



Water column dynamics during the last interglacial anoxic event in the Mediterranean (sapropel S5)

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[1] Marine organic-rich sediments are important for long-term carbon sequestration, and as a source of fossil hydrocarbons, but the processes underlying their formation remain elusive. We present a multiproxy analysis of a relatively recent, well-preserved Mediterranean organic-rich deposit known as sapropel S5, which formed 124–119 thousand years ago (ka B.P., where B.P. references year 1950). Combining planktonic foraminiferal abundance records with stable isotope and organic biomarker data, we reconcile the apparently contradicting coexistence of deep-dwelling foraminiferal faunas with green sulphur bacteria that indicate photic zone euxinia. S5 started with a rapid freshwater-induced increase of density stratification that inhibited vertical mixing and deepwater ventilation. Through the first 900 years, anoxic to euxinic deepwater conditions developed and extended up to roughly 200 m water depth. Then followed a 4100-year period of variability in the depth/intensity of winter mixing with a roughly 1400-year periodicity, coincident with fluctuations in water column stability. S5 terminated with a marked decrease in density stratification and consequently renewed convective overturn.

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1. Introduction

[2] Sapropels have developed in the Mediterranean (especially in the eastern Mediterranean; i.e., east of the Strait of Sicily) at regular time intervals during anoxic events, at times of orbitally induced insolation maxima [Rossignol-Strick *et al.*, 1982; Rossignol-Strick, 1983, 1985; Rohling, 1994]. These sapropels are organic-rich sediment layers intercalated between organic-poor sediments, and are often readily recognized by their distinctly dark color. The generally accepted numbering scheme for late Pleistocene sapropels counts them downward, starting with early to mid Holocene sapropel S1. Sapropel S5 formed during the insolation maximum of the last interglacial. The anoxic events that gave rise to sapropel formation reflect an indirect impact of African monsoon maxima on the basin's hydrography. This impact concerns the routing of African monsoon precipitation into the Mediterranean via the Nile River [Rossignol-Strick *et al.*, 1982; Rossignol-Strick, 1983, 1985] and other, currently dry, systems along the wider North African margin [Rohling *et al.*, 2002a].

[3] Sapropels S1 and S5 formed during the insolation/monsoon maxima of the current (Holocene) and last interglacial periods, respectively. Both monsoon maxima were interrupted by a dry interlude of several centuries, with concomitant cooling over the north of the basin [Rohling *et*

al., 1997, 2002a, 2002b, 2004; Emeis *et al.*, 2000; Gasse, 2000; Geraga *et al.*, 1999; Scrivner *et al.*, 2004].

[4] The enhanced surface buoyancy loss from the Mediterranean Sea due to the centennial-scale climatic deterioration (enhanced aridity and cooling) within the period of S1 deposition caused significant new deepwater formation, which reinvigorated oxygen advection into deeper levels of the basin. Sapropel deposition became interrupted/weakened, and there was repopulation of the seafloor at previously anoxic sites with oxygen-dependent benthic faunas, down to at least 1500 m depth [Rohling *et al.*, 1997; Mercone *et al.*, 2001; Casford *et al.*, 2003]. Numerical circulation models suggest that such deep-water reventilation may develop in response to even a relatively minor (2°C) climatic deterioration [Myers and Rohling, 2000].

[5] Sapropel S5 is generally more intensely developed than S1, with higher organic carbon accumulation rates than S1 [e.g., Fontugne and Calvert, 1992], stronger freshwater flooding than during S1 [Rohling, 1999; Rohling *et al.*, 2004], and a lack of benthic fossils that indicates persistent anoxia below 300 m water depth [Rohling and Gieskes, 1989; Rohling *et al.*, 1993]. The latter observation agrees with the common preservation of original sedimentary lamination in S5 [e.g., Pearce *et al.*, 1998; Kemp *et al.*, 1999; Corselli *et al.*, 2002], which reflects a lack of bioturbation and is seldom seen in S1. As a remarkable exception, the upper half of S5 contains benthic foraminifera in core GeoTüKL83 from 1433 m water depth on the easternmost (Israeli) margin of the basin (Figure 1) [Schmiedl *et al.*, 2003], suggesting less extreme conditions during S5 deposition in that region than in the rest of the open eastern Mediterranean. This is supported by observations that the onset of seafloor anoxia (benthic azoic

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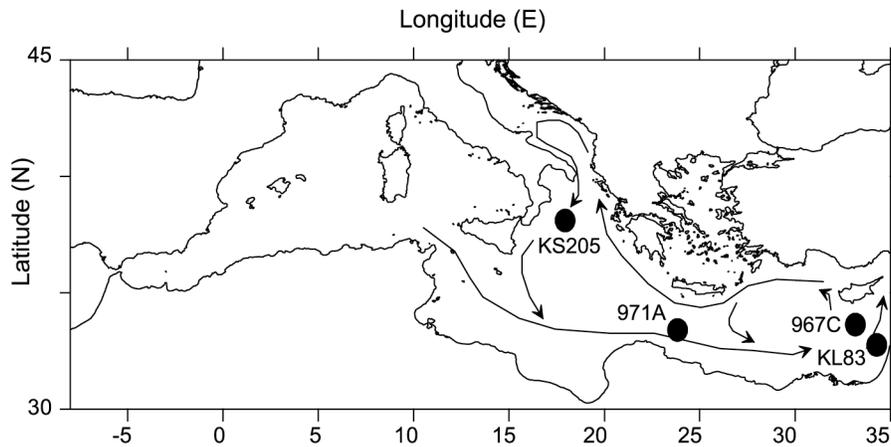


Figure 1. Map of the Mediterranean with schematic direction of surface circulation in the eastern Mediterranean basin and location of Ocean Drilling Program (ODP) Site 971A (data presented here). Also given are the locations of ODP Site 967C, core KS205, and core GeoTüKL83, which have yielded key results for comparison (see text).

conditions) was delayed by several centuries in ODP Site 967C (2554 m depth in the easternmost sector), relative to sites from similar water depths in the rest of the basin [Cane *et al.*, 2002]. Regional differences in the “intensity” of other sapropels have been ascribed to spatial differences in export production and associated deepwater oxygen utilization [Casford *et al.*, 2003].

[6] Contrary to the interruption of the Holocene monsoon maximum within S1, the interruption of the last interglacial monsoon maximum during S5 deposition does not seem to have caused sufficient reoxygenation at depth to allow widespread benthic recolonization [Cane *et al.*, 2002; Rohling *et al.*, 2002a], except at 1433 m depth in the easternmost sector (GeoTüKL83) where this period approximately marks the return of benthic foraminifera [Schmiedl *et al.*, 2003]. An absence of widespread deep-sea reoxygenation does not necessarily exclude the possibility that new deep waters were formed. Much, if not all, of the oxygen entrained within new deep waters into an ambient anoxic/euxinic water mass would be immediately used (chemically “titrated”) for oxidation of reduced compounds, which would leave no bioavailable oxygen (see, for example, the processes affecting Mediterranean inflow into the modern Black Sea [Konovalov *et al.*, 2003]). The displacements within the water column caused by a new deepwater “injection” would trigger enhanced upward advection of subsurface properties (e.g., nutrients), similar to processes observed in the basin today [Klein *et al.*, 1999]. The present paper evaluates whether such processes are manifest within S5.

2. Methods

[7] The present study compares new and existing data for sapropel S5, recovered in section 1H-3 of ODP Site 971A (central eastern Mediterranean; 24°41′N; 33°43′E; 2026 m) (Figure 1).

[8] Isorenieratene concentrations were determined by extraction with acetone and analysis with high-performance liquid chromatography as described elsewhere [Hopmans *et*

al., 2005]. Organic carbon concentrations were determined on decarbonated residues after solvent extraction by elemental analysis.

[9] The planktonic foraminiferal abundance counts are based on speciation and counting of random sample splits containing ~200 specimens in the 150- to 600- μm size fraction [Cane *et al.*, 2002]. The stable isotope analyses were performed with a Europa Geo2020 mass spectrometer following reaction, in individual acid baths, of 3–10 hand-picked and cleaned adult specimens within a 50 μm size window. Isotope ratios are expressed as $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$, in ‰ values relative to Vienna Peedee Belemnite (VPDB), and external precision was better than 0.06 ‰ for both $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ [Cane *et al.*, 2002; Rohling *et al.*, 2002a, 2004]. Previous work has discussed the stable oxygen isotope data alongside century-scale resolution sea surface temperature (SST) records, based on the alkenone unsaturation index, to unravel the nature of changes in surface buoyancy forcing through S5 [Rohling *et al.*, 2002a, 2004].

3. Results

[10] Figure 2 summarizes organic carbon concentrations, and stable oxygen and carbon isotopes for several planktonic foraminiferal species, along with isorenieratene concentrations and planktonic foraminiferal abundance counts through S5 in ODP Site 971A. All data have been obtained for a single set of samples. The black color that visually marks the sapropel in this core (shading) accurately delineates the interval devoid of benthic fossils, with an estimated total duration of 5000 years [Rohling *et al.*, 2002a]. Using this duration estimate, a constant sedimentation rate of 220 y cm^{-1} is assumed to obtain age/duration estimates for the various processes. The extent of S5 in this core coincides closely with the broad (63–40 cm) interval of very light $\delta^{18}\text{O}_{\text{ruber}}$ values, caused by enhanced freshwater influx during the last interglacial monsoon maximum; the interlude with heavier $\delta^{18}\text{O}_{\text{ruber}}$ values between 56 and 52 cm marks an 800–900 year interruption of the monsoon

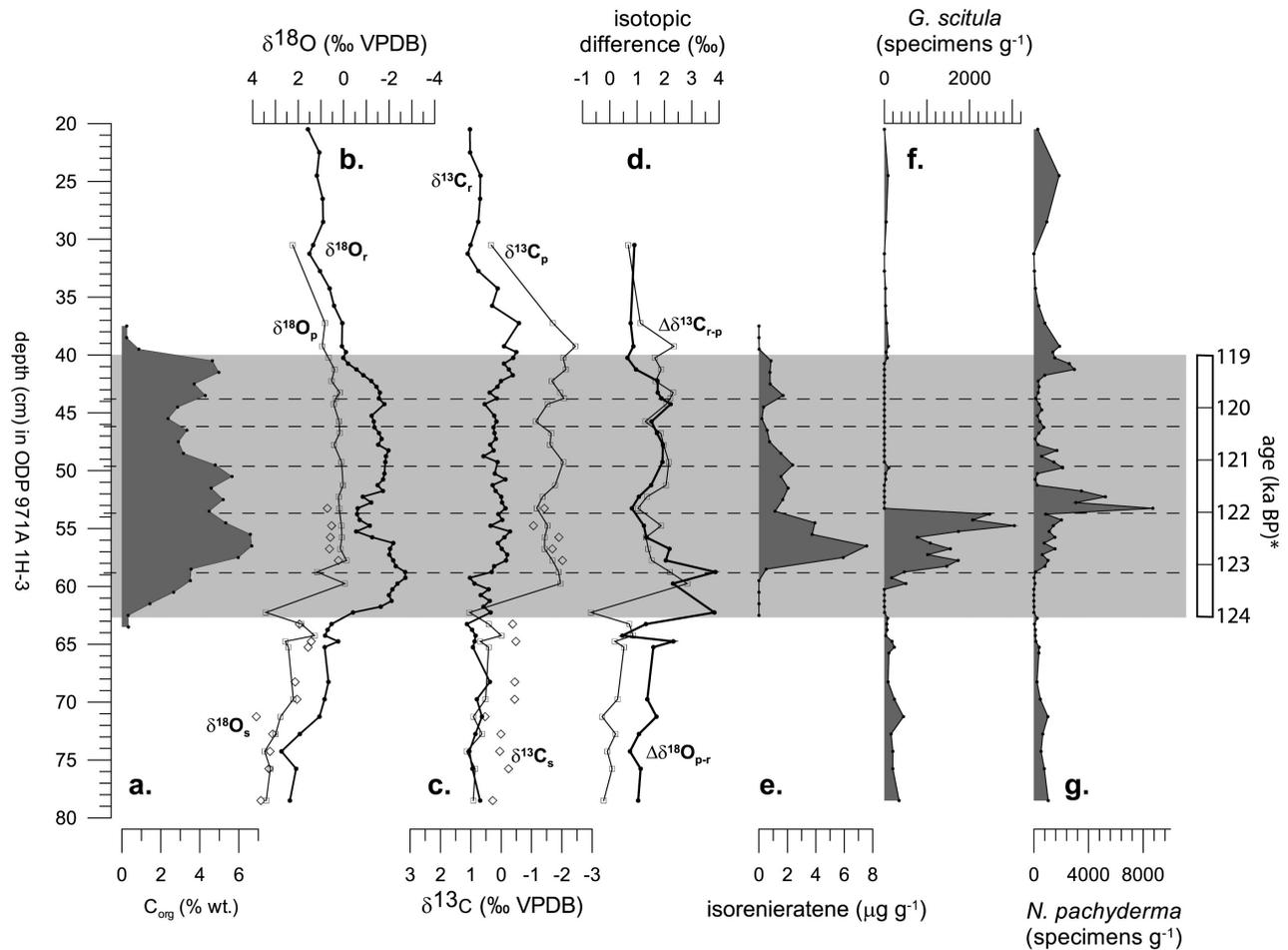


Figure 2. Records through sapropel S5 of ODP Site 971A (section 1H-3) for (a) organic carbon content; (b) planktonic foraminiferal $\delta^{18}\text{O}$ (dots and heavy line indicate *Globigerinoides ruber*, squares and thin line indicate right-coiling *Neogloboquadrina pachyderma*, and diamonds indicate *Globorotalia scitula*); (c) planktonic foraminiferal $\delta^{13}\text{C}$ (same species as in Figure 2b); (d) differences between isotopic values of *N. pachyderma* and *G. ruber*; (e) isorenieratene abundance; (f) number of *G. scitula* per gram dry weight; and (g) number of *N. pachyderma* per gram dry weight. Additional species abundance records are provided by Cane *et al.* [2002]. The age scale provided (at right) is provisional only and relies on a simple linear interpolation between estimated ages of 124 and 119 ka for the onset and end of S5, respectively, as discussed by Rohling *et al.* [2002a].

maximum [Cane *et al.*, 2002; Rohling *et al.*, 2002a, 2004; Scrivner *et al.*, 2004]. Organic carbon concentrations are generally high within S5, after an initial gradual increase to values over 3% within the basal 2.5 cm (about 550 years) (Figure 2a).

4. Depth of the Chemocline

[11] Our results include the first high-resolution record of isorenieratene within S5 (Figure 2e). Isorenieratene is a specific aromatic carotenoid of anaerobic, photolithotrophic green sulphur bacteria (Chlorobiaceae), which require both sulphide and light, albeit at very low intensity. It thus constitutes a proxy for shoaling of the chemocline, which marks the top of euxinic waters, toward the surface where sufficient light is available [Repeta *et al.*, 1989; Koopmans *et al.*, 1996; Passier *et al.*, 1999]. Today, green sulphur

bacteria thrive at the chemocline in the Black Sea, at light intensities as low as 0.0005% of surface irradiance [Overmann *et al.*, 1992].

[12] At present, the 1% irradiance level reaches 125 m depth in the eastern Mediterranean [Berman *et al.*, 1984; Abdel-Moati, 1990] and 80 m in the western basin [Algarra *et al.*, 1988]. On the basis of *e*-folding light attenuation with depth, an irradiance level 3 to 4 orders of magnitude below 1% occurs around 310–375 m in the eastern basin, or 200–240 m in the western basin. Productivity likely was enhanced throughout the eastern Mediterranean at times of sapropel deposition, causing stronger light attenuation than today. However, microfossil (foraminiferal, coccolith, diatom) and stable isotope records indicate that this productivity was focused in a deep chlorophyll maximum (DCM) below an oligotrophic mixed layer, which implies that deep light penetration to depths of roughly 100 m remained

important [cf. Rohling and Gieskes, 1989; Castradori, 1993; Kemp et al., 1999; Struck et al., 2001; Cane et al., 2002; Corselli et al., 2002; Rohling et al., 2004]. The low-light-intensity habitat of the green sulphur bacteria in the eastern Mediterranean at sapropel times may therefore have occurred at greater depths than the ~ 90 m at which it is currently found in the Black Sea.

[13] Other data help to constrain upper bounds to the shoaling of the chemocline. First, winter mixing reaches down to 150 m or more in the Mediterranean. Second, diverse planktonic foraminiferal faunas persisted throughout sapropel S5 on a basin-wide scale [Rohling et al., 1993; Corselli et al., 2002; Cane et al., 2002], which suggests a considerable thickness of habitable, oxygenated water. In the present-day Arabian Sea this is achieved in the 150- to 200-m-thick system above the oxygen minimum zone [cf. Ivanova, 1999; Peeters, 2000]. The presence of *Globorotalia scitula* in much of S5 offers some insight into the available depth habitats. Glacial faunas from the very closely spaced cores SL-11 and SL-21 in the Aegean Sea [Casford et al., 2002, 2003] differ in that *G. scitula* is absent in SL-11 (paleowater depth of 140 m) and continues in SL-21 (200 m), indicating that presence of *G. scitula* requires a water depth in excess of 140 m. Third, the rate at which climatically imposed oxygen isotope anomalies reduce with depth (because of volumetric dilution) in a box model representation of the mixed layer for S5 times suggests that *Neogloboquadrina pachyderma* (right coiling) and especially *G. scitula* lived at 100 m depth or more [Rohling et al., 2004]. Finally, S5 sapropels from the South Cretan slope show a complete absence of benthic foraminifera at and below 600 m (reflecting persistent anoxia), whereas persistence of benthics at 300 m depth [Rohling and Gieskes, 1989; Rohling et al., 1993] suggests regular oxygenation at that depth throughout S5. Benthic foraminiferal presence indicates that oxygen also was regularly available at 1433 m depth on the Israeli margin, during deposition of the latter half of S5 [Schmiedl et al., 2003]. The generally cyclonic circulation may have caused the chemocline to reside considerably deeper near the margins than in the open eastern Mediterranean, analogous to the “domed” configuration of the chemocline in the Black Sea today. In addition, chemocline depth variations might be expected in relation to spatial differences in export productivity during times of sapropel formation [cf. Casford et al., 2003].

[14] Overall, we estimate that the chemocline resided within a depth range of 150–300 m in the open eastern Mediterranean basin during S5 deposition. Much of the range would be due to temporal (seasonal?) and regional variability. A mean value around 200 m would imply that eastern Mediterranean light attenuation during S5 deposition was comparable to values observed today in the western Mediterranean. Some Pliocene sapropels contain organic carbon concentrations up to 30%, with very high abundances of isorenieratene and a near to total absence of planktonic foraminifera [Passier et al., 1999; Nijenhuis and De Lange, 2000], in stark contrast with the more moderate conditions reported here for S5 (which still is one of the most intensely developed sapropels of the late Quaternary). The Pliocene examples may represent extreme cases of

chemocline shoaling in the eastern Mediterranean, resulting in the elimination of planktonic foraminiferal faunas.

5. Surface Buoyancy Loss, Nutrient Advection, and Productivity

[15] The $\delta^{18}\text{O}$ values of the shallow-dwelling planktonic foraminifer *Globigerinoides ruber* (white) and the intermediate-water dwelling species *Neogloboquadrina pachyderma* (right coiling) (Figure 2b) may be used to characterize changes in the surface to intermediate water density gradient. These changes are regulated by buoyancy loss through cooling and/or net evaporation, and dominated by variations in (monsoonal) freshwater dilution of surface waters [Rohling et al., 2002a, 2004; Scrivner et al., 2004]. An increase (decrease) in the density gradient reduces (enhances) the potential for deepwater formation in the basin. Results for *G. scitula* are also shown and generally support the changes in *N. pachyderma*. At the onset and earliest part of S5, there is a particularly rapid decrease in $\delta^{18}\text{O}_{\text{ruber}}$ while $\delta^{18}\text{O}_{\text{pachyderma}}$ lags behind this change (poorly resolved because of absence of *N. pachyderma* and *G. scitula* in this interval). A similar (better resolved) lag has been observed around the onset of Holocene sapropel S1 [Casford et al., 2002]. Moreover, $\delta^{18}\text{O}_{\text{ruber}}$ shows a negative shift of about 3‰ from pre-S5 values to its peak around 59 cm, which exceeds the about 2‰ change in $\delta^{18}\text{O}_{\text{pachyderma}}$ (Figure 2b). Combined, the timing and magnitude differences define a rapid increase in $\Delta\delta^{18}\text{O}$ (Figure 2d), indicating that the onset of S5 coincided with a sharp increase in surface-intermediate water density stratification, which would have inhibited deep mixing. This is corroborated by the abundance variations of *Globorotalia inflata* through S5. This species, which thrives in conditions with extensive vertical mixing over several hundreds of meters, is present before S5, is absent within S5, and sharply reappears at the termination of S5 [Rohling et al., 1993; Corselli et al., 2002; Cane et al., 2002].

[16] Detectable amounts of isorenieratene are first encountered at about 4 cm (about 900 years) after the onset of S5 (Figure 2e). This lag likely reflects the period needed for depletion of the oxygen initially present in deep waters and subsequent extension of euxinic conditions into the zone of sufficient light availability. Even at modern oxygen utilization rates in the relatively warm (13° – 13.5°C) deep eastern Mediterranean [Schlitzer et al., 1991; Klein et al., 1999], all oxygen below 500 m depth would be consumed in about 600 years if deepwater formation failed [Rohling, 1994]. Hence the observation that development of euxinic conditions up to 200 m depth took about 900 years would imply only moderate enhancement of deep oxygen utilization. In other words, export production may have been enhanced during S5 but not to extraordinary levels, which agrees with our inference that light attenuation in the eastern Mediterranean during S5 was enhanced to values within range of present-day western Mediterranean observations.

[17] The vertical proximity of water masses laden with isotopically light respired CO_2 should, through upward advection, distinctly affect the subsurface $\delta^{13}\text{C}_{\text{pachyderma}}$. Very light $\delta^{13}\text{C}_{\text{pachyderma}}$ values are attained at about

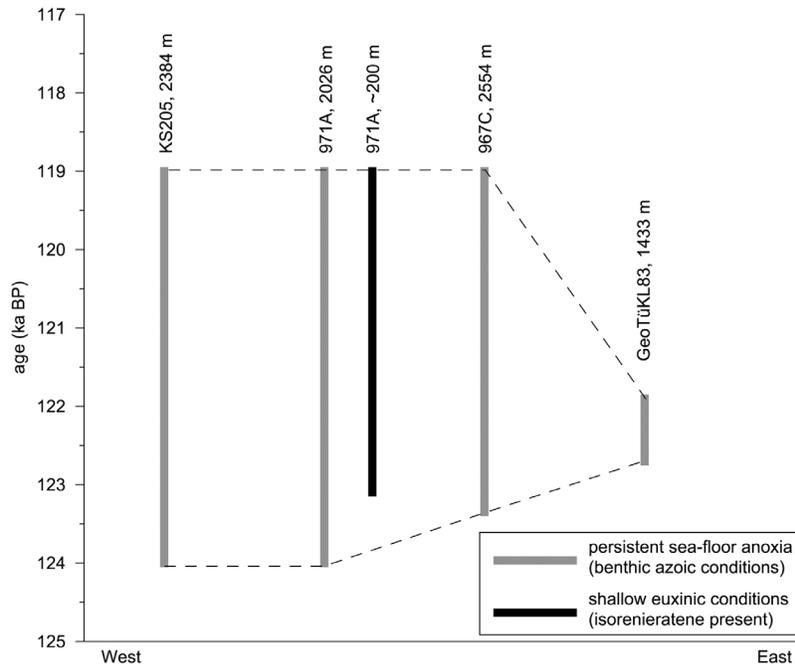


Figure 3. Approximate ages for the S5 onset and termination of persistent seafloor anoxia (benthic azoic conditions) in the various cores discussed here, in comparison with the onset and termination of both seafloor anoxia and shallow euxinic conditions (isorenieratene presence), in ODP Site 971A. Water depths for the observations are presented along with the core names. Age scale is as in Figure 2. Results have been presented along an approximately west-east transect (x axis), but note that spacing is not proportional to real distance between sites.

60 cm, some 2.5 cm (about 550 years) after the onset of S5. Hence advection from a substantial subsurface accumulation of respired CO_2 began to affect carbon isotope values near the base of the photic layer (where *N. pachyderma* thrives, see below) within 550 years. This estimate is remarkably close to the 600 years calculated for consumption of all oxygen below 500 m depth (above). Continuation of the processes of oxic and later anaerobic degradation of sinking organic matter over the next 350 years eventually culminated in the upward expansion of euxinic conditions registered by the isorenieratene record (Figure 2e).

[18] We note that the onset of benthic azoic conditions/seafloor anoxia in core GeoTüKL83 occurs near the top of the first major $\delta^{18}\text{O}_{\text{ruber}}$ minimum (monsoon maximum) in S5 [Schmiedl *et al.*, 2003]. This is roughly the same relative position as that occupied by the first peak of isorenieratene concentrations in S5 of ODP Site 971A (58.5–56 cm, Figure 2). However, it (1) detectably postdates the onset of seafloor anoxia within the lower part of the $\delta^{18}\text{O}_{\text{ruber}}$ minimum in nearby ODP Site 967C from 2554 m depth [Cane *et al.*, 2002; Rohling *et al.*, 2002a]; and (2) seriously lags the onset of seafloor anoxia in ODP Site 971A (2026 m) and core KS205 (2384 m) from the central and western sectors of the basin (Figure 1), where the onset of benthic azoic conditions coincides with the sharp $\delta^{18}\text{O}_{\text{ruber}}$ shift that marks the start of the monsoon maximum [Cane *et al.*, 2002; Rohling *et al.*, 2002a, 2004]. Given the vigorous wind-driven surface circulation, it is not realistic to assume significant diachroneity in the large $\delta^{18}\text{O}_{\text{ruber}}$ changes,

which demonstrates that the onset of seafloor anoxia was diachronous between the various records. Within the 2000–2500 m depth range, seafloor anoxia developed first in the central (971A) and western (KS205) sectors of the eastern Mediterranean, and several centuries later in the easternmost sector (967C) [Cane *et al.*, 2002]. Several centuries later again, seafloor anoxia reached depths around 1400 m in the easternmost sector (GeoTüKL83). This sequence of events (Figure 3) supports our inferred upward expansion and intensification of the anoxic/euxinic conditions during the first ~900 years of S5 development. It also highlights the existence of substantial superimposed spatial differences (possibly related to different rates of export production, [cf. Casford *et al.*, 2003]).

[19] In our records of ODP Site 971A, the earliest part of S5 sees a conspicuous drop to zero abundance of the deep chlorophyll maximum (DCM) indicator species *G. scitula* and *N. pachyderma* [Rohling and Gieskes, 1989; Rohling *et al.*, 2004] (Figures 2f and 2g). The reduced surface buoyancy loss during times of sapropel formation is thought to trigger shoaling of subsurface (intermediate) water toward the base of the photic layer, facilitating “new” nutrient advection from below and so fuelling a well-developed DCM [Rohling and Gieskes, 1989; Castradori, 1993; Rohling 1994; Myers *et al.*, 1998; Kemp *et al.*, 1999; Corselli *et al.*, 2002; Myers, 2002]. The importance of such upward nutrient advection for the presence of *N. pachyderma* and *G. scitula* is illustrated by the fact that the appearance of these species in substantial numbers coincides with the

development of very negative $\delta^{13}\text{C}_{\text{pachyderma}}$ values, at 2.5 cm (about 550 years) after the onset of S5. The appearance of *N. pachyderma* and *G. scitula* in substantial numbers was also found to coincide with the establishment of high diatom abundances within S5, in agreement with the known dietary preference of *N. pachyderma* for diatoms [Corselli et al., 2002]. No dietary insight exists for *G. scitula*, but its abrupt disappearance at 53.5 cm (a Mediterranean-wide feature [Rohling et al., 1993; Cane et al., 2002; Rohling et al., 2002a; Corselli et al. 2002]), while *N. pachyderma* continued in a similar depth habitat (according to isotope data), suggests that *G. scitula* relied on a different food source than *N. pachyderma*.

[20] Sustained presence of DCM indicator species at times with shallow euxinic conditions (i.e., intervals with isorenieratene) suggests that no nitrogen deficiency developed because of denitrification, which ought to inhibit development of a DCM in such a setting. This agrees with observations of a notable increase in atmospheric nitrogen fixation during times of sapropel deposition [Sachs and Repeta, 1999; Struck et al., 2001]. Mat-forming diatom species abundant within S5 are capable of migrating vertically through the photic layer by active buoyancy control [Corselli et al., 2002, and references therein], and are known to host nitrogen-fixing symbionts [Zehr et al., 2000; Wilson, 2003].

6. Temporal Variability Within S5

[21] Between the point when $\delta^{13}\text{C}_{\text{pachyderma}}$ attains very light values (at about 60 cm) and the terminal phase of S5 at about 40 cm, there is a striking apparent agreement (although, statistically, r^2 equals only 0.3 for $N = 19$) between changes in $\Delta\delta^{13}\text{C}$ and $\Delta\delta^{18}\text{O}$, driven mostly by fluctuations in $\delta^{13}\text{C}_{\text{pachyderma}}$ and $\delta^{18}\text{O}_{\text{ruber}}$ (Figure 2d). From 58.5 cm upward, the isorenieratene abundances appear to track this pattern. The isorenieratene maximum at 58.5–56 cm coincides with the light- $\delta^{18}\text{O}_{\text{ruber}}$ phase of maximum freshwater flux into the basin, but has a delayed start relative to $\delta^{18}\text{O}_{\text{ruber}}$ because of the initial 900 years needed for establishment of euxinic conditions. The collapse of the isorenieratene maximum at 56 cm coincides with a sharp termination of light $\delta^{18}\text{O}_{\text{ruber}}$ values. The differences in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ between *G. ruber* and *N. pachyderma* as well as the isorenieratene abundances subsequently reach minima at 53 cm, followed by increases until 49.5 cm, decreases until 46 cm, and another increase until 43 cm. Thereafter, $\delta^{18}\text{O}$ and isorenieratene abundance show another joint decrease until the end of S5 at 40 cm, while $\delta^{13}\text{C}$ follows another pathway.

[22] The described pattern suggests that a quasi-cyclic variability existed during the deposition of S5 with a periodicity of roughly 1400 years (6 to 6.5 cm in our record). It consists of a direct relationship between increases (decreases) in surface buoyancy/density stratification, and increases (decreases) in the flux of isotopically light carbon associated with shoaling (deepening) of the chemocline.

[23] There are two alternative mechanisms for depth changes of the chemocline and concomitant impacts on the upward flux of light carbon. The first and most intuitive

explanation consists of variations in winter mixing and associated oxygenation from above, on timescales within the resolution of our coarsest record (i.e., interannual to centennial). In this scenario, chemocline deepening would be in phase with shifts to heavier $\delta^{18}\text{O}_{\text{ruber}}$ values, and thus smaller $\Delta\delta^{18}\text{O}$ values.

[24] An alternative scenario was suggested earlier. It concerns generally enhanced upward advection in the water column associated with “injection” of new deep water into the interior of the basin, displacing ambient water masses. Similar to the Holocene S1 interruption, the interruption of light $\delta^{18}\text{O}_{\text{ruber}}$ values within S5 (56–52 cm in Figure 2b) has been found to represent a virtually complete collapse of excess monsoon flooding that likely coincided with a phase of cooling in the north [Rohling et al., 2002a, 2004; Scrivner et al., 2004]. Given this sharp increase in surface buoyancy loss, it is possible that the chemocline shoaling at that time reflects new deepwater injection into the basin’s interior, the entrained oxygen being rapidly removed by chemical titration in the ambient euxinic waters. The high initial abundances of reduced compounds within the euxinic water column during the deposition of the intense sapropel S5 might explain why reventilation events could not introduce sufficient bioavailable oxygen to sustain benthic repopulations within that sapropel (in contrast to, for example, S1 and S6 [Casford et al., 2003]). In such a scenario, chemocline shoaling would be most pronounced at times of reducing density contrasts in the basin (periods of highest ventilation potential), and would therefore be out of phase with changes in $\Delta\delta^{18}\text{O}$. In addition, local benthic repopulations would be expected within deepwater source regions. Validation of that expectation would require currently nonexistent high-quality S5 sapropels from the Adriatic and Aegean Seas. Intriguingly, there is one report of minute repopulations within S5 in core M40-4/67SL from 2158 m depth to the ESE of Crete [Schmiedl et al., 2003], which suggests that, very rarely, some oxygen may have reached the deep sea just outside one of Aegean Sea outlets.

[25] Although not 100% conclusive about which of the two possible scenarios for chemocline depth variations applies, our results do offer a strong suggestion. Apart from the offset timing of the first large isorenieratene peak (which we ascribe to delayed development of water column euxinia), the $\Delta\delta^{18}\text{O}$ and $\Delta\delta^{13}\text{C}$ variations appear closely in phase with the isorenieratene fluctuations (Figures 2d and 2e). This favors the explanation that invokes buoyancy-loss-related changes in the depth of winter mixing in the interval from 56 cm to the top of S5.

[26] Reappearance of *Globorotalia inflata* at the termination of S5 [Rohling et al., 1993; Corselli et al., 2002; Cane et al., 2002] signals the onset of deep mixing, in agreement with a reduction in the surface-intermediate water density stratification ($\Delta\delta^{18}\text{O}$, Figure 2d) due to a sharp reduction in monsoon flooding to pre-S5 values [Rohling et al., 2004]. Consequent reestablishment of intense convective overturn “cleaned up” the anoxic/euxinic conditions in the basin, causing enhanced advection of nutrients and isotopically light carbon. In part, this may have been transported out of the basin with intermediate water, through the Strait of Sicily. However, at least part appears to have reached

upward into the photic layer, causing enhanced numbers of *N. pachyderma*, reappearance of *G. scitula* [Rohling et al., 1993; Corselli et al., 2002; Cane et al., 2002], a strong low in $\delta^{13}\text{C}_{\text{pachyderma}}$, and even a lesser low in $\delta^{13}\text{C}_{\text{ruber}}$ (Figure 2c). Convective oxygen supply is an order of magnitude more intense than diffusive oxygen supply into the deep sea (for an overview, see Rohling [1994]), and the reestablishment of convective overturn that terminated sapropel formation caused rapid removal of the anoxic conditions from the water column.

[27] The sequence of events portrayed here on the basis of foraminiferal abundance and stable isotope data and organic biomarker abundances for S5 in ODP Site 971A corroborates the scenario proposed by Struck et al. [2001, Figure 10] on the basis of stable nitrogen and carbon isotopes through S5 in ODP Site 967.

7. Conclusions

[28] Combination of planktonic foraminiferal abundance and stable isotope records with organic biomarker data for eastern Mediterranean sapropel S5 illustrates an apparently contradicting coexistence of deep-dwelling foraminiferal faunas with green sulphur bacteria indicative of photic zone euxinia. However, the faunal indications that at least 150–

200 m of oxygenated surface waters must have existed are easily reconciled with the biomarker data, by accounting for the observed extremely low-light requirement of green sulphur bacteria in the present-day Black Sea (down to 0.0005% of surface irradiance). We thus reconstruct a chemocline depth range of 150–300 m for the time of S5 deposition. A mean depth around 200 m would imply that productivity and light attenuation values in the eastern Mediterranean during S5 times were roughly comparable to values observed today in the western Mediterranean basin. The onset of S5 deposition coincided with a rapid freshwater-induced increase of density stratification, which inhibited vertical mixing and deepwater ventilation. Anoxic to euxinic deepwater conditions then developed, and about 900 years after the onset of S5 these conditions reached up to the reconstructed mean depth of about 200 m. This was followed by a roughly 1400-year variability in the depth/intensity of winter mixing, which coincided with fluctuations in water column stratification/stability. A sharp decrease in density stratification, and consequently renewed convective overturn, occasioned the termination of S5 deposition.

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