



Position of the Polar Front along the western Iberian margin during key cold episodes of the last 45 ka

Frédérique Eynaud

*EPOC, UMR 5805, Université Bordeaux 1, Avenue des Facultés, F-33405 Talence, France
(f.eynaud@epoc.u-bordeaux1.fr)*

Lúcia de Abreu and Antje Voelker

Departamento de Geologia Marinha, LNEG, Estrada da Portela, 2721-866 Zambujal, Portugal

Also at CIMAR Associate Laboratory, 4050-123 Porto, Portugal.

Joachim Schönfeld

Leibniz Institute of Marine Sciences at University of Kiel (IFM-GEOMAR), Wischhofstrasse 1-3, D-24148 Kiel, Germany

Emilia Salgueiro

Departamento de Geologia Marinha, LNEG, Estrada da Portela, 2721-866 Zambujal, Portugal

Also at CIMAR Associate Laboratory, 4050-123 Porto, Portugal.

Jean-Louis Turon, Aurélie Penaud, and Samuel Toucanne

EPOC, UMR 5805, Université Bordeaux 1, Avenue des Facultés, F-33405 Talence, France

Filipa Naughton

Departamento de Geologia Marinha, LNEG, Estrada da Portela, 2721-866 Zambujal, Portugal

María Fernanda Sánchez Goñi and Bruno Malaizé

EPOC, UMR 5805, Université Bordeaux 1, Avenue des Facultés, F-33405 Talence, France

Isabel Cacho

Departament d'Estratigrafia, Paleontologia i Geociències Marines, Facultat de Geologia, University of Barcelona, Diagonal Sud, c/Martí Franques s/n, E-08028 Barcelona, Spain

[1] This paper documents the migration of the Polar Front (PF) over the Iberian margin during some of the cold climatic extremes of the last 45 ka. It is based on a compilation of robust and coherent paleohydrological proxies obtained from eleven cores distributed between 36 and 42°N. Planktonic $\delta^{18}\text{O}$ (*Globigerina bulloides*), ice-rafted detritus concentrations, and the relative abundance of the polar foraminifera *Neogloboquadrina pachyderma* sinistral were used to track the PF position. These three data sets, compared from core to core, show a consistent evolution of the sea surface paleohydrology along the Iberian margin over the last 45 ka. We focused on five time slices representative of cold periods under distinct paleoenvironmental forcings: the 8.2 ka event and the Younger Dryas (two recent cold events occurring within high values of summer insolation), Heinrich events 1 and 4 (reflecting major episodes of massive iceberg discharges into the North Atlantic), and the Last Glacial Maximum (typifying the highest ice volume accumulated in the Northern Hemisphere). For each event, we generated schematic maps

mirroring past sea surface hydrological conditions. The maps revealed that the Polar Front presence along the Iberian margin was restricted to Heinrich events. The sea surface conditions during the Last Glacial Maximum were close to those at present day, except for the northern sites which briefly experienced subarctic conditions.

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1. Introduction

[2] For more than 2.5 Ma, the Earth has experienced drastic climatic fluctuations among which the more pronounced occurred during the last 450 ka. Forcing factors at the origin of these major climatic seesaws are now fairly well identified throughout external (orbital) and internal parameters of the Earth's climatic system. However, phasing relationship and coupling phenomena between these parameters are still not well understood [e.g., *Lisiecki et al.*, 2008]. In this way, climate extremes are major challenges regarding our comprehension of the Earth's climate variability and stability. The abrupt cold excursions, i.e., stadial events [*Dansgaard et al.*, 1993] or Heinrich events [*Heinrich*, 1988], which occurred during the last glacial period, provide excellent archives of Pleistocene rapid climatic transitions. They, furthermore, raise major questions about the pace of the natural system. The extreme Heinrich events result from a massive calving of icebergs in relation to the sudden collapse of the periarctic ice sheets during the last glacial period [e.g., *Heinrich*, 1988; *Bond et al.*, 1993; *Hemming*, 2004]. They took place at the end of a sawtooth shaped cooling cycle of 4 successive stadials/interstadials (e.g., Bond cycles [*Bond et al.*, 1993]). However, the triggering of these events is still a matter of debate, as they imply complex mechanisms linking ice sheet, atmosphere and ocean dynamics. Their impact was recognized worldwide [e.g., *Leuschner and Sirocko*, 2000; *Wang et al.*, 2001] and each of

the Earth's surface reservoirs, i.e., the cryosphere, atmosphere, oceans, and biosphere, was affected. Some of the most severe Heinrich events impacted dramatically on the North Atlantic's overturning circulation [e.g., *Zahn et al.*, 1997; *Curry et al.*, 1999; *Roche et al.*, 2004]. These events potentially constitute catastrophic analogs of modern or near-future ice sheet (Antarctic especially) collapses in the context of global warming.

[3] The objective of this paper is to document the impact of some of these cold extremes on the surface water hydrology of the Iberian margin. For this purpose, we focus on the reconstruction of the position of the Polar Front over the last 45 cal ka B.P. (calibrated ¹⁴C kiloyears before present; hereafter referred to as "ka"). We selected five time slices (Table 2a in section 2.2) representative of well-known cold episodes under various climatic forcing mechanisms (insolation, continental ice storage, sea level values, vegetation cover). At Present, the Polar Front (PF) closely follows the Greenland and eastern Canadian continental margins (Figure 1). In the ocean it is defined as a physical oceanographic structure with high temperature and salinity gradients, which separates two different surface water masses: the relatively warm, high-salinity Atlantic Water, and the cold low-salinity modified Polar Water carried southward by the East and West Greenland and Labrador currents [*Holliday et al.*, 2007]. This front constitutes a strong ecological barrier in the North Atlantic Ocean and also physically constrains sea ice and iceberg drifting. The latitudinal migration

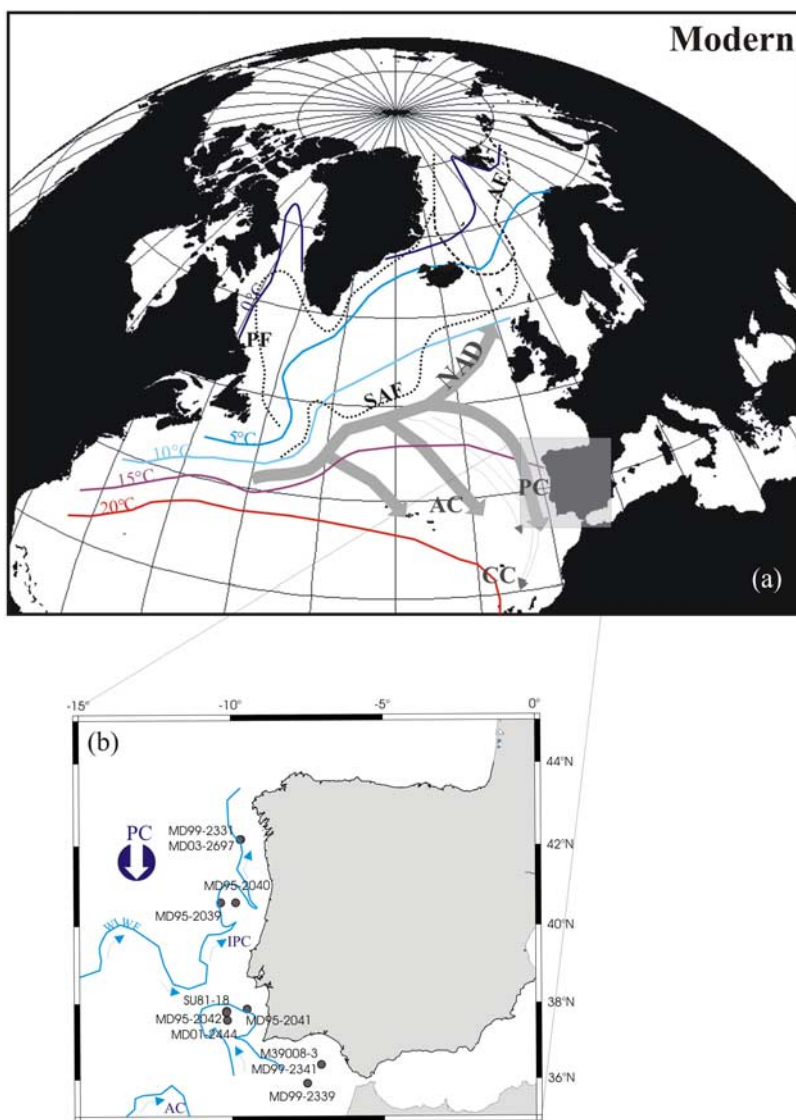


Figure 1. (a) Modern sea surface hydrographic conditions of the North Atlantic Ocean. Schematic pathway of the North Atlantic Drift (NAD) and its major boundary return currents (PC, Portugal Current; CC, Canary Current; AC, Azores Current) [after Sverdrup *et al.*, 1942; Crowley, 1981; Dickson *et al.*, 1988]. The SST isotherms are from Tchernia [1978] (SST annual mean). Dotted lines locate the major hydrographic fronts: PF, Polar Front; AF, Arctic Front; SAF, Subarctic Front [after Dickson *et al.*, 1988]. (b) Location of the studied cores along the Iberian margin with regard to the modern sea surface structures in winter (PC, Portugal Current; IPC, Iberian Poleward Current; AC, Azores Current; WIWF, Western Iberia Winter Front [after Peliz *et al.*, 2005]).

of the Polar Front as south as 40°N during late Pleistocene cold events has been documented by numerous studies [e.g., McIntyre *et al.*, 1976; Ruddiman and McIntyre, 1981; Duprat, 1983; Bard *et al.*, 1987; Abrantes *et al.*, 1998]. Our intention is to complement this information, since the Iberian margin offers by now an exceptional coverage of high-quality sequences, i.e., with diversified data sets, high time resolution and excellent stratigraphical control (Table 1). We choose

robust paleohydrological proxies, i.e., planktonic $\delta^{18}\text{O}$ (in *Globigerina bulloides*), ice-rafted detritus (IRD) concentrations and the relative abundance of the polar foraminifera *Neogloboquadrina pachyderma* sinistral (left coiled, *N. pachyderma* sinistral). These data sets were readily compared from core to core. They allowed us to create regional maps and estimate past changes in sea surface hydrology displaying the migration of the PF with

**Table 1.** Geographic Information of the Selected Cores

Core	Latitude	Longitude	Water Depth (m)	Core Length (m)
SU81-18	37.767	-10.183	3155	7
MD95-2039	40.580	-10.350	3381	35.71
MD95-2040	40.580	-9.860	2465	35.24
MD95-2041	37.883	-9.510	1123	17.62
MD95-2042	37.800	-10.170	3146	39.56
MD99-2331	42.150	-9.683	2120	37.20
MD99-2339	35.886	-7.528	1177	18.54
MD99-2341	36.389	-7.066	582	19.42
MD01-2444	37.561	-10.150	2642	27.48
MD03-2697	42.150	-9.700	2164	41.23
M39008-3	36.380	-7.072	577	5.8

a far better resolution than had been previously possible.

2. Iberian Margin

2.1. Present-Day Hydrography

2.1.1. Dynamic Features

[4] The hydrography at the western Iberian margin (Figure 1) is primarily influenced by the North Atlantic subtropical gyre dynamic and a complex current system [Relvas *et al.*, 2007]. The surface circulation regime is governed by the seasonal displacement of the Azores anticyclonic cell and its associated winds [e.g., *Fiúza*, 1984; *Fiúza et al.*, 1998]. Several major features have to be considered; they are described in sections 2.1.1.1–2.1.1.3.

2.1.1.1. Large-Scale Currents

[5] These include (1) the Portugal Current (PC), flowing southward from 45°N to 30°N; (2) the Azores Current (AC) derived from the southern branch of the Gulf Stream (GS) and flowing eastward to the Gulf of Cadiz at about 38°N; and (3) the Canary Current (CC) which is fed by both the AC and the PC. Together, these currents form the Eastern Boundary Current of the subtropical gyre.

2.1.1.2. Portugal Upwelling System

[6] The western Iberian margin belongs to the Portugal-Canary eastern boundary upwelling system [Peliz *et al.*, 2005]. Between April and October, upwelling cells are activated in response to strong northerlies which accompany the northward displacement and the strengthening of the Azores high-pressure cell [Fiúza *et al.*, 1982; *Fiúza*, 1984; *Sousa and Bricaud*, 1992; *Sánchez and Relvas*, 2003]. This leads to the formation of a cold and

nutrient-rich coastal water band [Relvas *et al.*, 2007], which may extend from between 30 and 50 km and 100–200 km offshore [e.g., *Sousa and Bricaud*, 1992; *Relvas et al.*, 2007]. Eastern North Atlantic Central Water (ENACW) and partially Mediterranean Outflow Water [Sánchez and Relvas, 2003] are the sources of the upwelled waters.

2.1.1.3. Iberian Poleward Current

[7] During winter months, weak northerlies interspersed occasionally with strong south-southwest winds, favoring coastal convergence conditions. The most relevant transport mechanism is then the northward flow of a warm surface and undercurrent denominated as the Iberian Poleward Current (IPC) or Portugal Coastal Counter Current [e.g., *Fiúza et al.*, 1998; *Peliz et al.*, 2005] (Figure 1b). This current is recognized northward up to the Bay of Biscay, where it is known as the “Navidad Current” [García-Soto *et al.*, 2002]. The poleward flow affects both the surface and intermediate waters down to 1500 m depth [Relvas *et al.*, 2007], adding warm and saline inputs into the study region [Haynes and Barton, 1990]. A connection with a northward recirculation of the AC has also been suggested [Peliz *et al.*, 2005].

2.1.2. Synoptic View

[8] Modern sea surface conditions (Figure 1) along the margin between 37° and 41°N are characterized by mean annual temperatures ranging from 18°C to 16°C [Levitus, 1994]. These isotherms move seasonally and zonal gradients may occur during some periods of the year [Peliz *et al.*, 2005]. During winter the Western Iberia Winter Front (WIWF), a recurrent feature between 10 and 11°W, separates the subtropical IPC waters from the cooler PC around 40°N [Peliz *et al.*, 2005] (Figure 1b). However, major features are generated with the



summer upwelling: a marked cold coastal band occurs, where the SST anomaly could reach an amplitude of -5°C (averaged values from June–September 2001–2005 [Relvas et al., 2007]).

[9] Intermediate circulation is dominated by the ENACW and the Mediterranean Outflow Water (MOW). Two main MOW branches are distinguishable, an upper core centered around 700–800 m, and a lower core at about 1200 m [Ambar and Howe, 1979; Iorga and Lozier, 1999]. The MOW follows a poleward route, but also propagates toward the west in the form of “meddies.” This water mass is supposed to have a dynamic impact regarding the ocean upper layers and it may notably constitute a complementary contributor to the AC [Jia, 2000].

2.2. Past Protrusions of Polar Waters

[10] Figure 2 presents the state of the art regarding mapping of past incursions of Polar waters in the North Atlantic for the Last Glacial Maximum (LGM [CLIMAP Project Members, 1981; Kucera et al., 2005a]) (Figure 2a) and for Heinrich events (HEs [Ruddiman, 1977; Cortijo et al., 2005]) (Figure 2b). It is worth noting that existing maps diverge regarding the kind of structure represented. This mismatch does not favor an objective representation of the events. In fact, the quality and resolution of the existing maps are linked to the proxies used for their construction, for instance the maximum abundances of lithic grains in the sediment [e.g., Hemming, 2004], $\delta^{18}\text{O}$ isolines for HEs [e.g., Cortijo et al., 2005], or seasonal sea ice limit for the LGM [e.g., CLIMAP Project Members, 1981; Kucera et al., 2005a; Byrkjedal et al., 2006]. They furthermore show poor details along the North Atlantic margins. This is especially true for the Iberian Margin despite the fact that this region was the focus of numerous paleoceanographic investigations, and that its sensitivity to abrupt climatic changes was recognized early [e.g., Turon, 1984; Bard et al., 1987; Zahn et al., 1997; Abrantes et al., 1998; Shackleton et al., 2000; Martrat et al., 2004, 2007].

[11] For the last glacial period, the southernmost latitudinal position of the Polar Front (PF) in the North Atlantic Ocean was initially drawn around 40°N [CLIMAP Project Members, 1981; Ruddiman and McIntyre, 1981; Bard et al., 1987] (Figure 2a). High-resolution studies revealed, however, contrasting hydrological patterns between the LGM interval (time slice: 19–23 ka B.P. in the sense of Mix et al. [2001]) and the surrounding HEs [e.g.,

Sarnthein et al., 1995, 2003; de Vernal et al., 2000]. Along the Iberian margin, this was first demonstrated by Abrantes et al. [1998], who compared the last 20 ka paleohydrology of six ENAM (European North Atlantic Margin) sites in between the Nordic seas and the Portuguese margin and concluded that the Polar Front did not reach the Iberian margin during the LGM. Instead, their SST reconstructions revealed the installation of hydrological conditions similar to those of the Holocene at that time. This was later corroborated by further studies also based on planktonic foraminifera [de Abreu et al., 2003; Kucera et al., 2005b]. Similar to the Younger Dryas [Bard et al., 1987], the most drastic hydrological changes along the Iberian margin were recorded during major glacial collapses, i.e., HEs [Abrantes et al., 1998], with a southward influence of icebergs testified up to 37°N and beyond [Kudrass, 1973; Schönfeld and Zahn, 2000; Cacho et al., 2001].

[12] In the “Ruddiman” IRD belt Heinrich layers are easily identified in deep-sea sediments as layers of coarse lithic grains, repetitively deposited over the last 70 ka [e.g., Heinrich, 1988; Bond et al., 1993; Grousset et al., 1993; Hemming et al., 1998]. The HEs also carry a strong micropaleontological signal there, with planktonic foraminifera assemblages becoming dominated by the polar species *N. pachyderma* sinistral, and a geochemical signature preserved in the polar foraminifera tests showing lightening in the surface water oxygen isotopic ratio due to the input of freshwater [Vidal et al., 1997; Elliot et al., 1998; Cortijo et al., 2005] (Figure 2b). Although the impact of HEs on the Iberian margin and adjacent continents has been recognized throughout different multiproxy studies as far south as 36°N [Bard et al., 1987; Lebreiro et al., 1996; Baas et al., 1997, 1998; Zahn et al., 1997; Abrantes et al., 1998; Bard et al., 2000; Sánchez Goñi et al., 2000; Thouveny et al., 2000; Schönfeld et al., 2003; de Abreu et al., 2003; Roucoux et al., 2001, 2005; Voelker et al., 2006; Narciso et al., 2006; Naughton et al., 2007, 2009; F. Naughton et al., Wet to dry climatic trend in north western Iberia within Heinrich events, paper presented at IX International Conference on Paleooceanography, Shanghai, China, September 2007], and over several climatic cycles [Sánchez Goñi et al., 1999; Eynaud et al., 2000; Moreno et al., 2002; Martrat et al., 2007], a discussion still exists whether or not large icebergs have prevailed in the seas off western Iberia. Diverse works have actually debated how to tackle the problem of what is purely ice-rafted material and what can be

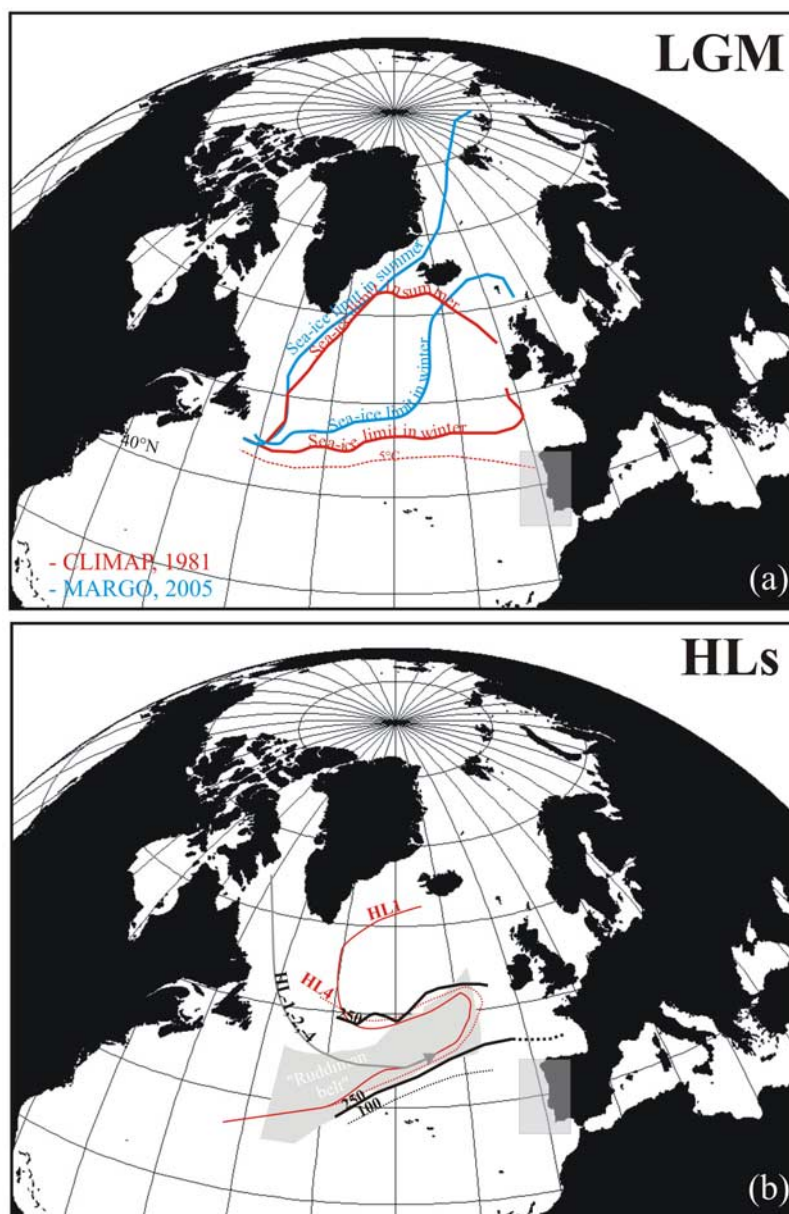


Figure 2. Past sea surface hydrographic conditions of the North Atlantic Ocean. (a) During the Last Glacial Maximum. Northern limit of ice-free conditions during winter and summer after *CLIMAP Project Members* [1981] (red lines) and after *Kucera et al.* [2005a] (blue lines). The 5°C SST isotherm in winter after CLIMAP is also indicated. (b) During Heinrich events. Ice-rafted detritus (IRD) belt or “Ruddiman belt” in gray with the 250 and 100 $\text{mg cm}^{-2} \text{ka}^{-1}$ flux lines [after *Ruddiman, 1977; Hemming, 2004*]. Preferential routing of icebergs during HE1, HE2, and HE4 [after *Grousset et al., 1993*]. Planktonic $\delta^{18}\text{O}$ 1‰ isoline for HE4 (dotted red line) and HE1 (full red line) [after *Cortijo et al., 2005*].

continent/platform-derived in this high-sedimentation margin [e.g., *Baas et al., 1997; Zahn et al., 1997; Bard et al., 2000; Toucanne et al., 2007*]. In fact, in this area heavier planktonic $\delta^{18}\text{O}$ values confirm colder SST rather than ice melting, as they consistently did during Dansgaard-Oeschger (DO) stadials [*Shackleton et al., 2000*], in agreement with other SST indicators like foraminifera and

dinocyst assemblages [e.g., *de Abreu et al., 2003; Schönfeld et al., 2003; Turon et al., 2003*], alkenone concentrations [e.g., *Pailler and Bard, 2002; Cacho et al., 2001*] and Mg/Ca [*Skinner and Elderfield, 2005*]. HEs, nevertheless, represent times of major protrusion of less saline polar waters on this margin as indicated by the co-occurrence of colder surface waters and ice-rafted

Table 2a. Stratigraphical Intervals of the Events Considered in This Paper^a

Event	8.2 ka ^b		YD ^c		HE1 ^d		LGM ^e		HE4 ^f	
	¹⁴ C ka B.P.	cal ka B.P.	¹⁴ C ka B.P.	cal ka B.P.	¹⁴ C ka B.P.	cal ka B.P.	¹⁴ C ka B.P.	cal ka B.P.	¹⁴ C ka B.P.	cal ka B.P.
TOP		8.0	10	11.5	13.4	15	16	19	33.9	38.5
BASE		8.6	11	13	15.1	18	19.5	23	34.9	40

^aThe ages of these intervals were reported from the most cited bibliographic sources dealing with the selected events.

^bLimits after Rohling and Pälike [2005].

^cLimits after Bard [1998].

^dThe ¹⁴C limits after Elliot et al. [1998, 2001] and CAL limits after Bard et al. [2000].

^eLGM EPILOG definition in the sense of Mix et al. [2001].

^fThe ¹⁴C limits after Elliot et al. [1998, 2001] and CAL limits after Sánchez Goñi et al. [2000].

debris (IRD [e.g., Lebreiro et al., 1996; Bard et al., 2000; Moreno et al., 2002; de Abreu et al., 2003]). The strong cooling in the surface waters overprinted the salinity signal (freshening due to ice-berg melting) in the planktonic $\delta^{18}\text{O}$ values.

[13] For this paper, we focused on 5 time slices representative of well-dated cold events in the Northern Hemisphere (Tables 2a and 2b), but which were characterized by distinct orbital configuration, atmospheric pCO₂ concentration, ice sheet extension, and vegetation cover. They are as follows:

[14] 1. Heinrich event 4 (HE4) is one of the most pronounced events regarding IRD fluxes [Cortijo et al., 1997, 2005; Schönfeld et al., 2003; Hemming, 2004], also marked by a nearly complete shutdown of the thermohaline circulation [Maslin et al., 1995; Elliot et al., 2002; Roche et al., 2004]. Decreasing summer insolation values (65°N) characterize this interval.

[15] 2. The LGM is the time of the most recent maximum in globally integrated ice volume and minimum in global sea level [Mix et al., 2001]. The temperate forests were substantially fragmented

and reduced in extent at that time, in a context of dryer and colder midlatitude climate [e.g., Harrison et al., 2001; Harrison and Prentice, 2003; Roucoux et al., 2005; Naughton et al., 2007]. Marine proxies suggest significantly colder than modern sea surface conditions in some domains of the North Atlantic, with a seasonal SST contrast larger than at present [de Vernal et al., 2005, 2006]. Also, a more zonal temperature pattern existed, with a steep 10°C SST gradient between 40 and 45°N [Duprat, 1983; Calvo et al., 2001; de Vernal et al., 2006]. The LGM was marked by a summer insolation at high latitudes with values similar to today.

[16] 3. Heinrich event 1 (HE1) is the last known major ice sheet collapse [Heinrich, 1988; Dowdeswell et al., 1995; Cortijo et al., 2005], and it coincides with the onset of the last termination and times of major climate reorganization toward modern conditions [Steffensen et al., 2008]. HE1 occurred in a context of increasing insolation. On the European margin, this event shows a double signature, with a first impact of European ice sheet melting (precursor events), followed by the Laurentide ice sheet collapse imprint [e.g., Grousset et al., 2000; Zaragosi et al., 2001; Peck et al., 2007].

Table 2b. Dates Obtained After Previous Works on the Iberian Margin for Heinrich Events^a

HE1 Along the Iberian Margin (cal ka B.P.)			HE4 Along the Iberian Margin (cal ka B.P.)		
Top	Base	Cores/References	Top	Base	Cores/Reference
14.85	17.64	MD95-2039 (Schönfeld et al. [2003]: after <i>N. pachyderma</i> sinistral %)			
14.96	17.84	MD95-2040 (Schönfeld et al. [2003] and de Abreu et al. [2003]: after <i>N. pachyderma</i> sinistral %)	37.95	39.93	MD95-2040 (Schönfeld et al. [2003]: after <i>N. pachyderma</i> sinistral %)
15.04	17.44	SU81-18 (Bard et al. [2000]: 255–330 cm after alkenone record and main IRD concentrations)			
15.38	17.17	MD99-2339 (Voelker et al. [2006]: SST)	38.5	39.5	MD99-2339 (Voelker et al. [2006]: SST)

^aRegarding the available chronology of these events in the area (and even if error bars inherent to datings could minimize this remark), cold spells appeared earlier on the northern margin and lasted longer there.



[17] 4. The Younger Dryas (YD) cold event, first described in palynological records from Scandinavia, is known as a true HE in the western North Atlantic (HE0 [e.g., *Andrews et al.*, 1995]). On the European margin, the YD is marked by the installation of subpolar conditions [*Baas et al.*, 1997] and the contraction of temperate tree populations [e.g., *Naughton et al.*, 2007].

[18] 5. The 8.2 ka event is known as a significant temperature anomaly during the early Holocene, which is attributed to a meltwater outflow into the North Atlantic Ocean via the Hudson Strait [*Barber et al.*, 1999; *Rohling and Pälike*, 2005; *Lajeunesse and St-Onge*, 2008] that perturbed the overturning circulation [*Ellison et al.*, 2006; *Kleiven et al.*, 2007]. Intrahemispheric impacts of this event were detected in many paleoclimatic records (continental or marine as well [*Alley and Ágústsdóttir*, 2005]).

3. Methodology

[19] The aim of this paper is to track the intrusions of polar waters over the Iberian margin and to delineate the Polar Front. Numerous high-resolution multiproxy records have been generated from sediment cores retrieved through international programs such as ENAM and IMAGES in this area. However, no synoptic synthesis between the different sequences has been carried out, yet. We have selected records adequate to both the existence of comparable proxy data among the cores and to a spatial distribution as wide as possible over the area (Table 1, see also auxiliary material), in order to tentatively draw a synthetic scenario for past sea surface hydrographical changes.¹

3.1. Stratigraphical Considerations

[20] The selected records are high-resolution records mainly from Calypso piston cores retrieved during the IMAGES I (1995) and V (1999) cruises with the R/V *Marion Dufresne* (IPEV). For ten of the eleven cores used, the age models were previously published (Table 3). The chronostratigraphies were based on calibrated ¹⁴C AMS datings and complemented by stratigraphical tie points from a correlation of the stable isotope data with Greenland ice core $\delta^{18}\text{O}$ records (GRIP, GISP2, NorthGRIP). The age model of core MD95-2041 was constructed by a correlation of the planktonic

oxygen isotope curve with those of the nearby sediment cores SU81-18, SO75-26KL, and MD99-2341, and with the GISP2 ice core. The lower part of the curve was tied to the stacked oxygen isotope record of *Martinson et al.* [1987]. For the other cores, we abstained from a further intercore fine tuning of the planktonic oxygen isotope records, and we decided to keep with the published age models, taking into account small stratigraphical discrepancies at certain intervals. These discrepancies were however at an infrascala time resolution than the one we used for our compilation, and therefore do not bias the results presented here. Further details regarding the methodology used to build the age model for each independent core are available in the papers cited in Table 3.

3.2. Best Proxies to Trace the Polar Front?

[21] The strategy to select the proxies to compile was motivated by the fact that among the existing set (see *Henderson* [2002] for a review), some of the most robust and “author to author consistent” paleohydrological proxies are those based on microfossil remains. We therefore focused on data from planktonic foraminifera, which benefit from more than 50 years of paleoceanographic expertise [*Kucera*, 2007], and which were thoroughly tested through both geochemical (stable isotopes [e.g., *Ravelo and Hillaire-Marcel*, 2007]) and ecological (transfer function [e.g., *Guiot and de Vernal*, 2007]) investigations. In particular, we used the planktonic $\delta^{18}\text{O}$ from *G. bulloides*, and the relative abundance of the polar foraminifera *N. pachyderma* sinistral counted in the $>150\mu\text{m}$ fraction (Figures 3 and 4). The choice to analyze this fraction was justified by the fact, that it has been conventionally adopted for large-scale paleoceanographical reconstructions [e.g., *CLIMAP Project Members*, 1981; *Pflaumann et al.*, 1996; *Kucera et al.*, 2005b]. Actually, it normally excludes morphologically immature forms (juvenile) and analysis is therefore less subjective. We combined these data with those of large lithic grain (LLG) analyses allowing us to identify ice-rafting activities (referred later in the text as LLG/IRD).

3.3. Mapping Southward Shifts of the Polar Front Throughout Time

[22] To support a synoptic view of the migration of the PF throughout time, we created maps reflecting the hydrological conditions over the Iberian margin for the 5 time slices. For mapping, we considered the $\delta^{18}\text{O}$ of *G. bulloides* together with the relative

¹Auxiliary materials are available at <ftp://ftp.agu.org/apend/gc/2009gc002398>.



Table 3. Chronological Constraints on the Selected Cores^a

Core	Number of AMS ¹⁴ C Dates (Last 45 ka)	Greenland Ice Core Based Time Scale?	Marine Core to Core Correlation/Comparison?	Successive References
SU81-18	17			<i>Bard et al.</i> [1987, 2000] and <i>Turon et al.</i> [2003]
MD95-2042	16	yes (GISP2 and GRIPSS09sea)	yes, with SU81-18	<i>Shackleton et al.</i> [2000], <i>Bard et al.</i> [2004], <i>Shackleton et al.</i> [2004], <i>Daniau et al.</i> [2007], and <i>Sánchez Goñi et al.</i> [2008]
MD95-2040	16	yes (GISP2)	yes, with SU81-18, MD95-2042, and MD95-2039	<i>Thomson et al.</i> [1999], <i>Moreno et al.</i> [2002], <i>de Abreu et al.</i> [2003], and <i>Schönfeld et al.</i> [2003]
MD95-2039	7	yes (GISP2)	yes, with MD95-2040	<i>Thomson et al.</i> [1999], <i>Thouveny et al.</i> [2000], and <i>Schönfeld et al.</i> [2003]
MD95-2041	17	yes (<i>J. Schönfeld et al.</i> , manuscript in preparation, 2009)		<i>Schönfeld et al.</i> (manuscript in preparation, 2009)
MD99-2341	12	yes (GISP2)	yes, with M39008-3, SU81-18, and MD95-2039	<i>Llave et al.</i> [2006] and <i>Toucanne et al.</i> [2007]
MD99-2339	20	yes (GISP2)	yes, with MD95-2043 and MD99-2343	<i>Voelker et al.</i> [2006]
MD99-2331	20	yes (GRIP, GISP2, and GRIPSS09sea)	yes, with MD95-2042	<i>Gouzy et al.</i> [2004], <i>Bard et al.</i> [2004], <i>Shackleton et al.</i> [2004], <i>Naughton et al.</i> [2007, 2009] and <i>Sánchez Goñi et al.</i> [2008]
MD01-2444			yes, with ODP1060	<i>Vautravers and Shackleton</i> [2006]
MD03-2697	7	yes (GRIP)	yes, with MD95-2042	<i>Naughton et al.</i> [2007] and <i>Sánchez Goñi et al.</i> [2008]
M39008-3	10		yes, with SU81-18	<i>Cacho et al.</i> [2001]

^aCompilation of the history of the construction of the age scales for each of the cores considered (see references). The chronology was established on the basis of ¹⁴C AMS dates but was also tied to Greenland ice cores, complemented by core to core correlation.

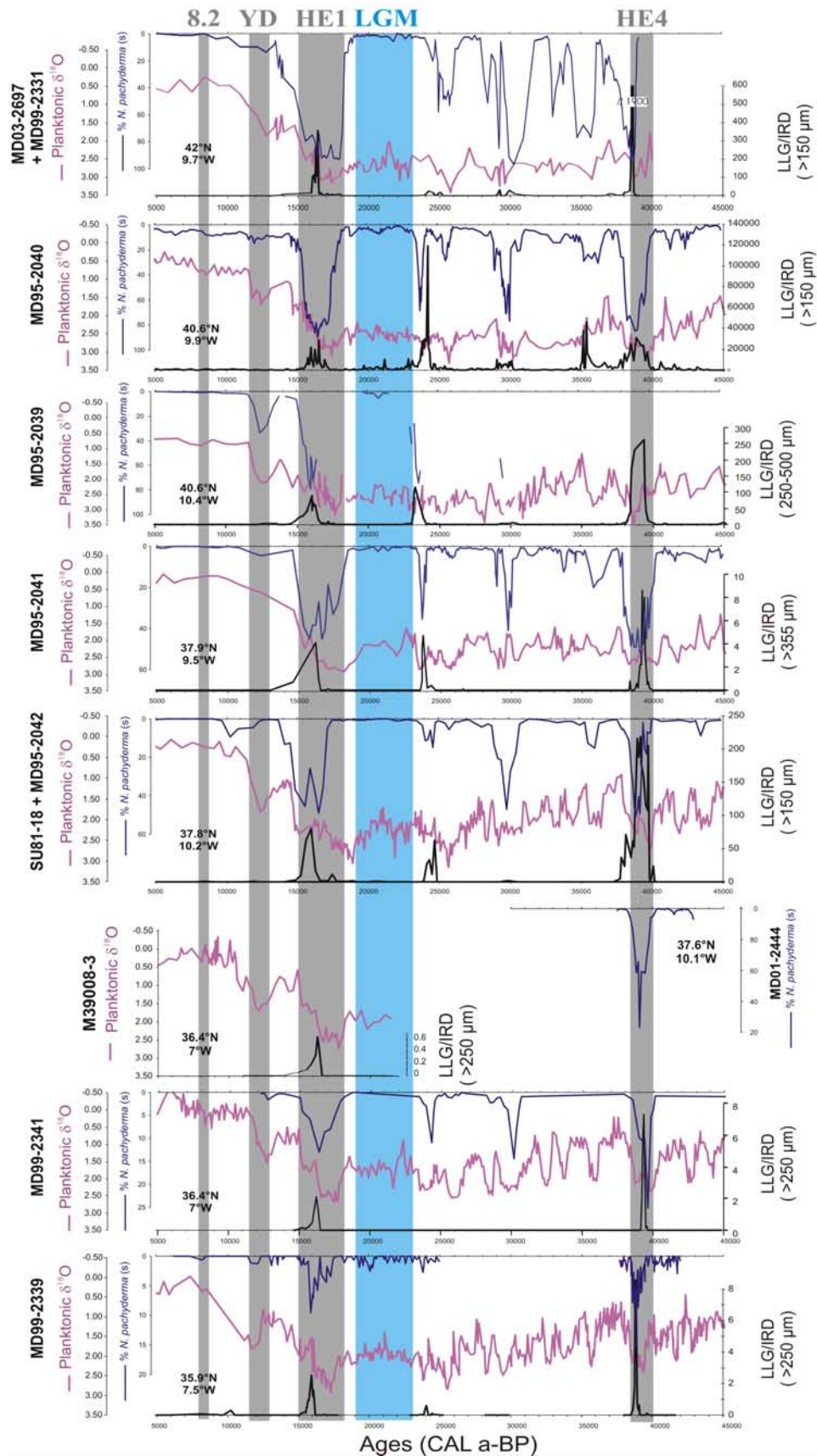


Figure 3

abundance of *N. pachyderma* sinistral only. The LLG/IRD concentrations were analyzed from different size fractions, in particular >150 μm , >250 μm , and >355 μm , and the values were given as different units, as proportion in the respective grain size fraction or as abundance of lithic grains per gram sediment. The IRD abundance in the >150 μm fraction can be by a factor of 100 higher than those >250 μm . Regarding these inconsistencies, we decided not to map the LLG/IRD data, but to use the information to mark ice-rafting periods.

[23] For HE4, HE1, the YD and 8.2 ka events we adopted the approach based on the $\delta^{18}\text{O}$ data developed by Cortijo *et al.* [2005]; that is, we regarded the difference between planktonic $\delta^{18}\text{O}$ values recorded prior to and during the event itself. During the LGM, however, such changes were not significant. We, therefore, decided to use values averaged over the whole interval. This approach was furthermore justified because of the 4 ka duration of the LGM, and also because previous LGM mapping exercises used mean values over the whole interval as well [CLIMAP Project Members, 1981; Kucera *et al.*, 2005a]. For the $\delta^{18}\text{O}$ mapping we decided to limit our interpretation to the plot of numerical values.

[24] Regarding the *N. pachyderma* sinistral abundances, we decided to plot the maximal abundances recorded for each event (Figure 3) in order to capture the true amplitude of the signal. Our stratigraphical constraint for each of the events is precise enough to exclude diachronism. We interpreted the plots drawing water mass limits with the intention to locate the PF. This was possible only because a good coherency exists between the modern distribution of *N. pachyderma* sinistral in surface sediments and the overlaying surface water masses (see Figure 4). Four major provinces were distinguished that, permit us to consider transitional settings and allow us to better qualify our observations: (1) *N. pachyderma* sinistral abundances of more than 90% prevail in the vicinity of the Polar Front, (2) abundances between 50 and 90% characterize a situation where modified polar water could have reached the margin and generated a front comparable to the present-day Arctic Front,

(3) relative abundances between 50 and 5% indicated that subarctic waters invaded the area, and (4) relative abundances below 5% depicted sea surface conditions close to the modern situation.

4. Polar Front Signature Over the Iberian Margin

[25] The selected proxies showed a good coherency through time and space, and they permit to distinguish specific events which appear synchronously along the margin. HEs especially exhibit a strong paleohydrological signature, with high *G. bulloides* $\delta^{18}\text{O}$ values and high *N. pachyderma* sinistral percentages, together indicating low SST. The different amplitudes recorded between the cores basically reflect the latitudinal distribution of the records. *N. pachyderma* sinistral occurs for instance, with up to 90% in quasi-monospecific assemblages at the northern part of the margin whereas its percentages range from 10 to 15% at maximum in the south (Figure 5). By analogy to the modern distribution of this species in surface sediments of the North Atlantic Ocean (Figure 4), this pattern suggests the penetration of polar surface waters in the north and, at least subpolar surface waters in the south during HEs. LLG/IRD concentrations also peak during HEs. Even if they were low as compared to those observed within the Ruddiman belt, LLG/IRD concentrations constitute a complementary evidence of the protrusion of polar waters on this margin. Regarding LLG/IRD, HE4 occurs as one of the more drastic events, whereas no lithic was recorded during the YD cold event. The LGM depicts mild sea surface conditions as supported by the near disappearance of *N. pachyderma* sinistral in the fauna. No LLG/IRD pulses were recorded. High *G. bulloides* $\delta^{18}\text{O}$ values during this interval mirror the change in water $\delta^{18}\text{O}$ related to sea level changes (global ice volume signal), rather than an invasion by cold surface waters [see Waelbroeck *et al.*, 2005]. The 8.2 ka event appears insignificant in the selected records in spite of an accurate resolution of sampling.

Figure 3. Compilation over the last 45 ka of the $\delta^{18}\text{O}$ planktonic records (*G. bulloides*) used for this work compared, when existing, with the large lithic grain abundances in the sediment (proportion of grains >150 μm or even >250 $\mu\text{m/g}$ of dried sediment, referred to in the text as LLG/IRD) and with the relative abundances of the polar foraminifera *N. pachyderma* sinistral within the planktonic fossil assemblages (>150 μm). From top to bottom, the records are displayed from the northern (42°N) to the southern one (36°N). Vertical bars delineate the events we consider for this paper (see also Table 3). For detailed information on the data used, see the auxiliary material.

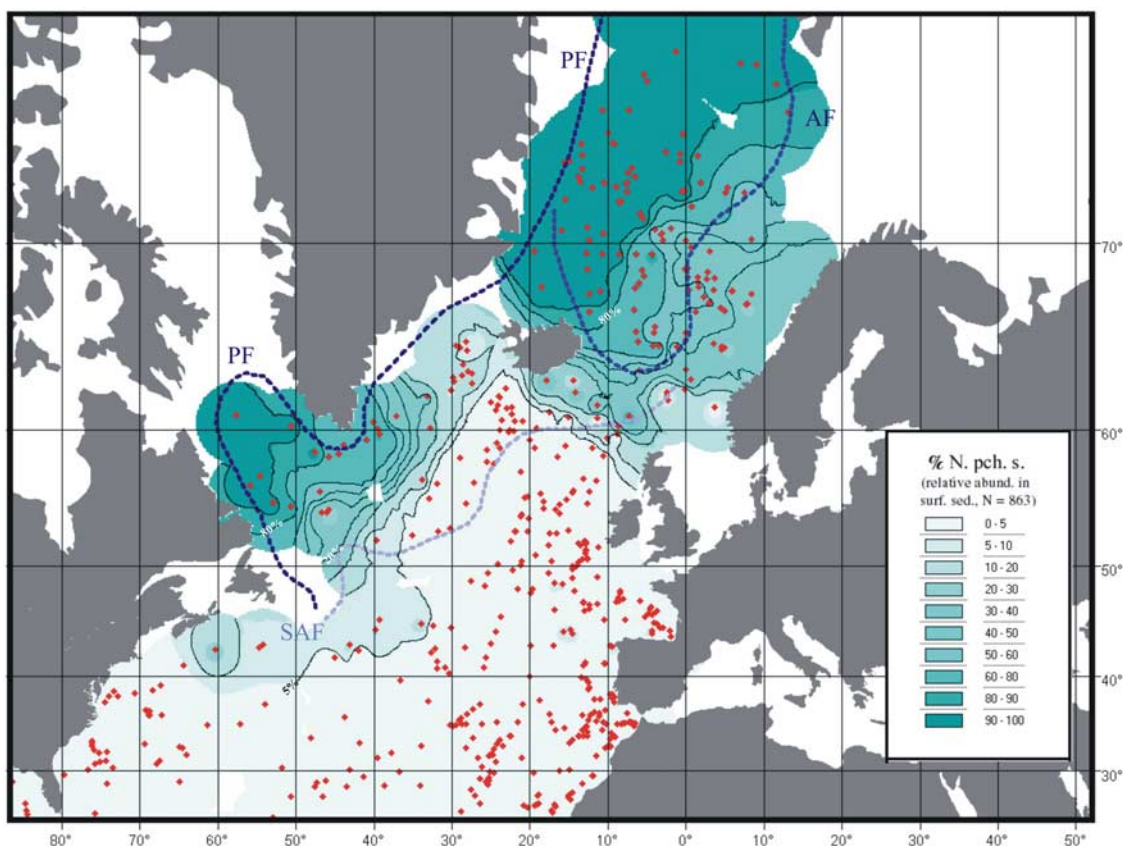


Figure 4. Present-day distribution of the polar foraminifera *N. pachyderma sinistral* in surface sediments of the North Atlantic Ocean (relative abundances mapped with ARCVIEW using the Pflaumann *et al.* [1996] database). Dotted lines locate the major hydrographic fronts (PF, Polar Front; AF, Arctic Front; SAF, Subarctic Front [after Dickson *et al.*, 1988]).

[26] The 2-D mapping exercise (Figure 5) allows us to constrain the latitudinal range where the PF could have been positioned. According to our compilation, it becomes clear that the PF reached the area only during HEs. The compilation for HE4 and HE1 produce a very consistent picture, with a PF located close to 42°N. Arctic and subarctic waters penetrated the area as far south as 36°N. The cold water may even invaded the western Mediterranean Sea [see Cacho *et al.*, 1999, 2006; Pérez-Folgado *et al.*, 2003; Sierro *et al.*, 2005]. Comparison with the $\delta^{18}\text{O}$ values of *G. bulloides* shows a coherent scheme of heavier values during HE1 and HE4, here showing typical stadial conditions as was demonstrated by Voelker *et al.* [2006]. However, it is worth noting that the northern margin records display only an attenuate increase of 1.5 to 0.5‰. This pattern is in agreement with the proximal position of the northern records to the Ruddiman belt, where a strong depletion in planktonic $\delta^{18}\text{O}$ values due to meltwater influx

overprinted the associated signal of low sea surface temperatures (in reverse to the signal directly off Iberia).

[27] The YD also records drastic changes, but the equatorward protrusion of colder water masses seems to have been limited to the installation of subarctic conditions only. This is coherent with the $\delta^{18}\text{O}$ values of *G. bulloides* that increased at that time, but only by 0.5 to 1‰ at maximum (Figures 3 and 5).

[28] The LGM and the 8.2 ka event appear to have had only minor impact on the regional SST, with no real distinctive features in the *N. pachyderma sinistral* records. These events display minor changes in the $\delta^{18}\text{O}$ signal of *G. bulloides* (Figure 5), reaching a maximum of 2.5 to 3‰ up to 40°N and of 1.5 to 2‰ in the southernmost sites during the LGM. Although for some cores, the lower sampling resolution throughout the Holocene seems to contribute to the weak representation of the 8.2 ka



Figure 5. Maps of the selected time snapshots showing potential hydrological indicators of the PF presence along the Iberian margin. (left) Seen by the relative abundance of the polar species *N. pachyderma sinistral* in the sediments (modern values in the area <5%). Delineations in between the different zones have been drawn by interpolation. Note that for HE4, the southern high value of 76% was recorded for only one level in core MD01-2444. (right) Seen by the amplitude of the calcite $\delta^{18}O$ of *G. bulloides* (before the event; during the event, methodology conforms to Cortijo et al. [2005]). For the LGM, we have chosen to indicate the averaged values of *G. bulloides* $\delta^{18}O$ over the interval to conform to the Kucera et al. [2005a] project. For detailed information on the data used, see the auxiliary material.



event, this is not the case for the LGM which furthermore encompasses a 4 ka interval.

5. Discussion

5.1. Impact of HEs Over the Iberian Margin

[29] Our compilation facilitated schematic views of the hydrological conditions prevailing along the Iberian margin during some cold events of the last 45 ka. It shows that the PF incursions were limited to episodes of major ice sheet collapses, i.e., Heinrich events. Further evidence of this drastic impact comes from other independent marine proxies, not compiled here as they were unfortunately not available for all the records considered. Alkenone data similarly revealed major decreases in SST, for instance as large as 5°C for HE4 at the southern part of the margin [e.g., Pailler and Bard, 2002; Martrat et al., 2007], and the incursion of less saline waters [Bard et al., 2000]. In the same way, other qualitative studies of plankton related groups, such as dinoflagellate cysts (dinocysts), coccolithophorids or diatoms, also registered major population changes during HEs. Dinocyst assemblages typify the migration of subpolar species over the area, in particular with the occurrence of *Bitectatodinium tepikiense* or *Nematosphaeropsis labyrinthus* [Eynaud, 1999; Sánchez Goñi et al., 2000; Boessenkool et al., 2001; Turon et al., 2003]. These species are known to be linked to low winter SST and low salinities in subboreal domains today. For coccolithophorids, the small (6–9 μm) morphotype of *Coccolithus pelagicus* s.l. was identified during HEs, in association with *N. pachyderma* sinistral and LLG/IRD peaks [Colmenero-Hidalgo et al., 2004; Parente et al., 2004]. It reveals a shift in the bioprovinces following the invasion of subpolar water masses. At an orbital scale, diatom assemblages and abundances were shown to be linked to oscillations in primary productivity related to the upwelling intensity. Major changes were identified during Terminations and at a lower amplitude during HEs [Abrantes, 1992; Thomson et al., 2000]. This was coherent also with the deposition of a typical lithofacies [Baas et al., 1997], slightly diverging from the lithology of Heinrich layers in the “Ruddiman belt,” but in accordance with the more southern position of the Iberian Margin. Benthic communities revealed a drawdown of intermediate and deep-water oxygenation during HEs [Baas et al., 1998; Schönfeld et al., 2003]. These events were furthermore all synchronous with drastic changes on the adjacent Iberian

continent, corresponding to the regression of the Mediterranean forest and the development of the steppic vegetation [e.g., Sánchez Goñi et al., 2000; Roucoux et al., 2001, 2005; Combourieu-Nebout et al., 2002; Daniaux et al., 2007; Naughton et al., 2007, 2009].

5.1.1. Upwelling System?

[30] If convergent data draw the scenario of a sudden installation of extreme environments over the area, we have, however, little evidence of how the local hydrology was affected by HEs. In particular, the dynamic of the local upwelling system is under debate [see also Salgueiro, 2006; Salgueiro et al., 2008]. For instance, the total organic content in sediments off the margin during HEs was interpreted by Pailler and Bard [2002] as indicating a decrease of the upwelling intensity, whereas Lebreiro et al. [1997] suggested the opposite, on the basis of the assumption that high amounts of heavy minerals during HEs result in an artifact in the calculation of organic carbon proportion. This assumption was supported by planktonic foraminiferal data, as high-productivity indexed species were observed in the assemblages during HEs (cores D11957P [Lebreiro et al., 1997] and MD01-2444 [Vautravers and Shackleton, 2006]). The same was demonstrated for dinocysts (core MD95-2042 [Eynaud, 1999; Sánchez Goñi et al., 2000]). Using microcharcoal fluxes to the ocean (core MD95-2042), Daniaux et al. [2007] conversely concluded to the prevalence of a southeasterly wind component over the southern Iberian Peninsula and then to a breakdown of the upwelling during HEs and D-O stadials. Their interpretation was coherent with a scenario of a compressed subtropical gyre due to the southward progression of polar water. The Azores High maintained at low latitudes and the prevailing winds from the south could not favor upwelling conditions.

[31] Regarding changes in productivity linked to HEs, Lebreiro et al. [1997] introduced an alternative solution with the idea of a system that was driven by the dynamics around icebergs rather than by the upwelling itself. Following Sancetta [1992], they hypothesized that the presence of drifting ice during HEs, supported high production by turbulent mixing and enhanced density stratification or by the supply of major or trace nutrients. This assumption is still conceivable as enhanced production has recently been observed in the vicinity of free-drifting icebergs in the Weddell Sea [Smith et al., 2007]. It is furthermore highly consistent



with our mapping, which supports the arrival of possibly nutrient-enriched polar water tongue.

5.1.2. Regional Currents?

[32] The shift of the PF and the protrusion of polar waters along the Iberian margin was coupled with a complete reorganization of the surface, intermediate and probably deep circulation in the northeastern Atlantic during HEs. Several surface current patterns around Iberia during HEs or during other cold climate phases in the Northern Hemisphere were proposed [Schönfeld and Zahn, 2000; de Abreu et al., 2003; Rogerson et al., 2004; Vautravers and Shackleton, 2006; Voelker et al., 2006; Llave et al., 2006]. Hypotheses about deep water changes in conjunction with surface water changes through their impact on the overturning circulation have been tested [Schönfeld et al., 2003; Skinner and Elderfield, 2007; Voelker et al., 2006]. These studies support a slowdown or shutoff of the Atlantic meridional overturning circulation (AMOC), coherent with studies from other areas and modeling results [e.g., Roche et al., 2004; Gherardi et al., 2005; Hall et al., 2006]. Synchronously, a drastic change in the Mediterranean Sea circulation happened [Sierro et al., 2005; Voelker et al., 2006]. Indeed, records of MOW paleointensities show that HEs, the YD, and D-O stadials were indeed periods of MOW acceleration [Faugères et al., 1985; Turon and Londeix, 1988; Schönfeld and Zahn, 2000; Cacho et al., 2000; Llave et al., 2006; Voelker et al., 2006; Rogerson et al., 2006; Toucanne et al., 2007].

[33] Our maps are picturing the events as time-static systems (Figure 5), it is then straightforward to distinguish features representative of a dynamic scheme in surface currents. However, strong thermal gradients during the HEs and the YD support a major sea surface hydrological reorganization in the area. The equatorward protrusion of polar or subpolar waters affected the margin in quite a similar way during the events considered. We can recurrently observe a demarcation at 40°N between the northern and southern part of the margin. This latitude has previously been suggested as the southern position of the PF during glacial times [Ruddiman and McIntyre, 1981; Duprat, 1983; Bard et al., 1987]. We demonstrate here that the PF probably oscillated along the northern margin only during HEs around this latitude, but a lens of coastal waters between Lisbon and Cape St Vincent recorded milder conditions at that time. This mirrors the scheme of an upper ocean winter circulation as synthesized by Peliz et al. [2005,

see Figure 18, p. 643] with warmer waters on the southern margin linked to a AC-like northward recirculation.

[34] Our maps reveal mean states of the climate episodes that will be useful for model validations. Nonetheless, high-resolution studies have revealed a much more complex pattern of climate/paleoceanographic changes within each individual HE [Bard et al., 2000; Sánchez Goñi et al., 2000, 2002; Voelker et al., 2006; Naughton et al., 2007, 2009, also presented paper, 2007].

5.2. LGM: Toward a Definitive Revision of CLIMAP Project Members [1981]

[35] Another important point emerging from our study is that the LGM was not a cold extreme but a period of warming as compared to the surrounding HEs. Following several works, both at the regional scale off the Iberian margin [Eynaud, 1999; Bard et al., 2000; Boessenkool et al., 2001; de Abreu et al., 2003; Turon et al., 2003], as well as at the scale of the North Atlantic Ocean [Sarnthein et al., 1995, 2003; Pflaumann et al., 2003; de Vernal et al., 2000, 2005, 2006], our compilation supports the presence of warm surface water masses between the continent and 10.5°W during the LGM. Only the offshore sites off Galicia recorded a slight sea surface cooling at that time (Figure 5). This cooling is, however, of relatively low amplitude, as percentages of the polar taxa *N. pachyderma* sinistral do not exceed 9%. This picture of warm, nearshore waters is coherent with previous studies which have shown the expansion of thermophilous planktonic taxa along the margin (dinocyst [Eynaud, 1999] and foraminifera [de Abreu et al., 2003; Salgueiro, 2006]) and estimated LGM-SST to be cooler by only 1 to 2°C relative to present day [de Abreu et al., 2003] or resembling averaged Holocene temperatures [Sánchez Goñi et al., 2008]. Close to the continent, a lens of subtropical waters with SST very similar to the modern ones, seems then to have prevailed, in particular south of 40°N. This draws the picture of dual subpolar and subtropical-sourced water masses at that time, depending on the latitude, and suggests the penetration of subarctic water masses, but definitely not the installation of the PF on the margin. Instead, we may expect the presence of a strong frontal structure, comparable to the modern Western Iberia Winter Front [Peliz et al., 2005] (Figure 1b). The same evidence of a tongue of warm water was previously related to a “glacial Azores Current” by Parente et al. [2004] and Rogerson et al. [2004]. A

striking point regarding this “warm” LGM sea surface signature is that the period conversely appears as a cold period on land [e.g., *Harrison et al.*, 2001]. This underlines a decoupling of atmospheric and sea surface processes, probably explained by the large volume and aerial extension of the continental ice at that time over the boreal latitudes, and moreover with a reduced vegetation cover in the vicinity of these large ice sheets. However, some recent demonstration of mitigated situations regarding pollen assemblages have been noted on the Iberian peninsula (Naughton et al., presented paper, 2007).

5.3. YD and 8.2 ka Events

[36] The YD experienced much more drastic conditions than the LGM, the later, regarding our mapping, resembling more the “8.2 ka event.” The recurrent feature which links these three events, is the partitioning of the margin at 40°N, with a northern part recording colder conditions. During the YD, penetration of subarctic waters was recognized offshore Galicia with a front splitting the nearshore and offshore sites around 40°N that was similar to the modern Western Iberian Winter Front (compare Figures 1 and 5). We can imagine that the milder conditions recorded along the shore, could be linked to a mixture of warmer subtropical waters from AC recirculations and potentially colder PC or upwelled waters.

6. Conclusions

[37] This compilation provides an integrated view of the sea surface conditions that have prevailed during the last 45 ka at the Iberian margin. Focusing on cold extremes of the late Quaternary, we have tentatively drawn schematic maps of sea surface hydrological structures, in order to visualize the migration of the PF. The maps indicated that the PF reached the Iberian margin only during HEs, prevailing in the vicinity of 40°N. The two selected HEs, HE1 and HE4, were major melting events of the Laurentian and European ice sheets. They left an imprint of very similar amplitudes regarding changes in the sea surface hydrology at the Iberian margin. Our reconstruction constrained the limits of the protrusion of polar surface waters. The maps revealed a partition of the margin into a northern area, where drastic conditions were recorded, and a southern sector where milder conditions prevailed, probably in relation to the influence of the local surface currents (Azores Current notably). The

presence of the PF at that time questions the dynamic of the Portuguese upwelling.

[38] Our compilation reveals that the 4 ka time interval of the LGM recorded quite stable sea surface conditions close to the modern ones, at maximum 1 to 2°C cooler than today, and thus a low-amplitude cooling when compared to the HEs. More drastic paleoenvironmental changes were recorded during the YD. For the LGM, the YD and also the 8.2 ka event, the northern margin experienced the most pronounced cooling, inferring further evidence for a different response of the northern and southern Iberian margin.

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