

# Northern Levantine and Adriatic Quaternary planktic foraminifera; Reconstruction of paleoenvironmental gradients

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## ABSTRACT

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A detailed study of the planktic foraminiferal records of three cores, recovered near Crete, shows very consistent changes. Unfortunately, the record of T87/2/13G, our shallowest core, is bioturbated too much for detailed comparisons, but it can still be used to discuss the general trends. The mutual record of the other two cores, T87/2/20G and T87/2/27G, extends back to about 125,000 yr B.P. Principal Components Analyses performed on the combined records of T87/2/20G and T87/2/27G yield two axes that can be interpreted paleoecologically, namely a temperature-axis and an “annual stability” axis. The latter is a measure of stratification within the euphotic layer.

In this study, we consider T87/2/20G and T87/2/27G to be reference cores for the eastern Mediterranean. Subsequently, we interpret the faunal variation in two cores from the Adriatic Sea (IN 68-5 and IN 68-9), containing records extending back to 15,000 and 18,000 yr B.P., in the context of these reference-cores. This is done by calculating scores of the Adriatic records along the two relevant PCA axes derived from the northern Levantine records. Also, cluster analyses of the Adriatic and northern Levantine records are compared.

On the average, the Adriatic Sea appears to have been considerably colder than the region around Crete, during the past 18,000 years. This temperature difference was largest during the pleniglacial and only small during the Holocene. We found no distinct expression of a Younger Dryas cooling event in the Adriatic records, neither on our PCA axis, nor in the oxygen isotope record. This contradicts previous paleoclimatic reconstructions, based mainly on paleobotanical evidence, which suggests that the Younger Dryas event greatly influenced circum-Mediterranean areas. As yet, we have no ready explanation for this discrepancy.

Low annual stability values dominate both the pleniglacial and the Holocene intervals of the Adriatic records. This suggests that homothermal conditions were as prevalent in the Adriatic euphotic layer at pleniglacial times as they are at present. In the present-day Adriatic, homothermal conditions are associated with deep convection and subsequent Eastern Mediterranean Deep Water (EMDW) formation, and therefore our data suggest that EMDW formation was as effective at pleniglacial times as it is at present. High annual stability values in the transitional interval, from 12,700 to 9600 yr B.P., in the Adriatic records are suggestive of low EMDW formation-rates at that time. In combination with high productivity throughout the eastern Mediterranean, which has previously been inferred from the planktic foraminiferal record, these low EMDW formation rates may have preconditioned the eastern Mediterranean for sapropel formation.

## 1. Introduction

In the marine pelagic ecosystem, planktic

foraminifera are essentially distributed, geographically as well as vertically, according to their preferences for specific temperature ranges and food requirements (e.g. Bé and Hamlin, 1967; Tolderlund and Bé, 1971; Kipp, 1976). Nearly all species thrive within the eu-

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photic layer. Within that layer, significant vertical differentiation of food quality and concentration may occur, so-called shallow and deep phytoplankton assemblages can be distinguished (for overview see Rohling and Gieskes, 1989). This vertical differentiation seems to be reflected in the assemblages of planktic foraminifera, which directly graze upon the phytoplankton or feed on other organisms that do so (Fairbanks and Wiebe, 1980; Fairbanks and Wiebe, 1982; Fairbanks et al., 1982; Hemleben et al., 1989).

Extensive studies on Northern Atlantic sediment cores have shown that the geographical distribution of planktic foraminifera has shifted through time, as a result of changes in hydrography and climate (e.g. Ruddiman and McIntyre, 1976; Crowley, 1981). Due to shifts in the biogeographical distribution, different North Atlantic assemblages successively occupied the waters in front of the Mediterranean gateway at Gibraltar. This invoked compositional changes in the planktic assemblages transported into the Mediterranean by the surface flow. Examples of this process are the repeated invasions into the Mediterranean of *Neogloboquadrina atlantica* in the late Pliocene. This species was continuously present in the middle to high latitude North Atlantic (Poore and Berggren, 1975). Zachariasse et al. (1990) demonstrated that, at times of northern hemisphere glaciations, the geographical distribution of *N. atlantica* periodically reached south of the Gibraltar passage, which enabled *N. atlantica* to migrate into the Mediterranean.

Within the Mediterranean, as anywhere, the abundance of planktic foraminiferal species is controlled by local environmental conditions. An outstanding example of abundance control by local environmental conditions is the distinct faunal contrast between the eastern and western Mediterranean subbasins (Thunell, 1978; Vergnaud Grazzini et al., 1986).

In this paper, we will analyze the Late Quaternary changes in the planktic foraminiferal records of three cores from the northern Lev-

antine Basin. According to our  $\delta^{18}\text{O}$  record, these cores extend back in time into the penultimate glacial (oxygen isotopic stage 6). Adding to the interpretation of Rohling and Gieskes (1989), who discussed the stratification history of the eastern Mediterranean euphotic layer on the basis of the record of the genus *Neogloboquadrina* in these cores, we will try to understand which processes accounted for variations in the record of the total planktic foraminiferal association.

We will then compare the records of two cores from the deep part of the Adriatic Sea to those of the northern Levantine cores. According to the  $\delta^{18}\text{O}$  record, the Adriatic cores reach into the last glacial maximum, which is confirmed by AMS  $^{14}\text{C}$  dates (Jorissen et al., 1993, this volume). By comparing the northern Levantine records to the two selected Adriatic records, we aim to show the history of environmental differences between the rather oceanic northern Levantine Basin and the Adriatic Sea, which has a much more continental setting and into which the Po river discharges huge amounts of freshwater loaded with nutrients.

## 2. Material and methods

Three gravity cores were taken in the northern Levantine Basin near Crete with the Dutch research vessel *Tyro* in May 1987 (Rohling and Gieskes, 1989). These cores (T87/2/13G, 306 m water depth; T87/2/20G, 707 m; and T87/2/27G, 607 m) were sampled at close intervals; the average sample distance is 3 cm. Geographic positions of the cores are shown in Fig. 1, the core lithologies in Fig. 2. All samples were sieved with mesh widths of 63, 150 and 595  $\mu\text{m}$ . We studied planktic foraminifera in the fraction between 150 and 595  $\mu\text{m}$ . The sample positions are indicated in Fig. 2a. Seven AMS  $^{14}\text{C}$  datings were performed on samples from the upper parts of the cores T87/2/20G and T87/2/27G (Fig. 2). Oxygen isotopes of the planktic foraminiferal species *Globigerinoides ruber* have been determined in 90 samples from core T87/2/20G, our most complete core. The

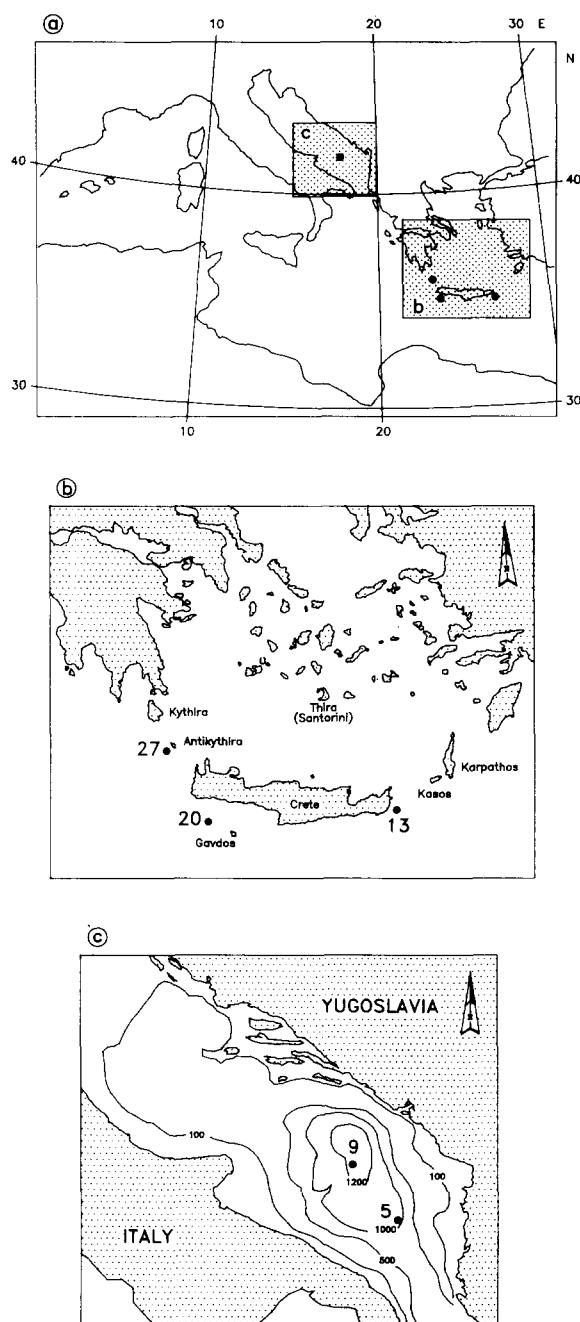


Fig. 1. Locations of the cores T87/2/13G, T87/2/20G, T87/2/27G, IN 68-5 and IN 68-9.

isotopic record and the isotopic stages are shown in Fig. 2.

Bioturbation has reduced the applicability of core T87/2/13G. We think, however, that the bioturbation should not be a reason to simply

discard the information from T87/2/13G, since the core recovered an exceptionally complete sequence from a shallow Levantine site. Therefore, we will use core T87/2/13G, but only when discussing general trends, and not for more detailed problems such as colour changes and occurrences of pink coloured morphotypes of *G. ruber*.

Two cores from the southern Adriatic Sea (Fig. 1) have been selected from a study by Jorissen et al. (1993, this volume). The lithology of the two Adriatic cores, IN 68-5 and IN 68-9, is shown in Fig. 3. The planktic foraminiferal records of IN 68-5 and IN 68-9 are based on determinations in the 150–595  $\mu\text{m}$  size-fraction. An oxygen isotope record on *G. ruber* has been constructed for core IN 68-5 (Fig. 3). The Adriatic cores have been accurately dated with the AMS  $^{14}\text{C}$  technique by Jorissen et al. (1993, this volume), who defined and dated three planktic foraminiferal biozones for the interval between 18,000 yr B.P. and Recent. The positions of the boundaries between these foraminiferal zones are indicated in Fig. 3.

For our interpretations of the fauna, we use habitat characteristics for each species as described in previous studies. These habitat characteristics, and the source references, are listed in the Appendix.

### 3. The northern Levantine cores

Our highly detailed oxygen isotope record of core T87/2/20G, along with the seven AMS  $^{14}\text{C}$  dates in T87/2/20G and T87/2/27G, endorses the tentative correlation of the northern Levantine cores with the eastern Mediterranean standard core RC9-181 (Vergnaud Grazzini et al., 1977) proposed previously by Rohling and Gieskes (1989). From top to bottom, the four sapropels present in each core are identified as  $S_1$ ,  $S_3$ ,  $S_4$  and  $S_5$ . In this paper, the term sapropel is applied to sediments deposited in dysoxic to anoxic bottom waters. Generally, these sediments are brown to black col-

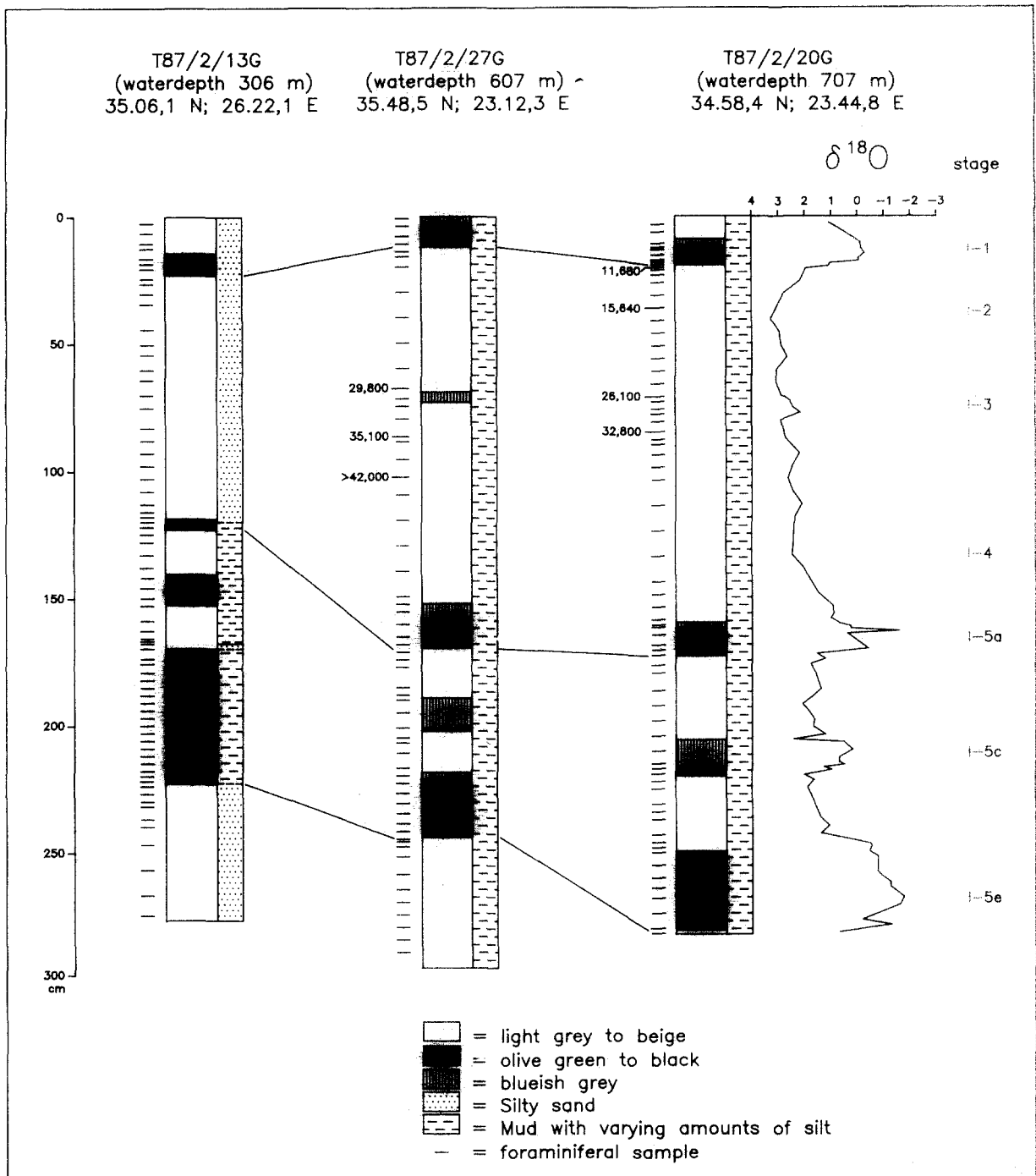


Fig. 2. Lithology and sample positions of cores T87/2/13G, T87/2/20G and T87/2/27G (modified from Rohling and Gieskes, 1989). Small numerals along T87/2/20G and T87/2/27G represent AMS  $^{14}\text{C}$  dates. The curve represents the oxygen isotopic record of T87/2/20G and a rough subdivision in isotopic stages is indicated.

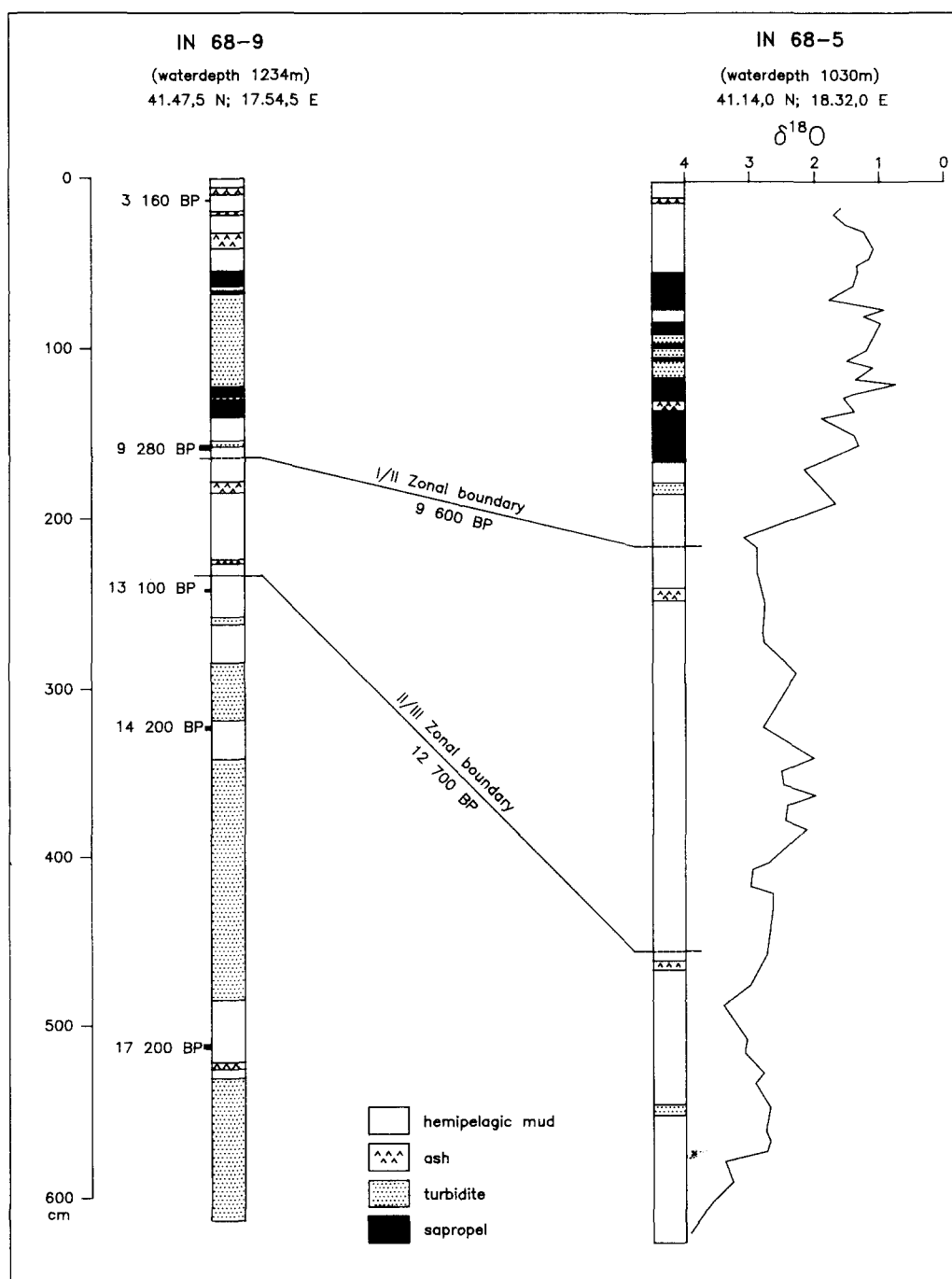


Fig. 3. Lithology of cores IN 68-5 and IN 68-9, with planktic foraminiferal zonal boundaries I/II and II/III, dated at 9600 and 12,700 yr B.P., respectively. Curve on right hand side represents oxygen isotope record of core IN 68-5, based on *G. bulloides*.

oured, and contain more than average amounts of pyrite and benthic foraminifera indicative of low bottom water oxygen conditions. Also, benthic foraminifera may be completely absent from sapropels (so-called "benthic desert" conditions). Often, sapropels are distinctly laminated. From a paleoenvironmental point of view, we do not consider the organic carbon content in sediments to be a useful means to discern what is (not) a sapropel. Factors such as variations in the sedimentation-rate and post-depositional oxydation may substantially influence the organic carbon content, in addition to the primary controls (export production and deep-water oxygen concentrations) which are highly indicative of environmental conditions at times of sapropel deposition.

In Fig. 2, hatched intervals have been indicated above several sapropels in the cores T87/2/20G and T87/2/27G. There, a sharp and distinct colour change, marking the visible top of a sapropel, is followed by a relatively minor colour change a few centimetres upcore. The

depths of these minor colour changes are listed in Table 1, and they are indicated with dashed lines in the planktic foraminiferal records displayed in Figs. 4–6. The hatched intervals represent blueish-grey clays which are lighter in colour than the sapropel below, but still somewhat darker than the clay above.

This might indicate that the sapropels initially were thicker, extending up to the (at present) minor colour change, and that they became partly oxydized downwards when the bottom waters were oxygenated again. Such a process has been described by De Lange et al. (1989) as "burning down". Rasmussen (1991) argued convincingly that "burning down" affected the Holocene sapropel  $S_1$  in cores from the Ionian Basin. Alternatively, Hilgen (1987) studied Pliocene sapropels and preferred to regard the blueish-grey layer as genuine and characteristic of a wet climatic phase, in which the incorporated sapropel would represent the extreme.

We favour the first interpretation, the blueish-grey (hatched) intervals being reoxi-

TABLE 1

The upper limits of the blueish-grey layer which occurs above most sapropels in the northern Levantine cores T87/2/20G and T87/2/27G

|                |   |
|----------------|---|
| Core T87/2/20G |   |
| Above $S_1$    | colour change between 8 and 9 cm  |
| Above $S_3$    | colour change at about 160 cm   |
| Above $S_4$    | colour change at 206 cm   |
| Core T87/2/27G |   |
| Above $S_1$    | no change visible in the 2 cm recovered above $S_1$   |
| Above $S_3$    | colour change between 152 and 153 cm  |
| Above $S_4$    | colour change between 188 and 189 cm  |
| Above $S_5$    | not obvious, sample at 217.5 cm may as well seem to be part of $S_5$ due to bioturbational mixing |

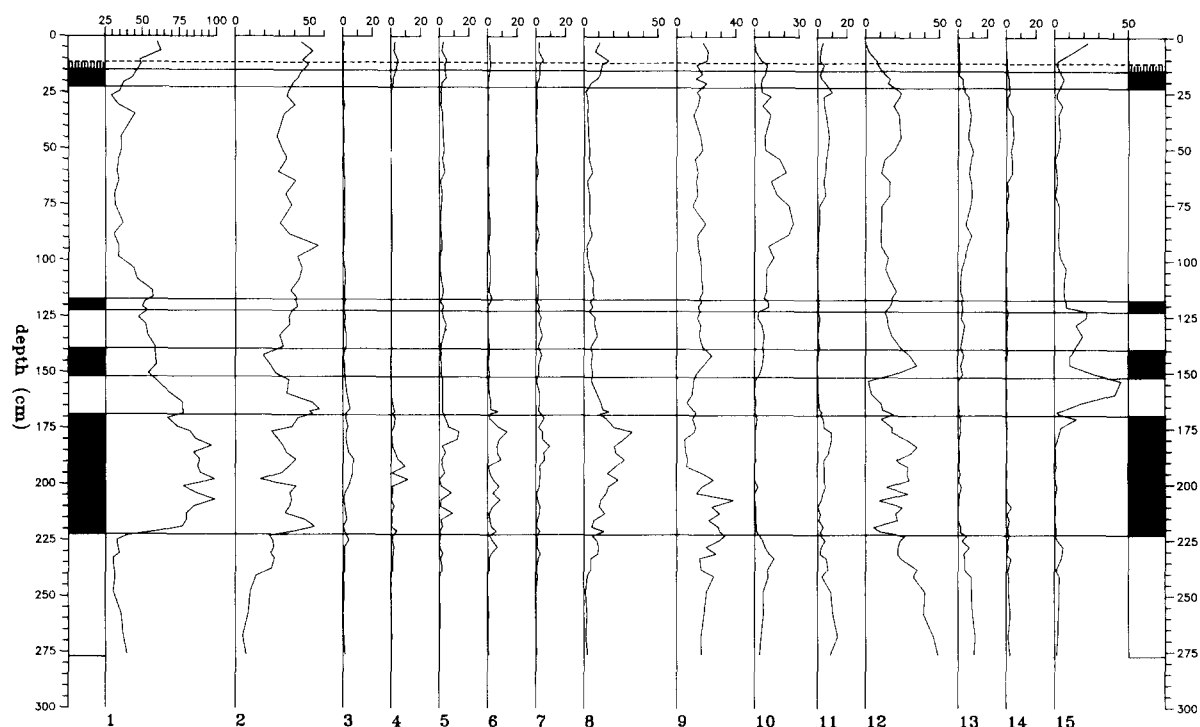


Fig. 4. Planktic foraminiferal percentages in T87/2/13G. 1 =  $100 \times P / (P + B)$ ; 2 = *G. ruber*; 3 = *O. universa*; 4 = *G. sacculifer*; 5 = *G. rubescens*; 6 = *G. tenella*; 7 = *G. siphonifera*; 8 = SPRUDTS-group; 9 = *G. bulloides*; 10 = *G. inflata*; 11 = *G. glutinata*; 12 = neogloboquadrinids; 13 = *T. quinqueloba*; 14 = *G. scitula*; 15 = indeterminate. Frequencies of *G. truncatulinoides* are too low to be plotted.

dized continuations of the sapropels. This interpretation is supported by the fact that intervals poor in benthic fauna extend above the currently visible tops of the sapropels [reflected by high values of the plankton percentage  $100 \times P / (P + B)$  in Figs. 4–6]. This suggests that dysoxic to anoxic bottom water conditions prevailed. Additionally, the more important faunal changes do not seem to correspond with the tops of the currently visible sapropels, but rather match with the minor colour changes marking the top of the blueish-grey clays. Table 2 lists the differences in core depth between the currently visible sapropels and such faunal changes. A fair correspondence is suggested at the sapropel bases, but offsets occur at the tops. Therefore, we will in the following consider the sapropels not to be delimited to the olive green to black coloured intervals, but we include the blueish-grey (presumably reoxidized) intervals.

### 3.1. Faunal records

The planktic foraminiferal records of the three northern Levantine cores are shown in Figs. 4–6. The present-day habitats of the taxa are listed, with references, in the Appendix. Figure 7 presents a dendrogram of the different taxa in the cores T87/2/20G and T87/2/27G. It illustrates the clustering of the individually rare species *Globigerinoides sacculifer*, *Hastigerina pelagica*, *Globoturborotalita rubescens*, *Orbulina universa*, *Globigerinella digitata*, *Globoturborotalita tenella* and *Globigerinella siphonifera* into the so-called SPRUDTS-group, which will in the following be treated as one entity (see Appendix).

There is a marked consistency between the three faunal records, especially between those of T87/2/20G and T87/2/27G. The consistent pattern of *Neogloboquadrina* abundances

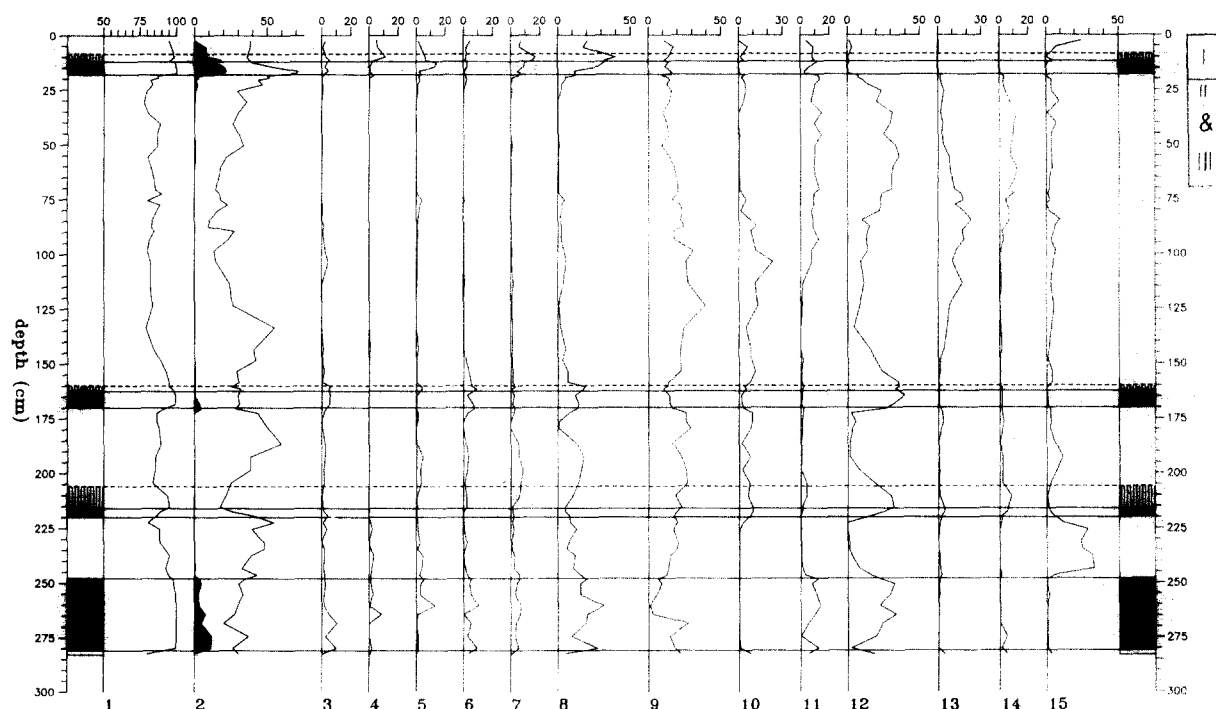


Fig. 5. Planktic foraminiferal percentages in T87/2/20G. Numbers correspond to those in the caption of Fig. 4. Within the distribution of *G. ruber*, the black coloured portion indicates the frequency of the pink variety. Roman numerals indicate the faunal zones defined by Jorissen et al. (1993, this volume).

in these cores has been discussed by Rohling and Gieskes (1989), who related it to previous observations of such a consistency (e.g. Kulenberg, 1952; Cita et al., 1977; Thunell et al., 1977; Vergnaud Grazzini et al., 1977). The frequency distribution of pink coloured *G. ruber*, determined in T87/2/20G and T87/2/27G, is very consistent as well. If this variety should indeed be considered as indicative of optimum warm conditions (see Appendix), temperature optima are suggested during the formation of  $S_1$  and (the lower part of)  $S_5$ .

The conspicuous peak of indeterminable specimens which occurs above sapropel  $S_5$  in all three cores (Figs. 4–6) is caused by the abundance of heavily carbonate-encrusted foraminifera. Nearly all foraminifera that could still be determined showed some coating as well. Moreover, encrustation has been observed on all species present in the interval, and we suspect that it is an early diagenetic feature.

Although we cannot rule out that some species are affected more than others, we think that our countings are still representative for the original associations.

On the average, a rough gradient appears in the percentages of *Turborotalita quinqueloba* and *Globorotalia scitula* versus *G. ruber* and the SPRUDTS-group, going from T87/2/13G via T87/2/20G to T87/2/27G (approximately from east to west). *T. quinqueloba* and *G. scitula* are indicative of low temperatures (e.g. Bé, 1969; Tolderlund and Bé, 1971; Thunell and Reynolds, 1984; Hemleben et al., 1989) and are most frequent in T87/2/27G (west). *G. ruber* and the SPRUDTS-group reflect higher temperatures (e.g. Tolderlund and Bé, 1971; Vergnaud Grazzini et al., 1986; Hemleben et al., 1989) and are most frequent in T87/2/13G (east). This suggests a general pattern of eastwards increasing temperatures throughout the last 125,000 years, the time-span covered



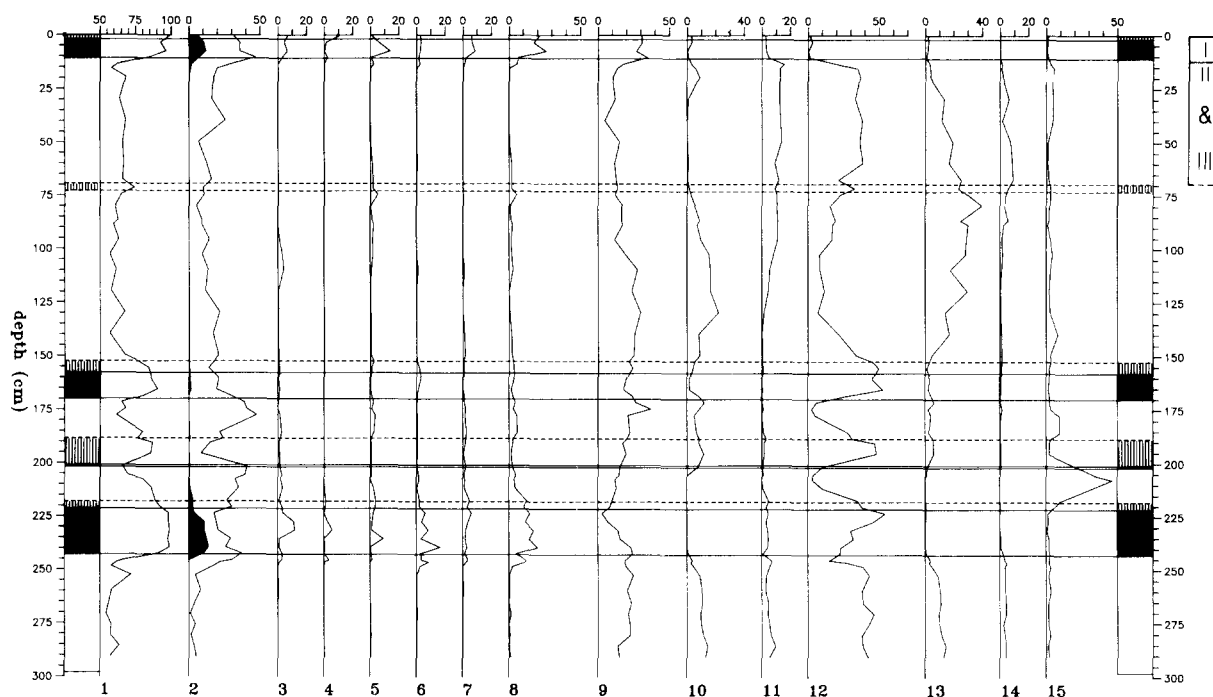


Fig. 6. Planktic foraminiferal percentages in T87/2/27G. Numbers correspond to those in the caption of Fig. 4. Within the distribution of *G. ruber*, the black coloured portion indicates the frequency of the pink variety. Roman numerals indicate the faunal zones defined by Jorissen et al. (1993, this volume).

by all three cores. This eastward temperature increase also exists at present (Vergnaud Grazzini, 1985; Stanev et al., 1989).

Apart from the consistent faunal changes in the three northern Levantine records, one important peculiarity was noticed, namely the marked increase of *Globigerina bulloides*, passing upwards from the last glacial maximum into S<sub>1</sub> in core T87/2/27G, whereas no such trend exists in the other two cores. T87/2/27G has been recovered from the Antikythira Straits, a passage between the western part of the Aegean Sea and the Levantine Basin. Therefore, we envisage that this increase of *G. bulloides* may be related to a similar increase in a core named SK1, which was recovered from the Sporades Basin, south of Skopelos Island (western Aegean; Zachariasse et al., 1991). Those authors argued that the *G. bulloides* increase in core SK1 resulted from increased nutrient concentrations during the S<sub>1</sub> formation, which resulted from enhanced river

runoff in northern Greece and extended southward by the western branch of the generally cyclonic Aegean circulation.

### 3.2. Principal Components Analysis

Because of the bioturbation in core T87/2/13G, we excluded this core from the statistical processing. The SPRUDTS-group was treated as one entity (see previous section and Appendix), and the sapropels are considered to incorporate the blueish-grey, presumably reoxidized, intervals. The habitat characteristics of the different species used in interpreting the results are discussed in the Appendix, and the reader is referred to that section for the source-citations.

A principal components analysis (PCA; Davis, 1973) has been performed on the combined records of T87/2/20G and T87/2/27G (total of 130 samples). Both the standardized and the non-standardized mode of PCA have

TABLE 2

The offsets between the limits of the presently visible sapropels and the major faunal changes in the northern Levantine cores T87/2/20G and T87/2/27G

| Core T87/2/20G |      |                   |                      |               |
|----------------|------|-------------------|----------------------|---------------|
| sapropel       |      | presently visible | major faunal changes |               |
| S <sub>1</sub> | top  | 12                | between              | 5.5 & 10.0    |
|                | base | 18                | between              | 17.5 & 18.5   |
| S <sub>3</sub> | top  | 162.5             | between              | 158.5 & 160.5 |
|                | base | 172               | between              | 170.5 & 172.5 |
| S <sub>4</sub> | top  | 216               | between              | 204.5 & 210.5 |
|                | base | 220               | between              | 219.5 & 222.5 |
| S <sub>5</sub> | top  | 248               | between              | 246.5 & 248.5 |
|                | base | 281               | between              | 280.0 & 282.5 |

| Core T87/2/27G |      |                   |                      |               |
|----------------|------|-------------------|----------------------|---------------|
| sapropel       |      | presently visible | major faunal changes |               |
| S <sub>1</sub> | top  | 2                 | above                | 0.5 (?)       |
|                | base | 11                | between              | 10.5 & 13.5   |
| S <sub>3</sub> | top  | 158               | between              | 149.5 & 152.5 |
|                | base | 170               | between              | 168.5 & 171.5 |
| S <sub>4</sub> | top  | 201               | between              | 185.5 & 188.5 |
|                | base | 202               | between              | 201.5 & 205.5 |
| S <sub>5</sub> | top  | 218               | between              | 211.5 & 217.5 |
|                | base | 243               | between              | 242.5 & 245.5 |

been applied; the first relying on the correlation-coefficients, and the second on covariances between the different taxa.

From the standardized PCA, the first two components are considered; PC1 which describes 32.27%, and PC2 which describes 22.31% of the variation (Table 3A). Table 4A lists the loadings of the taxa on these two axes. From the non-standardized PCA, as well, the first two components are considered; PC1 which describes 56.33%, and PC2 which describes 20.42% of the variation (Table 3B). Table 4B lists the loadings of the taxa on these two axes.

The standardized PC1 (Table 4A) shows highly positive loadings of *G. ruber* and the SPRUDTS-group, versus very negative loadings of *T. quinqueloba*, *G. scitula*, *Neogloboquad-*

*rina*, *G. glutinata* and *G. inflata*. This grouping corresponds quite closely to that found previously in principal components analysis on Quaternary eastern Mediterranean foraminiferal records by Thunell et al. (1977), although these authors found a stronger negative loading of *G. bulloides*, which probably results from the much larger area covered by their cores as compared to the restricted northern Levantine area focused on in the present study. As summarized in the Appendix, the positively loading taxa thrive in warm, subtropical waters, while the negatively loading taxa prefer cool/cold conditions or seem to be quite indifferent to temperature (*G. glutinata*, and neogloboquadrinids as a group). In agreement with Thunell et al. (1977), therefore, we interpret the standardized PC1 as a temperature axis,

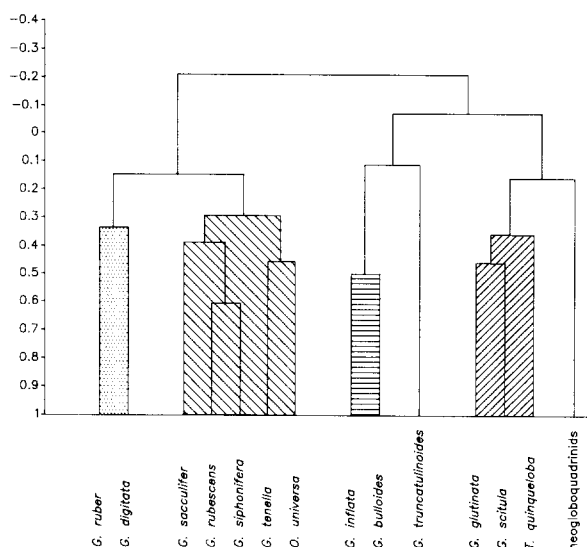


Fig. 7. Dendrogram of combined record of T87/2/20G and T87/2/27G. This illustrates the gathering of several rare species into the so-called SPRUDTS-group (see Appendix). Y-axis shows correlation coefficients.

with higher temperatures towards its positive side.

Note, however, that the influence of nu-

trient concentrations inevitably interferes with the temperature axis. This problem arises from the fact that the warm side is dominated by typical subtropical mixed layer dwellers. The mixed layer is usually deprived of nutrients due to the (seasonal) density discontinuity which marks its base, since this discontinuity hampers turbulent advection of nutrients from below. Usually, the density discontinuity is a (summer) seasonal thermocline, developing due to solar heating of the surface waters. Thus, apart from being warmer, the mixed layer waters are relatively oligotrophic, compared to the waters below the density discontinuity. In winter, there is no density discontinuity hampering nutrient advection from below. On the cool side of the temperature axis, species with negative loadings are found which do not prefer the warm, oligotrophic mixed layer conditions. Some thrive below the mixed layer, down to the base of the euphotic layer, when a (seasonal) pycnocline resides within the euphotic layer. Others prefer the winter situation of cool/cold temperatures and fairly eutrophic

TABLE 3

(A) The percentages of the variation explained by each PCA axis separately and the cumulative percentage. The result for the standardized run on the combined record of T87/2/20G and T87/2/27G. (B) The percentages of the variation explained by each PCA axis separately and the cumulative percentage. The result for the non-standardized run on the combined record of T87/2/20G and T87/2/27G

| A   |         |              | B   |         |              |
|---|---------|--------------|---|---------|--------------|
| standardized PCA on T87/2/20G + T87/2/27G |         |              | non-standardized PCA on T87/2/20G + T87/2/27G |         |              |
| axis                                      | % trace | % cumulative | axis  | % trace | % cumulative |
| 1   | 32.27   | 32.27        | 1   | 56.33   | 56.33        |
| 2   | 22.31   | 54.58        | 2   | 20.42   | 76.75        |
| 3   | 12.66   | 67.24        | 3   | 9.89    | 86.64        |
| 4   | 12.04   | 79.28        | 4   | 7.70    | 94.35        |
| 5   | 9.13    | 88.41        | 5   | 2.20    | 96.54        |
| 6   | 5.11    | 93.77        | 6   | 1.80    | 98.34        |
| 7   | 3.26    | 96.77        | 7   | 1.11    | 99.45        |
| 8   | 2.48    | 99.25        | 8   | 0.50    | 99.95        |
| 9   | 0.75    | 100.00       | 9   | 0.05    | 100.00       |

TABLE 4

(A) Ranking of the species along the first and second PCA axis, together with their loadings. The result for the standardized run on the combined record of T87/2/20G and T87/2/27G. (B) Ranking of the species along the first and second PCA axis, together with their loadings. The result for the non-standardized run on the combined record of T87/2/20G and T87/2/27G

| A                           |       | B                               |       |
|-----------------------------|-------|---------------------------------|-------|
| Cores T87/2/20G + T87/2/27G |       | Cores T87/2/20G + T87/2/27G     |       |
| ranking on standardized PC1 |       | ranking on non-standardized PC1 |       |
| <i>G. ruber</i>             | +0.49 | <i>G. ruber</i>                 | +0.68 |
| SPRUDTS-group               | +0.41 | SPRUDTS-group                   | +0.21 |
| <i>G. truncatulinoides</i>  | +0.05 | <i>G. bulloides</i>             | +0.03 |
| <i>G. bulloides</i>         | -0.03 | <i>G. truncatulinoides</i>      | 0.00  |
| <i>G. inflata</i>           | -0.24 | <i>G. glutinata</i>             | -0.06 |
| <i>G. glutinata</i>         | -0.26 | <i>G. scitula</i>               | -0.06 |
| neogloboquadrinids          | -0.32 | <i>G. inflata</i>               | -0.07 |
| <i>G. scitula</i>           | -0.38 | <i>T. quinqueloba</i>           | -0.22 |
| <i>T. quinqueloba</i>       | -0.46 | neogloboquadrinids              | -0.65 |
| ranking on standardized PC2 |       | ranking on non-standardized PC2 |       |
| <i>G. glutinata</i>         | +0.42 | neogloboquadrinids              | +0.60 |
| <i>G. scitula</i>           | +0.23 | SPRUDTS-group                   | +0.31 |
| neogloboquadrinids          | +0.22 | <i>G. ruber</i>                 | +0.30 |
| SPRUDTS-group               | +0.22 | <i>G. truncatulinoides</i>      | 0.00  |
| <i>G. ruber</i>             | -0.01 | <i>G. scitula</i>               | -0.03 |
| <i>T. quinqueloba</i>       | -0.10 | <i>G. glutinata</i>             | -0.04 |
| <i>G. truncatulinoides</i>  | -0.16 | <i>G. inflata</i>               | -0.25 |
| <i>G. inflata</i>           | -0.54 | <i>G. bulloides</i>             | -0.34 |
| <i>G. bulloides</i>         | -0.59 | <i>T. quinqueloba</i>           | -0.52 |

conditions throughout the euphotic layer. The quality of our standardized PC1 is, therefore, reduced to a rough approximation of temperature. The downcore scores are plotted in Figs. 8 and 9.

The non-standardized PC1 (Table 4B) predominantly separates *G. ruber* from the neogloboquadrinids. The downcore scores are plotted in Figs. 8 and 9. Evidently, the interference of different environmental parameters is much more extreme than on the standardized PC1. On the non-standardized PC1, temperature-induced modulations in the *G. ruber* record interfere with information on Deep Chlorophyll Maximum (DCM) development reflected by the record of *Neogloboquadrina*. Such an interference is not very informative, and we will not use this axis for paleoenvironmental interpretation. Temperature fluctuations can better be approximated with the standardized PC1. With respect to the history

of DCM development, better information can be obtained directly from the frequency variations of *Neogloboquadrina*, and this has been done already for the three northern Levantine cores by Rohling and Gieskes (1989).

The standardized PC2 and the non-standardized PC2 yield quite similar results (Figs. 8 and 9). Our interpretation, however, is confined to the non-standardized PC2 (Table 4B) since the standardized PC2 (Table 4A) shows highest positive loadings of *G. glutinata*, which is a cosmopolitan species from which no conclusive paleoenvironmental information can be derived yet.

The non-standardized PC2 shows high positive loadings of *Neogloboquadrina*, the SPRUDTS-group and *G. ruber*. The first prevail in areas with a well-developed "deep" (DCM) habitat and the latter two occur predominantly in the "shallow" (mixed layer) system, while some members of the SPRUDTS-group are

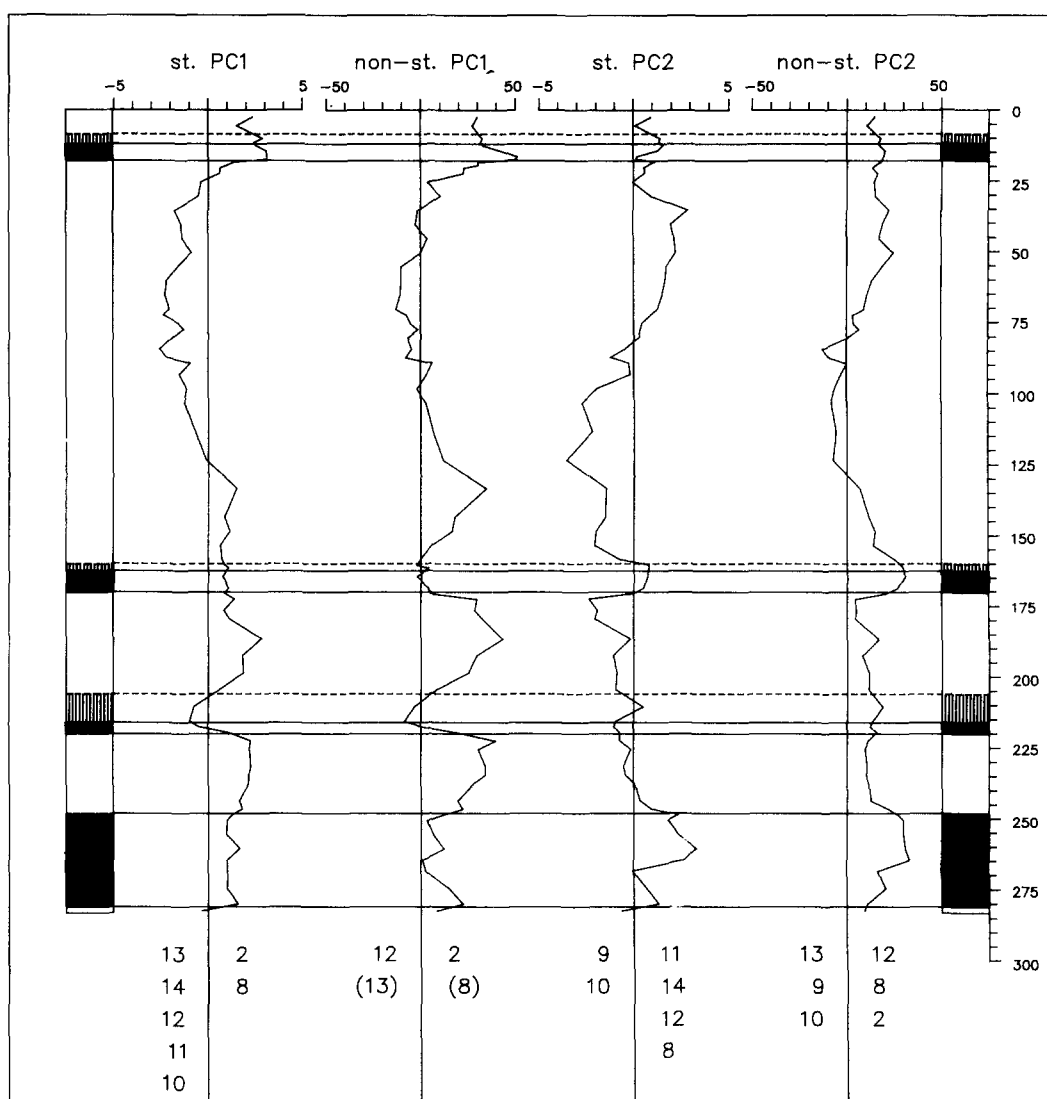


Fig. 8. Downcore scores on the four PCA axes for core T87/2/20G. Numbers below the axes correspond to those in the caption of Fig. 4 and indicate the species dominating the positive or negative sides of the axes (cf. Table 4).

known to thrive in the DCM during part of their life-cycle (see Appendix). Summarized, the association of these species is suggestive of a well-stratified euphotic zone, with a relatively oligotrophic mixed layer above the (seasonal) pycnocline and a DCM below it down to the base of the euphotic layer. On the negative side of the non-standardized PC2, a grouping occurs of *T. quinqueloba*, *G. bulloides*, and *G. inflata*. The first two species were grouped also by Van Leeuwen (1989), who in-

terpreted them as upwelling-indicators. Upwelling areas generally are characterized by a homothermal water column and high food levels. *G. inflata* is a species which requires a homothermal water column and intermediate food levels (see Appendix). Moreover, these three species are found to flourish in winter in the western Mediterranean (Vergnaud Grazzini et al., 1986), when the seasonal thermocline is disrupted and a homogeneous water column exists throughout the euphotic zone.

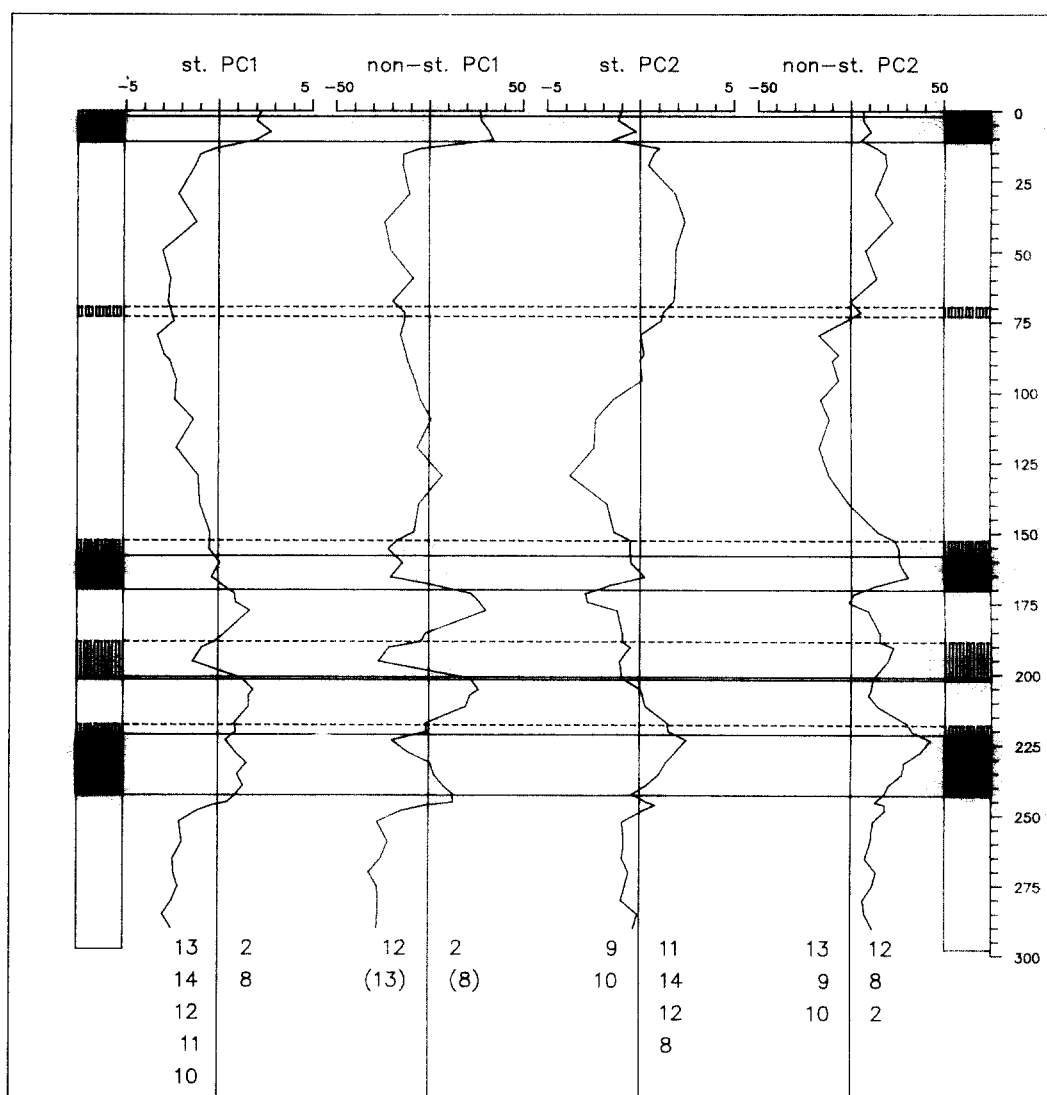


Fig. 9. Downcore scores on the four PCA axes for core T87/2/27G. Numbers below the axes correspond to those in the caption of Fig. 4 and indicate the species dominating the positive or negative sides of the axes (cf. Table 4).

The non-standardized PC2 may, therefore, be interpreted as a measure of “annual stability”. A shift towards more positive scores indicates that the euphotic layer was stratified over a longer proportion of the average year. A shift towards more negative values reflects prevalent homothermal conditions. In the vicinity of river mouths, abundances of *G. bulloides* may result from riverine discharge of particulate organic matter (see Appendix). In such settings, the flux of particulate organic

matter may, therefore, substantially influence the scores on the non-standardized PC2, pushing them to negative values. The complex interplay of environmental variations influencing the scores on the non-standardized PC2 is schematically represented in Fig. 10.

Summarizing, there are two axes which seem to yield relevant paleoenvironmental information. The standardized PC1 approximates temperature variations, and the non-standard-

ized PC2 probably approximates "annual stability".

### 3.2.1. Interpretation of the downcore scores on the two relevant PCA axes

The downcore fluctuations along the standardized PC1 (Figs. 8 and 9) are very similar to the temperature records of the cores T87/2/20G and T87/2/27G presented by Rohling and Gieskes (1989), which were based on a priori grouping of warm and cool water indicators. A similar conclusion was reached by Thunell et al. (1977). Thus, if interested only in determining paleotemperature variations, one might refrain from using the more elaborate PCA method, and favour a priori grouping (cf. Cita et al., 1977; Thunell et al., 1977; Rohling and Gieskes, 1989). In the present study, however, the results of the non-standardized PC2 are considered equally interesting, although less easily interpreted than those of the standardized PC1.

On the standardized PC1, our temperature

axis, glacial and interglacial intervals are clearly distinguishable, although neither the distinct peak glacials, nor the warmer interstadials show up. Also in the oxygen isotope record of core T87/2/20G (Fig. 2), the interstadial stage 3 cannot be distinguished from the glacial stages 2 and 4 without ambiguity. Thunell et al. (1977) found more obvious reflections of interstadials on their temperature axes, as well as in their isotope record. The isotope record of the eastern Mediterranean standard core RC9-181 (Vergnaud Grazzini et al., 1977) also showed a more prominent stage 3. As yet, we cannot provide a satisfactory explanation.

Apparently, the entire curve along the standardized PC1 for core T87/2/27G (Fig. 9) is displaced towards more negative (colder) values compared to that for T87/2/20G (Fig. 8), which suggests that the temperature near Antikythira Island has consistently been somewhat lower than that near Gavdos Island during the past 125,000 years.

Our temperature results endorse those of

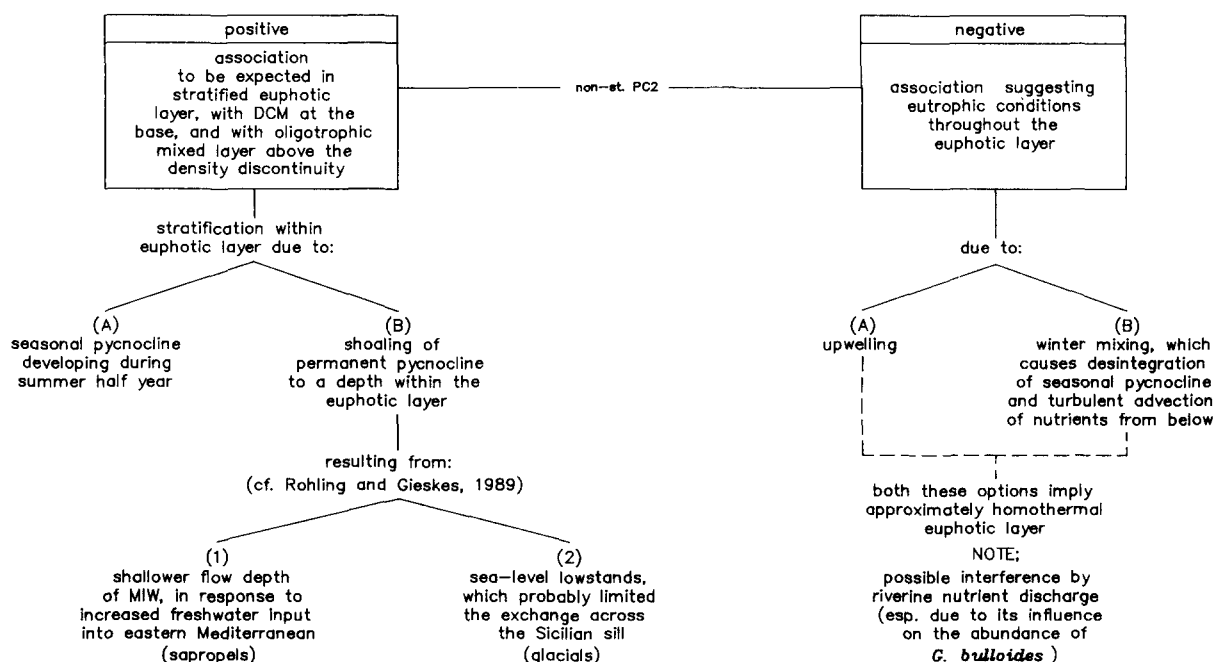


Fig. 10. Schematic representation of the complex interplay of environmental parameters influencing the scores along the non-standardized PC2.

Thunell et al. (1977), suggesting that sapropels  $S_1$  and  $S_5$  were deposited at highest temperatures,  $S_3$  during a slightly cooler phase and  $S_4$  during a short, remarkably cold period. A Holocene temperature optimum within  $S_1$  is strongly suggested in both T87/2/20G and T87/2/27G. No conspicuous temperature optimum is reflected within  $S_5$  (Figs. 8 and 9), in contrast to what we expected on the basis of high frequencies of pink *G. ruber*—a temperature optimum in the lower half of  $S_5$ .

As argued in the previous section, the non-standardized PC2 may be an indication of "annual stability". In the eastern Mediterranean, the annual stability within the euphotic layer may be increased in two ways (see Fig. 10). Firstly, by means of shortening of the cool and stormy season with deep vertical mixing, which implies a lengthening of the part of the year during which a seasonal thermocline is present. Secondly, by rising of the permanent pycnocline up to a depth within the euphotic layer (cf. Rohling and Gieskes, 1989). This permanent pycnocline essentially is a halocline, forming the transition between Mediterranean Intermediate Water (MIW) and surface water.

In interglacial non-sapropelic intervals, the values for annual stability are fairly high (Figs. 8 and 9). According to Rohling and Gieskes (1989), however, these periods were probably characterized by a permanent pycnocline at a position comparable to the present, namely below the base of the euphotic zone. Temperature was high during these periods, and therefore we envisage that the fairly high annual stability values reflect a significant persistence of the seasonal thermocline during a large part of the year.

At first sight, it is surprising that the upper half of the last glacial interval, which should contain the Last Glacial Maximum, displays a high annual stability, which is consistent in both cores. This may, however, be explained by persistence of the permanent pycnocline within the euphotic zone during this interval

of time. Such a shallow position of the permanent pycnocline would be a likely result of glacio-eustatic sea-level lowering (Rohling, 1991).

In the lower half of the last glacial interval, however, annual stability appears to be consistently low in both cores. Apparently the hydrographic conditions in the euphotic zone were not uniform throughout the last glacial and homothermality of the watercolumn within the euphotic layer was much more pronounced in the earlier part than in the later part. This may be interpreted as an indication of either a deeper position of the permanent pycnocline and/or a shorter thermally stratified season in the earlier part of the last glacial. As yet we cannot offer a sound explanation as to which of these two possibilities dominated.

The sapropels  $S_3$  and  $S_5$  are characterized by high annual stability values. Annual stability values in  $S_4$  are lower, especially in core T87/2/20G. Annual stability during the formation of  $S_1$  appears to have been about equal to that during the formation of  $S_4$  in core T87/2/20G, but less than that during the formation of  $S_4$  in core T87/2/27G. Rohling and Gieskes (1989) reasoned that the pycnocline probably resided within the euphotic layer during the formation of  $S_3$ ,  $S_4$  and  $S_5$ , which should cause high annual stability values in these sapropels. The relatively low values in  $S_4$ , therefore, may indicate a relatively short thermally stratified season during the formation of this sapropel as compared to that during the formation of  $S_3$  and  $S_5$ . On the standardized PC1, our temperature proxy (Figs. 8 and 9)  $S_4$  appeared significantly colder than  $S_3$  and  $S_5$ . Hence, we envisage that there is a link between the lower temperatures during the formation of  $S_4$  and its lower annual stability values relative to those of  $S_3$  and  $S_5$ .

Being devoid of *Neogloboquadrina*, apart from a brief recurrence in its topmost part, sapropel  $S_1$  bears no indication of a (shallow) position of the permanent pycnocline within the euphotic zone (Rohling and Gieskes,



1989), but still annual stability appears to have been fairly high (Figs. 8 and 9). We assume that this resulted from strong prevalence of the seasonal thermocline, which may be related to high temperatures during the  $S_1$  formation, as suggested by the standardized PC1.

#### 4. The Adriatic cores compared to the northern Levantine cores

Figures 11 and 12 show the planktic foraminiferal records of cores IN 68-5 and IN 68-9 from the southern Adriatic Sea. The core locations are shown in Fig. 1, the descriptions in Fig. 3. The time-stratigraphic framework for the Adriatic cores is provided by the oxygen isotope record of core IN 68-5 in combination with several AMS  $^{14}\text{C}$  dates (Fig. 3). The sapropel found in both cores IN 68-5 and IN 68-9 is identified as the Holocene sapropel  $S_1$ .

Within Figs. 3, 11 and 12, we indicated the

(bio-)zonal boundaries for intra-Mediterranean correlation as presented by Jorissen et al. (1993, this volume). Zone III is full glacial, the II/III boundary (12,700 yr B.P.) approximately matches the end of Termination 1a. Zone II spans the interval from 12,700 to 9600 yr B.P., and the Zone I/II boundary matches with Termination 1b. Zone I is the Holocene interglacial, containing sapropel  $S_1$ . The AMS  $^{14}\text{C}$  dates (Fig. 3) indicate that  $S_1$  formed between about 8300 and 6300 yr B.P., which is in fair agreement with previous estimates (Stanley and Maldonado, 1977, 1979; Vergnaud Grazzini, 1985; Troelstra et al., 1990, 1991; Rasmussen, 1991).

The faunal records of the two Adriatic cores are very similar, as expected since these cores were recovered from sites only about 100 km apart. They are remarkably different, however, from the records of the three northern Levantine cores. Due to the very low sedimen-

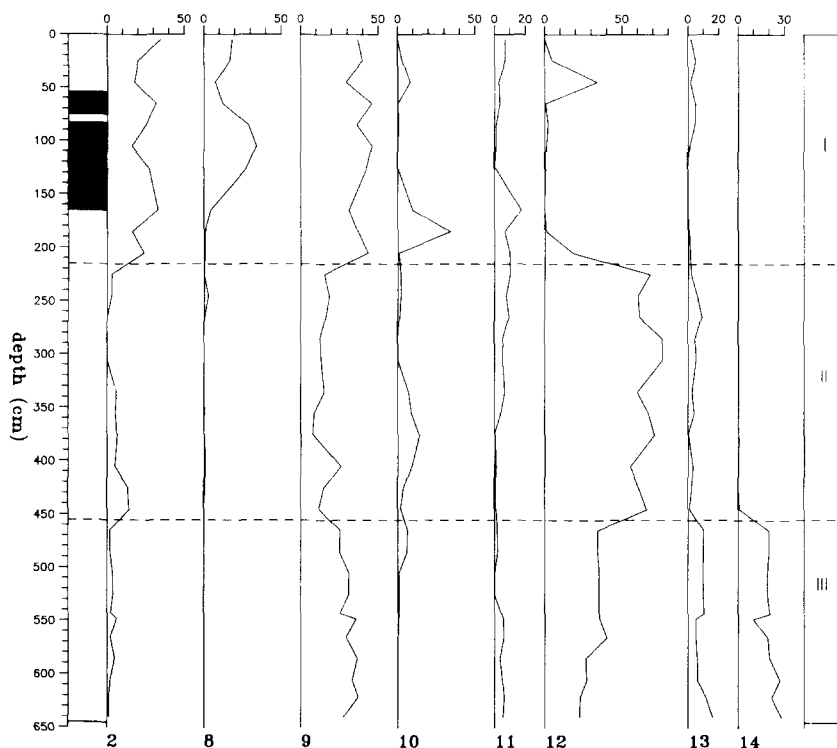


Fig. 11. Planktic foraminiferal frequency distribution in IN 68-5. Numbers correspond to those in the caption of Fig. 4. Roman numerals indicate the faunal zones defined by Jorissen et al. (1993, this volume).

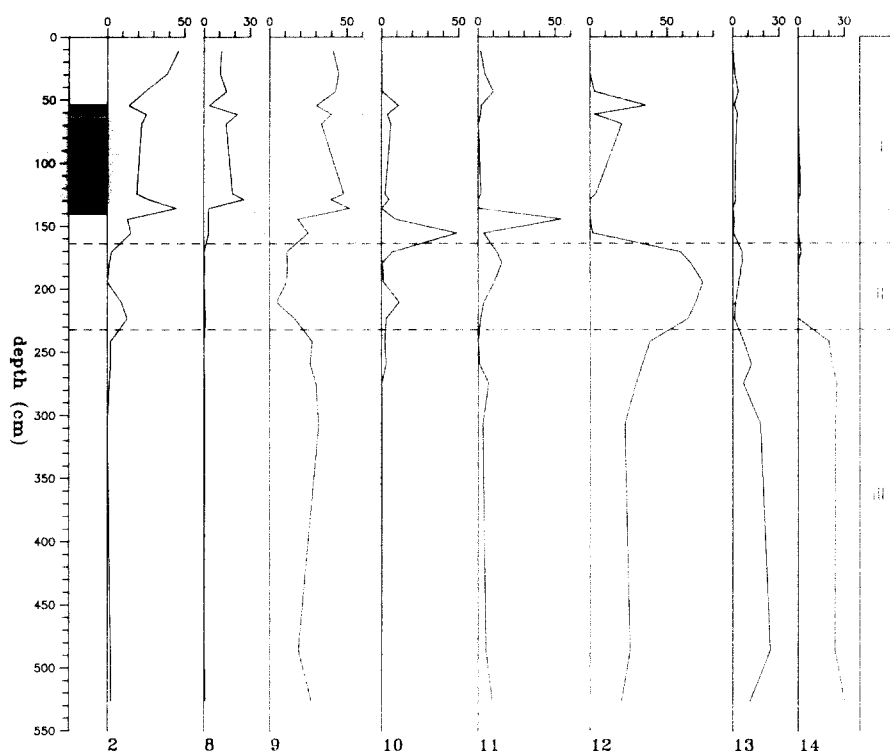


Fig. 12. Planktic foraminiferal frequency distribution in IN 68-9. Numbers correspond to those in the caption of Fig. 4. Roman numerals indicate the faunal zones defined by Jorissen et al. (1993, this volume).

tation rates in the northern Levantine cores, the three Adriatic biozones cannot be recognized without ambiguity. Especially the short Zone II (12,700–9600 yr B.P.) is difficult to discern in the northern Levantine cores. In the Adriatic cores, Zone II represents the interval between the top of Zone III (temporal exit *G. scitula*) and the base of Zone I (temporal exit *Neogloboquadrina* + marked increase *G. ruber*). These boundary criteria, however, suffered from bioturbation in the low sedimentation rate cores from the northern Levantine Basin. Especially in core T87/2/13G, the zonal discrimination appears impossible due to the intense bioturbation. For T87/2/20G and T87/2/27G (Figs. 5 and 6), we tentatively indicated the Zones I and III and also the interval which roughly corresponds in age to the Adriatic Zone II. From the Adriatic cores, no distinct criterion could be obtained to describe the base of Zone III. Based on cores T87/2/

20G and T87/2/27G, we propose to place the base of the glacial Zone III at the temporal exit of *G. inflata*, which roughly coincides with a marked increase of *G. scitula*. This proposed base of Zone III is also indicated in Figs. 5 and 6.

The species *G. ruber* is nearly absent in the Adriatic glacial Zone III (Figs. 11 and 12), whereas it is continuously present in the northern Levantine records (Figs. 5 and 6). *G. scitula* reaches higher percentages in the Adriatic Zone III than anywhere in the northern Levantine records. These differences suggest that the Adriatic Sea was characterized by substantially lower glacial temperatures than the northern Levantine Basin.

*G. bulloides* reaches high percentages in the Adriatic Holocene Zone I (Figs. 11 and 12), whereas it remains fairly constant in the northern Levantine cores (Figs. 5 and 6), from the last glacial to the Holocene (except in T87/2/

27G, as discussed in section 3.1). The frequency distribution of *G. bulloides* in the Adriatic seems likely to be influenced by variations in nutrient concentrations induced by the nearby Po river's discharge (see Appendix).

In the Adriatic records (Figs. 11 and 12) and in the northern Levantine records (Figs. 5 and 6), *Neogloboquadrina* abunds in the last glacial up to a marked frequency-drop to zero at the Pleistocene–Holocene transition. The extreme peak in the frequencies of *Neogloboquadrina* in the Adriatic Zone II (12,700–9600 yr B.P.), however, has no equivalent in the northern Levantine records. This discrepancy is genuine, since the almost continuous sampling below  $S_1$  in T87/2/20G should have revealed at least traces of a distinct frequency peak just prior to the Holocene (i.e. in Zone II), if such a peak were present. This discrepancy in (Zone II) neogloboquadrinid frequencies between the Adriatic and the northern Levantine records will be discussed in section 4.2.2.

#### 4.1. A brief comparison of cluster analyses

Cluster analyses (programme DENDRO; Drooger, 1982) were performed on 1. the combined records of T87/2/20G and T87/2/27G, which will serve as a northern Levantine standard, and 2. the combined records of IN 68-5 and IN 68-9, two standard cores for the Adriatic Sea. Unfortunately, it was not possible to run DENDRO on only the youngest part (last 18,000 years) of the northern Levantine records, since relatively few samples could be taken from that interval in these low sedimentation rate cores. Although we realize that there is a problem when we compare analyses based on records spanning different intervals of time, we apply DENDRO to show major differences and similarities between the general faunal groupings of the two areas.

The Adriatic dendrogram (Fig. 13) differs in two major aspects from the northern Levantine dendrogram (Fig. 14). The first differ-

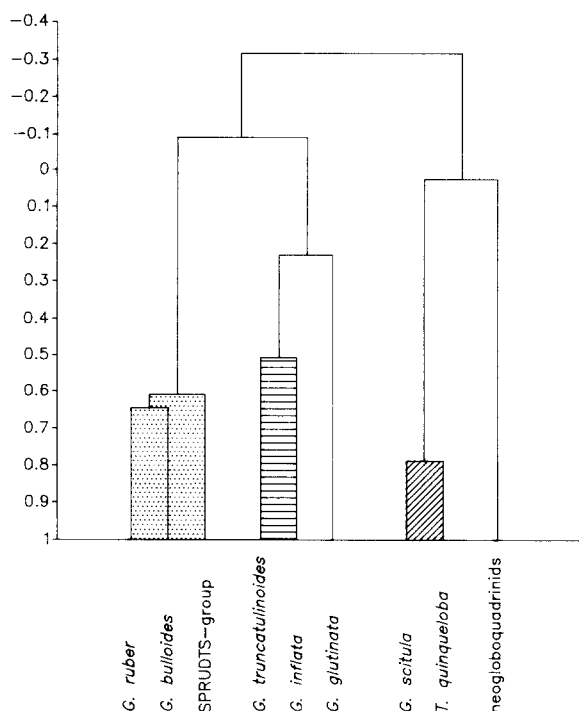


Fig. 13. Dendrogram of the combined record of the Adriatic cores IN 68-5 and IN 68-9. Y-axis shows correlation coefficients.

ence lies in the position of *G. bulloides*, an indicator of eutrophication due to upwelling or riverine influxes (see Appendix). In the Adriatic, *G. bulloides* appears strongly coupled to the warm-water cluster (i.e. *G. ruber* + SPRUDTS-group). This might suggest that the Po river discharge had the greatest influence during the warm (Holocene) interval in the Adriatic records. Secondly, the Adriatic dendrogram deviates from the northern Levantine dendrogram in as much as *G. inflata*, *G. glutinata* and *G. truncatulinoides* correlate negatively with the cool/cold-water cluster in the Adriatic records, whereas *G. inflata* and *G. glutinata* correlate negatively with the warm-water cluster in the northern Levantine records. Since these species are known neither as typical warm-water dwellers, nor as typical cold-water dwellers (Appendix), their affinity to the warm-water cluster in the Adriatic and to the cool/cold-water cluster in the northern

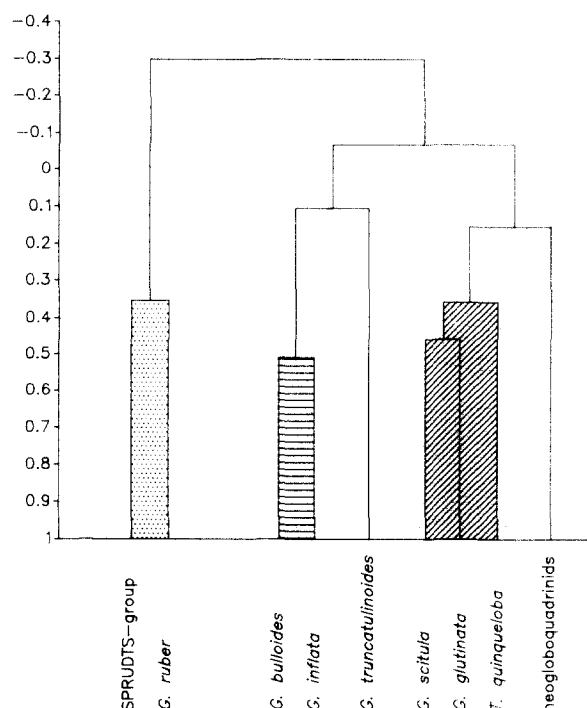


Fig. 14. Dendrogram of the combined record of the northern Levantine cores T87/2/20G and T87/2/27G. Y-axis shows correlation coefficients.

Levantine records might indicate a lower average temperature in the southern Adriatic Sea, relative to that in the northern Levantine Basin.

#### 4.2. Comparison of the northern Levantine and Adriatic records using PCA

We will not present separate principal components analyses of the Adriatic cores IN 68-5 and IN 68-9, since it would be difficult to compare with T87/2/20G and T87/2/27G owing to the relatively short time-span covered by the Adriatic records. Instead, we consider cores T87/2/20G and T87/2/27G as reference-sequences for the central and eastern Mediterranean, and subsequently interpret the faunal variation in the Adriatic cores in the context of the much longer northern Levantine records, by using the two paleoecologically relevant PCA axes of the latter.

In the non-standardized mode, we calcu-

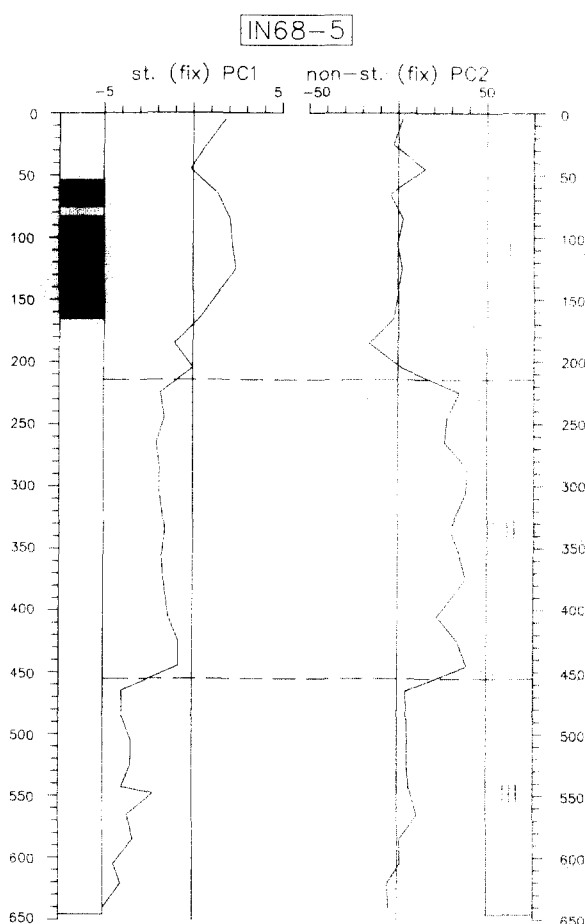


Fig. 15. Downcore scores on the two FPCA axes for core IN 68-5. Loadings of species along these axes are displayed in Table 4. Roman numerals indicate the faunal zones as defined by Jorissen et al. (1993, this volume).

lated the downcore scores of IN 68-5 and IN 68-9 by multiplying the relative frequencies of all taxa with their loadings on the PCA axes of the northern Levantine cores.

In the standardized mode, we standardized census data of IN 68-5 and IN 68-9 using the means and standard deviations from the matrix of the northern Levantine cores. Thereafter, the procedure was identical to the one in the non-standardized mode.

We realize that the described procedures, henceforth called FPCA (fixed principal components analysis), is only allowed in a region where specific paleoenvironmental fluctuations are expressed by similar faunal re-

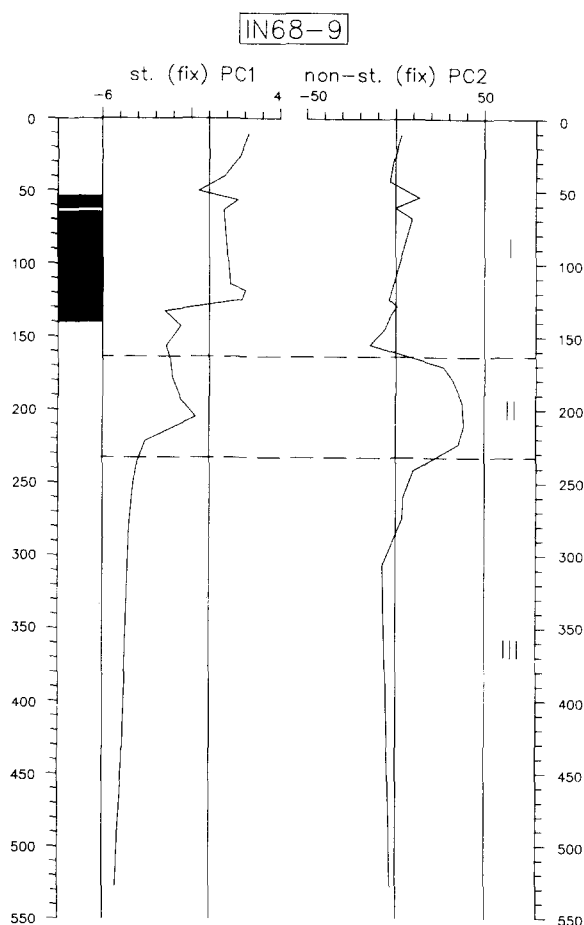


Fig. 16. Downcore scores on the two FPCA axes for core IN 68-9. Loadings of species along these axes are displayed in Table 4. Roman numerals indicate the faunal zones as defined by Jorissen et al. (1993, this volume).

sponges. We envisage that this is (approximately) the case for the central and eastern Mediterranean.

Figures 15 and 16 show the downcore FPCA scores of IN 68-5 and IN 68-9 (based on the fixed PCA axes of T87/2/20G + T87/2/27G). The only "fixed axes" presented are the standardized PC1 and non-standardized PC2 of T87/2/20G, since these are the only two yielding relevant paleoenvironmental information (see section 3.2).

#### 4.2.1. Interpretation of fluctuations along standardized FPC1; the temperature axis

The temperature differences between the Adriatic Sea and the northern Levantine Basin, which we inferred before (the former being cooler than the latter) are confirmed by the FPCA procedure. Especially the very negative values along the temperature axis (standardized FPC1) in the Adriatic full glacial interval (Zone III; Figs. 15 and 16) differ from those in the corresponding interval of the northern Levantine records (Figs. 8 and 9). This large difference might be an artifact of the method, since no such extremely negative values as found in the Adriatic full glacial interval occur anywhere in the Levantine records. We think, however, that the suggested trend, with more intense lowering of glacial temperatures in the land-locked Adriatic than in the more open oceanic sites of the northern Levantine Basin, may well be realistic. Stronger enrichment of  $\delta^{18}\text{O}$  in the Adriatic glacial Zone III (Fig. 3), relative to that in the corresponding interval of the northern Levantine records (Fig. 2), supports the trend observed with our temperature proxy. During the Holocene, our temperature proxy suggests that the temperature difference between the two areas appears to have been distinctly smaller than at glacial times. Nevertheless, the Adriatic Sea seems to have remained cooler than the northern Levantine Basin during the Holocene, as it is at present (Stanev et al., 1989). Hence, the glacial to interglacial temperature change appears to have been substantially larger in the Adriatic Sea than in the northern Levantine Basin.

Although the basal part of the interval between 12,700 and 9600 B.P. (Zone II) seems to be characterized by somewhat higher temperatures than the upper part of that zone (Figs. 15 and 16), in the Adriatic cores, no distinct Younger Dryas cooling event can be identified with our temperature proxy axis. Also, the Younger Dryas is poorly visible in the isotopic record of core IN 68-5 (Fig. 3), which endorses the information obtained from our

temperature proxy axis. This would suggest that the Younger Dryas, which so dramatically affected the Northern Atlantic realm (e.g. Berger et al., 1987; Mangerud, 1987), hardly had a noticeable effect on temperature in the Adriatic Sea. We find this difficult to believe, since previous paleoclimatic reconstructions suggested a great influence of the Younger Dryas event on the circum-Mediterranean climate (e.g. Guiot, 1987; Magaritz and Goodfriend, 1987; Pons et al., 1987; Rognon, 1987). As yet, we have no ready explanation for this discrepancy.

The temperature optimum during the formation of the Holocene sapropel S<sub>1</sub>, which is most prominent in the record of T87/2/20G, is much less conspicuous in the Adriatic records. In the Adriatic records, however, a distinct early Holocene temperature increase seems to precede the S<sub>1</sub> formation. Just above S<sub>1</sub>, a short recursion to lower values and a subsequent recuperation is suggested both in the Adriatic records and in that of T87/2/20G.

#### 4.2.2. Interpretation of fluctuations along non-standardized FPC2; "annual stability"

Along this axis, the Adriatic records (Figs. 15 and 16) display a very conspicuous tripartition. The glacial Zone III and the Holocene interglacial Zone I contain scores of comparable magnitude, whereas the transitional Zone II (12,700–9600 yr B.P.) is characterized by much higher scores. This pattern differs completely from that in the northern Levantine records (Figs. 8 and 9), where the last glacial to Holocene values remain more or less in the same range. The northern Levantine values are intermediate to the high values of the Adriatic Zone II and the low values of the Adriatic Zones I and III.

The high abundances of *Neogloboquadrina* in the last pleniglacial interval of the northern Levantine cores (Zone III; Figs. 5 and 6) are suggestive of a shallow position of the permanent pycnocline within the euphotic layer, which would be a likely result of glacio-eus-

tatic sea-level lowering (Rohling and Gieskes, 1989; Rohling, 1991). The presence of *Neogloboquadrina* in the Adriatic glacial interval (Zone III; Figs. 11 and 12) suggests that the pycnocline resided within the euphotic layer in that basin as well. Therefore, one would expect the annual stability scores to be as high in the Adriatic glacial interval as they are in the northern Levantine glacial interval. This expectation is, however, not supported by the actual values. The scores along the non-standardized FPC2 in the Adriatic full glacial interval (Zone III; Figs. 15 and 16) are much lower than those in the corresponding interval of the northern Levantine records (Figs. 8 and 9). Two processes may have caused this discrepancy.

Firstly, the Adriatic may have been characterized by a substantially shorter thermally stratified season, possibly as a result of its more continental setting, compared to the northern Levantine region. This would be supported by our previous observation that the Adriatic Sea probably was significantly colder than the northern Levantine Basin, at glacial times. Secondly, the influence of the Po river may have disturbed the signal along FPC2 by causing abundance increases of *G. bulloides*. Increased influence of the river's nutrient supply in the glacial interval is not necessarily due to enhanced discharge. It may also reflect the reduction of the distance between the river-mouth and the sites of IN 68-5 and IN 68-9 by, at least, 275 km, caused by the lower glacial sea level. Previously, however, we concluded from our cluster analysis that the influence of the Po river was relatively minor in the glacial interval and increased with temperature into the Holocene. Furthermore, Rohling and Gieskes (1989) argued that the higher than average primary production rates in the pleniglacial eastern Mediterranean should have resulted in a significant lowering of bottom water oxygen concentrations, unless substantial Eastern Mediterranean Deep Water (EMDW) was formed. Since the glacial sediments generally

appear to have accumulated under normal oxygenated conditions, this suggests that substantial EMDW formation indeed occurred at glacial times. Today, EMDW formation in the Adriatic Basin results from seasonal deep convection at times of reduced vertical density gradients and related low stability of the water column. Therefore, we envisage that the low glacial scores along FPC2 in the Adriatic, relative to those in the northern Levantine records, resulted primarily from a shorter stratified season in the Adriatic Sea compared to that in the northern Levantine Basin.

The very high scores on the non-standardized FPC2 in Zone II (12,700–9600 yr B.P.) of the Adriatic records (Figs. 15 and 16) are almost entirely due to extremely high frequencies of *Neogloboquadrina* in that interval (>60%; Figs. 11 and 12). As argued before, no such extreme peak is found just prior to the Holocene in the northern Levantine records. There, the neogloboquadrinid frequencies—just before the abrupt abundance drop at the base of the Holocene—are similar to those of the last pleniglacial interval.

Because of the general occurrence of *Neogloboquadrina*, we envisage that the pycnocline resided within the euphotic layer, and that a distinct DCM was developed throughout the eastern Mediterranean (including the Adriatic) during the pleniglacial and until about 9600 yr B.P. (i.e. top of Zone II). However, an additional factor must have contributed to the extreme dominance of neogloboquadrinids in the Adriatic between 12,700 and 9600 yr B.P. (Zone II). An argument inverse to the one applied to the pleniglacial section would imply that the thermally stratified season was longer in the Adriatic than in the northern Levantine Basin between 12,700 and 9600 yr B.P. This would mean that EMDW formation in the Adriatic was not very effective at that time. In combination with fairly high productivity rates throughout the eastern Mediterranean, low EMDW formation may have induced a grad-

ual decrease in the deep water oxygen concentrations, thus preconditioning the eastern Mediterranean for sapropel formation. This alleged preconditioning, starting at about 12,700 yr B.P., would be in agreement with previous studies, which suggested that the processes culminating in bottom water anoxia, and resultant deposition of S<sub>1</sub>, started already at about 13,800 yr B.P. (Troelstra et al., 1990, 1991), or 12,000 yr B.P. (Howell and Thunell, 1992).

The scores on the non-standardized FPC2 in the Holocene Zone I of the Adriatic records (Figs. 15 and 16) are markedly lower than those in the northern Levantine records (Figs. 8 and 9). This would suggest that, during the Holocene, thermal stratification was less prevalent in the Adriatic than in the northern Levantine Basin. Such an interpretation would agree with the fact that the Adriatic is the major site of EMDW formation. However, we have to be careful in interpreting the Holocene scores. Rossignol-Strick (1987) stated that from about 11,000 yr B.P. onwards and culminating during the formation of the sapropel S<sub>1</sub>, an increase of deciduous oak pollen occurred in the northern borderlands of the eastern Mediterranean. She interpreted this increase in terms of an increase in (summer) precipitation. Therefore, the frequencies of *G. bulloides* may have been enhanced by an increase of the Po river discharge and nutrient influx. Our cluster analysis of the Adriatic records showed a coupling of *G. bulloides* with *G. ruber* and the SPRUDTS-group, a major difference compared to its clustering with *G. inflata* in the dendrograms of the more oceanic northern Levantine records. We interpreted this anomaly by assuming that the Po river's influence had increased parallel to temperature. The low Holocene scores on the Adriatic non-standardized FPC2, therefore, may also be explained by an increased abundance of *G. bulloides* related to nutrient discharge of the Po river.

## 5. Conclusions

Throughout the past 125,000 years, an eastward increase of temperature seems to have existed from the site of T87/2/27G, near Antikythira island, via that of T87/2/20G, near Gavdos island, to that of T87/2/13G, just east of Crete.

Generally, the sapropels  $S_1$  and  $S_5$  seem to have been deposited at highest temperatures,  $S_3$  during a relatively cool phase and  $S_4$  during a remarkably cold period.

On the average, the Adriatic Sea has been significantly colder than the region around Crete during the last 18,000 years. This difference was largest during the pleniglacial and only small during the Holocene. The glacial to interglacial temperature gradient appears to have been much larger in the Adriatic Sea than in the northern Levantine Basin.

Apart from an axis expressing relative temperature changes, our PCA analysis also yields an axis which seems to reflect "annual stability", indicating the relative dominance of stratified or homothermal conditions in the euphotic layer.

In cores T87/2/20G and T87/2/27G, annual stability values are high in the upper part of the last glacial interval and in the sapropels  $S_3$  and  $S_5$ , probably as a result of a shallow position of the permanent pycnocline within the euphotic layer. Temperature and annual stability seem to have been low during  $S_4$  formation, which may indicate that, at that time, the thermally stratified season was shorter than during the formation of  $S_3$  and  $S_5$ . Annual stability during the formation of  $S_1$  appears to have been moderate, although there are no indications of a shallow permanent pycnocline position (within the euphotic layer). We think that the moderate annual stability during the formation of  $S_1$  is due to long persistence of the seasonal thermocline, since temperature was high at that time.

In the Adriatic cores IN 68-5 and IN 68-9, no distinct Younger Dryas cooling event could

be identified with our temperature proxy axis. Also, our detailed oxygen isotope record of IN 68-5 fails to show the Younger Dryas without ambiguity. This is not in accordance with previous paleoclimatic reconstructions of the circum-Mediterranean climate, which suggested a great impact of the Younger Dryas event. Furthermore, it is hard to discern interstadials within the last glacial interval of the northern Levantine records, even in the high resolution oxygen isotope record of T87/2/20G. As yet, we cannot offer a satisfactory explanation.

Low annual stability values in the last pleniglacial interval of the Adriatic cores contrast with high values in the corresponding interval of the northern Levantine cores. This may indicate that the thermally stratified season was shorter in the Adriatic, relative to that in the northern Levantine realm. If so, significant Eastern Mediterranean Deep Water (EMDW) formation may have occurred in the Adriatic, at pleniglacial times. High annual stability values in the interval between 12,700 and 9600 yr B.P. in the Adriatic records suggest strong prevalence of a seasonal thermocline, which could have resulted in relatively low EMDW formation rates. In combination with fairly high productivity throughout the eastern Mediterranean, this may have caused a gradual decrease in bottom water oxygen concentrations, thus preconditioning the eastern Mediterranean for sapropel formation.

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## Appendix

*Globigerinoides ruber*. Mixed layer species (Fairbanks et al., 1982; Hemleben and Spindler, 1983; Almogi-Labin, 1984; Thunell and Reynolds, 1984; Bé et al., 1985; Vergnaud Grazzini et al., 1986; Hemleben et al., 1989; Pujol and Vergnaud Grazzini, 1989; Van Leeuwen, 1989).  $\delta^{18}\text{O}$  reflects summer sea-surface temperatures (Ganssen, 1983).  $\delta^{13}\text{C}$  reflects that of surface  $\Sigma\text{CO}_2$  (Pujol and Vergnaud Grazzini, 1989). Temperature range 13.3° to 29.5°C, optimum above 21.3°C; pink variety most abundant at  $T > 24.4^\circ\text{C}$  (Tolderlund and Bé, 1971). Conclusions: reliable warm-water indicator; pink variety possibly reflects highest temperatures.

Although the following seven species do not live in completely identical habitats, their individually low frequencies and their tendency to group in cluster analyses, led us to lump them into one category, the SPRUDTS-group. Members of the SPRUDTS-group have been grouped in several other studies as well (e.g. Tolderlund and Bé, 1971; Thunell, 1978; Hemleben and Spindler, 1983; Almogi-Labin, 1984; Van Leeuwen, 1989).

*Globoturborotalita rubescens*, *Globigerinoides sacculifer* and *Hastigerina pelagica*. Often reported as mixed layer species (Hutson, 1977; Hemleben and Spindler, 1983; Almogi-Labin, 1984; Thunell and Reynolds, 1984; Bé et al., 1985; Vergnaud Grazzini et al., 1986; Pujol and Vergnaud Grazzini, 1989). Also deeper occurrences, however, for *G. sacculifer* (Fairbanks et al., 1982; Hemleben and Spindler, 1983; Bé et al., 1985) and *H. pelagica* (Hemleben and Spindler, 1983; Almogi-Labin, 1984; Bé et al., 1985). These probably represent specimens in reproducing stage, sunk due to spine shedding and/or breakdown of cytoplasm (Hemleben and Spindler, 1983).

*Orbulina universa*. Both in mixed layer (Bé et al., 1985; Vergnaud Grazzini et al., 1986; Reynolds and Thunell, 1989) and deeper waters (Fairbanks et al., 1982; Almogi-Labin, 1984; Bé et al., 1985). Lives surficially during spiral stage and deeper during spherical stage (Bé et al., 1985).

*Globigerinella siphonifera* and *Globoturborotalita tenella*. Thrive in deeper waters (Almogi-Labin, 1984; Bé et al., 1985).

*Globigerinella digitata*. Poorly known, but may be restricted to deeper waters (Hemleben et al., 1989).

Several members of SPRUDTS-group have optimum temperatures of about 20°C or higher (Tolderlund and Bé, 1971). Conclusions: SPRUDTS-group applicable as warm (subtropical) water indicator, especially when considered in combination with *G. ruber*.

*Globorotalia inflata*. Probably lives in deeper waters (Hutson, 1977; Pujol and Vergnaud Grazzini, 1989). However, seems to calcify in mixed layer (Fairbanks et al., 1980). Isotopic composition reflects winter sea surface temperatures (Ganssen, 1983). Peak abundances related to seasons of strong vertical mixing and a consequently homothermal water column (Tolderlund and Bé, 1971; Hutson, 1977; Pujol and Vergnaud Grazzini, 1989; Van Leeuwen, 1989). One of the authors (C.V.G.) recently observed high *G. inflata* abundances in a frontal system in the Alboran Sea. Distribution also reflects intermediate food levels (Van Leeuwen, 1989). Temperature range 2.2° to 22.6°C, optimum between 10.6°

and 17.3°C (Tolderlund and Bé, 1971). Conclusions: indicative of well-mixed, cool environment with intermediate nutrient levels.

*Globigerinita glutinata*. Most abundant in surficial (mixed layer) waters (Fairbanks et al., 1982; Bé et al., 1985; Thunell and Reynolds, 1984; Hemleben et al., 1989; Reynolds and Thunell, 1989), but may live in deeper waters as well (Almogi-Labin, 1984). Encompasses the entire temperature and salinity ranges studied by Tolderlund and Bé (1971). Greatest abundances occur at very different, not overlapping, temperature ranges. Conclusions: cosmopolitan species.

*Turborotalita quinqueloba*. Shallow (mixed layer) dweller (Hutson, 1977; Vergnaud Grazzini et al., 1986; Hemleben et al., 1989; Pujol and Vergnaud Grazzini, 1989; Reynolds and Thunell, 1989; Van Leeuwen, 1989). Considered as upwelling indicator in the Angola Basin, together with *Globigerina bulloides* (Van Leeuwen, 1989). Temperature range 2.2° to 16°C, although lower limit may even be about 1°C and upper limit about 21.5°C according to sparse findings; optimum between 4.6° and 10.8°C (Tolderlund and Bé, 1971). Test size seems inversely correlated to temperature (Kroon et al., 1988). Conclusions: tolerant to fairly low salinities and indicative of low temperatures and/or enhanced fertility in surficial waters (Ch. Hemleben, pers. commun., Cambridge, 1989).

*Neogloboquadrina*. Category in this study comprises predominantly dextrally coiled *pachyderma* types and quite tightly coiled *dutertrei* types. Both the reason to group them, and the records from three northern Levantine cores, are treated by Rohling and Gieskes (1989). Abundance closely linked to development of a Deep Chlorophyll Maximum (DCM; Rohling and Gieskes, 1989; and references therein). Often described as pycnocline-dwellers (also: Fairbanks and Wiebe, 1982; Fairbanks et al., 1980, 1982; Hemleben and Spindler, 1983; Thunell and Reynolds, 1984; Bé et al., 1985; Hemleben et al., 1989; Pujol and Vergnaud Grazzini, 1989; Reynolds and Thunell, 1989; Van Leeuwen, 1989). As a group, neogloboquadrinids have a very wide temperature range. Conclusions: thrive in deeper, eutrophic waters, in DCM and, therefore, considered indicative of stratified conditions within euphotic layer.

*Globorotalia scitula*. Probably thrives in very deep waters (Bé, 1969). Although little is known, it seems to behave more or less like *Globorotalia truncatulinoides* (see below), calcifying at very great depths. In spring, offshore Bermuda, *G. scitula* appear near surface as very thin shelled specimens and subsequently sink to deep water (Hemleben et al., 1989). Characteristic for subpolar waters, together with *T. quinqueloba* and *N. pachyderma* (Bé, 1969; Thunell and Reynolds, 1984). Occurs in temperate regions (Hemleben et al., 1989). Conclusions: occurs down to great depths, in cool to cold waters. *Globigerina bulloides*. Inhabits both mixed-layer (Bé, 1969; Bé et al., 1985; Hemleben et al., 1989; Reynolds and Thunell, 1989; Van Leeuwen, 1989) and deeper waters (Almogi-Labin, 1984; Hemleben et al., 1989; Pujol and Vergnaud Grazzini, 1989). Characterizes upwelling conditions (Kipp, 1976; Coulborn et al., 1980; Duplessy et al., 1981; Thiede, 1983; Zhang, 1985; Hemleben et al., 1989; Overpeck et al., 1989; Van Leeuwen, 1989). Also high standing stocks near river

mouths (Barmawidjaja et al., 1989; Van Leeuwen, 1989). Occurs in very warm tropical waters (Barmawidjaja et al., 1989) as well as in water with temperatures as low as 0°C (Tolderlund and Bé, 1971). May occur in warmer waters only when the food level is high, and in cooler waters also at more variable food levels (Van Leeuwen, 1989). Conclusions: no depth preference, but highly dependent on food levels enhanced by upwelling or river-input.

*Globorotalia truncatulinoides*. Very deep dwelling species (Tolderlund and Bé, 1971; Hutson, 1977; Hemleben and Spindler, 1983; Vergnaud Grazzini et al., 1986; Hemleben et al., 1989). Near Bermuda, *G. truncatulinoides* seems to reproduce in winter near the surface and continues to grow as it sinks to great depths (>1000 m; Hemleben et al., 1989). Lohmann and Schweitzer (1990), on the contrary, think that *G. truncatulinoides* reproduces at about 600 m water depth, and that vertical mixing down to that depth is required to return juveniles to surface. Size distribution may be indication of thermocline depth (Lohmann and Schweitzer, 1990). Temperature range between 10.5° and 28.1°C, optimum between 15.4° and 22°C (Tolderlund and Bé, 1971). Marked preference for coiling either sinistral or dextral (Tolderlund and Bé, 1971; Brummer and Kroon, 1988; Hemleben et al., 1989). Left-coiling shells spend only winter months in shallow water, while right-coiling shells persist there throughout the year (Lohmann and Schweitzer, 1990). Conclusions: occurs down to great depths, in fairly cool waters. Note: very rare in the studied eastern Mediterranean records.

## References

- Almogi-Labin, A., 1984. Population dynamics of planktonic foraminifera and pteropoda—Gulf of Aqaba, Red Sea. *Proc. K. Ned. Akad. Wet. Ser. B, Phys. Sci.*, 87(4): 481–511.
- Barmawidjaja, D.M., Van der Borg, K., De Jong, A.F.M., Van der Kaars, W.A. and Zachariasse, W.J., 1989. Kau Basin, Halmahera, a Late Quaternary paleoenvironmental record in a poorly ventilated basin. In: *Proc. Snellius II Symp. (Jakarta.) Neth. J. Sea Res.*, 24: 591–605.
- Bé, A.W.H., 1969. Distribution of selected groups of marine invertebrates in waters south of 35°S latitude. In: *Antarctic Map Folio Ser.*, 11: 9–12.
- Bé, A.W.H. and Hamlin, W.H., 1967. Ecology of recent planktonic foraminifera, 3, Distribution in the North Atlantic during the summer of 1962. *Micropaleontology*, 13: 87–106.
- Bé, A.W.H., Bishop, J.K.B., Sverdløve, M.S. and Gardner, W.D., 1985. Standing stock, vertical distribution and flux of planktonic foraminifera in the Panama Basin. *Mar. Micropaleontology*, 9: 307–333.
- Berger, W.H., Burke, S. and Vincent, E., 1987. Glacial Holocene transition: climate pulsations and sporadic shutdown of NADW production. In: W.H. Berger and L.D. Labeyrie (Editors), *Abrupt Climatic Change*. (NATO Adv. Study Inst., Ser. C, Math. Phys. Sci., 216.) Kluwer, Dordrecht, pp. 279–297.
- Brummer, G.J.A. and Kroon, D., 1988. Genetically controlled planktonic foraminiferal coiling ratios as tracers of past ocean dynamics. In: G.J. Brummer and D. Kroon (Editors), *Planktonic Foraminifera as Tracers of Ocean-climate History*. Free Univ. Press, Amsterdam, pp. 293–298.
- Cita, M.B., Vergnaud Grazzini, C., Robert, C., Chamley, H., Ciaranfi, N. and d'Onofrio, S., 1977. Paleoclimatic record of a long deep sea core from the eastern Mediterranean. *Quat. Res.*, 8: 205–235.
- Coulbourn, W.T., Parker, F.L. and Berger, W.H., 1980. Faunal and solution patterns of planktonic foraminifera in surface sediments of the North Pacific. *Mar. Micropaleontology*, 9: 329–399.
- Crowley, T.J., 1981. Temperature and circulation changes in the eastern North Atlantic during the last 150,000 years: evidence from the planktonic foraminiferal record. *Mar. Micropaleontology*, 6: 97–129.
- Davis, J.C., 1973. *Statistics and Data Analysis in Geology*. Wiley, New York, 550 pp.
- De Lange, G.J., Middelburg, J.J. and Pruysers, P.A., 1989. Discussion: Middle and Late Quaternary depositional sequences and cycles in the eastern Mediterranean. *Sedimentology*, 36: 151–156.
- Drooger, M.M., 1982. Quantitative Range Chart Analysis. *Utrecht Micropaleontology Bull.*, 26, 227 pp.
- Duplessy, J.-C., Bé, A.W.H. and Blanc, P.L., 1981. Oxygen and Carbon isotope composition and biogeographic distribution of planktonic foraminifera in the Indian Ocean. *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, 33: 9–46.
- Fairbanks, R.G. and Wiebe, P.H., 1982. Foraminifera and chlorophyll maximum: Vertical distribution, seasonal succession, and paleoceanographic significance. *Science*, 209: 1524–1525.
- Fairbanks, R.G., Wiebe, P.H. and Bé, A.W.H., 1980. Vertical distribution and isotopic composition of living planktonic foraminifera in the western North Atlantic. *Science*, 207: 61–63.
- Fairbanks, R.G., Sverdløve, M., Free, R., Wiebe, P.H. and Bé, A.W.H., 1982. Vertical distribution of living planktonic foraminifera from the Panama Basin. *Nature*, 298: 841–844.
- Ganssen, G., 1983. Dokumentation von küstennahen Auftrieb anhand stabiler Isotope in rezenten Foraminiferen vor Nordwest-Afrika. *Meteor. Forschungsberichte*, Reihe C, 37: 1–46.
- Guiot, J., 1987. Late Quaternary climatic change in France estimated from multivariate pollen time series. *Quat. Res.*, 28: 100–118.
- Hemleben, Ch. and Spindler, M., 1983. Recent advances in research on living planktonic foraminifera. In: J.E. Meulenkamp (Editor), *Reconstruction of Marine Paleoenvironments*. *Utrecht Micropaleontology Bull.*, 30: 141–170.

- Hemleben, Ch., Spindler, M. and Anderson, O.R., 1989. *Modern Planktonic Foraminifera*. Springer, New York, 363 pp.
- Hilgen, F.J., 1987. Sedimentary rhythms and high-resolution chronostratigraphic correlations in the Mediterranean Pliocene. *Newsl. Stratigr.*, 17: 109–127.
- Howell, M.W. and Thunell, R.C., 1992. Organic carbon accumulation in Bannock Basin: Evaluating the role of productivity in the formation of eastern Mediterranean sapropels. *Mar. Geol.*, 103: 461–471.
- Hutson, W.H., 1977. Variations in planktonic foraminiferal assemblages along north–south transects in the Indian Ocean. *Mar. Micropaleontol.*, 2: 47–66.
- Jorissen, F.J., Asioli, A., Borsetti, A.M., Capotondi, L., De Visser, J.P., Hilgen, F.J., Rohling, E.J., Van der Borg, K., Vergnaud Grazzini, C. and Zachariasse, W.J., 1993. Late Quaternary central Mediterranean biochronology. *Mar. Micropaleontol.*, 21: 169–189.
- Kipp, N.G., 1976. New transfer function for estimating past sea-surface conditions from sea-bed distribution of planktonic foraminiferal assemblages in the North Atlantic. In: R.M. Cline and J.D. Hays (Editors), *Investigation of Late Quaternary Paleoceanography and Paleoclimatology*. *Mem. Geol. Soc. Am.*, 145: 3–41.
- Kroon, D., Wouters, P.F., Moodley, L., Ganssen, G. and Troelstra, S.R., 1988. Phenotypic variation of *Turborotalita quinqueloba* (Natland) tests in living populations and in the Pleistocene of an eastern Mediterranean piston core. In: G.J. Brummer and D. Kroon (Editors), *Planktonic Foraminifera as Tracers of Ocean-climate History*. Free Univ. Press, Amsterdam, pp. 131–147.
- Kullenberg, B., 1952. On the salinity of the water contained in marine sediments. *Medd. Oceanogr. Inst. Göteborg*, 21: 1–38.
- Lohmann, G.P. and Schweitzer, P.N., 1990. *Globorotalia truncatulinoides* growth and chemistry as probes of the past thermocline: 1. shell size. *Paleoceanography*, 5: 55–75.
- Magaritz, M. and Goodfriend, G.A., 1987. Movement of the desert boundary in the Levant from latest Pleistocene to early Holocene. In: W.H. Berger and L.D. Labeyrie (Editors), *Abrupt Climatic Change*. (NATO Adv. Study Inst., Ser. C, Math. Phys. Sci., 216.) Kluwer, Dordrecht, pp. 173–183.
- Mangerud, J., 1987. The Allerød/Younger Dryas boundary. In: W.H. Berger and L.D. Labeyrie (Editors), *Abrupt Climatic Change*. (NATO Adv. Study Inst., Ser. C, Math. Phys. Sci., 216.) Kluwer, Dordrecht, pp. 163–171.
- Overpeck, J.T., Peterson, L.C., Kipp, N., Imbrie, J. and Rind, D., 1989. Climate change in the circum-North Atlantic region during the last deglaciation. *Nature*, 338: 353–357.
- Pons, A., De Beaulieu, J.-L., Guiot, J. and Reille, M., 1987. The Younger Dryas in southwestern Europe: an abrupt climatic change as evidenced from pollen records. In: W.H. Berger and L.D. Labeyrie (Editors), *Abrupt Climatic Change*. (NATO Adv. Study Inst., Ser. C, Math. Phys. Sci., 216.) Kluwer, Dordrecht, pp. 195–208.
- Poore, R.Z. and Berggren, W.A., 1975. The morphology and classification of *Neogloboquadrina atlantica* (Berggren). *J. Foraminiferal Res.*, 5: 692–694.
- Pujol, C. and Vergnaud Grazzini, C., 1989. Paleoceanography of the last deglaciation in the Alboran Sea (western Mediterranean). Stable isotopes and planktonic foraminiferal records. *Mar. Micropaleontol.*, 15: 153–179.
- Rasmussen, T.L., 1991. Benthonic and planktonic foraminifera in relation to the Early Holocene stagnation in the Ionian Basin, Central Mediterranean. *Boreas*, 20: 357–376.
- Reynolds, L.A. and Thunell, R.C., 1989. Seasonal succession of planktonic foraminifera: results from a four-year time-series sediment trap experiment in the northeast Pacific. *J. Foraminiferal Res.*, 19: 253–267.
- Rognon, P., 1987. Aridification and abrupt climatic events on the Saharan northern and southern margins, 20 000 Y BP to present. In: W.H. Berger and L.D. Labeyrie (Editors), *Abrupt Climatic Change*. (NATO Adv. Study Inst., Ser. C, Math. Phys. Sci., 216.) Kluwer, Dordrecht, pp. 209–220.
- Rohling, E.J., 1991. A simple two-layered model for shoaling of the eastern Mediterranean pycnocline due to glacio-eustatic sea level lowering. *Paleoceanography*, 6: 537–541.
- Rohling, E.J. and Gieskes, W.W.C., 1989. Late Quaternary changes in Mediterranean Intermediate Water density and formation rate. *Paleoceanography*, 4: 531–545.
- Rosignol-Strick, M., 1987. Rainy periods and bottom water stagnation initiating brine accumulation and metal concentrations: 1. the Late Quaternary. *Paleoceanography*, 2: 333–360.
- Ruddiman, W.F. and McIntyre, A., 1976. Northeast Atlantic paleoclimatic changes over the past 600 000 years. *Mem. Geol. Soc. Am.*, 145: 111–146.
- Stanev, E.V., Friedrich, H.J. and Botev, S.V., 1989. On the seasonal response of intermediate and deep water to surface forcing in the Mediterranean Sea. *Oceanol. Acta*, 12: 141–149.
- Stanley, D.J. and Maldonado, A., 1977. Nile Cone: Late Quaternary stratigraphy and sediment dispersal. *Nature*, 266: 129–135.
- Stanley, D.J. and Maldonado, A., 1979. Levantine Sea–Nile Cone lithostratigraphic evolution: quantitative analysis and correlation with paleoclimatic and eustatic oscillations in the Late Quaternary. *Sediment. Geol.*, 23: 37–65.
- Thiede, J., 1983. Skeletal plankton and nekton in upwelling water masses off northwestern South America and

- northwest Africa. In: E. Suess and J. Thiede (Editors), Coastal Upwelling. Plenum, New York, A, pp. 183–207.
- Thunell, R.C., 1978. Distribution of recent planktonic foraminifera in surface sediments of the Mediterranean Sea. *Mar. Micropaleontol.*, 3: 147–173.
- Thunell, R.C. and Reynolds, L.A., 1984. Sedimentation of planktonic foraminifera: Seasonal changes in species flux in the Panama Basin. *Micropaleontology*, 30: 243–262.
- Thunell, R.C., Williams, D.F. and Kennett, J.P., 1977. Late Quaternary paleoclimatology, stratigraphy and sapropel history in eastern mediterranean deep-sea sediments. *Mar. Micropaleontol.*, 2: 371–388.
- Tolderlund, D.S. and Bé, A.W.H., 1971. Seasonal distribution of planktonic foraminifera in the western North Atlantic. *Micropaleontology*, 17: 297–329.
- Troelstra, S.R., Ganssen, G.M. and Klaver, G.T., 1990. High resolution analysis of eastern Mediterranean sapropel S<sub>1</sub>; chronostratigraphy, oxygen isotopes and geochemistry. IX Regional Committee on Mediterranean Neogene Stratigraphy Congress. (Barcelona, 1990.) p. 349.
- Troelstra, S.R., Ganssen, G.M., Van der Borg, K. and De Jong, A.F.M., 1991. A Late Quaternary stratigraphic framework for eastern Mediterranean sapropel S<sub>1</sub> based on AMS <sup>14</sup>C dates and stable oxygen isotopes. *Radiocarbon*, 33: 15–21.
- Van Leeuwen, R.J.W., 1989. Sea-floor Distribution and Late Quaternary Faunal Patterns of Planktonic and Benthic Foraminifera in the Angola Basin. *Utrecht Micropaleontol. Bull.*, 38, 288 pp.
- Vergnaud Grazzini, C., 1985. Mediterranean late Cenozoic stable isotope record: Stratigraphic and paleoclimatic implications. In: D.J. Stanley and F.C. Wezel (Editors), Geological Evolution of the Mediterranean Basin. Springer, New York, pp. 413–451.
- Vergnaud Grazzini, C., Ryan, W.B.F. and Cita, M.B., 1977. Stable isotopic fractionation, climatic change and episodic stagnation in the eastern Mediterranean during the late Quaternary. *Mar. Micropaleontol.*, 2: 353–370.
- Vergnaud Grazzini, C., Glaçon, G., Pierre, C., Pujol, C. and Urrutiaguer, M.J., 1986. Foraminifères planctoniques de Méditerranée en fin d'été. Relations avec les structures hydrologiques. *Mem. Soc. Geol. Ital.*, 36: 175–188.
- Zachariasse, W.J., Gudjonsson, L., Hilgen, F.J., Langeris, C.G., Lourens, L.J., Verhallen, P.J.J.M. and Zijderveld, J.D.A., 1990. Late Gauss to early Matuyama invasions of *Neoglobobulimina atlantica* in the Mediterranean and associated record of climatic change. *Paleoceanography*, 5: 239–252.
- Zachariasse, W.J., Jorissen, F.J., Perissoratis, C., Rohling, E.J. and Tsapralis, V., 1991. Increased river discharge during the deposition of sapropel S<sub>1</sub> in the Aegean Sea. In: E.J. Rohling (Editor), Late Quaternary changes in the eastern Mediterranean climate and hydrography and implications for the formation of sapropels. Ph.D. Dissert., Univ. Utrecht, pp. 147–154.
- Zhang, J., 1985. Living planktonic foraminifera from the Eastern Arabian Sea. *Deep-Sea Res.*, 32: 789–798.