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Millennial surface water dynamics in the Ría de Vigo during the last 3000 years as revealed by coccoliths and molecular biomarkers

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Abstract

A combined study of coccolith assemblages and biomarkers in a gravity core collected from the Ría de Vigo (NW Spain) allowed us to reconstruct the paleoenvironmental conditions for the last 3000 years. The quantitative distribution of coccolithophore species points to three different intervals within the core, dated by AMS radiocarbon measurements. The first interval (ca. 975 BC–252 AD), characterized by high abundances of *Calcidiscus leptoporus* and *Gephyrocapsa muelleriae*, is thought to represent moderate water temperatures, suggesting a transition from a warmer to a cooler period. The second interval (ca. 252–1368 AD), characterized by the dominance of *Coccolithus pelagicus*, *Helicosphaera carteri* and *Syracosphaera* spp., and a high concentration of hexacosanol linked to terrestrial input, is interpreted as having been a humid period with fluvial input. The third interval (ca. 1368 AD–1950) is characterized by a high abundance of *Gephyrocapsa oceanica*, high alkenone values and low values of hexacosanol, and is thought to represent a period dominated by oceanic conditions within the Ría.

Taking into account the ocean–atmospheric system affecting the region studied, here we propose an alternation in the mean state of North Atlantic Oscillation (NAO) at millennial time scales. A well-developed upwelling system and an active Ría–ocean connection during the warmer interval I suggest a NAO+ phase influenced by a Hypsithermal period. The occurrence of the humid and relatively warm interval II is consistent with a negative phase in the NAO, as well as a relative restriction in ocean–Ría exchange. Interval III, which was drier and more productive, again suggests the dominance of a positive phase in the NAO, with a more intense oceanic connection and more energized upwelling.

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Keywords: North Atlantic; Ría de Vigo; Coccoliths; Coccolithophores; Molecular biomarkers; Alkenones; Holocene; Hypsithermal; North Atlantic Oscillation

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1. Introduction and oceanographic setting

A *ría* is a river valley that has been invaded by the sea (Derruau, 1983). The Ría de Vigo is located on the Spanish Atlantic coast and is the southernmost Ría of a group known as the “Rías Bajas” (Fig. 1). It is oriented along a central axis, direction N45°E, and occupies an area of 176 km². Along the major axis (33 km) is a central channel, with a maximum depth of 45 m at the mouth. The San Simón Inlet is the narrowest, most inland portion of the Ría (Fig. 1). This inlet is connected through the narrow (600 m) Rande Strait to the Ría. The Cies Islands, to the west, form a natural barrier to the sea, resulting in relatively calm conditions in the Ría de Vigo (Vilas et al., 1995; Fig. 1).

The Ría de Vigo overlies a bedrock that was heavily fractured during the Hercynian orogeny (IGME, 1981). These tectonic dynamics, together with subsidence and the Flandrian transgression (Derruau, 1983), resulted in the fiord-like morphology of the Rías. These characteristics are now seen in a complex system in which ocean and continent interact, leaving a mixed signature of both environments in the sediments of the Ría.

The Ría de Vigo area is affected by the East North Atlantic Water (ENAW) circulation pattern. This anticyclonic circulation produces an upwelling system at the mouth of the Ría, resulting in lower surface water temperatures, higher nutrient contents, and high primary production (Prego, 1993; García-Gil et al., 1995). The water dynamics inside the Ría de Vigo follow an estuarine circulation pattern, with a deep current flow into the Ría, and surface water export out of it (García-Gil et al., 1995). Overprinting this general pattern, a seasonal wind system affects this estuarine circulation. During summer, an anticyclonic wind circulation linked to the Azores High belt reinforces the estuarine pattern (Álvarez-Salgado et al., 1993), with an enhancement of upwelling due to stronger northern winds. Conversely, during winter, the southern position of the Atlantic Low belt, defines a dominant southerly wind regime. At this time, the interchange between the Ría and the ocean is either diminished or interrupted.

The dominant upwelling phases produce a mixture in the water column. Conversely, when upwelling is weakened, circulation between the Ría and the open ocean decreases, and the waters become stratified

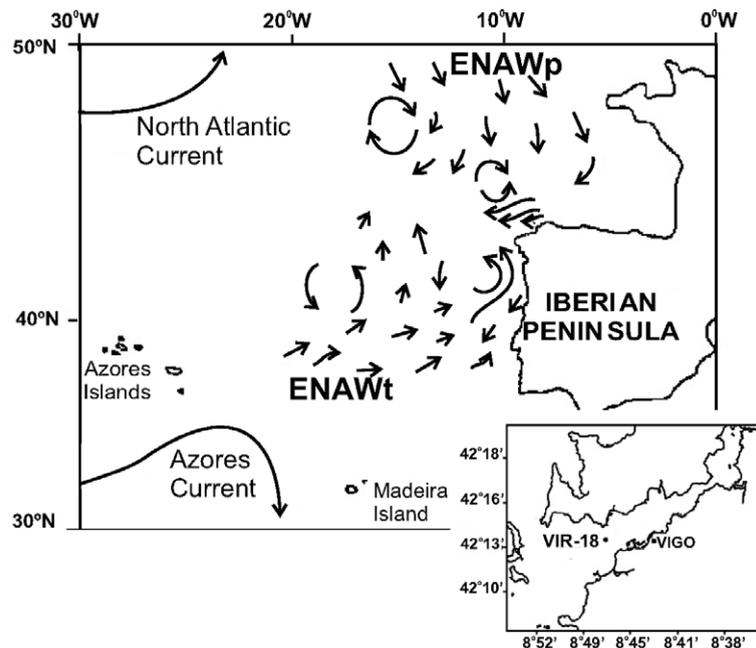


Fig. 1. Situation of core VIR-18 and general pattern of the Eastern North Atlantic Water (ENAW). ENAWp=polar origin; ENAWt=tropical origin.

(García-Gil et al., 1995). Modern river discharge has also introduced anthropogenically-produced nutrients from fertilizers that can produce peaks of higher productivity, although always in minor proportions as compared to the above process. Today, the sewer outlet of the City of Vigo accounts for most of the organic carbon delivered into the Ría de Vigo (Prego, 1993).

The Ría de Vigo is a high-productivity area where diatoms, dinoflagellates, coccolithophores and other microorganisms are abundant. Here we focus on coccolithophores: planktonic organisms surrounded by calcium carbonate scales. Their preservation in the sediment provides an excellent record for characterizing surface water-masses. Assemblages of this group respond to changes in environmental conditions and, together with biogeochemical analyses, provide a tool for reconstructing paleoenvironmental conditions. In the present study we combine the characterization of coccoliths and molecular biomarkers from a sediment core with the aim of elucidating changes in the surface water dynamics of the Ría de Vigo for the last 3000 years, monitoring sea–continent interactions in this particular transitional area.

2. Materials and methods

Core VIR-18 (380 cm long) was collected in 1990 with a “vibrocorer” from the central part of the Ría de Vigo (42°14.07N, 8°47.37W) at a water depth of 45 m (Fig. 1). Dark olive-grey clay and silt with bioclastic fragments of mollusks (mainly bivalves and gastropods) are the main constituents of these sediments. Short intervals of bioturbation are the only sedimentary structures observed. Samples for micropaleontological and biochemical analyses were taken systematically every 5 cm, providing a temporal resolution of about 30 years on average.

2.1. Age model

In this study we used the age model determined by Diz et al. (2002). The chronology is based on two Accelerator Mass Spectrometry (AMS) radiocarbon measurements: one in the 226–228 cm interval (ca. 907–890 AD) on bivalve shells (*Venus* sp.) found in

the same position as when they were alive, and the other at the core bottom, 380 cm, where bioclastic fragments are present. These analyses were carried out at the Paleobotany and Paleolimnology Laboratory of the University of Utrecht. Radiocarbon data were converted into calibrated ages using the Calib 4.3 radiocarbon calibration program (Stuiver et al., 2000, based on Stuiver and Reimer, 1993) (Table 1). The ages between these data points were obtained by linear interpolation.

2.2. Smear slide preparations

Smear slides were prepared using sediment solution decantation. The solution was distributed uniformly in Petri dishes by pumping it in and out several times with a micropipette (in order to generate a slight circulation of water in the dish). A coverslip was placed at the bottom of each dish and decantation was performed. The fluid was then withdrawn from each dish. After the dishes had been dried, smear slides were assembled with Canada balsam (Flores and Sierro, 1997).

For quantitative analyses, we used a light polarizing microscope (1250×). Around 500 coccoliths were counted per slide. To obtain percentages and absolute abundances we followed the technique of Flores and Sierro (1997). Additional Scanning Electron Microscope analyses were carried out on selected samples to pinpoint taxonomical aspects and to observe preservation features.

2.3. Coccolith preservation

The coccolith preservation in the samples analysed can be considered good to moderate (see for example Roth and Thierstein, 1972; Flores and Marino, 2002). Individual observations on the ring elements of small

Table 1
AMS ¹⁴C ages (modified after Diz et al., 2002)

Depth (cm)	Radiocarbon date (years BP)	Calibrated date	Calibrated age
226–228	1502±27	881 (902) 946 years AD	1069 (1048) 1004 years BP
380	3151±32	1010 (976) 9235 years BC	2959 (2925) 1872 years BP

placoliths and other fragile coccoliths allowed us to conclude that only a small degree of etching affected the assemblage within the VIR-18 core. In all cases, the dissolution features did not preclude taxonomical identification.

2.4. Molecular biomarkers

Variations in selected molecular biomarker abundances through the sedimentary record have often been used to assess the relative importance of the different organic matter sources over time. Amongst the compounds most studied are the C_{37} alkenones, which allow the establishment of paleo-sea surface temperatures (SST) by means of the U_{37}^K index (initially defined by Brassell et al., 1986 and later simplified as $U_{37}^{K'}$ by Prahl et al., 1988). These haptophyte-derived compounds are also frequently interpreted in terms of paleo-marine primary production (e.g. Villanueva et al., 1997a; Budziak et al., 2000). Other biomarkers of relevance are the higher plant-derived long chain alkanes and alcohols (Eglinton and Hamilton, 1967), which help elucidate changes in continental supply to the sediments (e.g. Ikehara et al., 2000; Calvo et al., 2001).

In this study, we present previously published data (Diz et al., 2002) on alkenone and *n*-hexacosan-1-ol concentrations as well as SSTs derived from the U_{37}^K index generated using published methods (Villanueva

et al., 1997b). Translation of $U_{37}^{K'}$ ratios into SST was achieved using the global core-top calibration equation of Müller et al. (1998) ($U_{37}^{K'} = 0.033 \text{ SST} + 0.044$).

3. Results

3.1. Coccolithophore assemblage

The coccolithophore assemblage of core VIR-18 is mainly represented by eight taxa: *Calcidiscus leptoporus*, *Gephyrocapsa muelleriae*, *Coccolithus pelagicus*, *Helicosphaera carteri*, *Syracosphaera* spp., *Gephyrocapsa oceanica*, *Emiliania huxleyi* and the group “small” *Gephyrocapsa* (this group includes species with a maximum coccolith diameter of $<3 \mu\text{m}$; *Gephyrocapsa aperta* and *Gephyrocapsa ericsonii*). The “small” *Gephyrocapsa* group (more than 50% of the total coccolith abundance) and *E. huxleyi* (27%) are the dominant taxa. *G. oceanica* reaches 7%, whereas the other species fluctuate between 1% and 2% (Fig. 2). The downcore accumulation rate of each taxon is shown in Fig. 3. *Calcidiscus leptoporus* and *G. muelleriae* have similar profiles and the maximum accumulation rate of both taxa occurs between 380 and 280 cm (ca. 975 BC–252 AD). From 280 cm to the top of the core the accumulation rate of both species diminishes. The trend of *E. huxleyi* is similar to those of

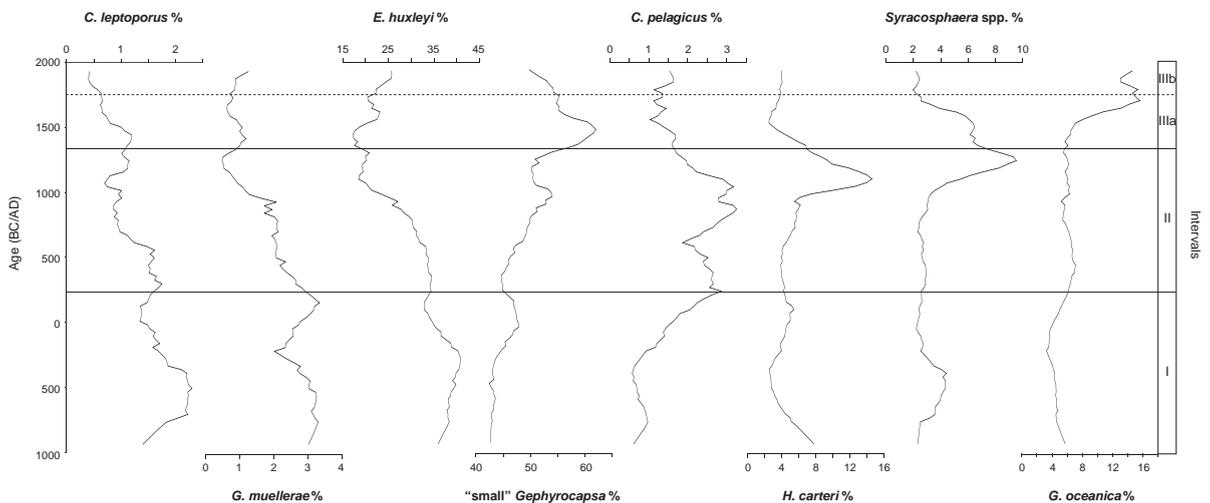


Fig. 2. Percentages of selected coccolithophore taxa in core VIR-18. For definition and characterization on intervals see text (Results section).

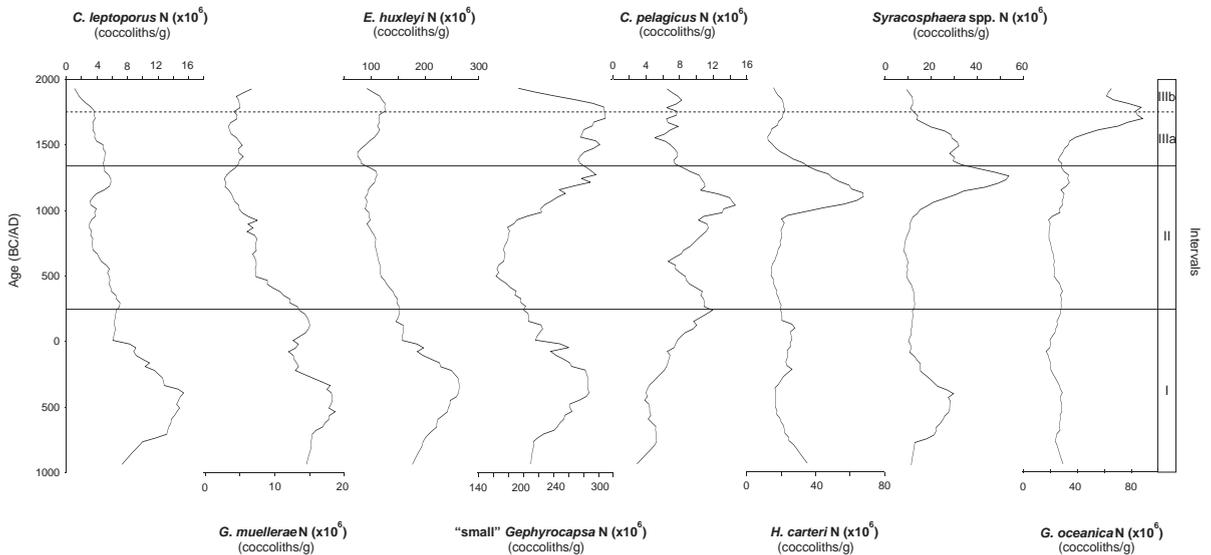


Fig. 3. Coccolithophore accumulation rates (N) of selected taxa identified in core VIR-18. For definition and characterization on intervals see text (Results section).

the previous two species, although its accumulation rate is higher. The profile of the “small” *Gephyrocapsa* group shows two peaks, one in the lower part of the core and the other close to the core top; the lowest accumulation rate occurs in the central part. The accumulation rates of *C. pelagicus*, *H. carteri* and *Syracosphaera* spp. are low in the lower and upper parts of the core, while the maximum accumulation rates of these species occur within the interval between 280 and 125 cm (ca. 252–1368 AD). Interestingly, each of these species reaches a maximum at different times within this period in the following sequence: first, *C. pelagicus* peaks at ~1000 years AD, followed by maxima of *H. carteri* at ~1100 years AD and *Syracosphaera* spp. at ~1300 years AD, the accumulation rate of *G. oceanica* was low throughout the core except for an abrupt increase starting from 125 cm (ca. 1368 AD).

In order to compare our data with reconstructions based on other proxies, the core was divided into three intervals, taking into account the total abundance of coccolithophores as well as the record of selected species (Figs. 3 and 4). Interval I is defined from the core bottom to 280 cm (ca. 975 BC–252 AD), and is characterized by a high abundance of *Calcidiscus leptoporus* and *Gephyrocapsa muelleriae*. Interval II, from 280 to 125 cm (ca. 252–1368 AD), corresponds

with the lowest values in total abundance of coccoliths and highest values in the proportion of *Coccolithus pelagicus*, *Helicosphaera carteri* and *Syracosphaera*. Finally, interval III, from 125 cm to the core top (ca. 1368 AD–1950), is defined by moderate values in the total abundance of coccoliths and a dominance of *Gephyrocapsa oceanica* in the upper part; this variation in interval III allowed us to separate two different subintervals, IIIa and IIIb, respectively (Fig. 3).

4. Ecological interpretation of the coccolithophore assemblages

The coccolithophores are autotrophic organisms living in the euphotic layer. Assemblage structure is controlled by different factors, such as water, light intensity, water temperature, nutrient content and salinity (Brand, 1994; Young, 1994). In this section we summarize the ecological characteristics of the most significant species identified in this study on the basis of data provided by different specialists.

Emiliana huxleyi is a cosmopolitan and highly eutrophic species (Roth, 1994). When *E. huxleyi* appears together with “small” *Gephyrocapsa* (*Gephyrocapsa ericsonii* and *Gephyrocapsa aperta*) as part of

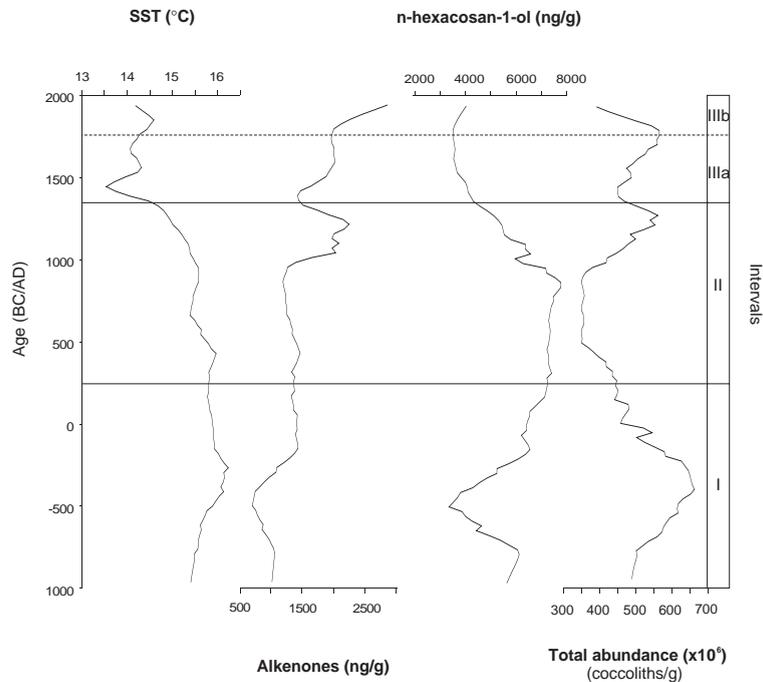


Fig. 4. Estimated paleotemperature, molecular biomarker concentrations, and coccolith total abundances (per visual field, $\times 1250$) vs. age (years BC/AD). For definition and characterization on intervals see text (Results section).

the coccolith assemblage, it is considered as an upwelling proxy (Wells and Okada, 1997).

Calcidiscus leptoporus is a tropical species usually living between 20 and 30 °C in oligotrophic conditions (Giraudeau and Rogers, 1994), whereas *Gephyrocapsa muelleriae* is a cold Atlantic species (Knappertsbusch et al., 1997; Flores et al., 1997) equivalent to the “*Gephyrocapsa* Cold” morphotype of Bollmann (1997). Also, *G. muelleriae* is related to subarctic cold waters with mean temperatures lower than 20 °C and with moderately productive surface waters (Bollmann, 1997).

Coccolithus pelagicus is a cold water species (McIntyre and Bé, 1967; Okada and McIntyre, 1979), although other factors such as a constant input of nutrients in a moderate turbulence regime may control the presence or abundance of this species (Cachão, 1995; Cachão and Moita, 2000). *Helicosphaera carteri* preferentially occurs in tropical and subtropical waters (Okada and McIntyre, 1979) but can also be found in high-productivity waters (Pujos, 1992; Flores et al., 1995), whereas *Syracosphaera* spp. prefers warm and stratified waters (Jordan et al.,

1996). In the Gulf of Cadiz Colmenero-Hidalgo et al. (2004) observed that increases in *H. carteri* and *Syracosphaera* spp. are related to the input of turbid and fresher waters.

Gephyrocapsa oceanica is recorded in upwelling areas, or is at least relatively abundant in stratified waters (Winter, 1982), and is equivalent to the “*Gephyrocapsa* Equatorial” morphotype of Bollmann (1997), strongly correlated with water temperatures above 25 °C.

5. Discussion

5.1. Interval I: 380–280 cm (ca. 975 BC–252 AD)

This interval is characterized by the highest abundance of coccoliths. The dominant species are *Emiliania huxleyi* and “small” *Gephyrocapsa*, although *Calcidiscus leptoporus* and *Gephyrocapsa muelleriae* also have high values (Figs. 2–4). Taking into account the environmental preferences of the above-mentioned species, it is possible to interpret an

oceanic influence, with a mixture of relatively warm and cold-water markers.

For this period, the calculations obtained using biomarkers give higher temperature values (16.5 °C at 325 BC) (Fig. 4), the highest recorded throughout the core, although it is important to note that the temperature fluctuation range is only around 3 °C, coinciding with data previously reported by Bond et al. (1997) and Desprat et al. (2003).

For the same core, Diz et al. (2002) failed to detect significant changes in the data from foraminifera, and they described only one interval from the core base to 1000 years AD. By contrast, Desprat (2001) and Desprat et al. (2003) distinguished two periods, based on pollen fluxes, between ca. 950 years BC and 450 years AD with a relatively cool period from 1000 to 250 years BC and a warm period between 250 BC and 450 AD (Table 2). The first cold period recorded in the VIR-18 core can be correlated with a cold and humid period recorded in Northern Europe and corresponds to the Subboreal–Subatlantic transition (ca. 850 BC–450 AD) (van Geel et al., 1996).

Interval I seems to be a transition period between the end of a warmer period and the beginning of a colder one. The warmer period could be the “Hypsithermal” period, dated between 8000 and 2000 BP (corresponding to 6050–50 BC) (Levac, 2001; Boudreau et al., 2001). The bottom of core VIR-18 (975 years BC), with higher SSTs than present and a high total abundance of coccoliths, may correspond to the last part of the “Hypsithermal” period. The interpretation of this interval as a transition period has also been proposed by Martínez-Cortizas et al. (1999) (Table 2), who calculated the temperature index in Galicia using the accumulation of atmospheric mercury in a peat. Strong temperature fluctuations are the most significant characteristic feature in the interval between 1000 and 250 years BC. Moreover, strong sea level oscillations and changes in the prevailing winds have been reported for 1050–800 BC in northern Spain (Goy et al., 1996). The relatively low values recorded in the abundances of *Coccolithus pelagicus* are in agreement with the relatively high temperature; conversely however, the high values of *Gephyrocapsa muelleriae* are interpreted here as being due to oceanic influence.

5.2. Interval II: 280–125 cm (ca. 252–1368 AD)

During this interval the lowest values in the accumulation rate of total coccoliths are recorded, together with a net increase in *Coccolithus pelagicus*, *Helicosphaera carteri* and *Syracosphaera* spp. (Fig. 3).

The SST estimated for this interval reach mean values of 15.5 °C. Alkenones display intermediate abundances between ~1500 ng/g at ~1000 AD, thereafter increasing to reach values of ~2500 ng/g. During this period, the terrestrial marker hexacosanol shows the opposite pattern to that of C37 alkenones, with maximum values (~7000 ng/g) in the earlier part of the interval (252–1059 AD) and a decreasing trend thereafter. There is an interesting correlation between the abundance of *Coccolithus pelagicus* and hexacosanol (Fig. 5) that holds for most of the core. The quasi-

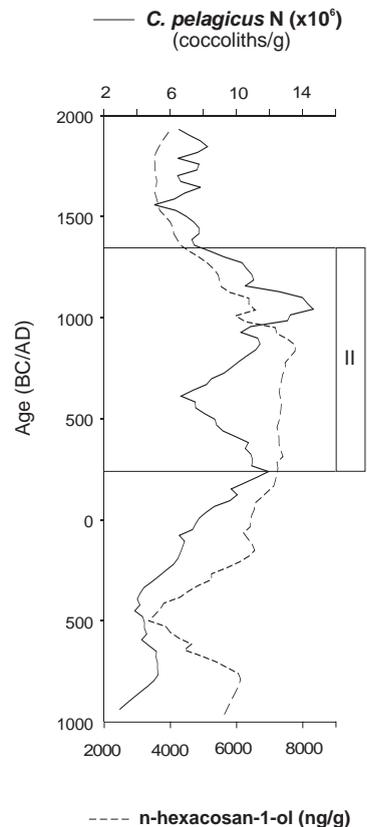


Fig. 5. Accumulation rate (N) of *Coccolithus pelagicus* vs. hexacosanol (ng/g). For definition and characterization on intervals see text (Results section).

parallel record observed between hexacosanol and *C. pelagicus* is not well understood; a suggested explanation, considering the observed relationship between this taxon and nutrient content (Cachão, 1995; Cachão and Moita, 2000) could be an increase in the fluvial nutrient supply to the Ría that may have increased the production of *C. pelagicus*, although a decrease in the SST could have produced a similar effect.

Similarly, the total abundance of coccoliths depicts an inverse correlation to hexacosanol, particularly during this interval but also throughout most of the core. This could be the result of dilution of the coccolith signal by terrestrial input.

The observed peaks of *Helicosphaera carteri* and *Syracosphaera* spp. are interpreted here as episodes of an increase in surface water stratification. Colmenero-Hidalgo et al. (2004) observed the same response of these taxa in the Mediterranean sea. In the Ria de Vigo, fluvial input may produce stratification by changing the mixing regime (Diz et al., 2002). We cannot dismiss, however, a possible connection with the Medieval Warm Period (Table 2), which could not be detected in the SST record.

Two pollen zones have been defined by Desprat (2001) and Desprat et al. (2003) in this interval: from 450 to 950 AD and from 950 to 1400 AD. The first interval is considered cold and humid, whereas the second is interpreted as warmer. The cold and humid period coincides with a peak in the hexacosanol record.

In summary, interval II indicates a stronger continental influence, with a decrease in SST. A weak upwelling, with episodes of stratification from increased river input in the Ría, stopped the estuarine circulation. This explanation is in agreement with the wind pattern provided by Goy et al. (1996) for this period, which indicated prevailing winds blowing from west–southwest and drastic reductions in estuarine connections with the open sea.

5.3. Interval III: 125 cm-core top (ca. 1368–1950 AD)

This interval is characterized by high abundances of “small” *Gephyrocapsa*, with a clear peak of *Gephyrocapsa oceanica* at 1690 AD. (Fig. 6), corresponding to a return of oceanic influence, especially prominent at the top of the core (between 1691 and 1950 AD).

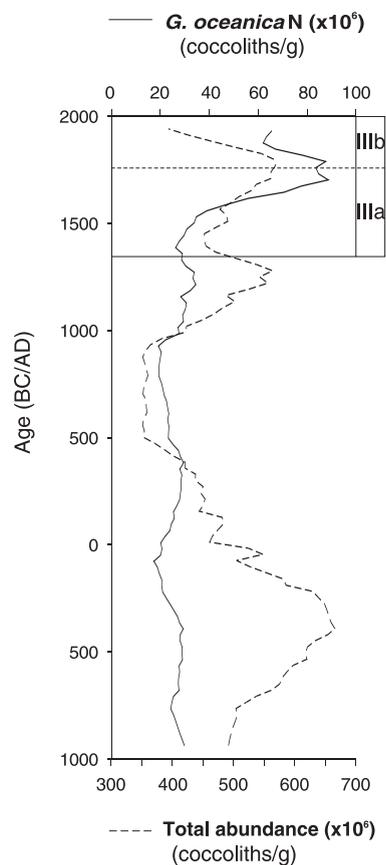


Fig. 6. Accumulation rate (N) of *Gephyrocapsa oceanica* vs. total abundance of coccolithophores per visual field. For definition and characterization on intervals see text (Results section).

For this interval, the paleotemperature record shows the lowest values along the core, 13.3 °C at 1424 AD (Fig. 4). The hexacosanol profile shows lower values in this interval, whereas the alkenones also have high concentrations in the core top.

Based on pollen data, Desprat (2001) and Desprat et al. (2003) defined subintervals from 1400 to 1860 AD and 1860 to 1983 AD, corresponding to relatively warm and cold periods, respectively. These pollen intervals coincide with those defined by us for coccolithophores and are in agreement with the historical data provided by Font Tullot (1998), who reported a cold period at around 1760 AD.

The coccolithophore assemblage observed in this interval suggests a reinforcement in the connection between the Ría and the ocean, in agreement

Table 2

Interpretation of different proxies and their relationship with climatic events in the Ria de Vigo

	Foraminifers Diz et al., 2002	Pollen Desprat et al., 2003	Cocolithophores This study	Mercury Martinez et al., 1999	Climatic periods
2000	Intensification of coastal upwelling processes	Relatively cold	Enhanced oceanic connection Upwelling developed NAO+ dominant	Cold	LIA Little Ice Age
1860					
1691					
1400	Restricted environment	Relatively warm	High continental runoff and surface-water stratification	Thermal stability	MWP Medieval Warm Period
1368					
1000	Exchange with open ocean waters restricted	Relatively cold	NAO - dominant		Dark Ages
950					
450	Frequent salinity changes	Relatively warm	Transitional period with oceanic influence	Warm	RWP Roman Warm Period
252					
0					
150		Relatively cold	Upwelling developed NAO + dominant	High variability in temperature	High Subatlantic
250					
1000					

with previous interpretations based on benthic foraminifera and molecular biomarkers (Diz et al., 2002).

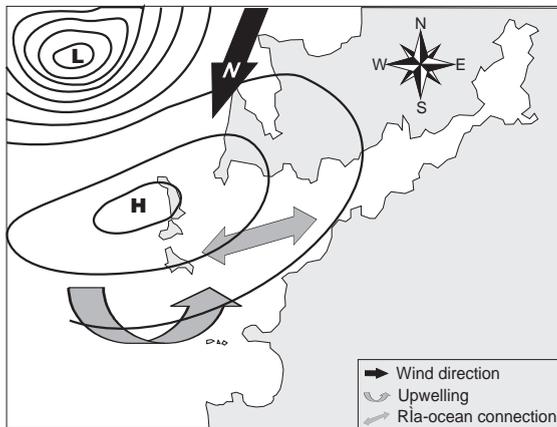
A change in the wind direction from the north to the southwest, as suggested by Goy et al. (1996) and Diz (1998) and a consequent intensification in the upwelling caused a decrease in the SST and an increase in nutrient fluxes to surface waters.

In Table 2 we summarize our results as well as the correlation with intervals/episodes defined by other proxies. It is interesting to note that, whereas the continental record (especially the pollen, but also the mercury in peat bogs) is easy to correlate with the climatic episodes described during the late Holocene in Europe, in the Ria the organisms react in a different way; not always in direct correspondence with the global temperature.

5.4. Possible inferences about the mean state of the North Atlantic Oscillation

The North Atlantic Oscillation (NAO) is an alternation of air masses that occurs between sub-tropical regions (centered over the Azores and Portugal), and the sub-polar regions of the North Atlantic (centered over Iceland) (Fromentin and Planque, 1996). During periods of positive NAO, when a well-developed high atmospheric pressure close to the Azores and a low pressure close to Iceland exist, strong westerly winds develop across the mid-latitude North Atlantic region. In Europe, this situation results in a general warming, with enhanced precipitation in the North and dryness in the South (Fig. 7a). Conversely, during periods of negative NAO, when the high atmospheric system over the

a) NAO+



b) NAO-

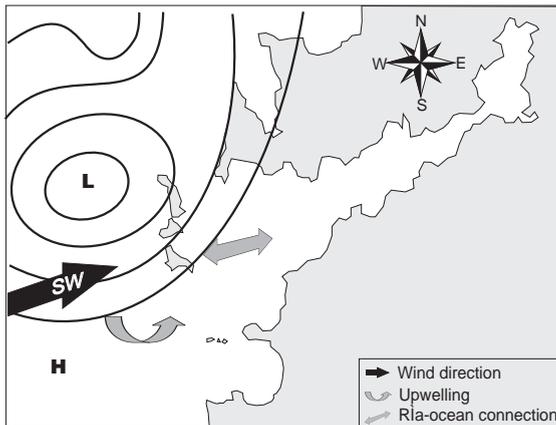


Fig. 7. North Atlantic Oscillation (NAO). General model. a) NAO+: intervals I and III. b) NAO-: interval II. Black arrow: direction of prevailing winds; double arrow: connection between North Atlantic and Ría; bended arrow: upwelling. In all cases, the size of the arrow denotes intensity of the event.

Azores is weakened, westerlies across the North Atlantic region are also weakened, resulting in cold temperatures in northern Europe but increased rainfall over southern Europe (Fig. 7b).

The NAO exhibits prolonged periods of both positive and negative phases. During the last century, for instance, the positive phase of the NAO dominated the atmospheric circulation from the 1900s until about 1930 (Hurrell, 1995). From the early 1940s until the early 1970s, the negative phase was dominant, and later another positive phase was recorded up to the 1990s (Hurrell, 1995). Longer-term records based on

data from ice cores and tree rings have allowed the reconstruction of decadal and even secular variations in the NAO (Glueck and Stockton, 2001; Cullen et al., 2001).

The influence of NAO in palaeoclimate reconstructions has been recognized in the Mediterranean region (Sánchez-Goñi et al., 2002), in the North Atlantic (Bond et al., 1997; Bianchi and McCave, 1999; Chapman and Shackleton, 2000; Giraudeau et al., 2000) and also in our studied area (Desprat et al., 2003), demonstrating a large degree of millennial-scale climatic variability during the Holocene.

As a hypothesis, we propose that the surface water variations observed in the Ría de Vigo may have been controlled by changes in the atmospheric cell gradient than modified by the NAO on a millennial scale. As a result of this, we interpret that our results respond to changes in intensity and wind direction, as well as precipitation. As explained above, interval I, a transitional period, is characterized by relatively high temperatures and an oceanic influence. The higher temperatures are related to a progressively weaker Hypsithermal period and the Azores High position controls the atmospheric circulation in this zone (Fraga, 1991). Therefore, oceanic influence, related to a well-developed Azores High, would elicit a positive NAO (+) situation.

During interval II, surface-water stratification was the dominant scenario in the Ría de Vigo, related to an increase in runoff (Desprat, 2001) and changes in the wind regime. The available data suggest a negative phase of the NAO (-).

During interval III, a strong oceanic influence prevailed in the region, with a weak freshwater input into the Ría and a well-developed upwelling. These data are consistent with a lower humidity on the continent linked to a dominant positive phase of the NAO (+), with different characteristics than those observed in interval I.

6. Conclusions

Coccolith abundances and the molecular biomarker record obtained from core VIR-18 from the Ría de Vigo point to changes in the hydrographic and atmospheric regimes and interactions between the Ría and the open ocean over the last 3000 years.

Our data have allowed the identification of events with different environmental conditions which are interpreted here as follows.

Interval I (ca. 975 years BC–252 AD) is characterized by a maximum accumulation rate of *Calcidiscus leptoporus* and *Gephyrocapsa muellerae*, suggesting a transition from a warmer to a colder period. This transitional warm interval with oceanic influence might be indicative of a prevalence of the positive phase of NAO.

Interval II (ca. 252–1368 AD) is characterized by asynchronous peaks in the accumulation of *Coccolithus pelagicus*, *Helicosphaera carteri*, and *Syracosphaera* species. This interval is associated with high continental runoff and surface-water stratification, as indicated by high hexacosanol concentrations. This humid interval could be associated with a prevailing negative phase in the NAO, which is commonly associated with enhanced precipitation in southern Europe. Interval III (ca. 1368–1950 AD) is characterized by a significant and progressive increase in the distribution of *Gephyrocapsa oceanica*, together with higher C₃₇ alkenone values. This interval shows an enhanced marine influence in the Ría, concomitant with a reduction in fluvial input, as evidenced by a clear decrease in hexacosanol concentrations. This situation is in agreement with a prevailing positive phase of the NAO.

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Appendix A. Taxonomic appendix

From Jordan and Kleijne, (1994).

Calcidiscus leptoporus (Murray and Blackman, 1898) Loeblich and Tappan, 1978 f. *leptoporus*,

Coccolithus pelagicus (Wallich, 1877) Schiller, 1930 f. *pelagicus*,

Emiliania huxleyi (Lohmann, 1902) Hay and Mohler, in Hay et al., 1967 var. *huxleyi*,

“small” *Gephyrocapsa* group (this group includes species with a maximum coccolith diameter of <3 µm; *Gephyrocapsa aperta* Kamptner, 1963 and *Gephyrocapsa ericsonii* McIntyre and Bé, 1967),

Gephyrocapsa muellerae Bréhéret, 1978,

Gephyrocapsa oceanica Kamptner, 1943,

Helicosphaera carteri (Wallich 1877) Kamptner, 1954 var. *carteri*,

Syracosphaera spp. Lohmann, 1902.

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