,000 yr B.P., rather than 11,000 yr B.P. Note that I assumed an age of roughly 0 yr B.P. for the core-top and 8300 yr B.P. for the base of S_1. In case TR171-27 would be characterized by incomplete recovery of the core-top, or if a reservoir-age of several hundreds of years would be incorporated in the calculations, the average sedimentation rate

would be somewhat higher and the extrapolated ages, therefore, would tend to become younger, increasing the offset with the dates proposed by Tang and Stott (1993).

Detailed sedimentological descriptions indicate that the depositional sequence of sapropels is characterized by a gradual build-up, with an onset well below the base of the actual sapropel, resulting in the formation of so-called organic oozes, or protosapropels (Stanley and Maldonado, 1979; Anastasakis and Stanley, 1984; Murat and Got, 1987; De Lange et al., 1989; Murat, 1991). On the basis of chemical arguments, De Lange et al. (1989) reasoned that the sequence of gray-green oxidized clays, via increasingly darker hues of green in the protosapropel, to the dark green/black color of the sapropel, resulted from gradually decreasing bottom water oxygen concentrations. Changes in benthic fauna preceding the deposition of a sapropel seem to indicate, at least in the case of several sapropels in various study areas, reduced deep water ventilation already before the deposition of the true sapropel (a.o., Van Straaten, 1972; Cita and Podenzani, 1980; Mullineaux and Lohmann, 1981; Vismara-Schilling, 1984; Jorissen et al., 1993).

The base of the protosapropel below S_1, as determined from X-ray photographs, was dated by Troelstra et al. (1991) at about 11,000 yr B.P. ($^{14}C_{nc}$; nc = not corrected for reservoir age). On the basis of color characteristics, however, those authors considered this to be some boundary within a more extensive protosapropel, which would extend even further down, beyond the level they dated at 13,800 yr B.P. Jorissen et al. (1993) reported distinct numbers of benthic foraminifera typical for low oxygen conditions already before the sapropel itself developed. A more detailed appraisal of the benthic record in the well dated Adriatic Sea core IN68-9 showed that low-oxygen conditions developed about 1000 years before the formation of S_1 (around 9300 yr B.P. $^{14}C_{nc}$), while severe dysoxia developed 200–300 years before the onset of S_1, namely around 8500–8600 yr B.P. $^{14}C_{nc}$ (Rohling et al., submitted). Note that from the $^{14}C_{nc}$ datings, a reservoir age of about 400 years should be substracted at these mid-latitudes

(Bard, 1988; Broecker et al., 1988; Bard et al., 1990), giving $^{14}C_c$ values.

The ending of sapropel formation has recently been the focal point of various studies. Sedimentological arguments led Murat and Got (1987) to conclude that sapropel formation ended abruptly. However, the sharp color change, which Murat and Got (1987) considered to be the top of the sapropel, may well be a relict redox front produced by downward progressing oxidation ("burning down") after the return of oxygenated bottom waters (a.o., De Lange et al., 1989; Pruysers et al., 1991, 1993; Higgs et al., 1994). This interpretation was more or less endorsed by micropaleontological evidence, which showed that associations typical for the sapropel continued in some cases above the presently visible top (Lander Rasmussen, 1991; Rohling et al., 1993a). Higgs et al. (1994) proposed that the original top of sapropel S_1 resided at the location of the youngest of two Mn peaks they found above that sapropel in an Ionian and an eastern Levantine core, so that sapropel formation would have continued until about 5000 yr B.P. (age on the basis of assumed 8600 yr B.P. for base of S_1 and constant accumulation rate), but also suggested that this inference ought to be tested by analyses of benthic foraminiferal associations. In contrast to Higgs et al. (1994), Pruysers et al. (1993) related the original top of the sapropel to the older of the two Mn peaks in their core 9G1 recovered southeast of Crete. They explained the second (youngest) Mn peak with a process of non-steady state formation of double peaks in response to an increased organic carbon accumulation between 1900 and 1350 yr B.P. In a recent study, Rohling et al. (submitted) did what Higgs et al. (1994) suggested to solve the problem, and presented results of high resolution benthic foraminiferal countings across S_1. The analyses were performed on the well (AMS ^{14}C) dated core IN68-9 from the Adriatic Sea with a sedimentation-rate of about 15 cm kyr^{-1} (Jorissen et al., 1993). In short, the benthic foraminiferal record of Rohling et al. (submitted) shows low-oxygen stress building up since about 9300 yr B.P. ($^{14}C_{nc}$), some 1000 years before the onset of S_1 (8300 yr B.P. $^{14}C_{nc}$), and ending at the very top of the lithological expression of S_1 (6300

yr B.P. $^{14}C_{nc}$). Therefore, the results contradict the hypothesis of Higgs et al. (1994) that sapropel formation lasted until about 5000 yr B.P., instead showing that sapropel formation (revealed by low-oxygen stressed benthic faunas) ended around 6300 yr B.P., or 5900 yr B.P. when corrected for reservoir age.

Today, with the precession cycle near a maximum, the deep eastern Mediterranean is well oxygenated, except for some small, brine-filled, anoxic basins (De Lange and Ten Haven, 1983; Jongsma et al., 1983; Scientific Staff Cruise BAN84, 1985). Oxygenated conditions returned in the deep eastern Mediterranean around 6500–6000 yr B.P. ($^{14}C_{nc}$) (Troelstra et al., 1991; Lander Rasmussen, 1991; Jorissen et al., 1993). On the basis of changes in the Adriatic benthic foraminiferal record, Jorissen et al. (1993) argued that modern conditions were at least achieved by about 4000–3500 yr B.P.

A minimum in the precession cycle, characteristic of times of sapropel formation, implies stronger insolation in Northern Hemisphere summer and weaker insolation in Northern Hemisphere winter, compared to the present, causing increased seasonal and land–sea temperature contrasts on the Northern Hemisphere (Kutzbach, 1985; Kutzbach and Guetter, 1986). As a consequence, monsoonal circulation would be intensified in summer, resulting in increased precipitation over eastern equatorial Africa, which, in turn, would cause increased discharge of the Nile river into the eastern Mediterranean. The importance of this chain of events for the generation of anoxia in the Mediterranean has been described first by Rossignol-Strick et al. (1982), and more elaborately by Rossignol-Strick (1985).

Rohling and Hilgen (1991) reviewed reconstructions of past climatic variations in the Mediterranean realm. They summarized that, at times of minima in the precession cycle, not only monsoonal precipitation over eastern equatorial Africa had increased, but precipitation in the Northern Borderlands of the Eastern Mediterranean (NBEM) as well (Shaw and Evans, 1984; Rossignol-Strick, 1987; Cramp et al., 1988; Wijmstra et al., 1990). The review emphasized that increased precipitation in the NBEM can not be

explained by intensification and areal expansion of the Indian Ocean summer monsoon into the eastern Mediterranean, so that the influence of monsoonal intensification on the eastern Mediterranean hydrography seems to have been restricted to that exerted by increased Nile discharge. It was concluded that increased precipitation in the NBEM, at times of minima in the precession cycle, most likely resulted from increased activity of Mediterranean depressions.

Today, Mediterranean depressions develop over the northern part of the western Mediterranean, as a result of incursions of cold (polar origin) air into that region, in the wake of larger Atlantic depressions tracking eastwards north of the Alps (Trewartha, 1966; Boucher, 1975). At times of sapropel formation, the westerly track of cyclones (depression track) over the Atlantic Ocean would have shifted considerably northwards, up to about 70°N (see also Kutzbach and Guetter, 1986), much as it had at medieval times (Lamb, 1966, 1969, 1977; Flohn, 1981). As a consequence, a secondary depression track could have developed across the Mediterranean (Lamb, 1966, 1969, 1977; Boucher, 1975). In addition, increased northward heat transport in the North Atlantic, in response to seemingly increased northward advection of heat in the South Atlantic and into the equatorial Atlantic at times of minima in the precession cycle (McIntyre et al., 1989), would tend to increase the temperature contrast across the Gulf Stream and North Atlantic Drift, and the evaporation from warmer waters. Combined, these increases could account for increased formation of Atlantic depressions with enhanced moisture transport towards the east.

1.4. Approaches to explain sapropel formation

Numerous studies have focused on explaining the formation of eastern Mediterranean sapropels, since their first discovery in cores collected during the 1947–1948 Swedish Deep Sea Expedition (Kullenberg, 1952). Olausson (1961) reasoned that development of stagnant deep water layers, resulting in deep water anoxia and consequent formation of sapropels, was induced by massive outflow of relatively fresh water from the Black Sea, in response to establishment of a marine

connection between the eastern Mediterranean and the Black Sea by post-glacial sea-level rise. Recent AMS ^{14}C datings allow detailed comparison of differences in the timing of events in the eastern Mediterranean and the Black Sea. Jorissen et al. (1993) dated the onset of sapropel formation in the eastern Mediterranean at about 8300 yr B.P. ($^{14}C_{nc}$), i.e., 7900 yr B.P. ($^{14}C_c$). Jones and Gagnon (1994) dated the onset of the recent sapropel formation in the Black Sea at about 7350 yr B.P. ($^{14}C_c$). This onset would be a response to re-establishment of a marine connection (sea level at -34 m) with the Mediterranean through the Bosporus Strait that would have occurred around 9500 yr B.P. ($^{14}C_c$; cf. Fairbanks, 1990; Jones and Gagnon, 1994). These figures do not exclude a possible importance of Black Sea overflow in the case of S_1, although it would remain to be explained why—considering the $^{14}C_c$ ages—the opening of the marine connection through the Bosporus Strait would have preceded sapropel formation in the Black Sea by some 2150 years, and in the eastern Mediterranean by only 1600 years.

Regarding sapropel formation in general, faunal and oxygen isotope studies showed that sapropels formed not only during Quaternary warming phases, but also (less frequently) during cool episodes (a.o. Cita et al., 1977; Vergnaud-Grazzini et al., 1977; Cita and Grignani, 1982; Thunell et al., 1983; Thunell et al., 1984; Muerdter and Kennett, 1984; Vergnaud-Grazzini, 1985). Nevertheless, oxygen isotope depletions in many sapropels suggest that eastern Mediterranean surface water salinities were relatively low (Cita et al., 1977; Vergnaud-Grazzini et al., 1977; Williams et al., 1978; Thunell et al., 1987; Thunell and Williams, 1989). The study of Tang and Stott (1993) suggests that during the deposition of the Holocene sapropel S_1, the main depletion was located at the very surface, while its effects were still noticeable down to about 75 m depth. The fact that sapropels have formed both during interglacial and glacial intervals at times of minima in the precession cycle suggests that the major disturbances in the freshwater budget inducing the formation of most sapropels were presumably not related to Black Sea overflow. Rather, these distur-

bances should be ascribed to increased discharge from the Nile (Rossignol-Strick et al., 1982; Rossignol-Strick, 1985) and Eurasian rivers (Shaw and Evans, 1984; Rossignol-Strick, 1987; Cramp et al., 1988; Wijmstra et al., 1990). In addition, increased activity of Mediterranean depressions would have caused increased precipitation over, and possibly decreased evaporation from, the eastern Mediterranean surface waters.

De Lange and Ten Haven (1983) concluded that organic production was substantially increased in the eastern Mediterranean at times of sapropel formation. Calvert (1983) reported high concentrations of Ba in Pleistocene sapropels, which he interpreted as evidence of increased production during their deposition. Van Os et al. (1991), who sampled several sapropels on millimeter scales, argued that the "high Ba" argument for increased production during sapropel formation may well be correct, but also called for careful consideration of Ba-relocation and precipitation processes during diagenesis, before using the distribution of Ba in sediments as a strict measure of (paleo)productivity. An evaluation of recent sapropel formation in the Black Sea suggested that high productivity, rather than bottom water anoxia, controls the enrichment of organic carbon in sediments, which would imply a need for high productivity during the deposition of eastern Mediterranean sapropels (Pederson and Calvert, 1990). Boyle and Lea (1989) concluded on the basis of Cd/Ca ratios in the planktic foraminiferal species *Globigerinoides ruber*, that near surface phosphate concentrations were substantially higher in the eastern Mediterranean at times of sapropel deposition than today. Van Os et al. (1994) presented a high-resolution micropaleontological and geochemical study across the first sapropel-containing carbonate cycle in the Pliocene of Sicily, and inferred that high production of siliceous organisms prevailed during deposition of that sapropel, in spite of the absence of opal in the sediment (dissolved). The authors presented this conclusion following a scenario that convincingly unites their wide range of analyses with previous work on grain size and clay-mineralogy (a.o., De Visser et al., 1989).

Since various studies of Late Quaternary sapropels suggest a general increase in river discharge

(see above), riverine nutrient influxes are often considered to be rather important for the development of the increased production that caused sapropel formation, even if the various authors disagree on the importance of a concomitant decrease in deep water ventilation (a.o. Rossignol-Strick et al., 1982; De Lange and Ten Haven, 1983; Ten Haven, 1986; Boyle and Lea, 1989; Rohling and Gieskes, 1989). Indications for increased river-discharge were also listed by Rohling et al. (1993b), based on a high resolution study across Upper Pliocene sapropel C2 from the Singa section (southern Italy).

Alternatively, a large number of studies proposed a potential reversal of the eastern Mediterranean circulation, from anti-estuarine to estuarine, in response to an alleged switch in the eastern Mediterranean freshwater budget from negative to positive (a.o., Stanley et al., 1975; Stanley, 1978; Calvert, 1983; Thunell et al., 1983, 1984; Muerdter and Kennett, 1984; Ten Haven, 1986; Buckley and Johnson, 1988; Thunell and Williams, 1989; Howell and Thunell, 1992). Such a reversed (estuarine) type of circulation would cause a "nutrient-trap" situation, and a possible concomitant decrease in oxygen advection to deeper water in the eastern Mediterranean, resulting in the deposition of sapropels. The most coherent case for this reversed scenario, made by Sarmiento et al. (1988) on the basis of a box model for phosphate distribution, will be discussed extensively in Section 1.5.

The depletion of $\delta^{18}O$ values found in most sapropels (a.o., Cita et al., 1977; Vergnaud-Grazzini et al., 1977; Thunell and Williams, 1983; Muerdter and Kennett, 1984; Thunell et al., 1984; Vernaud-Grazzini, 1985; Thunell and Williams, 1989; Tang and Stott, 1993) has frequently been used to infer the presence of a "low-salinity cap", consisting of sea water with lower salinities than commonly found in the Mediterranean. In the various theories, the formation of a "low-salinity cap" would have led to highly stable stratification with stagnant deep waters (Olausson, 1961; Cita et al., 1977; Vergnaud-Grazzini et al., 1977; Cita and Grignani, 1982; Vergnaud-Grazzini, 1985), or to a reversal of circulation (see references in previous paragraph). Tang and Stott (1993) suggested

that the intensity of surface water $\delta^{18}O$ depletion may have been highly variable on seasonal time scales, to the point that it may at times have been completely absent. They concluded that there is no reason to assume that a "low-salinity cap" was present without interruption. Rather, Tang and Stott (1993) suggested that returns to relatively normal salinity conditions occurred seasonally every year or seasonally on an interannual basis. Vergnaud-Grazzini et al. (1988) ascribed the isotope depletions in the early Holocene sapropel S_1 to influx of relatively low salinity Atlantic water, in addition to increased run-off or Black Sea overflow.

1.5. Estuarine or anti-estuarine circulation during sapropel formation?

1.5.1. Nutrient distribution and implications for the estuarine scenario

Using a phosphate based box model, Sarmiento et al. (1988) concluded that sapropel formation in the eastern Mediterranean may be explained only by assuming an estuarine circulation, or an extremely weak anti-estuarine circulation. Their key argument for this conclusion was the fact that phosphate export from the eastern into the western basin should have remained very low, to account for the absence of sapropels in the western Mediterranean, which they incorrectly inferred from the literature. Rohling (1991b) commented that limited phosphate export could be equally well fulfilled in an anti-estuarine circulation that was still distinct, although weakened relative to the present, provided that the decrease of water exchange across the sill resulted predominantly from decreased contribution of EMDW to the subsurface outflow (in response to decreased new formation of EMDW).

More significantly, however, the assumption of limited phosphate export itself seems rather debatable. More phosphate may have been exported by subsurface outflow of MIW across the Sicilian sill than assumed by Sarmiento et al. (1988), without enhancing production throughout the western Mediterranean. This would have been possible if the nutrients in the western basin remained shielded from the euphotic layer by a relatively

deep pycnocline. In such a scenario, nutrient enrichment in MIW would only boost western Mediterranean production in certain specific areas where MIW would reach the euphotic layer. This seems to be confirmed by peak abundances of the planktic foraminiferal genus *Neogloboquadrina* in the interval corresponding roughly to S_1, in core BS78-12 from the northern Tyrrhenian Sea (Jorissen et al., 1993). Also supporting that scenario, Vergnaud-Grazzini et al. (1989) concluded from stable isotope records on both sides of the Strait of Gibraltar, that MIW was nutrient enriched and contained less oxygen at times of sapropel formation in the eastern Mediterranean, relative to the present.

The above discussed findings of sapropels and sapropel-like intervals in the western Mediterranean appear to provide further counter-evidence for the limited phosphate export assumption of Sarmiento et al. (1988). Apparently, more phosphate was actually exported from the eastern Mediterranean than assumed by those authors. This, in turn, indicates that there is no need for a severe constraint on exchange transports through the Strait of Sicily, as formulated by Sarmiento et al. (1988). Therefore, sapropel formation may have occurred in an anti-estuarine circulation. In other words, to induce the formation of sapropels, less dramatic changes in the eastern Mediterranean freshwater budget would seem to be required than those proposed by Sarmiento et al. (1988).

1.5.2. Microfossils and implications for the anti-estuarine scenario

Alternative to the circulation reversal hypothesis, Rohling and Gieskes (1989) provided a working model for sapropel formation in a semi-enclosed basin with a distinct anti-estuarine circulation, albeit weakened in comparison to that of today. The authors proposed a combination of increased production and decreased deep water ventilation to explain the formation of sapropels in a basin with anti-estuarine circulation. On the basis of the common abundance (Fig. 3) of *Neogloboquadrina* in most late Quaternary sapropels throughout the eastern Mediterranean basin (see also Kullenberg, 1952; Cita et al., 1977; Thunell et al., 1977; Vergnaud-Grazzini et al.,
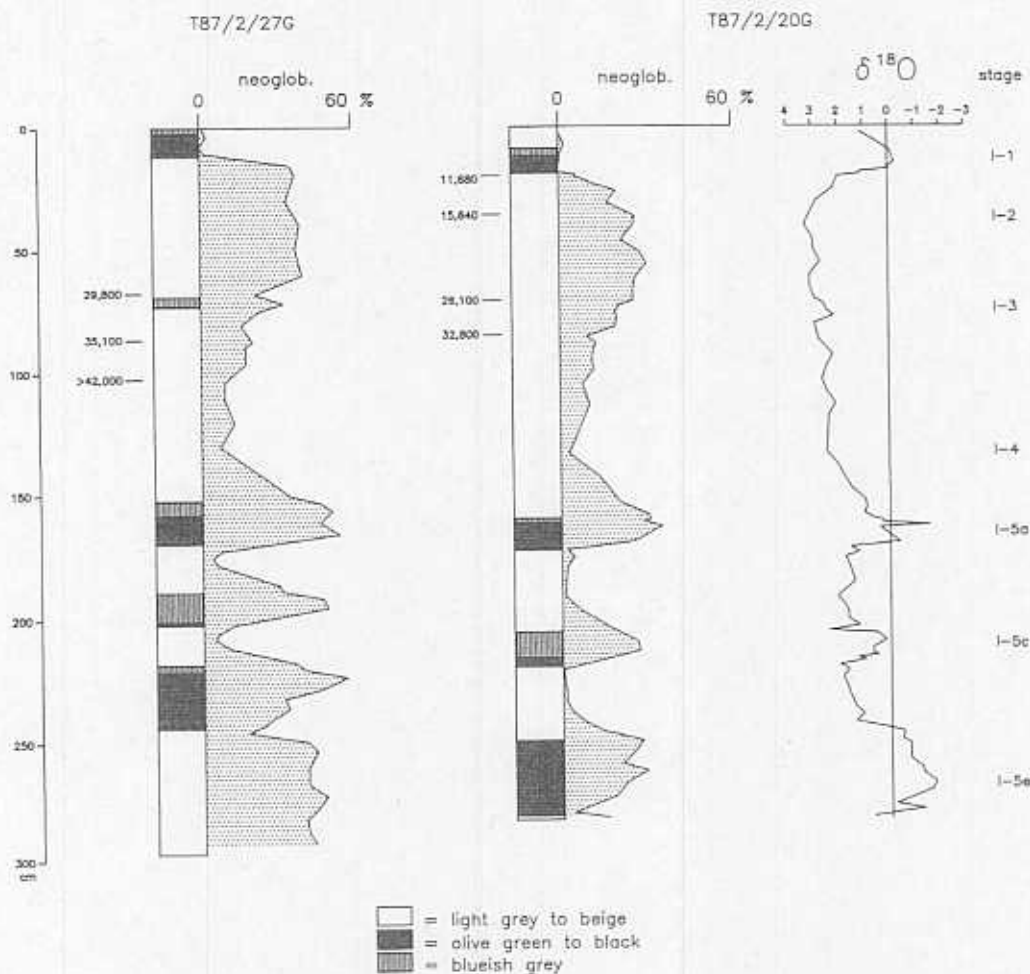
Fig. 3. Downcore neogloboquadrinid percentages in two cores from the Northern Levantine basin, with stable oxygen isotope stratigraphy for core T87/2/20 (after Rohling et al., 1993a). Note the commonly high abundances in most sapropels except the youngest (S₁). This distribution pattern was found to be typical for the entire eastern Mediterranean (for table of comparison see Rohling and Gieskes, 1989). Small numbers refer to AMS ¹⁴C datings.

1977), and relying on studies on the present-day habitat of this group of planktic foraminifera (Fairbanks et al., 1982; Hemleben and Spindler, 1983; Thunell and Reynolds, 1984; Bé et al., 1985; Hemleben et al., 1989; Pujol and Vergnaud-Grazzini, 1989; Reynolds and Thunell, 1989), it was argued that a basin-wide Deep Chlorophyll Maximum (DCM) had developed in response to general shoaling of the eastern Mediterranean (halo-) pycnocline (which separates the intermediate and surface waters) and the associated nutricline, to a depth within the euphotic layer (Fig. 4).

The development of a DCM, in which nutrient recycling is far less efficient than in the oligotrophic mixed layer, would provide an effective means of increasing export production and, thus, the flux of organic matter towards the sea floor. The hypothesis of pycnocline shoaling at times of sapropel formation was supported by studies of the total planktic foraminiferal fauna (Rohling et al., 1993a) and of calcareous nannofossil records (Castradori, 1993), while a stable isotope study using planktonic foraminifera with different depth-habitats suggested similar conditions during the deposition
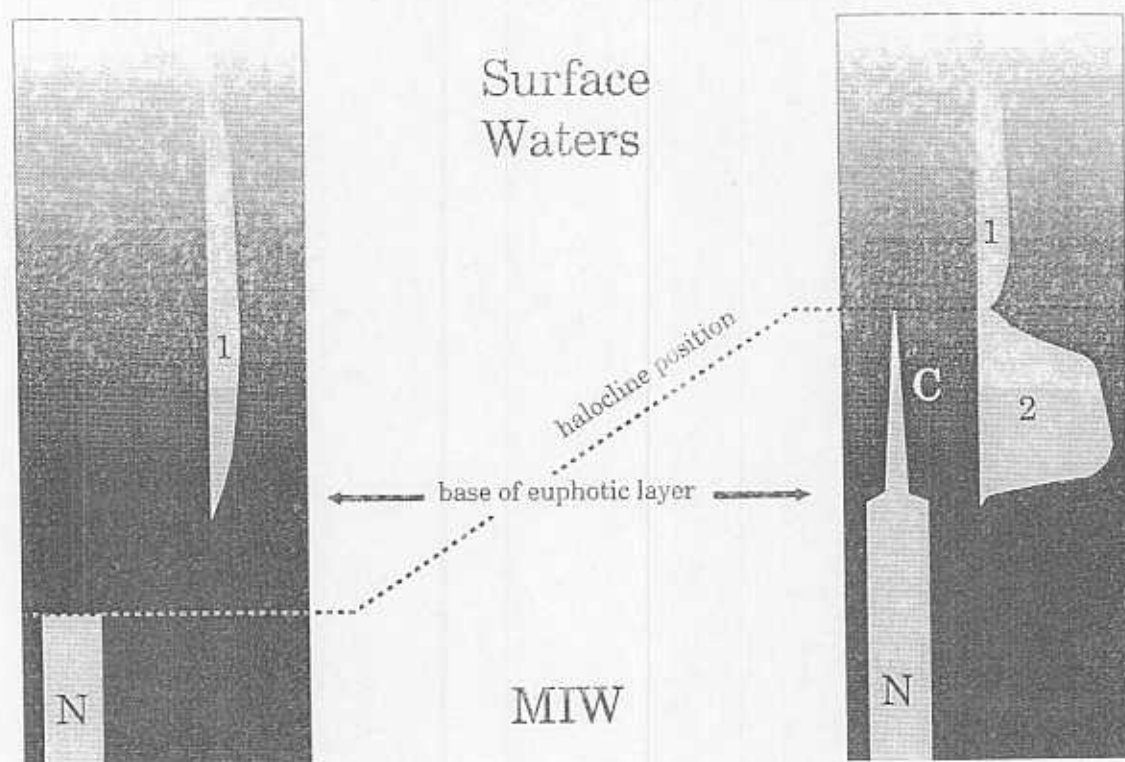
Fig. 4. Schematic representation of the influence of depth variations of the boundary between MIW and surface waters relative to the base of the euphotic layer, on the development of a distinct Deep Chlorophyll Maximum (cf. Rohling and Gieskes, 1989). *1* = mixed layer phytoplankton association; *2* = Deep Chlorophyll Maximum association; *C* = in situ autotrophic consumption of nutrients which causes the nutricline to remain at the base of the euphotic layer.

of sapropels S_6 and, especially, S_7 (Ganssen and Troelstra, 1987).

Rohling and Gieskes (1989) also inferred pycnocline shoaling for full glacial times, also characterized by abundant *Neogloboquadrina*. Rohling (1991a) calculated that glacial pycnocline shoaling would be a very likely result of sea-level lowering, while Rohling (1991b) showed that pycnocline shoaling at times of sapropel deposition would be primarily related to decreased excess evaporation. These conclusions were based on simple two-layer models for circulation and hydrography in the semi-enclosed Mediterranean (see next chapter). The fact that glacial intervals are not entirely characterized by anoxic sedimentation indicates that pycnocline shoaling only provides a means to increase export production, and not a unique mechanism to trigger sapropel formation.

Apparently, some additional factor is needed to cause sapropel formation, and this additional factor was not (sufficiently) operative throughout the glacial periods. It was suggested that this additional factor consists of reduction of the density of newly formed deep water, in response to decreased buoyancy loss.

There is one seemingly disturbing peculiarity, namely the absence of *Neogloboquadrina* from the early Holocene sapropel S_1. In contrast to sapropel S_5, which is also positioned in the aftermath of a major deglaciation (Termination II), and which contains abundant *Neogloboquadrina*, sapropel S_1 is completely barren of this group of planktic foraminifera, with the exception of a brief recurrence in its topmost part (Rohling and Gieskes, 1989; Rohling et al., 1993a; Jorissen et al., 1993). The virtual absence of neogloboquadrinids in S_1

demonstrates that development of a DCM is not a strict prerequisite for sapropel formation, but only a convenient way for increasing export production. Note that the inferred stronger development of export production at times of the formation of S_5 seems to be supported by much higher preserved organic carbon contents in S_5, relative to S_1 (Cita and Grignani, 1982; Pruysers et al., 1991; Fontugne and Calvert, 1992; Calvert et al., 1992).

A recent study of calcareous nannofossil records, especially that of *Florisphaera profunda*, from four cores along a NW–SE transect in the eastern Mediterranean (Castradori, 1993), provided strong support for the scenario for sapropel formation proposed by Rohling and Gieskes. Moreover, this study included sapropels S_6, S_7, and S_8, and Castradori (1993) argued that those sapropels also bear evidence of distinct pycnocline shoaling. In the cases of S_6 and S_8, Castradori argues that pycnocline shoaling would have been most extreme. Castradori furthermore concluded that some influence of pycnocline shoaling is reflected within S_1 in the easternmost core of his transect, whereas no such influence was recorded in S_1 more towards the west. As a result, the pycnocline position at times of the formation of S_1 could be much better constrained than done by Rohling and Gieskes (1989), bearing in mind that there is a gentle average westward deepening of MIW (Wüst, 1961). During the formation of S_1, shoaling of the pycnocline appeared to have been less pronounced than during the formation of S_3 to S_8, so that only its shallowest (easternmost) part reached a position within the euphotic layer, whereas more westward the pycnocline failed to reach the base of the euphotic layer. The lack of confirmation for full glacial pycnocline shoaling in the record of *Florispaera profunda* (Castradori, 1993) possibly results from too low temperatures for that species.

1.5.3. Summary of arguments against the estuarine, and in favor of the anti-estuarine scenario

The popularity of the circulation reversal theory tends to conceal a couple of arguments pleading against it.

Firstly, the presence of sapropels in the western

Mediterranean would necessitate circulation reversal in the entire Mediterranean. This is, however, very unlikely in view of recent reconstructions of the intensity of subsurface Mediterranean Outflow through the Strait of Gibraltar during the past 140,000 years. These reconstructions suggest that subsurface outflow continued, although weakened at times of sapropel formation (Zahn et al., 1987; Vergnaud-Grazzini et al., 1989). In the anti-estuarine scenario for sapropel formation, on the contrary, the presence of western Mediterranean sapropels is not surprising. Not only in the eastern, but also in the western basin would deep water formation be affected by changes in the salinities of surface and intermediate waters. Nutrients contained by MIW could equaly well boost western Mediterranean production, if "transported" into the euphotic layer. Actually, the anti-estuarine scenario would be rather debatable if no sapropels, or other indications of increased production, were found in the western Mediterranean.

Secondly, changes in not only intensity, but also in nutrient content and oxygenation of MIW and Mediterranean Outflow Water (Zahn et al., 1987; Vergnaud-Grazzini et al., 1989), contradict with a circulation reversal and strongly support the anti-estuarine scenario. Moreover, Ross and Kennett (1984) argued that low-oxygen intermediate water flowed from the eastern into the western Mediterranean, at times of eastern Mediterranean sapropel formation. Their conclusions were based on a benthic foraminiferal study of cores from the Strait of Sicily, and suggest that a more or less present-day type of circulation prevailed.

Thirdly, to discriminate on the basis of $\delta^{18}O$ values whether or not circulation reversal occurred, $\delta^{18}O$ values in sapropels should be documented versus time-equivalent values for MIW (using deep living foraminifera) and/or time-equivalent values for surface water in the western Mediterranean. Only a few studies in this direction have been presented (for instance Jenkins and Williams, 1984; Thunell and Williams, 1989). Since $\delta^{18}O$ depletions observed to date have been used in the literature to underscore both the reversed and the anti-estuarine scenario, it seems that the traditional approach is not sufficiently conclusive to discriminate between these two options. A promising new

development concerns the analyses of the isotopic composition of planktic foraminiferal species with different depth-habitats from individual samples, to reconstruct vertical gradients in the eastern Mediterranean (Ganssen and Troelstra, 1987). Following this approach, Tang and Stott (1993) concluded that their results are in no way suggestive of a permanent low-salinity "cap" during sapropel formation, but rather point to season-bound returns to relatively normal surface water salinities. Even more complicating, some studies suggest that not all sapropels contain depletions in $\delta^{18}O$ (Cita and Grignani, 1982; Jenkins and Williams, 1984; Emeis et al., 1991). In a preliminary study, Van Os and Rohling (1993) calculated that an interrelated combination of (1) weakened anti-estuarine circulation with both (2) reduced evaporation and (3) increased precipitation, can provide the necessary boundary conditions to explain even the largest of Late Quaternary depletions ($-2.5\permil$ in S_5).

Fourthly, the apparent asymmetry in the depositional sequence of sapropels, as discussed in Section 1.3, has not yet been sufficiently addressed in the existing models for sapropel formation. Asymmetry seems hard to explain with a circulation reversal scenario, whereas it is implicitly related to the anti-estuarine scenario, as I will demonstrate in this paper. A similar argument holds for the observed increase of preserved organic carbon contents with increasing depth of deposition, within individual sapropels (Anastasakis and Stanley, 1984; Anastasakis, 1988; Murat, 1991).

1.6. Towards a comprehensive model

In the present paper, I use a simple and schematical graphic presentation to demonstrate how essentially asymmetrical sapropel sequences may be explained in a basin with anti-estuarine circulation. It is discussed how deglaciation (sea-level rise + global surface water salinity decrease) in combination with periodically decreased excess evaporation, or periodically decreased excess evaporation by itself, would result in development of stable stratification, isolating deep water layers from the vertical circulation. Then, a simple

parameterization of eddy (turbulent) diffusion is used to break down this stable stratification with time, until the stability is reduced sufficiently to enable the return of convective mixing throughout the water column.

The results describe the development and disruption of stable stratification with time. First, such a description is made for the interval spanning the past 18,000 years, containing the early Holocene sapropel S_1. That scenario serves as the example for interaction between deglaciation and changing excess evaporation. Then, a hypothetical case of a full glacial sapropel is discussed, in a scenario forced exclusively by reduced excess evaporation.

Both scenarios show a distinctly asymmetrical sequence for oxygen advection to bottom waters; even in the second scenario, which is forced by a completely symmetrical disturbance in excess evaporation.

2. Method

2.1. Concept

Since faunal and floral evidence points towards some basin-wide process of eutrophication in the eastern Mediterranean during the formation of most sapropels and also during full-glacial intervals, Rohling and Gieskes (1989) and Castradori (1993) proposed a basin-wide mechanism that would bring nutrients to the euphotic layer. Direct local input from, for example, rivers would not explain the vast geographical extent of the observed eutrophication. Moreover, the eutrophication should be caused by a sufficiently persistent long-term process, to account for the vast time-spans during which its influence has been continuously recorded [the oligotrophic conditions characterizing the basin at present are the exception rather than the rule, considering the record of the past 150,000 years (Fig. 3)]. The proposed mechanism of pycnocline shoaling fulfils these requirements. Note that the pycnocline discussed in these studies, and the related two-layered models (Rohling, 1991a,b), refers to the permanent (halo-)pycnocline between the intermediate and surface waters. The studies were not meant to

address or model the complexities of the seasonal mixed layer, for which accurate estimates of (preferably monthly) values of wind-stress and temperature would be required, both of which are extremely difficult to assess from the geological record. Instead, the emphasis was on the long-term average state. That long-term average state was argued to be expressed by the depth of the permanent (halo-)pycnocline. In other words, the thermal fluctuations associated with the summer–winter alternations and resulting in the development and disappearance of a seasonal thermocline (Fig. 2) are ignored initially. These fluctuations, however, have been discussed at a later stage, after the long-term response of the basin had been assessed. This discussion of seasonal variability was one of the main problems addressed by Rohling et al. (1993a), using planktic foraminiferal records from the Adriatic and Levantine Basins.

On the basis of mass and salt conservation statements and a simple parameterization of buoyancy loss by net evaporation, Rohling (1991a) developed a model to determine the depth of the eastern Mediterranean pycnocline between surface and intermediate waters during glacial times with a sea-level lowering of 100 m, relative to the present. In this model, the freshwater budget of the eastern Mediterranean was considered to have remained equal to its present-day value, despite fossil evidence of possibly increased aridity around the Mediterranean at glacial times (Rognon and Williams, 1977; Sarnthein, 1978; Street and Grove, 1979; Guiot, 1987; Magaritz and Goodfriend, 1987; Wijmstra et al., 1990; Cheddadi et al., 1991). In a study complementary to the former, Rohling (1991b) focused on variations in the eastern Mediterranean pycnocline depth that would result from changes in the basin's freshwater budget, assuming that sea level remained constant. The models are two-layered, with upper layer (surface) water flowing in through the Strait of Sicily, and deeper layer (intermediate) water flowing out of the basin through that strait. Surface layer water is transformed into deeper layer water by a general loss of buoyancy, related to the excess of evaporation over freshwater input. The effects of ignoring thermal changes were considered afterwards.

Rohling and Bryden (1994) combined the two earlier models, using a hydraulic control model for exchange through the Strait of Gibraltar (Bryden and Kinder, 1991) to determine to what extent sea-level variation would influence the volume of surface water flow into the eastern Mediterranean. Sea-level variations for the past 18,000 years were used as described by Fairbanks (1989). The resulting model (summarized in Fig. 5) relates sea-level fluctuations and changes in the freshwater budget to the depth of the pycnocline in the eastern Mediterranean. The model was applied to estimate the conditions characterizing the basin at 9600 yr B.P. At that time, the abrupt disappearance of *Neogloboquadrina* from the planktic foraminiferal record suggests that the eastern Mediterranean pycnocline crossed the base of the euphotic layer, sinking from a shallow glacial position to a deeper Holocene position below the base of the euphotic layer. It was calculated that excess evaporation was in the order of 20% higher than at the present. Unfortunately, the authors overlooked the fact that they were comparing the dating of an event in $^{14}C_{nc}$ years (cf. Jorissen et al., 1993) with a sea-level curve plotted versus $^{14}C_c$ years (Fairbanks, 1989). After correction for a reservoir age of 400 years, the proper age for comparison of the disappearance of *Neogloboquadrina* with the sea-level curve is 9200 yr B.P. instead of 9600 yr B.P., corresponding to a sea-level position of about −40 m instead of −50 m. Thus, the calculated excess evaporation should be only 5% higher than present, instead of 20% (see Table 1).

In the present paper, the same method (Fig. 5) is applied to more time-slices for which the (approximate) depth of the pycnocline can be constrained, between the Last Glacial Maximum (LGM) and Present. I rearrange the equations to solve for the salinity of Mediterranean Intermediate Water (S_{MIW}). Since the high salinity MIW is a major component of the Eastern Mediterranean Deep Water formed in the Adriatic Basin, S_{MIW} strongly determines the salinity of EMDW (S_{EMDW}).

On the basis of Wüst's (1961) description of Mediterranean water masses, Mangini and Schlosser (1986) argued that a 0.2‰ lowering of S_{EMDW} would suffice to lower its density such that

Rohling and Bryden [1994] model for the eastern Mediterranean.

Two-layered. Inflow of surface layer water and outflow of deeper layer water, through Strait of Sicily.

Net buoyancy loss (B) proportional to the average thickness of upper layer (H) and to density contrast between deeper and upper layers ($\Delta\rho_{uic}$): $B \propto H\Delta\rho_{uic}$.

ΔS_{uic} was substituted for $\Delta\rho_{uic}$, and X for B. ΔS_{uic} is salinity difference between deeper and upper layers (i.e., intermediate and surface water), and X is excess of evaporation over freshwater input.

Ignoring of temperature effects introduces estimated error of only $\pm 5\%$.

Mass and salt budgets in basin are maintained by exchange through Strait of Sicily. Variation in exchange related to sea level variation expressed by parameter Φ (constrained using model of Bryden and Kinder, 1991). Φ responds quasi-linearly by about 4.4×10^{-3} per meter sea level change (at present, $\Phi = 1$, and Φ decreases with sea level lowering).

Parameter γ describes past excess evaporation relative to the present value $\gamma = (X/X^{p})$.

S_{u} is salinity of upper layer inflow through Strait of Sicily.

Then (simplified after Rohling and Bryden, 1994):

$$\Delta S_{uic} = \frac{S_{u}}{23.5\dfrac{\Phi}{\sqrt{\gamma}} - 1}$$

and

$$\gamma = \left(11.75\Phi - \frac{1}{2}\sqrt{(23.5\Phi)^{2} - 0.016HS_{u}}\right)^{2}$$

Fig. 5. Summary of the Rohling and Bryden (1994) model for the influence of sea-level change and variations in the freshwater budget on the depth of MIW in the eastern Mediterranean.

it would overflow the deep water masses in the Ionian and Levantine Basins formed in the Adriatic before the salinity lowering. In other words, a reduction of S_{MIW} would result in the formation of EMDW with a density lower than that of the previously formed EMDW. Such a situation would be conducive of poor deep water ventilation. The EMDW formed prior to the reduction of salinity will in the following be referred to as Old Deep Water (ODW; Fig. 6). Only during severe winters or distinct dry spells might newly formed EMDW acquire sufficiently high densities to mix with the ODW mass. Moreover, the highest density bottom waters in the deepest parts of the ODW mass would be the least likely to be replaced by newly formed EMDW. In this scenario, the newly formed deep water would settle above, or

within the upper parts of the ODW mass, so that it would be more properly referred to as Eastern Mediterranean deep Intermediate Water (EMdIW; Fig. 6). This might explain the reported increase in preserved organic carbon contents within individual sapropels, with increasing depth of deposition (Anastasakis and Stanley, 1984; Anastasakis, 1988; Murat, 1991), which could, however, also be explained in terms of diffusive oxygen supply from above to the deep sea that would be rather insufficient to compensate for the total demand, but still sufficient to show an influence on the preserved organic carbon contents (Béthoux, 1992).

Shielded from deep convective overturn, the deep layers (consisting of ODW) would only loose their characteristics gradually, through vertical eddy (turbulent) diffusion. Eddy diffusion coeffi-
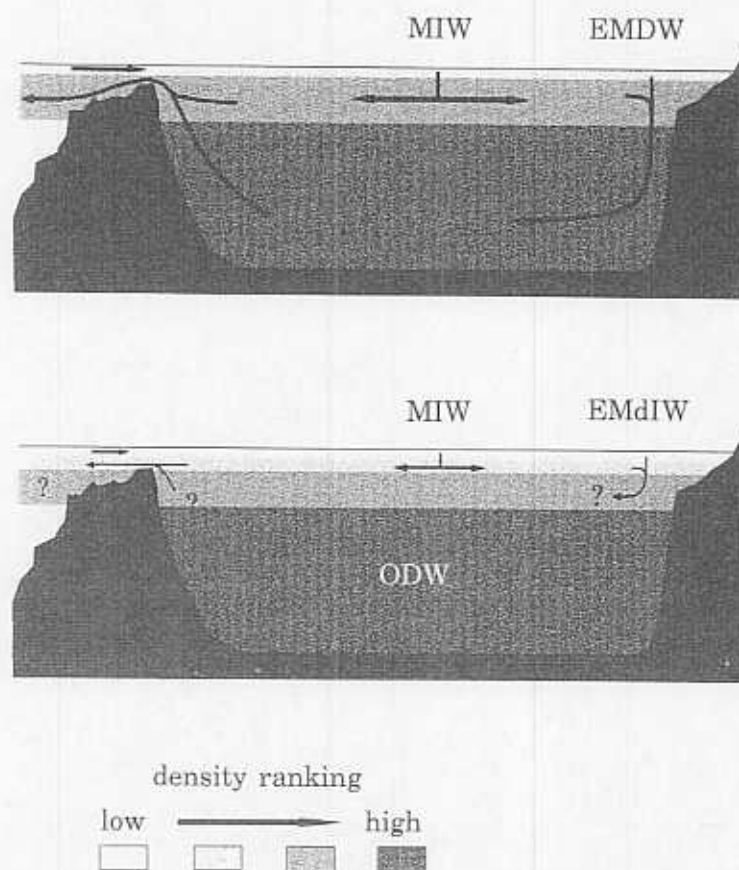

Fig. 6. Schematic presentation of the vertical circulation in the eastern Mediterranean at present (upper) and at times of sapropel formation (lower). Concept underlying the reconstruction is explained in the text, as are the abbreviations.

cients in the present-day Mediterranean range between 1 and 3 $cm^2 s^{-1}$ (Mangini and Schlosser, 1986). Using 2 $cm^2 s^{-1}$ for the diffusion coefficient (D), and 3000 m as a typical vertical scale for the eastern Mediterranean (H), the time needed for diffusion to erase the difference between characteristics of the deep and intermediate waters by a factor e (to $100 \times 1/e \approx 37\%$ of the original) may be calculated at roughly 1500 years ($t = H^2/D$; Mangini and Schlosser, 1986), provided that the intermediate water characteristics remained constant after the first disturbance. When the stability of the water column would have decreased sufficiently to permit deep convection again, bottom water oxygenation would improve rapidly, perhaps even on time scales in the order of 50 to 80 years (the present-day flushing time).

2.2. The past 18,000 years and sapropel S_1

The following reconstruction of the eastern Mediterranean stratification history of the past 18,000 years focuses on the general trends. The relatively short-term fluctuation associated with the Younger Dryas event can only be roughly estimated, because the different parameters needed in the model are insufficiently constrained.

Using the equation for ΔS_{sic}, the outflow–inflow salinity difference (Fig. 5), the salinity of intermediate water (S_{MIW}) is the simply the sum of upper layer salinity (S_u) and ΔS_{sic} ($S_{MIW} = S_u + \Delta S_{sic}$). The non-dimensional coefficient Φ needed for calculation of ΔS_{sic} is described in Fig. 5, as is γ, which describes changes in the excess of evaporation over freshwater input. These coefficients indicate variability relative to the present, and, therefore, their present-day value is 1. To find a value for S_{MIW}, S_u and γ need to be determined, in addition to the parameter Φ, which Rohling and Bryden calculated using the sea-level curve for the past 18,000 years described by Fairbanks (1989). The sea-level curve, and the corresponding values of Φ, are plotted in Fig. 7, and the values at the "key time-slices" used in this study are listed in Table 1. Estimates for S_u in different time-slices can be derived from oxygen isotope studies. These values, and the references from which they are derived, are summarized in Table 1.

The values used for the parameter γ (Table 1) need some explanation. Rohling and Bryden (1994) calculated that γ_{9600} (the value of γ at 9600 yr B.P.) ≈ 1.2, which, as explained in the previous section, should be changed to $\gamma_{9200} \approx 1.05$. After the arid Last Glacial Maximum (LGM; 18,000 yr B.P.), an overall increase in humidity seems to have started in the circum-Mediterranean countries, although this trend may be quite complicated on local scale (Adamson et al., 1980; Goodfriend and Magaritz, 1988; Pons and Reille, 1988; Rossignol-Strick et al., 1992; review in Hastenrath,
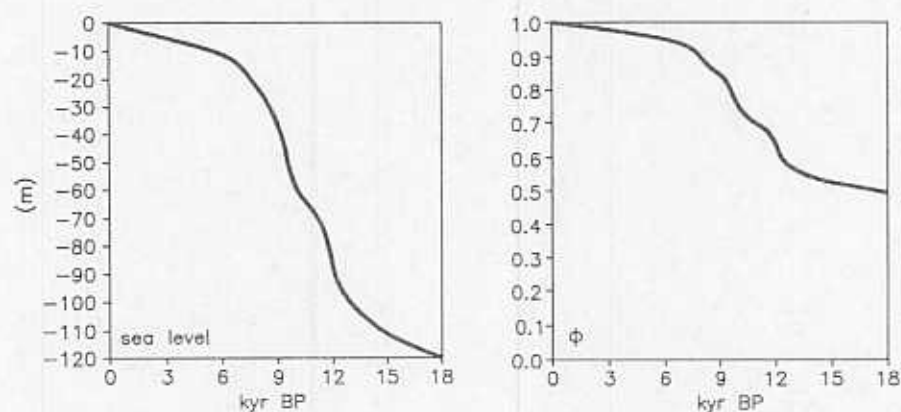

Fig. 7. The change of parameter Φ (measure of the volume of inflow into eastern Mediterranean relative to the present) as related to the change in global sea level since the Last Glacial Maximum determined by Fairbanks (1989) (after Rohling and Bryden, 1994). Note that both curves are plotted versus $^{14}C_c$ years [corrected for a reservoir effect of 400 years (Fairbanks, 1989)].

Table 1

Values used for the different parameters in solving eq. 2. References from which S_u was derived: (1) Thunell et al. (1987); (2) Rohling and Bryden (1994); (3) this study, and equal to present-day values; and (4) Wüst (1961). Values of Φ as determined by Rohling and Bryden (1994), see also Fig. 7. Values of γ as discussed in this study (cf. Fig. 5). Also listed: the values calculated for S_{MIW}, using eq. 1; the values for ΔS_{ric} ($=S_{MIW}-S_u$); and the values for S_{EMDW}. The calculated values of S_{EMDW} are minimum values, representing the salinity required to achieve an EMDW density equal or greater than MIW density, assuming that the temperature difference between the two watermasses (EMDW–MIW) remained about 1°C, as it is at present (POEM group, 1992)

age	age (BP) corr. ^{14}C	Φ	S_u (‰)	γ	ΔS_{ric} (‰)	S_{MIW} (‰)	S_{EMDW} (‰)
LGM	18,000	0.49	40.5[1]	1.25	4.4	44.9	44.6
9600 BP	9200	0.81	38.4[2]	1.05	2.2	40.6	40.3
8300 BP	7900	0.90	36.0[1]	0.85 -1.05	1.6 - 1.8	37.6 - 37.8	37.3 - 37.5
6300 BP	5900	0.94	36.0[1]	0.7	1.4	37.4	37.2
4000 BP	3600	1.00	37.2[3]	1.0	1.7	38.9	38.6
Present	0	1.00	37.2[4]	1.0	1.7	38.9	38.6

1985). Cheddadi et al. (1991) argued that humidity did not increase, but that a peak of aridity occurred at Younger Dryas times, preceded by a short, somewhat more humid event between 13,000 and 11,000 yr B.P. After the Younger Dryas, most records agree that maximum humidity coincided with the deposition of S_1 (8300–6300 yr B.P. $^{14}C_{nc}$, i.e., 7900–5900 yr B.P. $^{14}C_c$). Although the vegetational information still appears to be rather inconclusive, $\gamma_{18,000}$ seems to have been at least equal and presumably larger than γ_{9200}. Since no quantitative estimate of $\gamma_{18,000}$ could be made (yet), I use a hypothetical value of $\gamma_{18,000}=1.25$. Future research may prove this tentative estimate to be too high, since LGM surface temperatures along the northern edge of the eastern Mediterranean may have been lowered by as much as 7°C in winter and 3°C in summer (Wijmstra et al., 1990). A probably comparable lowering of sea surface temperature (SST) in the northern parts of the basin would have caused a decrease in evaporation, which would suggest that $\gamma_{18,000}$ may have been quite similar to γ_{9200} ($=1.05$). In that case, ΔS_{sic} would be about 4‰, and S_{MIW} about 44.5‰,

instead of 4.4‰ and 44.9‰, respectively (Table 1). Apparently, even major adjustment of $\gamma_{18,000}$ would not be of much consequence for the argument presented in this paper, besides causing a minor adjustment in the amplitude of the variations between 18,000 and 9200 yr B.P. in Fig. 8a and b.

When the deposition of S_1 started at about 8300 yr B.P. $^{14}C_{nc}$ (7900 yr B.P. $^{14}C_c$), the reported progressive increase in humidity since the LGM, or more specifically since the Younger Dryas, would imply that $\gamma_{7900}<\gamma_{9200}$, i.e., $\gamma_{7900}<1.05$. Furthermore, no indications of a distinct DCM are found at the base of S_1. The pycnocline and associated nutricline probably resided below the base of the euphotic layer. Rohling and Bryden (1994) assumed that the depth of base of the euphotic layer remained constant through time at its present level (about 120 m). In that case, the absence of a DCM at the onset of S_1 formation suggests that $H_{7900}>120$ m. Using that value, with $\Phi_{7900}=0.90$, and $S_{u7900}=36$‰, it is found that $\gamma_{7900}>0.85$ (Fig. 5). Therefore, $0.85<\gamma_{7900}<1.05$ (Table 1).

In the very top of S_1, dated about 6300 yr B.P.

$^{14}C_{nc}$ (5900 yr B.P. $^{14}C_c$), *Neogloboquadrina* shows a brief recurrence that probably reflects the development of a distinct DCM (Jorissen et al., 1993; Rohling et al., 1993a), which would suggest that $H_{5900} \approx 120$ m. Using that value, with $\Phi_{5900} = 0.94$, and $S_{u5900} = 36‰$, it is found that $\gamma_{5900} \approx 0.7$ (Table 1).

Around 4000 yr B.P. $^{14}C_{nc}$ (3600 yr B.P. $^{14}C_c$), modern conditions were achieved in the Mediterranean, suggesting $\gamma_{3600} \approx 1.0$, as it is at present (Table 1).

The rate of change in surface water salinity of the North Atlantic Ocean resulting from the addition of meltwater, is approximated according to the rate of sea-level change during the past 18,000 years, as portrayed in Fig. 7 after Fairbanks (1989). Between 18,000 and 13,000 yr B.P., a gradually increasing trend of sea-level rise is suggested, with a rapid acceleration to a 1000 year period of maximum rise centered around 12,000–12,500 yr B.P. [Meltwater peak 1A; Fairbanks (1989)]. Decelerated sea-level rise is found between 11,500 and 11,000 yr B.P., followed by a second acceleration centered at about 9500 yr B.P. [Meltwater peak 1B; Fairbanks (1989)]. As explained before in this section, the influence of decelerated sea-level rise associated with the Younger Dryas cooling event can only be tentatively estimated in the presented schematic reconstruction. Distinct deceleration of sea-level rise started at about 7000 yr B.P., with near modern conditions being reached at about 4000–3500 yr B.P.

The influence of deglaciation (sea-level rise + global reduction of surface water salinity), combined with that of the above described local Mediterranean humidity history, would have induced a trend in MIW salinity towards decreasing values, in a gradually decreasing trend between 18,000 and about 13,000 yr B.P., after which time S_{MIW} started a very rapid decrease, which was more or less interrupted at Younger Dryas times. The eastern Mediterranean surface water salinity values based on oxygen isotope studies (Table 1) suggest a rapid decrease in the rate of change in MIW salinity at about 8200 yr B.P. Minimum values were reached during deposition of S_1, until about 5900 yr B.P. After that time, a rapid increase followed, until near modern conditions were

reached at about 4000–3500 yr B.P. The MIW salinity curve in Fig. 8a schematically represents the discussed trends. The most obvious effect of humidity changes in Fig. 8a is the excessive downward bulge between 12,000 and 4000 yr B.P. This bulge would not appear if only deglaciation effects were considered, in which case the curve would look more or less like a smoothed version of the sea-level curve in Fig. 7.

Together with the above described line for S_{MIW}, the estimate of 1500 years needed for diffusion (see Section 2.1) allows for a rather qualitative, but illustrative, graphic presentation of the hydrographic changes culminating in the formation of S_1. As a first step, a line for S_{EMDW} is drawn in relation to that for S_{MIW}, as if S_{EMDW} would always adjust immediately to variations in S_{MIW} ("EMDW direct" in Fig. 6a). The minimum salinity of EMDW (listed in Table 1) is derived from the salinity of MIW by the requirement that the resultant density of EMDW should be equal to, or greater than, that of MIW (assuming that the temperature of EMDW was about 1°C lower than that of MIW, as it is at present (POEM group, 1992).

As mentioned before, stable stratification tends to develop when S_{MIW} decreases, as a result of which diffusion would be the only practical means of change in EMDW characteristics; the process of diffusion gives an estimated time-lag of roughly 1500 years (see Section 2.1). Therefore, the S_{EMDW} line is shifted forward by 1500 years, for those parts where S_{MIW} decreases ("EMDW lag 1500 y" in Fig. 8a). Here, it should be mentioned that this 1500 year time-lag is valid only for conservative properties (such as salinity), and that the lag will be longer or shorter for substances that are produced/released or consumed internally within the watermass, respectively. When S_{MIW} increases or remains constant with time, deep convective mixing is favoured, and the basin tends to ventilate on time scales as short as 50 to 80 years (the present-day turnover time). In those intervals, the EMDW characteristics will change in rapid response to changes in S_{MIW}, and the S_{EMDW} line to be followed in Fig. 8a is that named "EMDW direct". Fig. 8b shows the resultant plot of S_{MIW} and S_{EMDW} during the past 18,000 years.
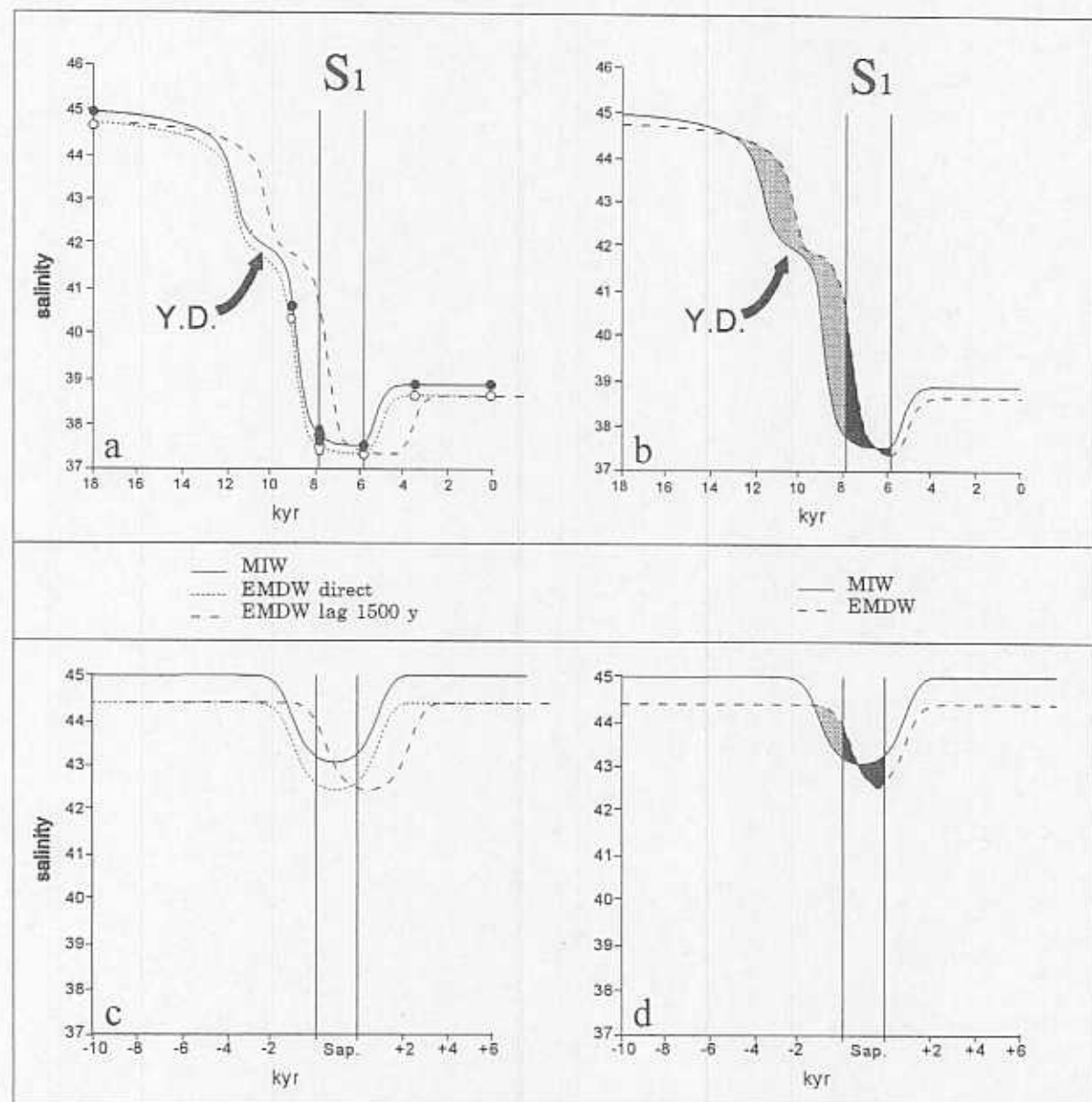
Fig. 8. (a). Schematical representation of changes in the salinity of MIW (solid line) since the Last Glacial Maximum, as constrained by the black circles which represent calculated values listed in Table 1 and explained in the text. The dotted line corresponds to the minimum salinity values for EMDW determined in relation to the MIW salinities (method explained in caption Table 1). The dashed line indicates the EMDW salinity line with a lag of 1500 years, determined on the basis of turbulent diffusion over a typical vertical scale of 3000 m (see text), and therefore provides a measure for the change of EMDW salinity with time in the absence of deep convection. (b) The resultant of Fig. 8a, with MIW and EMDW salinity since 18,000 yr B.P. EMDW salinity is taken according to the 1500 year lagged line of Fig. 8a in intervals where there is a decreasing trend with time in MIW salinity, and according to the "EMDW direct" line of Fig. 8a in intervals where MIW salinity is stable or increasing (see further explanation in the text). Hatching indicates the interval where poor deep water ventilation would be likely, and darker hatching is used in the interval covered by S₁. As explained in the text, the closing of the two lines in the Younger Dryas (Y.D.) interval would be

2.3. Sapropel formation independent of deglaciation

In the above scenario, changes in S_{MIW} resulted from a combination of deglaciation and variation of humidity. In the following scenario, only changes in humidity will be used, by focusing on sapropel formation at full glacial times, without substantial changes in sea level (such as, for instance, S_6). For this hypothetical scenario, portrayed in Fig. 8c, I use full glacial values of S_u, S_{MIW}, and S_{EMDW} that are comparable in magnitude to values of the LGM. Then, a disturbance of those values, which would have resulted from humidity changes, is positioned symmetrically over the sapropel. The magnitude of that disturbance is chosen in agreement with the humidity-imposed disturbance associated with S_1. The duration of formation of the hypothetical full glacial sapropel is taken at 2000 years, similar to that of S_1. Subsequently, a line for S_{EMDW} ("EMDW direct") is drawn in relation to S_{MIW}, in the same way as discussed above. Then, a 1500 year lagged line for S_{EMDW} ("EMDW lag 1500 y") is plotted.

As discussed above, the line "EMDW lag 1500 y" is followed in the intervals where S_{MIW} decreases, whereas the line "EMDW direct" is followed where S_{MIW} is constant or increasing. Fig. 8d shows the resultant plot of S_{MIW} and S_{EMDW} across the hypothetical full glacial sapropel.

3. Discussion

First, I emphasize that all datings mentioned in this discussion are AMS ^{14}C datings, corrected for a reservoir age of 400 years, or interpolated values thereof. The original, not corrected, datings and the references from which these were obtained have been mentioned extensively in previous sections.

The above described graphic presentation (Fig. 8a and b) demonstrates how the combined influences of deglaciation (causing sea-level rise + global reduction of surface water salinity) and astronomically induced reductions in eastern Mediterranean excess evaporation, cause the development of a poorly ventilated deep water mass. Since a more or less distinct pycnocline/nutricline was persistently present in the basin (which requires advection), it may be concluded that intermediate water formation remained active during sapropel formation. Also, some process similar to the modern deep water formation may have existed. However, the shallow Upper Depth Limit of anoxic sediments (< 500 m) suggests that such a two or three-layer circulation in the basin covered only a narrow depth-range. Therefore, a possible water mass settling below the intermediate water of Levantine origin (MIW) should be referred to as Eastern Mediterranean deep Intermediate Water (EMdIW), rather than as Eastern Mediterranean Deep Water (Fig. 6). Still deeper, poorly ventilated ODW (= deep water formed before the reduction of surface water salinities) would gradually become dysoxic, to anoxic, through mineralization of organic matter sinking from the surface layers. The ODW characteristics would be gradually modified by eddy diffusion and possibly occasional influxes of "extra dense" EMdIW resulting from sustained periods of relatively cold and/or dry winters.

In Fig. 8a and b, a threshold value is reached between about 12,500 and 12,000 yr B.P., the salinity of MIW becoming lower than that of EMDW. After about that time, newly formed deep water would not have been capable of displacing that formed previously, thus causing the separation of an ODW mass. As a result, poor ventilation, and consequently rapid oxygen depletion, would have characterized the ODW mass. With an average oxygen concentration below 1500 m depth of 180 mol m^{-3}, and an average deep water oxygen utilisation rate in the eastern Mediterranean of 0.5

conducive to possible convection during periods of sustained winter cooling. The time scales of Fig. 8a and b are in [14]C$_c$ years, corrected for a reservoir age of 400 years. (c) Schematical representation of salinity changes in MIW and EMDW across a hypothetical glacial sapropel formed exclusively by a change in the freshwater budget that is assumed fully symmetrical in time, in the absence of sea level change. Details explained in the caption of (a) and in the text. (d) The resultant of Fig. 8c. Details explained in the caption of (b) and in the text.

mol m^{-3} yr^{-1} (Schlitzer et al., 1991), the oxygen below 1500 m depth would be removed after only about 360 years, provided that oxygen supply had ceased entirely. The latter assumption is not correct, because of a supply of oxygen by eddy diffusion, and probably also occasional convective events. Moreover, anoxia were found as shallow as 500 m (or less), rather than being restricted to depths below 1500 m, and convective events would likely penetrate these upper reaches more frequently than the greater depths. As a result, the roughly calculated 360 years required to deplete all oxygen should be viewed as a minimum estimate, providing information only about the order of magnitude of that period.

Assuming for the moment that ODW prevailed below 500 m waterdepth, then diffusive oxygen input (M) into those waters (i.e., into a column of about $3000 - 500 = 2500$ m height), with an eddy diffusion coefficient of 2 cm^2 s^{-1}, and a surface water oxygen concentration of 220 mol m^{-3}, amounts to about 555 mol O_2 m^{-2} yr^{-1} ($M = D \times C \times t/L$; where $D = 2$ cm^2 s^{-1}, $C = 220$ mol m^{-3}, $t =$ seconds in a year, and $L = 2500$ m). This quantity contrasts sharply with the advective supply of almost 4000 mol O_2 m^{-2} yr^{-1} below 1500 m waterdepth effectuated by modern EMDW formation in the Adriatic (cf. Schlitzer et al., 1991). The supply of 555 mol O_2 m^{-2} s^{-1} into a water column with a height of 2500 m accounts for a supply of $555/2500 = 0.22$ mol O_2 m^{-3} yr^{-1}. This is less than half the average oxygen utilisation rate characteristic of the present-day deep eastern Mediterranean, which is about 0.5 mol O_2 m^3 yr^{-1} (Schlitzer et al., 1991). It is important here, to emphasize that this oxygen utilisation rate is a net average based on the total oxygen supply to deep water and the resultant deep water oxygen concentration. However, even if the diffusive supply were not already accounted for in the Schlitzer et al. (1991) value, the diffusive supply of 0.22 mol O_2 m^{-3} yr^{-1} to the ODW mass would be much lower than the utilisation rate of 0.5 mol O_2 m^{-3} yr^{-1}. In that case, all oxygen below 500 m would be consumed by mineralisation of organic matter within $180/(0.5 - 0.22) = 640$ years.

There is yet another complication. It may not be ruled out that dysoxic conditions could be maintained in large portions of the ODW mass, while true anoxia developed only in a "blanket" near the sediment–water interface. Such a blanket would result from mineralization of organic matter that rapidly sank to the sea floor (e.g. pellets). Note, in this respect, that the oxygen utilisation rate of Schlitzer et al. (1991) represents an average value, and that a rather substantial part of the actual utilisation may take place at the sediment–water interface. Cramp et al. (1988) suggested that such a blanket of anoxia prevailed during the deposition of S_1 in the western Aegean Sea, while Rohling et al. (1993b) found indications for such a configuration during the deposition of Upper Pliocene sapropel C2 in the Singa section (southern Italy, northwest Ionian Basin).

Occasional spells of drought and/or severe winters would result in the formation of EMdIW with relatively increased densities, which might accordingly penetrate and supply oxygen into the ODW mass. Because of an increase of density with depth within the ODW mass, such short-term flushing would likely be more frequent at shallower levels, resulting in the observed increase of preserved organic carbon contents with increasing depth. Under favorable conditions, such as high sedimentation rate and little bioturbation, the more severe of these short-term flushing events may be detectable as short-term interruptions of the sapropel by intervals of sediments deposited under (dys-)oxic conditions, which may be recognized in benthic faunas (e.g., Jorissen et al., 1993; Rohling et al., 1993b, submitted). Sapropels S_6 and S_8 commonly contain intercalations of grey muds (e.g. Cita and Grignani, 1982; Thunell et al., 1983), which might also be reflections of temporarily improved oxygenation, but have not yet been investigated as such. Note that in the context of a reversed circulation scenario for the formation of sapropels, such short-term flushing would be very hard to explain.

If the above described rather severe (i.e. detectable) flushing events indeed took place, then it seems likely that less severe and/or persistent (i.e. hardly or not detectable) flushing events existed as well. In that case, the period of time in which the ODW mass would become oxygen-depleted would be lengthened by occasional advection. However,

even if such ventilation events managed to double or triple that period (minimum duration ≈ 360 years; see above), severe environmental stress should show in fossil benthic communities within about 1000 years after high stability developed in the water column. In the case portrayed by Fig. 8a and b, therefore, one would expect benthic fauna to reflect low bottom water oxygen concentrations around 11,000 yr B.P. In reality, however, benthic faunas reflecting increasingly stressed conditions in the well dated Adriatic core IN68-9 (Jorissen et al., 1993) appear only about 1000 years before the formation of S_1, at about 8900 yr B.P. Influence of severe dysoxia is found only 200–300 years before the onset of S_1 (Rohling et al., submitted). In other words, the model predicts low-oxygen stress around 11,000 yr B.P., whereas the record shows the onset of such conditions around 8900 yr B.P. This lag may well be related to the Younger Dryas, as indicated in Fig. 8a and b. The progressive deterioration of deep water ventilation and related lowering of the deep water oxygen content, starting around 12,000 yr B.P., could have been "reset" by the Younger Dryas, after which a new trend developed starting around 10,000 yr B.P. This new trend resulted in detectable stress on the benthic community after some 1000 years (namely around 8900 yr B.P.), ultimately causing severe oxygen depletion around 8100–8200 yr B.P. and true anoxia marking the onset of the formation of S_1 at 7900 yr B.P.

Another major point of interest concerns the roles of increased production and decreased oxygen advection, relative to one another. As mentioned before, a Deep Chlorophyll Maximum (DCM) was apparently present during the deposition of most sapropels, but some (like, for instance, S_1) hardly contain evidence of a well-developed DCM. This demonstrates that the increased export production associated with a DCM is not a basic prerequisite for the formation of sapropels. It would imply that sapropels could develop as a result of severe reduction of deep oxygen advection alone, with export production rates equal to those of today, or only slightly raised. However, in the absence of substantial increases in export production, only sapropels with relatively low organic carbon contents may develop

(Calvert, 1983). Higher organic carbon contents in eastern Mediterranean sapropels should, therefore, be considered to reflect a significant increase in export production, in addition to the effects of severely reduced deep oxygen advection. Organic carbon contents of well over 10% have been reported (among others, Calvert, 1983; Murat, 1991). Of course, one should compare the organic carbon contents of the various sapropels only in settings with approximately similar sedimentation rates. Such a comparison in one single core, MD-84641 (Fontugne and Calvert, 1992), showed that S_5 contains the highest organic carbon content, and S_1 the lowest, among the series of sapropels $S_{1,3-10}$ (S_2 is absent, as it nearly always is). In the scenario for sapropel formation described in the present paper, sapropels with high organic carbon contents should ideally contain abundant *Neogloboquadrina*, reflecting the presence of a distinct DCM with consequently increased export production. Also the nannofossil assemblage should reflect these conditions (Castradori, 1993).

The crucial role of reduced deep oxygen advection for the process of sapropel formation is perhaps best illustrated by the full glacial intervals, when a DCM seems to have been present without disturbance of the normal oxygenated environments in the deep sea. This may be explained in terms of the probable high density (high salinity and relatively low temperature) of EMDW formed at glacial times, promoting ventilation down to the greatest depths. The scenario for eastern Mediterranean sapropel formation outlined in the present paper is strongly supportive of that proposed by Mangini and Schlosser (1986).

Emeis et al. (1991) performed carbon–oxygen budget calculations for the Tyrrhenian Basin (western Mediterranean), and concluded that reduced deep water ventilation, with consequently reduced oxygen advection, was the primary factor responsible for the formation of sapropels in that basin as well. The scenario for sapropel formation in the eastern Mediterranean presented here may probably be transposed to provide a basic understanding of the simultaneous formation of sapropels in the western Mediterranean, although true sapropels seem to occur much less frequently in that basin.

and are also less well developed and preserved than contemporary eastern Mediterranean sapropels. The eastern basin may be more likely to develop into the "sapropel-mode" than the western basin because of less effective deep water formation, which for the eastern basin occurs in a specific marginal basin (Adriatic Sea), whereas open-ocean convective processes cause formation of deep water in the western basin. Even at present, flushing is 2.5 to 4 times more rapid in the western basin than in the eastern basin.

4. Conclusions

Review of previous studies on the mechanisms of sapropel formation suggests that an anti-estuarine scenario for sapropel formation would fit the accumulated evidence better than an estuarine ("reversed") scenario. The anti-estuarine circulation would have been to some degree weakened relative to the present, in response to reduction of the excess of evaporation over freshwater input. This reduction seems to have been related to increased intensity of the Indian Ocean summer (SW) monsoon, influencing the eastern Mediterranean via the Nile river, and to increased activity of Mediterranean depressions, as an element of the westerly Atlantic system. Both the monsoon and the westerly Atlantic system would have been intensified in response to the occurrence of distinct minima in the cycle of precession.

In addition to its implications for the magnitude of variations in the freshwater budget, which are relatively moderate and therefore seem realistic, the presented anti-estuarine scenario also accounts for (1) pycnocline depth variations, (2) asymmetric sequences of sapropel deposition, (3) increasing preserved organic carbon contents with increasing depth of deposition, and (4) brief intervals of improved oxygenation within sapropels. Moreover, the anti-estuarine circulation is endorsed by the quoted findings of sapropels in the western Mediterranean, and by isotopic and faunal reconstructions of the exchange patterns through both the Strait of Gibraltar and the Strait of Sicily.

A schematic graphic representation is used to demonstrate that poorly oxygenated deep water conditions, with development of anoxic conditions especially near the sea floor, would have resulted from "isolation" of deep waters from actively ventilating surface and intermediate layers. Consequently diminished oxygen advection down from a few hundred meters in the water column would have favoured preservation of sinking organic matter. This could, however, have resulted only in the formation of sapropels with very low organic carbon contents. The high organic carbon contents observed in various sapropels are argued to reflect superimposed increases of export production.

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