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The last glacial–interglacial transition (LGIT) in the western mid-latitudes of the North Atlantic: Abrupt sea surface temperature change and sea level implications

Teresa Rodrigues [a,b,*], Joan O. Grimalt [b], Fátima Abrantes [a], Filipa Naughton [a], José-Abel Flores [c]

[a] Marine Geology Unit, LNEG, Estrada da Portela-Zambujal, Apartado 7568, 2721-866 Amadora, Portugal
[b] Institute of Chemical and Environmental Research (C.S.I.C.), Jordi Girona, 18, 08034-Barcelona, Catalonia, Spain
[c] Department of Geology, Faculty of Sciences, University of Salamanca, 37008-Salamanca, Spain

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ABSTRACT

High resolution reconstructions of sea surface temperature ($U^{k'}_{37}$-SST), coccolithophore associations and continental input (total organic carbon, higher plant *n*-alkanes, *n*-alkan-1-ols) in core D13882 from the shallow Tagus mud patch are compared to SST records from deep-sea core MD03-2699 and other western Iberian Margin cores. Results reveal millennial-scale climate variability over the last deglaciation, in particular during the LGIT. In the Iberian margin, Heinrich event 1 (H1) and the Younger Dryas (YD) represent two extreme episodes of cold sea surface condition separated by a marine warm phase that coincides with the Bølling–Allerød interval (B–A) on the neighboring continent. Following the YD event, an abrupt sea surface warming marks the beginning of the Holocene in this region. SSTs recorded in core D13882 changed, however, faster than those at deep-sea site MD03-2699 and at the other available palaeoclimate sequences from the region. While the SST values from most deep-sea cores reflect the latitudinal gradient detected in the Iberian Peninsula atmospheric temperature proxies during H1 and the B-A, the Tagus mud patch (core D13882) experienced colder SSTs during both events. This is most certainly related to a supplementary input of cold freshwater from the continent to the Tagus mud patch, a hypothesis supported by the high contents of terrigenous biomarkers and total organic carbon as well as by the dominance of tetra-unsaturated alkenone ($C_{37:4}$) observed at this site.

The comparison of all western Iberia SST records suggests that the SST increase that characterizes the B-A event in this region started 1000 yr before meltwater pulse 1A (mwp-1A) and reached its maximum values during or slightly after this episode of substantial sea-level rise. In contrast, during the YD/Holocene transition, the sharp SST rise in the Tagus mud patch is synchronous with meltwater pulse 1B. The decrease of continental input to the mud patch confirms a sea level rise in the region. Thus, the synchronism between the maximum warming in the mid-latitudes off the western Iberian margin, the adjacent landmasses and Greenland indicates that mwp-1B and the associated sea-level rise probably initiated in the Northern Hemisphere rather than in the South.

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1. Introduction

The increase of high-latitude summer insolation that followed the Last Glacial Maximum (LGM), favoured the northern hemisphere ice sheets retreat and triggered a 120–130 m increase in global sea level (Fairbanks, 1990; Fairbanks et al., 2005; Peltier and Fairbanks, 2006; Stanford et al., 2006). Superimposed on this orbitally induced long-term warming trend, abrupt widespread millennial-scale climate variability have punctuated the last deglaciation in Greenland (Alley et al., 1993; Dansgaard et al., 1993; Johnsen et al., 2001), in the North Atlantic (Ruddiman and McIntyre, 1981; Boyle and Keigwin, 1987; Duplessy et al., 1992; Bond et al., 1993; Keigwin and Lehman, 1994; Andrews et al., 1995; Hughen et al., 1996; Andrews et al., 1999; Rühlemann et al., 1999; Peterson et al., 2000; McManus et al., 2004) in North America and Europe (Iversen, 1954; Mangerud et al., 1974; Peteet et al., 1993; Naughton et al., 2007a). Within this millennial scale climate variability, extreme cold episodes such as H1 and YD have been detected in most of the northern hemisphere archives. H1 is likely the result of abrupt and massive icebergs discharges into the north Atlantic region (Bond et al., 1993) while the YD is related to catastrophic drainage episodes of the proglacial Lake Agassiz (Teller et al., 2002). Although resulting from different processes, climate simulations suggest that the input of freshwater in the North Atlantic region disturbed the general pattern of thermohaline

* Corresponding author at: Laboratório Nacional de Energia e Geologia (LNEG), I. P., Unidade de Geologia Marinha, Estrada da Portela, Bairro do Zambujal - Alfragide Apartado 7586, 2720-866 Amadora, Portugal. Tel.: +35 966785586.
E-mail address: teresa.rodrigues@ineti.pt (T. Rodrigues).

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circulation and triggered a substantial atmosphere and oceanic cooling in both cases (Paillard and Labeyriet, 1994; Fawcett et al., 1997; Seidov and Maslin, 1999; Ganopolski and Rahmstorf, 2001; Knutti et al., 2004; Stouffer et al., 2006).

Northern hemisphere archives detect a short warm phase that reflects the Bølling–Allerød (B-A) episode and occurs between H1 and YD cold events (Iversen, 1954; Mangerud et al., 1974; Bond et al., 1993; Dansgaard et al., 1993; Peteet et al., 1993; Keigwin and Lehman, 1994; Hughen et al., 1996; Johnsen et al., 2001; Teller et al., 2002; McManus et al., 2004; Naughton et al., 2007a). Another warm episode occurred after the YD and marks the onset of the present-day interglacial, the Holocene. The millennial-scale climate variability that characterizes the last deglaciation had also a great impact on the northern hemisphere ice sheet dynamics (McCabe and Clark, 1998; Dyke, 2004; McCabe et al., 2005; Millera et al., 2005; McCabe et al., 2007) and sea level changes (Fairbanks 1989; Bard et al., 1990; Bard et al., 1996; Lambeck and Chappell, 2001). This last deglaciation is characterized by episodes of sudden sea level rise, called meltwater pulses (mwp), 2A, 1B and 1A (Fairbanks, 1990; Bard et al., 1996; Yokoyama et al., 2000; Fairbanks et al., 2005), however, controversy surrounds the precise timing, sources and global impact of these meltwater pulse episodes (Bard et al., 2010).

This millennial scale climate variability has also been recorded along the Portuguese continental Margin (Lebreiro et al., 1997; Bard et al., 2000; Shackleton et al., 2000; Abreu et al., 2003; Martrat et al., 2007; Salgueiro et al., 2010). However, none of these works show the impact of the continental input during these particular events that characterizes the last deglaciation. Also, although this margin records a gradual latitudinal sea surface temperatures decrease during the last 150 ka (Salgueiro et al., 2010) and in particular during Last Glacial Maximum (LGM) (Morey et al., 2005), none of these studies provide information about the longitudinal pattern of SST from the continent to the open sea.

This paper presents the first high-resolution multi-proxy study ($U^{k'}_{37}$ and coccolithophore associations, total of organic carbon and higher plant *n*-alkanes and *n*-alkan-1-ols) of core (D13882) retrieved from a Tagus mud patch of the shallow inner shelf. The reconstruction of continental input and its implication on the SST allows to further explore sea level and climatic variability on the south-western Iberian margin during the last glacial-interglacial transition. Furthermore, the fact that our site is located on the continental shelf provides an integrated picture of the marine and continental conditions, and its comparison to core MD03-2699 and other available deep-sea records from western Iberian margin allows for a latitudinal and cross-shelf evaluation.

2. Material and methods

Cores D13882 and MD03-2699 were retrieved from 88 m and 1865 m water depth from the western Iberian margin (38°38.07′N, 9°27.25′W and 39°02.20′N, 10°39.63′W, respectively) using the National Oceanography Centre long-piston coring system and the giant CALYPSO corer during the Discovery 249 and PICABIA oceanographic cruises on board of R/V Discovery and Marion Dufresne (Fig. 1; Table 1). Core D13882 is 12 m long and covers the last 13.1 ka providing the highest resolution palaeoclimatic record of the western Iberian margin for the Last Glacial–Interglacial Transition (LGIT). In contrast, core MD03-2699, is 26.5 m long and covers the last 570 ka (from Marine Isotopic Stages (MIS 1 to 15). In this paper, we will focus this study in the first 3 m of core MD03-2699.

102 and 103 samples from cores D13882 and MD03-2699, respectively, were used to quantify haphtophyte-synthesized C_{37} alkenones and terrigenous biomarkers, such as C_{23}–C_{33} *n*-alkanes and *n*-alkan-1-ols C_{20}–C_{30}, following the methods described in Villanueva and Grimalt (1997). Samples (2 g) were freeze-dried and extracted by sonication with dichloromethane and then hydrolysed with 6% potassium hydroxide in methanol. After derivatization with bis(trimethylsilyl)trifluoroacetamide, they were analyzed with a Varian gas chromatograph Model 3800 equipped with a septum programmable injector and a flame ionization detector. The concentrations of each compound were determined using *n*-nonadecan-1-ol, *n*-hexatriacontane and *n*-tetracontane as internal standards. Sea surface temperatures were estimated from the relative composition of C_{37} unsaturated alkenones using the $U^{k'}_{37}$ index and the core top calibration equation described in Müller et al. (1998).

Total organic carbon (TOC) was determined using a CHNS-932 Leco elemental analyzer on aliquots (2 mg) of dry and homogenized sediments. Organic Carbon content was calculated using the loss of ignition procedure (heat up to 500 °C during 3 hours). For coccolithophore analysis smear slides were prepared using the methodology described in (Flores and Sierro, 1997) and identified and counted various species with the use of a light polarizing microscope (1250×).

2.1. Age model

The age model of core D13882 is based on four accelerator mass spectrometer (AMS) ^{14}C dates obtained on *Turritella sp.* and bivalve shells (Table 2, Fig. 2A). AMS ^{14}C dates were calibrated using CALIB Rev 5.0 program and the "global" marine calibration dataset (marine 04.14c) (Stuiver and Reimer, 1993; Hughen et al., 2004, 2006).

It is known that the local reservoir ages have changed in the past in particular during the last glacial interglacial transition at high latitudes of the eastern North Atlantic (e.g. Bard et al., 1994; Waelbroeck et al., 2001; Sarnthein et al., 2007) and of the western North Atlantic (Waelbroeck et al., 2001; Sarnthein et al., 2007). The Iberian margin is located far away from those sites, and, therefore, we assumed a regional marine age reservoir of −400 yr for this region as previously suggested by Bard et al. (2004) and Stuiver et al. (2005).

Calibrated kilo years before present will be referred as ka and were determined using the 95.4% (2 sigma) confidence intervals and the relative areas under the probability curve as suggested by Stuiver et al. (2005). Our sampling interval is of about 1 cm reflecting a temporal resolution of approximately 2.5 yr.

To construct the age model of core MD03-2699, we used AMS ^{14}C dating and SST correlation between this core and core MD01-2443 (Martrat et al., 2007). Eight AMS ^{14}C dates were obtained on monospecific foraminifers' samples (*Globigerina inflata*) and one on a multiespecies foraminifer sample (Table 2). However, only four dates were included in the age model since the other four seem contaminated with old material (Fig. 2B). A contamination that can be the consequence of increasing the slope instability derived from changes in the sea level during the last deglaciation, as proposed by Lebreiro et al. (2009).

For the bottom part our SST record was correlated with other south-western Iberian margin SST records (Bard, 2001; Martrat et al., 2007) (correlation coefficient of 0,97%) (Fig. 3).

The sedimentary rates are in the order of 14 cm/ka allowing an average resolution of 140 year for a 2 cm sampling interval.

3. Results

3.1. Sea surface temperature

SST has been reconstructed in two cores, the MD03-2699 located at 1865 m water depth, and a shallow core D13882 collected

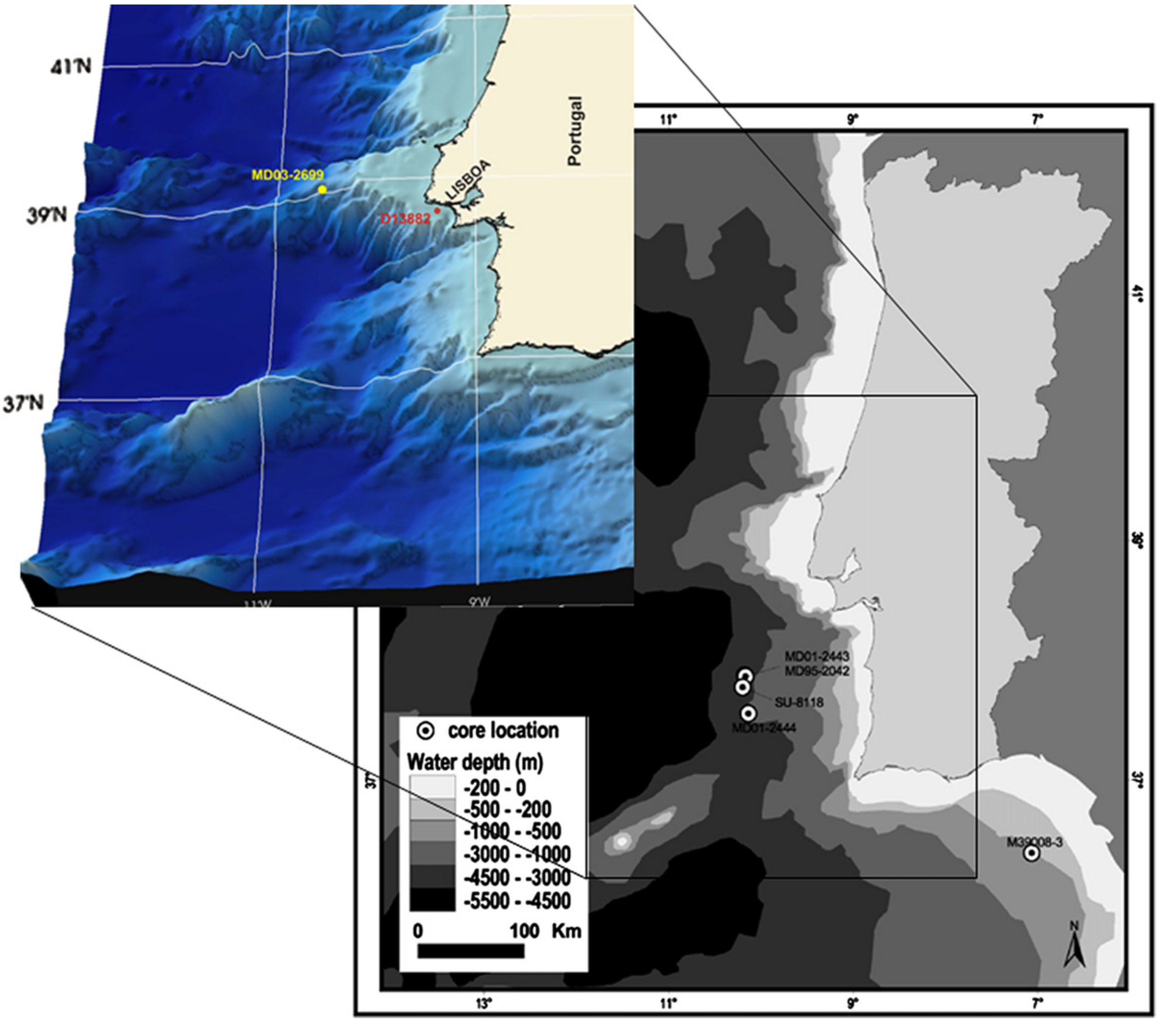

Fig. 1. Area of study indicating the position of western Iberian margin cores D13882 and MD03-2699 (this study) and other discussed in this manuscript.

at 88 m water depth for the last 13.5 ka. Core MD03-2699 shows SST changes between 13 °C to 18 °C, the coldest temperature is recorded at around 18 ka and the maximum at the beginning of the Holocene and warmer conditions (11.2 ka). During the last deglaciation our records show a gradual SST increase from 13 °C at 16.95 ka to 17 °C at 13.5 ka. After that, this SST increase was interrupted by a drop of 1 °C at around 12.6 ka. Around 12 ka the SST increases again to 18.3 °C in 800 years. During the Holocene SST shows a slight and gradual decrease trend towards the Present. However, short cooling episodes of 1–1.5 °C punctuate the general cooling trend at around 8.2, 6.2, 4.3 ka.

Core D13882 SST record for the last 11.5 ka is published in (Rodrigues et al., 2009), for this study only the data for the period between 13.5 and 11.5 ka is considered. The record reveals a cold period that is actually composed by two cold phases framed by a relatively warmer period (11 °C). The first plateau with SSTs of 8 °C occurs from 12.5 to 12 ka, while the second phase, with 1 °C warmer temperatures, goes up to 11.6 ka. The final transition to the present interglacial is marked by a change of 7 °C (from 10 °C to 17 °C) in a short period of time, 40 years, from 11.6 to 11.56 ka.

Table 1
Location, water depth and length of the shallow and deep sea cores from western Iberian margin.

Core ID	Core type	Water depth (m)	Core length (m)	Latitude (N)	Longitude (W)
D13882	Piston	88	13.61	38°38.07′	9°27.25′
MD03-2699	Piston	1865	26.5	39°02.20′	10°39.63′

3.2. Other proxies analyzed in D13882

For the shallow core D13882 we also generated the marine and terrigenous biomarker signal for the study period. The terrigenous biomarkers (*n*-alkanes and *n*-alcohols) show a decrease trend with the maximum values around 2000 ng/g at 13.5 ka and lower values around 300 ng/g after 11.2 ka. The ratio C_{29}/C_{31} used to identify the vegetation type in the source, shows higher and stable values until 11.5 ka, indicating a dominance of the C_{29} alkane typical of scrubs and other lower vegetation. After the 11.5 ka the ratio shows a decreasing trend to values close to 1.

The organic content of the sediments was quantified by TOC measurements and shows higher values close to 0.8%wt, on average, until 11.5 ka, followed by a decrease to values around 0.4% wt. Total alkenones concentration shows lower values before 11.5 ka and a slight increase in the Holocene (<11.5 ka). Within the alkenones concentration we highlight the presence of the $C_{37:4}$ only during the period between 13.5 and 11.5 ka, a trend also observed for the coccolithophore specie *G. mullerae* which has its highest percent abundance during this period.

Table 2
Radiocarbon ages of cores D13882 and MD03-2699.

Lab Code	Sample ID	Depth (cm)	Material	14C Age (BP)	Calendar Age BP	Error
OS- 37707	D13882	798–799	*Turritella sp.*	10450	11578	75
KIA 27307	D13882	820–821	Bivalve	10490	11627	70
OS- 37708	D13882	975–976	Shell F > 2 mm	11100	12749	50
OS- 37709	D13882	1140–1141	Bivalve	11500	13011	70
KIA 30539	MD03-2699	3–4	*G. ruber white & pink; G. inflata*	2025	1189.5	25
KIA 30540	MD03-2699	25–26	*G. inflata*	3650	3078	30
KIA 30541	MD03-2699	61–62	*G. inflata*	5810	5785.5	35
KIA 31227	MD03-2699	99–100	*G. inflata*	10165	11168.5	45
KIA 30542	MD03-2699	113–114	*G. inflata*	13770	15336	60
KIA 30543	MD03-2699	145–146	*G. inflata*	25270	29720	+210/−200
KIA 31228	MD03-2699	164–165	*G. inflata*	23320	27770	160
KIA 35296	MD03-2699	210–211	*G. inflata*	24930	29530	+230/−220
KIA 35297	MD03-2699	278–279	*G. inflata*	>44910		

4. Discussion

4.1. Climate variability in the western Iberian margin

For the last 21 ka, the Iberian margin experienced the most extreme cold sea surface conditions during H1 as demonstrated by alkenone based SST estimates for cores MD03-2699, SU81-18 (Bard et al., 2000), MD95-2042 (Pailler and Bard, 2002), MD01-2444 (Martrat et al., 2007) and M39-008 (Cacho et al., 2001) (Fig. 4). Although the SST values recorded for H1 are quite similar (11° to 13 °C) in most of the western Iberian margin cores (Fig. 4), the southernmost one (core M39-008) is marked by 1 °C warmer SSTs (>14 °C), suggesting a latitudinal warming gradient along the Iberian margin during Heinrich event 1. A similar air temperature gradient was also detected on the continent where the temperate and pine forest contraction, reflecting extreme cold temperatures, was more pronounced in the north rather than in the south Iberian Peninsula during the Oldest Dryas (terrestrial equivalent of H1 event) (Naughton et al., 2007b). The sharp increase in SST around 15 ka marks the beginning of the warm Bølling–Allerød interstadial (B-A) in the western Iberian margin but maximum SSTs peak at around 14 ka (Fig. 4). This warm event is marked by alkenone based SST estimates of about 15° to 17 °C in cores: MD03-2699, SU81-18 (Bard et al., 2000), MD95-2042 (Pailler and Bard, 2002), and MD012444 (Martrat et al., 2007) (Fig. 4). Again, SST values are highest (about 18° to 19 °C) in core M39-008 (Cacho et al., 2001) located in the southernmost area of the Iberian margin. This climate improvement that characterizes the B-A interstadial, is well expressed in marine and terrestrial pollen sequences from the Iberian margin and adjacent landmasses (Allen et al., 1996; Peñalba et al., 1997; Turon et al., 2003; Naughton et al., 2007b). The expansion of temperate and humid forest suggests increasing temperatures and precipitation in the southern part of the Iberian Peninsula, (Allen et al., 1996; Peñalba et al., 1997; Turon et al., 2003; Naughton et al., 2007b). The SST record of core D13882 does not cover the entire B-A but only the final stages of this event (Fig. 4). This final stage is marked by SST values around 11°C, that is, 3 °C to 5 °C lower than those recorded in the other western Iberian margin sites (Fig. 4). This core is located closer to the continent, and in the area of influence of the Tagus River (Fig. 1). Indeed, although the total organic carbon (TOC) curve parallels the total alkenones record of core D13882 in most of the sequence (Fig. 4), the high TOC content (1% W) contrasts with low total alkenone concentrations (less than 100 ng/g) during the B-A event. This suggests that the TOC increase is likely the result of higher terrigenous input as in fact indicated by the *n*-alkanes and *n*-alkan-1-ols increase in concentrations (about 2500 ng/g and 5000 ng/g, respectively; Fig. 5) supporting the hypothesis of higher terrestrial input during the B-A event. The percentage of $C_{37:4}$ recorded has high values during extreme cold events such as Heinrich events. In the north Atlantic their presence is attributed to an increase of cold water input from iceberg melting (Bard et al., 1987; Rosell-Melé et al., 1998; Schulz et al., 2000; Rosell-Melé, 2004), the dominance of $C_{37:4}$ during the B-A in core D13882 suggests that substantial amounts of cold

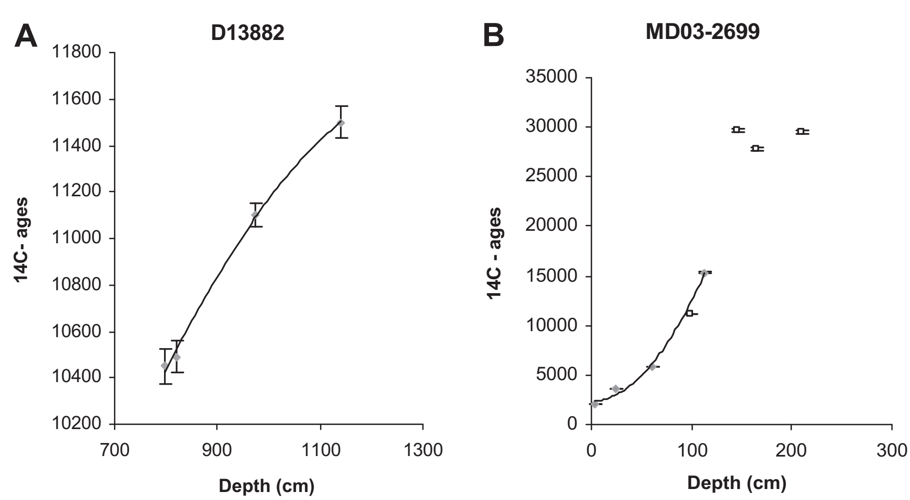

Fig. 2. Chronostratigraphy of cores D13882 (A) and MD03-2699 (B). Grey diamonds with error bars represent the radiocarbon ages. Three ages have been rejected (see explanation in the text). A second-order polynomial adjustment performed in each core is represented by the continuous dark line.

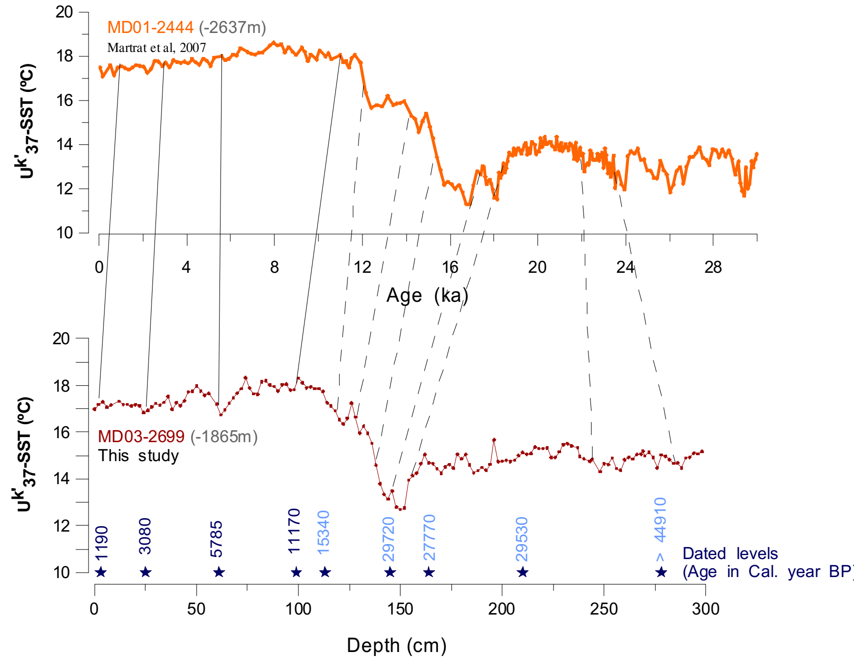

Fig. 3. Correlation of the $U^{k'}37$-SST values in cores MD03 2699 on depth scale (this study) and MD01-2444 on age scale (Martrat et al., 2007). The black lines are the dated levels and the dashed lines are the tuning between the two cores.

freshwater were reaching the study area. However, this freshwater input cannot be explained by ice melting processes but rather by increasing fluvial input during the B-A event.

Following these warm episode, cold conditions return to most of Iberian margin deep sea cores which reflect the YD event at around 12.9 ka (Fig. 4). Temperate forest decline, reflecting atmospheric cooling, was less extreme during the YD than during H1 event but occurred both in the north- and southwestern Iberian Peninsula (Allen et al., 1996; Peñalba et al., 1997; Turon et al., 2003; Naughton et al., 2007b) In addition, the expansion of semi-desert plants' associations in the Iberian Peninsula vegetation cover during the YD event indicates also lower humidity conditions relatively to the

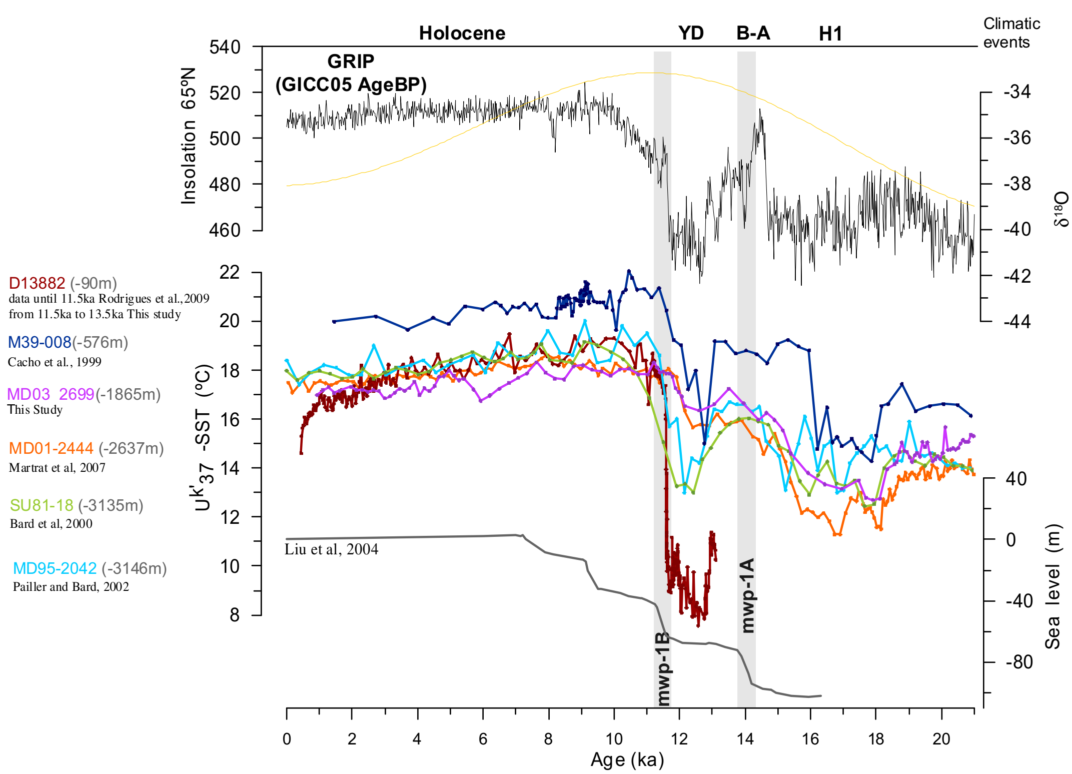

Fig. 4. Alkenone-based SST values of the last 21 ka obtained in the Iberian margin cores: D13882 and MD03 2699 (this study); M39-008 (Cacho et al., 2001); MD01-2444 (Martrat et al., 2007); SU81-18 (Bard et al., 2000); MD95 2042 (Pailler and Bard, 2002). Black line, $\delta^{18}O$ record of Greenland ice core GRIP (GICC05 age scale). In yellow the Isolation curve at 65°N (Berger, 1978). High resolution Sea level curve (Liu et al., 2004).

previous B-A warm phase (Allen et al., 1996; Turon et al., 2003). SST was colder in the shallow (about 8 °C at 90 m depth; core D13882: Fig. 4) and deepest sea cores (about 12 to 13 °C at depths > 3000 m; SU8118 (Bard et al., 2000), MD95-2042 (Pailler and Bard, 2002)) than in mid-water depth cores (Fig. 4). It is believed that the catastrophic episode of Lake Agassiz freshwater discharges into the North Atlantic lead to a substantial cooling of the North Atlantic and Europe, resulting in a significant reduction in the Meridional Overturning Circulation (MOC) (McManus et al., 2004; Eltgroth et al., 2006). However, this mechanism alone cannot explain the observed SST differences between the shallow, deep and mid-depth cores. As for the B-A event, the shallow core was probably also affected by cold freshwater input by the Tagus river during the YD event. The percentages of $C_{37:4}$ are higher than those recorded in offshore cores such as in core SU8118 (Bard et al., 2000), where the occurrence of $C_{37:4}$ above 5% is taken as reflecting low-salinity water (Bard et al., 1987; Rosell-Melé et al., 1998; Schulz et al., 2000) and has been used for the reconstruction of the North Atlantic polar front position during the H1 (Bard et al., 2000). Furthermore *G. mullerae* abundances are synchronous with high percentages of $C_{37:4}$ (Fig. 5), which comes as a confirmation that the area was under the influence of cold freshwater masses also during the YD

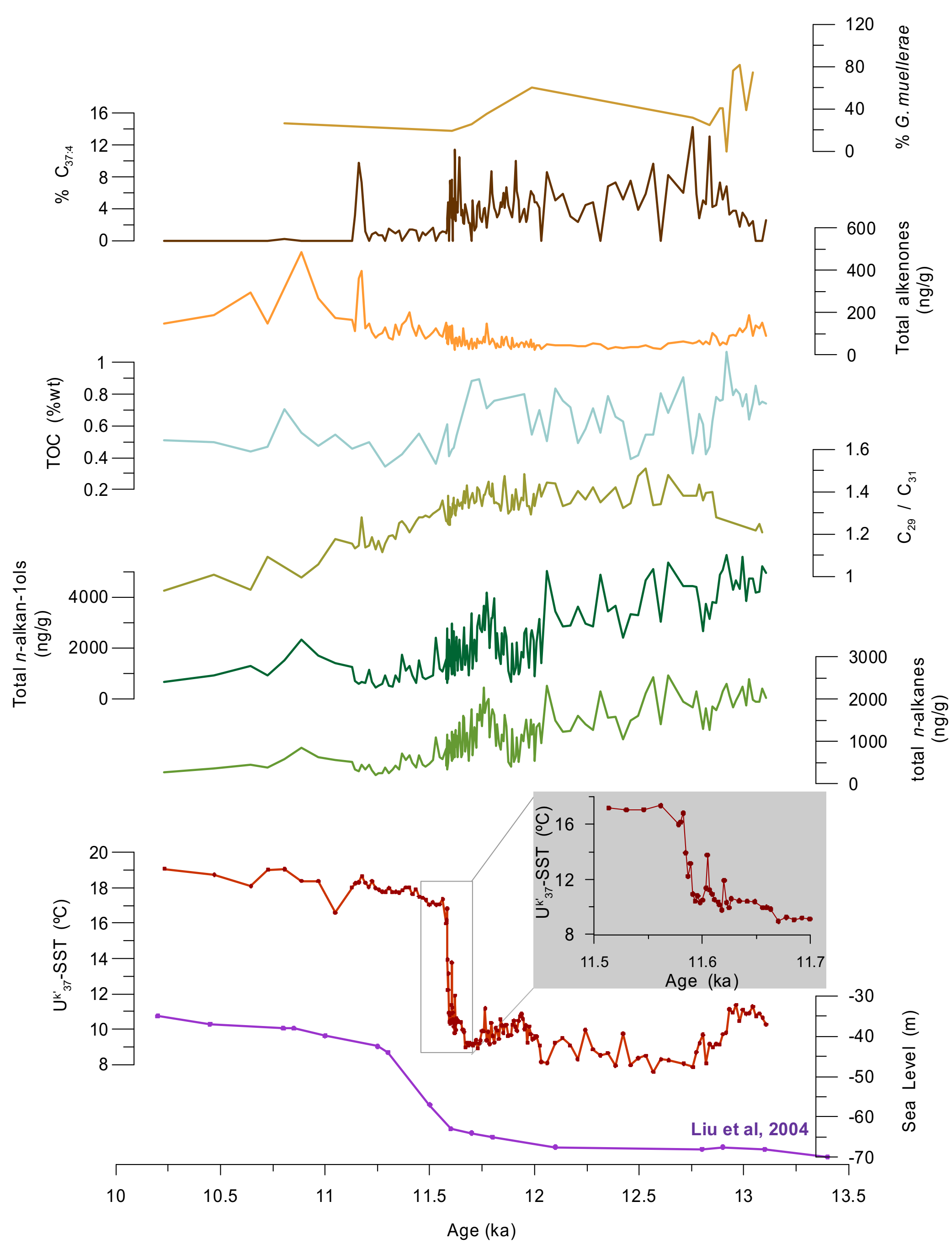

Fig. 5. Multi-proxy results obtained in core D13882. From the top to the bottom: % of *G. mullerae*; % of tetra-unsaturated alkenone ($C_{37:4}$); total alkenone concentration (ng/g); total organic carbon content (%wt); Alkanes ratio of C_{29}/C_{31}; total n-alkanes concentration (ng/g); alkenone-based $U^{k'}_{37}$-SST including a zoom (200 years); high resolution sea level curve (Liu et al., 2004).

(Fig. 5) (Flores and Sierro, 1997; Wells and Okada, 1997; Flores et al., 1999).

The high resolution alkenone based SST estimates of core D13882 further show a complex pattern within the YD event, composed by two cold phases framing one relatively warm period (Figs. 3 and 4). A 3 °C cooling episode detected at around 12.9 ka is followed by a continuous increase in temperatures up to 11 °C at 11.9 ka and by a second phase cooling (about 2 °C) after 11.8 ka. A similar complex pattern has been detected in the temperature record of Greenland (Alley, 2000; Lowe et al., 2008), in the $\delta^{18}O$ contained in ostracods shells from lake Ammersee (VonGrafenstein et al., 1999), and in the Alboran sea surface temperatures (Cacho et al., 2001). Concomitantly marine pollen records from western Iberian reveal changes in the deciduous forest cover in the neighbor continent (Turon et al., 2003; Naughton et al., 2007b). The comparison of western Iberian margin SST records (cores D13882 and MD03-2699) with the $\delta^{18}O$ NGRIP record from Greenland (Johnsen et al., 2001; North Greeland Ice Core project member, 2004; Rasmussen et al., 2006) demonstrates that the timing of the YD event is similar at both areas, occurring between 12.8 ka and 11.6 ka (about 1200 year long) (Fig. 4). These age limits coincide with those obtained from $^{230}Th/^{234}U$ measurements in the Barbados corals (Fairbanks, 1990) and in varved sediments from European lakes (Goslar et al., 2000). The timing of the YD in northwestern Scotland is also synchronous with that of NGRIP as shown by the presence of the Verdi ash layer in both records (Kroon et al., 1997), as well as the YD episode recorded in Cariaco basin (Hughen et al., 1998).

The transition zone between the YD and the Holocene is marked by a steep increase of SST in most Iberian margin records (cores: D13882, SU81-18, MD95-2042 and M39-008; Fig. 4) (Bard et al., 2000; Cacho et al., 2001; Pailler and Bard, 2002) at 11.6 ka which is synchronous with the expansion of temperate trees in the neighboring continent (Turon et al., 2003; Naughton et al., 2007b) In addition, this sharp SST increase is contemporaneous with a marked sea level rise and subsequent retreat of the coastline in the Tagus mud patch as revealed by the western Portugal shore line evolution (Dias et al., 2000). A sea level rise likely to favour sedimentation within the estuary and preclude the input of terrigenous material into the Iberian shelf. This is in fact the case at the Tagus mud patch region where TOC decrease from 1W% to less than 0.5W % and the higher plant proxies concentration also decrease markedly (*n*-alkanes: 2000 ng/g to 500 ng/g and *n*-alkane-1-ols: 5000 ng/g to 1000 ng/g) during the YD/Holocene transition (Fig. 5). Furthermore, the decreases of $C_{37:4}$ and G. *mullerae* percentages in core D13882 also testify to the decrease of continental cold freshwater input into this area (Fig. 5).

4.2. Is there evidence of the meltwater pulses in the Iberian Margin?

In recent years, many studies have been carried out to understand the timing, sources and global impact of the mwp episodes during the last deglaciation (Clark et al., 1996; Clark et al., 2002; Weaver, 2003; Liu and Milliman, 2004; McManus et al., 2004; Bassett et al., 2005; Peltier, 2005; Tarasov and Peltier, 2005; Stanford et al., 2006; Roche et al., 2007). Based on AMS ^{14}C (Fairbanks, 1989) and U-Th (Bard et al., 1990; Bard et al., 1996) dating of the Barbados and Tahiti coral records, two major mwp have been identified after the well-known Heinrich 1 event, the mwp-1A and mwp-1B. These events occurred at around 14.5–13.8 ka (mwp -1A) and 11.5–11.1 ka (mwp -1B) reflecting a sea level rise of about 20 and 15 m, respectively (Liu and Milliman, 2004).

Studies from the South China Sea suggest that the mwp-1A was synchronous with the warm Bølling episode, which has contributed to massive meltwater discharges and subsequent weakening of deep-water formation in the North Atlantic region (Pelejero et al., 1999; Kienast et al., 2003). In addition, it has been proposed that the abrupt resumption of the MOC at the beginning of B-A event together with the increase of high latitude summer insolation accelerated the Laurentide Ice sheet melting and triggered the mwp-1A (McManus et al., 2004; Peltier, 2005). These hypotheses have been corroborated by (Flower et al., 2004) based on $\delta^{18}O$ records of the Gulf of Mexico, which suggest that the SST warming precede the Laurentide Ice Sheet decay and consequent input of freshwater into the ocean. This mechanism should induce the subsequent reduction or a shutdown of the conveyor belt (Roche et al., 2007). However, the mwp-1A occurred more than 1000 years before the next significant change in the MOC which arises during the Younger Dryas cold interval (Clark et al., 2002). This implies that the mwp occurring during the B-A event primarily originated from Antarctica reducing its impact on the Atlantic MOC (Clark et al., 2002). Other studies suggest that the mwp-1A did not occur during the fast warming phase of the Bølling but rather through the Older Dryas cooling event (Stanford et al., 2006).

The Iberian margin cores show that the rise in sea surface temperatures, which characterizes the B-A event, started 1000 yr before the mwp-1A and attain its maximum values during or slightly after this substantial sea level rise episode (Fig. 4). A result in accordance with the findings of (Clark et al., 2002), where the mwp-1A occurred more than 1000 years before the next significant change in the MOC which arises during the Younger Dryas cold interval, leading the authors to consider that this mwp event was primarily originated from Antarctica what reduced its impact on the Atlantic MOC. Although we cannot infer about the origin/source of this meltwater pulse, the fact is that the western Iberian margin sea surface temperatures show a gradual increase synchronous with the increase of the high latitude summer insolation and attains its maximum values after the Older Dryas event.

The YD/Holocene transition is marked by a gradual increase of SST in the lowermost resolution records (MD03-2699, SU81-18, MD95-2042, MD012444 and M39-008) and by an abrupt SST increase in the higher resolution D13882 record (Fig. 4). This SST abrupt increase is contemporaneous with a substantial warming over Greenland (Johnsen et al., 2001) and with a maximum in high-latitude summer insolation (Berger, 1978). Additionally, the decrease of continental input into the Tagus mud patch, as revealed by D13882 terrigenous biomarkers' and TOC record, suggest that the sea level rise was already favoring the deposition of continental material in the estuaries rather than in the ocean. Thus, the abrupt SST increase of 7 °C in 40 years, simultaneous with the decrease of continental input in the Tagus mud patch, occurs synchronously with the well known mwp-1B (Fig. 5). This event, occurred between 11.5 and 11.2 ka and is marked by a sea level rise of about 13 m (Figs. 3 and 4) (Liu and Milliman, 2004; Bard et al., 2010). Besides the substantial sea level rise, the expansion of temperate and humid trees in the adjacent landmasses (Naughton et al., 2007b) would have also precluded the land-sea transport of continental material.

All these evidences show that the timing of the maximum marine and terrestrial warming at the mid-latitudes of western Iberia, together with maxima temperatures over Greenland is synchronous with that of the mwp-1B. This suggest that this drastic melting episode (mwp-1B) must have been initiated in the Northern Hemisphere rather that in the South.

5. Conclusions

The comparison between marine climatic indicators (U^{k-}_{37}-SST), coccolithophore associations and proxies of terrestrial input (total organic carbon and higher plant *n*-alkanes and *n*-alkan-1-ols) of core D13882 retrieved from the inner-shelf Tagus mud patch provides the highest resolution record of the last deglaciation and in particular of the last glacial-interglacial transition (LGIT) for western Iberia. Sea surface temperatures downcore have been compared with those obtained for deep-sea core MD03-2699 and other deep-sea records from the western Iberian margin.

The SST records of the deeper cores demonstrate H1 as the extreme cold event (11° to 13 °C) of the last 21 ka. During this period the SST is less cold in the southernmost site, suggesting a latitudinal gradient along the Iberian margin.

The beginning of the Bølling–Allerød interstadial is marked by a sharp SST increase to 15°–17 °C, around 15 ka. This SST increase has started 1000 yr before the mwp-1A and attain its maximum values during or slightly after this substantial sea level rise episode.

The shallow core D13882 only covers the last part of this warm interstadial (B-A) and shows lower temperatures (11 °C) when compared to the deeper cores. This site reflects the dominance of $C_{37:4}$, higher concentration of terrigenous biomarkers and TOC provide evidence of the arrival of cold freshwater to the area. This freshwater input although possibly associated with the melting of ice sheets is most certainly reflecting the strong input of freshwater by the Tagus River.

Returning cold conditions mark the Younger Dryas event at around 12.9 ka. During the YD, as for B-A, SSTs were lower by about 4–5 °C in the Tagus mud patch than in the open sea sites, indicating the continuity of the cold freshwater input by the River.

The high-resolution alkenone based SST estimates of core D13882 further show that the YD event is composed by two cold phases surrounding a relatively warm one. That is, a 3 °C cooling episode detected at around 12.9 ka is followed by a continuous temperature increase up to 11 °C at 12.0 ka and followed by a second cooling phase (about 2 °C) after 11.8 ka.

The glacial-interglacial transition is marked by a sharp SST increase from 10 °C to 17 °C in the shallow core, contemporaneous with a strong sea level rise and consequent retreat of the coastline in the Tagus mud patch which probably precluded the input of terrigenous material to the inner-shelf.

The abrupt SST increase of 7 °C in 40 yr together with the decrease of continental input to the Tagus mud patch occurs simultaneously with the well-known mwp-1B. The synchronous timing between the drastic melting episode 1B and the abrupt marine and terrestrial warming at the mid-latitude of western Iberia, together with warm temperatures over Greenland, suggest that this melting water episode and consequent sea level rise were initiated in the Northern Hemisphere.

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