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Palaeoclimate and the formation of sapropel S1: inferences from Late Quaternary lacustrine and marine sequences in the central Mediterranean region

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Abstract

Synchronous responses to climatic changes during the Late Pleistocene–Holocene transition are inferred from marine and lacustrine stratigraphic records in the central Mediterranean region. New stratigraphic data are presented from well-dated sequences in the Meso-Adriatic Depression (MAD), two crater lakes in the Lazio region, and in the Tyrrhenian Sea. The sequences all span the last termination and the Holocene, but we focus here on the evidence in each record for the time period during which sapropel S1 formed in the Mediterranean (ca. 9.0 to 6.8 cal kyr B.P.). The new records provide evidence of palaeoenvironmental changes on land and sea that can be reconstructed at a high temporal resolution, and which throw some light on the processes which led to the formation of S1. The collective evidence indicates that: (i) organic-rich sediments occurred in both the marine and the crater lake sites during the time of formation of the S1 sapropel; (ii) there is evidence of increased stratification and anoxia in the sea-water column during the period of S1 formation; (iii) the S1 period in the study area is divisible into two sub-phases (S1a and S1b), which reflect short-term variations in oceanographic conditions (stratification and anoxia); (iv) changes in stratification in the marine column were contemporaneous with regional climate variations that are inferred from the terrestrial records. We conclude that the key factor that initiated the formation of S1 was increased discharge of freshwater into the Mediterranean following a change post-9.0 cal kyr B.P. to a warmer and wetter climate. Furthermore, the period of S1 formation was interrupted by a short-lived episode (ca. 500 years) of comparatively cooler and drier conditions during the Early–Middle Holocene transition (EMHT). © 2000 Elsevier Science B.V. All rights reserved.

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1. Introduction

Sapropels are distinctive layers of organic-rich sediment commonly observed within marine sediment cores recovered from many localities throughout the Mediterranean Basin (e.g. Olausson, 1961; Ryan, 1972; Cita et al., 1977; Cita and Grignani, 1982; Vergnaud-Grazzini et al., 1986; van Straaten, 1972, 1985; Thunell et al., 1977). A number of sapropel layers have been observed in marine sedimentary sequences that extend over several glacial–interglacial cycles, which suggests that the conditions under which sapropels form are linked, directly or indirectly, to Quaternary climate forcing mechanisms. Early theories on their mode of formation proposed that increased precipitation and freshwater run-off during glacial intervals led to stagnation in the water-column and hence rapid accumulation of organic detritus (Bradley, 1938; Kullenberg, 1952). More recent theories, based on multi-proxy palaeo-environmental investigations, link the formation of sapropels to changes in stratification and/or the chemistry of the water column, which are thought to lead to increased anoxia in bottom water, increased primary productivity, and more complete preservation of organic matter (e.g. Rossignol-Strick et al., 1982; Calvert et al., 1992; Howell and Thunell, 1992; Sancetta, 1994; Rohling, 1994). Others (e.g. Jan Bosch et al., 1997) consider that sapropels are formed by anoxia in the photic zone.

The most recent (Holocene) sapropel (S1), widespread throughout the eastern Mediterranean Basin and the Adriatic but less well-developed in the western Mediterranean, formed between ca. 9.8 and 6.8 cal kyr B.P. However, this is a simplification, because what is termed the ‘S1 layer’ is often a complex feature, one that is difficult to detect in some places because it is susceptible to extensive oxidation, resulting in considerable thinning, or even its complete removal (van Straaten, 1972). Where this is observed, the terms ‘missing sapropel’ or ‘sapropel-like’ sediment are often employed (Higgs et al., 1994; Thomson et al., 1995; van Santvoort et al., 1997). Furthermore, Rohling et al. (1997) and De Rijk et al. (1999) have shown that sapropel S1 is not a continuous feature in the eastern Mediterranean, and they

concluded that conditions necessary for sapropel formation were discontinuous during the period in which sapropel S1 was formed. The formation of sapropels may thus be sensitive to subtle changes in palaeoceanographic conditions, which in turn depend on regional climatic variations (see e.g. Kallel et al., 1997a,b).

During the last 20,000 years, the magnitude of the influence of Atlantic waters flowing into the Mediterranean through the Strait of Gibraltar has not been constant due to changing sea levels. The current consensus, however, is that the Mediterranean circulation pattern was similar to the modern one even during glacial stages (e.g. Béthoux, 1984; Thunell et al., 1987). Nevertheless, it is possible that short-term changes in circulation occurred, altering the salinity of surface waters, as a result of abrupt changes in sea-level and of regional climate conditions during the last deglaciation and the Holocene. Short-term salinity changes and current reversals have been proposed as possible ‘trigger’ mechanisms for sapropel formation (Stanley, 1978; Sarmiento et al., 1988).

It is possible therefore that there is a close association between sapropel formation and regional climatic changes. However, this hypothesis is difficult to test in any rigorous way because: (i) most detailed studies of sapropels have been concentrated in the eastern Mediterranean, and comparatively few have been undertaken in other parts of the Mediterranean Sea; (ii) studies of the palaeoclimate of the Mediterranean area are predominantly based upon palynological data, and few quantitative estimates of past climatic changes have been made (Rossignol-Strick and Planchais, 1989; Aksu et al., 1995; Zonneveld, 1996); and (iii) most studies have either focused on changes in the sea water column (palaeoceanographic changes) or on terrestrial records, and very little direct comparison between the two realms has been reported (e.g. Rossignol-Strick and Planchais, 1989; Zonneveld, 1996).

In this paper we examine the extent to which the S1 sapropel layer coincides with the evidence for regional climatic changes in the Mediterranean region. We present stratigraphical and palaeoclimatic data from two marine and two terrestrial sequences that have been securely dated and which

can be correlated precisely. Sedimentary cores extending over the Holocene and the last termination have been obtained from the Meso-Adriatic Depression (MAD), two crater lakes in central Italy (Lakes Albano and Nemi) and a Circum-Tyrrhenian basin. Stratigraphical comparisons can thus be made along a transect from the Adriatic, through central Italy to the Tyrrhenian Sea. The marine records are located close to the mainland, facilitating land–sea correlations. Furthermore, we adopt a multi-proxy approach to study the conditions under which sapropel has formed on marginal areas where dark, organic-rich sediment rarely occurs.

2. Study area and site locations

The Adriatic Basin is an elongated and narrow epicontinental shelf that contains the MAD, an almost-closed basin about 250 m deep (Fig. 1). This basin contains a continuous sequence of Late Quaternary marine sediments (Trincardi et al., 1996) because it remained subaqueous even during the lowest eustatic sea level of the last glacial maximum (LGM). The modern circulation pattern is such that surface waters from the western Mediterranean reach the Adriatic Sea via the Ionian Gyre and the Strait of Otranto (e.g. Wust, 1961). Fresh, nutrient-rich waters are fed in from the north by the Po River and this influx encourages high primary productivity in the upper part of the water column. The circulation of these plumes of low-salinity Adriatic waters can be detected along the coast of Italy as far south as Sicily (Fig. 1; Artegiani et al., 1989, 1997; Orlic et al., 1992). A subsurface contribution of LIW also enters the Adriatic. Finally, cold and dense waters are produced in winter by the action of the cold and dry north-easterly Bora wind, which leads to deep water flow to the Mediterranean via the southern Adriatic Basin (Zore Armanda, 1963; Artegiani et al., 1989).

The Tyrrhenian Sea occupies a small, triangular-shaped back-arc basin. The basin comprises a central, deep abyssal plain where new oceanic crust forms, fringed by a narrow continental shelf and complex slope, the latter indented by several minor

slope basins (Circum-Tyrrhenian basins) within which sediment accumulation rates are high (see Wezel et al., 1982; Trincardi et al., 1995). Modern surface waters of the Tyrrhenian Sea are dominated by the influx of less saline Atlantic waters (Béthoux, 1984).

Lakes Albano and Nemi occupy two calderas in the Albani Hills in Latium, central Italy, ca. 25 km south–east of Rome (Fig. 1). Both lakes are hydrologically-closed basins, receiving water mainly from precipitation and underwater springs (Chondrogianni et al., 1996a). Lake Albano, located at an altitude of ca. 293 m a.s.l., is the largest and deepest of the two lakes, with an area of 6 km², and a maximum depth of 175 m. Located at ca. 320 m a.s.l. altitude, Lake Nemi has an area of 1.8 km², and a maximum water depth of 32 m.

Extensive seismic surveys have been conducted in the MAD and in the two crater lakes as part of the PALICLAS project funded by the EU (Oldfield and Guilizzoni, 1996). Sedimentary cores recovered from various water depths at each site provide comprehensive stratigraphical information which has been analysed for various fossil types, sediment magnetic properties, stable isotope variations and major element geochemistry. A continuous high-resolution record of the last glacial–Holocene transition and of the entire Holocene has been obtained from core CM92-43 in the MAD (Figs. 2 and 3). The multi-proxy data from this core are compared with those obtained from core MC82-12, which was recovered from the Palinuro Circum-Tyrrhenian basin as part of the CNR project ‘Oceanografia e Fondi Marini’ (Wezel et al., 1982). The Tyrrhenian Sea core provides a continuous marine record from marine oxygen isotope Stage 2 until the present day (Fig. 1).

Several cores up to 14 m long were retrieved from Lakes Albano and Nemi in water depths ranging between 77 and 252 m and between 30 and 20 m, respectively (Table 1). The sedimentary infill in the two basins spans the period from oxygen isotope Stage 3 to the present time (Chondrogianni et al., 1996b). We present here the data from each core sequence that provide the best resolution of part of the last termination and the Holocene. Core PALB94-3A, retrieved at

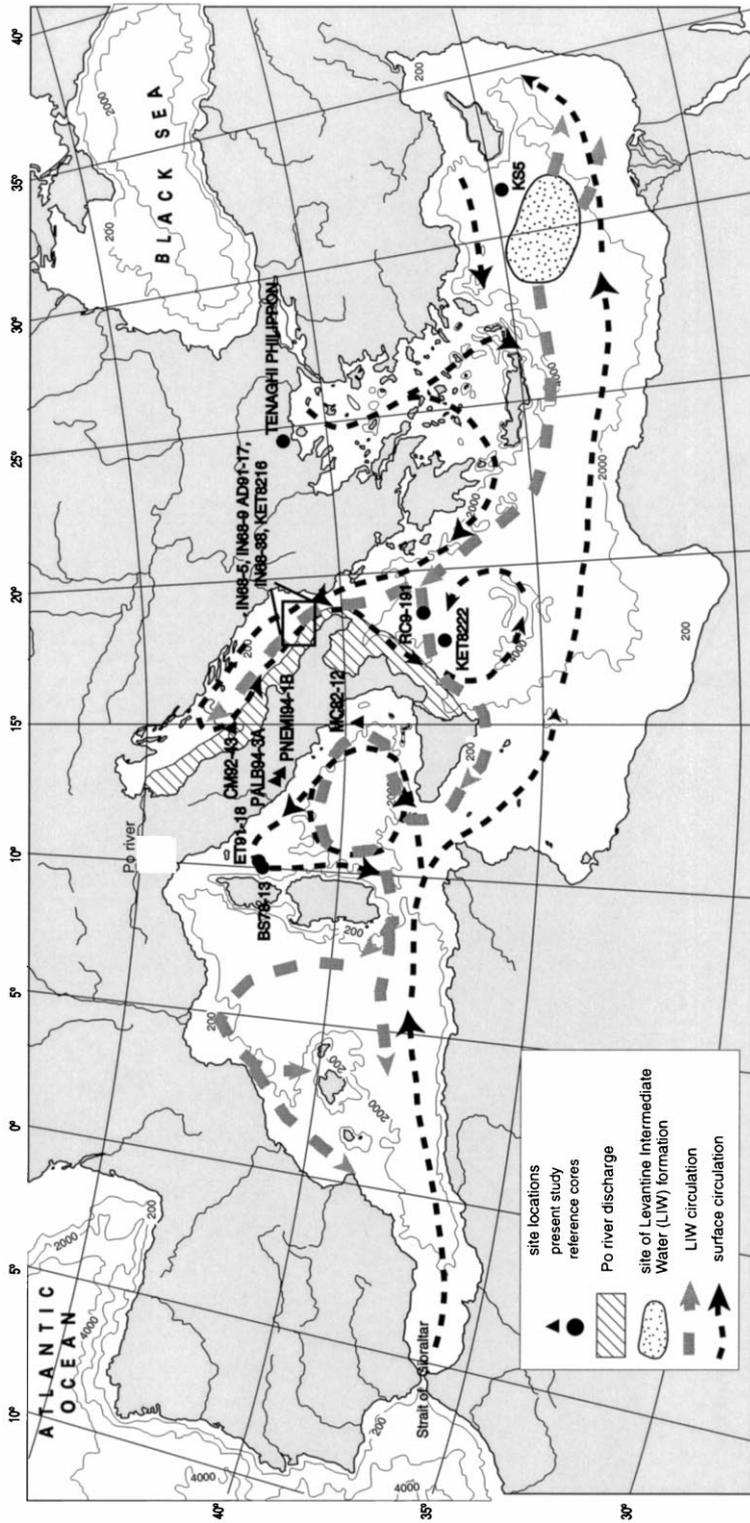


Fig. 1. Simplified bathymetry of the Mediterranean Basin showing the modern circulation and the main drainage pattern on the surrounding land mass. The hatched area indicates the spread of fresh water from the Po discharge in the Adriatic and Ionian Seas. Dark, dashed arrows indicate the surface circulation of Atlantic waters and thick grey arrows depict the circulation pattern of the Levantine intermediate waters (LIW); this water mass forms in the eastern Mediterranean (dotted area). Triangles indicate the location of the cores studied in this paper; solid dots indicate core locations from other published marine and terrestrial sites discussed in this paper.

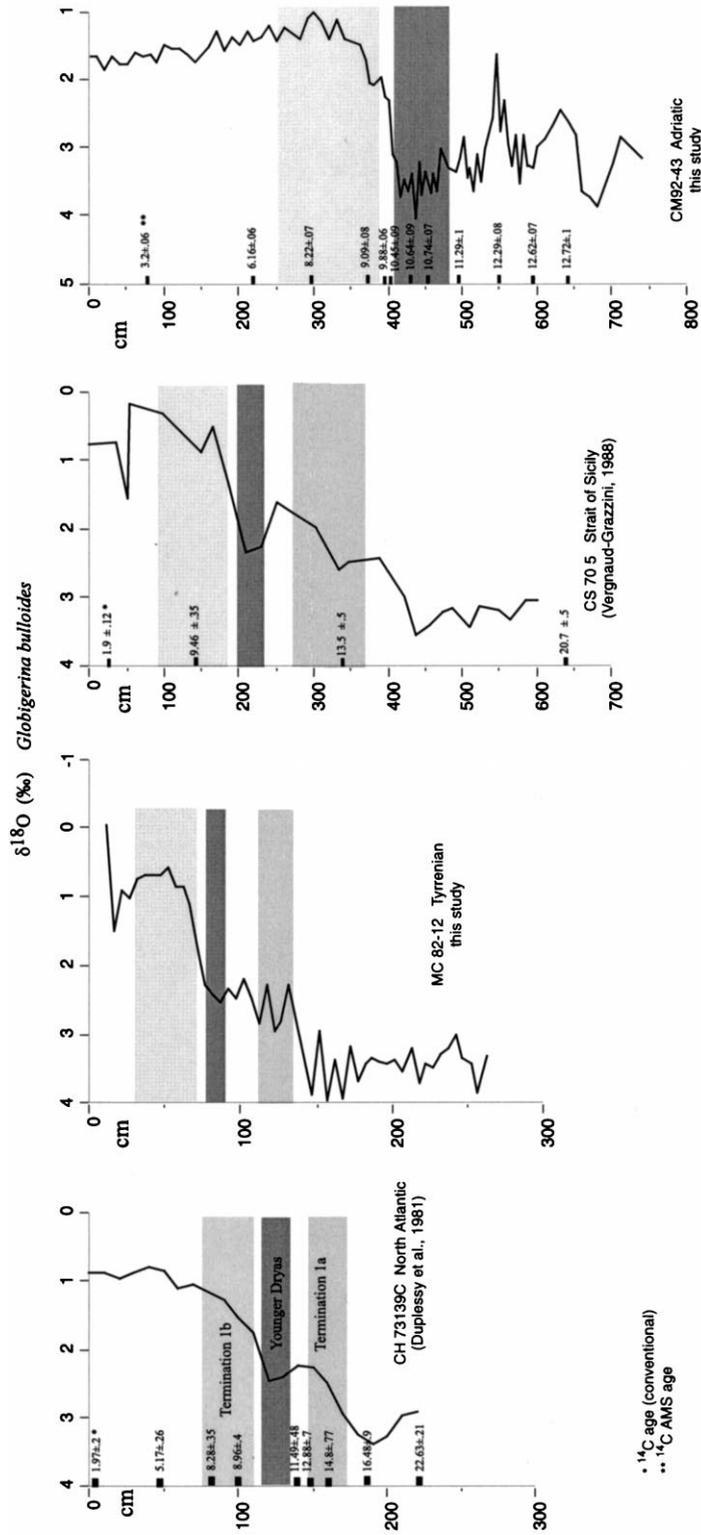


Fig. 2. Correlation based on oxygen isotope data between our marine cores (MC82-12 and CM92-43) and other key cores in the Mediterranean (Vergnaud-Grazzini et al., 1988) and Atlantic (Duplessy et al., 1981) records. Depths are in centimetres in all cores and ^{14}C ages are reported, where available. The chronology in core CM92-43 is based on AMS dates (Langone et al., 1996), whereas the other records have been dated using the conventional ^{14}C method. The main patterns of correlation palaeoceanographic events are evident in all cores: light grey indicates melt-water intervals and dark grey depicts the YD or GSI cold interval.

core CM92-43

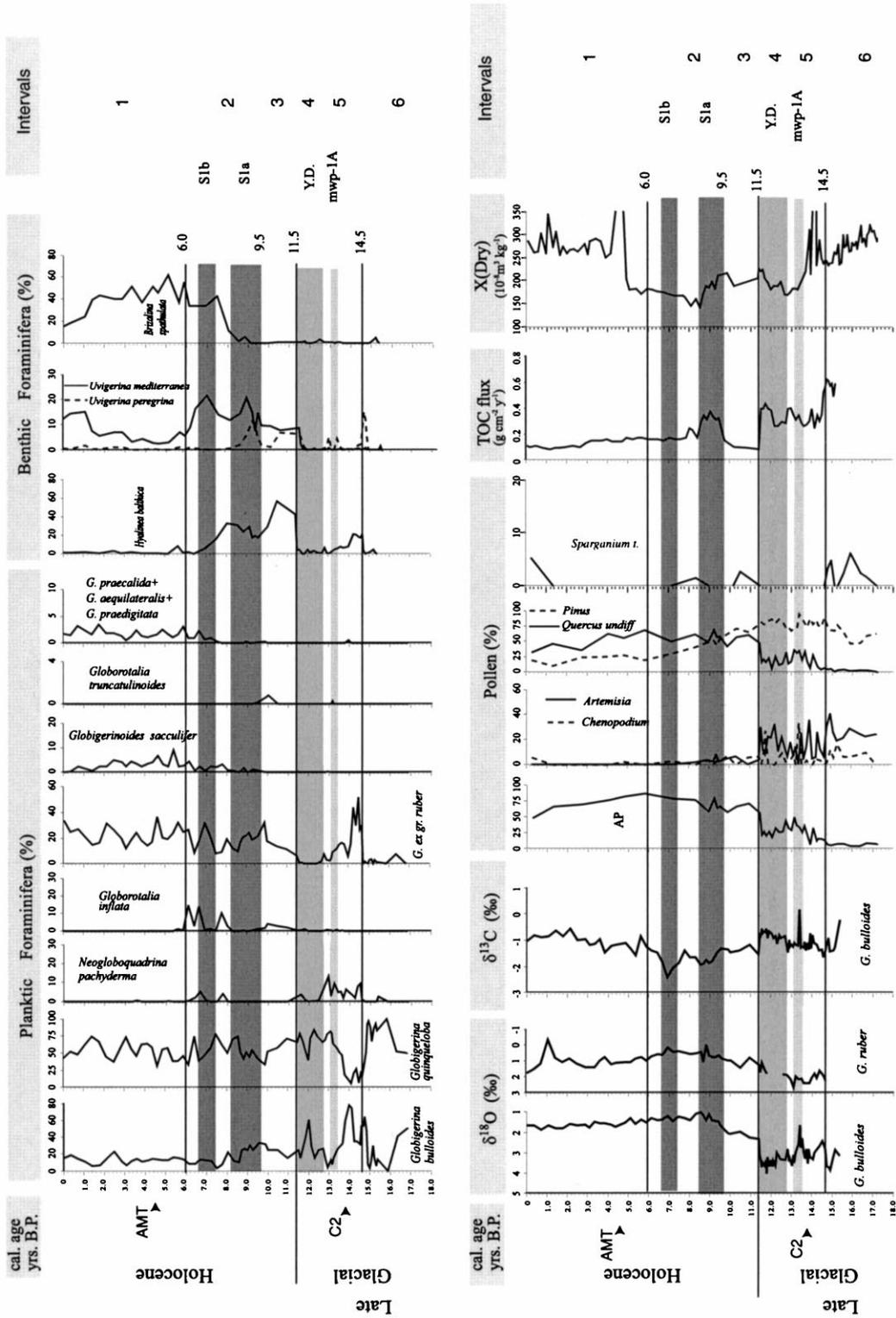


Fig. 3. Multi-proxy data from core CM92-43 plotted versus age in calendar years. AMT: Agnano Montespina Tephra; C2: Neapolitan Yellow Tuff. Note the short interval of depleted $\delta^{18}\text{O}$ on *Globigerina bulloides* centred at 13.5 kyr B.P., most probably corresponding to mwp-1A (Fairbanks, 1989). During the time of sapropel S1 formation the sediments from the Adriatic Basin show the most depleted $\delta^{18}\text{O}$ composition in both surface and intermediate water planktic foraminifers, high TOC fluxes and depleted $\delta^{13}\text{C}$ values. The interruption within the unit equivalent to the S1 sapropel is marked by a peak of *Globigerina inflata* and an interval of relatively more enriched $\delta^{13}\text{C}$ compositions.

Table 1

Geographic location, water depths (or altitude above mean sea level) for the cores discussed in the text and appropriate references

Core number	Coordinates	Location	Water depth (m)	References
CM92-43	14°43'N/42°53'E	Central Adriatic	252	Ariztegui et al. (1996a); Asioli (1996); Trincardi et al. (1996); this study
MC82-12	39°43'N/14°25'E	Tyrrhenian Sea (Palinuro Basin)	1657	Tamburini et al. (2000); this study
PALB94-3A	see Chondrogianni et al. (1996a,b)	Lake Albano at 293 m a.s.l.	120	Ariztegui et al. (1996b); this study
PNEMI94-1B	see Chondrogianni et al. (1996a,b)	Lake Nemi at 320 m a.s.l.	31	Chondrogianni et al. (1996a,b); this study
KS5	34°49'N/31°44'E	Ride de Florence	1560	Znaidi Rivault (1982)
IN68-5	41°14'N/18°32'E	Southern Adriatic	1030	Jorissen et al. (1993); Capotondi et al. (1999)
IN68-9	41°47'N/17°54'E	Southern Adriatic	1234	Rohling et al. (1997)
AD91-17	40°52'N/18°38'E	Southern Adriatic	844	Capotondi et al. (1999)
IN68-38	41°07'N/17°34'E	Southern Adriatic	716	Jorissen et al. (1993)
KET8216	41°31'N/17°59'E	Southern Adriatic	1166	Fontugne et al. (1989)
RC9-191	38°11'N/18°02'E	Ionian Basin	2345	Fontugne et al. (1989)
KET8222	37°56'N/16°53'E	Ionian Basin	1691	Fontugne et al. (1989)
ET91-18	42°36'N/9°52'E	Tyrrhenian Sea (Corsica Basin)	651	Capotondi (1995)
BS78-13	42°22'N/9°53'E	Tyrrhenian Sea (Corsica Basin)	829	Benvenuti (1989)
AC85-4	41°45'N/11°46'E	Tyrrhenian Sea	662	Capotondi et al. (1989)
TENAGHI PHILIPPON		Southern Macedonia (NE Greece) at 40 m a.s.l.		Wijmstra (1969)

120 m water depth in Lake Albano, consists of 1200 cm of well-laminated and mostly fine-grained (mud) deposits. Core PNEMI94-1B, recovered from Lake Nemi at a water depth of 30 m, consists of ca. 900 cm of fine-grained deposits, mainly with diatom-rich, distinct laminations.

3. Methods

Subsamples for foraminiferal analysis were extracted every 5 or 10 cm throughout the marine cores. Since sediment accumulation rates were not constant, however, the temporal resolution of the foraminiferal records is variable. The subsamples were weighed, washed and sieved with a mesh width of 63 μm (core CM92-43) or 125 μm (core MC82-12) and then split into aliquots containing at least 300 individuals each of planktic and benthic taxa (Asioli, 1996; Tamburini et al., 1998). The

data are expressed as percentages of the total number of planktic or benthic foraminifers.

Stable oxygen and carbon isotopic compositions of selected planktic foraminifers were measured following the standard procedures of the Stable Isotopes Laboratory, ETH, Zürich. The results are reported as per mil deviation with respect to the international standard V-PDB. The reproducibility of the measurements is $\pm 0.1\%$. Foraminiferal samples were ultrasonically cleaned in order to remove contamination and/or diagenetic alterations, such as coccoliths, overgrowth and detrital fill. Hand-picked foraminifers showing no signs of diagenetic alteration were selected under a binocular microscope. None of the isotopic data presented here have been corrected for the ice volume effect. Two planktic species were selected for isotopic measurement: *Globigerina bulloides* and the near-surface-dwelling *Globigerinoides ex gr. ruber*. The first species is present throughout the entire

core sequences, while the second is missing from some critical intervals of the record, such as, for example, the sections assigned to the Younger Dryas cold event (zone GS-1 of the GRIP ice-core record — see Björck et al., 1998). Measurements have also been obtained from one benthic species, *Bulimina marginata*.

Sediment samples from both the Adriatic and crater lake sediment sequence were prepared for pollen analysis following the methodology described by Lowe et al. (1996b). For the lake sequences, pollen sums varying between 200 and 700 of 'dry-land' taxa have been adopted, and all percentages are expressed as percentages of total 'dry-land' pollen. Because marine deposits frequently contain very high percentages of coniferous pollen, due to their resistance to decomposition, ease of transport and buoyancy, pollen of *Pinus* have been excluded from the pollen sums of all the pollen diagrams presented here, including those relating to the crater lake sequences.

Variations in organic carbon content of the Adriatic sediment sequence were measured using a Fison CHN elemental analyser. Prior to measurement, samples were treated with 2 N HCl to eliminate the carbonate fraction. The hydrogen index (HI), measured in the Lake Albano sediments and expressed in milligrams hydrocarbon per gram of organic carbon, is an estimate of the amount of hydrogen contained in the hydrocarbon component of sedimentary organic matter (i.e. chemical quality of the organic matter). Extraction and measurements of the concentrations of pigments follows the methodology described by Ryves et al. (1996). Total pigments were extracted with 90% acetone and specific algal pigments were determined by ion pairing reverse-phase HPLC and expressed as nanomole per gram organic matter.

Rock magnetic parameters are a diagnostic tool for monitoring environmental changes in sedimentary sequences, especially for identifying intervals of reductive diagenesis such as sapropel layers (e.g. Thompson and Oldfield, 1986; Oldfield, 1996; van Santvoort et al., 1997; Vigliotti, 1997). Magnetic parameters from cores CM92-43 and PNemi-1B were measured on discrete samples at the palaeomagnetic laboratory of the IGM,

Bologna. The low-field magnetic susceptibility was measured using a Bartington MS1 susceptibility meter. Anhysteretic remanence (ARM) was induced in the sample by applying 99 mT of alternating field and a 0.05 mT biasing field for the Adriatic samples and 0.1 mT for the Nemi samples. Isothermal remanence (IRM) was induced up to 1 T. ARM and IRM were measured using a JR-4 spinner magnetometer for the CM92-43 samples, whereas a Minispin spinner magnetometer was used for analysis of the Nemi core.

4. Stratigraphic correlations and chronology

The marine cores can be correlated using oxygen isotope stratigraphy and foraminiferal data. The oxygen isotope stratigraphy for the last glacial–interglacial cycle from cores in the Mediterranean Sea matches well the characteristic changes in the Atlantic isotope record (Vergnaud-Grazzini et al., 1986, 1989). Indeed, the $\delta^{18}\text{O}$ signal in Mediterranean cores also shows the two steps of the last termination seen in some Atlantic cores (Termination IA at 14.8–11.5 ^{14}C kyr B.P. and Termination IB at 9.5 ^{14}C kyr B.P., separated by an interval of more positive oxygen isotopic composition between 10.0 and 11.0 ^{14}C kyr B.P.; Duplessy et al., 1981; Fairbanks, 1989). These features can also be observed in the Adriatic and Tyrrhenian cores discussed in this paper (Fig. 2), while oxygen isotope Stages 1 and 2, along with a reversal which is dated to the Younger Dryas (YD or GS-1) event, are also clearly identifiable.

Several foraminiferal marker events, already well-recognised for the central Mediterranean (Jorissen et al., 1993; Capotondi, 1995; Asioli, 1996; Capotondi et al., 1999), were used to constrain the age of core MC82-12 and to strengthen the correlations between the Adriatic and Tyrrhenian sites. This approach was preferred to linear extrapolation between successive ^{14}C ages, since that assumes linear sedimentation rates, which are unlikely in this context. Some of the sedimentary units in the Adriatic successions are laterally discontinuous or have accumulated very slowly, so that the temporal resolution of succes-

sive units is highly variable (Trincardi et al., 1996). Linear extrapolation between radiocarbon ages may therefore result in spurious differences in age between what are effectively time-parallel markers (e.g. the biozone marker horizons) in the various marine basins. The following is a summary of the main foraminiferal marker events that were used to establish correlations between the marine sequences.

(1) A strong increase of *Globigerinoides ex gr. ruber*, the near absence of *Globorotalia scitula*, an increase of *Globorotalia inflata* and the appearance of *Globorotalia truncatulinoides* all coincide at ca. 13.0 ¹⁴C kyr B.P. (ca. 14.5 cal kyr). The equivalent event in the central Adriatic record is reflected in an abrupt increase of *G. ex gr. ruber* only; *G. inflata* and *G. truncatulinoides* were absent in the MAD mainly because the shallow depth of the basin prevented the development of the life cycle of these deep dwelling species.

(2) A distinct peak in *Globigerina bulloides* has been dated to 10.74 ¹⁴C kyr B.P. (11.8 cal kyr B.P.) using planktic foraminifers obtained from core CM92-43 (Asioli, 1996; Langone et al., 1996). An equivalent event in a core sequence from the Tyrrhenian Sea (core ET91-18 in Capotondi, 1995) has been dated to 10.63 ¹⁴C kyr B.P. (AMS dating on planktic foraminifers). This *G. bulloides* peak occurs during the YD (GS-1) cold episode.

(3) A conspicuous shift in the *Globigerina bulloides* $\delta^{18}\text{O}$ curve towards lighter values is dated to 10.5 ¹⁴C kyr B.P. (ca. 11.5 cal kyr B.P.; Langone et al., 1996) and post-dates the upper boundary of the YD (GS-1) event. This feature is considered to correspond to the onset of Termination IB.

(4) A strong decrease of *Neogloboquadrina pachyderma* and *Globorotalia inflata*, accompanied by a concurrent increase of *Globorotalia truncatulinoides* in the Tyrrhenian Sea, is a distinctive mid-Holocene feature (core AC85-4 in Capotondi et al., 1989; core BS78-13 in Benvenuti, 1989; core ET91-18 in Capotondi et al., 1999). This bioevent marks the last occurrence of *N. pachyderma* and *G. inflata* in the Adriatic Sea (Jorissen et al., 1993; Asioli, 1996; Capotondi et al., 1999; Tamburini et al., 1998). An increase of *Globigerinoides sacculifer* (sensu Hemleben et al., 1989) follows this bioevent in both the Adriatic and Tyrrhenian Seas.

This shift in dominance within the planktic foraminiferal assemblages has been dated to ca. 58 ¹⁴C kyr B.P. (ca. 6.0 cal kyr) and corresponds to the time when the modern sea level high-stand was attained (Trincardi et al., 1996). This transition also corresponds to an episode of attenuated sedimentation which marks the time that the sea reached its maximum stand during the Holocene. The condensed sediment units which formed at this time provide a distinctive stratigraphic marker which can easily be traced between the shallow, marginal successions of the Adriatic. The transition corresponds to the boundary between ecozones 2 and 3 of Capotondi et al. (1999), the age of which they suggest to be ca. 4.0 ¹⁴C kyr B.P. in the Tyrrhenian and 5.8 ¹⁴C kyr B.P. in the Adriatic. In our opinion, this age discrepancy is an artefact introduced by lack of adequate ¹⁴C control and by an assumption of constant sediment accumulation rates between dated horizons. Since the dates correspond to a time of condensed deposition, the actual time represented may vary markedly between the two marine basins (Loutit et al., 1988). In the light of this, linear extrapolation between ¹⁴C ages is inadvisable. This transition is also characterised by a sharp increase in $\delta^{13}\text{C}$ following a period of depleted $\delta^{13}\text{C}$ levels recorded throughout the Mediterranean basins between ca. 9.0 and 5.0 ¹⁴C kyr B.P. (see Vergnaud-Grazzini and Pierre, 1992; Pujol and Vergnaud-Grazzini, 1989; Vergnaud-Grazzini et al., 1989). This feature has also been detected in the *Globigerina bulloides* $\delta^{13}\text{C}$ data obtained from both the MC82-12 and CM92-43 cores.

The chronostratigraphy of the central Adriatic Sea is based on 40 AMS ¹⁴C dates obtained from planktic foraminifers from several core sequences which span the last termination and the Holocene (Langone et al., 1996; Trincardi et al., 1996). The dates were corrected for the apparent ¹⁴C age of surface sea water (reservoir age), estimate to about 570 years in the Adriatic Basin (Langone et al., 1996) and calibrated using the Stuiver and Reimer (1993) CALIB 3.0 program. Independent tests of the geochronology of the central Adriatic sequences is provided by tephrochronology and the identification of distinct features in oxygen isotope stratigraphic records (Trincardi et al.,

1994, 1996; Chondrogianni et al., 1996b; Calanchi et al., 1998).

The chronology of the crater lake sequences is based on the combined results of AMS radiocarbon dating of macrofossils, argon/argon dating, tephrochronology, varve counting and palaeomagnetic correlations (for further details see Oldfield, 1996).

5. Palaeoenvironmental reconstructions

The full sequence of stratigraphical changes recorded in the four sites is not discussed in this paper. Details can be found in papers referenced in this section but especially in Oldfield and Guilizzoni (1996) where the results of the PALICLAS project are reported. Here we focus on environmental conditions and developments in both the Mediterranean Sea and on land at around the time of deposition of sapropel S1 (9.8 to 6.8 cal kyr B.P.). However, the marine foraminiferal stratigraphic record is examined for a longer period of time because it provides essential context for linking the new evidence to previously published accounts of sapropel S1.

5.1. Adriatic and Tyrrhenian records

The stratigraphical succession of the two marine cores is summarised by considering developments in six intervals (Figs. 3 and 4), recognised on the basis of major changes in planktic foraminiferal assemblages, in oxygen and carbon stable isotope ratios and in pollen assemblages. The likely habitats indicated by the main species of planktic foraminifers used in our interpretations are based on the interpretations of Pujol and Vergnaud-Grazzini (1995) and references therein. Note that *Globigerinoides* ex gr. *ruber* comprises *Globigerinoides ruber* (pink and alba varieties), *Globigerinoides gomitulus* and *Globigerinoides elongatus*, while *Globigerinoides sacculifer* includes *Globigerinoides trilobus*, *Globigerinoides sacculifer* and *Globigerinoides quadrilobatus* (Hemleben et al., 1989). The chronology is expressed in calendar years.

Interval 6 (before 14.5 cal kyr B.P.). The plank-

tic assemblage is dominated by *Globorotalia scitula*, *Neogloboquadrina pachyderma*, *Globigerina bulloides* and *Globigerina quinqueloba* while ‘warm’ species, such as *Globigerinoides* ex gr. *ruber* and *Orbulinai*, are virtually absent. During this interval, cold and productive waters characterised both the Adriatic and Tyrrhenian Seas.

Interval 5 (14.5–12.8 cal kyr B.P.). At the beginning of this phase an abrupt warming of the surface waters is recorded (increase of *Globigerinoides* ex gr. *ruber* per cent), followed by gradual cooling which is well-defined in the record from CM92-43 (Asioli, 1996; Asioli et al., 1997). This cooling trend is also evident in other Tyrrhenian and Adriatic cores (IN68-9 in Rohling et al., 1997; ET91-18 and IN68-5 in Capotondi et al., 1999), although it is not evident in the record from core MC 82-12 because of the lower resolution of that sequence. The initial abrupt warming (between 15.0 and 14.6 cal kyr B.P.) is equivalent in age to zone GI-1e of the GRIP ice-core record (Björck et al., 1998), while the cooling trend between ca. 14.6 and 13.0 cal kyr B.P. equates approximately with zones GI-1d to GI-1a in GRIP. This cooling trend in the Mediterranean appears to have been characterised by three distinctive steps (Asioli et al., 1998), which bears a striking resemblance to the sequence of events in the GRIP record between zones GI-1d to GI-1a. During the second of these steps, an episode of fresh water influx is inferred from the $\delta^{18}\text{O}$ composition of *Globigerina bulloides* in core CM92-43. This peak is contemporaneous with the mwp-1A record from the Atlantic (sensu Fairbanks, 1989 and Clark et al., 1996). This episode of freshwater influx into the Mediterranean was first recognised by Vergnaud-Grazzini and Pierre (1992) in cores from the southern Adriatic. This event is not so well-defined in the isotopic data from the Tyrrhenian Sea. The foraminiferal assemblage corresponding to this event is dominated by *Globigerina quinqueloba* and *Globigerinoides* ex gr. *ruber* in the central Adriatic and by *Globigerina inflata*, *Neogloboquadrina pachyderma* and *G.* ex gr. *ruber* in the southern Adriatic (core IN68-5, winter assemblage in the western Mediterranean, apparently requiring a well-mixed and cool-mixed layer (Pujol and Vergnaud-Grazzini, 1995).

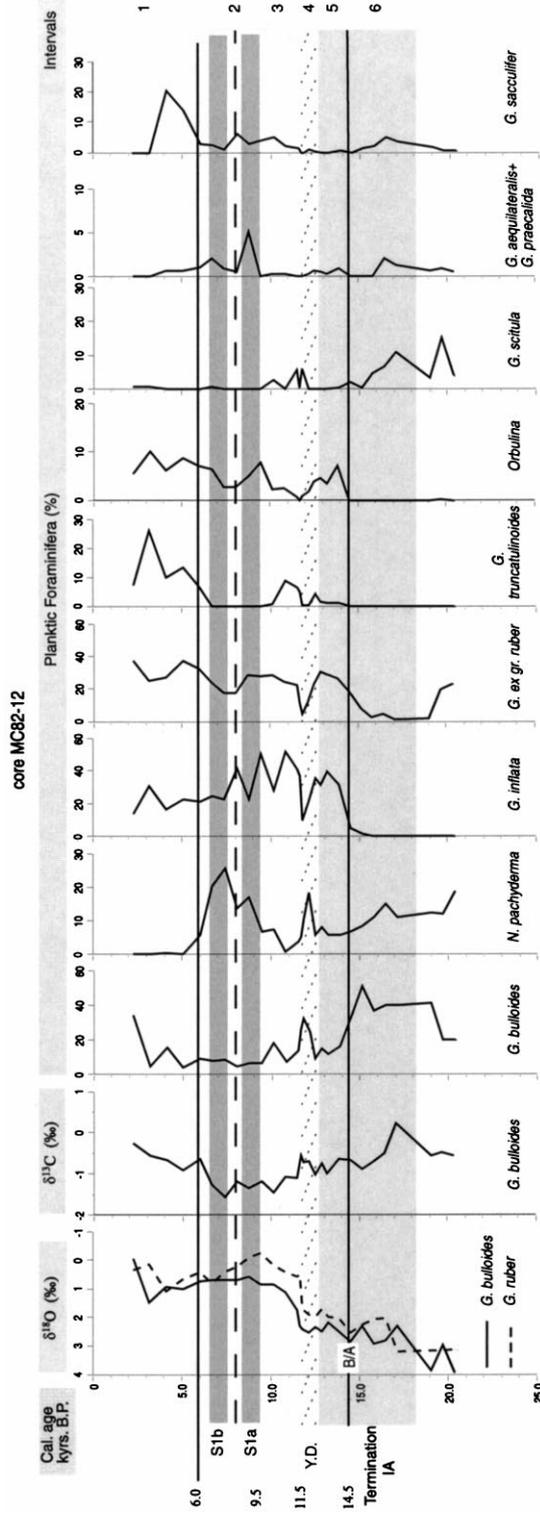


Fig. 4. Planktic foraminifer assemblages and oxygen and carbon isotope compositions in *Globigerina bulloides* and *Globigerinoides ruber* in core MC82-12. Ages are expressed in calendar years after correlation to bioevents dated in core CM92-43. Although the temporal resolution in this core is not very high, a break in the deposition of the episode that is equivalent to sapropel S1 is well-defined as a peak of *Globigerina inflata* and *Globigerinoides sacculifer*. These peaks coincide with minimum percentages of *Globigerinella aequilateralis* and *Orbulina*.

Interval 4 (12.8–11.5 cal kyr B.P.). Subtropical species disappear in the Adriatic (core CM92-43) or abruptly decrease to a very low frequency in the Tyrrhenian (core MC82-12). These species are replaced by *Globigerina bulloides*, *Globigerina quinqueloba* and *Neogloboquadrina pachyderma*. The interval is also characterised by a peak of *Artemisia* and Chenopodiaceae pollen (Lowe et al., 1996a,b), a positive peak in $\delta^{13}\text{C}$ and maximal TOC fluxes. All these indicators point to a cold episode characterised by relatively arid conditions on land and cold and productive ocean waters. This interval corresponds to the cold YD chronozone (GRIP zone GS-1). During this phase, the oxygen isotopic composition of *G. bulloides* changed to more positive values.

Interval 3 (11.5–9.5 cal kyr B.P.). The lower part of this interval is characterised by either the reappearance of (core CM92-43), or substantial increase in (core MC82-12), *Globigerinoides* ex gr. *ruber*, together with high abundance of the herbivorous species *Globigerina inflata* and *Globorotalia truncatulinoides*, while *Neogloboquadrina pachyderma* is almost absent. The data for this interval indicate phytoplankton blooms, strong vertical mixing of the water column during the winter and stratification during summer; the same event is reflected in the records from the southern Adriatic and other Tyrrhenian sequences (Benvenuti, 1989; Rohling et al., 1997; Capotondi et al., 1999). A concomitant rapid increase in arboreal pollen percentages during this interval indicates that steppe genera were replaced by forests of *Quercus* and other deciduous trees (Lowe et al., 1996a,b), while the $\delta^{18}\text{O}$ values become more negative. These developments coincide with evidence for enhanced erosion on the shelf after the termination of the YD, and by reduced deposition in the deeper basins (Cattaneo and Trincardi, 1999).

Interval 2 (9.5–6.0 cal kyr B.P.). During this interval, which corresponds approximately to the time of deposition of sapropel S1, further diagnostic changes in the foraminiferal assemblages are recorded. In the central Adriatic (core CM92-43) as well as the southern Adriatic (Rohling et al., 1997), *Globorotalia truncatulinoides* is absent, while there are renewed increases in *Neogloboquadrina pachyderma* and *Globigerina inflata* in two phases:

one between 8.0 and 7.5 cal kyr B.P. and another between 7.0 and 6.0 cal kyr B.P. A similar fluctuation is observed in the $\delta^{13}\text{C}$ values of *Globigerina bulloides* over this time period. This episode of generally-depleted $\delta^{13}\text{C}$ values is interrupted by a short-lived phase of relative enrichment lasting only a few hundred years and corresponding to a temporary re-increase in *G. inflata*. In the benthic assemblage of the core obtained from the MAD, some species indicative of organic matter enrichment and/or oxygen depletion, such as *Uvigerina peregrina* and *Uvigerina mediterranea* (Lutze and Coulbourn, 1984), increase during the time interval of S1 deposition, while *Hyalinea balthica* is also quite abundant.

High-resolution seismic profiles indicate an increased supply of muddy sediment to the shelf between ca. 9.5 and 6.0 cal kyr B.P. (Fig. 5; Cattaneo and Trincardi, 1999). A thickness of as much as 15 m of mud accumulated in shore-parallel depocentres during this interval, with an average sediment accumulation rate exceeding 4 mm/yr (Cattaneo and Trincardi, 1999). It was during this interval that sapropel S1 was deposited in the Levantine Basin as two phases of organic-rich sediment, one from 9.5 to 8.2 cal kyr B.P. (S1a) and one from 7.8 to 7.0 cal kyr B.P. (S1b). In the Adriatic (MAD) core, the increased TOC flux values and the lowest magnetic susceptibility measure (X in Fig. 3) are associated with the earliest of these two phases. Magnetite dissolution related to reductive diagenesis is typical of sapropel layers (e.g. van Santvoort et al., 1997) and has already been recognised in the Adriatic sediments by Oldfield (1996).

There is no clear lithological equivalent in the Tyrrhenian Sea record (in the form of a black, sapropelitic clay) to the S1 sapropel found elsewhere in the Mediterranean. However, in the part of the Tyrrhenian record that is of equivalent age, there are changes in faunal, isotopic and palaeomagnetic stratigraphy similar to those typically associated with S1 layers. The strongest similarities are evident in the foraminiferal stratigraphies. *Globorotalia truncatulinoides* and *Globorotalia scitula* disappear from both the Tyrrhenian and the Adriatic records from ca. 10.0 cal kyr B.P. This is followed by peaks in *Orbulina*, *Globigerinella aequi-*

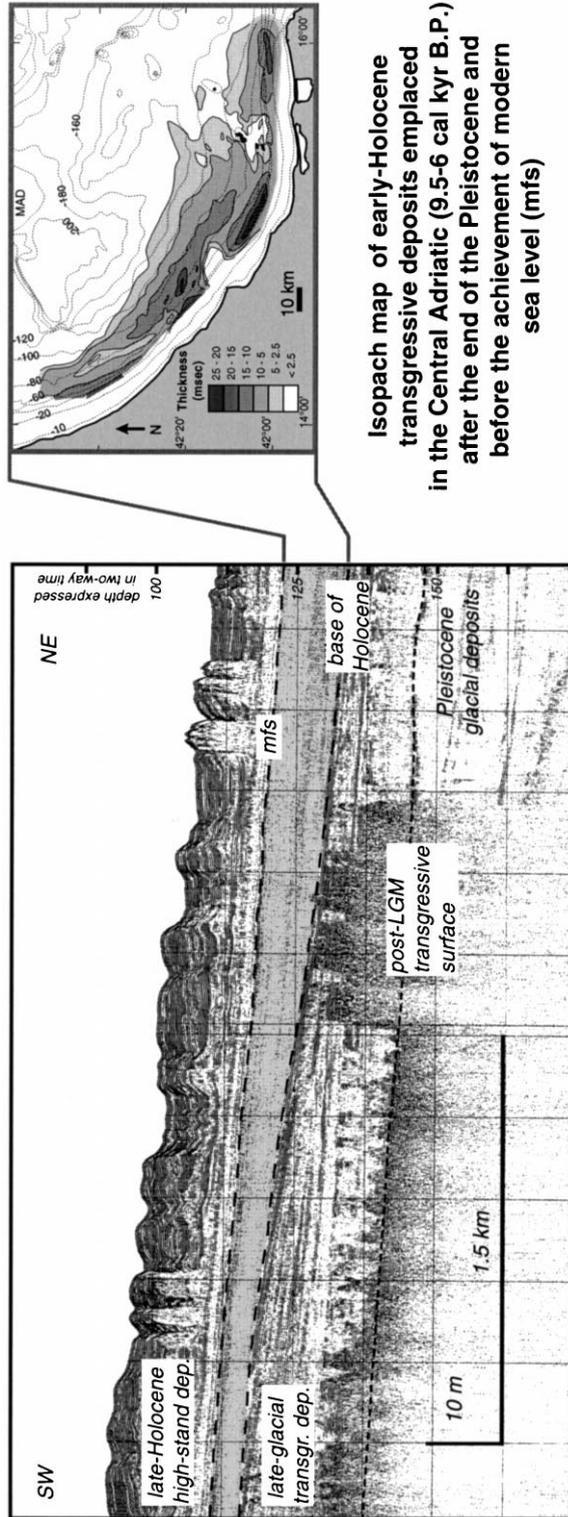


Fig. 5. High-resolution 3.5 kHz seismic profile in the MAD (Adriatic Sea) showing the sedimentary body deposited between 9.5 and 6.0 cal kyr B.P. This unit (grey-shaded) represents the upper unit of the Late Quaternary transgressive system tract (TST) as defined by Trincardi et al. (1996) and Cattaneo and Trincardi (1999). The isopach map of this unit shows a distribution of elongated muddy sediments parallel to other shoreline with thickness that locally exceeds 15 m.

lateralis and *Globigerina praecalida*, which coincide with a marked decrease in *Globigerina inflata*. This biostratigraphical sequence is in accordance with that recorded for the sapropel S1 interval in the Ionian Sea (Lander Rasmussen, 1991), in the Levantine Basin (Rohling et al., 1993) and in time-equivalent deposits in the Corsica Channel (Benvenuti, 1989; Capotondi, 1995). Indeed, a significant drop in percentages of *G. inflata* appears to be diagnostic of S1 sapropel layers (Williams and Thunell, 1979). Following these developments, *G. inflata* re-expands, while *Globigerinoides ex gr. ruber*, *Orbulina*, *G. aequilateralis* and *G. praecalida* all decrease. *G. truncatulinoides* is also recorded in higher percentages in other Tyrrhenian sites over the same time period (core MC82-2 in Tamburini et al., 1998; and core ET91-18 in Capotondi, 1995). This short-lived episode coincides with the reoccurrence of *G. inflata* in the Adriatic Basin between 8.0 and 7.5 cal kyr B.P., which Rohling et al. (1997) equated with a break in sapropel deposition. Since *G. inflata* and *G. truncatulinoides* prefer a cool and well-mixed layer with intermediate to high nutrient levels (Pujol and Vergnaud-Grazzini, 1995), the data may reflect a short-lived episode of cooling and/or increased contrast in seasonal conditions with vertical mixing during the winter. After this event (from ca. 7.5 cal kyr B.P.), *G. aequilateralis*, *G. praecalida* and *Orbulina* increase once again, accompanied by the strong representation of *Neogloboquadrina pachyderma* during interval S1b. In the Tyrrhenian Sea records, the final peak in *G. inflata* is dated to ca. 6.0 cal kyr B.P. An equivalent feature may occur in core MC82-12, although it is not as clear. Here, the final peak in *G. inflata* precedes the disappearance of *N. pachyderma* and the reoccurrence of *G. truncatulinoides*. We equate this bioevent with the final *G. inflata* peak in the MAD records, which is dated to 6.0–7.0 cal kyr B.P.

Interval 1 (6.0 cal kyr B.P.–present): *Globigerina inflata* and *Neogloboquadrina pachyderma* disappeared in the Adriatic Basin during the latter part of the Holocene and decreased to very low frequency in Tyrrhenian Sea assemblages. *Globorotalia truncatulinoides*, which disappeared in the Adriatic basin at ca. 10.0–9.8 cal kyr B.P., persisted into the late Holocene in the Tyrrhenian

Basin. Increased frequencies of *Globigerinoides sacculifer* and *Globigerinoides ex gr. ruber* characterised the late Holocene assemblages in both seas. These changes indicate that, during the late Holocene, the surface waters at both locations were dominated by an oligotrophic mixed layer for most of the year.

5.2. Crater lake records

Core PALB94-3A (120 m water depth; see Table 1) from Lake Albano contains a continuous Holocene record with a clear glacial–Holocene boundary dated to 11.48 cal kyr B.P. (Oldfield, 1996; Chondrogianni et al., 1996b). The Holocene sediments are laminated on a millimetre scale with the exception of the middle Holocene part of the sequence which is characterised by more frequent diatom blooms and higher sedimentation rates. Although high productivity levels characterise the entire Holocene sequence as estimated by reconstructed trophic status (Ryves et al., 1996), the early Holocene deposits are particularly organic-rich, as shown by high TOC as well as by relatively high HI values (Fig. 6). The results of investigations of a number of biological indicators, including pigment concentrations and algal remains, indicate that these variations more probably reflect changes in lake productivity (Ryves et al., 1996; Ariztegui et al., 1996c). During the period in which sapropel S1 developed in the Mediterranean, TOC values were relatively stable (average 1.6%) and the laminated nature of the lake deposits suggests that hypolimnetic anoxia developed in relatively deep waters. Between 8.1 and 7.6 cal kyr B.P., however, there is a decrease in TOC fluxes and in HI values, a substantial decrease in pigment concentrations, a marked decrease in isorenieratene/TOC ratios and dominance of diatoms over other types of algae (Ryves et al., 1996; Ariztegui et al., 1996c), all of which suggests a short-lived episode of reduced anoxia contemporaneous with a decline in primary productivity and/or preservation. In addition, Manca et al. (1996) reported a drop in concentration and diversity of chydorids (type of chironomids) during this short interval to values characteristic of the Würm late-glacial period, suggesting that a short-lived cold event affected the

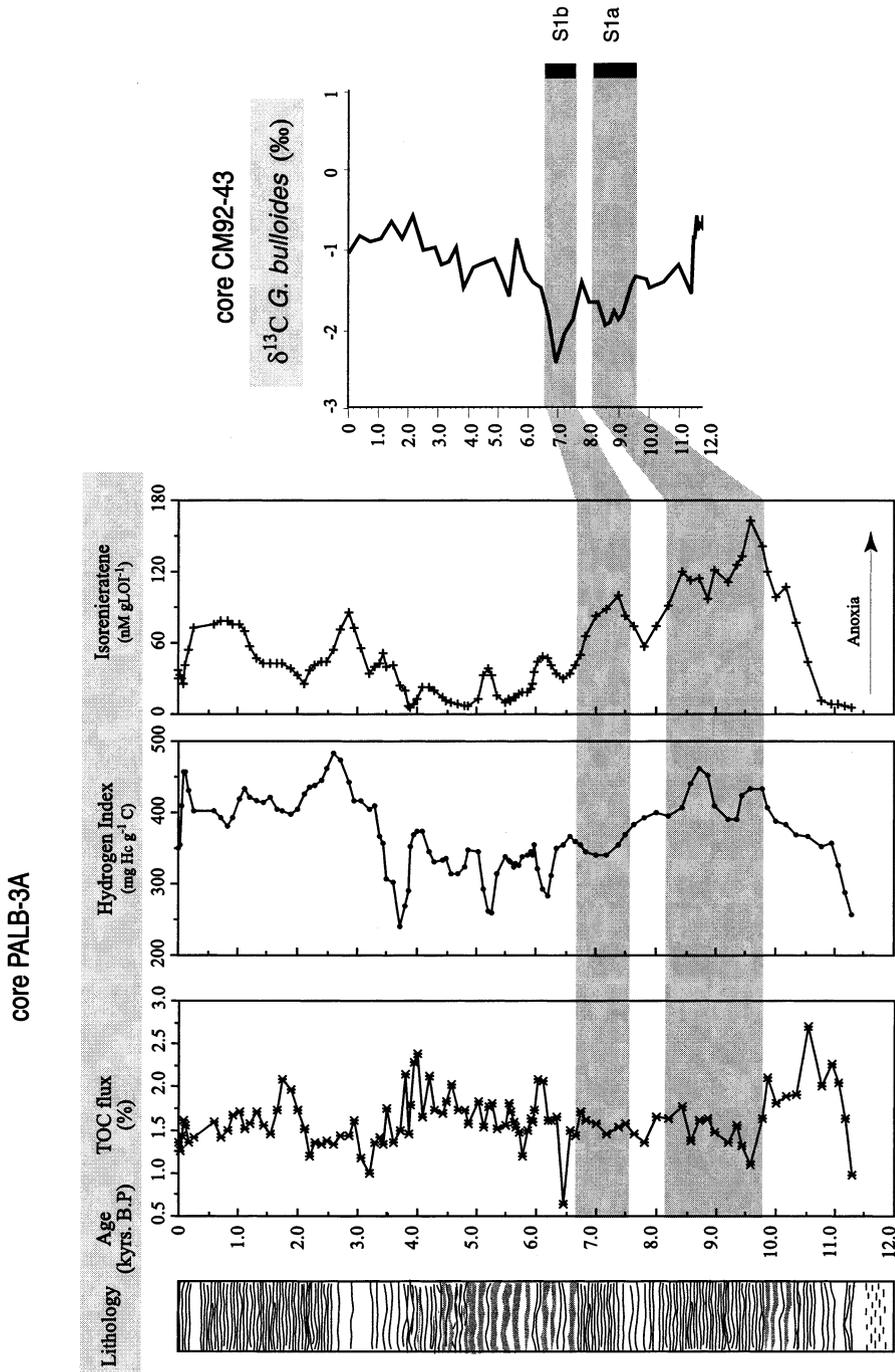


Fig. 6. Core PALB94-3A from Lake Albano and correlation with the $\delta^{13}\text{C}$ record on *Globigerina bulloides* from marine core CM92-43. The age scale in both cores is expressed in calendar years B.P. TOC flux and isorenieratene concentrations in the lake sediments show a shift to lower values contemporaneous with the observed interruption of S1 deposition in the Mediterranean.

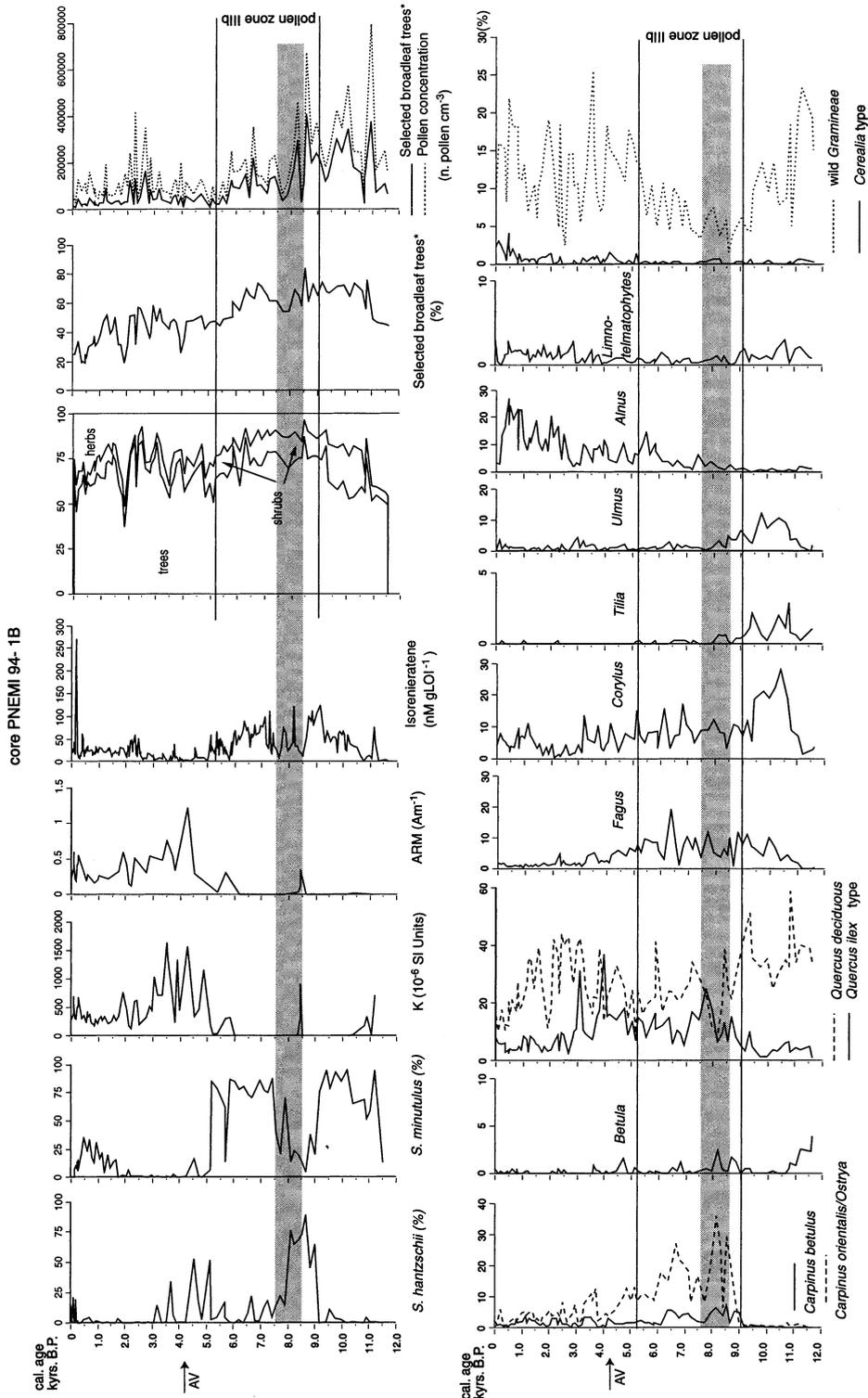


Fig. 7. Pollen, magnetic and pigment data from core PNEMI 94-1B from Lake Nemi (see Table 1). Notice the changes observed in the different data during the time of interruption of SI deposition. This interval is particularly well-defined by minima in percentages of selected broadleaf trees. Although highly variable, the conception of isorenieratene displays a minimum contemporaneous with this interruption of sapropel in the Mediterranean. See text for complete discussion.

* *Carpinus betulus*, *Carpinus orientalis/Ostrya*, *Betula*, *Fagus*, *Corylus*, *Quercus deciduous*, *Tilia* and *Ulmus*

region during the mid-Holocene. This point is returned to below.

Comparable evidence can be found in the biostratigraphical records obtained from Lake Nemi. Core PNemi-1B was retrieved from the deepest part of Lake Nemi, in a water depth of 30 m (Fig. 7 and Table 1). The sediment sequence at this locality consists of three distinct lithological units, all of which contain diatom-rich beds and laminations intercalated with massive muds (Chondrogianni et al., 1996a). A well-developed set of laminations characterises the early Holocene part of the sequence, with a diatom flora dominated by *Stephanodiscus minutulus* (Fig. 7), a taxon typical of high lacustrine primary productivity (Ryves et al., 1996). As in the Lake Albano record, the interval contemporaneous with the deposition of sapropel S1 is characterised by high values of isorenieratene and a marked decline in magnetic content associated with a larger magnetic grain size, strongly suggestive of reductive diagenesis (Alvisi and Vigliotti, 1996; Vigliotti et al., 1999). Between 8.3 and 8.5 cal kyr B.P., however, a significant amount of magnetic content is reflected in high values of concentration-dependent parameters such as susceptibility (K), anhysteretic and isothermal remanence (ARM, SIRM) (Fig. 7). We interpret this increased content of ferrimagnetic minerals as related to environmental changes leading to the interruption of dominant anoxic conditions in the water column and the consequent preservation of primary ferrimagnetic minerals within the sediment. The coincidence with variations in biological proxies (diatom assemblages) and pigment content (isorenieratene) support this interpretation.

Pollen-stratigraphic evidence from Lakes Albano and Nemi also suggest that there were significant changes in terrestrial vegetation at around the time of formation of sapropel S1. The most detailed of the two palaeobotanical records for the Holocene is that from Lake Nemi (Fig. 7). This record displays an increasing abundance of *Quercus deciduous* and *Corylus* contemporaneous with the beginning of S1 deposition in the Mediterranean (ca. 10.0 cal kyr B.P.). Both taxa suggest warm and wet conditions. A clear decrease

in pollen abundance of a number of broadleaf trees can be observed for the period ca. 8.5–7.5 cal kyr B.P. Taxa that were already well-represented in the pollen assemblages prior to 8.5 cal kyr B.P., such as deciduous *Quercus*, *Ulmus*, *Tilia*, *Fagus* and *Corylus*, show a sharp decline, while there are also reduced representations of those taxa that had only begun to expand in the region at around that time (e.g. *Carpinus orientalis/Ostrya*, and *Carpinus betulus*). There is some pollen-stratigraphical evidence that suggests anthropogenic activities in the area at the same time (e.g. pollen of Cerealia type, such as the *Hordeum* group and the *Avena-Triticum* group). The pollen evidence as a whole, however, indicates cooling of climate associated with a decrease in regional precipitation as the most probable natural cause of the simultaneous demise of such a wide range of arboreal taxa between ca. 8.5 and 7.5 cal kyr B.P. (Lowe et al., 1996a). Additional evidence supporting this interpretation is an increase in the representation of *Alnus* pollen, whereas percentages of pollen of aquatic and semi-aquatic taxa decrease, probably reflecting lower water levels and the spread of alder trees onto exposed marginal lake deposits. A significant decrease in pollen of the holly oak (*Quercus ilex* type) and a re-increase in pollen of *Betula* also support the notion of climatic cooling after ca. 8.1 cal kyr B.P. A similar decrease in pollen percentages and in concentrations of some of the broadleaf trees is also evident in the pollen diagram from the Lake Albano sequence at about 8.5 cal kyr B.P. (Lowe et al., 1996a).

5.3. Subdivision of sapropel S1: evidence from other records

The marine and the crater lake sequences discussed in the preceding section show high levels of accumulation of organic-rich sediments during the period ca. 9.0 to 6.8 cal kyr B.P., equivalent to the time of formation of sapropel S1 in the Mediterranean region. There is also evidence for a short-lived period of reduced organic matter deposition (i.e. relatively low productivity and/or preservation), dated to ca. 8.0 to 7.5 cal kyr B.P.

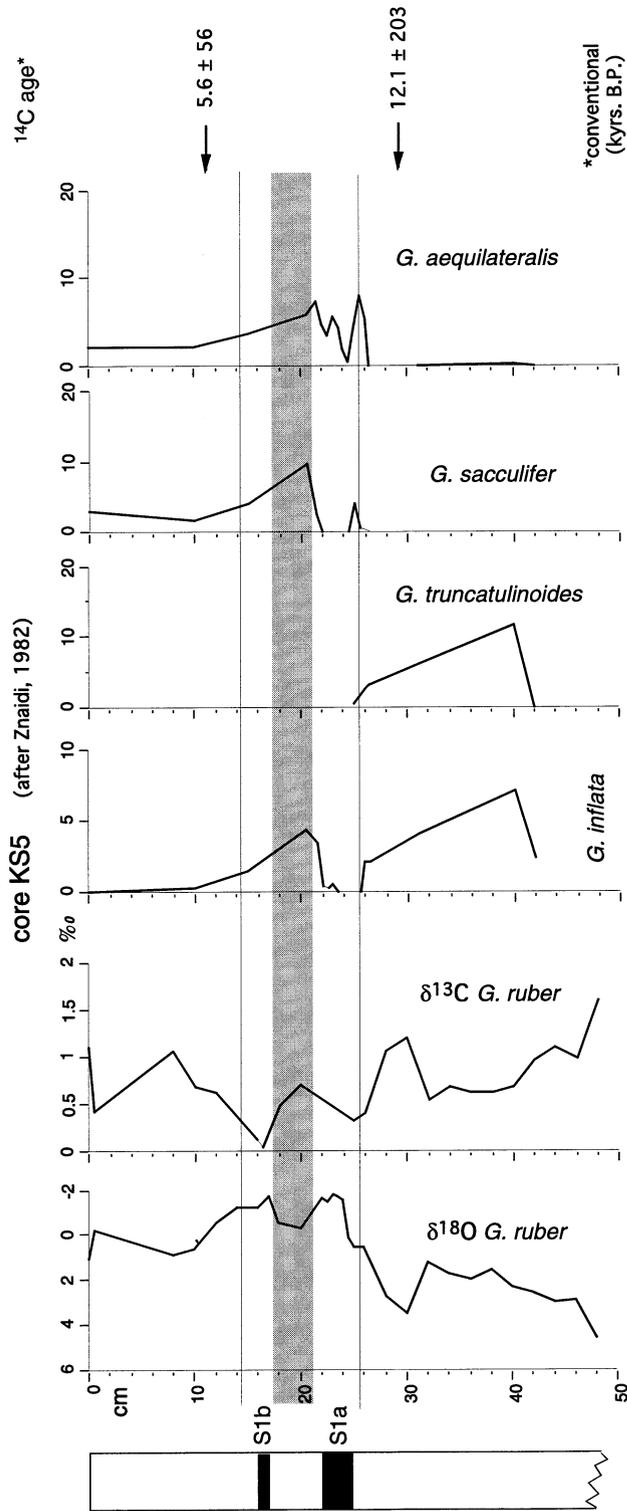


Fig. 8. Core KS5 after Znaidi (unpublished Ph.D. Thesis, 1992) in the eastern Mediterranean showing clear evidence of an interruption during the deposition of sapropel S1; this break is reflected as a prominent change in lithology indicative of oxidised sediments, and relative peaks in *Globobulimina inflata* and *Globobulimoides sacculifer* accompanied by shifts to more positive δ¹⁸O and δ¹³C values on *Globobulimoides ruber*.

This is considered to reflect regional climatic cooling associated with reduced levels of precipitation that interrupted a period of generally higher precipitation. A similar subdivision of the sapropel S1 phase can be detected in a number of published stratigraphical records from sites located in various parts of the Mediterranean Basin.

A division of S1 into two phases separated by a short interval with increased frequencies of *Globigerinoides sacculifer* can be detected in cores from the southern Adriatic (IN68-38, Jorissen et al., 1993; AD91-17, Capotondi et al., 1999; IN68-9, Rohling et al., 1997) and the eastern Mediterranean (Znaidi Rivault, 1982). Znaidi Rivault (1982) interpreted the short-term increase in *G. sacculifer* in the middle of S1 (e.g. in Core KS5, Fig. 8) as evidence of reworking of older fossils. However, the fact that a contemporaneous increase in *G. sacculifer* is found in the eastern Mediterranean, the southern Adriatic and, from our studies, the Meso-Adriatic Depression (e.g. Core CM92-43, Fig. 3), supports the idea that this is a regionally coherent signal, and that it reflects a widespread change in palaeoceanographic conditions.

The stable isotope stratigraphy of several records from the Mediterranean also shows a subdivision of the S1 event. Fluctuations in $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ compositions in the southern Adriatic (Core KS-5) as well as in the Ionian Basin (RC-191, KET82-16 and KET82-22) occur in the middle of the S1 layer (Fontugne et al., 1989; Siani et al., 1997). This is reflected particularly by increased $\delta^{18}\text{O}$ values of *Globigerinoides ruber*. At about the same time (8.0–7.5 ^{14}C kyr B.P.) there is a shift towards more negative $\delta^{13}\text{C}$ values in accumulated organic matter (Fontugne et al., 1989), which is ascribed to a decrease in the influx of terrestrial organic material, and which could be explained by reduced precipitation levels in adjacent continental catchments.

Terrestrial records also display a short-lived event at around 8.0 to 7.5 cal kyr B.P. Pollen-stratigraphic records from marine cores and continental sites in the eastern Mediterranean region show a characteristic *Pistacia* phase which reflects marked climatic warming in the region between ca. 9.0 and 6.0 ^{14}C kyr B.P., when sapropel S1

was deposited (Rossignol-Strick, 1995). In several records, however, such as those from sites in Greece and Syria (e.g. Wijnstra, 1969; Niklewski and Van Zeist, 1970; Bottema, 1974), a notable reduction in *Pistacia* and *Quercus* percentages, associated with an increase in *Artemisia* pollen, occurs in the middle of this phase.

There is therefore widespread evidence in both the marine and continental records for significant palaeoenvironmental changes during the S1 interval, and the period is thus subdivided into an early (S1a) and later phase (S1b), these being separated by a short-lived episode which dates to around 8.0 to 7.5 cal kyr B.P.

6. Discussion

Several key issues emerge from the new evidence and discussions presented here.

(1) A peak in freshwater influx at about 13.5 cal kyr B.P. is revealed in the MAD and, although much less pronounced, in the southern Adriatic as well (Vergnaud-Grazzini and Pierre, 1992). The signal appears to be magnified in the central Adriatic under the influence of local factors, such as the reduced size and volume of the Adriatic Basin, the limited connection to and water exchange with the rest of the Mediterranean, and increased supply by rivers, especially the Po. The isotopic signal of this peak in freshwater discharge is best developed in the $\delta^{18}\text{O}$ data obtained from *Globigerina bulloides*. This species dwells in deeper waters than *Globigerinoides ruber*. Isotopic data from the latter show no significant variations during this interval. This evidence suggests that fresh waters entered the Adriatic Basin as hyperpycnal flows; increased sediment load entered the central Adriatic at that time (Trincardi et al., 1996; Cattaneo and Trincardi, 1999), and may have triggered density flows. Both the influx of freshwater and the increased sediment loads may be explained by melting of glaciers in the Alps and Apennines at the last termination (Maisch, 1992). A high freshwater influx may also explain the records of *Sparganium* (a freshwater macrophyte — see Fig. 3) encountered in late-

glacial sediments in several core sequences from the MAD (Lowe et al., 1996a,b).

(2) Melt waters during the late-glacial interval may have induced water stratification but did not lead to the conditions favourable for sapropel formation. The faunal records for this interval of freshwater discharge in the southern Adriatic show a peak of *Globigerina inflata* that indicates strong vertical mixing during the winter season. A similar signal is also found in the Tyrrhenian Basin record (Fig. 4). Fresh water discharge appears therefore to have been seasonal and not to have persisted throughout the year.

(3) Conditions favourable for sapropel formation occur during the early Holocene, from ca. 9.5 cal kyr B.P. onwards. Clearly, some threshold critical for the accumulation and preservation of organic detritus was crossed at that time. The contemporaneous disappearance of *Globigerina inflata* from the Adriatic and from the eastern Mediterranean is likely to have been in response to a lack of mixing of the water column, with year-round stratification. This could have been induced by increased discharges by rivers in response to wetter climatic conditions in the region. The absence of *Neogloboquadrina pachyderma* from the Adriatic and Ionian records indicates that a deep chlorophyll maximum (DCM) was not able to develop (cf. Rohling and Gieskes, 1989). A strong density stratification must therefore have characterised the water column during this interval. However, *N. pachyderma* and *G. inflata* are recorded throughout the S1 interval in the Tyrrhenian sequences, although with significant fluctuations in frequency. Therefore, the Tyrrhenian Basin appears to have been characterised by oceanographic conditions somewhat different from those of the eastern Mediterranean during the S1 interval: winter mixing was probably active and DCM formed on a seasonal basis. Moreover, the isotopic data obtained from planktic foraminifers in Tyrrhenian core MC82-12 record an excursion to more depleted $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ compositions, suggesting an increase of fresh-water input. In summary, therefore, the records from the two basins suggest water stratification and reduced winter convection at the time of S1 formation, but the differences in faunal assem-

blages indicate different palaeoceanographic conditions. In the Tyrrhenian Sea, in particular, conditions did not favour the formation of dark organic-rich sediment, perhaps because of prolonged ventilation of bottom waters. This view accords with that of Kallel et al. (1997a), who regard *N. pachyderma* as an indicator of cold and deep water and concluded that deep water formed in the Tyrrhenian Sea during the sapropel interval.

(4) The relative importance of *Globigerina inflata* in the faunal assemblages is not only indicative of the vertical structure of the water mass, but also allows us to make inferences about the horizontal exchange of water masses throughout the Mediterranean. According to Pujol and Vergnaud-Grazzini (1995), the modern distribution of *G. inflata* in the Mediterranean Sea coincides with both the path of the modified Atlantic waters (MAW), entering the Alboran Basin (and generating cyclonic gyres in the Balearic and Tyrrhenian Seas), and the westward path of Levantine intermediate waters. Therefore, the pathway of MAW to the easternmost part of the Sicilian Strait in winter may be inferred from an assemblage dominated by *G. inflata*. The presence of this species in the Tyrrhenian Sea during the S1 interval may therefore indicate the inflow of Atlantic waters. Any such inflow was probably quite weak, however, as *G. inflata* shows very low frequencies in the Alboran Basin (Pujol and Vergnaud-Grazzini, 1989; Rohling et al., 1995). The absence of *G. inflata* in the eastern Mediterranean suggests a limited exchange between the western and eastern Mediterranean during S1, perhaps reflecting light surficial water in the eastern basin. However, the abundance of species such as the *Globigerina calida-praecalida*, *Globigerina digitata-praedigitata*, and the *Globigerinella aequilateralis* group (spinose and carnivorous species typical of the Levantine Basin, Parker, 1958; Thunell, 1978) during this interval, in both the Ionian and Tyrrhenian Basins, may indicate that intermediate water was active and probably flowing to the western Mediterranean, as it does in the modern circulation. Contrary to some published views (Sarmiento et al., 1988), therefore, we find no evidence for a reversal of marine circulation during the S1 interval.

(5) In common with other records from the Mediterranean Sea, we have found evidence for a subdivision of sapropel S1 into two sub-units (S1a and S1b) separated by a mid-S1 interruption, dated to ca. 8.0 to 7.5 cal kyr B.P. The short interruption in formation of S1 is marked by a very similar faunal turnover in both marine basins. In the foraminifer assemblage the reoccurrence of *Globigerina inflata* between 8.0 and 7.5 cal kyr B.P. suggests that vertical mixing during winter was re-established in the Adriatic and Levantine Basins. Moreover, the peak percentages recorded for this species during this interval may indicate an enhanced inflow of MAW; evidence for this increased inflow can be detected throughout almost the entire Mediterranean Basin. In the Alboran Sea a substantial increase of *G. inflata* between 7.0–8.0 ^{14}C kyr B.P. has been interpreted as the onset of more or less modern hydrographic conditions (Pujol and Vergnaud-Grazzini, 1989; Rohling et al., 1995). The constrained stratigraphic resolution in that area, however, does not allow secure correlation of this event with the base of the S1 interruption evident in our records or with the younger peak of *G. inflata* at 7.0 cal kyr B.P. The concurrent occurrence of *Neogloboquadrina pachyderma* may indicate some deep-water formation in the Adriatic region during this break. Moreover, the increase of *Globigerinoides sacculifer*, a symbiont bearing species dwelling in warm and oligotrophic mixed layer, during the S1 interruption implies that waters at that time must have been characterised by a scarcity or absence of turbidity. Similar evidence is recorded throughout the Mediterranean, which suggests that there was reduced fluvial inflow from the time of the S1 interruption.

(6) Several lines of evidence suggest a marked contrast between sub-units S1a and S1b. Benthic foraminiferal assemblages indicate that deeper waters remained cool and oxygenated in the MAD during S1a, while the southern Adriatic was azoic and not ventilated (Jorissen et al., 1993; Rohling et al., 1997). On the other hand, ventilation as well as a high degree of productivity in the water column appears to have characterised both basins during the deposition of S1b. The evidence from the MAD is of sufficient resolution to show that

Hyalinea balthica gradually disappeared during S1b while *Brizalina spathulata* was spreading into the basin. The former taxon indicates cold and relatively well-oxygenated bottom waters (Ross, 1984), whereas the high frequency of the latter taxon indicates low-oxygen waters and marked stratification throughout most of the year (Barmawidjaja et al., 1995a,b). Fluxes of TOC were greater during S1a and S1b. Furthermore, $\delta^{13}\text{C}$ of *Globigerinoides ruber* is in general more depleted in S1b than in S1a, probably due to a comparatively lower productivity during S1b. The planktic evidence discussed above also points to significant differences between the S1 sub-units.

(7) The termination of S1 must have resulted from a change in seasonality and an increased influx of Atlantic water into the Mediterranean, because it coincides with the reappearance of *Globigerina inflata*, which suggests that both the Adriatic and Levantine Basins became well ventilated again.

(8) The stratigraphic information obtained from the crater lake basins provides evidence for significant environmental changes during the time of S1 deposition in the marine basins. Relatively stable levels of productivity and preservation of organic matter characterise the Lakes Albano and Nemi records during the S1 interval, and here too there is evidence for a short-lived episode of declining productivity and/or organic matter preservation, which we equate with the Early–Middle Holocene transition (EMHT) (Stager and Mayewski, 1998).

7. Conclusions

The formation of sapropel S1 and variations in its development, both in a temporal (differentiation of S1a and S1b) and spatial (contrasts between western and eastern basins) context, are intricately linked to palaeoceanographic changes. The new evidence presented here shows that S1 formation and variations within the S1 interval also correlate with evidence in the terrestrial records for major environmental changes that affected the Italian mainland. The collective marine and terrestrial data indicate that both the marine and continental

realms were probably responding together to regional climatic changes, and the formation of S1 in the sea coincides with a significant increase in precipitation levels. This is inferred mainly from the evidence for the spread of broadleaf vegetation across central Italy, and is an interpretation that is supported by other pollen-stratigraphical studies of mid-Holocene successions in Italy and France (Rossignol-Strick and Planchais, 1989; Guiot et al., 1989). Such an increase in precipitation, we suggest, also explains the contemporaneous increase in sediment and nutrient supply, as well as the salinity changes, inferred from the marine records presented in this paper. The high resolution records from the MAD and from Lakes Albano and Nemi also suggest that this episode of warm and humid conditions was interrupted by an abrupt cooling event that lasted ca. 500 years. This event, which resulted in an interruption in the deposition of S1, is well-defined in marine and continental records in and around the western and eastern Mediterranean, and seems to be coeval with the EMHT cooling event reflected in records from the equator and from the poles, and attributed by Stager and Mayewski (1998) to marked changes in solar irradiance or in cosmic ray flux.

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