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Abstract
Paleoenvironmental records in Europe describing paleofires extending back to the Last Interglacial have so far been unavailable. Here we present paleofire results from the combined petrographic and automated image analysis of microcharcoal particles preserved in marine core MD95-2042 retrieved off southwestern Iberia and covering the last climatic cycle. The variability of microcharcoal concentrations reveals that the variability of fire emissions is mainly imprinted by the 23 000 yrs precessional cycle. A focus on the Last Glacial Period further shows that paleofires follow the variability of Dansgaard-Oeschger oscillation and Heinrich events and, therefore, parallel the variability of atmospheric temperatures over Greenland detected in ice cores. There is no evidence for fire increase related to human activity. The variability of fire emission by-products for the Last Glacial Period is interpreted in terms of changes in biomass availability. Low fire activity is associated with periods of drought which saw the development of semi-desert vegetation that characterised stadial periods. Fire activity increased during wetter interstadials, related to the development of open Mediterranean forests with more woody fuel availability.

Keywords: marine sequence; microcharcoal; fire; Iberian Peninsula; Dansgaard-Oeschger oscillations; Heinrich events

1. Introduction

Fires occur today in both hemispheric regions at every period of the year and affect the chemistry of the atmosphere by the release of radiatively and chemically active gases i.e. greenhouse gases (carbon dioxide, methane, nitric oxide), and particulates (Crutzen et al., 1979; Lobert et al., 1990; Thonicke et al., 2005; Carmona-Moreno et al., 2005). Biomass burning appears therefore to be a significant driver for climate change (Clark et al., 1997). The study of fire regime in the past is therefore an instructive approach to tackling changes in atmospheric conditions and discusses the feedback mechanism through fire-derived greenhouse gases.

Several studies on marine cores from the Southern Hemisphere have shown a link between fire history and orbital scale climatic variability over last Glacial/Interglacial cycles (Beaufort et al., 2003; Kershaw et al., 2002; Bird and Cali, 1998; Verardo and Ruddiman, 1996; Thevenon et al., 2004). In Europe, due to the temporally limited archives (lake, peat, swamp), fire activity studies cover only the Holocene time period, and mainly focus on boreal and temperate regions (Bradshaw et al., 1997). They show local climatic control over fire regime at a centennial scale, the latter exerting in turn an influence on vegetation development at a regional scale. Human interference with fire regime has been also proposed (Bradshaw et al., 1997).

At present, one of the most fire sensitive regions in Europe is Iberia, related to severe anticyclonic summer drought contemporaneous with high ENSO activity (Carmona-Moreno et al., 2005). However, no studies so far show the impact of the sub-orbital climatic changes on fire regime even in a key region such as the Mediterranean. Here, we present for the first time a high resolution record of wildfire regime in southwestern Iberia covering the last climatic cycle from the microcharcoal analyses of the deep-sea core MD95-2042 off Lisbon. It has been shown that this core, retrieved not far from the mouth of the Tagus and Sado rivers, contains fine terrestrial material transported through canyons (e.g. pollen and microcharcoal) from both hydrological basins (Jouanneau et al., 1998; Naughton et al., 2007) and, therefore, should reveal fire regime changes of the southwestern Iberian Peninsula.

Charcoal particles result from incomplete combustion of woody debris during fires. They are defined as an intermediate stage between vegetation debris and soot (total combustion). The result of this combustion is a shrinkage of cellular walls, although the structure and microstructure of wood are weakly affected (Clark, 1984; Chabal et al., 1999). Microcharcoal comprises all charred particles with a size less than 250 microns. It is relatively resistant to chemical decomposition (classified as inertite) (Hart et al., 1994; Habib et al., 1994; Quénéa et al., 2006) and microbial decomposition is minimal (Verardo, 1997; Hockaday et al., 2006) especially if microcharcoal burial occurs in an environment with high sedimentation rate such as that of the marine core MD95-2042.

Microcharcoal is generally analysed with the point-count method (Clark, 1982) based on an optical counting of particles. While this method has been significantly improved, it is time-consuming and does not permit the analysis of the morphology and structure of the particles (Tolonen, 1986) and thus fails to verify whether counted particles are actually derived from wood combustion. For these reasons, we have developed a new method based on a coupled analysis involving automated image (Beaufort et al., 2003) and petrographic analyses.

The direct comparison between microcharcoal data and marine and terrestrial climatic records from the same core allows the pattern of variability of paleofire record to be interpreted in terms of paleoclimate. Previous studies of core MD95-2042 and of the twin core SU81-18 have shown the influence of millennial-scale climatic variability (Dansgaard-Oeschger oscillations (D-O) and Heinrich events) in the ocean and on land (Shackleton et al., 2000; Pailler and Bard, 2002; Cayre et al., 1999; Thouveny et al., 2000; Sánchez Goñi et al., 2000; 2002; Turon et al., 2003). Notably, the millennial-scale climatic variability of Greenland and the North Atlantic ocean is synchronous with an alternation of semi-desert vegetation during D-O stadials and the development of open Mediterranean forest during D-O interstadials in southwestern Iberia. Further, the correlation of this paleofire record with the NorthGRIP isotopic curve (NGRIP members, 2004) will shed light on the links between fire, vegetation and climate in the North Atlantic region.

2. Environmental setting

2.1 Present-day fires, vegetation and climate

The Tagus basin today is characterised by a Mediterranean climate, with mild winters and hot and dry summers with a high level of precipitation in its western part (oceanic conditions) (Peinardo Lorca and Martinez-Parras, 1987). This region is colonized by vegetation composed of Atlantic and Mediterranean floristic elements (Blanco Castro et al., 1997). Winds are characterised by a weak westerly component interrupted by a relatively strong northerly and northwesterly component in summer related to the intensification of Trade Portuguese Winds and causing the formation of a seasonal upwelling from July to September (Fiuza et al., 1982).

At present, fires in Portugal and Spain occur during the dry summer season (June to August) (Carmona-Moreno et al., 2005; Pérez et al., 2003; http://www.incendiosforestales.org/estadisticas.htm). Pereira et al., 2005 have reported rare and large wildfire events in Portuguese forests responsible for burning large areas of Portugal. These extreme fires are triggered by a long and intense dry period with an absence of precipitation in late spring and early summer and the occurrence of days of extreme synoptic situations characterised by south-easterly winds on the Iberian Peninsula (figure 1) and a strong anomalous advection of hot and dry air from Northern Africa via central Iberia.

2.2 Microcharcoal production and deposition

Vegetation fires produce different sizes of particles of which the smallest, classified as fine particles, are deposited far from the source (Patterson et al., 1987). Aeolian and fluvial processes are the main agent responsible for the transport of microcharcoal from the combustion site to the sedimentation basin where they are preserved. Large fires are characterised by a convection column which is made visible even at large distances from the fire by smoke particles (Viegas, 1998). Particle diameters in these smoke plumes vary in size between the submicron scale and several centimetres (Schaefer, 1976), as fragmentation by collision can occur in these convective currents (Walsh and Li, 1997). These microcharcoal particles remain in the atmosphere and are moved long distances (Clark, 1988), carried by low atmospheric winds (<10km) and deposited a few days or weeks after their formation (Palmer and Northcutt, 1975; Clark and Hussey, 1996).

In water, after a short period of bedload transport, charred fragments break down into relatively resistent, somewhat rounded pieces, and thereafter remain stable. They exhibit the same behaviour as fragments of highly vesiculated pumice, which initially floats and sinks as it becomes waterlogged (Nichols, 2000). The small charcoal particles may sink in a matter of hours, but the larger ones may float for months or years before becoming waterlogged and sinking. Whitlock and Millspaugh (1996) have reported charcoal introduced into deep lake sedimentary record within a few years after a fire event.

Griffin and Goldberg (1975) have shown that coarse particles, generally deposited close to the fire source, can be washed by rivers and transported close to the coasts. However, the study of Jouanneau et al. (1998) on sediment transport at the location of core MD95-2042, have shown that suspended fine material (including pollen and microcharcoal) from Tagus and Sado estuaries can be transported through canyons to the deep ocean. Microcharcoal sedimentation,

which may be comparable to pollen sedimentation behaviour, can be deposited in several weeks on the ocean floor as a part of the marine snow (Hoogmiestra et al., 1992; Chmura et al., 1999).
The relatively rapid aeolian and fluvial transport, in the order of days or weeks, suggests that no significant timelag exists between production and deposition of microcharcoal at the study site. Although charcoal may undergo a period of storage in fluvial, alluvial, estuarine and soil sediments, we do not anticipate a significant temporal bias at the basin scale. Masiello and Druffel (1998) have reported a timelag of thousand years in the Pacific and Southern oceans between Black carbon (fire proxy) and bulk sedimentary organic carbon, but hypothetise that the age difference would be even smaller closer to the continents. For our study area, evidence of no significant timelag between production and deposition is further supported by the similar AMS ^{14}C age obtained from level 290 cm of core MD99-2331 (42°09'00N, 09°40'90W) on the planktonic foraminifera specimen *Globigerina bulloides* (15 770 yr ±130, ECLIPSE) and on bulk organic matter including microcharcoal and pollen (15 010 yr ±130, eth n° 28540, Zurich, Hajdas, pers. com.).
Experimental studies (Heusser, 1985) have shown that cores located close to the river mouth, as in the case of core MD95-2042, recruite pollen mainly from the river basins. The Tagus basin, which crosses the southern and central part of Iberia from the east to the west, is the major contributor of fine material to our study area (Jouanneau et al., 1998) and therefore core MD95-2042 provides a reliable high resolution record of microcharcoal to explore fire variability of southwestern Iberia at a centennial to millennial scale.

3. Material and methods

3.1 Core location, sampling and chronostratigraphy

Deep-sea core MD95-2042 (37°45'N, 10°10'W; 3146 m water depth) (figure 1) was collected during the 1995 IMAGES cruise (Bassinot and Labeyrie, 1996) on the Iberian margin off Lisbon using the CALYPSO Kullenberg corer aboard the Marion Dufresne vessel. The site is located 140 km from the nearest coast line (figure 1) on a nearly flat continental rise with slope <2%, downstream of the Tagus, Setubal and Sado canyons and upstream of the Tagus abyssal plain (Thouveny et al., 2000). The sediments are mainly composed of clayey mud and carbonate (Cayre et al., 1999), with 20-40% of carbonate and <1% of organic matter (Pailler and Bard, 2002).

The age model of core MD95-2042 is based on 12 age control points from Shackleton et al. (2000-revised June 2002, 2003, 2004) and 16 AMS ^{14}C ages from Bard et al. (2004) and follows that of Sánchez Goñi (2006) (table 1).
Between 0 and 4.34 m (column 1) only two age control points are available, obtained by tuning the δ^{18}O of the planktonic foraminifera *Globigerina bulloides* to the δ^{18}O record of GISP2 ice record (Shackleton et al., 2000). For the core-depth interval 4.58-13.78 m we used 15 AMS ^{14}C ages (column 2) on *Globigerina bulloides* with a calendar time scale built by tuning the alkenone record of Iberian margin sediments with the δ^{18}O record of the GISP2 ice core (Bard et al., 2004). Similar AMS ^{14}C ages to those of Bard et al. (2004) were obtained by Shackleton et al. (2004) for this interval by the tuning of the δ^{18}O of the planktonic foraminifera from MD95-2042 with the GISP2 time-scale and with the GRIP SS09sea time scale, the latter being the most recent. The age obtained for level 15.48 m is the mean between those in column 2 and column 3 (table 1). From 15.80 to 19.92 m depth, seven age control points (column 3) were used, obtained by tuning the planktonic δ^{18}O with the GRIP SS09sea time scale (Shackleton et al., 2004). For the bottom part of the core (21.40-29.90 m), six age control points (column 1) were defined by Shackleton et al. (2003) using the correlation of

equivalent sea level stands identified in the $\delta^{18}O$ of benthic foraminifera of core MD95-2042 with those detected and dated by U/Th in coral terraces from the Pacific ocean.
For the purpose of this study, core MD95-2042 was sampled every 10 cm down to 2700 cm, and every 5 cm between 1300 and 1419 cm, this last interval corresponding to the time of the arrival of Anatomically Modern Humans (AMH) in this region, giving an average resolution of 500 years. Moreno (2000) described the perturbation of the magnetic fabric of this core over the first 15 m due to the elongation of the core during coring process resulting in a higher accumulation rate in the upper part of the core.

3.2 Microcharcoal analyses

3.2.1 Petrographic analysis

The erosion of organic-enriched sediments (including coals) from the sedimentary basins may be a source of non-burnt and oxidized particles, i.e vitrinite, which can appear black in transmitted light and be misidentified as burnt particles. A number of sites containing coal are reported in Spain and Portugal (Arche et al., 2004; Flores, 2002; Jiménez et al., 1999; Suárez-Ruiz and Jiménez, 2004; Suárez-Ruiz et al., 2006), but are however out of the hydrological basins of the Tagus and Sado rivers. Moreover, pollen analysis of core MD95-2042 revealed only a sporadic presence of putative prequaternary trilete spores (Sánchez Goñi M.F., pers. com.) which supports our inference of a negligible input of material reworked from geologically ancient deposits. In order to check whether black particles in this core derive from fire, we applied petrographic analysis to 29 levels representing maxima and minima of microcharcoal concentration, as revealed by automated image analysis.
Petrographic analysis in white reflected light has been performed on polished slides following a slightly modified preparation technique developed at the ISTO laboratory (Orléans University). Chemical treatments with HCl (10 mL 25%; 50% - followed by centrifugation); HF (25 mL 70%; centrifugation); HCl (10 mL 25%; centrifugation) were applied for demineralisation of about 0.2 g of dried sediment. We added 5 mL of 68% HNO_3 and 33% H_2O_2 leaving it for 24 hours. Despite a possible microcharcoal fragmentation during centrifugation, this step was crucial to remove all HF solution. We applied a dilution of 0.1 before filtering on cellulose acetate filters containing nitrocellulose (diameter 25 mm, porosity 0.45 µm). Filters were dissolved on Plexiglas (5 mm thickness) with ethyl acetate. We also verified that centrifugation did not substantially remove particles. After hand polishing, each particle appearing black in transmitted light was analysed in white reflected light using a Leica DM6000M microscope at x500 magnification using a squared grid eye piece. Four categories of particles were identified based on criteria traditionally used for petrographic studies and described in the work by Noël (2001):
a) burnt particles (BP), showing visible plant structures characterised by thin cell walls and empty cellular cavities with sharp edges, a low to relatively high reflectance, sometimes with the presence of devolatilisation vacuoles.
b) homogeneous particles (without visible structures) identified by their low to medium reflectance in grey (HGP).
c) homogeneous particles with a high reflectance (in white) (HWP).
d) particles which cannot be classified as either BP and HGP because the position of these particles were not under focus (especially on the edge of the filter) (Unclassified Particles UcP).

3.2.2 Image analysis for microcharcoal counting

288 samples were analysed using automated image analysis. Microcharcoal preparation technique was adapted from the work of Beaufort et al. (2003). About 0.2 g of dried sediment

was directly submitted to chemical treatment without centrifugation because fragmentation occurs during this procedure (Rhodes, 1998) and without rinsing between acids to reduce the loss of microcharcoal. We applied combined acid-base oxidation for removing all organic contamination from the samples (Bird and Gröcke, 1997). This treatment was performed over 24 hours on the dried sediment and involved the addition of 5 mL of 37% HCl for carbonate hydrolysis; 5 mL of 68% HNO_3 for pyrites and humic material removing (Clark, 1984 ; Winkler, 1985); 10 mL of 33% H_2O_2 for labil or less refractory organic matter and for kerogen oxidation (Wolbach and Anders, 1989). This procedure does not blacken unburned plant materials (Clark, 1984). Microcharcoal extracted by this chemical treatment is most likely equivalent to organic resistant elemental carbon (OREC), a reliable combustion/pyrolysis product (Bird and Gröcke, 1997). For this study we used the term microcharcoal for small carbonised particles produced during fires, following Jones et al. (1997).

Dilution of 0.1 was then applied to this residue and this suspension was filtered onto a cellulose acetate membrane of 0.45 µm porosity and 47 mm in diameter. A portion of the membrane was mounted onto a slide with Canadian balsam.

The slides were scanned with an automated Leica DMRBE microscope in transmitted light at a x500 magnification. In order to have a good statistical representation of each sample, 100 view-fields (100 images) of 1.7 mm² were "grabbed" in black and white with a standard 756x582 pixels digitising camera (Beaufort et al., 2003). Images were analysed using an adapted program of image analysis written in Scion Image software, allowing microcharcoal recognition given by a threshold in grey level. This threshold was determined by visual identification of microcharcoal following criteria from Boulter (1994) i.e. black debris, opaque, angular with sharp edges.

Several measured variables were obtained for each particle of microcharcoal: surface area, length and width of the ellipsoid of the particle, as well as the number of microcharcoal particles for each level. All particles less than 15 pixels in area were removed and the variable perimeter was not used because of the large range of calculated values which may be obtained for the same object by this software depending on the orientation of the particle in the plane (Francus P., pers. com.).

From these measurements we calculated for each sample (for the calculation method see Beaufort et al., 2003):

a) the concentration of microcharcoal (Char number): number of microcharcoal particles per gram ($nb.g^{-1}$).

b) the concentration of microcharcoal surface area (Char surface), which is the sum of all surface areas of microcharcoal in one sample per gram ($\mu m^2.g^{-1}$). This is to avoid the overrepresentation of microcharcoal concentration as the result of potential fragmentation during production (Théry-Parisot, 1998) or transport.

c) the morphology of the microcharcoal based on the elongation degree (length divided by width). Umbanhowar and McGrath (1998) have demonstrated experimentally that the elongation degree can be a good indicator for the determination of the burnt vegetation type. Charcoal fragmentation occurs along axes derived from the anatomical structure of plant species and the elongation degree is preserved even when the particle is broken (Clark, 1984; Umbanhowar and McGrath, 1998). A near-square morphology (ratio of one) characterises particles produced during the burning of deciduous forest. A high ratio (elongate particles) identifies the burning of herbaceous vegetation.

d) the mean microcharcoal surface area per sample (Char surface divided by Char number, µm², Char mean surface) in order to estimate the intensity of fires (fires of high or low temperature). Particles produced during a fire of low or medium temperature are generally bigger than those produced during an intense fire (Théry-Parisot, 1998; Komarek et al., 1973).

e) the microcharcoal fluxes ($\mu m^2.cm^{-1}.yr^{-1}$).
McDonald et al. (1991) have experimentally validated the use of image analysis showing a good correlation between visual counts using Clark's method and image analysis. However, image analysis appears to underestimate the surface area of microcharcoal due to the effect of light "bleeding" onto the edges of fragments, lowering the optical density of the periphery of microcharcoal particles.

4. Results

4.1 Petrographic analysis

Petrographic analysis has revealed the presence of burnt particles in all the samples (figure 2). Figure 2a shows that the microcharcoal number concentration values were quite high all along the profile (500 to 100 000 $nb.g^{-1}$). The BP represent up to 20% of the total number of counted particles. The HGP dominate the total number of particles (51% on average) with concentration values ranging between 25 000 to 170 000 $nb.g^{-1}$. The HWP are quite rare (figures 2a and 2b).
The surface area concentration values of BP were globally lower (from 0.07 to 3.46 $mm^2.g^{-1}$) than those of HGP (figure 2b), but contribute as much as 5 to 40% (16% on average) of the total surface area which is not negligible for the marine domain. Indeed, Noël (2001) has reported surface area of BP representing 40% of the total surface area for ancient soil regularly burnt. The HGP contribute to the high values (54.5% on average) of the total surface area. UcP number concentration values represent in average the 36% of the total particle number (from 15 000 to 103 000) and only the 27% of the total surface area.
The HGP could be classified as vitrinite particles following identification with petrographic criteria, i.e. homogeneous gelified particles with a grey reflectance. However, Noël (2001) has reported carbonised particles showing the same characteristics as vitrinite probably due to a less severe combustion of the plant debris. Our own observations carried out on charcoals of pine and oak, the most important trees found in the Mediterranean forest, collected from domestic fireplace and on moorland surface samples, confirmed these features. As a matter of fact, microscopic observations (in incident light) revealed two distinctive morphologies within the same particle: a characteristic burnt particle structure (with highly reflected thin cell walls and empty cavities) progressively replaced by a homogeneous and gelified structure. Broken particles can therefore lead, because of the fragility of the cellular cavities, to one part with visible structures and the other with a homogeneous aspect. These two types of particles will be counted apart, one in BP and the other one in HGP, although they come from the same particle of ash. Some apparently homogeneous particles have also been observed at a x625 magnification and have revealed the presence of small remnants of cellular cavities not obviously visible at a x500 magnification. Moreover, the surface area distribution (figure 3) of these particles showed that BP are larger than HGP and HWP: the median, the quartiles at 25% and 75% and the maximum are higher for BP. From these observations we believe that HGP, which form the majority of particles, are debris from burnt particles, for which the original structures were broken during transport by air or water. The variation of reflectance for the different categories can be explained by different temperatures of combustion.

4.2 Image analysis: wildfire regime in southwestern Iberia

Cayre et al. (1999) have shown that core MD95-2042 records the glacial Marine Isotopic Stages (MIS) 6, 4, 3, 2 and the interglacials MIS 5 and 1 (figure 4), i.e. the last climatic cycle. A visual inspection of the record reveals that Char surface and Char number covary and have undergone rapid fluctuations from low to high concentration values. Char surface varies between 89.7 and 652.5 x10^3 µm².g^{-1} and Char number between 23.5 and 181.4 x10^3 nb.g^{-1}. The alternation of minima and maxima of Char surface and Char number are punctuated by episodes characterised by very low concentration values indicated by the grey bands. These intervals correspond to ages of 135.3-126.5 ka (transition between MIS 6 and 5); 86.5-82.7 ka (cold period of Mélisey II, end of OIS5 (Sánchez Goñi et al. 1999)); 12.9-11.8 ka (Younger Dryas identified by Lézine and Denèfle (1997) in the twin core SU81-18); 62.8-59.6 ka; 48.8-46.4 ka; 40-38.5 ka; 30.7-28.9 ka; 25.2-23.3 ka; 17.6-15 ka during the Last Glacial Period. The Char mean surface record reflects a similar variability pattern, the biggest particles, indicating rapid fires of low intensity, characterising the episodes of very low values of Char surface and Char number.

The organic carbon (OC) content has been interpreted as a marine productivity proxy since $CaCO_3$ and OC contents were positively correlated (Pailler and Bard, 2002). However, a part of the OC content may be of terrestrial origin, including organic matter, charcoal and coal (see section 3.2.1 for the latter). The weak correlation ($r=0.22$; $R^2=0.05$) between the OC content and Char surface (figure 5) supports our conclusions from petrographic analysis that microcharcoal were uniquely isolated during the extraction procedure and are a proxy for biomass burning (Thevenon et al., 2004).

The similar shape of the microcharcoal influx curve and the mean sedimentation rate curve (figure 4) suggests that major changes in the accumulation rate (Moreno, 2000) strongly affected the calculated influx, in particular, for the upper 15 meters of the core. However, if we do not consider absolute values but focus on the variability of influx trends, the influx maxima (and minima) correspond to maxima (and minima) of concentration. The fact that the same variability pattern is reported for influx, concentration and mean surface area of microcharcoal (free from changes in sedimentation rate), suggests that the microcharcoal concentration record represents the most reliable picture of fire regime variability in this region. For this reason, we will use only concentration curves for the following discussion. Core MD95-2042 thus identifies fluctuations in the fire regime, with an alternation of periods with high frequency and intensity of fire characterised by high values of Char surface/number, and periods with relatively weak frequency and intensity of fire identified by low value of Char surface/number.

To extract the significant periodicities contained in the microcharcoal signal, spectral analysis was performed using different algorithms (Blackman Tuckey, Maximum Entropy, Thomson and Singular spectrum analysis) on Char surface and Char mean surface. The results show the presence of a period close to 23 ka (22-23 ka) (precession) as the main cycle of temporal variability in the charcoal record. Spectral analysis conducted on Char mean surface reveals periods of 8.5 ka and 6.1 ka, 5.4-5.7 ka, 2.8-2.9 ka and a less pronounced period of 4.6-4.7 ka (figure 6b).

The Char surface also displays millennial periods, one around 8.8 ka particularly between 75 000 and 105 000 years, 6.3 ka, 4.8 ka (4.5-4.9) and 2.9 ka (2.8-3) (figure 6a). Our fire record is mainly imprinted by the precession, and periods which are multiples of the millennial variability (c. 1.5 ka) of D-O (8.5, 6.1, 4.8, 2.9 ka). The 8.5 ka and 6.1 ka are probably related to the precessional harmonic p3 (p3=7.4 ka) reported in the equatorial Atlantic by Pokras and Mix (1987) on the abundance of the freshwater diatom *Melosira* spp. and the 5.4-5.7 ka to p4 (p4= 5.5 ka).

No substantial contribution of human fire activity is recorded over the studied time period and, in particular, at the moment of the arrival of Anatomically Modern Humans in this region at around 33 000 years BP (d'Errico and Sánchez Goñi, 2003). This result differs from that of Beaufort et al. (2003), who inferred a substantial increase in the amplitude of the microcharcoal concentration at the time of the arrival of Anatomically Modern Humans in

Indonesia (50 ka). We can propose as a working hypothesis that AMH arriving in this region did not use fire as a tool for modifying their environment as has been proposed for the first inhabitants of Indonesia. An alternative hypothesis could be that their number was low and any human-fire signal was masked by stronger variability in fire regime driven by climate through changes in fuel availability.

5. Wildfires and climatic variability in southwestern Iberia for the Last Glacial Period (70-14 ka)

For the purpose of this study, we have focused on the Last Glacial Period to explore how fire regime was affected by the sub-orbital climatic changes.

Previous studies (figure 7) carried out on different proxies from core MD95-2042 have revealed that the southwestern Iberian margin preserves a strong imprint of the D-O oscillations and Heinrich events. D-O stadials are evidenced in core MD95-2042 by a decrease of sea surface temperature (SST) (Pailler and Bard, 2002, not shown), and heavy planktonic oxygen isotopic values from *Globigerina bulloides* (Shackleton et al., 2000) while D-O interstadials are characterised by warmer SST and light planktonic $\delta^{18}O$ values. Extreme cooling episodes corresponding to Heinrich events were characterised by the occurence of ice rafted debris (IRD) (Cayre et al., 1999; Gendreau, 1999; Sánchez Goñi et al., 2002; Sánchez Goñi, 2006), the strongest increase in planktonic $\delta^{18}O$ values, the development of the polar foraminifera *Neogloboquadrina pachyderma* (s.) left coiling and peaks in magnetic susceptibility (Thouveny et al., 2000, not shown). A visual inspection of the record reveals that the fire regime covaries with the planktonic $\delta^{18}O$ value (figure 7), i.e. weak fire activity is associated with heavy $\delta^{18}O$ anomaly during stadials whereas strong fire activity with light $\delta^{18}O$ anomaly during interstadials. The lowest fire activity (62.8-59.6 ka; 48.8-46.4 ka; 40-38.5 ka; 30.7-28.9 ka; 25.2-23.3 ka; 17.6-15 ka) corresponds to Heinrich events. Fire regime also parallels the variability of atmospheric temperature shifts over Greenland indicated by $\delta^{18}O$ from NGRIP. High and low fire regimes correspond to D-O interstadial and stadial periods, respectively. However, a contrasting trend is observed: while the increase of $\delta^{18}O$ is abrupt in Greenland at the onset of interstadials, fire activity increases gradually.

Today, fire activity is enhanced during severe summer drought associated with persistent anticylonic conditions (Carmona-Moreno et al., 2005; Pereira et al., 2005). During the Last Glacial Period, interstadials are characterised by a high fire activity whereas stadials and Heinrich events (HE) are associated with low fire activity. We could hypothesize therefore using the classical interpretation for the Holocene (Millspaugh et al., 2000; Whitlock, 2001; Whitlock and Millspaugh 1996; Carcaillet et al., 2001; Bradshaw et al., 1997) that periods of high fire regime indicate episodes of severe drought. However, our record shows the opposite. Indeed, climatic conditions of the southwestern Iberia determined by pollen analysis (Sánchez Goñi et al., 1999, 2002; Turon et al., 2003) show that the D-O climatic variability has strongly affected this region, producing an alternation of semi-desert vegetation mainly composed of *Artemisia*, Chenopodiaceae, *Ephedra* during stadials indicating dry atmospheric conditions, and open Mediterranean forest (mainly evergreen and deciduous oak) during interstadials identifying wetter and warmer periods (figure 7).

The dry climatic conditions in southwestern Iberia during stadials would be expected to favour fire ignition and, therefore, high fire activity should be found in our record, which is not the case. This apparent mismatch between low fire regime and dry climatic conditions on the continent could be explained either by a different transport of microcharcoal during stadials (fluvial or wind direction and intensity shifts) or by a low biomass availability (Black et al., 2006).

The North Atlantic Oscillation (NAO) is a characteristic feature of the wintertime temperature and precipitation patterns across the North Atlantic region. During the positive phase of NAO, the Mediterranean region experiences major drought because of the shift of the westerlies northwards, whereas a negative phase brings humidity to this region due to weaker westerly winds being guided southwards to mid-latitudes. For the Last Glacial Period, a prevailing phase of a positive NAO-like situation during stadials has been already suggested for the Iberian Peninsula (Sánchez Goñi et al., 2002; Moreno et al., 2002a; Combourieu-Nebout et al., 2002). Nowadays, years with a high NAO index lead to the development of orographic low pressures over North Africa (Moulin et al., 1997; Hamonou, 2000). Based on these works and that of Trigo et al. (2002), we suggest that this atmospheric pattern creates a prevailing south-easterly wind component over the Iberian Peninsula during stadial events similar to that observed today and described by Pereira et al. (2005) (figure 1) during episodes of large wildfires in Portugal. We assume that this rare atmospheric pattern (Pereira et al., 2005), leading to southeasterly winds, was enhanced and more frequent during D-O stadials because of the prevalence of a positive NAO-like situation (figure 8). Moreover, the low microcharcoal concentration during stadials is contemporaneous with haematite-rich dust layers in core MD95-2042 which have been related to aeolian transport by strong winds blowing from the Iberian Peninsula to the Atlantic ocean (Moreno et al., 2002b). Finally, the Portuguese margin is characterised at present by an upwelling during summer related to winds of North or Northwest component (Trigo and DaCamara, 2000). The increase of percentages of the coccolith species *Florisphaera profunda* (Bouldoire et al., 1996) and the decrease of productivity (Pailler and Bard, 2002) revealed by the same core provide evidence for a breakdown in upwelling activity during D-O stadials and particularly Heinrich events on the Portuguese margin. This also indicates a synoptic situation with a dominant southeasterly wind.

The presence of Saharan dust found in the Alboran Sea (Mediterranean sea) during stadials (Moreno et al., 2005) is furthermore in agreement with this situation.

The prevailing south-easterly wind conditions would have favoured the transport of microcharcoal to our study area during stadials. Wind direction does not appear therefore to have been a limiting agent for microcharcoal transport during these cold periods.

During D-O interstadial periods, some authors have reported an increase in frequency of a negative NAO-like situation (Sánchez Goñi et al., 2002; Moreno et al., 2002a; Combourieu-Nebout et al., 2002). From Trigo et al. (2002), this atmospheric situation will induce winds coming from the southwest to the Portuguese coast. The microcharcoal could be therefore displaced inland of the Iberian Peninsula because of this dominant wind pattern. However, a high concentration of microcharcoal is found in our record off western Iberia during interstadials. During these periods, Combourieu-Nebout et al. (2002) suggest that the climatic conditions are close to present-day ones, and therefore fires can develop in summer under atmospheric conditions comparable to the rare situation described by Pereira et al. (2005). In winter, enhanced precipitation increasing river flow (Trigo et al., 2004) can wash microcharcoal from the land surface and the atmosphere transporting microcharcoal to the Atlantic ocean.

Microcharcoal concentration variability in core MD95-2042 does not appear therefore to be controlled by the efficiency of aeolian and fluvial transports. The synchroneity between forest cover and microcharcoal concentration shifts suggest a strong relationship between microcharcoal production and fuel availability (figure 7).

We infer that during Heinrich events and D-O stadials, fires propagate in the semi-desert vegetation, characterised by *Artemisia*, Chenopodiaceae and *Ephedra*, producing a low

quantity of microcharcoal and bigger and elongate particles (figure 7). Indeed, studies on different type of steppic formation in Iberia have already shown that the semi-desert vegetation has the lowest biomass (Gauquelin et al., 1998). In grassland ecosystems, residence times of fire and fire intensities are typically low (Johansen et al., 2001) and can produce bigger particles (Komarek et al., 1973; Théry-Parisot, 1998; Umbanhowar and McGrath, 1998). The highest values of the mean elongation degree of our particles during stadials, except for Greenland Stadial (GS) 8 and GS12 (figure 7) also confirm that fires propagate in herbaceous vegetation (Umbanhowar and McGrath, 1998).
During interstadials, the smaller, wider particles indicate fires of high intensity which propagate slowly. This kind of fire, occuring today in temperate forest and heathland, is consistent with vegetation reconstruction from pollen preserved in the same levels. Moreover, the study of modern fires (Blackford, 2000) has shown very high charcoal concentrations and small charred particles as the result of intense fires in heathland (Théry-Parisot, 1998).We infer therefore that during interstadials, fires propagated in the Mediterranean forest and heathland, the high input of microcharcoal being derived from the increased availability of woody fuel.

The variability of the fire regime in southwestern Iberia thus likely reflects changes in biomass availability accompanying the millennial-scale climatic variability of the D-O oscillations, leading to the apparent paradox that periods characterised in western Mediterranean region by major drought (D-O stadials), are associated with low fire activity.

6. Conclusions

Microcharcoal analyses of core MD95-2042 document in detail the fire history of southwestern Iberia during the Last Glacial Period. The combined methods that we applied in this work using petrographic and automated image analyses have shown that microcharcoal in marine core can be used as a proxy to detect fire regime variation at a regional scale. The variability of fire regime in southwestern Iberia during the Last Glacial Period follows the millennial scale oscillations of D-O and is related to shifts in biomass availability. There is no apparent relationship between the arrival of Anatomically Modern Humans and fire activity in the Iberian Peninsula. During periods of drought associated with D-O stadials and especially during Heinrich events, the low fire activity registered in southwestern Iberia is explained by the weak biomass of the semi-desert vegetation, whereas the warmer and wetter periods of interstadials are characterised by higher fire activity related to the increasing fuel availability due to the spread of the Mediterranean forest and heathland. The fire regime seems therefore indirectly controlled by climatic variability. The Iberian Peninsula is today one of the most important region for fire activity in Europe and substantially contributes to the release of atmospheric greenhouse gases. In the past, shifts in fire regime emissions between D-O interstadial and stadial periods could have contributed to the atmospheric greenhouse gas concentration variations identified for the last climatic cycle in Greenland and Antarctic ice cores.

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10. Figure legends, tables, figures, schemes

Figure 1. Map of the Iberian Peninsula showing core location of MD95-2042 (filled circle). Black arrows represent the wind direction anomaly during large summer wildfires after Pereira et al. (2005).

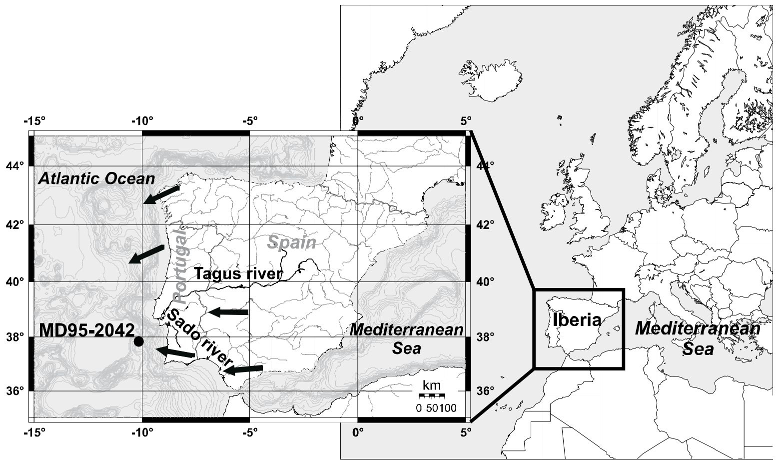

Figure 2. (a) Histograms showing number concentration and associated percentages for Burnt Particle (BP), Homogeneous Grey Particle (HGP), Homogeneous White Particle (HWP) and Unclassified Particle (UcP) by samples from petrographic analysis of core MD95-2042. Logarithmic scale for BP number concentration.
(b) Idem but for surface area concentration.

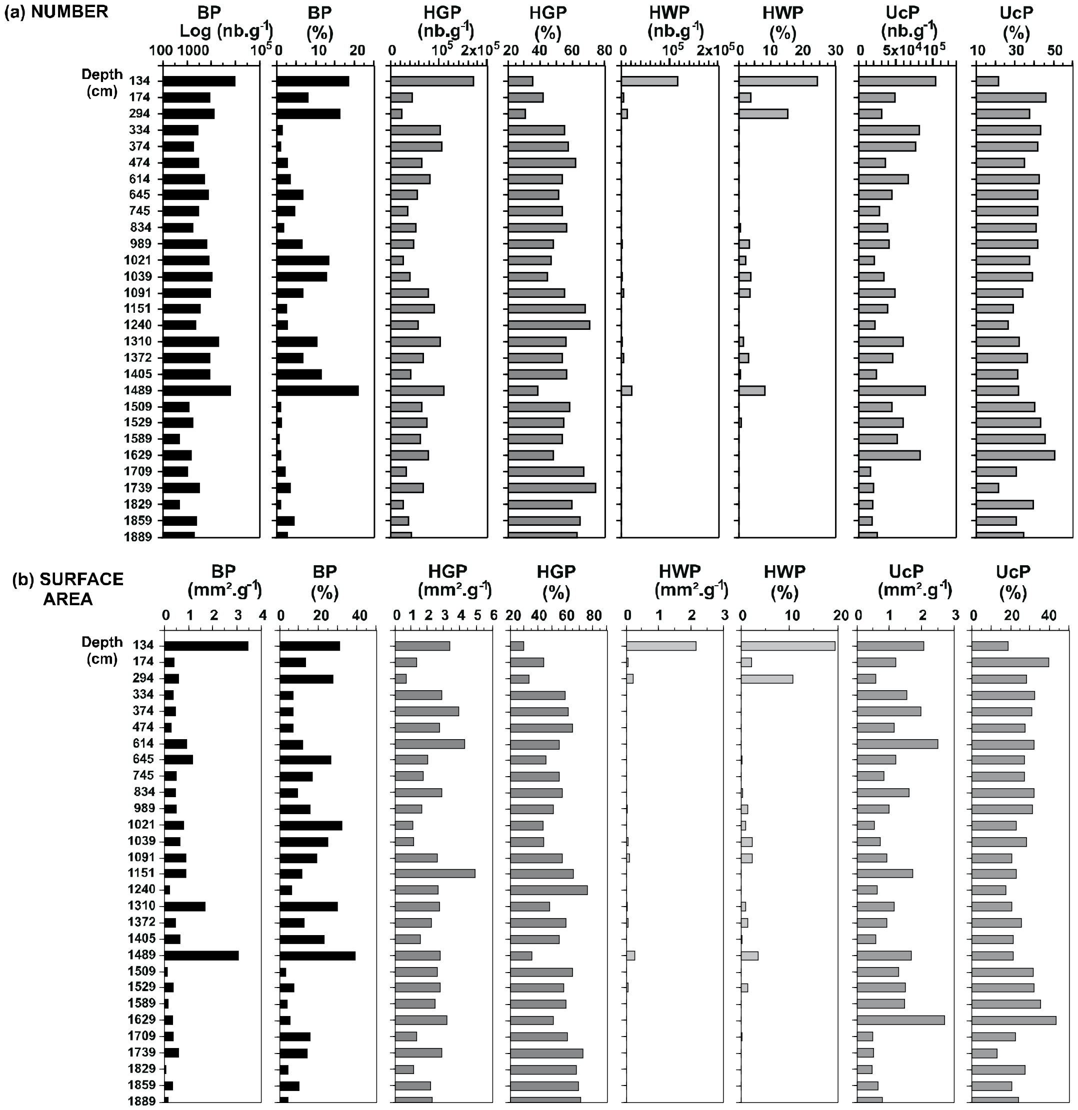

Figure 3. Box plots showing the surface area distribution (logarithmic scale) for Burnt Particle (BP), Homogeneous Grey Particle (HGP), Homogeneous White Particle (HWP) and Unclassified Particle (UcP) from petrographic analysis of core MD95-2042. Black horizontal lines represent the median. Vertical lines represent the minimum and maximum of surface area. The lower (and upper) box limits represent the first (and third) quartile. n is the number of particles.

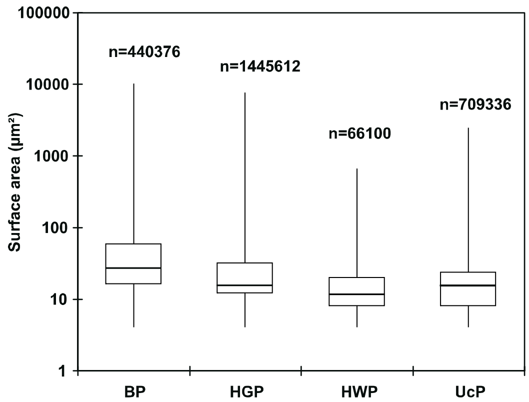

Figure 4. Results from microcharcoal image analysis (this work) and comparison with the organic carbon content (OC) from Pailler and Bard (2002) of core MD95-2042. All records are plotted versus depth. a) the microcharcoal surface area concentration curve (Char surface), b) the microcharcoal number concentration curve (Char number), c) the microcharcoal mean surface area curve (Char mean surface), d) the mean sedimentation rate obtained from the age model of this study, e) the microcharcoal influx curve, f) the percentage curve of organic carbon content (OC) from Pailler and Bard (2002). Grey intervals indicate the lowest values of microcharcoal concentration. Limits of marine isotopic stage (MIS) are from Cayre et al. (1999). The Younger Dryas (YD) and Mélisey II (MeII) chronostratigraphic intervals are indicated.

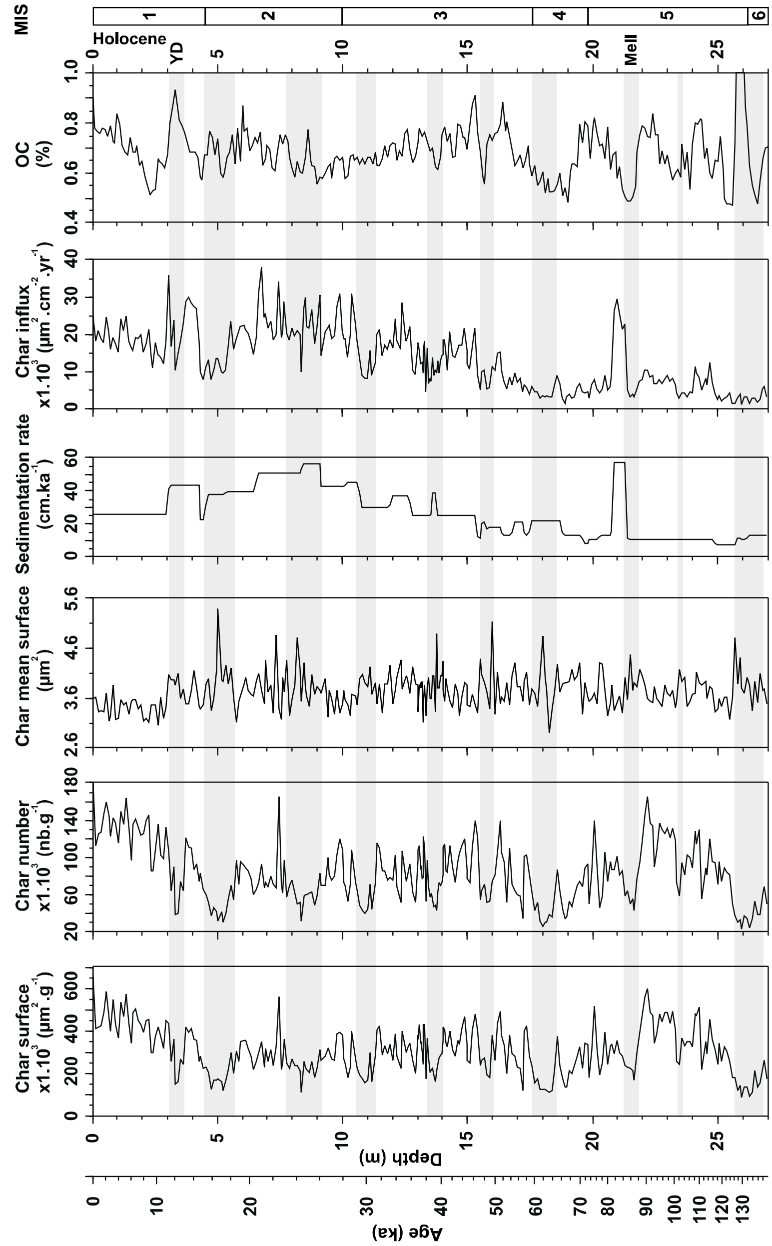

Figure 5. Correlation between microcharcoal concentration (this work) and organic carbon content (Pailler and Bard, 2002) of core MD95-2042.

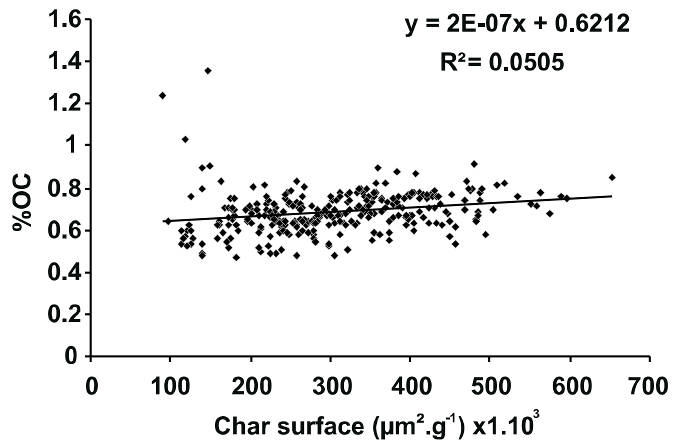

Figure 6. Amplitude of the wavelet transform of the Char surface and Char mean surface time series. The X-axis is the time scale (ka) and the Y-axis corresponds to periods (ka). The color reflects relative changes in the amplitudes, i.e. darker color represents larger amplitudes, while the lighter one indicates weaker amplitude. This figure shows the frequency evolution of the fire regime over the last 140 000 ka.

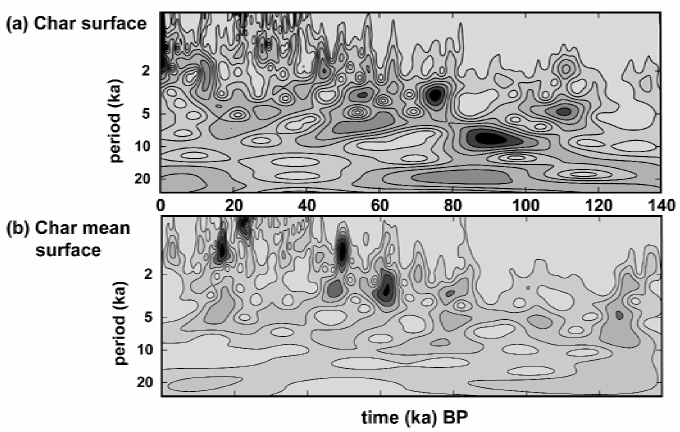

Figure 7. Comparison between microcharcoal trends and climatic proxies of core MD95-2042. All records are plotted versus age. From left to right: a) the NorthGRIP isotopic curve b) the planktic isotopic curve reflecting sea surface temperature and salinity changes (Shackleton et al., 2000), c) the concentration curve of the Ice Rafted Debris (IRD), d) the percentage curve of the polar foraminifera *Neogloboquadrina pachyderma* l.c. (left coiling), e) the surface area concentration curve of microcharcoal (Char surface), f) the elongation degree of microcharcoal, g) pollen percentage curve of the temperate forest including Mediterranean plants, h) pollen percentage curve of the semi-desert vegetation (*Artemisia*, Chenopodiaceae, *Ephedra*), i) pollen percentage curve of Ericaceae (heather). The pollen data are from Sánchez-Goñi et al. (2000) and Sánchez-Goñi (2006) except for the interval 14,000 and 25,000 years where they come from the twin core SU81-18 (Lézine and Denèfle, 1997). Grey bands indicate Heinrich events and other D-O stadials. The age model (calendar age) for core MD95-2042 derives from Shackleton et al. (2000-revised June 2002, 2003, 2004) and Bard et al. (2004). GIS n° indicates Greenland interstadial number.

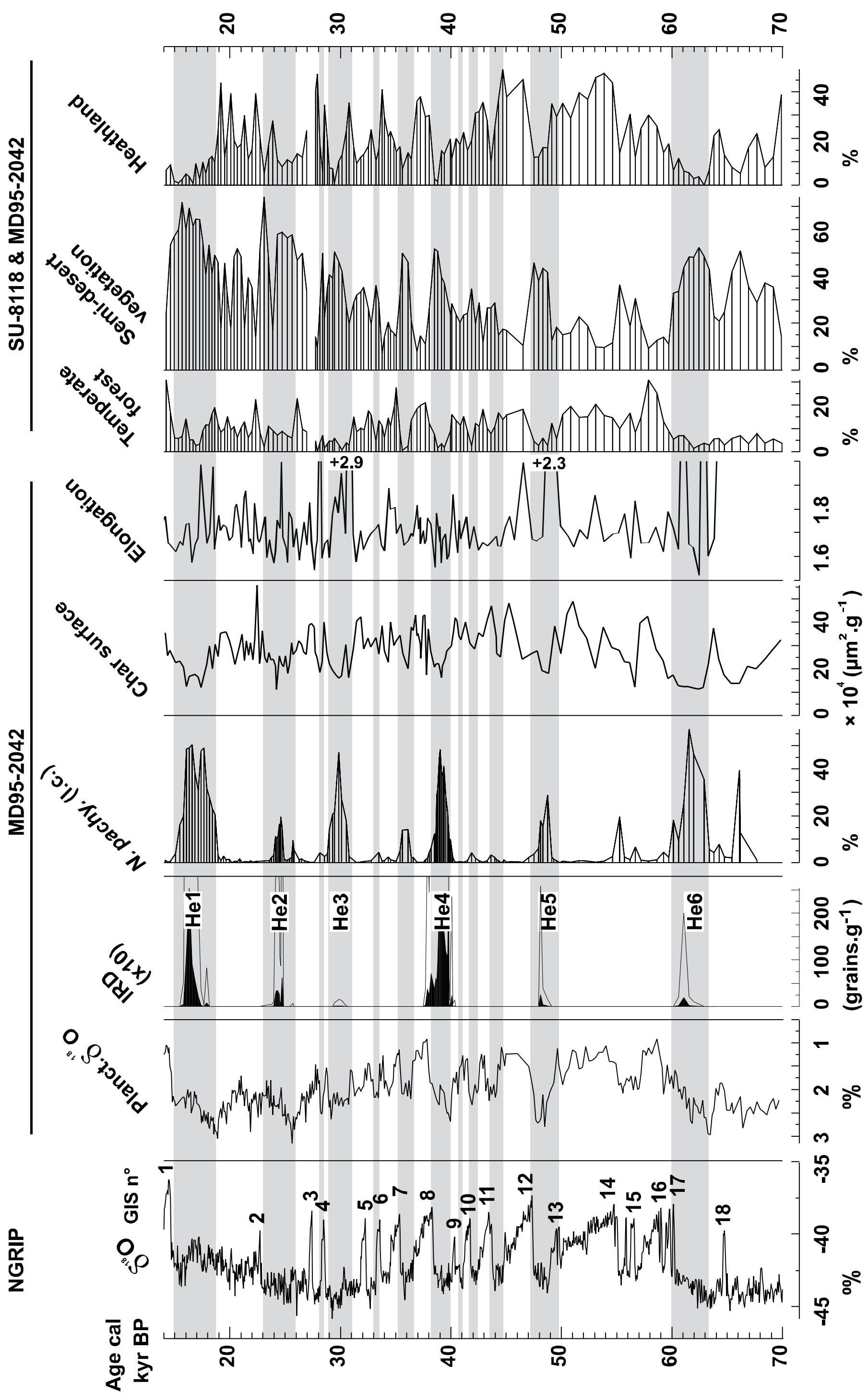

Figure 8. Synoptic scheme illustrating the location of high and low pressures (HP and LP) and main wind directions over the eastern North Atlantic region for a positive North Atlantic Oscillation situation, according to Moulin et al. (1997) and Hamonou (2000). Location of marine cores: 1. MD95-2042, 2. MD95-2043. D-O stadials and Heinrich events would be the result of prevailing positive NAO-like situation blocking the upwelling off Portugal (Moreno et al., 2002b; Bouldoire et al., 1996; Pailler and Bard, 2002), and increasing Saharan dust input in the Alboran sea (Moreno et al., 2005).

Dansgaard-Oeschger stadials and Heinrich events

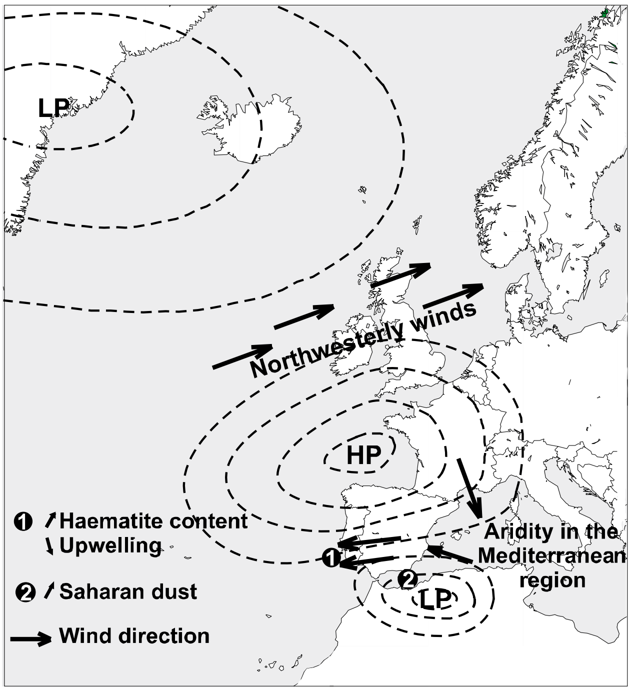

Table 1. Control points for the age model of core MD95-2042. Column 1: Shackleton et al. (2000-revised June 2002, 2003). Column 2: Bard et al. (2004). Column 3: Shackleton et al. (2004), Age GRIP SS09sea.

Depth MD95-2042 (m)	Age (ka) (this study)	Events	Age (ka) 1.	Age (ka) 2.	Age (ka) 3.
3.05	11.63	base Holocene	11.63		
4.34	14.59	base Bölling	14.59		
4.58	15.635	H1/GIS1		15.635	
4.59	15.659	H1/GIS1		15.659	
5.38	17.743	H1		17.743	
5.39	17.77	H1		17.77	
6.58	20.766	LGM		20.766	
8.41	24.364	H2/GIS2		24.364	
9.18	25.734	H2		25.734	
9.21	25.808	H2		25.808	
10.19	28.096	H3/GIS4		28.096	
10.78	29.395	H3		29.395	
10.79	29.424	H3		29.424	
11.99	33.433	end of GIS6		33.433	
12.79	35.578	GIS7		35.578	
13.61	38.877	H4		38.877	
13.78	39.318	H4		39.318	
15.48	46.019	H5/GIS12		45.268	46.77
15.8	48.328	H5			48.328
16.44	51.9	GIS14			51.9
16.65	53.52	GIS15			53.52
17.31	57.23	GIS17			57.23
17.8	60.146	H6/GIS17			60.146
18.6	63.812	base H6			63.812
19.92	74	GIS20			74
21.4	82.9	MIS 5a	82.9		
24.92	116	Upper limit of MIS5e "plateau"	116		
25.8	128	Beginning of MIS5e "plateau"	128		
25.94	129.1	Interglacial high stand	129.1		
26.26	132	MIS6/5	132		
29.9	160	base MIS6	160		