Circulation changes and nutrient concentrations in the late Quaternary Aegean Sea: A nonsteady state concept for sapropel formation

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1The modern Aegean Sea is an important source of deep water for the eastern Mediterranean. Its contribution to deep water ventilation is known to fluctuate in response to climatic variation on a decadal timescale. This study uses marine micropaleontological and stable isotope data to investigate longer-term variability during the late glacial and Holocene, in particular that associated with the deposition of the early Holocene dyssoxic/anoxic sapropel S1. Concentrating on the onset of sapropel-forming conditions, we identify the start of “seasonal” stratification and highlight a lag in 18O response of the planktonic foraminifer N. pachyderma to termination T1b as identified in the 18O record of G. ruber. By use of a simple model we determine that this offset cannot be a function of bioturbation effects. The lag is of the order of 1 kyr and suggests that isolation of intermediate/deep water preceded the start of sapropel formation by up to 1.5 kyr. Using this discovery, we propose an explanation for the major unresolved problem in sapropel studies, namely, the source of nutrient supply required for export productivity to reach levels needed for sustained sapropel deposition. We suggest that nutrients had been accumulating in a stagnant basin for 1–1.5 kyr and that these accumulated resources were utilized during the deposition of S1. In addition, we provide a first quantitative estimate of the diffusive (1/e) mixing timescale for the eastern Mediterranean in its “stratified” sapropel mode, which is of the order of 450 years. INDEX TERMS: 1050 Geochemistry: Marine geochemistry (4835, 4850); 3030 Marine Geology and Geophysics: Micropaleontology; 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); 4267 Oceanography: General: Paleoceanography; 4870 Oceanography: Biological and Chemical: Stable isotopes; KEYWORDS: sapropel, Mediterranean, Holocene, foraminifera, climate variability

1. Introduction

2The present-day Aegean Sea (Figure 1) is an important source of deep water for the eastern Mediterranean [Lacombe et al., 1958; Miller, 1963; Roether et al., 1996; Lascaratos et al., 1999]. Aegean Intermediate Water (AEIW) is derived from Levantine Intermediate Water (LIW), with its source in the Rhodes Gyre. As this travels north along the Turkish Coast, prevailing offshore winds allow upwelling of the intermediate water to the surface [Lascaratos, 1989; Yuce, 1995]. In these shallow eastern shelf areas the AEIW consequently forms a single uniform water mass from the surface to the seafloor. As the upwelled AEIW progresses northward, its salinity continues to increase due to evaporation. Winter winds across the Athos Basin (Figure 1) in the far north further enhance the salinity of AEIW, and this together with winter cooling increases its density. This buoyancy loss drives the formation of Aegean Deep Water (AEDW) [Bruce and Charnock, 1965; Burman and Oren, 1970; Theocharts, 1989; Yuce, 1995]. Today, AEDW settles in the deeper parts of the Aegean Basin, below 300 m. Traditionally, AEDW formation was considered of minor importance to the deep water ventilation of the open eastern Mediterranean [Wüst, 1961]. However, recent studies show that specific (cold) climatic forcing over the Aegean has throughout the 1990s caused AEDW to replace Adriatic Deep Water (ADW) as the main deep water in the open eastern Mediterranean [Roether et al., 1996; Samuel et al., 1999]. The Aegean’s rapid response to atmospheric forcing makes it an ideal case study for the analysis of deep water formation and its relationship with climatic change. [3] Dramatic deep water ventilation changes on longer timescales are also witnessed in the sedimentary record of the Mediterranean by the presence of sapropels. These dark organic-rich layers are found throughout the eastern Mediterranean. Sapropel formation is related to slowing down of deep water ventilation in response to climate related reductions in buoyancy loss [Rossignol-Strick et al., 1982; Jenkins and Wiliams, 1983; Rossignol-Strick, 1983, 1985, 1987; Parisi, 1987; Cramp and Collins, 1988; Cramp et al., 1988; Perissoratis and Piper, 1992; Rohling, 1994]. These reductions are thought to be caused by changes to much wetter climatic conditions at times of increased Northern Hemisphere insolation (precession cycle minima) [Rossignol-Strick et al., 1982; Rossignol-Strick, 1983, 1985; Rohling and Hilgen, 1991; Hilgen, 1991; Lourens et al., 1996]. In addition to suppression of deep water production, sapropel formation has also been associated with increases in export productivity [Rohling and Gieskes, 1989; De Lange et al., 1990; Rohling, 1994; Rohling and Hilgen, 1991; Thomson et al., 1995; Cramp and O’Sullivan, 1999].

4Modeling of circulation during deposition of the most recent, Holocene, sapropel (S1) has suggested that relatively small increases in surface buoyancy can lead to suppression of deep water circulation in the eastern Mediterranean. An increase of 20–30% in the freshwater budget is thought to be enough to allow interruption of deep water production [Myers et al., 1998; Rohling and De Rijk, 1999; Rohling, 1999b]. Using the Myers [Myers et al., 1998] circulation model, the biogeochemical implications of

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influxes. Were we present evidence to suggest that the steady state approach is seriously flawed in that a significant period of potential nutrient accumulation in a stagnant basin may have preceded the actual sapropel deposition.

[6] We use abundance variations of planktonic foraminifera together with species-specific oxygen and carbon stable isotope ratios in these forams to derive a picture of the oceanographic processes leading up to, during, and after the most recent sapropel deposition in the Aegean. We observe a conspicuous change in hydrography, starting ~6 kyr prior to S1. We validate our observations with a simple bioturbation model and use the validated records to suggest that long-term (~1.5 kyr) storage of nutrients may have occurred in the Aegean Basin. When this reservoir became available for production, the formation of S1 commenced.

2. Methods and Materials

[7] We present results for two gravity cores from the Northern Aegean Basin (SL-21 and SL-31) and an additional gravity core (SLA-9) together with one piston core (LC-21) from the southern Aegean (Figure 1). All four cores are comprised of microfossil-rich hemipelagic ooze, with a clearly defined darker band of sapropelic material.

[8] Each core was sampled in a continuous sequence: SL-21, SL-31, and SLA-9 were sampled at 0.5 cm intervals, and LC-21 was sampled at 1 cm intervals. The samples were freeze-dried and weighed, and selected (weighed) subsamples were disaggregated and wet sieved using demineralized water. The sieved fractions were collected on 600, 150, 125, and 63 μm mesh sizes. The >150 μm fractions were subdivided using a random splitter to provide an aliquot of ~200 individual planktonic foraminifera. These were then determined and sorted on Chapman slides and counted. Results were obtained as numbers g⁻¹ and as percentages (Figure 2).

[9] Several AMS radiocarbon dates were obtained for cores LC-21, SLA-9, and SL-31 using only hand-picked clean planktonic foraminiferal tests with no evidence of pyritization or overgrowth. The samples were too small for monospecific dating, but no systematic differences would be expected for such dates relative to our results (see Jorissen et al.’s [1993] comparison of planktonic versus benthic dating). The picked material was submitted for analysis at the Natural Environment Research Council (NERC) radiocarbon laboratory at SURRC (LC-21) and at the Leibniz AMS Laboratory at Kiel (Germany) (SLA-9 and SL-31). Radiocarbon convention ages obtained were calibrated using the marine mode of the program Calib 4.2 [Stuiver and Reimer, 1993]. A reservoir age correction of 149 ± 30 years for the Aegean was used [Facorellis et al., 1998]. The results are listed in Table 1. All ages in this paper are reported in calibrated kyr B.P. unless otherwise stated.

[10] Detailed stable oxygen and carbon isotope records have been constructed for individual planktonic foraminiferal species in cores LC-21, SLA-9, SL-21, and SL-31, with resolutions on the order of 1 cm (Figure 3). The species chosen were the very shallow, surface-dwelling Globigerinoides ruber; the deeper-dwelling Globorotalia inflata associated with deep (winter) mixing; and the deep-living species Neogloboquadrina pachyderma, which has been associated with the deep chlorophyll maximum at the base of the euphotic layer. This selection follows global and specific Mediterranean habitat descriptions by Hemleben et al. [1989], Pujol and Vergnaud-Graziini [1995], Rohling et al. [1993a, 1995, 1997], De Rijk et al. [1999], and Hayes et al. [1999]. The analyses were performed at two separate intercalibrated facilities: the Europa Geo 20–20, with individual acid bath preparation, at the Southampton Oceanography Centre (SOC), and the VG-Optima with a common acid bath preparation at NERC Isotope Geoscience Laboratory (NIGL), Keyworth.
Figure 2. Relative and absolute abundances of planktonic foraminifera in cores SLA-9, SL-21, SL-31, and LC-21. The vertical shaded box indicates the extent of the benthic sapropel, identified from the presence of low-oxygen-tolerant benthic foraminifera and/or total absence of benthic fauna that is equated with truly anoxic bottom waters [Fontanier, 2000]. Warm/cold plots are percent warm species/(percent warm + percent cold), following the method of Rohling et al. [1997]. The previously defined biozonal boundaries I/II and II/III are also shown [Jorissen et al., 1993]. All dates in Figure 2 are expressed as uncalibrated radiocarbon convention ages, uncorrected for reservoir effect, discussed further in Table 1.
Isotope results are reported as per mil standardized to Vienna Pee Dee belemnite. Machine error are is of the order of <0.06‰ (standard deviation).

3. Results

3.1. Planktonic Foraminifera

[11] Five distinct assemblages were identified in this study (Figure 2). The main boundaries equate well with the previously identified biozonal boundary II/III, and biozonal boundary I/II [Jorissen et al., 1993] and this nomenclature is used. The three remaining assemblages consist of subdivisions of biozone I of Jorissen et al. [1993]. Relevant Mediterranean habitat characteristics have been summarized by Rohling et al. [1993a, 1993b, 1995], Pujol and Vergnaud-Grazzini [1995], De Rijk et al. [1999], and Hayes et al. [1999]. From old to young we identify the following five assemblages. (1) Assemblage III is an assemblage dominated by N. pachyderma and Turborotalita quinqueloba, with lower densities of Globorotalia scitula and Globigerinita glutinata, and a generally low to very low abundance of G. ruber. This fauna dominates the cool “glacial” intervals. (2) Assemblage II is an intermediate assemblage characterized by presence of T. quinqueloba, somewhat enhanced numbers of N. pachyderma and G. glutinata, especially in the central Aegean cores, and the presence of G. ruber and G. inflata. (3) Assemblage I is an assemblage dominated by the warm subtropical species G. ruber and the SPRUDTS group (including but not requiring Globigerinella siphonifera, Hasterigia pelagica (absent in the Aegean), Globoturborotalita rubescens, Orbulina universa, Globigerinella digi-
tata, Globoturborotalita tenella, and Globigerinoides sacculifer [see Rohling et al., 1993a, 1993b]. In addition it includes abundant Globigerina bulloides and somewhat increased T. quinqueloba compared with the preceding assemblage II. (4) Assemblage Ic is characterized by elevated absolute and relative abundances of the pink morphotype, G. ruber rosa. A peak in G. inflata is also present at the base of this assemblage, followed by a marked absence in this species in the remainder of Ic. Core LC-21 shows an interruption in assemblage Ic with a return to a fauna resembling assemblage II, which corresponds to an “interruption” of the darker-colored sapropel [Hayes et al., 1999; De Rijk et al., 1999; Mercenote et al., 2000]. (5) Assemblages Ib and Ia are diverse assemblages dominated by G. ruber and G. bulloides, with SPRUDTS, G. inflata, and G. glutinata. Ib is characterized by elevated numbers of G. inflata and appears to be of short duration. Ia shows a fauna similar to that seen in the present-day Mediterranean (core top data of Thouvenel [1978]). Assemblage Ia shows slightly lower abundances of “warm” preferring species than assemblage Ic.

3.2. Stable Isotopes

[12] Combination plots of the monospecific isotopic profiles of δ¹⁸Oruber, δ¹⁸Oinflata, and δ¹⁸Opachyderma (Figure 3) show several distinct, previously unreported features: Oxygen isotopes initially show a high degree of synchronicity, being similar in value and variation. This is followed by a depletion in δ¹⁸Oruber that separates it from the unaffected signal of N. pachyderma. This first separation between δ¹⁸Oruber and δ¹⁸Opachyderma coincides with the biozonal boundary II/III and corresponds in age with glacial termination T1a. After a short interval of little change, δ¹⁸Oruber shows a second rapid depletion to its typical Holocene values, while δ¹⁸Opachyderma remains at pre-Holocene values or even shows an enrichment. This enrichment, while generally small, is well within the sensitivity of our equipment and must be regarded as real. It is most obvious in cores SL-9 (0.8% over five samples), SL-21 (0.5% over 10 samples), and SL-31 (0.4% over four samples), while the available resolution leaves the signal in LC-21 inconclusive in this respect. The δ¹⁸O signal of G. inflata shows an inflection to lighter values at the same time as this second depletion in δ¹⁸Oruber, while the absolute δ¹⁸Oinflata values remain
intermediate between those of $\delta^{18}O_{\text{rubber}}$ and $\delta^{18}O_{\text{pachyderma}}$. This second sharp depletion in $\delta^{18}O_{\text{rubber}}$ equates with termination T1b. Finally, after a period on the order of ~1 kyr, the values of $\delta^{18}O_{\text{pachyderma}}$ also start depleting to this species’ Holocene values.

13Shortly after the start of the depletion in $\delta^{18}O_{\text{pachyderma}}$ to Holocene values, we observe a general depletion in $\delta^{13}C$. The $\delta^{13}C$ records show an initial synchronous drop in both $\delta^{13}C_{\text{rubber}}$ and $\delta^{13}C_{\text{pachyderma}}$, which is followed in SL-21, SL-31, and SLA-9 by a separation of values as $\delta^{13}C_{\text{pachyderma}}$ continues to deplete after $\delta^{13}C_{\text{rubber}}$ has leveled out. This separation in the $\delta^{13}C$ records coincides with the onset of sapropel deposition.

### 3.3. Bioturbation Model

14The offset in the isotopic responses for $\delta^{18}O_{\text{rubber}}$ and $\delta^{18}O_{\text{pachyderma}}$ around T1b is conspicuous, and we need to assess whether this is a genuine feature or the result of bioturbation. Owing to the rapid fall in numbers of *N. pachyderma* before the depletion, bioturbation might mix a relatively large proportion of undepleted foraminifera with the comparatively few depleted forams. This could potentially shift the resultant isotopic composition to less depleted values. To test whether such processes could explain the observed trends, we developed a simple model, which assumes a hypothetical step change in the isotopic compositions of both *G. ruber* and *N. pachyderma* at the same point in time. The model then simulates a progressive homogenization (bioturbation) of each successive 0.5 cm of deposited sediment with the previous 5 or 10 cm (separate model runs). Two versions of this simulation were run. An extreme version based on a large step change in numbers of both *G. ruber* and *N. pachyderma* at the same point as the isotopic shift and a run based on the actual numbers observed for these species. The results of the bioturbation model are shown in Figure 4.

15All simulations smoothed the imposed step-like $\delta^{18}O$ change into more gradual depletions similar to those seen in the sedimentary record, with differences in the profiles for the two species. However, in all cases the inflection points for the $\delta^{18}O$ values of both species is the same, with both values start to deplete at the same point. This differs markedly from the observed data with a separation in the inflection points of these species by up to >15 cm. Only in our most extreme scenario do we approach the real data, with a lag on the order of 5–10 cm, although even here the inflection points are synchronous on closer observation. The models also totally failed to reproduce the enrichment trend in *N. pachyderma* (5–10 data points) that is seen in the actual data during the major depletion in *G. ruber* (this is most clearly seen in cores SL-21, SL-31, and SLA-9).

16The bioturbation model leads us to suggest that the actual $\delta^{18}O$ depletion may have been more step-like than is preserved in the sedimentary record. We also deduce that the offset in inflection points seen in our cores reflects a real change in isotopic gradients within the water column during this period.

17Interestingly, the signal from *G. inflata* does appear to parallel the *N. pachyderma* signal shape seen in our model. It too sees a marked decline in numbers over the period of isotopic depletion. This signal may therefore provide an indication of the strength of bioturbation below the sapropel, suggesting that homogenization of foraminiferal sized particles due to bioturbation was <5 cm. This would agree with observations in the faunal counts where species *N. pachyderma*, *G. inflata*, and *G. glutinata* fall abruptly to zero in between consecutive samples at 2.5 cm spacing.

### 4. Discussion

18Changes in stable isotope composition have led to considerable speculation on the variability of Mediterranean freshwater budgets [Huang and Stanley, 1972; Cita et al., 1977; Ryan and Cita, 1977; Williams et al., 1978; Rossignol-Strick et al., 1982; Thunell and Williams, 1989; Kallel et al., 1997; Rohling and De Rijk, 1999]. Recent work points out that oxygen isotopic ratios from planktonic foraminifera cannot be used to determine absolute salinities in any straightforward way on geological timescales. Rather, they show responses to hydrographic changes that may be several times greater than the corresponding changes in conservative properties, i.e., salinity [Rohling and De Rijk, 1999; Rohling, 1999a, 1999b]. Such problems associated with temporal gradients are avoided in comparisons of isotopic compositions of different species within an individual sample since foraminifera are then analyzed from an area in a single hydrological regime. Thus differences in the signal between species will reflect real, contemporaneous differences between their preferred habitats.

19We propose that our multiproxy records are best interpreted in combination with the foraminiferal abundance records, as a series of successive changing, climatically driven, dynamic regimes. These are illustrated as a series of transitory states together with a schematic summary of the main isotopic and faunal changes recognized in our Aegean records (Figures 5 and 6). Each state represents a single point in time, which may be considered
4.1. State A

State A is interpreted as typical of the glacial Aegean Sea. This state is characterized by the absence of warm mixed layer species. We observe coinciding values of $\delta^{18}O_{ruber}$ and $\delta^{18}O_{pachyderma}$ as well as $\delta^{13}C_{ruber}$ and $\delta^{13}C_{pachyderma}$, suggesting that there were no isotopic gradients between these shallow- and subsurface-living species. Thus it is deduced that the water column during state A comprised of a single homogenized water mass, with intermediate water undistinguishable from surface water. The presence of *G. inflata* is regarded as indicative of deep seasonal mixing and is hence shown in association with winter mixing ([Hemleben et al., 1989; Rohling et al., 1995; Pujol and Vergnaud-Grazzini, 1995] summarized by Rohling et al. [1993a] and Reiss et al. [2000]). The faunal assemblage comprised of predominantly cool-water species, with only very rare occurrences of the warm dweller *G. ruber*. *T. quinqueloba* is shown as the principal surface dweller [cf. Rohling et al., 1993a]. *G. scitula* and *N. pachyderma* are shown living at depth. *N. pachyderma* is known to thrive at or just above the base of the euphotic zone and generally prefers stable stratified environments ([Hemleben et al., 1989; Rohling and Gieskes, 1989; Rohling et al., 1993a, 1993b; Reiss et al., 2000]). *G. scitula* in particular is tolerant of low temperatures ([Hemleben et al., 1989]), and we therefore show it as present exclusively in winter, although we cannot exclude its possible presence from other seasons on the basis of the data available.

4.2. State B

This state is typical of the regime we believe marks the appearance of distinct seasonal stratification in the post-glacial Aegean. The earliest depletion in $\delta^{18}O_{ruber}$ represents termination 1a. The onset of a separation between the $\delta^{18}O_{ruber}$ and $\delta^{18}O_{pachyderma}$ records suggests that *G. ruber* lived in an isotopically different water mass than *N. pachyderma*. Where available, the $\delta^{18}O_{inflata}$ values clearly follow the $\delta^{18}O_{pachyderma}$ record. Therefore we show these two species in state B as inhabiting the same water mass: *G. inflata* in the deep winter mixed season and *N. pachyderma* below the seasonal thermocline in the previous winter’s water. The increased abundance of *G. ruber* over termination 1a itself also indicates the development of seasonal stratification with a warm mixed layer since *G. ruber* has a minimum temperature requirement of $\sim 14^\circ$C ([Hemleben et al., 1989; Bijma et al., 1990a, 1990b; Reiss et al., 2000]). Hence we infer that termination 1a was associated with significant development/strengthening of the summer thermocline. This is further corroborated by the increase in *N. pachyderma* abundances since this species is known to prevail in stable stratified settings with a well-developed deep chlorophyll maximum ([Hemleben et al., 1989; Rohling and Gieskes, 1989; Rohling et al., 1993a, 1993b; Reiss et al., 2000]). Plankton tow samples from the modern NW Mediterranean clearly illustrate *N. pachyderma*’s preference for such hydrographic conditions ([Rohling et al., 1995]. The continued presence of the *G. inflata* and its peak in abundance toward the end of biozone II strongly suggests the persistence of seasonal mixing to considerable depth. The shallower-living (127 m) species *G. glutinata* [Reiss et al., 2000] also occurs in state B. This species is a specialist diatom feeder ([Hemleben et al., 1989] and is normally associated with the spring bloom, triggered by the newly available nutrients at the end of winter mixing and increased solar irradiation. Thus we interpret the overall evidence for state B as indicative of seasonal, thermal stratification, alternating with vigorous seasonal overturn of the water column in the colder months.

4.3. State C

State C in particular represents a transitory phase, marking a “snapshot” within the changing conditions from the start of termination 1b to the onset of sapropel production. This regime sees the first occurrence of *O. universa* and other members of the
SPRUDTS group. These taxa prefer warm conditions and are shown in Figure 5 above the thermocline. *O. universa* lives in a temperature range of 12°C–31°C, and although its photosynthetic symbionts show a dominant habitat in shallower waters, with sufficient light penetration, it can be found down to 250 m, [Hemleben et al., 1989]. We interpret the increase in SPRUDTS and of *G. ruber* as indicative of an increase in depth and extent of the thermocline. The regime in state C is also characterized by the distinct decoupling of the δ¹⁸O*ruber* and δ¹⁸O*pachyderma* signals at the biozone I/II boundary, marking an increasing (isotopic) isolation of intermediate/bottom waters from the surface system. There is a synchronous onset of depletions in δ¹⁸O*ruber* and δ¹⁸O*inflata*, but δ¹⁸O*pachyderma*, on the contrary, responds initially with a small enrichment. As δ¹⁸O*inflata* reflects with δ¹⁸O*ruber*, we deduce that the winter water (δ¹⁸O*inflata*) responded to the same climatic trend as the summer water (δ¹⁸O*ruber*). This is what would be expected for a substantial perturbation: the effect is seen first and strongest in the shallow summer mixed layer and subsequently passed on to the more voluminous winter mixed layer. Any species living below the summer mixed layer effectively lives in the previous year’s winter water, and hence even subsurface summer species should reflect the main isotopic change. However, as δ¹⁸O*pachyderma* shows a completely independent and opposite response, we infer that *N. pachyderma* lived in a water mass “isolated” from the surface water system. For this reason, we indicate *N. pachyderma* in an intermediate water mass (IW) that was strongly differentiated from the surface system, with the mixing indicators restricted to levels above the IW boundary. Such a condition could have resulted from an invasion of intermediate water from a remote source. Despite the presence of a physically stable environment for *N. pachyderma* we see a rapid reduction in this species’ abundance, suggesting that its habitat underwent rapid deterioration. This would fit with our proposal of remotely derived intermediate water since an increase in the pathway from its source relative to the previous locally produced AEIW would result in poorer oxygenation for intermediate water masses in the Aegean. State C also shows a marked reduction in numbers of *G. inflata*, implying that seasonal water column homogenization became inhibited. Shortly following state C, *G. inflata* numbers dwindle into insignificance. The only species showing a real increase is *G. glutinata*. *G. glutinata* may require less dramatic vertical mixing and may hence have replaced *G. inflata*. However, as *G. glutinata* has a short reproductive cycle [Hemleben et al., 1989] it’s occurrence may also be an opportunistic response to any increase in nutrient availability. We also identify an increase in numbers of the eutrophic species *G. bulloides*. On the basis of its year-round occurrence in the area today [Pujol and Vergnaud-Grazzini, 1995] we have represented it with a year-round occurrence. State C therefore represents a progressive warming of climate and a resulting reduction in local intermediate and deep water production. This allows the increasing isolation of the intermediate water, which in turn allows the

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**Figure 5.** Summary of major changes in isotopic signals δ¹⁸O and δ¹³C. The thick shaded line indicates δ¹⁸O*ruber*, thin solid line indicates δ¹⁸O*pachyderma*, and the dashed line indicates δ¹⁸O*inflata*. Depth scale is based on core SL-21, and the vertical shaded box indicates the extent of the benthic foraminiferal defined sapropel. Bold letters refer to state summaries in Figure 5. All dates are expressed as calibrated radiocarbon convention ages, corrected for reservoir effect (kyr B.P.) and derived from average ages for the event from cores in this study (Table 2). The hatched area of dysoxic water may be truly anoxic at greater depths (e.g., LC-21, 1500 m). The positions of the Younger Dryas and the Bolling-Alerød are also indicated.

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**Figure 6.** Schematic reconstruction of the history of Aegean circulation. The states illustrated are transitory, and this schematic represents the typical changes in deep water mixing and faunal distribution in the Aegean since the last glaciation. The timing of these transitions and placement of these typical assemblages are summarized in Figure 5.
Table 2. Timing of Events

<table>
<thead>
<tr>
<th>Event Description</th>
<th>Core</th>
<th>Depth (m)</th>
<th>Calibrated kyr B.P.</th>
<th>Youngest Age ± (Range/2)</th>
<th>Calibrated kyr B.P.</th>
<th>Youngest Age ± (Range/2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Start of δ18Oruber depletion T1a</td>
<td>LC-21</td>
<td>Depth 1520 m</td>
<td>1.57</td>
<td>1.54 ± 0.2</td>
<td>1.5</td>
<td>1.5 ± 0.2</td>
</tr>
<tr>
<td>Start of δ18Opachyderma depletion</td>
<td>SL-31</td>
<td>Depth 430 m</td>
<td>1.58</td>
<td>1.48 ± 0.2</td>
<td>1.5</td>
<td>1.5 ± 0.2</td>
</tr>
<tr>
<td>Start of δ13C minimum</td>
<td>SL-21</td>
<td>Depth 630 m</td>
<td>1.6</td>
<td>1.1 ± 0.2</td>
<td>1.5</td>
<td>1.1 ± 0.2</td>
</tr>
<tr>
<td>Start of δ13Cpachyderma depletion difference</td>
<td>SL-21</td>
<td>Depth 630 m</td>
<td>1.6</td>
<td>0.9 ± 0.2</td>
<td>1.5</td>
<td>0.9 ± 0.2</td>
</tr>
<tr>
<td>Age of maximum δ13Cruber</td>
<td>SL-21</td>
<td>Depth 630 m</td>
<td>1.6</td>
<td>0.9 ± 0.2</td>
<td>1.5</td>
<td>0.9 ± 0.2</td>
</tr>
<tr>
<td>Benthic sapropel</td>
<td>SL-21</td>
<td>Depth 630 m</td>
<td>1.6</td>
<td>0.9 ± 0.2</td>
<td>1.5</td>
<td>0.9 ± 0.2</td>
</tr>
</tbody>
</table>

*This table details the interpolated starting ages of important horizons in cores LC-21, SL-21, and SL-31. Ages are given in ka, with the range between cores expressed as an error. The mean age of each event is given in the last column, with the range between cores expressed as an error.

4.4. State D

This represents the deposition of sapropel S1 and is the culmination of the sequence of changes that started with termination 1b. Roughly 1 kyr after the start of the T1b depletion in δ18Oruber, a similar depletion begins to show up in δ18Opachyderma. The longer time lag suggests that diffusive mixing was the major mechanism for transfer of the isotopic depletion from the surface (G. ruber) to the deeper environments (N. pachyderma) since convective mixing would have caused a virtually instantaneous response between the two species (see, for example, the coincidence between the onsets of depletion in δ18Oinflata and δ18Oinflata). The anomalous response of δ18Opachyderma led us to conclude that it was living subsurface in a water mass unaffected by the local seasonal homogenization (see state C). The δ13Cpachyderma in state D becomes strongly depleted relative to δ13Cinflata and δ13Cruber, suggesting that the strongly reduced N. pachyderma population that could survive did so subsurface in an "isolated" poorly ventilated water mass with accumulation of 13C-rich remineralization products. We therefore contend that the eventual depletion in δ18Opachyderma resulted from a slow diffusive mixing process. Our inference of a halt in convective mixing is supported by the presence of dysoxic indicators in the benthic foraminiferan fauna within S1 since a lack of convective overturn results in poor ventilation and consequently dysoxia in bottom waters. For some intervals, at depth (LC-21, 1500 m), no benthic species survive at all, suggesting that bottom waters became totally anoxic. In the planktonic foraminifera, assemblage δ dominates, which consists of predominately warm-water species with a notable absence of fall/winter/spring mixing indicators. Hence we infer that there was a strongly developed, possibly year-round, thermocline/halocline. During this regime we see depletion in δ13C for all species recorded. This may be the effect of influx of terrestrial dissolved organic carbon (DOC) [Asu et al., 1999] since this period is known to coincide with a widespread increase in humidity. This is also evidenced by high North African lake levels, high abundance of humidity markers in the local palynological records, and the isotopic anomalies in speleothem data [Rossignol-Strick, 1995; Edmunds et al., 1999; Bar-Matthews et al., 1999; Tzedakis, 1999; deMenocal et al., 2000]. The resultant increase in freshwater input is schematically represented by a rainfall symbol in state D, even though much of the fresh water would have arrived in the form of river runoff rather than direct precipitation [Jenkins and Williams, 1983; Shaw and Evans, 1984; Thunell and Williams, 1989; Rohling and Hilgen, 1991; Rohling, 1994, 1999b]. As mentioned previously, δ13Cpachyderma depletes more than δ13Cruber and δ13Cinflata, which suggests that such any DOC explanation needs to be combined with the concept that N. pachyderma survived in an ageing water mass. With the intermediate water isolated and (virtually) stagnating, it would accumulate an excess of 13C from remineralization. In contrast, the shallow δ13Cruber signal is continually being equilibrated with contact with the atmosphere. Note that the separation in δ13C signals coincides with the onset of sapropel formation and therefore with the appearance of benthic markers for very poor bottom water oxygenation, suggesting a culmination of subsurface/deep water stagnation.

4.5. State E

Here we recognize the establishment of a modern circulation regime. The δ18Opachyderma and δ18Oinflata signals have come back together, suggesting that both live in isotopically undifferentiated winter water. The δ18Oruber is also similar to δ18Opachyderma and δ18Oinflata in SL-21 but is slightly more depleted in the remaining cores. This suggests that intermediate and surface waters...
in this area are once again very similar, which implies direct local communication between these water masses. This may indicate that (1) intermediate waters are upwelling to the surface and/or (2) surface waters directly contribute to intermediate water formation. Slight δ13C differentiation between N. pachyderma and G. ruber/ G. inflata suggests that N. pachyderma continued to live subsurface at levels more affected by remineralization than the surface/ mixed layer environments preferred by G. ruber and G. inflata. The return of deep mixing indicator G. inflata and the presence of the spring bloom indicator G. glutinata suggests that seasonal mixing is again well developed. In addition, we see a dominance of G. bulloides, giving an overall faunal aspect that is very similar to that seen in modern records [Thunell, 1978; Pujol and Vergnaud-Grazzini, 1995].

4.6. Deductions on Mechanism and Timing of Circulation Change

[25] Having identified a general trend of reduced potential for deep overturn in the circulatory system during the 6 kyr leading up to sapropel production, we now consider the mechanism and timing of these changes. The trend starts from a glacial environment characterized by a single well-mixed water mass with strong accordance between isotopic signals in all species analyzed. This suggests little density contrast in the water column, while the presence of deep mixing species like G. inflata points to regular homogenization/overturn. We have interpreted this as indicative of glacial circulation. This glacial circulation appears to alter with the start of termination 1a. At that time, a shift in environment is clearly indicated in both the fauna and the isotopic data. Our interpolated dates suggest the onset of this change at ~15.9 ka (Table 2). This change associated with T1a appears to have occurred very rapidly, within a few hundred years.

[26] The subsequent regime is characterized by the start/ strengthening of summer stratification. We believe this change in circulation was driven by the combination of climatic warming and rising sea levels at the onset of the Bölling-Allerød. During that interval we see the first substantial occurrence of warm mixed layer species, suggesting a general sea surface warming. Together with the increase in N. pachyderma, this leads us to suggest the development/strengthening of a stable summer thermocline in a generally well oxygenated environment. Studies of African lake levels and aeolian dust influxes [Edmunds et al., 1999; deMenocal et al., 2000] suggest that this period also saw the start of a regional humidity increase (African humid phase (AHP)). Any increase in freshwater budget would have increased surface buoyancy, helping to establish (seasonal) stratification. This state persisted until the onset of the Younger Dryas (YD) at ~12.5 ka.

[27] In common with many Mediterranean records, the YD is represented in this study by a plateau in the δ18O records. The fauna shows increases in the mixing indicator G. inflata and a decrease in relative abundance of the warm mixed layer species. This suggests a strengthening of winter convection during this period. The YD has also been identified as an interruption in the AHP [deMenocal et al., 2000], and probably represented a cool and climatic event [Rassoulzadegh-Strick, 1993, 1995, 1999]. The YD conditions continue until the start of termination 1b marked at ~11.3 ka in our records.

[28] The start of T1b marks a clear dissociation between δ18O_pachyderma and δ18O_ruber/δ18O_inflata, from which we deduce that the (intermediate) water mass in which N. pachyderma lived was no longer locally connected to the surface system. In addition, we argue that the major source region of intermediate water at this time was somewhere outside of the Aegean. Adriatic intermediate water would be one possible source of our proposed “foreign” intermediate water in the Aegean. This agrees with suggestions of Myers et al. [1998] that, the Adriatic Sea was a major source of intermediate water to the eastern Mediterranean during sapropel formation and that this under certain circumstances could enter the Aegean Sea [Myers and Rohling, 2000]. This introduction of foreign intermediate water also has implications for the dating of foraminiferal material, a proportion of which will be living in this older water. Correlation between the LC-21 timescale and the GISP II data suggests we see the start of a 350 year offset in LC-21 dates within the anoxic phase. This offset appears to be largely finished by the time of the deposition of the Santorini ash layer [Rohling et al., 2002]. T1b marked the end of the peak in G. inflata abundance and a shift toward the dominance of the shallower-living species G. glutinata and of warm mixed layer species. This signals the inhibition of deep mixing and the isolation of intermediate waters, corroborating our interpretation of the separation between δ18O_ruber and δ18O_pachyderma. From this point on, remineralization products were able to build up in the subsurface to deep waters.

[29] With the ending of the Younger Dryas, the AHP recommenced and persisted until 5.5 ka [deMenocal et al., 2000]. In the Aegean cores we note a depletion in δ13C for all species analyzed, starting at ~10.5 ka, which we interpret as a possibly the result of increasing humidity in the Aegean Sea. Coinciding with this depletion in δ13C, G. glutinata replaces G. inflata. This may be tentatively explained in terms of increased freshwater input, which reduced surface buoyancy loss and hence suppressed mixing.

[30] With the suppression of convective mixing in the Aegean, diffusion-type processes become the main driver for property exchange through the water column. At ~10.2 ka, we identify an indication in δ18O_pachyderma, which seems to mark the start of the conveyance of the termination 1b signal to the isolated intermediate water. This allows us to determine a timescale of diffusive mixing. By assessing the time between the maximum δ18O_ruber-δ18O_pachyderma gradient and its 1/e-fold reduction we can roughly estimate the diffusive timescale for the basin in its stratified sapropel mode. We calculate this to be ~450 years (Table 2).

[31] By ~9.8 ka, deposition of sapropel S1 has commenced (Table 2). This is ~400 years after the total suppression of mixing inferred from the start of the overall δ13C depletion and the faunal shift to a virtually complete dominance of mixed layer species. The start of sapropel production also occurs ~1500 years after the isolation of the deep/intermediate waters, inferred from the separation of δ18O_ruber and δ18O_pachyderma. The isolation of subsurface waters would have allowed subsurface accumulation of remineralization products over a period of up to 1.5 kyr before these products became available for production in the euphotic zone. We suggest that this long-term accumulation provided a major source of excess nutrients that could sustain enhanced productivity during sapropel deposition.

[32] On the basis of benthic foraminiferal indicators, S1 persisted until 6.5 ka (Table 2). At its termination the return of G. inflata indicates the restart of seasonal mixing. The warm/cold plots (Figure 2) suggest this may have been a cooling event. In LC-21 the end of the benthic sapropel is marked in the planktonic δ13C, 14C, and δ18O records. The LC-21 timescale and the GISP II record by the occurrence of δ18O_pachyderma. From this point on, remineralization products were able to build up in the subsurface to deep waters.
Around that time the fauna also settled into its modern abundance distributions (Ib-Ia transition, Table 2).

5. Conclusions

[33] There is a clear link between the Aegean hydrographic regime and the global deglaciation phases. Seasonal stratification is weak to nonexistent before the onset of termination 1a, while intermediate water was virtually indistinguishable from shallow waters. After the onset of termination 1a, we identify a distinct seasonal stratification alternating with vigorous overturning seasons. This resulted in a winter mixed layer of similar characteristics to the intermediate water and a summer mixed layer that was distinguished from this by a marked seasonal thermocline. Tlb marks the next distinct hydrographic change when intermediate water became dissociated from the summer/winter mixed layers in the study area. This noncommunication between the surface and intermediate system indicates reduced/curtailed ventilation of intermediate and deeper waters. This implies that property exchanges would have been dominated by flow (diffusive mixing) and we estimate a 1/e-fold diffusive mixing timescale of ~450 years. This gives the first ever observation-based quantitative estimate of this timescale in the stratified (sapropel mode) Mediterranean.

[34] Furthermore, the dissociation of surface and intermediate systems allowed remineralization products to accumulate in intermediate and deep waters, over a period of up to ~1.5 kyr prior to sapropel deposition. Meanwhile, the isolated intermediate water would have become progressively more dysoxic, a process augmented by the observed increase in (year-round?) stratification. The sapropel therefore appears to represent the culmination of a dynamic nonsteady state process. This has important consequences for the currently accepted steady state approach to circulation and property budget calculations for the eastern Mediterranean in sapropel mode. The sapropel mode ended with the reoccurrence of winter mixing indicative species at ~6.5 ka, while the return to a modern faunal assemblage was completed by ~4.9 ka.

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