Ocean climate variability in the eastern North Atlantic during interglacial marine isotope stage 11: A partial analogue to the Holocene?

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[1] Similar orbital geometry and greenhouse gas concentrations during marine isotope stage 11 (MIS 11) and the Holocene make stage 11 perhaps the best geological analogue period for the natural development of the present interglacial climate. Results of a detailed study of core MD01-2443 from the Iberian margin suggest that sea surface conditions during stage 11 were not significantly different from those observed during the elapsed portion of the Holocene. Peak interglacial conditions during stage 11 lasted nearly 18 kyr, indicating a Holocene unperturbed by human activity might last an additional 6–7 kyr. A comparison of sea surface temperatures (SST) derived from planktonic foraminifera for all interglacial intervals of the last million years reveals that warm temperatures during peak interglacials MIS 1, 5e, and 11 were higher on the Iberian margin than during substage 7e and most of 9e. The SST results are supported by heavier $\delta^{18}O$ values, particularly during 7e, indicating colder SSTs and a larger residual ice volume. Benthic $\delta^{13}C$ results provide evidence of a strong influence of North Atlantic Deep Water at greater depths than present during MIS 11. The progressive ocean climate deterioration into the following glaciation is associated with an increase in local upwelling intensity, interspersed by periodic cold episodes due to ice-rafting events occurring in the North Atlantic.

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1. Introduction

[2] Stage 11 (428–360 kyr B.P.) is the interglacial period often considered to be the closest analogue to the Holocene, both in orbital forcing conditions and degree of climate variability/stability. The importance of this interglacial for understanding climate changes lies in its insolation geometry. The orbital parameters during the early part of this isotope stage were very similar to those prevailing at the present day, with low eccentricity, high obliquity and low precessional amplitude [*Berger and Loutre*, 1991; *Loutre and Berger*, 2003]. The levels of greenhouse gases were also similar to the preindustrial Holocene, at least during its later part [*Petit et al.*, 1999; *EPICA Community Members*, 2004]. Deep-sea records suggest the persistence of warm interglacial conditions for longer than any other interglacial of the late Quaternary [*McManus et al.*, 1999; *Hodell et al.*, 2000; *Federici and McManus*, 2003]. This is considered to be controlled by the eccentricity modulation of precession: low eccentricity (more circular orbit) leads to a dampening of the precessional variations, resulting in the absence of extensive cold substages [*Berger and Loutre*, 1991; *Berger et al.*, 1996].

[3] In spite of recent studies of stage 11, several ambiguities remain. These are related to the magnitude of the warming associated with this interglacial and its possible influence on ice sheet reduction and sea level changes, and the role of thermohaline circulation on the preservation of warm conditions during such an unusually long period.

[4] Data from the Labrador Sea [*Aksu et al.*, 1992] suggest a more significant warming in stage 11 than during the Holocene, but in this particular work the Holocene is anomalously cold when compared even with previous interglacials such as stages 7 and 9. In the western part of the Nordic Sea, evidence for lower surface temperatures during stage 11 has been published by *Bauch et al.* [2000] and *Bauch and Erlenkeuser* [2003] and *Helmke et al.* [2003]. On the other hand northeastern Atlantic data suggest a warm early middle stage 11 not very dissimilar to the present interglacial in terms of the range of planktonic $\delta^{18}O$ values and sea surface temperature variation (1°–2°C) [*Oppo et al.*, 1998; *McManus et al.*, 1999, 2003].

[5] Modeling work [*Loutre and Berger*, 2003] indicates a reduction in global ice volume during the first 20 kyr of stage 11 but the geological evidence for higher sea level is somewhat controversial [*Rohling et al.*, 1998; *Bowen*,

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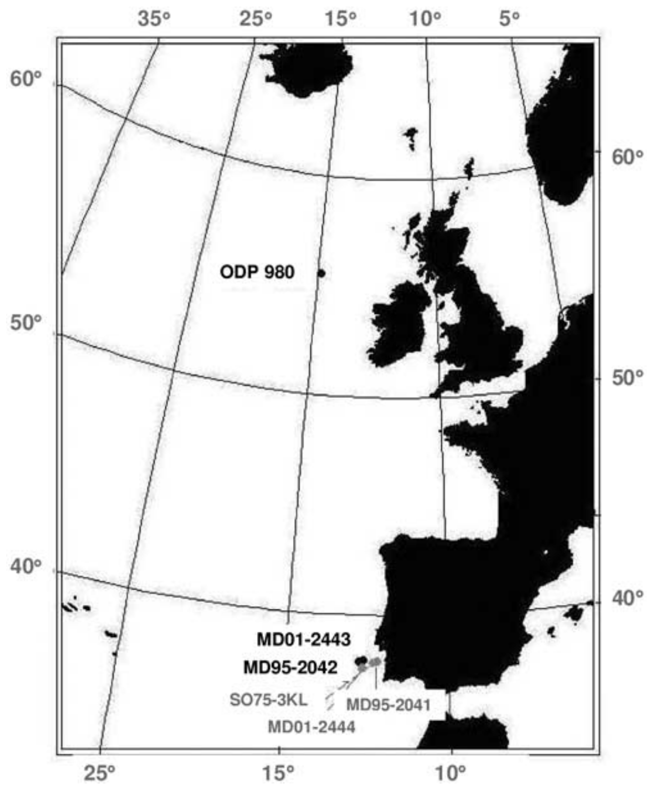

Figure 1. Location map of core MD01-2443 on the Iberian margin and Ocean Drilling Program Site 980 (Feni Drift) [*McManus et al.*, 2003]. Cores MD95-2042 [*Cayre et al.*, 1999; *Shackleton et al.*, 2000], MD01-2444 (N. J. Shackleton et al., unpublished results, 2003), SO75-3KL [*Schönfeld and Zahn*, 2000], and MD95-2041 (J. Schönfeld and R. Zahn, unpublished results, 2001) are plotted for comparison between stage 11 and earlier interglacials.

2003]. Several data sets indicate that sea level was at least thirteen [*Brigham-Grette*, 1999; *Bowen*, 1999] and possibly as much as 20 m [*Hearty et al.*, 1999; *Kindler and Hearty*, 2000] above the modern levels. The mechanism proposed for explaining such a sea level rise is a bipolar ice sheet disintegration, partially affecting the Greenland ice sheet [*Stanton-Frazee et al.*, 1999] and the West Antarctic ice sheet [*Scherer et al.*, 1998; *Scherer*, 2003]. Although a 20-m sea level rise cannot be accounted for by these two ice sheets alone, about 5 m could be derived from a partial melt of the East Antarctic ice sheet [*EPICA Community Members*, 2004]. The isotopic composition of carbonate shells and the air content record of the Vostok ice core [*Raynaud et al.*, 2003] do not support a significant melting of polar ice and suggest an ice volume (sea level) similar to the present day, an hypothesis corroborated also by diatom records from the East Antarctic circumpolar region [*Becquey and Gersonde*, 2002; *Gersonde and Zielinski*, 2000]. A vigorous thermohaline circulation during stage 11 may be one possible explanation for the apparently contradictory results. As postulated by *Oppo et al.* [1998] and *Poli et al.* [2000], this would increase the benthic isotopic gradients between the Atlantic and the Pacific basins, along with changes in the path of surface currents. The combined influence of factors such as temperature, salinity and $\delta^{18}O_{seawater}$ may have generated a climate record close in some aspects to the Holocene, but the exact mechanisms behind this partial similarity are not yet disclosed.

[6] As a contribution toward the improvement of palaeoclimatic reconstructions for the North Atlantic during stage 11, this paper aims to describe the palaeoclimatic evolution of stage 11 on the Iberian margin to point out regional differences and similarities relative to the Holocene record, and also to compare it with other interglacial intervals in the same region. Finally, we compare our new records from the Iberian margin to published records from ODP Site 980 (55°29′N, 14°42′W; 2189 mbsl) collected from the Feni Drift [e.g. *Oppo et al.*, 1998; *McManus et al.*, 1999, 2003].

2. Study Area

[7] Detailed micropalaeontological, sedimentological and geochemical data were obtained for the lower sections of core MD01-2443 (37°52.89′N, 10°10.57′W; 29.49 m long) (Figure 1), a Calypso giant piston core collected from the western Iberian margin at 2941 m water depth, during the MD123Geosciences Scientific Cruise on board the R/V *Marion Dufresne*. This core was collected as part of the "POP" Project (Pole-Ocean-Pole: A Global Stratigraphy for Millennial-Scale Variability). Previous work from this region has provided high-resolution records of ocean-climate variability, comparable to both Northern and Southern Hemisphere ice cores [e.g., *Shackleton et al.*, 2000; *de Abreu et al.*, 2003], and showing that climate variability occurred in a systematic way through interhemispheric teleconnections.

[8] Off Iberia, distinct eastern boundary seasonal surface circulation regimes are governed by meridional displacements of the Azores anticyclonic cell and its associated large-scale wind pattern [e.g., *Fiúza*, 1984; *Fiuza et al.*, 1998]. During summer months, coastal upwelling of deep cold, nutrient-rich waters occurs, in association with the northward displacement and strengthening of the Azores high-pressure cell and the weakening of the Iceland Low [*Fiúza*, 1984]. The strong northerlies along the Portuguese coast are one of the factors responsible for the onset of the upwelling, which may extend from 30–50 km to 100–200 km offshore (e.g., E. D. Barton, Eastern boundary of the North Atlantic–northwest Africa and Iberia, course notes from Advanced Study Course on Upwelling Systems, 15 pp., University de Las Palmas de Gran Canaria, Las Palmas de Gran Canaria, Spain, 1995). Eastern North Atlantic Central Water (ENACW) is the source of the upwelled water. It consists of two branches with diverse origin and distinct thermohaline characteristics: the Subtropical Eastern North Atlantic Central Water, formed along the Azores Front mainly during the winter, and the Subpolar Eastern North Atlantic Central Water, which flows under the subtropical ENACW and is derived from the descending branch of the North Atlantic Drift [*Fiúza*, 1984; *Fiuza et al.*, 1998]. This summer circulation pattern features a group of

mesoscale structures such as jets, meanders, eddies and upwelling filaments. The latter features are not simply superficial structures and have clear biological and chemical signatures. During the rest of the year, in particular during winter months, coastal convergence conditions prevail, and the most relevant transport mechanism is the northward flowing warm undercurrent, which takes the designation of Portugal Coastal Counter Current [e.g., *Fiuza et al.*, 1998]. The Poleward Current advects subtropical (and partly Mediterranean Water) northward, adding warm and saline inputs into the study region [*Haynes and Barton*, 1990].

[9] Intermediate circulation is dominated by North Atlantic Central Water (NACW) and the Mediterranean Outflow Water (MOW). The present influence of MOW has long been documented in the vicinity of the Portuguese margin, not only following a poleward route, but also propagating toward the west in the form of "meddies." MOW is represented off Iberia into two main branches: An upper core is centered around 700–800 m, and a lower core is at about 1200 m [*Ambar and Howe*, 1979]. Below Mediterranean water, at about −2000 m, deep water masses are controlled by the interplay between North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW), which currently circulates in the eastern Atlantic basin at depths greater than 4000 m [*Emery and Meinke*, 1986; *van Aken*, 2000]. At the present day, the cored site is under the influence of NADW.

3. Methods

3.1. Stable Isotope Measurements

[10] In order to assess variations in the surface and near-surface water mass properties, stable isotope measurements have been performed at the Godwin Laboratory on five species of planktonic foraminifera for the interval between 360–420 ka, which includes most of stage 11. These taxa represent a set of surface (*Globigerinoides ruber*), to subsurface (*Globigerina bulloides* and dextral *Neogloboquadrina pachyderma*) and deep dwellers (*Globorotalia inflata* and dextral *Globorotalia truncatulinoides*). To avoid size-related offsets, samples containing an average of 30 planktonic foraminifera were picked from the sediment fraction larger than 250 μm and prepared following the standard procedure described by *de Abreu et al.* [2003]. Depending on the sample size, they were analyzed using different mass spectrometers and preparation systems. The larger samples were reacted with 100% orthophosphoric acid at 90°C using a VG Isotech Isocarb common acid bath system, and the carbon dioxide produced during this reaction was then analyzed by a VG Isotech SIRA-Series II mass spectrometer. For the smaller samples (less than 100 μg) analyses were performed using a PRISM mass spectrometer with a Micromass Multicarb sample preparation system. Results were calibrated to VPDB (Vienna PeeDee belemnite) using the repeated analysis of an internal carbonate standard (Carrara Marble). The analytical long-term precision was better than 0.08‰ for oxygen and 0.06‰ for carbon. The standard deviation of the analyses of replicate samples is 0.14‰ for $\delta^{18}O$ and 0.08‰ for $\delta^{13}C$.

[11] Stable isotope analysis of benthic foraminifera was carried out so as to detect ventilation changes in the deep-sea environment, associated with variations in the production of NADW. For the benthic $\delta^{18}O$ record several different species were analyzed because of the relative scarcity of benthic specimens in the sediments of core MD01-2443. Where possible, two or three separate measurements of different species were made per sample; a correction factor was applied according to the species and the average of all the corrected values at each level is shown in Figure 2d (and Figure 6a). A composite record constitutes the final result. The following species were analyzed and calibrated to *Uvigerina peregrina* as indicated: *Cibicidoides robertsonianus*, +0.50; *Uvigerina peregrine*, 0.0; *Globobulimina affinis*, −0.30; *Cibicidoides wuellerstorfi*, +0.64; *Cibicidoides kullenbergi*, +0.51; *Hoeglundina elegans*, −0.60; *Oridorsalis umbonatus*, 0.0; and *Cassidulina carinata*, 0.0. For the benthic $\delta^{13}C$ record, we used the values for *Cibicidoides* spp. However, the values for endobenthic genera were not used because these do not maintain a constant $\delta^{13}C$ offset with respect to the overlying deep water. To directly compare the benthic $\delta^{18}O$ records of the cores discussed in this paper, a calibration to *Uvigerina peregrina* of 0.64‰ was also applied to the published measurements of site 980 [*McManus et al.*, 2003].

3.2. Planktonic Foraminifera Fauna and Sea Surface Temperature Reconstruction

[12] The identification of the planktonic foraminiferal species in the sediment fraction larger than 150 μm is based on the work of *Kennett and Srinivasan* [1983]. Twenty-five different species of planktonic foraminifera have been identified all belonging to the living fauna in the area [*Duprat*, 1983; *Levy et al.*, 1995; *Martins and Gomes*, 2004]. Species relative abundance data were used to identify six distinctive assemblages based on species distributions in surface water masses and surface sediments [e.g., *Kipp*, 1976; *Chaisson et al.*, 2002]. The six assemblages are polar, subpolar, temperate/transitional, subtropical, tropical and gyre margin. Sea surface temperature (SST) was estimated from the planktonic foraminifera using a modern analogue technique, SIMMAX 28, as described by *Pflaumann et al.* [1996], but using the updated 2004 version, with a larger data set with 947 samples.

4. Chronological Framework

[13] The chronology employed in this paper for core MD01-2443 (Table 1) was developed by aligning the benthic $\delta^{18}O$ record to the Antarctic Vostok deuterium record [*Petit et al.*, 1999] as described by *Tzedakis et al.* [2004]. This was based on the implications of the work of *Shackleton et al.* [2000] who showed the similarity between benthic $\delta^{18}O$ records off Portugal and temperature over Antarctica. The improved Vostok timescale used as the basis for our comparison has been developed by F. Parrenin (personal communication, 2004) through alignment with the European Programme for Ice Coring in Antarctica (EPICA) Dome C record [*EPICA Community Members*, 2004]. The

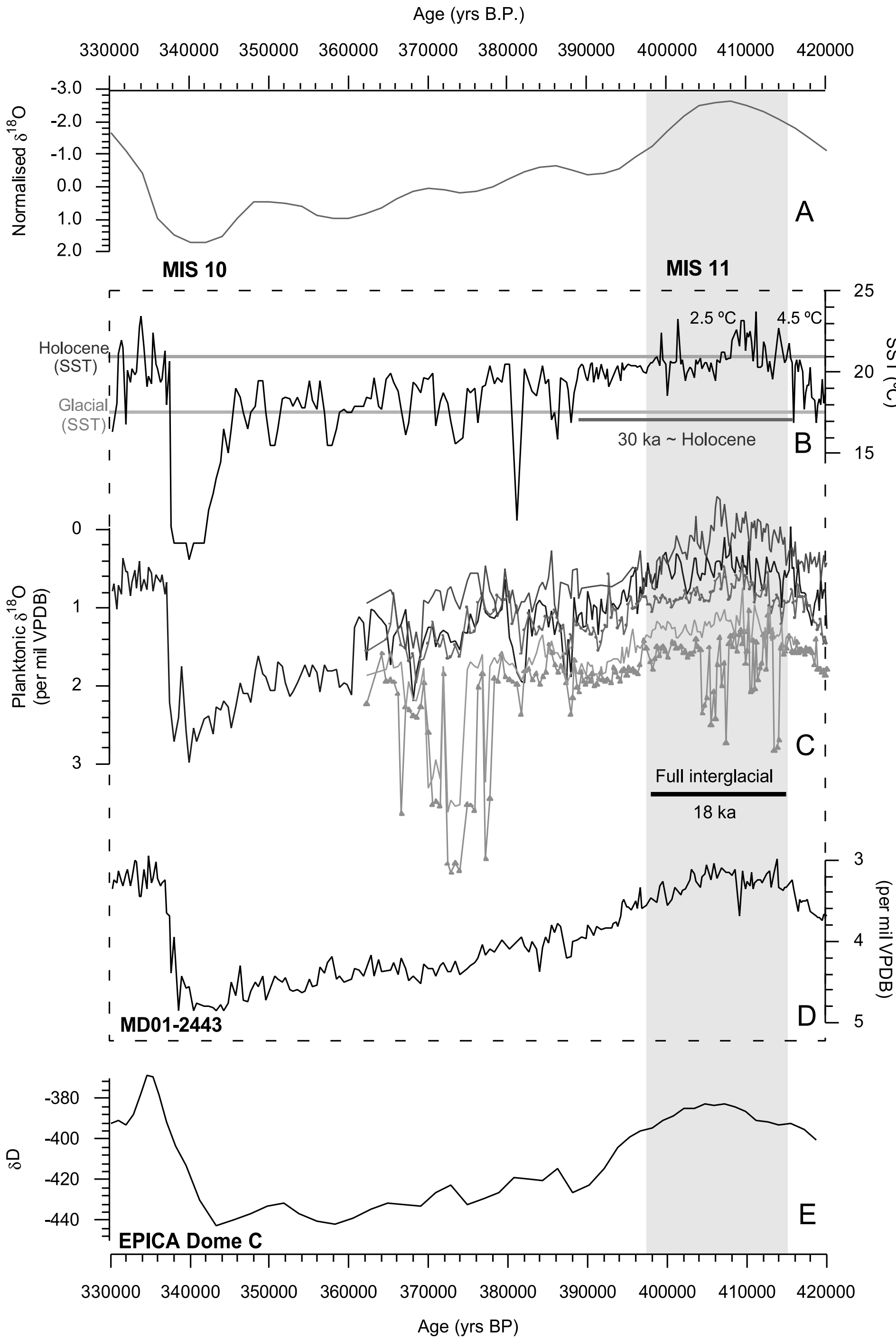

Figure 2. Down-core results for stages 10 and 11 from MD01-2443. (a) Bassinot normalized curve [*Bassinot et al.*, 1994]. (b) Estimated summer sea surface temperature. (c) Multispecies planktonic $\delta^{18}O$ records (red, *Gs. rubber*; dark blue, *G. bulloides*; green, dextral *N. pachyderma*; grey, *Go. inflata*; and light blue, *Go. truncatulinoides*). (d) Benthic $\delta^{18}O$ based on the average between measurements for each sample and corrected for their respective equilibrium offsets [*Shackleton and Hall*, 1995; *Shackleton et al.*, 2000]. (e) European Programme for Ice Coring in Antarctica (EPICA) Dome C deuterium record as given by *EPICA Community Members* [2004]. Grey shaded area marks the period where full interglacial conditions were maintained. See color version of this figure at back of this issue.

Table 1. Chronological Model Presented for Core MD01-2443 (for Stage 11)[a]

Core Depth, cm	Age, years B.P.	Proxies Used for Correlation	Source
2468	338,000	benthic $\delta^{18}O$ and D/H Vostok[b]	*Petit et al.* [1999]
2660	385,170	benthic $\delta^{18}O$ and D/H Vostok[b]	*Petit et al.* [1999]
2676	387,000	benthic $\delta^{18}O$ and D/H Vostok[b]	*Petit et al.* [1999]
2948	420,000	benthic $\delta^{18}O$ and Dome C D/H	*EPICA Community Members* [2004]

[a]Sediment rates are calculated between time control points; the two extremities per run between temporal controls were not considered.
[b]Vostok records were tuned to the new EPICA Dome C record, so that both are comparable within the same time frame.

base of MD01-2443 was defined by comparison with the EPICA Dome C deuterium record [*EPICA Community Members*, 2004] as this extends in time the previously available Vostok records. Mean linear sedimentation rates for this core were estimated, assuming a constant accumulation rate between dated levels. The sampling at 2-cm intervals gives an average resolution of 330 years for stage 11 (Figure 2).

[14] For comparison, the independently created chronology of ODP Site 980 [*McManus et al.*, 1999] was revised by tuning Termination V to the astronomically calibrated low-latitude stack [*Bassinot et al.*, 1994], consistent with the time interval defined through $^{40}Ar/^{39}Ar$ dating for Termination V [*Karner and Morra*, 2003] and not very far from the limit attributed to the new Antarctic ice core record [*EPICA Community Members*, 2004].

5. Results

5.1. Benthic Isotopic Measurements

[15] Benthic $\delta^{18}O$ measurements indicate that the corer did not penetrate into MIS 12 at site MD01-2443. Near the beginning of our record, the benthic $\delta^{18}O$ decreases from 3.7‰ to 3.1‰, indicating a reduction in ice volume (Figure 2). According to the present age model, the period of lowest ice volume, and thus establishment of peak interglacial conditions lasted nearly 7–8 kyr, until ~402 ka, with short-term variations of ~0.2‰. The next dominant feature is the long-term increase in $\delta^{18}O$ (up to a maximum of 3.5‰), with a superimposed higher-frequency variability from 396 ka onward toward glacial stage 10, coincident with the growth of ice sheets. This is indicated by an increase of nearly 1.4‰ in the benthic $\delta^{18}O$ values up to the start of the glacial. In stage 10, a further increase is recorded off the Iberian margin, reaching the highest benthic isotope ratio of 4.8‰ between 341 and 345 ka.

5.2. Multispecies Planktonic Isotopic Record

[16] The interpretation of the stable isotopic composition of foraminiferal shells in upwelling areas, such as the Iberian margin, is difficult and often hampered by the dynamic character of such settings [e.g., *Peeters et al.*, 2002]. In such a complex hydrological regime, the differences in the signal between species must not only be interpreted taking into consideration both their sequence of development, but the contemporaneous differences between distinct habitats. Since there are currently few sediment trap or water column data for the Iberian margin available, we support our observations with ecological data from adjacent North Atlantic regions, as well as with the behavior described for some species in other upwelling areas [e.g., *Levy et al.*, 1995; *Thunell and Sautter*, 1992; *Peeters et al.*, 2002]. Future sediment trap data from ESF Euromargins project SEDPORT might shed light on the microfauna seasonal variability.

[17] Most of the planktonic species chosen for this comparison characterize distinct seasons [e.g., *Levy et al.*, 1995; *Schiebel and Hemleben*, 2000]: *Gs. ruber* is a warm season, surface dweller, while *G. bulloides* occurs in the upper 50 m of the water column, associated with the spring/summer upwelling system when the degree of mixing of the upper water column is high. Recent observations indicate that *Gs. ruber* can also be found during the upwelling season (F. Peeters, personal communication, 2003) but unlike *G. bulloides* it survives in the distal areas of the upwelled water. Toward the end of the upwelling season, the dextral *N. pachyderma* appear in great numbers, when the influence of warmer Azores Current waters increases off Portugal [*Rogerson et al.*, 2004]. This species has been observed at depths of up to 250 m. Deep dwellers *Go. inflata* (100–250 m) and dextral *Go. truncatulinoides* (250–300 m up to 1000 m) were selected with the purpose of detecting possible changes in the structure of the deep thermocline [e.g., *Hemleben et al.*, 1989; *Abrantes et al.*, 2001; *Matzumoto and Lynch-Stieglitz*, 2003].

[18] Bearing in mind the complexity and the limitations of this approach, the combination of the monospecific profiles of $\delta^{18}O$*ruber*, $\delta^{18}O$*bulloides*, $\delta^{18}O$*pachyderma*, $\delta^{18}O$*inflata* and $\delta^{18}O$*truncatulinoides* (Figure 2) show several distinct, previously unreported features. Through an interval of circa 20 kyr (from ~416 to 396 ka), the oxygen isotope values from intermediate-depth dwellers generally did not vary more than 0.25–0.28‰, which is equivalent to approximately 1°C. However, some excursions of up to 0.5‰ occurred, suggesting occasional greater sea surface temperature changes or even possible salinity anomalies at the Iberian margin. In contrast, surface dweller *Gs. ruber* displays a significant shift toward heavier values from 402 ka, which became more accentuated after 396 ka. The lack of a comparable initial increase in intermediate to deeper dwellers suggests some type of seasonal or depth habitat compensating effect.

[19] The beginning of our sequence (at 420 ka) up to the peak interglacial conditions as defined by the lightest $\delta^{18}O$ *Gs. ruber* values (416 to 402 ka) suggest a well-defined upper water column stratification, as surface, intermediate and deep dwellers show progressively higher isotopic ratios. Surface dweller *Gs. ruber* is typically lighter throughout stage 11 (−0.3 to 0.2‰), and the same trend is followed by the intermediate dwellers in particular up to the end of the peak interglacial. The deeper dwellers exhibit a clearly

heavier oxygen isotope ratio (by as much as 0.8 to 1‰). The deep-living foraminifera *Go. truncatulinoides* dextral shows a trend similar to the other deep species, but with several rapid excursions in $\delta^{18}O$ of over 1‰, centered at 413 ka and 405 ka, respectively. These isotopic excursions during the progression to peak interglacial conditions, although not seen in any other record, coincide with a change in the *Go. truncatulinoides* population toward the dominance of left coiling forms, which preferentially live in cold waters and typically display heavier isotopic ratios. It is then possible that we have different genotypes throughout our stage 11 samples, that because of morphological convergence are not fully distinguishable under the binocular microscope [*de Vargas et al.*, 2001], and this in turn may account for the observed isotopic shifts. With the decline of interglacial conditions, an increase in the levels of upper water column mixing is suggested by the smaller range in the $\Delta\delta^{18}O$ between surface/intermediate and deeper dwellers. The larger amplitude variability which characterizes the progression toward a new glacial (stage 10) is one of the main features of the record. *G. bulloides* and *N. pachyderma* display very similar $\delta^{18}O$ values (between 0.9 and 1.2‰), occasionally also close to *Go. inflata*, with the exception of three prominent excursions toward heavier values of *G. bulloides*, respectively at 388, 381 and 369 ka.

[20] The hydrological complexity of this region appears to be more evident during the cool intervals of stage 11. The $\delta^{18}O$ of dextral *N. pachyderma* show the same trend as data from *G. bulloides* and, in most warm periods, to *Go. inflata*, possibly as a function of the ice volume effect on seawater composition. However, during cool intervals they show lighter $\delta^{18}O$ values, with differences of up to 0.4‰. One possible explanation for this feature lies in the narrowly overlapping season that these two species exist off Iberia and the possibility of a dual surface water mass influence at the site of MD01-2443. Assuming that *G. bulloides* usually thrives in cooler waters (being a subpolar species and an upwelling form), an isotopically heavier signal might be evidence of upwelling. If dextral *N. pachyderma*, which dominates toward the end of the upwelling season, thrives more under the influence of a warmer water mass of subtropical origin – derived from the Azores Current [e.g., *Fiúza*, 1984] then it records the cool events in a more attenuated form. On the other hand, living toward the Autumn/Winter months, the decrease in dextral *N. pachyderma* $\delta^{18}O$ may be related to periodic salinity anomalies.

[21] At the end of stage 11 both deep dwellers, *Go. inflata* and dextral *Go. truncatulinoides*, show much more significant variations, reaching over 2‰. As such variations are not recorded in the remaining species, and given that the left coiling variety of *Go. truncatulinoides* is once again dominant, we believe there is sufficient evidence to implicate a change in the ecological conditions and the additional influence of an oligotrophic intermediate water mass during this period [e.g., *Peeters et al.*, 2002; *Matzumoto and Lynch-Stieglitz*, 2003]. Further studies of these particular sediment samples will shed some light into the possibility that these heavier values are due to the presence of secondary calcite, formed at deeper water levels, and thus corresponding to colder conditions.

5.3. Faunal Changes

[22] Grouped into faunal assemblages (Figure 3) [e.g., *Bé*, 1977; *Ottens*, 1991], the 25 different species of planktonic foraminifera show evidence of three main surface environmental "modes" within stage 11. The development of progressively warmer sea surface conditions is shown by the presence of an increasing number of tropical (36–40%) species toward the peak interglacial (Figure 3), coinciding with the lightest planktonic $\delta^{18}O$ values. The succession of warm water species include *N. dutertrei* (up to 1.5%) which is considered a postupwelling species, and can be used as a proxy for the stabilization of the upper water column (following trap data of *Thunell and Honjo* [1983] and *Thunell and Sautter* [1992] as a potential analogue).

[23] From ~402 ka onward, the warm water planktonic fauna was progressively replaced by cold subtropical, transitional and even subpolar species (such as *G. bulloides* and dextral *N. pachyderma*) that dominate throughout the remaining interglacial interval (50%), and become progressively more important toward stage 10. The decrease in tropical species, after 396 ka, agrees with the enrichment in both *Gs. ruber* $\delta^{18}O$ and benthic $\delta^{18}O$, marking the end of full interglacial conditions and the gradual decline of climate.

[24] A third mode of faunal succession is detected in the periodic increase of the polar species left coiling *N. pachyderma*, which at times reaches 23% of the planktonic population. Increases in transitional/temperate species (by over 10%) indicate the recovery of sea surface conditions up to an interglacial mode and strong oscillation in the position of the subpolar front. A closer look into the variation of individual species reveals an extra degree of complexity, which points toward distinct trophic situations as the interglacial progressed. Species usually associated with the seasonal upwelling system, such as *T. quinqueloba* and *G. bulloides* show a progressive increase toward the end of stage 1l, suggesting an enhanced upwelling and increased nutrient levels. Additionally, the increasingly high abundance of dextral *N. pachyderma* (from 10% at the peak interglacial to over 40% at the end of stage 11), usually associated with the end of the upwelling events, may be indicative of a more continuous degree of mixing of the upper water column. This hypothesis is supported by the presence of *Orbulina bilobata*, considered an indicator of high nutrient levels [e.g., *Robbins*, 1988; *Hemleben et al.*, 1989], the convergence of some of the multispecies planktonic $\delta^{18}O$ records, and an increase in the organic carbon by ~0.5%.

5.4. Sea Surface Temperature Changes

[25] As indicated by the planktonic isotope record, the sea surface temperature reconstructions suggest that this long interglacial (stage 11) evolved in three distinctive phases on the Iberian margin. The summer sea surface temperature (SST) rise from deglacial conditions to the interglacial maximum was about 4.5°C. This rise occurred over a period

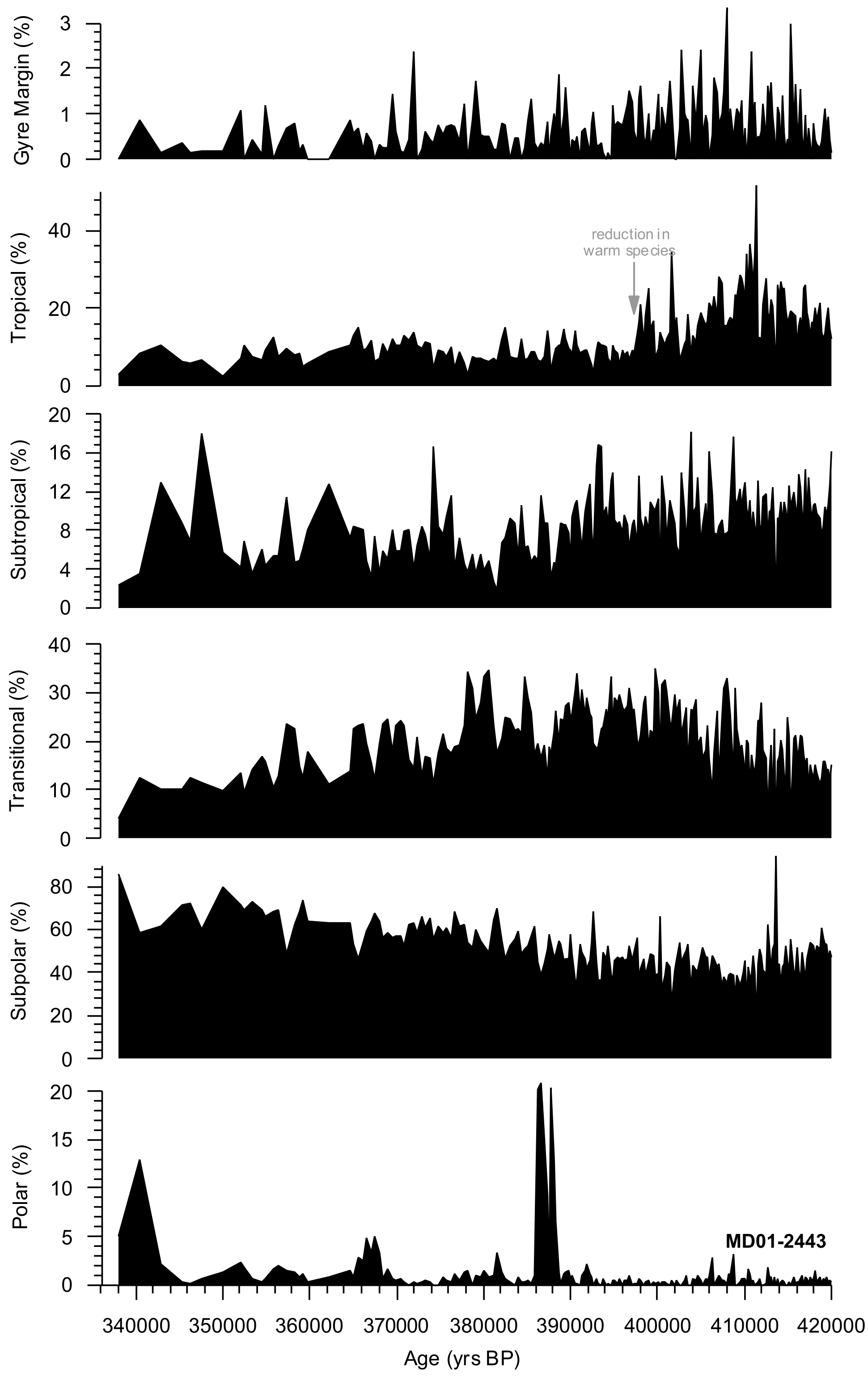

Figure 3. Relative abundance of the different planktonic foraminifera assemblages throughout stage 11 in core MD01-2443.

of nearly 5–6 kyr but appears to have been disrupted several times, in particular during the interval of highest insolation (between 408 and 412 ka). Peak interglacial temperatures (SSTs around 20.5°–23.3°C) occurred between 412 and 406 ka. However, if the average Holocene temperature (21°C) for this region is considered as a threshold, the duration of this interglacial phase lasted for nearly 27 kyr (from 416 up to 389 ka) in good agreement with previous estimates of *McManus et al.* [2003] and *Bauch and Erlenkeuser* [2003].

[26] Only after 389 ka do the surface records start to show a greater magnitude of variability, with sporadic abrupt SST decreases of up to 7°C. These are values close to the average value found for the last glacial period (17°C) in this region (Figure 2). However, the average value may not be the best way to compare the two records; in fact, both the temperature estimates and the periodic pattern of large cooling events resemble the palaeoceanographic changes described for stages 2 and 3 in nearby cores [e.g., *Cayre et al.*, 1999; *Bard et al.*, 2000; *de Abreu et al.*, 2003]. The pronounced SST decrease during peak stage 10 was comparable to the surface water cooling recorded during Heinrich events 2, 3 and 5, whereas SSTs reached even more extreme values (with differences of up to 2°C) during Heinrich events 1 and 4 [*Cayre et al.*, 1999].

6. Discussion

6.1. Iberian Margin: Regional Expression of Interglacial Stage 11

[27] High-frequency excursions are more evident in the records of surface δ^{18}O and foraminifera-derived SSTs than in the benthic δ^{18}O record, since they are directly dependent on surface hydrology, in particular on temperature and salinity. Regarding the magnitude of these variations in surface water indicators, stage 11 can be roughly interpreted in terms of three distinct phases: (1) the deglacial interval which extended up to the establishment of full interglacial conditions at ~416 ka; (2) the peak interglacial interval where the range of oscillations in the planktonic record was small, reaching up to 0.1–0.25‰, and changes in SST were of the order of 1°–2°C; and (3) the progressive climate deterioration, starting at ~396 ka when excursions in δ^{18}O began to gradually increase in amplitude, surpassing the variability recorded in the early part of stage 11, and the transition into MIS 10. Both the planktonic foraminiferal fauna (in particular the significant decrease in tropical species), the surface dwelling *Gs. ruber* isotope record, and the benthic δ^{18}O record show the end of the full interglacial conditions and new ice volume increase. This occurred, according to the present age model, nearly 7 kyr before a significant planktonic enrichment was recorded by intermediate-depth dwelling species (Figure 2).

[28] The discrepancy between planktonic isotopic values from surface and deeper dwellers suggests some kind of compensating effect, either related to the different seasonal distribution of the selected species, or perhaps a function of their distinct depth habitat. Species that are usually associated with stronger mixing conditions and thus greater homogenization of the water column, such as *G. bulloides* and *N. pachyderma*, display a restricted amplitude of the isotopic anomalies in response to hydrological events, and may indicate the presence of a dampened NACW signal. These faunal successions could also be influenced by the migration of the subpolar front southward and/or the strengthening of the gyre/Eastern Boundary Current, with the Portugal Current transporting more water to the south during the spring/summer season. On the other hand, *Gs. ruber* is the first planktonic species recording climatic change possibly of subtropical origin, as it calcifies in the Azores current.

6.2. Comparison With Succeeding Interglacials: Role of Ventilation Changes and Influence of Southern Ocean Waters

[29] A comparison of peak interglacial planktonic δ^{18}O and SST records of stage 11 with those of younger warm intervals (substages 9e and 7e from core MD01-2443, and substage 5e and the Holocene along nearby core MD95-2042) reveals pronounced warm conditions, with stages 1, 5e and 11 generally 1°C warmer than the maximal SSTs obtained for substage 7e and most of 9e (Figure 4). Faunal analysis shows a distinct pattern in the assemblage composition during peak interglacials, in particular during substages 7e and 9e, where along with a temperate to subtropical fauna, subpolar species are abundant. The results for substages 7e and 9e gain further support from both the planktonic and benthic δ^{18}O records (Figure 4) which show heavier values, in particular during substage 7e, indicating colder surface temperatures and a larger global ice volume, when compared to other interglacials [*McManus et al.*, 1999]. This agrees with *Kandiano and Bauch* [2003] in the sense that the closest interglacials in terms of absolute temperature are the Holocene, substage 5e and stage 11, with SSTs being slightly higher in the latter.

[30] In addition, benthic δ^{18}O values in stage 11 are lighter by up to 0.3‰ than the average Holocene isotopic ratio [*Shackleton et al.*, 2000], which might correspond to a smaller ice volume and/or a potentially warmer deep ocean [e.g., *McManus et al.*, 2003]. Overall estimates for both interglacials agree well within the error of our calculation (Figure 4). These proxy data alone suggest that the ice volume/sea level was not significantly different than at present, assuming that the ice volume was the single factor affecting the isotopic record.

[31] The comparison between benthic δ^{13}C records from the Holocene and stage 11 presented in Table 2 can give some clues toward a better understanding of the influence of water masses with distinct characteristics in this region. In agreement with other published North Atlantic records [e.g., *Venz et al.*, 1999; *Flower et al.*, 2000], the compiled results are interpreted to indicate an enhanced contribution of high-δ^{13}C NADW during peak interglacial 11 at intermediate and deep sites in the northeast Atlantic basin, versus a stronger influence of less ventilated deep waters at 37°N during the Holocene. Similarities in the behavior of some of the palaeoceanographic proxies used in this paper may thus result from a changing influence of distinct sourced intermediate and deep water masses.

[32] Expanding our geographic coverage, a comparison between the high-resolution benthic isotope records from sites MD01-2443 and ODP 980 indicates the presence of water masses with uniform characteristics at intermediate and deepwater eastern North Atlantic sites during stage 11 (Figures 5 and 6). Maximum $\delta^{13}C_{benthic}$ values of ~1.3‰, higher than Holocene records from neighboring cores [*Shackleton et al.*, 2000; *Schönfeld and Zahn*, 2000; *Schönfeld and* Zahn, unpublished results], agree within

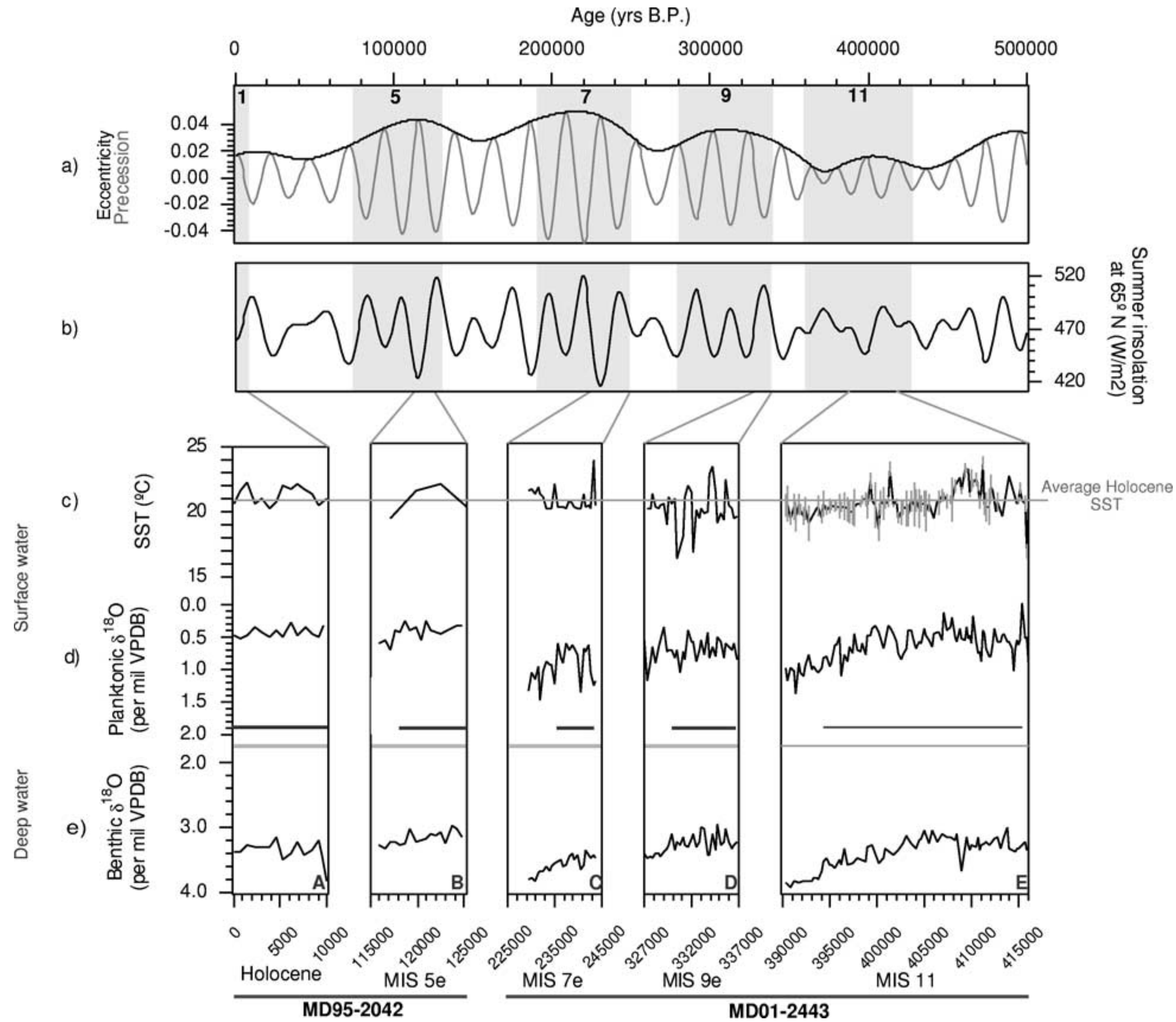

Figure 4. Comparison of the stage 11 records from MD01-2443, with the Holocene (labeled A) and substage 5e (labeled B) data from nearby core MD95-2042, substage 7e (labeled C) and substage 9e (labeled D) from MD01-2443. (a) Eccentricity and precession curves for the last 500 ka. (b) Summer insolation at 65°N [*Berger and Loutre*, 1991]. (c) Estimated summer sea surface temperatures (SST). (d) Planktonic $\delta^{18}O$ records from *G. bulloides*. (e) Benthic $\delta^{18}O$ values corrected to *Uvigerina peregrina*. Proposed age model–dependent duration of each peak interglacial marked by a grey segment.

±0.2‰ with results from Site 980 taken as a closer reference end-member for NADW [*Venz et al.*, 1999; *Flower et al.*, 2000]. Almost no difference between both $\delta^{13}C_{benthic}$ records exists during most of the long warm interval (~27 kyr), indicating the maintenance of a strong thermohaline circulation throughout this period. This way, northern ocean waters would have spread their influence to deeper sectors of the eastern North Atlantic, whereas at present there is the dual influence of two different deep water masses, as indicated by the ~0.3–0.4‰ difference in $\delta^{13}C$ between the two sites.

[33] During the establishment of interglacial conditions, upper NADW would have had a strong contribution from Labrador Seawater (LSW) and possibly MOW, which could have compensated for the lower ventilation at intermediate and deep ocean levels. The North Atlantic Current transported warm, salty surface waters into the Nordic Seas and benthic $\delta^{13}C$ values in the deep North Atlantic increased with an increased production of NADW and its spreading to deeper waters, therefore replacing the influence of Southern Ocean Waters, as seems to be pointed out by the record of MD01-2443. Several short-term, $\delta^{13}C$ depletion events of ~0.2‰ were recorded that occasionally reached values lower than those seen during stage 3 at the Iberian margin [e.g., *de Abreu et al.*, 2003]. These episodes of reduced ventilation seem to be associated with periods of ice volume decrease throughout stage 11 (Figures 2d and 6b) as suggested by the benthic $\delta^{18}O$ record, indicating an inter-

Table 2. Comparison of Benthic $\delta^{13}C$ Values Between the Holocene and Stage 11 From Different Depth Iberian Margin Cores

Core	Latitude	Longitude	Depth, m	Range of Benthic $\delta^{13}C$, ‰	Time Period	Source
MD95-2042	37.811	−10.150	3146	0.4–0.92	Holocene	*Shackleton et al.* [2000]
MD01-2444	37.574	−11.468	2656	0.85–1.09	Holocene	N. J. Shackleton, unpublished results (2003)
SO75-3K	37.920	−9.900	2430	0.4–1.0	Holocene	*Schönfeld and Zahn* [2000]
MD95-2041	37.833	−9.518	1123[a]	0.96–1.09	Holocene	J. Schönfeld and R. Zahn, unpublished results (2001)
MD01-2443	37.891	−10.183	2941	0.9–1.3	stage 11	this paper

[a]Records mainly Mediterranean Outflow Water.

ference with deepwater production/circulation. Cold events detected during the transition between stages 11 and 10 are also associated with benthic $\delta^{13}C$ depletions of 0.4–0.5‰ in MD01-2443, indicative of a more diluted influence of NADW at this site. These shifts are more attenuated on the Iberian margin than at Site 980, which is closer to the deepwater source region. The shift to higher amplitudes occurs when the ice volume threshold suggested by *McManus et al.* [1999] is passed ($\delta^{18}O_{benthic}$ of ~4.1‰). Beyond this value climate variability seems to be amplified by feedback mechanisms. The progressive enrichment of the benthic $\delta^{18}O$ records may derive from a composite influence of ice volume increase and a renewed cooling of the deep oceanic environment [*McManus et al.*, 1999]. The increase in the isotopic gradient between our two different depth sites and the changes in $\delta^{13}C_{benthic}$ during late stage 11 and glacial stage 10 are consistent with the redistribution of deep and intermediate water masses, associated with a distinct chemical stratification.

[34] During stage 10, the benthic record of ODP Site 980 shows an enrichment in $\delta^{18}O$, suggestive of the presence of colder deep water during the glacial interval, in a similar fashion to the variability described during the last glacial period [*Skinner et al.*, 2003], and the decrease in poleward heat flux. The slower glacial circulation may have contributed to the enhancement of the thermal/isotopic gradient between these sites. However, once again we cannot ignore that these sites would be affected by the combined, variable influence of two distinct main deep water masses.

6.3. Duration and Variability of Interglacial Intervals

[35] In core MD01-2443, a deglacial phase leading toward peak interglacial values occurred during the first 4–5 kyr of the record. Bearing in mind that by comparison with Site 980 we may be missing a sediment record equivalent to nearly 7 kyr, our estimates for the length of this deglacial warming are not far from independent modeling results from *Loutre and Berger* [2003], who proposed that interglacial conditions were achieved after a period of ~20 kyr. According to our timescale, a convincing relationship between the two sites is observed, with full interglacial conditions prevailing at the Iberian margin for ~18 kyr (Figure 5), surpassing the length of the elapsed portion of the Holocene. The foraminifera-derived SSTs point to the maintenance of the mild sea surface conditions for a longer period (of nearly 27 kyr) in agreement to previously published work [e.g., *McManus et al.*, 1999] but the high temperatures estimated may derive from the dominance of subtropical and transitional species immediately after surface and deep records point to the end of interglacial conditions. See Figure 6.

[36] The observed differences between the last five interglacial periods are probably a reflection of the distinct orbital configurations and different insolation values that characterize each one of these warm intervals (Figures 4a and 4b). For example, while there is a very similar insolation geometry between stage 11 and the Holocene, the insolation parameters for substage 5e show large differences in terms of total insolation and amplitude. These results suggest that, on the basis of comparison to stage 11, the Holocene may last another 6–7 kyr.

[37] In addition to insolation, another important mechanism for maintaining the warm conditions in the North Atlantic would be to have a strong thermohaline circulation [e.g., *Oppo et al.*, 1998; *McManus et al.*, 1999] which continuously supplies warm waters to the northern regions during a period when insolation levels were low [*McManus et al.*, 2002]. A third factor that needs to be studied further is the variable influence of differently sourced deep water masses to the deep ocean environment, and whether or not their properties were similar to modern days.

[38] In stage 11 evidence points to gradual ice volume decay up to ~402 ka, and characterized by a transient high-frequency variability. In the case of Termination II, the benthic $\delta^{18}O$ shows a very rapid shift (less than 5 kyr) toward lighter values, very similar to those of stage 11. Future work should focus on the possibility either that deep water was slightly warmer than during the Holocene and substage 5e, or that sea level was a few meters above present, implying less ice on Antarctica and/or Greenland. In addition, work is needed to establish the distribution of ice that generated the IRD that started to appear at about 390 ka.

Figure 5. Expression of northern Atlantic ice-rafting events off Iberia in comparison with the Feni Drift. (a) Relative abundance of left coiling *N. pachyderma* and ice-rafted debris (IRD) in Site 980 [e.g., *Oppo et al.*, 1998; *McManus et al.*, 1999]. (b) The $\delta^{18}O$ records of dextral *N. pachyderma* from Site 980. (c) Onboard magnetic susceptibility (plotted on reverse scale). (d) Total carbonate content. (e) Organic carbon. (f) Estimated summer sea surface temperatures. (g) Relative abundance of left coiling *N. pachyderma*. (h) The $\delta^{18}O$ record of *G. bulloides*.

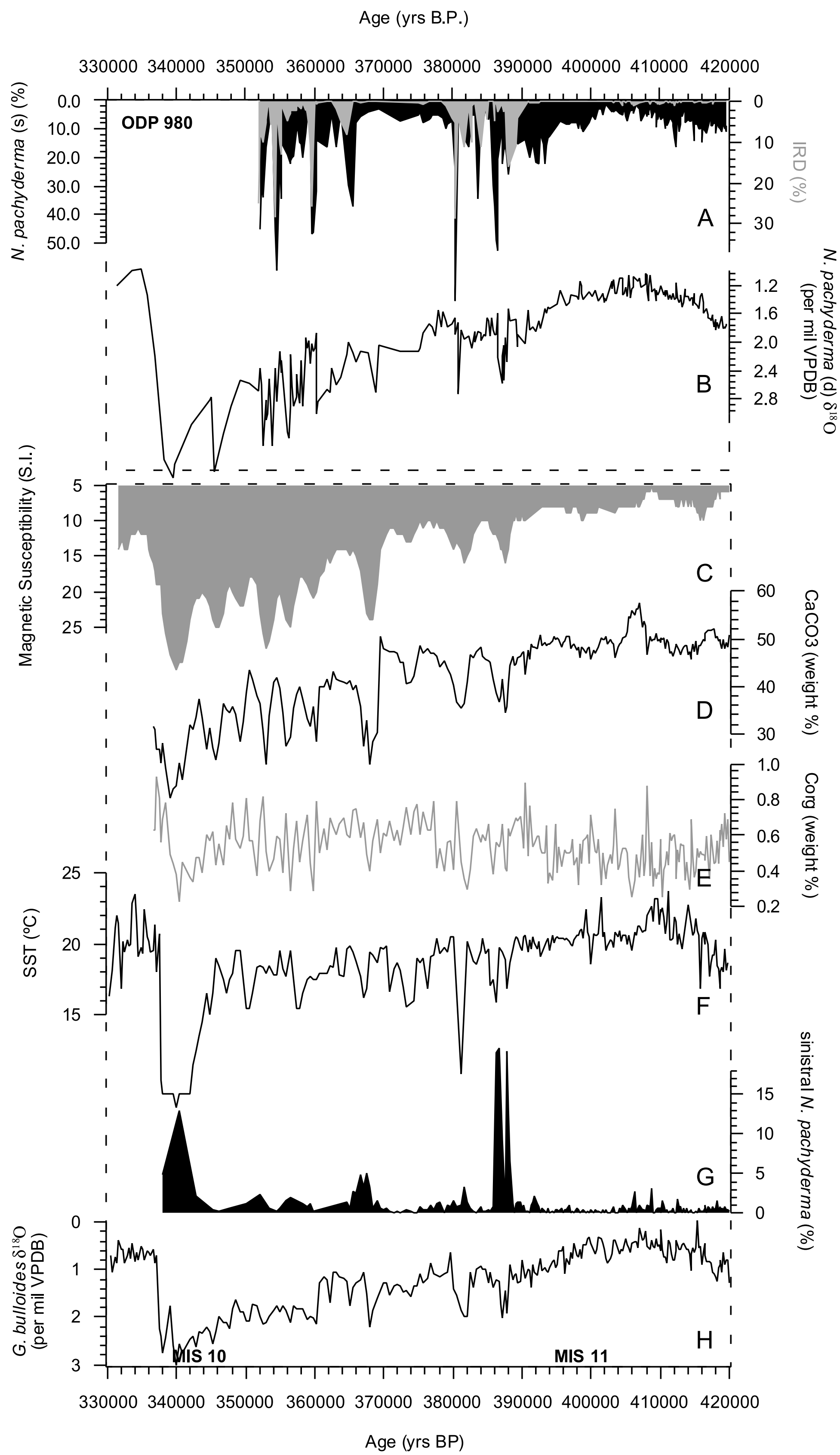
Age (yrs B.P.)
330000 340000 350000 360000 370000 380000 390000 400000 410000 420000
ODP 980
N. pachyderma (s) (%)
IRD (%)
A
N. pachyderma (d) δ18O (per mil VPDB)
B
Magnetic Susceptibility (S.I.)
C
CaCO3 (weight %)
D
Corg (weight %)
E
SST (ºC)
F
sinistral N. pachyderma (%)
G
G. bulloides δ18O (per mil VPDB)
H
MIS 10
MIS 11
Age (yrs BP)

Figure 5

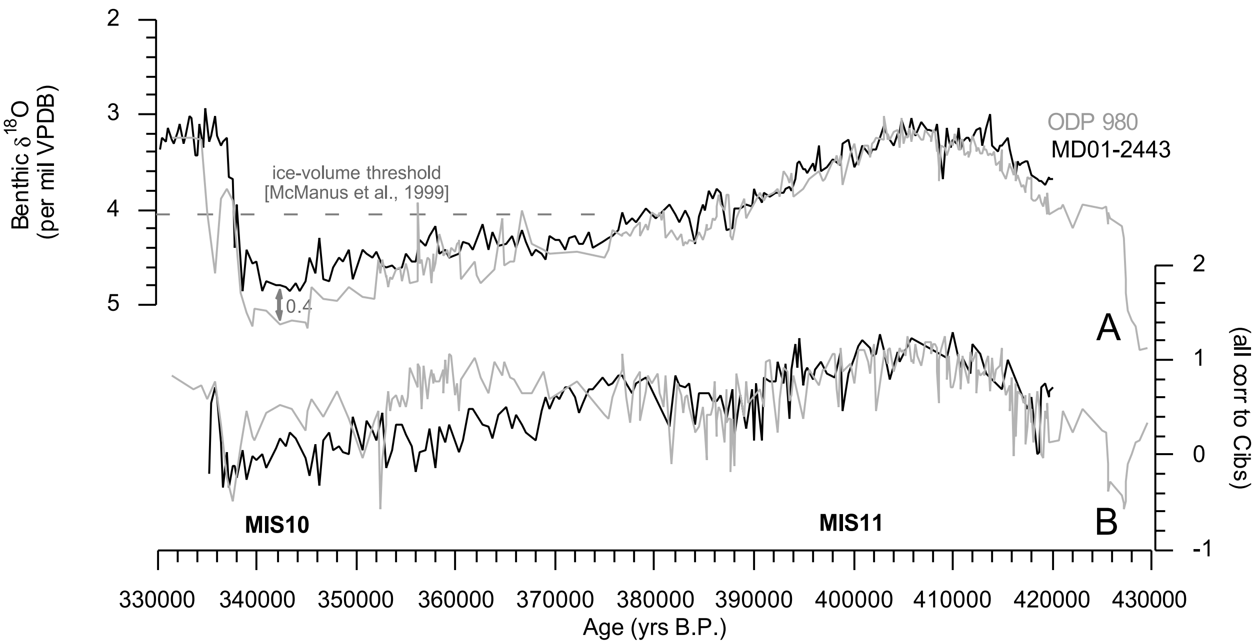

Figure 6. Benthic isotope records from MD01-2443 and ODP Site 980. (a) Similarity of $\delta^{18}O$ values throughout MIS 11. (b) Establishment of a distinct geochemical stratification ($\delta^{13}C$) between the two sites once the ice volume threshold [*McManus et al.*, 1999] was surpassed.

6.4. Rapid Cooling Events at the End of Interglacial Stage 11

[39] In contrast to other peak interglacials, stage 11 experienced a long gradual cooling, inferred from the steadily increasing subtropical, temperate and subpolar species (Figure 3). Several rapid cooling events (ranging from 4°–8°C) were detected in the progressive transition from stage 11 into glacial stage 10 (Figure 3). These coolings reached temperatures lower than the regional Last Glacial Maximum (LGM) estimates of 16°–21°C from foraminifera-derived reconstructions [e.g., *Cayre et al.*, 1999] and 14°–15°C for alkenone-derived temperature [*Pailler and Bard*, 2000].

[40] In a process similar to that described for the last glacial period, the greater amplitude of the surface thermal and isotopic variability seems to be associated with the effects of minor deglaciation observed in the benthic isotopic records (Figure 2). This can be linked with the diluted influence of cold water off the Iberian margin during periodic collapse of portions of the North Hemisphere ice sheets [e.g., *Zahn et al.*, 1997; *Schönfeld et al.*, 2003; *de Abreu et al.*, 2003].

[41] The influence of ice rafting detected at the Feni Drift [*Oppo et al.*, 1998] at the end of stage 11 can be roughly followed up to the latitude of core MD01-2443 (Figure 5). Their Iberian equivalents cannot be defined by the presence of coarse ice-rafted particles (IRD) as during the last glacial [*Bard et al.*, 1987; *Skinner et al.*, 2003], but the existence of quartz grains in the sediment fraction larger than 90 μm along with volcanic particles (obsidian and basalt lithoclasts) during periods of climate deterioration within late stage 11 was noted. The deposition of finer sediment is particularly marked on the shipboard magnetic susceptibility record (Figure 5c) which shows several peaks coinciding with the main periods of rapid SST decrease. These periods are also marked by reduction in the very high total carbonate content (Figure 5d), in part related to dilution by the fine lithics transported up to 37°N. During full interglacial conditions, the total carbonate content reaches values of over 50%, higher than the typical Holocene range of 32–38% for nearby core MD95-2042 [*Pailler and Bard*, 2000].

7. Summary

[42] The similarity of surface water conditions during stage 11 and the Holocene inferred from northeastern Atlantic cores was only partially confirmed by the Iberian margin sediments, which show that this was not an "extreme" interglacial in absolute terms. Stable isotope data for the Holocene and stage 11 show similar values for both the surface and deep ocean. In light at the hypothesis that global ice volume was reduced for part of stage 11, one possible explanation for the similarity of the isotopic values is this is a result of a different combination of water mass temperature, salinity and seawater isotopic composition. Only when we manage to sort out the variability in these factors will we have a better understanding of the climate evolution during stage 11, and its true duration.

[43] Off the Iberian margin our multiproxy approach suggests three upper water column palaeoenvironmental modes: (1) from ∼416 to 402 kyr, a warm, generally well stratified water column; (2) from ∼396 kyr toward stage 10, a trend toward slightly colder SSTs with an associated increase in the accumulation of organic matter and a

convergence of the oxygen isotope ratios of *G. bulloides*, dextral *N. pachyderma* and occasionally *Go. inflata*, indicating a possible increase in local upwelling intensity and stronger trade winds; and (3) periodic cold events, bringing SSTs nearly to local glacial levels (Heinrich events), with the lowest temperatures associated with the presence of polar and subpolar species. This third mode, characteristic of the later part of stage 11 coincides with the ice-rafting episodes detected for the same interval at 55°N (ODP Site 980), and shows that even during an interglacial interval North Atlantic ice-rafting events can have a detectable expression on the sensitive Iberian margin palaeoenvironment.

[44] Interglacial duration was recorded differently by surface and intermediate dwellers, because of possible compensation mechanisms related to seasonal or depth distribution. Thus the selection of proxies used to characterize full interglacial conditions is important in order to establish a more accurate comparison with the Holocene. Indications from both the surface and benthic stable isotope data are that the interglacial conditions were preserved for a period of 18 kyr. However, intermediate water dwellers display a separate behavior, with light isotope ratios recorded for a period of nearly 20–27 kyr. A third proxy, foraminifera-derived SST, extends the estimated duration of warm surface conditions to nearly 27 kyr, suggesting that the modern analogue technique may be too sensitive to the presence of a high percentage of subtropical and temperate species after 396 kyr, when the tropical assemblage suffers a reduction of nearly 20%.

[45] In a broader geographic context, no deep water mass isotopic gradient during stage 11 was detected between MD01-2443 and ODP Site 980, and the $\delta^{18}O_{benthic}$ and $\delta^{13}C_{benthic}$ records show very similar values and trends, indicating the influence of a deep water mass of uniform characteristics. These results suggest that NADW would have been present at intermediate and deep levels, but the dynamic equilibrium between different sourced deep water masses must have been kept for most of stage 11. The greater variability displayed by the $\delta^{13}C$ record of ODP Site 980 may point to changes in the source and/or properties of NADW, in a similar way to the one observed for the Holocene [e.g., *Skinner et al.*, 2003]. An increase in bathymetric $\delta^{13}C$ gradient by <0.4‰ between Site 980 and MD01-2443 during late stage 11 and glacial stage 10 indicates a distinct combination of UNADW, decrease in LNADW and stronger influence of nutrient-rich SOW. Oscillations in deepwater production were synchronous with the main changes in sea surface temperature in the eastern North Atlantic.

[46] One important conclusion that stands out from our work is the necessity to compare and interrelate similar reconstructions from different areas of the world, not just in palaeoceanographic terms, but as an expression of a coupled influence of the atmospheric circulation and ocean water masses. Only via a larger study, both geographically and with the use of complementary proxies focused on the evolution of the two main Atlantic deep water masses during this long interglacial, will we be able to evaluate the still controversial apparent analogy to the elapsed portion of the Holocene.

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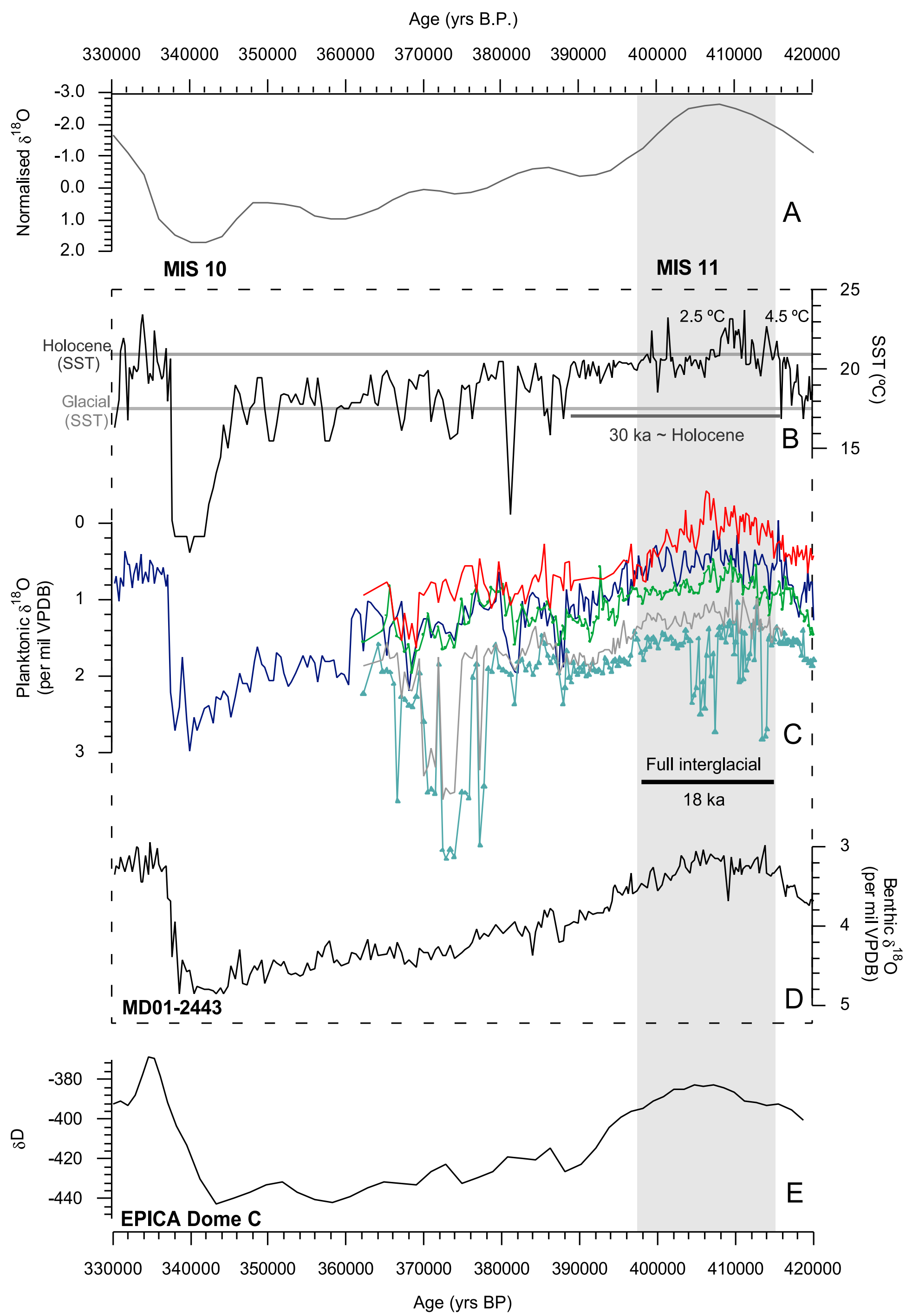

Figure 2. Down-core results for stages 10 and 11 from MD01-2443. (a) Bassinot normalized curve [*Bassinot et al.*, 1994]. (b) Estimated summer sea surface temperature. (c) Multispecies planktonic $\delta^{18}O$ records (red, *Gs. rubber*; dark blue, *G. bulloides*; green, dextral *N. pachyderma*; grey, *Go. inflata*; and light blue, *Go. truncatulinoides*). (d) Benthic $\delta^{18}O$ based on the average between measurements for each sample and corrected for their respective equilibrium offsets [*Shackleton and Hall*, 1995; *Shackleton et al.*, 2000]. (e) European Programme for Ice Coring in Antarctica (EPICA) Dome C deuterium record as given by *EPICA Community Members* [2004]. Grey shaded area marks the period where full interglacial conditions were maintained.