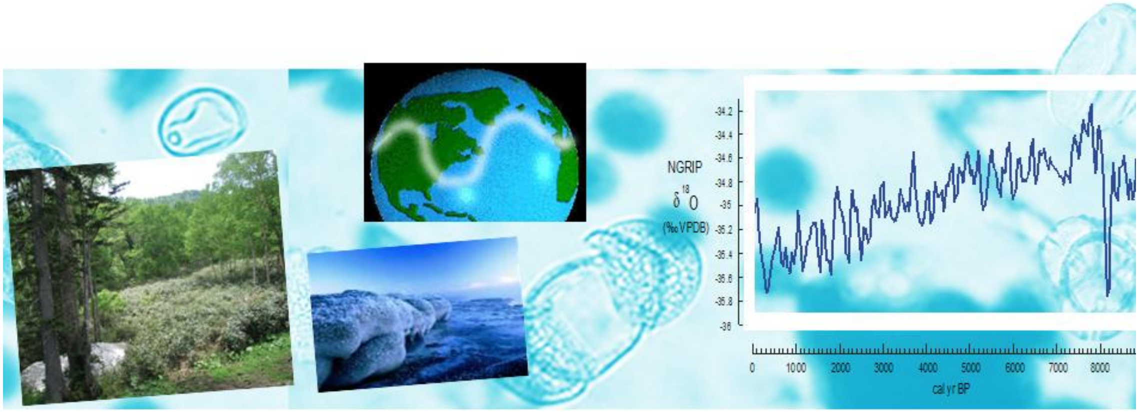
Universidade de Lisboa
Faculdade de Ciências
Departamento de Geologia

Vegetation response to Holocene climate variability in south-western Europe



Dulce da Silva Oliveira

Mestrado em Ciências do Mar

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Vegetation response to Holocene climate variability in south-western Europe

Dissertação orientada pelo Professor Doutor Ricardo Machado Trigo e co-orientada pela Doutora Ana Filipa Naughton Henriquez Andrez

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To my great grandfather Joaquim

Somewhere, something incredible is waiting to be known.

Carl Sagan

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SUMMARY

Understanding past climate variability, especially abrupt climate events, is essential for predicting future climate, as they may provide crucial information about the climate system's sensitivity to perturbations. Accordingly, this research is focused on documenting the vegetation response to the natural evolution of the current interglacial period, the Holocene, and on evaluating the anthropogenic contribution to it. Also, we intend to identify the nature, timing and causes of Holocene climate variability at orbital and suborbital time scales in a key region of the North Atlantic region.

The present study reveals the vegetation and climate changes in southwestern France and northern Spain for the last ca. 9000 cal. yr BP in a well dated shelf core, KS05-10, retrieved in the southwestern margin of the Bay of Biscay (Basque country). The continuous high resolution pollen record shows orbital and suborbital climate fluctuations contemporaneous with those noticed for the North Atlantic region, Greenland and Europe. The gradual decline of pine and oak trees and the general increase of herbaceous plants, reflecting a gradual cooling between 9000 and 1000 yr cal. BP, follows the cooling in Greenland as well as the decrease of mid-latitude summer insolation. The gradual replacement of the oak forest by beech also reveal the reduction of seasonality, probably triggered by the gradual increase of the precession, and the increase of moisture conditions in mid- to late Holocene.

Superimposed on the orbitally induced long-term cooling, KS05 10 pollen record detects an abrupt millennial scale climatic event between 8.3 and 8.1 ka in the southwestern Bay of Biscay, which is related to the well-known 8.2 ka event. The vegetation changes (reduction of temperate and humid trees, particularly *Corylus*, increase of ubiquist plants, principally Cyperaceae, and the presence of *Carpinus*) point to a cold and wet episode. The relatively cold conditions were probably the result of the weakening of the Meridional Overturnig Circulation triggered by the final catastrophic drainage of the Laurentide Lakes and consequent input of freshwater in the North Atlantic region. However this mechanism can not explain the wet conditions detected in the KS05 10 pollen record. These wet conditions could probably be the result of the influence of the Atlantic Westerly Jet stream and prevalence of strong zonal flow and frequent low pressure systems (associated with less blocking events located in the

northern Iberian Peninsula and southwestern France). The blockage of sunlight by clouds, which is associated to high precipitation, may be responsible for the particular decline of *Corylus* (light-demanding tree) during this climatic downturn event.

Small-amplitude millennial-scale cooling events after the 8.2 ka event and until the late Holocene may be reflected in the oscillations of the hazel trees. Spectral analysis of *Corylus* percentages shows a climatic cyclicity of ~500yr from 9 to 3 ka, comparable with those recognized in the North Atlantic region and Greenland ice cores, suggesting common climate forcing mechanisms such as changes in solar activity and perturbation of the North Atlantic circulation. The impact of human activity on vegetation over the last 1000 years is superimposed on the climatic natural changes.

RESUMO

O aquecimento global é na atualidade inequívoco, sendo evidentes o aumento das temperaturas médias do ar e do oceano à escala global, o degelo de neve e gelo e o aumento dos eventos meteorológicos extremos tais como: secas, cheias, ondas de calor, vagas de frio e furacões. Dada a gravidade das consequências que as alterações climáticas acarretam, o estudo destas temáticas constitui uma prioridade na agenda de diversas nações a nível socio-económico e científico. É vital, portanto, compreender o sistema climático ampliando o conhecimento sobre os mecanismos forçadores de clima e respetivas consequências nas condições climáticas no Atlântico Norte. Neste contexto, o estudo das variações climáticas registradas no presente período interglacial, o Holocénico, representa especial relevância.

Estudos sobre os interglaciários e em particular sobre o Holocénico (últimos 11500 anos) são um dos principais temas de investigação atuais. Nos últimos anos foram efetuados estudos em variadíssimos registos naturais (ex. lagos, sedimentos marinhos) de modo a compreender a natureza, duração e causas das oscilações climáticas que ocorreram durante o Holocénico. Todavia, muitas das reconstituições climáticas existentes até à data, não se baseiam na correlação direta entre o oceano, o continente e o gelo, tornando difícil obter com precisão o conhecimento das interacções entre os sistemas atmosfera-oceano-continente e do seu real impato na variabilidade climática global. Acresce ainda, à impossibilidade de se estabelecer uma correlação direta, o fato de nenhum destes registos isolados ser adequado para identificar a variabilidade temporal e espacial necessária à comparação das variações climáticas regionais com modelos climáticos. Consequentemente, o tipo de mecanismos responsáveis pela variabilidade climática Holocénica está longe de ser reconhecido.

Os principais objetivos deste trabalho são a) determinar e caraterizar a evolução do clima e da vegetação no Holocénico no sudoeste da margem continental Francesa/Norte de Espanha ; e b) detetar e compreender a frequência, duração e amplitude da variabilidade climática no Holocénico, assim como inferir sobre os principais mecanismos forçadores. Para tal, foi efetuado um estudo polínico de alta resolução temporal numa sondagem colhida num ponto geograficamente estratégico Atlântico Norte: norte da Península Ibérica/sudoeste da margem continental Francesa.

Este estudo mostra que a vegetação na região de estudo ao longo do Holocénico respondeu à variabilidade climática orbital e sub-orbital, e em particular ao evento abrupto designado por 8.2 ka.

A diminuição gradual da floresta temperada, em particular do *Pinus* e do *Quercus* decíduo, acompanhada de um aumento sucessivo de plantas herbáceas, sugere um arrefecimento progressivo compatível com a diminuição da insolação de verão das médias latitudes do Hemisfério Norte e a diminuição gradual do $\delta^{18}O$ nos registos de gelo na Gronelândia. Durante o Holocénico médio e superior, a substituição do *Quercus* decíduo e *Pinus* pelo *Fagus*, sugere, além do arrefecimento progressivo, um aumento das condições de umidade e uma diminuição da sazonalidade. A redução da sazonalidade é contemporânea com o aumento geral da precessão.

Superimposta a esta variabilidade climática orbital, verificou-se um episódio caracterizado pela diminuição da floresta temperada, especialmente de *Corylus*, juntamente com um aumento significativo das herbáceas, sobretudo Cyperaceae e a presença de *Carpinus*. Estes indicadores atestam a presença do evento frio e húmido designado por "evento 8.2 ka" no norte da Península Ibérica/sudoeste da margem continental Francesa. Todas as evidências apontam para os episódios terminais de expulsão dos lagos de "Agassiz" e de "Ojibway" e a consequente redução gradual da "MOC" (meridional overturning circulation), como as principais causas para o súbito arrefecimento durante o evento 8.2 ka. A diminuição da intensidade da circulação termohalina terá impedido o transporte de calor para as altas latitudes provocando a diminuição da temperatura registada no Atlântico Norte e na Europa. Este trabalho propõe que o mecanismo atmosférico que explica as condições húmidas durante este evento nas latitudes médias da Europa envolve alterações na atividade ciclónica e na posição da Corrente de Jato no Atlântico, e a prevalência de situações de forte circulação zonal com frequentes sistemas depressionários tipicos de uma ausencia de eventos de bloqueio na zona de estudo. Além disso, o aumento na quantidade de nuvens durante este evento abrupto pode ter induzido à particular diminuição de *Corylus* (árvore dependente de bastante luz para o seu desenvolvimento) através do bloqueio da luz solar e consequente diminuição da sua disponibilidade.

A variabilidade climática sub-orbital não é muito evidente após o evento 8.2 ka no nosso registo polínico. No entanto, as percentagens de todos os *taxa* foram submetidas a uma análise espetral (Wavelet), de forma a determinar a evolução temporal das

amplitudes e periodicidades prevalentes das variações climáticas holocénicas na região da Atlântico Norte em estudo. Foi obtida uma ciclicidade de ~500 anos para o *Corylus*. Esta ciclicidade é semelhante à detectada em registos na Gronelândia e no Atlântico Norte, o que sugere que esta espécie, em particular, terá respondido aos mesmos mecanismos climáticos forçadores (variações na atividade solar e/ou perturbações da circulação termohalina). Contudo, o nosso registo não possui resolução temporal suficiente para explorar esta possibilidade, sendo necessário para isso efetuar estudos adicionais.

No último milénio, tornou-se evidente que o impato antropogénico através da presença contínua de espécies indicadoras de atividade antropogénica, como *Castanea sativa*, *Juglans* e cereais. O impacto humano aparenta ter sido sobreposto à variabilidade climática natural milenar durante este milénio.

Este estudo contribuiu para a reconstrução das condições paleoclimáticas e a resultante resposta da vegetação ao longo do Holocénico no Norte da Península Ibérica/Sul de França; bem como para a compreensão dos mecanismos forçadores responsáveis por esta variabilidade climática orbital e sub-orbital. Os resultados desta pesquisa serão integrados nos dados existentes de alta resolução de várias regiões geográficas "chave" do Atlântico Norte incluídas no projeto CLIMHOL " Variabilidade climática Holocénica registada no Atlântico Norte e continente adjacente: correlação directa oceano-continente" (referência PTDC/AAC- CLI/100157/2008), financiado por fundos nacionais através da FCT/MCTES (PIDDAC) e co-financiado pelo Fundo Europeu de Desenvolvimento Regional – FEDER- através do COMPETE – Programa Operacional Factores de Competitividade – POFC.

É importante realçar que a avaliação do tempo e a natureza de resposta da vegetação a eventos abruptos como o 8.2 ka é de particular importância pois os modelos climáticos preveem uma redução na intensidade da "MOC" devido ao aquecimento global.

Em termos de investigação no futuro, pretende-se continuar a aprofundar o conhecimento dos mecanismos envolvidos nas alterações climáticas. Deste modo, é essencial detectar e compreender a frequência, duração e amplitude e os mecanismos responsáveis pela variabilidade climática natural em períodos interglaciares com condições análogas ao Holocénico, mas que não são influenciadas pelas atividades humanas.

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CHAPTER 2

HOLOCENE CLIMATE VARIABILITY IN THE MID-LATITUDES OF THE EASTERN NORTH ATLANTIC REGION

APPENDIX 1

Images of the pollen grains presented in the KS05 10 synthetic pollen diagram

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Chapter I

1. MAIN OBJECTIVES AND MOTIVATION

Climate is changing significantly since the last few decades, affecting people and the environment worldwide. In particular, increasing air and ocean temperatures, widespread melting of snow and ice, and rising sea levels result in a shift in the Earth's climate system equilibrium. Natural events and human activities are believed to be contributing to this climatic trend. Differentiating natural from anthropogenic forcing of climate change is one of the most important challenges of future climate prediction. There is therefore an urgent need to improve our documentation and understanding of natural variability for periods stretching back beyond the instrumental record. From this perspective, it is of extreme importance to know more about natural climate variations that occurred during the current interglacial, the Holocene, because its boundary conditions are similar to those experienced now and in the near future.

Growing evidence suggests that variations in Holocene climate were larger than previously believed. In the last few years, numerous studies have been performed on several naturally occurring archives such as lake and marine sediments, tree rings, speleothems and ice cores to understand the nature, timing and causes of Holocene natural climate oscillations. Such studies have shown that superimposed on the orbitally-induced long-term cooling sub-orbital millennial-scale climate variability has affected this interglacial.

The most extreme short-lived cold episode noticed in the Greenland Ice cores, the "8.2-kyr-BP event", lasted 100-200 years and has been detected elsewhere in the North Atlantic and in Europe. Several hypotheses have been invoked to explain this cooling such as: a) significant alterations of solar activity or b) changes in general circulation pattern of the North Atlantic region. The first hypothesis suggest a reduction of the solar activity as the result of increasing sunspots while the second hypothesis suggest that this cooling was triggered by the final catastrophic drainage of the Lakes Agassiz and Ojibway which contributed to the introduction of large amounts of freshwater into the North Atlantic Ocean, disturbed the thermohaline circulation and cooled both Europe and North America. In recent years, few studies have shown that this event was longer and more complex than previously believed. Yet, the spatial distribution, timing, amplitude and their impact in the ecosystems remains one of the outstanding mysteries of climate variability. Also, the 8.2 ka event is seen as the best analogue for the "worst case" scenario for the future. Thus, the global warming is accelerating the melting of the Greenland ice sheet which is contributing to a drastic increase of the global sea level,

threatening low-lying areas around the globe with beach erosion, coastal flooding, and contamination of freshwater supplies. As this ice melts, voluminous amounts of cold, fresh water dump into the world's oceans disrupting the general trend of the global oceanic circulation leading to an extreme cooling over Europe and North America. Understanding the changes in the frequencies and intensities of extreme climate events and weather, as well as in sea level rise, is therefore a necessity for reducing adverse impacts on natural and human systems.

In contrast, the signature of millennial-scale climate changes during the mid- and late Holocene (after the “8.2 ka event”) is not easily detected in paleoclimatic records. However, this variability was strong enough to affect human societies, particularly during the last millennium, as historically documented for the Little Ice Age (LIA) and the Medieval Warm Period (MWP). The global impact, amplitude, periodicities and causes triggering these short-lived climatic oscillations have been widely discussed in the last decade.

There is therefore an urgent need to understand the nature, timing and causes of climate oscillations and to determine how widespread, systematic and abrupt they may have been.

Most of the available Holocene climatic reconstructions are however, not based on good correlation between terrestrial, marine and ice records making it difficult to get an accurate understanding of the interactions of the atmosphere-ocean-land systems and their impact on global climate variability. For this reason, the mechanisms that control Holocene climate variations in the North Atlantic and adjacent landmasses are far from being resolved. We propose, therefore, to establish for the first time, high-resolution sea-land correlation on a shelf core from southwestern France (highlighted by the IPCC models as one of the most sensitive region to the ongoing global climatic changes) covering the Holocene.

This master thesis aims to improve the understanding of the nature, timing and causes of Holocene natural climate oscillations and to determine how abrupt they may have been. Furthermore, this thesis aims to document how changes in the behaviour of coupled atmosphere-ocean-land systems have affected climate in the North Atlantic region during the Holocene. In particular we pretend:

a) to document the response of vegetation to Holocene climate changes in terms of long and millennial scale and in particular:

- to determine the impact of long term cooling in the vegetation cover of both northern Iberian Peninsula and southwestern France;

- to determine the impact of the 8.2 ka event in both northern Iberian Peninsula and southwestern French margin;

- to determine the trigger mechanisms involved in the climatic signal left by the 8.2 ka in southwestern French margin/northern Spain;

- to determine the impact of less extreme events in the northern Iberian Peninsula and southwestern France and to determine, if possible, the cyclicities within the sub-orbital climatic oscillations in terrestrial environments and to determine the trigger mechanisms involved in those changes;

b) to compare the obtained marine pollen data with other marine paleoclimatic and ice records to further understand the nature, amplitude and timing of millennial-scale climate variability in northern Iberian Peninsula and southwestern France;

c) to understand if the sub-orbital millennial-scale climate oscillations had contributed to an amplification or reduction of the long-term climatic signal;

To achieve theses aims, I have organised the work with the following sequence of steps:

1. Obtain high-resolution data and high-quality chronology

High time-resolution analyses enabled by both high sedimentation rate marine cores and well chronologically constrained records (using several AMS ^{14}C dating) are a prerequisite to investigate the rapid climatic variability of the Holocene.

2. Document the Holocene southern European vegetation changes

Pollen, representing an integrated image of the regional vegetation of the borderlands, will be analysed.

3. Integrate terrestrial and marine data results

An accurate correlation of marine and terrestrial settings is of prime importance for time equivalent documentation of environmental changes and to evaluate the synchronicity of occurring climatic shifts and events in both environments.

4. Implement cyclicity analysis of the marine and terrestrial records

This task will allow us to determine the main trigger mechanisms involved in the Holocene millennial scale climatic variability.

5. Evaluate the natural and anthropogenic contributions to the Holocene climatic changes.

2. CLIMATE VARIABILITY DURING THE LATE QUATERNARY

During the late Quaternary (roughly the last million years) global climate has changed dramatically due to a number of linked physical, chemical and biological processes occurring in the atmosphere, land and ocean.

Changes in Earth's orbit (Milankovitch cycles) as well as oscillations in solar activity determine the temporal and spatial distribution of insolation, being considered the ultimate forcing of the long-term climate oscillations of the Quaternary. Additionally, Earth's internal mechanisms that result from the interaction between atmosphere-ocean-lithosphere-cryosphere-land systems have a huge contribute to the climatic changes. The variability of these global phenomena may trigger, amplify, sustain or globalize rapid climatic fluctuations (Peixoto and Oort, 1992).

2.1 Long term climate variability (Glacial-Interglacial cycles)

Over the last million years, the earth's climate system has experienced several long term climatic shifts between glacial and interglacial conditions which were mainly controlled by solar irradiance variations linked to changes in the Earth's orbit around the sun (Milankovitch, 1920; Berger, 1978; Imbrie et al., 1992; Berger and Loutre, 2004; Ruddiman, 2006). Changes in the amount of incoming solar radiation, as well as their temporal and spatial distribution are determined by three types of orbital parameters (Milankovitch, 1920) (Fig. 1):

(1) Eccentricity, reflects the shape of Earth's orbit around the Sun, ranging from a quasi-circular (low eccentricity of 0.0006) to a slightly elliptical shape (high eccentricity of 0.0535) and with two periodicities of about 100 and 400 thousands of years (kilo years, hereafter referred to as ka) (Berger and Loutre, 1992; Berger and Loutre, 2004). Differences in solar radiation received on earth of about 30% may occur between perihelion (in early January, when the Earth is closest to the sun) and aphelion (in early July, when the Earth is further from the sun) during eccentricity maxima (Goodess et al., 1992). In contrast, during episodes of low eccentricity (0.016) as the present-day, the difference in insolation between perihelion and aphelion is around 6.4 % over the year (Berger, 2001).

(2) Obliquity of Earth's axis in relation to the orbital plan, tilting from 22° and 25° over a period of about 41 ka (Berger and Loutre, 2004), determine the differences in seasonal contrast on earth (Buchdahl, 1999). Increased obliquity implicates a higher seasonal contrast in both hemispheres at high latitudes, since summers receive more solar radiation and winters less (i.e. warmer summers/colder winters). Inversely, decreased obliquity origin temperate summers and milder winters, which is the most likely mechanism promoting the onset of glacial conditions favoring the ice cap growth in the high latitudes. Fluctuations in obliquity have less influence at low latitudes, as the strength of the effect decline towards the equator (Buchdahl, 1999).

(3) precession (change in the orientation of the Earth's rotational axis) is modulated by eccentricity, which splits the precession into two periods of about 23 ka and 19 ka, leading to an average period of 21 ka (Berger, 2001; Berger and Loutre, 2004). This cycle has two components: an axial precession, caused by the gravitational forces exerted on Earth of all other planetary body's in our solar system, and an elliptical precession, in which the elliptical orbit of the Earth itself rotates about one focus (Buchdahl, 1999). Changes in axial precession modify the times of perihelion and aphelion, and consequently increase the seasonal contrast in one hemisphere and decrease in the other hemisphere. The hemisphere at perihelion experiences an increase in summer solar radiation and a cooler winter, while the opposite hemisphere will have a warmer winter and a cooler summer. Presently, the Earth is at perihelion in the northern hemisphere (hereafter referred to as NH) winter, which makes the winters and summers less severe in this region (Ruddiman, 2001).

Orbital forcing is the only forcing that is fully understood, and can be calculated not only for the past but also for the future several million years (Berger and Loutre, 2004). Isolated or combined together, the orbital parameters shape the distribution of solar radiation in Earth's surface. Whereas eccentricity is the only Milankovitch cycle that modify the annual-mean global solar insolation at aphelion and perihelion; the seasonal and latitudinal variation of the incoming radiation is balanced by precession and obliquity respectively.

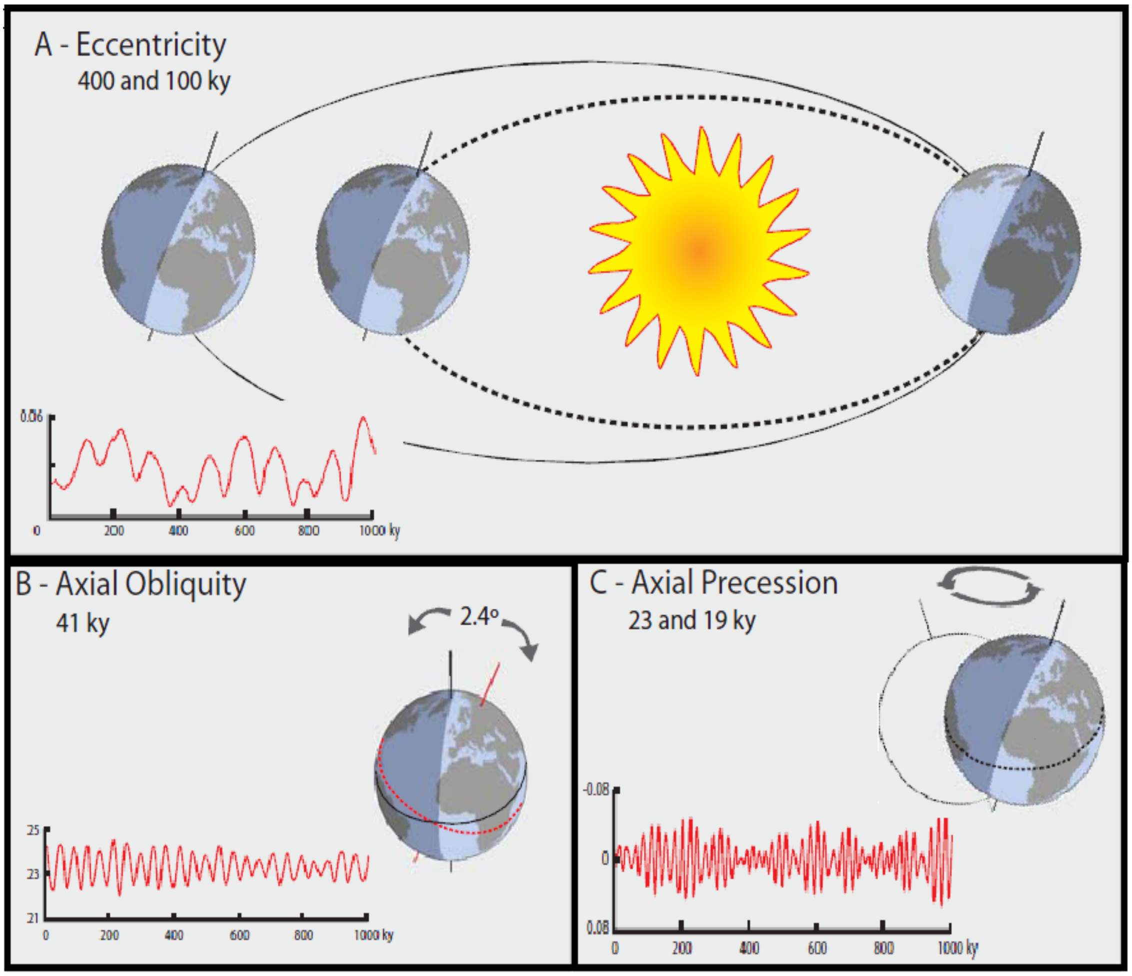


Fig. 1. Schematic representation of the Earth's orbital cycles (adapted from Zachos et al., 2001). (A) Changes in the eccentricity of the Earth's orbit, with periods of 400 and 100 ka. (B) Variations in the obliquity, or tilt of the Earth's axis, with amplitude of 2.4° every 41ka. (C) Precession, that corresponds to changes in the direction of the Earth's axis relative to the fixed stars, with periods of 23 and 19 ka.

The timing, duration and amplitude of major glacial-interglacial cycles have been modulated by changes in the prevailing astronomical parameters. Thus, Early Pleistocene low-amplitude 41 ka obliquity-forced climate cycles were gradually replaced by later Pleistocene 100 ka eccentricity-forced climate cycles (Head and Gibbard, 2005) (Fig. 2). The transitional period occurring between the 41 ka and 100 ka cyclicities is known as the Mid Pleistocene Transition (MPT) (1.25-0.7 Ma) (Clark et al., 2006).

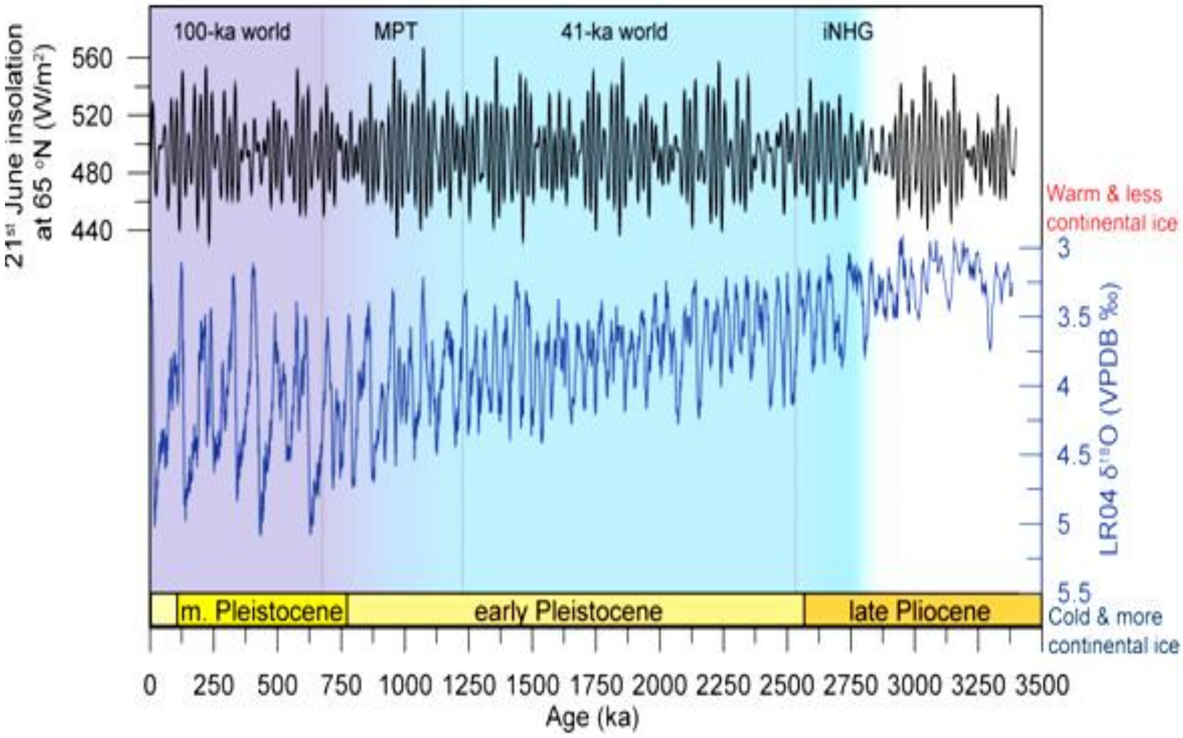


Fig. 2. Orbitally induced 100 ka and 41 ka climatic oscillations. In the top of the figure, Summer solstice insolation at 65 ºN (Laskar et al., 2004). The Benthic foraminiferal $\delta^{18}O$ stack of the last 3.5 Ma (Lisiecki and Raymo, 2005) shows a general cooling trend and an increase of global ice volume over the past 3 Ma (from Naafs, 2011). iNHG refers to the intensification of the NH glaciations and MPT refers to the middle Pleistocene transition.

This transition favored the gradual intensification (ice volume increase) and longer duration of glacial states (Imbrie et al., 1993) (Fig. 3). Interglacials before the Mid-Brunhes Event (MBE) at ~425 ka were less intense than post-MBE interglacials, meaning that temperatures were probably colder and sea level and CO_2 concentrations in the atmosphere were lower (Lisiecki and Raymo, 2005) (Fig.3). Thus, the last 1 Ma is characterized by long cold periods with massive continental ice-sheets and polar ice caps followed by short warmer stages with ice-free poles and sea level rising (Zachos et al., 2001).

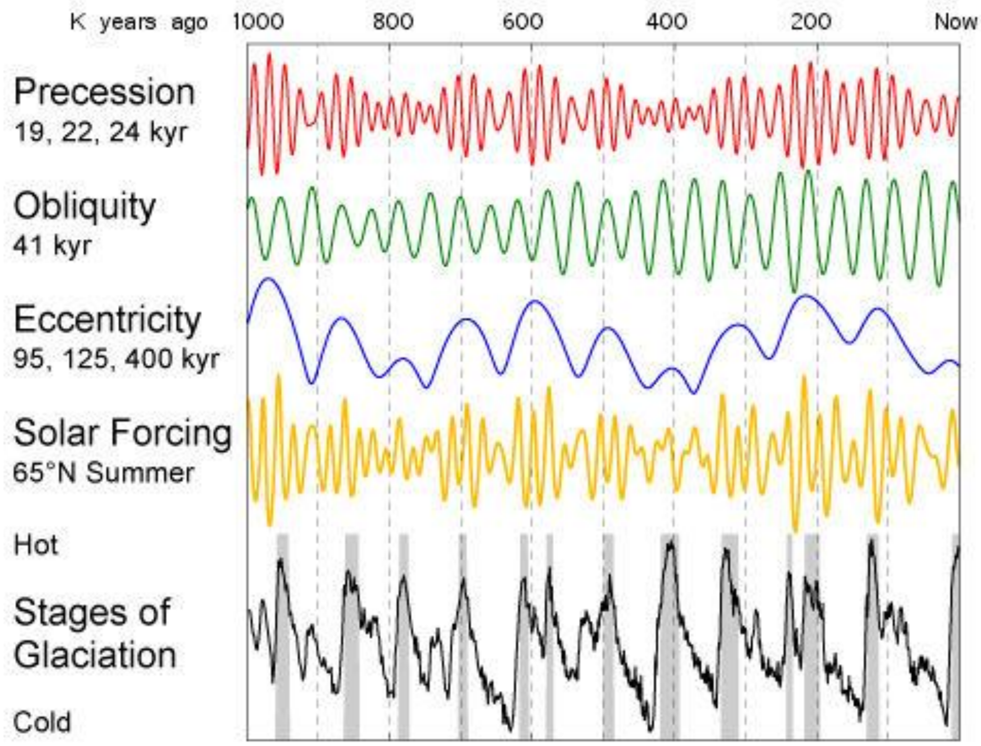


Fig. 3. Each orbital parameter is shown over the last 1000 ka along with the insolation and stages of glaciation. The Milankovitch cycles have influenced climate change in 100 ka periods and determined the frequency of Quaternary glaciations (Berger, 1978).

2.2 Millennial-scale climate variability

Superimposed on the long term variability, the last glacial cycle is characterized by large, abrupt and widespread millennial-scale climatic changes, known as Dansgaard-Oeschger (D-O) events (Dansgaard et al., 1993) (Fig. 4). The D-O events are characterized rapid warming (Greenland interstadials - GIS) and gradual cooling episodes (Greenland stadials - GS), occurring every 1500 years (Dansgaard et al., 1993). During the last climate cycle, twenty five D/O events were detected in Greenland with atmospheric temperatures presenting a warming of 8ºC to 16ºC within several decades (e.g. Severinghaus and Brook, 1999; NGRIP members, 2004).

Some of those D-O stadials were extremely cold, occurring every 5-10 ka (Elliot et al., 1998), and are known as Heinrich events (Bond and Lotti, 1995). Heinrich events were first documented in several North Atlantic deep-sea cores between 45 and 50° N (Ruddiman belt) (Ruddiman, 1977; Heinrich, 1988; Bond and Lotti, 1995) and identified by the anomalous presence of ice-rafted detritus (IRD) that were transported

to the ocean by drifting icebergs from the Laurentide and northern European ice sheets as well as by synchronous peaks of polar foraminifera, *N. pachyderma (s)* (e.g. Bond and Lotti, 1995; Hemming, 2004), sea surface temperature decreases (Bond and Lotti, 1995; Cortijo et al., 1997) and magnetic susceptibility peaks (Grousset et al., 1993). These coarse fraction intervals, representing the well known IRD layers, were also detected beyond the Ruddiman belt i.e. north of 50°N (e.g. Elliot et al., 1998; Fronval et al., 1995; Rasmussen et al., 1996; Van Kreveld et al., 2000; Voelker et al., 1998) as well as below 40°N (e.g. Baas et al., 1997; Bard et al., 2000; Chapman et al., 2000; de Abreu et al., 2003; Lebreiro et al., 1996; Zahn et al., 1997; Naughton et al., 2009). The thickness of the IRD layers and the magnetic signal is, however, smaller in the mid-latitude sites than in the northern ones (Thouveny et al., 2000). Also, the duration of the impact of these extreme events on the sea surface temperatures (SST) in this region is longer than that of the IRD layers (e.g. Bard et al., 2000; Chapman et al., 2000; Sánchez Goñi et al., 2000; Naughton et al., 2009). Moreover, the vegetation patterns within these extreme cold events are also complex in southwestern Europe (Naughton et al., 2009).

Internal mechanisms have been invoked to explain Heinrich climatic anomalies. Thus, the introduction of anomalous freshwater pulses into the North Atlantic have affected the general pattern of the global thermohaline circulation (THC), by forcing the THC to slowdown (almost shutdown) (e.g. Ganopolski and Rahmstorf, 2001; Knutti et al., 2004), triggering a substantial SST drop in the North Atlantic region and extreme cooling in Europe (e.g. Paillard and Labeyrie, 1994; Seidov and Maslin, 1999). The slowdown/shutdown of the THC also precludes the moisture transfer to Europe. Following this, rapid oceanic and atmospheric reorganizations favoured the transfer of cold conditions everywhere on earth suggesting that although Heinrich events were mainly North Atlantic (hereafter referred to as NA) phenomena they had a global impact (Leuschner and Siroko, 2000; Voelker et al., 2002). Several other hypotheses have been invoked to explain the response of the ocean-land-ice to Heinrich events (e.g. Flückiger et al., 2006; Sánchez Goñi et al., 2002; Naughton et al., 2009). However, despite the recent paleodata acquisition and modelling efforts, understanding of the mechanisms that give rise to theses instabilities is far from being completely understood (Hemming, 2004).

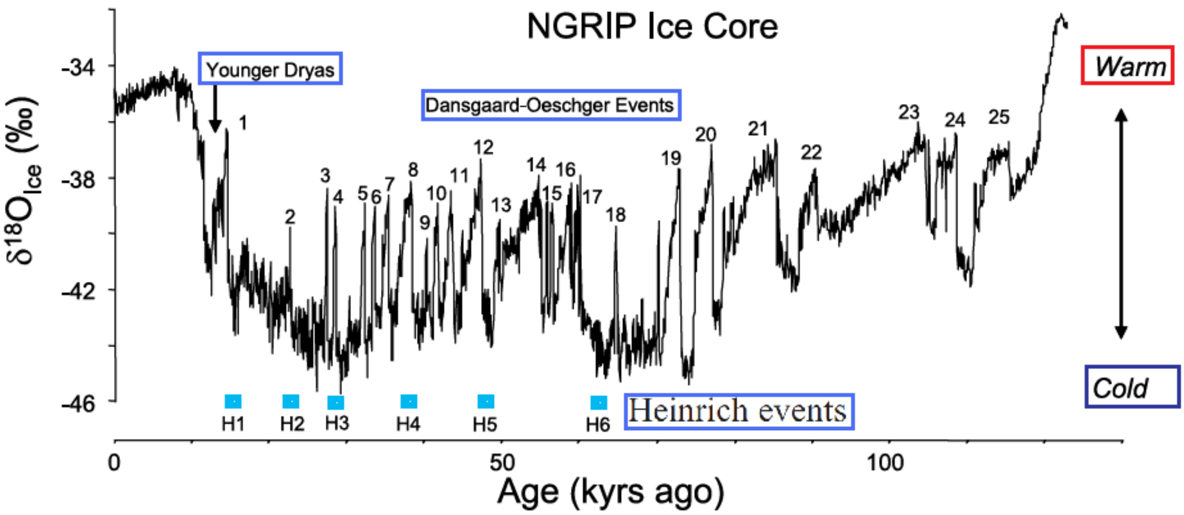


Fig. 4. North Greenland Ice Core Project (NGRIP) ice core $\delta^{18}O$ record for the last 123 ka. The abrupt cooling reflecting the Younger Dryas corresponds to the Greenland stadial 1. The rapid warmings that characterize the D/O interstadial events are numbered from 1 to 25. Heinrich events are labeled from H1 to H6. Some stadials coincide with Heinrich events (from NGRIP members, 2004).

The millennial-scale fluctuations were largest when glacial ice sheets existed in the NH, and they have been less significant during interglacial climates as the present one (Ruddiman, 2001). Nevertheless, the mechanisms and causes that drive these abrupt changes are not totally understood and remain one of the greatest challenges in climate research.

2.3 The Holocene

For a long time the Holocene (present-day interglacial; past 11,7 ka) was considered a period of stable warm climatic conditions (e.g. Dansgaard et al., 1993; McManus et al., 1999). However, in the last few years a wide array of different high-resolution paleoclimatic records such as lake and marine sediments, peat bogs, tree rings, speleothems and ice cores together with climate simulations have shown that a long term and millennial scale climate variability has affected this interglacial (e.g. Kutzbach and Gallimore, 1988; Bond et al., 1997; 2001; Crucifix et al., 2002; Weber and Oerlemans, 2003; Mayewski et al., 2004; Renssen et al., 2005; Wanner et al., 2008; 2011).

2.3.1 Holocene long term climatic changes

The decrease in the northern high-latitudes summer insolation along the Holocene induced a long-term cooling trend in the NH confirming that orbitally induced changes in NH insolation is the major driver of long-term climate variations during the Holocene (Kutzbach and Gallimore 1988; Braconnot et al,. 2000; Duplessy et al., 2001; Johnsen et al., 2001; Crucifix et al,. 2002; Marchal et al., 2002; Andersen et al., 2004; Kim et al., 2004; Moros et al., 2004; Solignac et al., 2004; Weber et al., 2004; Renssen et al., 2005; Keigwin et al., 2005).

During the Holocene, the insolation changes are dominated by the effect of precession and obliquity astronomical parameters. Fig. 5 clearly shows the seasonal hemispheric insolation changes as a function of latitude, with the summer insolation curves of both hemispheres exhibiting an opposite behavior. In the NH the summer insolation slightly increase by the beginning of the Holocene with a peak around 11-10 ka BP (BP means "before present" and 0 BP represents 1950 AD) and afterward decreases towards modern values. The NH winter insolation shows a change in the opposite direction with a steady raise since the beginning of the Holocene.

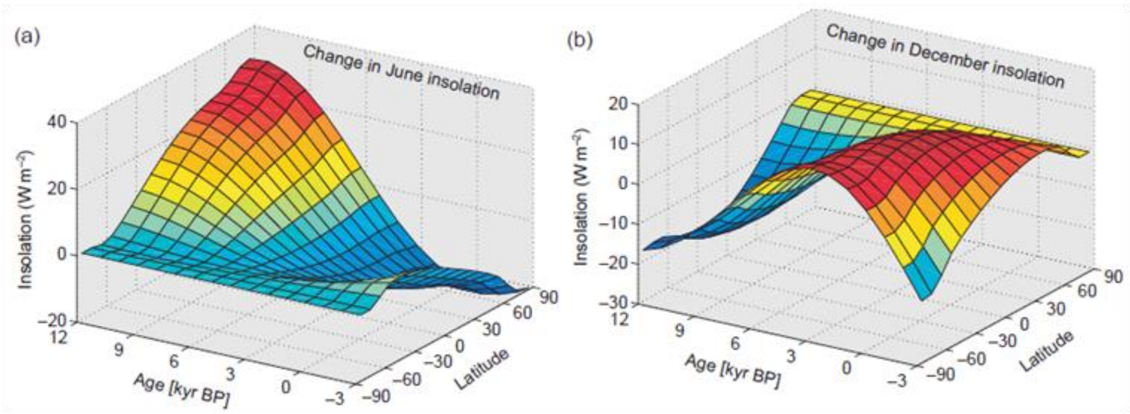


Fig. 5. Variations in orbital forcing during the Holocene (from 12 ka BP to 3 ka in the future) in June (a) and December (b) as a function of latitude (from Beer and Geel, 2008). In June there is a strong decreasing trend in the north (a). In December, the insolation on the southern Hemisphere first increases and then reaches its maximum between 2 and 4 cal. ka BP before decreasing again (b).

The gradual increase of precession consequently induces a reduction of the NH seasonality during the Holocene. This is in agreement with climatic models of Holocene climate evolution that reveal an early optimum (9–8 ka BP) in the high northern latitudes followed by a gradual decrease in summer temperatures (1-3°C), a reduction in summer precipitation and the expansion of sea-ice cover (Renssen et al., 2005).

Nevertheless, the greenhouse gas forcing partly counteracts (+ 0.5° C) the decrease in summer insolation in the areas north of 60ºN over 9 ka (Crucifix et al., 2002; Renssen et al., 2005).

The orbital induced NA long term cooling of the Holocene is also accompanied by a southern shift of the Intertropical Convergence Zone (ITCZ) and a weakening of the NH summer monsoon systems (Mayewski et al., 2004; Braconnot et al., 2007; Wanner et al., 2008; 2011).

Besides the long term cooling, the Holocene is divided in three main periods (Table 1): The early Holocene deglaciation (~ 11.7 - 7 ka BP), Holocene Thermal Optimum (~ 7 - 4.2 ka BP) and Neoglacial (~ 4.2 ka BP - pre-industrial time) (e.g. Nesje and Dahl, 1993; Marchal et al., 2002; Wanner et al., 2008; 2011).

Period	Main features
Early Holocene deglaciation	Climatic amelioration when compared with the previous glacial period. This includes the North European “Preboreal” and “Boreal” chronozones. - High NH summer insolation. The insolation maximum occurred at around 11 ka BP due to a coincidence of the appropriate phases of both precession and obliquity cycles. - Deglaciation of the North America and Eurasia ice sheets triggering a slight cooling effect at least at a regional scale. - High monsoon activity in Africa and Asia. - Strong sea level rise.
Holocene Thermal Optimum	Relatively warm period called “Holocene Climate Optimum” and “Hypsi- or Altithermal”, which also includes the North European “Altantic” chronozone. - Decreasing NH summer insolation, southward displacement of the ITCZ. - Summer temperatures in the NH mid- and high-latitude regions are higher than during the pre-industrial period (prior to year 1900 AD). - North American ice sheet is no longer large enough to influence climate at a hemispheric scale. - Most of the global monsoon systems still active, but weaker. - Major sea level changes ceased after ~6 ka BP.

Neoglacial	Rather cold phase that coincides with the North European "Subboreal" and "Subatlantic" chronozones. - Low values of NH summer insolation leading to a decrease in summer temperatures. - Occurrence of several cold relapses with remarkable glacier advances in different areas of the globe. - Sea ice increases in the high latitudes. - The Neoglacial period was interrupted by the global warming driven by anthropogenic forcing (increased anthropogenic greenhouse effect), therefore it lasted until the beginning of industrialization.

Table 1. The Holocene three main periods (adapted from Nesje and Dahl, 1993; Marchal et al., 2002; Wanner et al., 2008; 2011).

Note: The Holocene has been subdivided into five time intervals, or chronozones, based on Northern European climatic stratigraphies (Preboreal, Boreal, Atlantic, Subboreal and Subatlantic). The Hypsi- or Altithermal refers to warm conditions in northern mid- to high latitudes. Nevertheless these terms are related to regional climatic fluctuations and have not been consistently applied (Wanner et al., 2008).

Although summer insolation in the NH decreased after ~11-10 ka BP, the timing and magnitude of the Holocene Thermal Maximum (hereafter referred to as HTM) was not synchronous across the hemisphere (Fig. 6), suggesting the involvement of extra feedbacks and forcings (e.g. Kaufman et al., 2004; Renssen et al., 2005; IPCC, 2007; Renssen et al., 2009). The HTM was detected at the beginning of the Holocene (~11-8 ka BP) in several paleoclimatic studies of the NA and adjacent Arctic (Duplessy et al., 2001; Andrews and Giraudeau, 2002; Marchal et al., 2002; Kaufman et al., 2004; Kim et al., 2004; Knudsen et al., 2004; Vernal et al., 2005; Kaplan and Wolfe, 2006). Also, the Iberian margin records, located at the mid-latitudes of the NA, clearly show an early Holocene thermal maximum (Bard et al., 2000; Pailler and Bard, 2002; Rodrigues et al., 2010) synchronous with the maximum expansion of temperate trees in the adjacent landmasses (Turon et al., 2003; Naughton et al., 2007a). However, other proxy data from northern high latitudes indicate that the HTM occurred considerably later, after ~8 ka (Dahl-Jensen et al., 1998; Korhola et al., 2000; Johnsen et al., 2001; Levac et al., 2001; Kaufman et al., 2004; Solignac et al., 2004; Keigwin et al., 2005; Kaplan and Wolfe, 2006). In Europe and north-western North America the Holocene maximum warmth was observed between 8 and 4 ka BP (Davis et al., 2003; Kaufman et al., 2004).

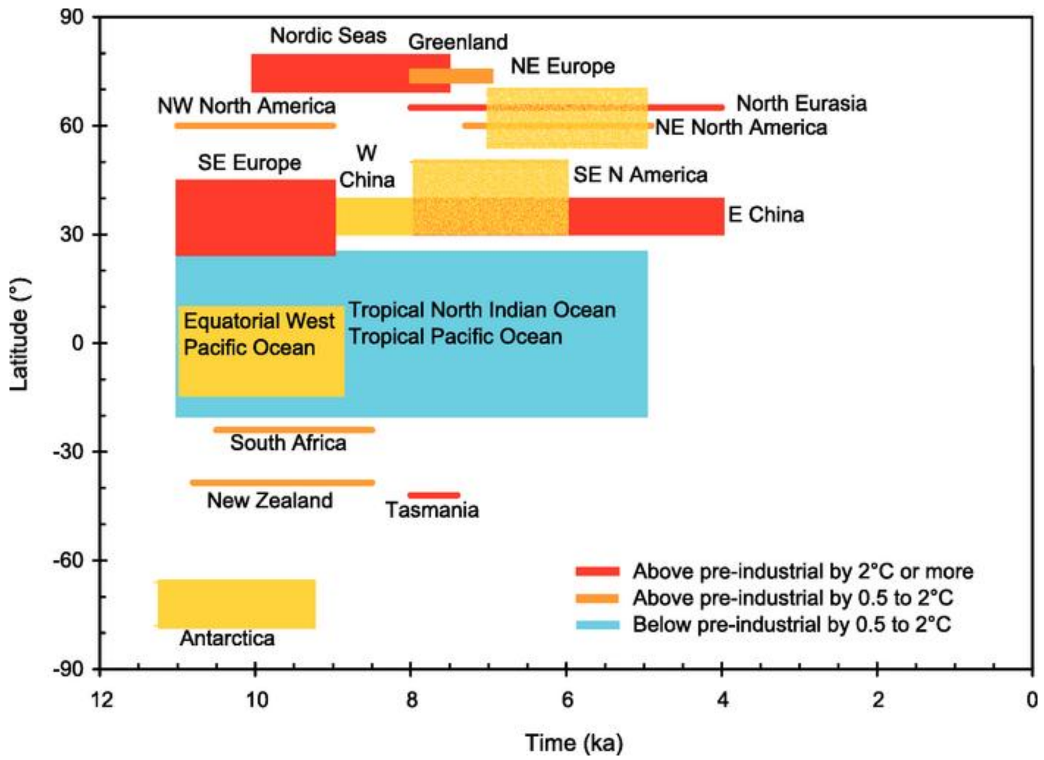


Fig. 6. Timing and intensity of maximum temperature deviation from pre-industrial levels, as a function of latitude and time (from IPCC, 2007). It is suggested a possible south to north pattern, with southern latitudes showing HTM a few millennia earlier than the NH regions (IPCC, 2007).

2.3.2 Holocene millennial to sub-millennial scale climate variability

Superimposed on the orbital induced long-term cooling trend a series of relatively cold events punctuated the Holocene. Denton and Karlén (1973) have suggested that advances of the Northern Europe and America mountains glaciers had responded to millennial scale climate coolings; O'Brien et al., (1995) have showed that Holocene atmospheric circulation above the ice cap of Greenland ice sheets was disturbed by a series of millennial scale shifts. Encouraged by these evidences Bond et al., (1997) have launched a deep sea sedimentary research in the NA revealing the presence of IRD (Ice rafted detritus) and a decrease in the sea surface temperatures during the coldest phases of this millennial scale variability. Following these pioneer works, these short-lived climate oscillations have been detected in a variety of paleoclimatic archives such as ice, marine and terrestrial records (Fig. 7) (e.g. Mayewski et al. 2004; Wanner et al., 2008; 2011).

The timing of millennial-scale climate oscillations detected in the NA region during the Holocene is of about 2800-2000 and 1500 yr (e.g. Denton and Karlén, 1973; O'Brien et al., 1995, Bond et al., 1997; 2001, Mayewski et al., 2004).

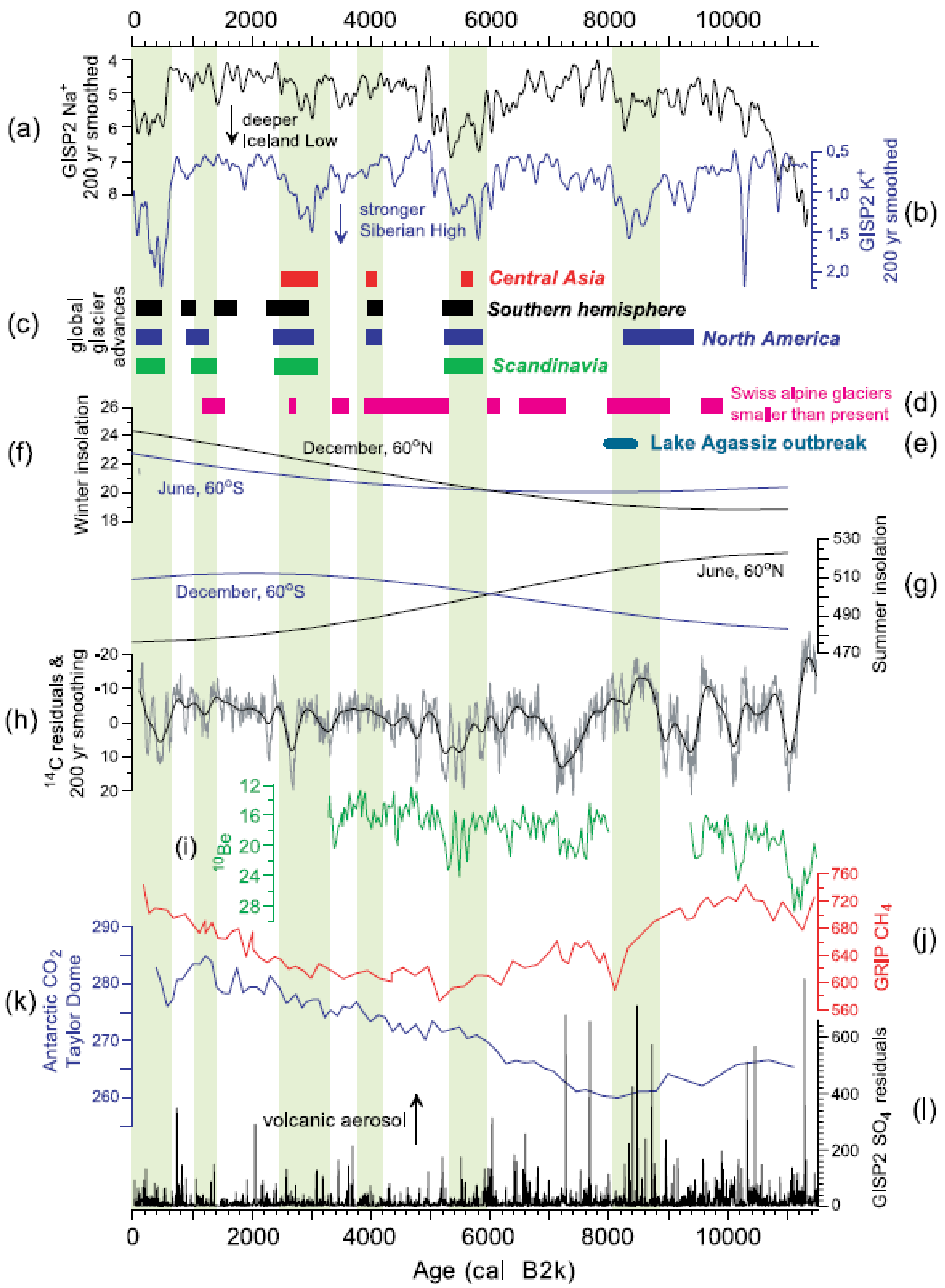


Fig. 7. Globally distributed glacier fluctuation records and climate forcing time series (cosmogenic isotopes reflecting solar variability, orbital insolation changes, volcanic aerosols, and greenhouse gases). Green bands represent timing of rapid climate change (RCC) identified by Mayewski et al. (2004) by tuned to GISP2 (Greenland Ice Sheet Project 2) record. For more details see Mayewski et al. (2004).

A number of processes such as solar activity, volcanic eruptions and internal oscillations of the climate system driven by land-ocean-atmosphere interactions have been put forward to explain the millennial-scale variability (e.g. Mayewski et al., 2004; Wanner et al., 2011) and references therein for a comprehensive overview) (Fig. 8).

In the NH the most important internal variations of the climate system include the Arctic Oscillation/North Atlantic Oscillation (e.g. Hurrell, 1995, hereafter referred to as AO/NAO), the Atlantic Multidecadal Oscillation, the Pacific Decadal Oscillation and the variations in the NA thermohaline circulation (hereafter referred to as THC). In terms of the major phenomena in the Atlantic, it is important to note that during the past 6 ka there is evidence for a continuous SST decrease in the NA and a negative trend (weakening) in the NAO index (Rimbu et al., 2003; Wanner et al., 2008). According to Rimbu et al. (2003) NAO may have played a role in generating millennial-scale SST trends, with the positive (negative) phase accompanied by relatively mild (cold) winters over northern Europe and a relatively cold (warm) climate in the eastern Mediterranean and the Middle East.

Fig. 8 shows the four most important external climate forcing factors during the Holocene and six cold relapses with no clear cyclicity (8200, 6300, 4700, 2700, 1550 and 550 years BP) recognized by Wanner et al. (2011) at multidecadal to multicentury scale in the NH. The complex spatiotemporal pattern of the cold relapses may be a consequence of different dynamical processes such as low solar activity, combined with a meltwater flux into the NA, a possible slowdown of the thermohaline circulation and, in some cases, series of large tropical volcanic eruptions (Wanner et al., 2011). Moreover, feedback mechanisms may have been important, in particular when regional or even local phenomena are studied. Thus, although the global impact, amplitude, periodicities and causes triggering these short-lived climatic oscillations have been widely discussed in the last decade (e.g. Risebrobakken et al., 2003; Mayewski et al., 2004; Andrews, 2006; Debret et al., 2007; Wanner and Bütikofer, 2008; Wanner et al., 2008; 2011) the spatial distribution of those millennial-scale climatic patterns and the origin of this pacing during the Holocene remains one of the outstanding mysteries of climate variability. There is therefore an urgent need to understand the nature, timing and causes of climate oscillations and to determine how widespread, systematic and abrupt they may have been.

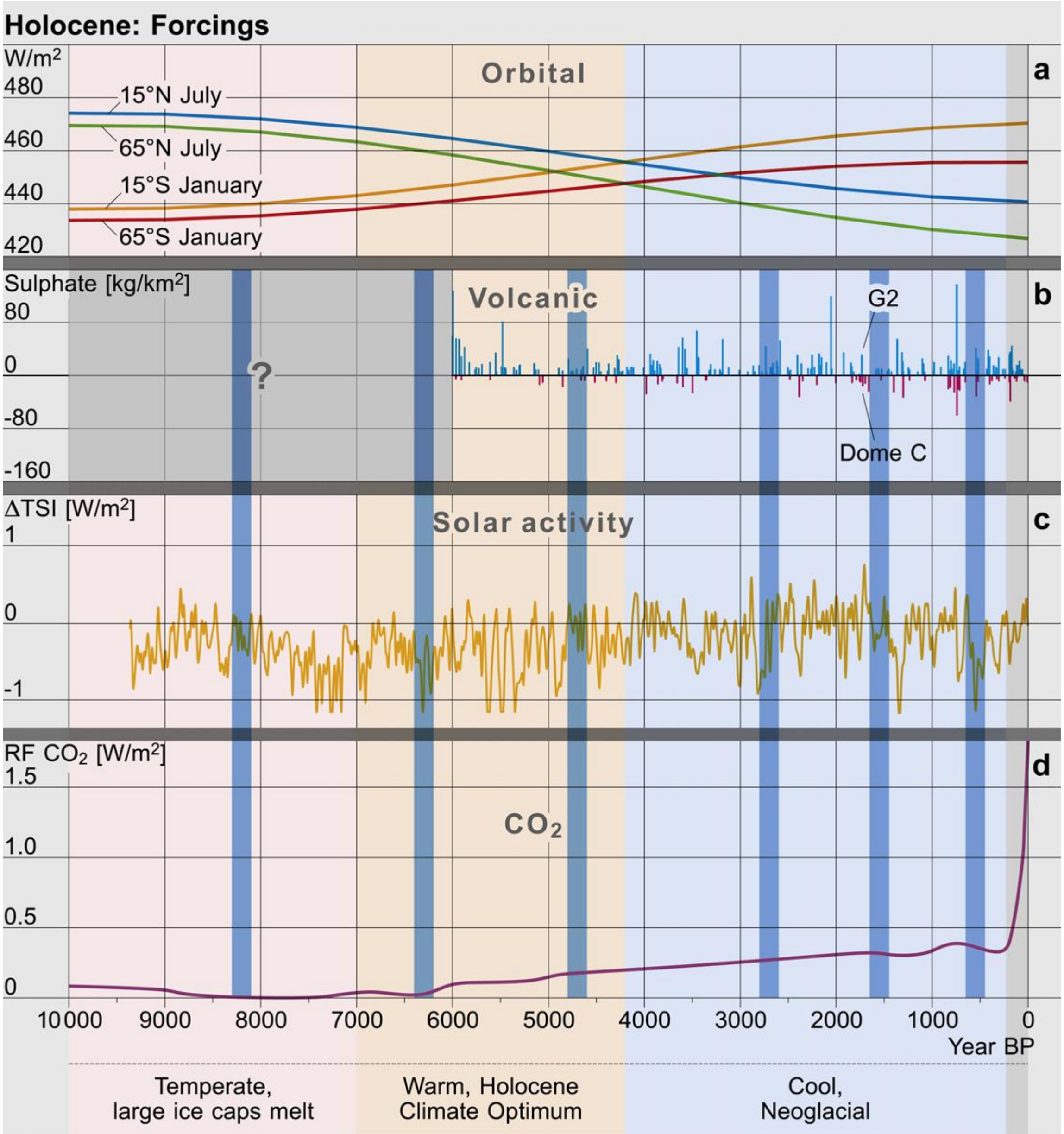


Fig. 8. Main external forcings driving Holocene climate changes (from Wanner et al., 2011). Holocene long-term trend: (a) Solar insolation due to orbital changes for two specific latitudes in the Northern and Southern Hemispheres during the corresponding summer and d) Forcing due to rising CO_2 concentrations. Holocene millennial scale climate changes: (b) Volcanic forcing during the past 6 ka depicted by the sulphate concentrations of two ice cores from Greenland (blue vertical bars) and Antarctica (red vertical bars); (c) Solar activity fluctuations reconstructed based on ^{10}Be measurements in polar ice. The six vertical blue bars indicate the timing, but not the length of the six cold periods during the last 10 ka in the NH.

2.3.2.1 Particular cases of well known Holocene short-lived climatic events

The most extreme Holocene short-lived cold episode occurred at around 8200 yr BP and is known as the 8.2 ka cooling event. The 8.2 ka event is characterized by a evident cooling of about 3.3 +/-1.1ºC during ~160 yr in Greenland ice core records (Fig. 9), coinciding with a reduction in ice accumulation rate, increasing wind speeds and a decline in atmospheric methane concentrations. (O'Brien et al., 1995; Alley et al., 1997; Muscheler et al., 2004; Kobashi et al., 2007; Thomas et al., 2007). This event was also detected in many other terrestrial (e.g. Klitgaard-Kristensen et al., 1998; Von Grafenstein et al., 1998; Magny et al., 2003) and marine records (e.g. Bond et al., 1997; 2001; Keigwin and Boyle, 2000; Alley and Ágústsdóttir, 2005 and references therein; Naughton et al., 2007b) of the NA region, as well as in areas influenced by monsoons suggesting the widespread signature of the abrupt 8.2 ka event (Alley and Ágústsdóttir, 2005; Rohling and Pälike, 2005; Fleitmann et al., 2007).

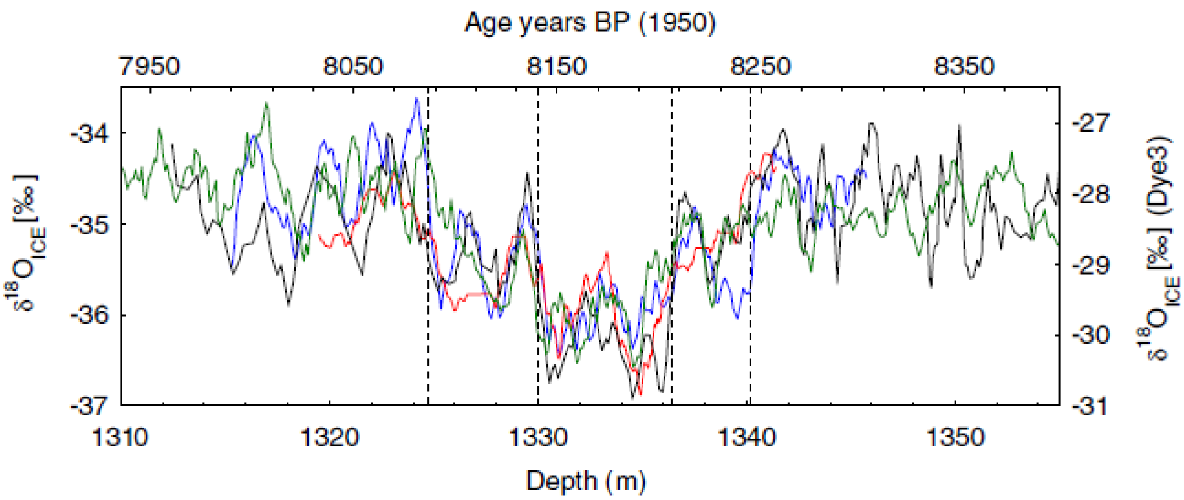


Fig. 9. Oxygen isotope ratios from GRIP (Greenland Ice Core Project) (red), GISP2 (black), NGRIP (blue), and Dye 3 (green) all plotted on the GRIP depth scale and the GICC05 age scale (from Thomas et al., 2007).

The causes triggering the 8.2 ka cold event have been largely debated in the last 15 years. Several authors suggest that "8.2 event" was triggered by the rapid collapse of the Hudson Bay dome of the Laurentide Ice sheet which result in a final catastrophic drainage of the Laurentide Lakes (Lake Agassiz and Ojibway) at around 8470 yr BP (Fig. 10) (Barber et al., 1999; Clarke et al., 2004). This phenomenon contributed to the introduction of large amounts of freshwater into the NA Ocean which disturbed the thermohaline circulation and the formation of the North Atlantic deep water (NADW)

(e.g. Alley et al., 1997; Clark et al., 2001; Clarke et al., 2004). The slowdown of the THC prevented the maintenance of the heat transport by the Gulf and NA currents to the high latitudes as showed by models simulations (Fawcett et al., 1997) generating a cooling and dryness of the NA and Europe (Broecker et al., 1985; 1990; Bond et al., 1997; Barber et al., 1999; Broecker., 2000; Cubasch et al., 2001; Renssen et al., 2001; Rind et al., 2001; Vellinga and Wood, 2002; Nesje et al., 2004; Wiersma and Renssen, 2006; Wiersma et al., 2006; Flesche Kleiven et al., 2008). This caused the weakening of the THC and associated northward oceanic heat flux, leading to a significant cooling in the NA (Fawcett et al., 1997; Rahmstorf, 2002). Additionally, these processes may have caused the spread of winter sea ice across the NA, thus causing the northern region to experience much colder winters which induced an increase in seasonality (Denton et al., 2005).

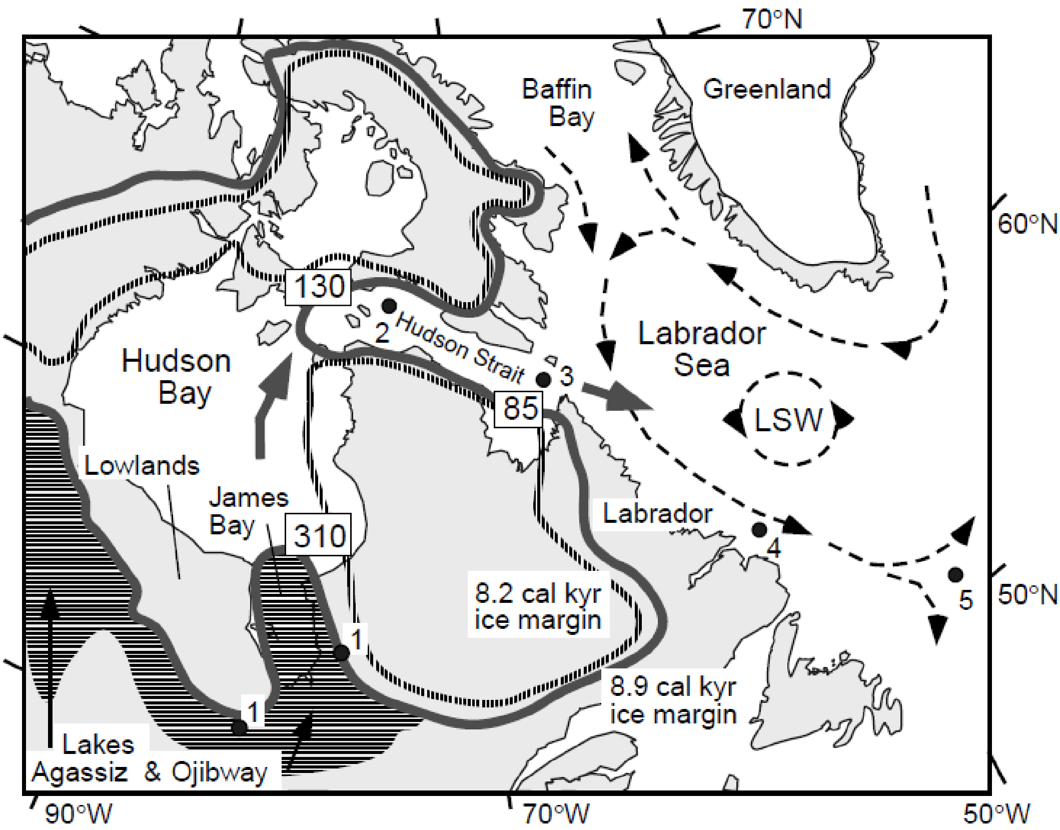


Fig.10. Configuration of the Northeast Canada and adjacent seas (from Barber et al., 1999). Former ice-sheet margins are shown for 8.9 ka and 8.2 ka prior to present time (vertical hatched line and thick grey line respectively), before and after disintegration of ice in Central Hudson Bay, respectively. Simultaneously, northward drainage is shown through the Hudson Bay and Hudson Strait (dark grey arrows). Horizontal hatching shows Lake Agassiz and Ojibway. Labrador Sea current patterns and the area of Labrador Sea Intermediate Water (LSW) formation is indicated by arrows with dashed lines show. Numbers in boxes are regional mean DR values (years). Sites discussed in Barber et al. (1999) are numbered from 1 to 4.

Other authors, however, suggested that this event was caused mainly by variations related to a reduction in solar irradiance (Denton and Karlen, 1973; Bond et al., 2001; Van Geel et al., 2003). These studies argue that changes in solar activity might have triggered the observed climate changes, most likely involving changes in ocean circulation. However, the fact that this event is more prominent in the NA region; that it follows outbursts flooding episodes; and that the existing similarities between reconstructed anomaly patterns and patterns expected following a NA freshening seems to favour the freshwater pulse mechanism as the major trigger for the 8.2 event (Alley and Ágústsdóttir, 2005).

Nevertheless, it is possible that the two hypothesis are compatible with recent publications (Rohling and Pälike, 2005; Ellison et al., 2006; Naughton et al., 2007b) that suggest the occurrence of the 8.2 ka event within a long climate cooling anomaly of multicentennial scale, between 8600 and 8000 years ago.

The millennial scale climate oscillations after the 8.2 ka event have not left a global "fingerprint" and there is no agreement on the amplitude and frequency of these oscillations. Nevertheless, although the amplitude of natural climate variations was weaker after the 8.2 event, they were strong enough to affect ecosystems and human societies (Mayewski et al., 2004), as documented for the last two millennia.

The last two millennia are marked by 5 climatic events in the NH: the Roman Warm Period (RWP) (~1–300 AD), the Dark Age Cold Period (DACP) (~300-800 AD), the Medieval Climate Anomaly (MCA) (AD 800–1300), the Little Ice Age (LIA) (1300-1900 AD) and the recent warming (RW) (Fig. 11) (Ljungqvist, 2010). The temperature anomaly reconstruction of Ljungqvist (2010) is relatively consistent, in terms of amplitude and timing of warm and cold periods, with those of: Lamb (1977), Moberg et al. (2005) and Mann et al. (2008), except for the DACP.

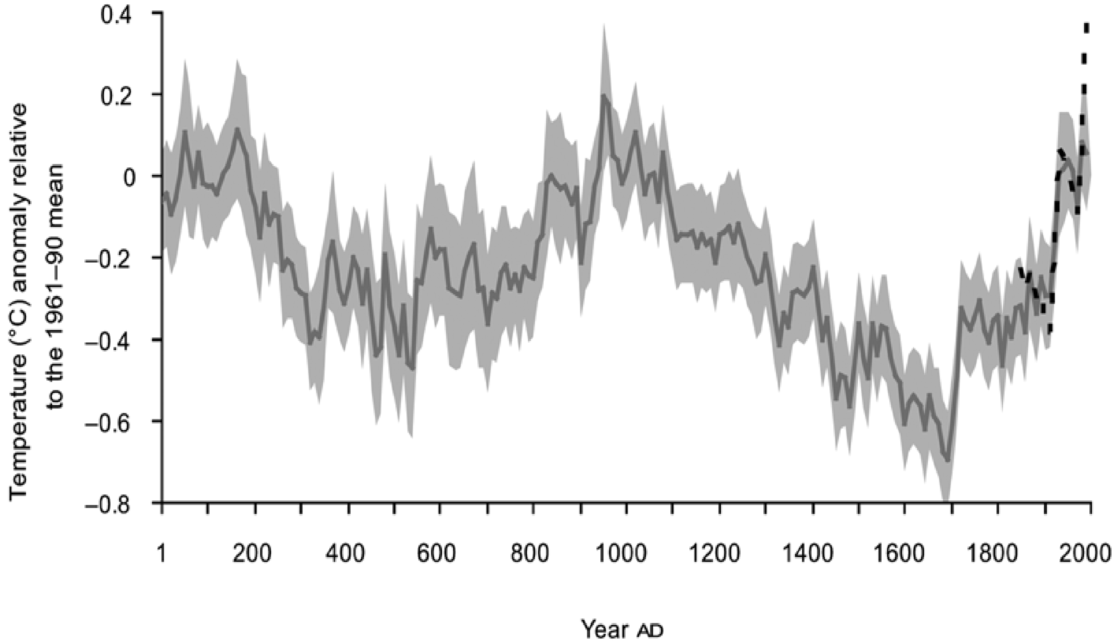


Fig.11. Extra-tropical NH (90–30°N) temperature anomaly reconstruction of the last 2000 years (from Ljungqvist, 2010).

The Roman Warm Period (RWP) was warm and wet (e.g. Martinez-Cortizas et al., 1999; Reale and Dirmeyer, 2000; Mcdermott et al., 2001; Diz et al., 2002; Desprat et al., 2003; Abrantes et al., 2005; Eiríksson et al., 2006; Lebreiro et al., 2006; Martin-Chivelet et al., 2011; Wanner and Brönnimann, 2012) at least in Europe and regions influenced by changes in the monsoon system (e.g. Ji et al., 2005; Liu et al 2006).

Several hypothesis have been invoked to explain such a warming as: high solar activity and weak volcanic activity (e.g. Steinhilber et al. 2009; Wanner and Brönnimann, 2012); increase of near-bottom water speed flow in the south Iceland basin (Bianchi and McCave, 1999). These conditions were also associated with the prevalence of the negative mode of the North Atlantic Oscillation (NAO-) (e.g. Diz et al., 2002; Lebreiro et al., 2006). The persistence of the NAO- pattern would be responsible for the increase of moisture over Europe. In Europe this period is marked by an important episode of deforestation as the result of Roman culture practices (Wanner and Brönnimann, 2012).

The Dark Age Cold Period was relatively cold and dry (Mcdermott et al., 2001; Desprat et al., 2003; Eiríksson et al., 2006 ; Lebreiro et al., 2006; Andrade et al., 2011; Martin-Chivelet et al., 2011; Wanner et al., 2011). The nature of this period is still

unclear, nevertheless the solar forcing curve shows a negative peak shortly after this event, and a larger volcanic activity started as well after AD 600 (Zielinski, 2000). Historically this period is marked by strong human migration in Europe (Wanner et al., 2011).

The Medieval Climatic Anomaly MCA), also called Medieval Warm Period (MWP), marks the return to relatively warm but relatively dry conditions over southern Europe (Diz et al., 2002; Desprat et al., 2003; Álvarez et al., 2005; Lebreiro et al., 2006; Abrantes et al., 2011, Andrade et al., 2011; Martin-Chivelet et al., 2011). The northern Europe was also warmer but wet (e.g. Trouet et al., 2009). In contrast the tropical pacific was cold (Trouet et al., 2009). This globally complex climatic pattern can be explained, particularly over the Atlantic/European region, by the prevalence of the positive mode of NAO during the MCA (Trouet et al., 2009).

The transition from MCA to the subsequent Little Ice age (LIA) is one of the most debated questions in Late Holocene climate variability; however there is evidence that it was due to a shift from prevailing positive to a most negative mode of the NAO (Hughes and Diaz, 1994; Bradley et al., 2003; Trouet et al., 2009; Mann et al., 2009).

In the NH, LIA was probably the coldest multidecadal to multicentury long Holocene period since the 8.2 ka event (e.g. Kaufman et al., 2009). These cold conditions were probably triggered by a series of Grand Solar Minima and high number of strong tropical volcanic eruptions (see Fig. 8) (Renssen et al., 2007; Wanner et al., 2011). The LIA has however a complex climatic signature in the Iberian margin and adjacent landmasses. Thus, most records from NW Iberia Peninsula and margin show a sea surface and terrestrial cooling, a general continental dryness and enhanced upwelling regime during this period associated with a prevailing NAO+ mode (Diz et al., 2002; Desprat et al., 2003; Álvarez et al., 2005; Lebreiro et al., 2006; Abrantes et al., 2011, Andrade et al., 2011; Martin-Chivelet et al., 2011) while the Tagus Prodelta is characterized by storminess and excess in precipitation and river discharges associated with prevailing NAO- mode (Abrantes et al., 2005b; Eiríksson et al., 2006; Lebreiro et al., 2006). Interestingly, this apparent contradictory links with the prevailing NAO mode during the LIA are not restricted to the Iberian Peninsula and reflect a much wider potential inconsistency described in detail by Trouet et al. (2012).

The unfavorable climate conditions during the LIA had an extensive impact in human societies in Europe, e.g. decline of health and economic wealth of medieval societies triggered by the Black Death, the abandonment of the Norse settlements in Greenland or the abandonment of permanent settlements in the Alps (Pfister, 1995; Fagan, 2000; Grove, 2004; Patterson et al., 2010: Bűntgen et al., 2011).

Despite the considerable progresses that have been made in recent years in late-Holocene palaeoclimatology (for a review, see Wanner et al. (2008), IPCC (2007) report and Jones et al. (2009)), the world-wide nature, timing and amplitude of these events are not completely understood yet.

3. MATERIAL AND ENVIRONMENTAL SETTING

3.1 Sediment core

To attain the objectives of this master a sediment core was retrieved in a selected key site from the continental shelf of the Basque Country (southwestern Europe) (Fig. 12; Table 2). This geographical region is particularly sensitive to climate changes, since it is directly affected by hydrological changes of the North Atlantic Drift and by a temperate Atlantic climate regime in the adjacent landmasses.

Core KS05 10 was retrieved from the shelf mud patch of southwestern margin of the Bay of Biscay (Basque country) in 2005, during the cruise EUSKA 3 on board N/O Côtes de la Manche.

This sequence is characterized by relatively homogeneous silty-clay sediments (Fig. 13) and the chronological framework indicates that covers the last ca. 9000 yr. The sedimentation rate varies between 0.01 and 0.05 cm/yr allowing a centennial time-scale resolution.

Core Ref.	**KS05 10 core**
Longitude (W)	2°16’744W
Latitude (N)	43°22’765N
Water depth (m)	114
Core lenth (m)	2,5
Ship	N/O Côtes de la Manche
Cruise	EUSKA 3
Sampler	Gravity corer

Table 2. Main features of the cored site.

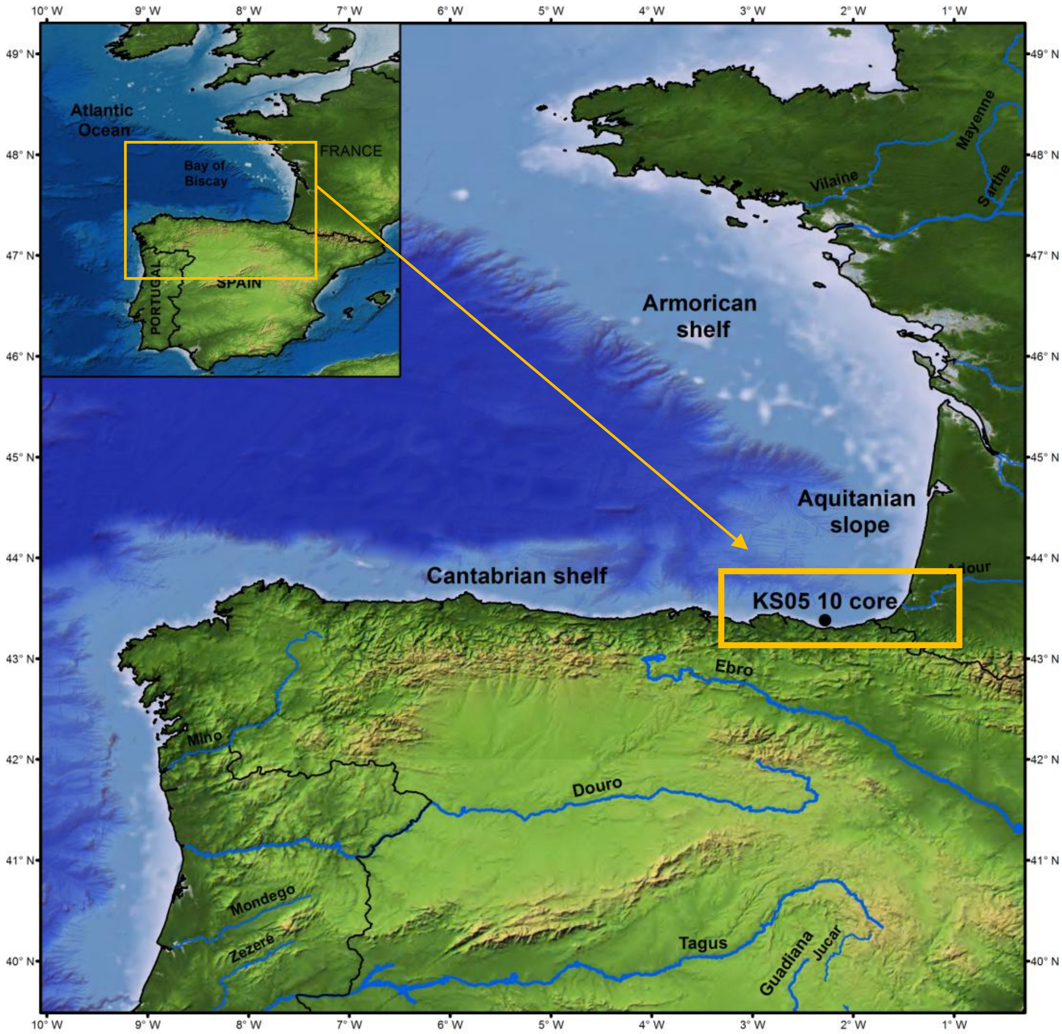


Fig. 12. Map showing the location of the studied core. Sources: The bathymetry is derived from the Digital Bathymetry as produced in the EMODNet Hydrography (http://www.emodnet-hydrography.eu). The elevation data is derived from SRTM (Shuttle Radar Topography Mission) 90m Digital Elevation Database v4.1 (Farr et al., 2007). The drainage system is derived from the Europe and North Asia (EURNASIA) Vmap Level Zero (VMAP0 - Digital Chart of the World) (http://webgis.wr.usgs.gov/globalgis/metadata_qr/metadata/perennial_rivers.htm).

Top

Base

Fig. 13. Photos of KS05 10 core.

3.2 Environmental Setting

The selected key-site is strongly influenced by the complex interactions between the ocean, atmosphere, and cryosphere of the NA realm. Accordingly, the following section presents a brief review of the main features of the ocean and atmospheric circulation of the NA, as well as the present day climate and vegetation of the studied region.

3.2.1 Ocean and Atmospheric Circulation

The NA is one of the few regions in the world where deep water formation occurs and is characterized by a continuous poleward flow of warm and salty surface waters that constitutes the upper part of the Meridional Overturning Circulation (MOC) (Fig. 14). The THC currents carry water, that has been warmed in the tropics, to the NA Ocean by means of the Gulf Stream and North Atlantic Current (NAC), causing a large influence in the climate by the release of both latent and sensible heat to the atmosphere. When surface waters reach high latitudes, they have cooled and augmented in salinity, leading to an increase in their density. The cold and salty surface waters sinks to the deep ocean to form the oxygen-rich and nutrient-poor NADW that is transported southwards in the deep ocean. Together with the wind driven surface currents, this circulation is collectively referred to as the MOC (Rahmstorf, 2007).

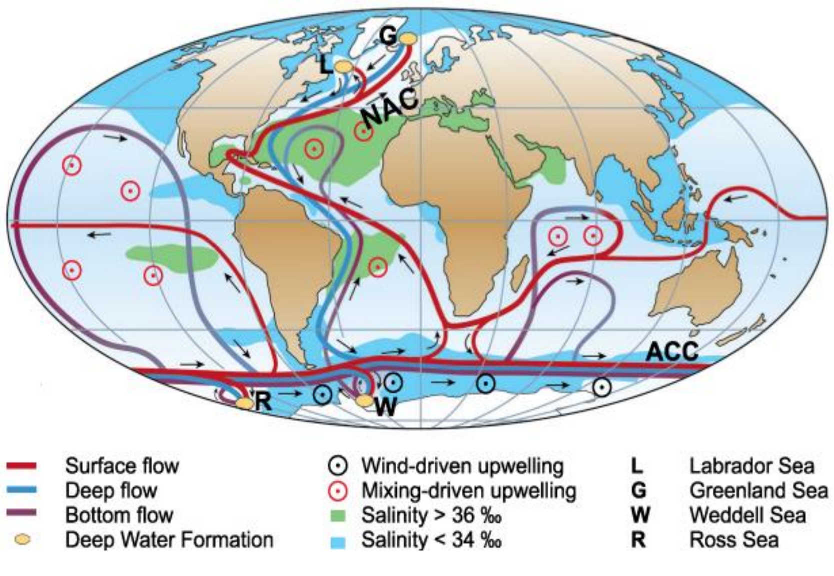


Fig. 14. Schematic representation of the Meridional overturning circulation (from Rahmstorf, 2007). Global circulation system in the world ocean that shows the northward transport of warm and salty surface waters in the NA and the return flow to the south of cold and dense NADW in the abyssal ocean. ACC corresponds to the Antarctic Circumpolar Current.

The main characteristics of the water circulation in the Bay of Biscay includes a weak water circulation, along with the frequent presence of eddies (Koutsikopoulos and Le Cann, 1996). In terms of water masses (Fig. 15), below the mixed layer flows the Eastern North Atlantic Central Water (ENACW), which is originated by convection during the winter on a broad region from northeast Azores to the European margin in the area. This central water is bounded at north by a branch of the NAC that enters in the Bay of Biscay from northwest favoring an anticyclonic circulation, and the Azores Current (AC) that flows southward (Pollard and Pu, 1985; Pollard et al., 1996). The intermediate layer, filled by the northward bound Mediterranean Water (MW), is located just below the ENACW (Pollard et al., 1996). In the deeper layer, below 1500 m, the prevailing deep-water masses are controlled by the termohaline equilibrium between the NADW and the carbonate corrosive Antarctic Bottom Water (AABW).

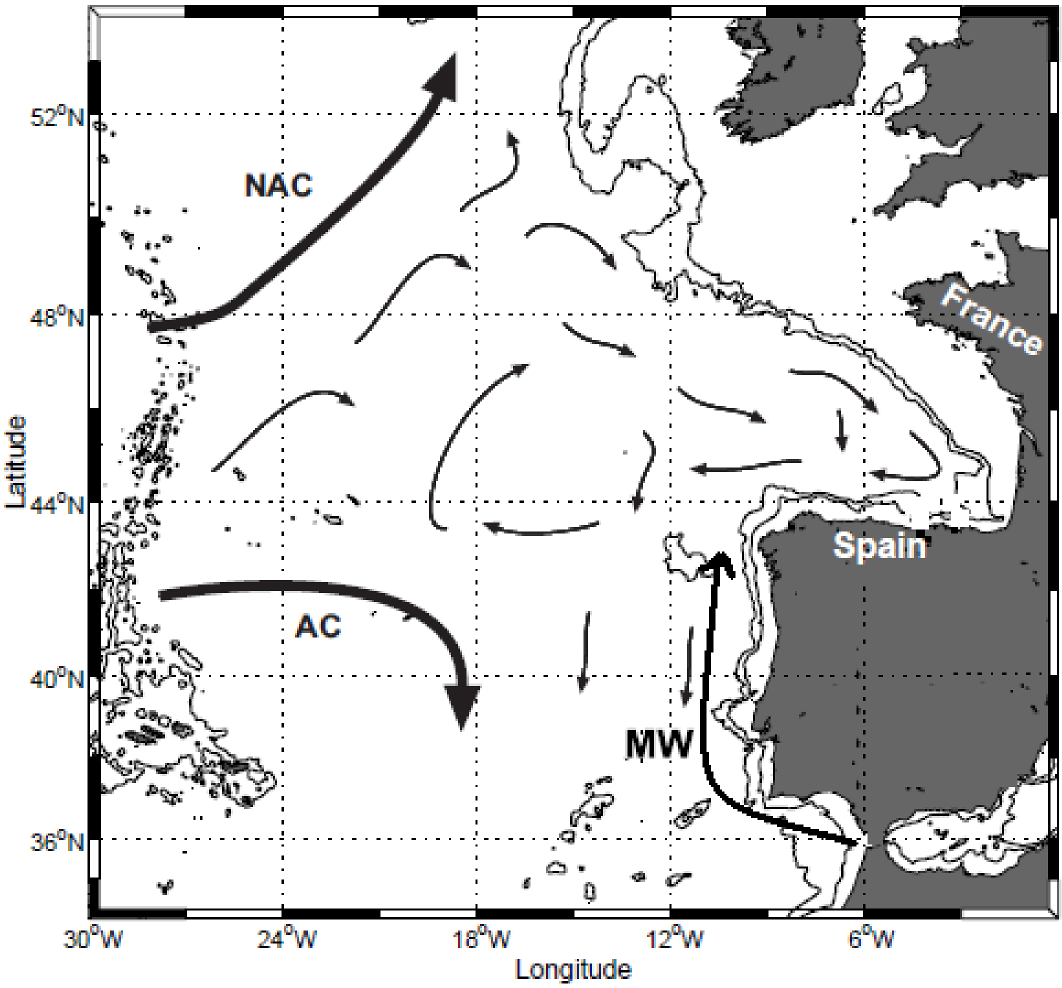


Fig. 15. Composition of ENACW circulation in the Bay of Biscay and the nearest part of the Atlantic at the level of ENACW (adapted from González-Pola et al., 2005).

The vegetation changes reflected in the KS05 10 record are associated to environmental and climatic conditions registered in the closest landmasses, i.e. the area corresponding to modern day northern Spain and southern France. This region is particularly sensitive to interannual shifts in the trajectories of mid-latitude cyclones that can lead to remarkable anomalies of precipitation and, to a lesser extent, of temperature (Trigo et al., 2008). This area is influenced by a relatively small number of large-scale modes of atmospheric circulation in the Northern Hemisphere that have been described previously in literature (Wallace and Gutzler, 1981; Barnston and Livezey, 1987), namely: a) the North Atlantic Oscillation (NAO), b) the Eastern Atlantic (EA) pattern and c) the Scandinavian pattern (SCAND) also known as Blocking pattern. In recent years most climate variability publications use the teleconnection indices derived and maintained by the Climate Prediction Center (CPC) from NOAA (http://www.cpc.ncep.noaa.gov/data/teledoc/telecontents.shtml). Their methodology is based on Rotated Principal Component Analysis applied to monthly mean standardized 500-mb geopotential height anomalies (Barnston and Livezey, 1987).

The spatial signature of the three main large-scale patterns affecting the Euro-Atlantic region for the period that spans between 1960 and 2000 can be observed in Fig. 16, representing respectively the NAO (top), EA (middle) and SCAND (bottom) patterns. In fact, Fig. 16 represents the temporal correlation between the monthly standardized height anomalies at each point and the monthly teleconnection pattern time series from 1960 to 2000. It should be always kept in mind that these modes are seasonally dependent, i.e. their signature can vary throughout the year. The seasonal dependence of these modes and the impact on surface climate (temperature and precipitation) can be appreciated in the CPC site as well.

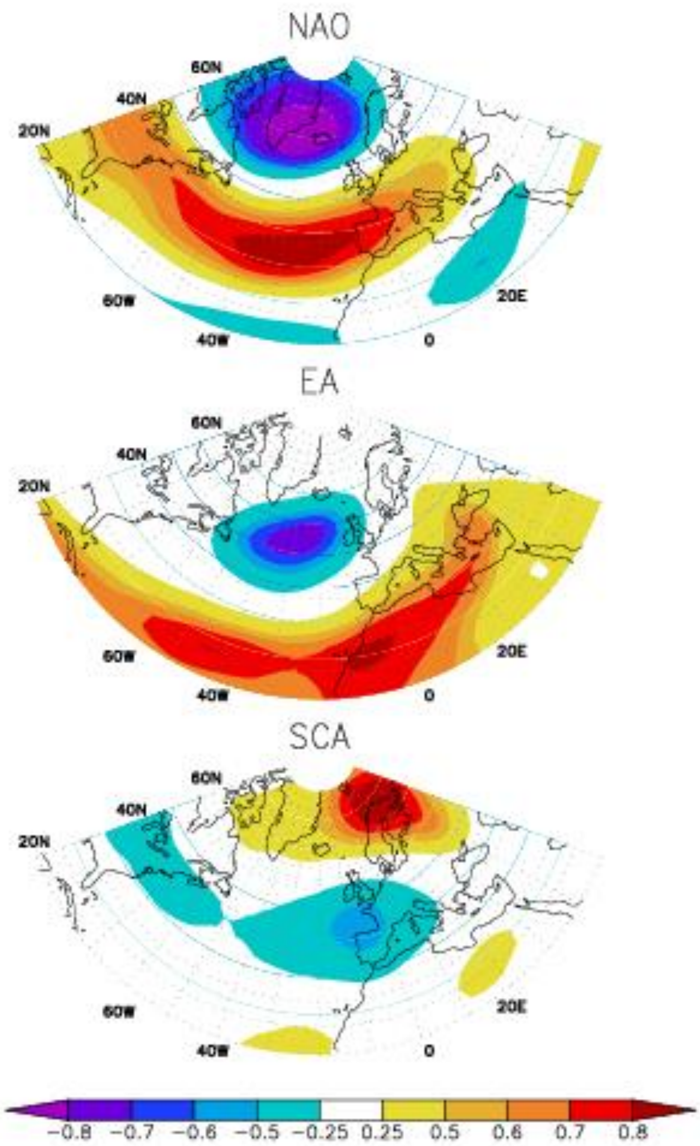
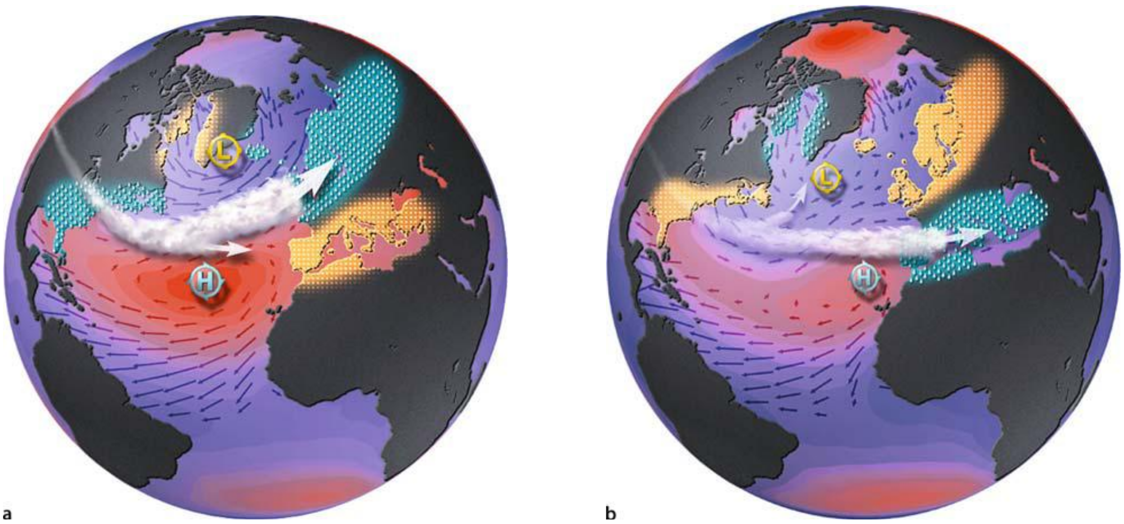


Fig. 16. Based on the method described by Barnston and Livezey (1987) it is possible to compute the spatial patterns associated to the teleconnection indices NAO, EA, and SCAND for DJF and for the period 1960-2000 (from Trigo et al., 2008).

It is widely accepted that NAO corresponds to the dominant mode of atmospheric circulation across the NH especially over the NA region, representing roughly 40% of sea level pressure variability of the area (Fig. 16) dictating to a large extent the climate of the studied region, particularly during winter time (e.g. Trigo et al., 2002). This mode is associated with the strength of the meridional pressure gradient along the North Atlantic sector. Some authors consider the NAO mode as a regional manifestation of the hemispheric Arctic Oscillation (AO, Thompson and Wallace, 1998). Usually the NAO Index is determined by the difference between sea level pressure measured over Iceland

and Ponta Delgada (Azores), although in recent years other stations have been proposed for the southern sector of the pattern, namely Lisbon (Hurrell, 1995) or Gibraltar (Jones et al., 1997). When the NAO is in a positive phase, the Icelandic low pressure center and high-pressure center at Azores are both enhanced causing a large wintertime meridional pressure gradient over the NA (Fig. 17) (Hurrell, 1995). In the Negative NAO phase both centers are weakened, i.e. weak subtropical high and a weak Icelandic low (Hurrell, 1995).

Oscillations from one phase to another of the NAO pattern causes variations in mean wind speed and direction over the NA, seasonal mean heat and moisture transport over the ocean, the path and number of storms, and can be responsible for important changes in ocean temperature, and heat content, current patterns, and sea ice cover in the Arctic (Trigo et al., 2002; Hurrell, et al., 2003). For example, a positive phase is associated with more frequent and stronger winter storms across the mid-latitudes of the Atlantic onto Europe, generating an intense upwelling along the western Portuguese Margin (Abrantes et al., 2005). In the opposite situation, the reduced pressure gradient induces southerly/westerly winds during cold periods and occurs an increase in precipitation over western Iberia (e.g. Hurrell et al., 2001; Trigo and DaCamara, 2000; Trigo et al., 2004a).



a b

Fig. 17. The two phases of the North Atlantic Oscillation variability mode. a) Positive phase, stronger than average westerlies and mild and wet winters over N-Europe and dry conditions in the Mediterranean; b) Negative phase, leading to the opposite conditions such as weaker than average westerlies, cold, dry winters in N-Europe and rainy winters in S-Europe (from http://www.ldeo.columbia.edu/NAO by Martin Visbeck).

However, the climate of the region where KS05 10 record is located (southwestern France and Basque country) is particularly complex not being well related with a unique mode of variability (Sáenz et al., 2001a). In particular it is possible to state that this region is less influenced by the NAO pattern (Trigo et al., 2004a) being more dependent on the EA and SCAND modes (Xoplaki, 2002; Trigo et al., 2006).

The EA pattern is structurally comparable to the NAO, and consists of a north-south dipole of anomaly centers spanning the NA from east to west; however the SCAND pattern explains a higher fraction of European precipitation variability than the EA mode (Trigo et al., 2008). The SCAND pattern consists of a primary circulation center over the Scandinavia Peninsula and an additional weaker center with opposite sign located over southwestern Europe (Fig. 16) being responsible for a large fraction of precipitation registered over central and western Mediterranean basin (Quadrelli et al., 2001; Xoplaki, 2002; Trigo et al., 2008) as depicted in Fig. 18. It should be emphasize that the SCAND circulation mode corresponds to the typical configuration of the European blocking pattern, usually described in studies using sub-monthly scales (e.g. Tibaldi et al., 1997). Blocking episodes are known to produce significant impacts on both the precipitation and temperature fields of the Mediterranean Region (Trigo et al., 2004b).

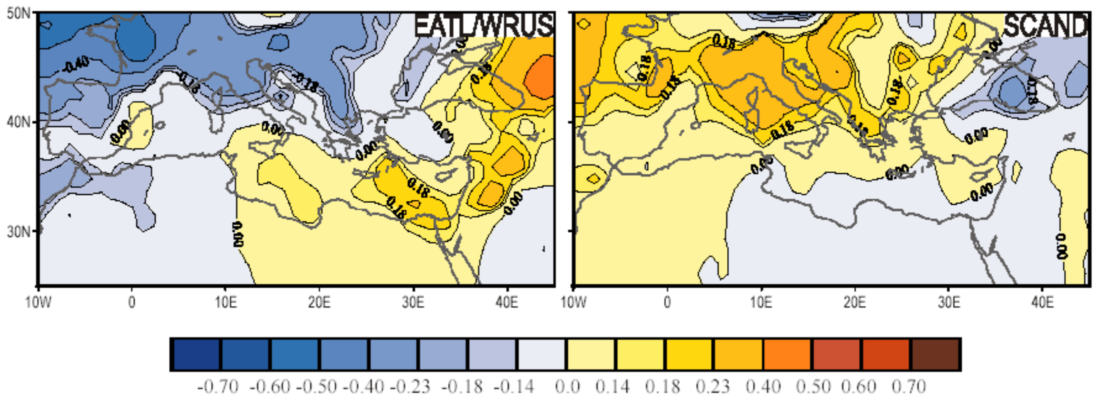


Fig. 18. Spatial correlation between the patterns EA/WRUS and SCAND patterns and winter (NDJF) Mediterranean station precipitation for the period 1950-1999. Correlations |r|≥0.14 indicate significance at the 95% level, |r|≥0.18 at the 99% level and |r|≥0.23 at the 99.9% level (n = 4x50 = 200 months), respectively (adapted from Trigo et al., 2006).

3.2.1 Present-day vegetation

The Iberian Peninsula encompasses two macrobioclimatic and biogeographic areas: the Temperate macrobioclimatic zone mostly in the north, which corresponds to the Eurosiberean biogeographic region, and where is located the studied site; and the Mediterranean macrobioclimatic/biogeographic area, that includes a large area of the centre and south of the peninsula (Rivas-Martínez et al., 2004) (Fig.19).

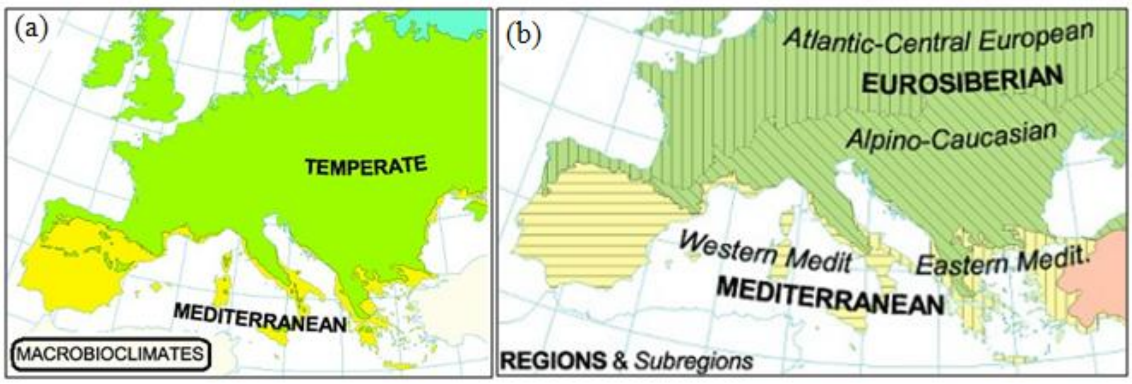


Fig. 19. Iberian Peninsula Bioclimatic (a) and Biogeographic map (b) (from Worldwide Bioclimatic Classification System, 1996-2009). The Iberian Peninsula show marked differences between the two bioclimatic zones. The north and the northwest of the Peninsula are characterized by a Temperate bioclimate, with colder temperatures and higher precipitation than that of the Mediterranean bioclimatic. The small patches of Temperate bioclimate enclosed in the Mediterranean type correspond to large Mountain ranges in the centre and north of Iberia.

The Basque Country, which is located within the middle latitudes of the northeastern Atlantic, is subjected to the influence of the Gulf Stream and the atmospheric westerlies in the middle and upper troposphere, presenting as a consequence an annual mean temperature higher than 10ºC (Usabiaga et al., 2004). The climate is temperate, oceanic, with moderate winters and warm summers (Usabiaga et al., 2004). Also, it is wet, with over 1,500 mm of rainfall each year (Usabiaga et al., 2004), exhibiting a more variable distribution of precipitation than of temperature (Sáenz et al., 2001a, b). According to Koppen's classification, the studied region is associated with a Cfb climate (marine west coast-mild). As aforesaid above, this climate is also under the influence of large-scale modes of atmospheric circulation (NAO, EA and SCAND).

The climate in the studied region allows the development of Eurosiberian vegetation type. The Eurosiberian region is noted for the abundance of deciduous oak forests (*Quercus robur*, *Q. petraea*, *Q. ilex*), heath communities (Ericaceae and *Calluna*) and *Ulex*. There are also locally birch (*Betula pubescens subsp. celtiberica*) and hazel (*Corylus avellana*) groves, and brooms (*Genista*) (Alcara Ariza et al., 1987). The presence of some species, such as *Juglans, Castanea, Cerealia*, *Eucalyptus, Vitis* and pine (*P. sylvestris* and *P. pinaster*), reflect human activities (e.g. agriculture, ruderalization or arboriculture) (Carrion et al., 2010 and references therein).

4. POLLEN AS A PROXY FOR PALEOCLIMATIC RECONSTRUCTION

Paleoclimatology main target is to describe and understand earth's climatic history through as many time scales as possible. Part of this aim can be achieved with climate proxy indicators enclosed in the deep-sea sediments sequences.

A proxy climate indicator is defined as a local record that is interpreted, using physical and biophysical principles, to represent some combination of climate-related variations back in time (IPCC, 2001). A single deep-sea sediment core can encompass several proxies (e.g. stable isotopes; trace elements; microfossil assemblages, magnetic and physical properties), that provide a reliable and extensive climatic reconstructions (Cronin, 1999).

Pollen grains included in sediments are considered the most important tool to understand vegetation responses to the Holocene climate variability, being therefore used in the present study in the northern Iberian Peninsula and southwestern France.

4.1 Basic principles of pollen analysis

Palynology is one of the most useful tools in Quaternary paleoenvironmental and paleoclimatology studies. This branch of science is concerned with the study of pollen grains (produced by seed plants, angiosperms and gymnosperms) and spores (produced by pteridophytes, bryophytes, algae and fungi) (Moore et al., 1991). Although Palynology refers to the study of both groups, these will be referred together as 'pollen' for the sake of brevity.

Faegri and Iverson (1989) describe pollen analysis as "a technique for reconstructing former vegetation by means of the pollen grains it produced". Since Palynology embraces the uniformitarian principle, *the present is the key to the past*, it is reliable to use the relationship between modern pollen distribution and climate as a "guide" to understand pollen patterns recorded in the past and make paleoclimatic reconstructions.

The vegetation reconstruction from fossil pollen spectra are based in the following general principles of pollen analysis (Birks and Birks, 1980):

- Pollen grains are abundantly produced during the natural reproductive cycle of plants;

- In the atmosphere pollen grains are mixed by atmospheric mechanisms, causing a uniform pollen rain over a certain area;
- Pollen grains are preserved in anaerobic environments (e.g. bogs, marshes, lakes, fens, ocean floor) and reflect the natural vegetation at the time of pollen deposition. Consequently, the composition of the pollen rain is a function of the vegetation composition;
- Fossil grains can be extracted from sediments and identification is possible at the level of genus or family, and may be achievable to species level;
- Pollen spectra obtained through a sediment sequence provide a picture of vegetation variability over time, which can yield information about past climatic conditions.

4.2 Pollen grain features

Pollen grains and spores differ in their function, however both result from cell division involving a reduction by half of the chromosomes content (meiosis) and need to be dispersed in order to carry out their functions (Moore et al., 1991).

Spores are reproductive structures that contain the necessary genetic material for dispersal and growth of plants. On the other hand, pollen grains are unicellular and microscopic (15–100 μm) organs that contain the male genetic material of the angiosperms and gymnosperms; sexual reproductive success is assured only if this material reaches a female receptacle of the same plant species.

The dispersal of pollen grains to the female reproductive structure – pollination – can occur in different ways depending on the plant taxa. Although most pollen is either wind-dispersed (anemophilous) or insect-dispersed (entomophilous) (Faegri and van der Pijl, 1979), there are others agents of pollination such as water (hidrophily) and vertebrates (zoophily).

Both types of grains comprehend an outer layer, the exine, made of a mixture of cellulose and a complex polymer called sporopollenin. Sporopollenin is resistant to most forms of chemical (including enzymatic) and physical degradation, except oxidation (Faegri and Iversen, 1989). As a result, pollen may be preserved in anaerobic environments and strong chemicals can be used to remove other components in laboratory.

Sculptural elements are developed in the exine, which is divided into two sublayers, the inner is the endexine and the outer is the ektexine (Fig. 20) (Faegri, 1956; Punt et al., 2007). While the ektexine consists of a basal foot layer with projecting columellae, that may be free distally (intectate) or partially connected by a tectum (semitectate); the endexine is an unstructured layer (Hesse et al., 2009). The interior wall that surrounds the cytoplasm, identified as intine, is largely composed of cellulose and other substances that are easily destroyed and thus hardly fossilized.

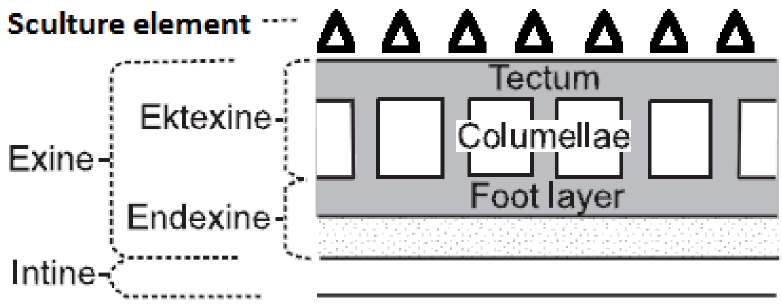


Fig. 20. Details of angiosperm pollen wall structure as defined by Faegri (1956) (adapted from Heusser, 2005).

Pollen grains have different morphology depending on the taxa; therefore they can be distinguished by their size, shape, apertures, surface sculpture and wall structure. The surface sculpture, wall structure and apertures are considered the most important features for pollen identification.

An aperture is a region of the pollen that is thinner than the remainder of the sporoderm and generally differs in ornamentation and/or in structure (Erdtman, 1947). There are two types of apertures, the pores (isodiametric apertures) and colpi (elongate or furrow aperture).

Fig. 21 shows the several groups of pollen according to the number, type and position of apertures.

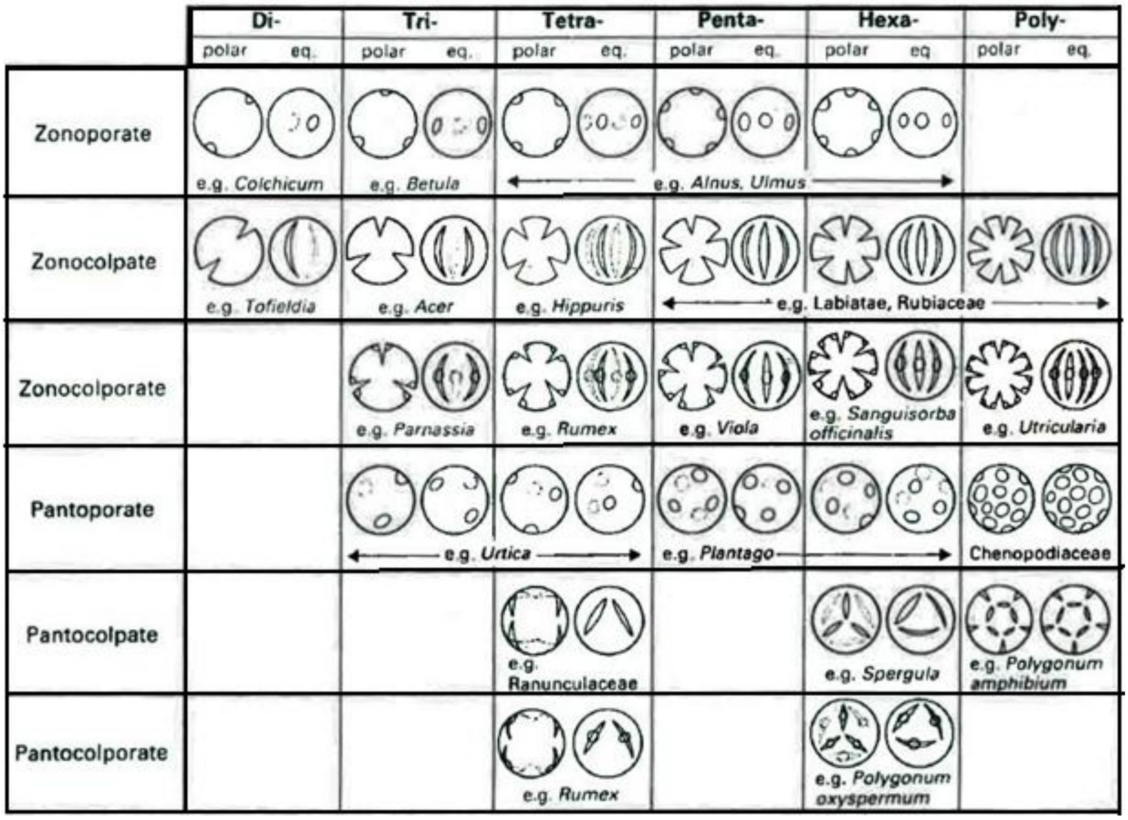


Fig. 21. Groups of pollen according to the number, type and position of apertures (from Moore et al., 1991). Examples are shown in equatorial view (view of a pollen grain where the equatorial plane is directed towards the observer) and polar view (the polar axis is directed towards the observer).

The structure (internal construction of the pollen wall) and ornamentation of the exine are also extremely important in pollen identification. LO analysis is used to detect patterns of exine organization with a light microscopy; "L" means lux/light and "O" means obscuritas/darkness (Erdtman, 1952; Punt et al., 2007). A variation of the focus allows the understanding of all the sculptural and structural elements of a pollen grain; at high focus raised exine elements appear bright and at low focus they become dark. Fig. 22 represents the main sculpture and ornamentations forms that can be found in Palynology: psilate, verrucate, echinate, striate, reticulate, fossulate, baculate and clavate.

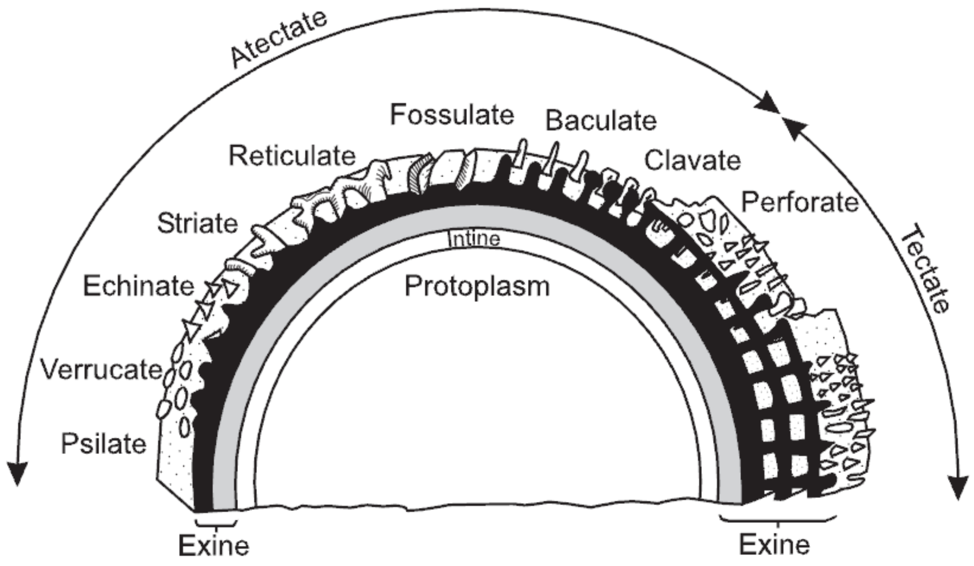


Fig. 22. Wall structure and ornamentation of angiosperm pollen (from Heusser, 2005).

4.3 Marine Palynology

The terrestrial pollen records are normally discontinuous, often restricted to relatively short periods of time and their pollen content generally represent local rather than an integrated image of the regional vegetation (Groot and Groot, 1966; Manten, 1966; Cooke, 1979). Furthermore, different chronostratigraphic tools are used to date marine and continental sequences preventing a good and direct correlation between both environments (Balsam and Heusser, 1976). In contrast, pollen included in marine sediments provides an integrated image of the regional vegetation of the close continent over a large span of time (e.g. Manten, 1966; Stanley, 1969; Heusser and Shackleton, 1979; Heusser and van de Geer, 1994) and can be directly correlated with the marine proxies in the same stratigraphic level of a marine sequence (e.g. Heusser and Balsam, 1977; Turon, 1984; Aksu and Mudie, 1985; Hooghiemstra et al., 1992; 2006; Naughton et al., 2007a; Combourieu-Nebout et al., 2009).

Therefore, pollen grains preserved in marine sediments is the most important tool to reconstruct past vegetation and climate in both terrestrial and marine environements (e.g. Heusser and Balsam, 1977; Turon, 1984; Aksu and Mudie, 1985; Hooghiemstra et al., 1992; 2006; Naughton et al., 2007a; Combourieu-Nebout et al., 2009).

There are different kinds of pollen suppliers to the sea depending essentially on the environmental conditions of each place (e.g. Groot and Groot, 1966; Dupont et al., 2000; Hooghiemstra et al., 2006). Arid zones with small hydrological systems, like NW of Africa, evidences for a mainly windy role in the pollen transfer to the marine environments (e.g. Hooghiemstra et al., 2006). Other zones like Gulf of Guinea (Lézine and Vergnaud-Grazzini, 1993) and Alboran Sea (Moreno et al., 2002) have shown a mixture of stream and windy pollen transport. In coastal zones with complex fluvial systems, pollen grains are principally transported by rivers and streams to the sea (e.g. Muller 1959; Bottema and Van Straaten, 1966; Heusser and Balsam, 1977; Naughton et al., 2007a). The southwestern French/Northern Spain margin is affected by north westerly prevailing winds which impede direct airborne transport of pollen seaward. This region is also drained by several rivers suggesting that pollen grains are probably transported to the studied region essentially by rivers (see Chapter 2; section 2. Environmental Setting, Present-day pollen deposition for more detail).

Once the pollen grains have entered the water column, they are transported and incorporated in the surface sediments of the ocean floor, behaving similarly to fine sedimentary particles (Muller, 1959; Chmura and Eisma, 1995). Nevertheless, their quantity and preservation in marine sequences can be influenced by several factors related to the marine environment (e.g. marine currents, oxygen content and temperature of water, sedimentary rate) and to pollen features (primary pollen productions, type of pollen transporters) (Bottema and Van Straaten, 1966). However, pollen assemblages in marine cores are usually similar in their general trends to the pollen records of terrestrial deposits allowing a good correlation between both pollens signatures (Heusser and Florer, 1973; Naughton et al., 2007a).

5. METHODOLOGY

5.1 Chronology

Establishing accurate dating is essential for paleoclimatic studies. Without reliable chronology it is impossible to investigate the climatic response driven by external and internal forcings; neither is it conceivable to compare our sequences with other marine and continental records.

Radiocarbon dating is considered the most useful dating method in paleoclimatic research of the late Quaternary. Radiocarbon (^{14}C) is produced in the Earth's atmosphere by neutron bombardment of atmospheric nitrogen atoms (^{14}N) and is used by vegetable and animal life forms; as Libby (1955) mentioned "all living things will be rendered radioactive by the cosmic radiation". The death of the organisms closes off the carbon exchange with the atmosphere and starts the ^{14}C decay "clock".

Considering that the atmospheric ^{14}C levels have remained constant during the period useful for ^{14}C dating (Libby, 1952), the amount of radioactive carbon left in the sample gives a reproducible indication of how old it is. However, several studies show that variations in the production of ^{14}C in the atmosphere occurred over time (e.g. de Vries, 1958; Linick et al., 1986; Stuiver et al., 1991). These variations can be caused by several factors, such as: changes in the solar magnetic field (de Vries, 1958, 1959; Stuiver, 1961; Stuiver and Quay, 1980); variations in solar activity (Goslar et al., 2000) and carbon cycle changes tied to deep ocean circulation (Edwards et al., 1993; Hughen et al., 2000). As a result, accurate calibration of conventional radiocarbon ages (yr BP) to calendar years (cal. yr BP) is indispensable.

Several standard calibration curves are available, based on independently dated samples such as tree rings, ice-core annual layers, varved sediments, speleothems, corals and historical records. In the present study it was used the international calibration curve IntCal09 (NH atmospheric; Reimer et al., 2009) and Marine09 (for marine dates; Reimer et al., 2009), which can be found in the CALIB Radiocarbon Calibration software Version 6.0 (http://calib.qub.ac.uk).

Radiocarbon dating using an accelerator coupled to a mass spectrometer (AMS dating) is the most widely used dating technique, allowing the dating of very small organic samples (e.g. Muller, 1977; Fairbanks et al., 2005). Recent advances AMS and sample preparation techniques have reduced the sample-size requirements by a factor of

1000 and decreased the measurement time from weeks to minutes (Fairbanks et al., 2005).

In the present study, the age-depth model is based on eight accelerator mass spectrometry (AMS) radiocarbon dates obtained on *Turritela communis* and other mollusk shells and in one historically well-dated botanical event (see Chapter 2; section 3. Material and Methods, Chronology).

AMS ^{14}C ages are calibrated using Calib 6.0 software and the Marine09 dataset (Marine09.^{14}C) (Stuiver and Reimer, 1993; Hughen et al., 2004; Stuiver et al., 2005; Reimer et al., 2009). This dataset represents the "global" ocean, incorporating a time-dependent global ocean reservoir correction of about 400 years (Stuiver et al., 2005; Reimer et al., 2009). To accommodate local effects, the difference ΔR in reservoir age of the studied region is incorporated (Stuiver et al., 2005). The statistical significance of results is based on the use of 95.4% (2 sigma) confidence intervals and their relative areas under the probability curve as well as the median probability of the probability distribution (Telford et al., 2004), as suggested by Stuiver et al. (2005).

5.2 Pollen analysis procedures

5.2.1 Pollen concentration technique

The sample preparation technique followed the procedure described by de Vernal et al. (1996) and improved at the UMR CNRS 5805 EPOC (Unité mixte de Recherche 5805, Centre National de la Recherche Scientifique/Environnement et Paléoenvironnements Océaniques), (Desprat, 2005; (http://www.epoc.u-bordeaux.fr/index.php?lang=fr&page=eq_paleo_pollens):

1. 4 to 5 cm^3 sediment samples were washed through sieves to recover the fraction inferior to 150 µm. The sediment of the recovered fraction was separated from the supernatant by decantation during 48 hours.
2. The residue was transferred into a polypropylene centrifuge tube and balanced with distilled water. After centrifugation (7 min at 2500 r.p.m), exotic pollen tablets (*Lycopodium*) were added to each sample. The purpose of this step is to add a known number of exotic marker pollen grains to a known volume of sample, allowing the determination of pollen concentrations. The target is to add about as much exotic

pollen as there is fossil pollen, to minimize counting effort and maximize precision of results (Maher, 1981).

3. To remove carbonates, treatment with cold hydrochloric acid (HCl) was carried out successively; first with 10% HCl, then with 25% HCl until complete cessation of effervescence and at last with 50% HCl.
4. To avoid pollen being obscure when mounted, it was essential to remove silica and silicates (Moore et al., 1991). Hence, two attacks of cold hydrofluoric acid (HF) were performed, first with 25% HF for 2.5 hours and the second with 70% HF for 36 hours.
5. Two further successive attacks with 25% cold HCl were used to eliminate colloidal silica and silicofluorides formed during the treatment with HF.
6. The residue was washed with distilled water and then filtered through a sieve of 10 nylon mesh screens to recover the fraction between 10 and 150 μm (Heusser and Stock, 1984).
7. The final residue was mounted unstained in glycerol. This method has the advantage of being optically suitable, allowing the three-dimensional pollen grains to be rotated below the coverslip by applying a gentle pressure (Moore et al., 1991).

5.2.2 Pollen identification and counting

Pollen were observed and counted on a light Nikon microscope at x 500 (oil immersion) magnification, accompanied by the use of x1000 magnification (oil immersion) for confirmation/identification of critical grains. The slides were scanned along parallel equidistant lines and identifications were achieved based in morphological characters and comparison with specialist atlases (for example Reille, (1992) and (Moore et al.) (1991)).

Pollen grains and spores were identified to the most precise taxonomic category possible given preservation and morphology. Pollen taxa includes family, genus and species, whereas pollen type (referred by suffix '-type') is a pollen morphological category, subsidiary to a pollen class and including pollen grains which can be recognized by distinctive characters (Punt, 1971). Into consideration that pollen grains are easily damage during the fossilization and laboratorial treatments, pollen grains that were hidden, crumpled, corroded and broken were include in the indeterminate group.

A minimum of 200 pollen grains (excluding *Pinus*, aquatic plants, spores, indeterminate and unknown pollen grains), 100 *Lycopodium* grains and 20 pollen types were counted in the 83 samples analyzed. This method was performed to provide a representation of the total population in the original sample and ensure statistically reliable pollen spectra (McAndrew and King, 1976; Maher, 1981; Rull, 1987).

5.2.3 Pollen percentage and concentration

The relative frequencies (pollens percentage) of different pollen types were calculated as follows:

$$\text{\% pollen type ``Y''} = \frac{\text{nº of pollen type ``Y'' counted}}{\text{main sum}} \times 100$$

Results were expressed as percentages of main sum, which excludes aquatic plants, spores, indeterminate and unknown pollens. For the aquatic plants and spores percentages it was used the total sum including aquatic plants, spores and the grains that were not identifiable.

In the KS05 10 core, *Pinus* was excluded from the main sum since it was over-represented in the pollen assemblage (Heusser and Balsam, 1977; Turon, 1984). *Pinus* is a large pollen producer, and because of their particular morphology they can be easily transported (wind, rivers and ocean currents), which favours their over-representation in the marine records (Heusser and Balsam, 1977; Turon, 1984). The percentages of pine were calculated by using the total sum.

The pollen concentration values were determined as follows:

$$\underset{(\text{Grains/cm}^3)}{\text{Concentration}} = \frac{\text{nº of tablets added} \times \text{nº of exotics per tablet} \times \text{nº of exotics counted}}{\text{nº of exotics counted} \times \text{volume of sediment in cm}^3}$$

5.2.4 Pollen diagrams

Pollen data is typically displayed graphically through a vertical sequence of samples in a form of a pollen diagram. This diagram is composed of pollen spectra, expressed as relative frequencies or concentrations of the different taxa, from each level sampled.

The pollen diagram is divided into pollen zones with the aim of obtain a homogeneous assemblage of pollen types (Moore et al., 1991). According to Pons and Reille (1986) the pollen zones can be established using qualitative fluctuations with a minimum of two curves of ecologically important taxa. However, the definition of what constitutes a zone is a subjective concept that depends of the researcher and the purpose of the study (Tzedakis, 1994).

The pollen data can be interpreted with detailed and/or in terms of synthetic pollen diagrams based in the relative frequencies or concentrations. A detailed pollen diagram includes all taxa and a synthetic pollen diagram is constructed after the ecological arrangement of the taxa presented in detailed pollen diagrams (Suc, 1984). In the present study the pollen data are presented in synthetic diagrams which contain the most representative pollen taxa.

Conventionally, the pollen diagram is also arranged into groups of taxa, with arboreal types followed by non-arboreal pollen types, although the precise arrangement within any group differs subtly among the scientists (Moore et al., 1991). The pollen taxa identified in the KS05 10 pollen record of this thesis were grouped as follows:

- Temperate and humid trees and shrubs: *Pinus*, *Acer*, *Ilex*, *Hedera*, *Alnus*, *Betula*, *Carpinus betulus*, *Corylus*, *Buxus*, Cupressaceae, *Hippophae*, *Castanea*, *Vitis*, *Fagus*, deciduous *Quercus*-type, *Juglans*, *Fraxinus excelsior*-type, *Salix*, *Tilia*, *Ulmus*, *Daphne*.
- Pioneer trees: *Betula*,Cupressaceae and *Hippophae*.
- Mediterranean plants: *Pistacia*, *Cistus*, *evergreen Quercus*-type, *Olea*, *Fraxinus ornus*-type.
- Semi-desert plants: *Ephedra fragilis*-type, *Artemisia*, and Chenopodiaceae.
- Ubiquist plants: Apiaceae, *Taraxacum*-type, *Aster*-type, *Anthemis*-type, *Centaurea cyanus*-type, *Centaurea nigra*-type, *Centaurea scabiosa*-type, Boraginaceae, Brassicaceae, Campanulaceae, Caryophyllaceae, Crassulaceae, Cyperaceae, *Scabiosa*-type, Ericaceae, *Calluna*, *Euphorbia*-type, *Mercurialis*-type, Fabaceae, *Ulex*-type, Gentianaceae, *Geranium*, *Erodium*, *Mentha*-type, Liliaceae-type, *Asphodelus*, *Cerealia*-type, *Plantago*, Plumbaginaceae, Poaceae, *Polygonum aviculare*-type, *Rumex*-type, *Ranunculaceae*, *Thalictrum*, Rosaceae, *Filipendula*, *Sanguisorba minor*, *Sanguisorba officinalis*, *Galium*-type, Urticaceae and Valerianaceae.
- Anthropogenic: *Cerealia*-type, *Castanea*, *Juglans* and *Vitis*.
- Aquatic plants: *Myriophylllum verticillatum*-type.

- Spores: *Isoetes*, *Polypodium vulgare*-type, Monolete psilate, Monolete ornamented, *Botrychium*-type, *Cryptogramma*, *Osmunda*, *Pilularia globulifera*, Trilete ornamented and Trilete psilate.

Pollen data can now be easily displayed using one of various computer programs, including Tilia (Grimm, 1993), Psimpoll (Bennett, 2000), and Polpal (Walanus and Nalepka, 1997). The pollen diagrams developed in this thesis were constructed using the Psimpoll program (Bennett, 2000).

Images of the pollen grains presented in KS05 10 synthetic pollen diagram are available in Appendix 1.

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Chapter II

Holocene Climate variability in the mid-latitudes of the eastern North Atlantic region

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ABSTRACT

Vegetation and climate changes in southwestern France/northern Spain are documented for the last ca. 9000 cal. yr BP in a well dated shelf core, KS05 10, retrieved in the southern margin of the Bay of Biscay (Basque country) (43°22’765N, 2°16’744W). The continuous high resolution pollen record shows orbital and suborbital climate fluctuations contemporaneous with those noticed for the North Atlantic region, Greenland and Europe.

A gradual long-term oak and pine forest decline along with an increase in herbaceous plants, reflecting a gradual cooling between 9000 and 1000 yr cal. BP, replicates the cooling trend in Greenland and the general decrease of mid-latitude summer insolation. The replacement of oak and pine forest by beech in the mid-Holocene suggests besides the cooling, an increase of moisture conditions and a decrease in seasonality mimicking the general increase of the precession.

Superimposed on the orbitally induced long-term cooling, KS05 10 pollen record detects an abrupt cool and wet episode, marked by the contraction of hazel trees, the expansion of Cyperaceae and the presence of hornbean, between 8.3 and 8.1 ka in the southwestern France/northern Spain, which is related to the well-known 8.2 ka event. The relatively cold conditions were probably the result of the weakening of the Meridional Overturnig Circulation triggered by the final catastrophic drainage of the Laurentide Lakes and consequent input of freshwater in the North Atlantic region. However, this mechanism can not explain the wet conditions detected in the KS05 10 pollen record. These wet conditions could probably be the result of the influence of the Atlantic Westerly Jet stream and prevalence of strong zonal flow and frequent low pressure systems (associated with less blocking events) over southwestern France/northern Spain. The blockage of sunlight by clouds, which is associated to high precipitation, may be responsible for the particular decline of *Corylus* (light-demanding tree) during this climatic downturn event.

Small-amplitude millennial-scale cooling events until 3 ka are reflected by oscillations of the hazel trees within a ~500yr cyclicity. The impact of human activity on vegetation over the last 1000 years is superimposed on the climatic natural changes.

Key words: Holocene, 8.2 ka event, Climate, Vegetation, Southwestern Europe, Atlantic Westerly Jet stream, Blocking events, *Corylus*

1. INTRODUCTION

Climate is changing significantly since the last few decades, affecting people and the environment worldwide (IPCC, 2007). In particular, increasing air and ocean temperatures, widespread melting of ice sheets, and rising sea levels result in a shift in the Earth's climate system equilibrium. As the climate of our planet appears to be heading for unusual rates of climate change, whether due to natural variability or as a result of human activity, a deeper understanding of the mechanisms that drive the Earth's climate, by studying past climate changes, is therefore crucial for predicting future climate. Hence, there is an urgent need to improve our documentation and understanding of natural variability for periods stretching back beyond the instrumental record. From this perspective, it is of extreme importance to know more about natural climate variations that occurred during the current interglacial, the Holocene (last 11.7 ka), because its boundary conditions are similar to those experienced now and in the near future.

Growing evidences suggest that variations in Holocene climate were larger than previously considered. In the last few years, numerous studies have been performed on several naturally occurring archives such as lake and marine sediments, tree rings, speleothems and ice cores, together with climate simulations, to understand the nature, timing and causes of Holocene natural climate oscillations. Such studies have shown that superimposed on the orbitally-induced long-term cooling sub-orbital millennial-scale climate variability has affected this interglacial. (e.g. Kutzbach and Gallimore, 1988; Bond et al., 1997; 2001; Klitgaard-Kristensen et al.,1998; Von Grafenstein et al., 1998; 1999; Nesje and Dahl, 2001; Tinner and Lotter, 2001; 2006; Baldini et al., 2002; Crucifix et al., 2002; Davis et al., 2003; Weber and Oerlemans, 2003; Magny et al., 2004; Mayewski et al., 2004; Renssen et al., 2005; Lorenz et al., 2006; Naughton et al., 2007a; Wanner et al., 2008; 2011; Seppä et al., 2009).

However, the mechanisms that control Holocene climate variations are far from being resolved. Thus, several questions remain unsolved namely on the role played by the different dynamical processes involved, such as solar activity, volcanic eruptions, meltwater discharge and consequent disruption of the oceanic thermohaline circulation (THC), internal climate variability and feedback mechanisms, in the Holocene millennial scale climate variability (e.g. Bond et al., 1997; 2001; Mayewski et al., 2004; Renssen et al., 2005; 2009; Wanner et al., 2008; 2011). The most extreme and

widespread Holocene short-lived cold event corresponds to the well known 8.2 ka event, which has been detected in the Greenland ice cores and in several high resolution terrestrial and marine paleoclimatic records, especially in the North Atlantic region (e.g. O'Brien et al., 1995; Alley et al., 1997; Bond et al., 1997; 2001; Tinner and Lotter, 2001; 2006; Magny et al., 2003; Muscheler et al., 2004; Veski et al., 2004; Alley and Ágústsdóttir, 2005; Rohling and Pälike, 2005; Kobashi et al., 2007; Naughton et al., 2007a; Seppä et al., 2007; Thomas et al., 2007). Several hypotheses have been invoked to explain this cooling including: a) significant alterations of solar activity (e.g. Bond et al., 2001; van Geel et al., 2003) or b) changes in the general circulation pattern of the North Atlantic region (e.g. Barber et al., 1999; Von Grafenstein et al., 1999; Alley et al., 2003; Ellison et al., 2006). The first hypothesis suggest a reduction of the solar activity as the result of increasing sunspots (e.g. Bond et al., 2001; van Geel et al., 2003) while the second assumption suggest that this cooling episode was triggered by the final catastrophic drainage of the Lakes Agassiz and Ojibway which contributed to the introduction of large amounts of freshwater into the North Atlantic Ocean, disturbed the thermohaline circulation and cooled both Europe and North America (e.g. Barber et al., 1999; Teller et al., 2002; Clarke et al., 2004; Alley and Ágústsdóttir, 2005, and references therein; Ellison et al., 2006; Flesche Kleiven et al., 2008), as supported by several climate models (e.g. Renssen et al., 2001, 2002; Alley and Ágústsdóttir, 2005; LeGrande et al., 2006; Wiersma and Renssen, 2006; Wiersma et al., 2006; LeGrande and Schmidt, 2008; Li et al., 2009).

In contrast, the signature of millennial-scale climate changes during the mid- and late Holocene (after the “8.2 ka event”) is not easily detected in paleoclimatic records. Nevertheless, this variability was strong enough to affect human societies, particularly during the last millennium, as historically documented for the Little Ice Age (LIA) and the Medieval Warm Period (MWP) (e.g. Desprat et al., 2003; Abrantes et al., 2005; Jones and Mann, 2004; Lebreiro et al., 2006; Osborn and Briffa, 2006; Mann et al., 2008; Ljungqvist, 2010; Trouet et al., 2012). The global impact, amplitude, periodicities and causes of these short-lived climatic oscillations are still a matter of debate (Wanner et al., 2008, and references therein; Jones et al., 2009).

In the last few years, several studies have been performed in the Northern Iberian Peninsula and Pyrenees in order to document the Holocene climate variability and particularly the vegetation response to the Holocene climate oscillations (e.g. García-

Antón et al., 2006; González-Sampériz et al., 2006; 2009; Davis and Stevenson, 2007; Muñoz-Sobrino et al., 2007; 2009; García-Amorena et al., 2008; López-Merino et al., 2008, Pèlachs et al, 2009; Iriarte, 2009; Morellón et al., 2009; Carrión et al., 2010; Morales-Molino et al., 2011; Moreno et al., 2011; Rius et al., 2011). However, there are few continuous palaeoclimate studies in that region with sufficient temporal resolution and robust chronological framework to detect Holocene millennial-scale climatic events. Furthermore, high-resolution pollen analysis in marine cores reflecting the regional vegetation of the neighbor continent only exist in the northwestern France (Naughton et al., 2007a) and in the northwestern Iberian margin (Naughton et al., 2007b). Also, there are still some controversies about the climatic signal left by the 8.2 ka in that region. While some authors suggest that this episode was dry (González-Sampériz et al., 2006; 2009; Davis and Stevenson, 2007) others point out for a wet event (Muñoz Sobrino et al., 2009).

In order to improve our understanding on the vegetation response either to the orbitally induced long-term cooling or to suborbital millennial-scale climate variability in northern Iberian Peninsula and southwestern France, we performed a high resolution pollen analysis in a shelf core, KS05 10, retrieved in the southern margin of the Bay of Biscay (Basque country). The Bay of Biscay is considered a key region, being particularly sensitive to changes in the North Atlantic drift and in the atmosphere dynamics being, therefore, able to provide crucial information about the nature, timing and causes of the Holocene climate variability in northern Iberian Peninsula and southwestern France.

2. ENVIRONMENTAL SETTING

2.1 Morphology and recent sedimentation

The selected key-area for studying the Holocene climate variability is located in the southwestern French/northern Spain margin and, in particular, in the continental shelf of the Basque Country (Fig. 1). This continental shelf is narrow, ranging from 7 km off Matxitxako Cape to 20 km off the Oria River estuary (Uriarte, 1998) and is composed of a mudpatch extending from East to West along 56 km (Jouanneau et al., 2008a).

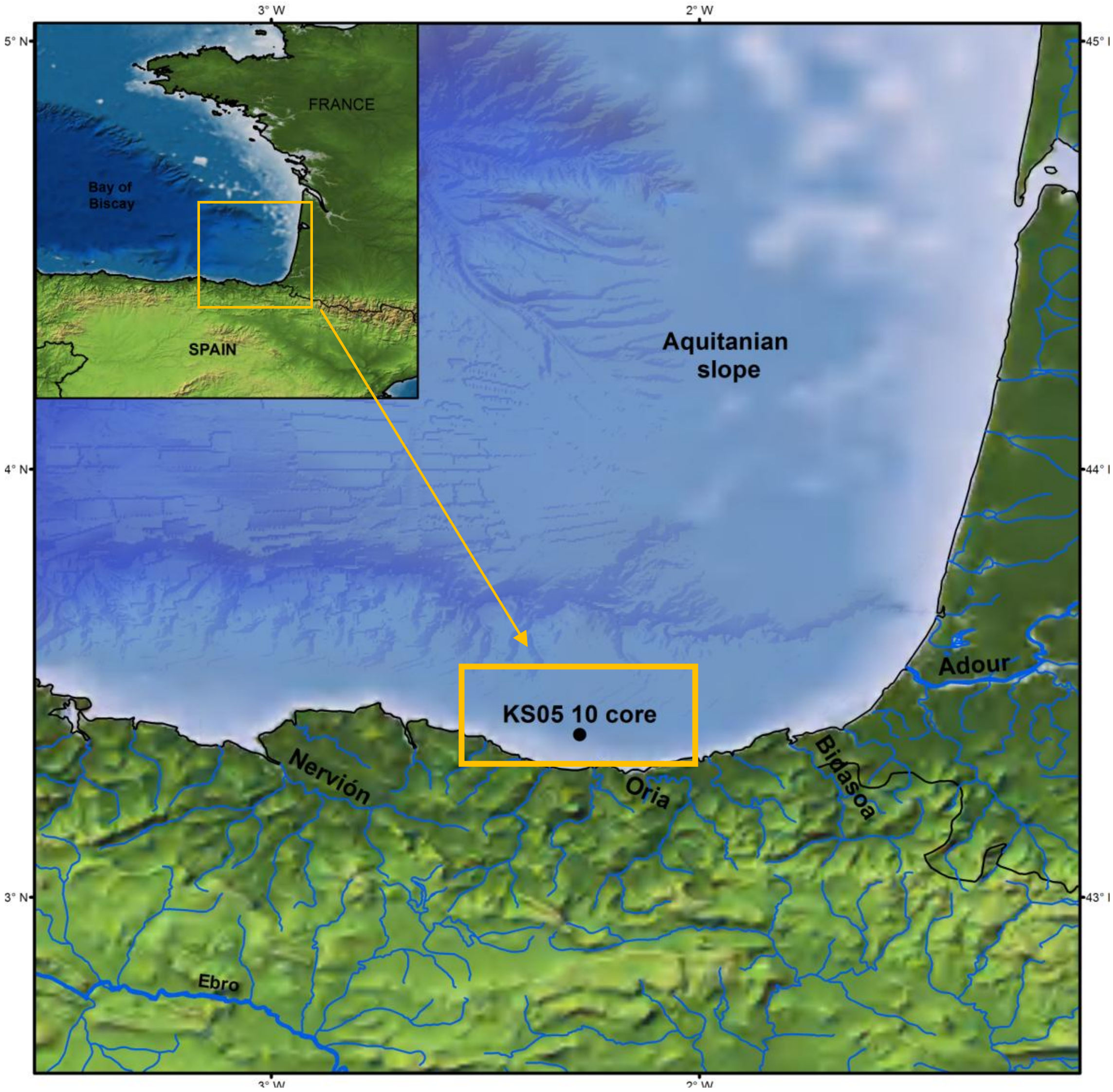

Fig. 1. Location of study area and site (KS05 10; 43°22’765N, 2°16’744W). Sources: The bathymetry is derived from the Digital Bathymetry as produced in the EMODNet Hydrography (http://www.emodnet-hydrography.eu). The elevation data is derived from SRTM (Shuttle Radar Topography Mission) 90m Digital Elevation Database v4.1 (Farr et al., 2007). The drainage system is derived from the Europe and North Asia (EURNASIA) Vmap Level Zero (VMAP0 - Digital Chart of the World) (http://webgis.wr.usgs.gov/globalgis/metadata_qr/metadata/perennial_rivers.htm).

The Basque continental shelf receives directly 1.57×10^{6} t yr^{-1} of sediments from twelve main Basque rivers (Fig. 2) (Madelain, 1970; Frouin et al., 1990; Uriarte et al., 2004; Jouanneau et al., 2008a; Ferrer et al., 2009). The Nervión River is the main sediment supplier to the Basque shelf followed by the Adour (southwestern French river), Oria and the Bidasoa (Fig. 2) (Jouanneau et al., 2008a). Also, the northwestern Iberian Peninsula Rivers can contribute indirectly, by lateral currents such as the poleward current “Navidad” during winter, with sediment to this shelf (Pingree and Le Cann, 1990; 1992; Le Cann and Pingree, 1995; García-Soto et al., 2002; 2004).

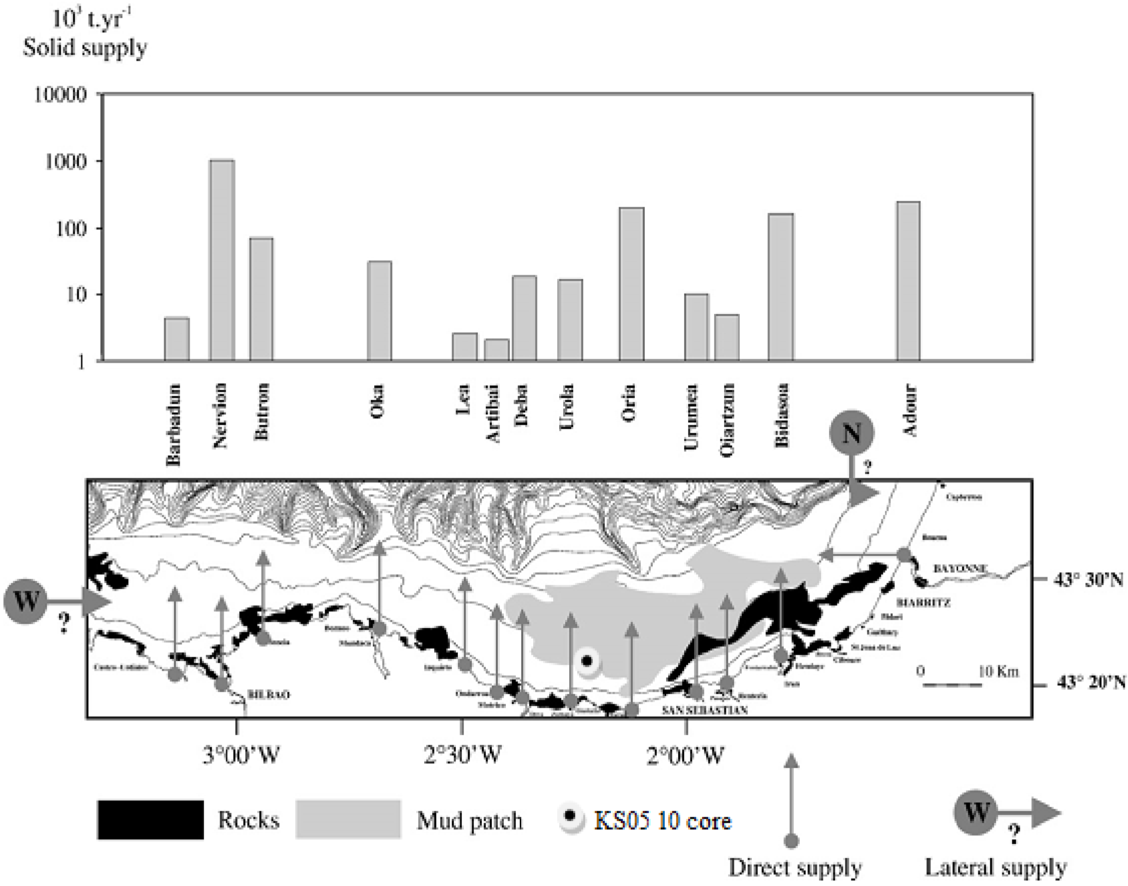

Fig. 2. Sources and solid fluxes to the Spanish Basque country shelf (adapted from Jouanneau et al., 2008a).

The annual sedimentary rate varies from 0.13 to 0.50 cm yr^{-1} in the main mud patch with maximum values in the central mud patch and minimum values close to the rocky outcrops (Jouanneau et al., 2008a).

The western Basque shelf (including the Viscaia and Cantabria shelfs) is caractherised by mostly sandy deposits while the east (Gipuzkoa shelf) by fine sediments (Jouanneau et al., 2008a). The studied core site (KS05 10) is located in the former sector of this mudpatch (Fig. 2).

2.2 Present-day pollen deposition

Numerous studies in coastal areas with well developed hydrographic basins and prevailing offshore winds have shown that pollen grains after being produced and initially dispersed by the wind can sink down and be mainly transported to the sea by rivers and streams (Muller, 1959; Bottema and Van Straaten, 1966; Peck, 1973; Heusser and Balsam, 1977; Heusser, 1978; Turon, 1984; Dupont and Wyputta, 2003; Naughton et al., 2007b). In particular, Turon (1984) has shown that in the French margin, which includes the Basque shelf, the fluvial systems are the primary seaward pollen suppliers, since the westerly prevailing winds and regional complex topography do not favour direct airborne transport of pollen to the sea. Also, Turon (1984) has demonstrated that the marine pollen signature from the French margin reflects an integrated image of the regional vegetation of the close continent. Hence, pollen grains preserved in core KS05 10 are essentially recruited by the rivers from the southern part of the Bay of Biscay and in particular, by those considered as the main sediment suppliers of the Basque shelf (Nervión, Adour, Oria and the Bidasoa rivers). Moreover, and following the experimental work made by Turon (1984) in the French margin, we assumed that pollen grains included in the Basque mud patch represent an integrated image of the regional vegetation of the adjacent continent and in particular that of the Basque country and southwestern France including the Pyrenees.

2.3 Present-day climate and vegetation

The climate of southwestern France and Basque country is particularly sensitive to interannual shifts in the trajectories of mid-latitude cyclones that can lead to significant anomalies of precipitation and, to a lesser extent, of temperature (Trigo et al., 2008). Nevertheless, unlike most of the Iberian Peninsula, the climate of this region is not related with a single mode of variability (Sáenz et al., 2001a). In particular, this region is more affected by the Eastern Atlantic (EA) and Scandinavian (SCAND) modes (Xoplaki, 2002; Trigo et al., 2006) than by the North Atlantic Oscillation mode (NAO) (Trigo et al., 2004a). The SCAND pattern consists of a primary circulation center over the Scandinavia Peninsula and an additional weaker center with opposite sign located over southwestern Europe, being responsible for a large fraction of precipitation registered over central and western Mediterranean basin (Quadrelli et al.,

2001; Xoplaki, 2002; Trigo et al., 2008). This circulation mode corresponds to the typical configuration of the European blocking pattern, usually described in studies using sub-monthly scales (e.g. Tibaldi et al., 1997). Blocking episodes are known to produce significant impacts on both the precipitation and temperature fields of the Mediterranean Region (Trigo et al., 2004b).

The present day climate of this region is temperate, humid and oceanic, with a mean annual temperature higher than 10ºC and a mean annual precipitation of 1500 mm per year (Usabiaga et al., 2004). It corresponds to a Cfb climate (marine west coast-mild) according to Koppen's classification, and presents a more variable distribution of precipitation than of temperature (Sáenz et al., 2001a; b). These climatic conditions are favourable to the expansion of a temperate deciduous mixed forest mainly composed of deciduous oak woodlands (*Quercus robur*, *Q. pyrenaica* and *Q. petraea*) with ash (*Fraxinus excelsior*), birch (*Betula alba*), alder (*Alnus* glutinosa), sycamore (*Acer pseudoplatanus*), hazel (*Corylus avellana*) and elm (*Ulmus*) (García-Antón et al., 2006; García-Amorena et al., 2008). There are also shrubs of *Ulex europaeus*, heaths with different species of the genus *Erica* and brooms (*Genista*). Further back from the coast and at higher altitudes, beech forests (*Fagus sylvatica*) can be found. The Mediterranean influence is noted by the presence of some scattered evergreen sclerophyllous species (*Quercus ilex*, *Phillyrea* spp., *Olea europaea*).

The Basque Country vegetation has been affected by intense human intervention throughout the last millennium. The development of agriculture and arboriculture has favored the expansion of cereals and numerous species of trees, such as *Castanea sativa* (chestnut) and *Juglans regia* (walnut). Nevertheless, after centuries of continuous deforestation, the traditional human-use has undergone a change since the early 19th century, and has been largely replaced by intensive plantations of haloctonous trees, essentially pine (*Pinus radiate*, *P. pinaster*.) and eucalyptus (*Eucalyptus globulus*) (Irizar et al., 2004; Rigueiro-Rodríguez et al., 2005). Currently, forests represent more than half of the total Atlantic-influenced area of the Basque Country, of which nearly two-thirds are planted haloctonous conifers and less than 15% are broad-leaved forests (mainly beech, followed by oak and chestnut) (Irizar et al., 2004).

3. MATERIAL AND METHODS

3.1 KS05 10 sediment sequence

Core KS05 10 was retrieved in the Basque mud patch (southern Bay of Biscay) (43°22’765N, 2°16’744W), at 114 m water depth, using a gravity corer during the cruise EUSKA 3 on board N/O Côtes de la Manche, in 2005. The KS05 10 core is 2.5 m long and is mainly composed of relatively homogeneous silty-clay. Sedimentological analyses including grain size, carbonate content and organic matter were performed by Jouanneau et al. (2008b).

In addition, radiographical analysis using X-ray equipment (SCOPIX® image-processing; Migeon et al., 1999) show a well preserved sedimentary sequence, undisturbed by bioturbation. The pollen record of KS05 10 shelf core also confirms the absence of sedimentological gaps, since it presents the same vegetation succession recorded by continental sequences from the northern Spain and southwestern of France.

3.2 Chronology

The depth-age model of core KS05 10 is based on a combination of eight accelerator mass spectrometry (AMS) ^{14}C dates and one historically well-dated botanical event (*Pinus* expansion) documented in northern Spain (Tables 1 and 2) (Fig. 3). The sedimentatation rate varies between 0.01 and 0.05 cm yr^{-1}, allowing the obtention of a temporal resolution from 60 to 221 years. Thus, the KS05 10 sedimentary record provides essencial information about vegetation and climate variability on centennial-to-millennial time scales in the northern of Spain/southwestern of France (Fig. 3) for the last ca. 9030 cal. yr BP.

The eight ^{14}C ages were measured by accelerator mass spectrometry (AMS) on *Turritela* sp. and other mollusk shells, at Beta Analytic Inc. (Beta) in US and "Laboratoire de Mesure du Carbone 14" in Saclay (SacA) (Table 1). The AMS^{14}C ages were converted from radiocarbon dates (yr BP conv.) into to calendar ages (cal. yr BP) using Calib 6.0 software with the Marine09 calibration dataset (Marine09.14c) (Stuiver and Reimer, 1993; Stuiver et al., 2005; Reimer et al., 2009). This dataset represents the "global" ocean, incorporating a time-dependent global ocean reservoir correction of about 400 years (Stuiver et al., 2005; Reimer et al., 2009). To accommodate local

effects, the geographically dependant reservoir correction of the closest area to our site was incorporated (Δr=3±40 for the Bay of Arcachon, France) (Stuiver et al., 2005). For the depth-age model construction we used 95.4% (2 sigma) confidence intervals and their relative areas under the probability curve as well as the median probability of the probability distribution (Telford et al., 2004), as suggested by Stuiver et al. (2005). The radiocarbon dating of two samples from the bottom of the sequence (levels at 236 and 242 cm) shows reversal values (Table 1). However, we excluded Beta225433 date from the age model because: a) the results obtained by Beta analytic have intrinsic larger errors associated than those of the Laboratoire de Mesure du Carbone 14 (SacA) and b) Beta225433 was the only AMS ^{14}C date that was not obtained on *Turritella* sp..

The botanical event, detected at 7.5 cm of core depth, marks the expansion of *Pinus* in northern Iberian Peninsula at ca. 300 yr cal. BP, reaching its maximum expansion during the last century due to successive reforestation policies and forest management following the suggestion of Valdès and Gil Sánchez (2001) and Desprat et al. (2003). ^{14}C ages included in other available published terrestrial data used for comparision with the KS05 10 record have been converted into calendar years using the program Calib 6.0 (Stuiver and Reimer, 1993) and INTCAL09 calibration curve (Reimer et al., 2009).

AMS ^{14}C dates

Sample Code	Core Depth (cm)	Material	Conventional AMC ^{14}C age (yr BP)	Delta R Arcachon Bay, France	Calibrated date (BP) Two Sigma Ranges (95,4%)	Calibrated date (BP) Median probability
SacA 7392	38	*Turritella* sp.	2800±30	3±40	2365 : 2689	2543
SacA 7393	89	*Turritella* sp.	4695±30	3±40	4805 : 5068	4921
SacA 7394	112	*Turritella* sp.	5185±30	3±40	5437 : 5647	5541
Beta223757	124	*Turritella* sp.	5450±50	3±40	5653 : 5953	5816
SacA 7395	151	*Turritella* sp.	6130±30	3±40	6424 : 6681	6558
SacA 7396	236	*Turritella* sp.	8280±30	3±40	8643 : 8978	8828
Beta225433	*242*	*Mollusk shell*	*8260±50*	*3±40*	*8599 : 8977*	*8797*
Beta223758	246	*Turritella* sp.	8420±60	3±40	8784 : 9246	9028

Table 1. Results of AMS dating of core KS05 10. Level in *italic* corresponds to the age not considered for the age model.

Pollen stratigraphical date

Core Depth (cm)	Age (yr AD)	Botanical event
7,5	1650	*Pinus* expansion

Table 2. Historical botanic event.

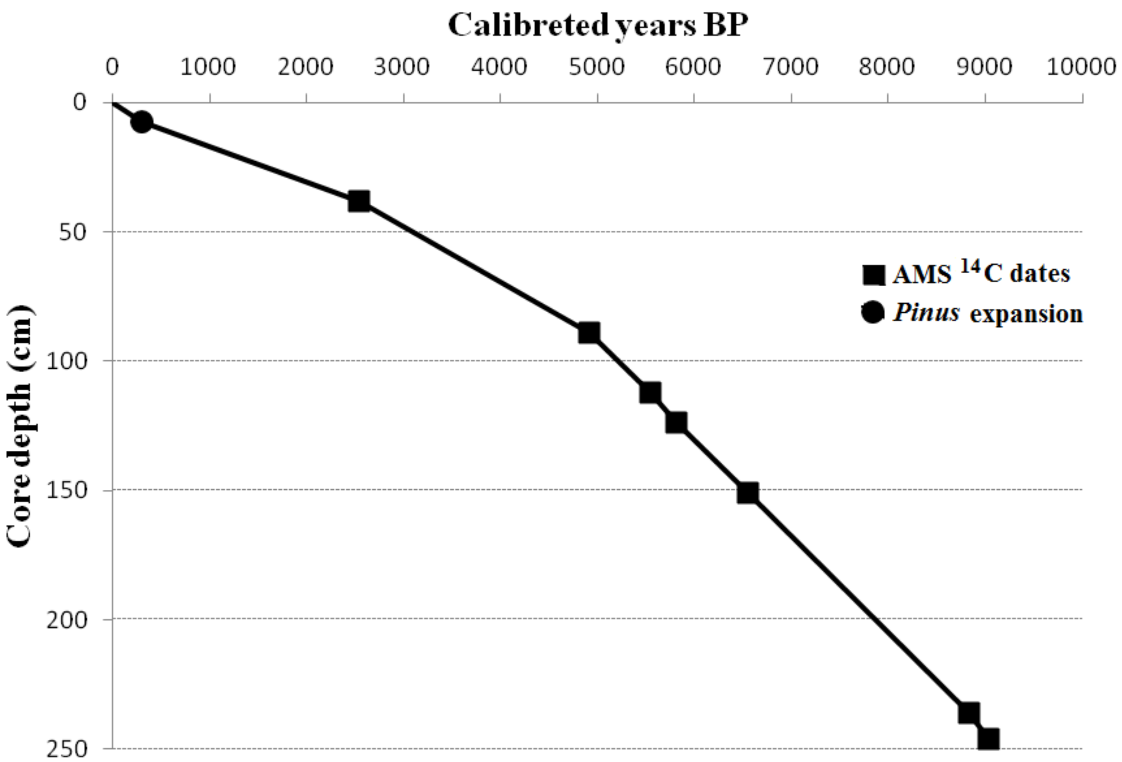

Fig. 3. Depth-age model of KS05 10 record.

3.3 Pollen analysis

The KS05 10 core was subsampled for pollen analysis every 3 cm. The sample preparation technique followed the procedure described by de Vernal et al. (1996) and improved at the UMR CNRS 5805 EPOC (Unité mixte de Recherche 5805, Centre National de la Recherche Scientifique/Environnement et Paléoenvironnements Océaniques) (Desprat, 2005; http://www.epoc.u-bordeaux.fr/index.php?lang=fr&page=eq_paleo_pollens). Exotic pollen tablets (*Lycopodium*) were added to each sample, allowing the determination of pollen

concentrations (pollen grains per unit of sediment volume). To remove carbonates, treatment with cold hydrochloric acid (HCl) was carried out successively at 10%, 25% and 50%. To avoid pollen being obscure when mounted, silica and silicates were eliminated by chemical digestion using cold HF at 25% and 70% (Moore et al., 1991). Colloidal silica and silicofluorides were removed with two successive attacks with 25% cold HCl. The final residue was sieved through 10 µm nylon mesh screens (Heusser and Stock, 1984) and mounted unstained in glycerol, which allowed the understanding of the pollen three-dimensional structure (Moore et al., 1991).

Pollen slides were counted on a light Nikon microscope at x 500 (oil immersion) magnification, accompanied by the use of x1000 magnification (oil immersion) for confirmation/identification of critical grains. The slides were scanned along parallel equidistant lines and identifications were achieved through comparison with specialist atlases (Reille (1992) and Moore et al. (1991)). A minimum of 200 pollen grains (200 to 243, excluding *Pinus*, aquatic plants, spores and indeterminable pollen grains), 100 *Lycopodium* grains and 20 pollen types were counted in the 83 samples analyzed. This method provides a representation of the total population in the original sample and ensures statistically reliable pollen spectra (McAndrew and King, 1976; Maher, 1981; Rull, 1987).

Pollen data was expressed as percentages of the main sum, which excludes *Pinus*, aquatic plants, spores, indeterminate and unknown pollens. *Pinus* was excluded from the main sum since it is strongly over-represented in most of marine deposits (Heusser and Balsam, 1977; Turon, 1984; Naughton et al., 2007b). The percentages of pine were calculated by using the total sum (pollen+spores+indeterminable+unknowns). The pollen diagrams were performed by using the Psimpoll program (Bennett, 2000).

The overall trends of pollen data are similar whether calculated as percentages, concentrations or accumulation rates, thus for the sake of brevity we only present the pollen diagram expressed in percentages. The pollen percentages are plotted against age in a synthetic diagram, which only includes the most representative taxa and the ecologically meaningful groups of taxa (Fig. 4). These ecological groups indicate different climatic conditions: (1) Deciduous *Quercus, Acer, Alnus, Betula, Corylus, Hedera, Ulmus, Tilia, Fraxinus excelsior*-type and *Fagus* are the most important components of the temperate group, indicating a humid and temperate climate; (2) the ubiquist plant group, principally formed by *Taraxacum,* Poaceae, Cyperaceae and

Ericaceae, may reflect cold/cool conditions or, on the other hand, open forests or the development of pasturelands; and (3) the semi-desert group, composed by *Ephedra fragilis*-type, *Artemisia* and Chenopodiaceae, record dry conditions. Cereals, *Juglans* and *Castanea* constitute the group of anthropogenic indicators, representing the influence of human activities in the vegetation composition of the studied region.

The arrangement of the taxa within each group firstly followed the order of appearance in the KS05 10 pollen record and after that of alphabetic order. Arboreal pollen (AP), non-arboreal pollen (NAP) and *Pinus* curves are presented at the bottom of the figure (Fig. 4).

We established a number of pollen zones by using qualitative and quantitative fluctuations of a minimum of 2 curves of ecologically important taxa (Pons and Reille, 1986).

The synthetic pollen diagram is presented jointly with a table which includes a detailed description of the diagram (see below).

3.4 Spectral analysis

Wavelet spectral analysis (Torrence and Compo, 1998) was applied to the obtained pollen data to detect potential cyclicities of vegetation and climate changes. Analysis was undertaken using interactive software available at http://paos.colorado.edu/research/wavelets/ and selecting 90% confidence level against a red noise spectrum.

4. RESULTS AND DISCUSSION

4.1 Vegetation history and climatic variations inferred from KS05 10 pollen record

Pollen analysis of the KS05 10 record allows the recognition of eight major pollen zones numbered from bottom to the top and designated by the sequence name (KS05 10) plus number.

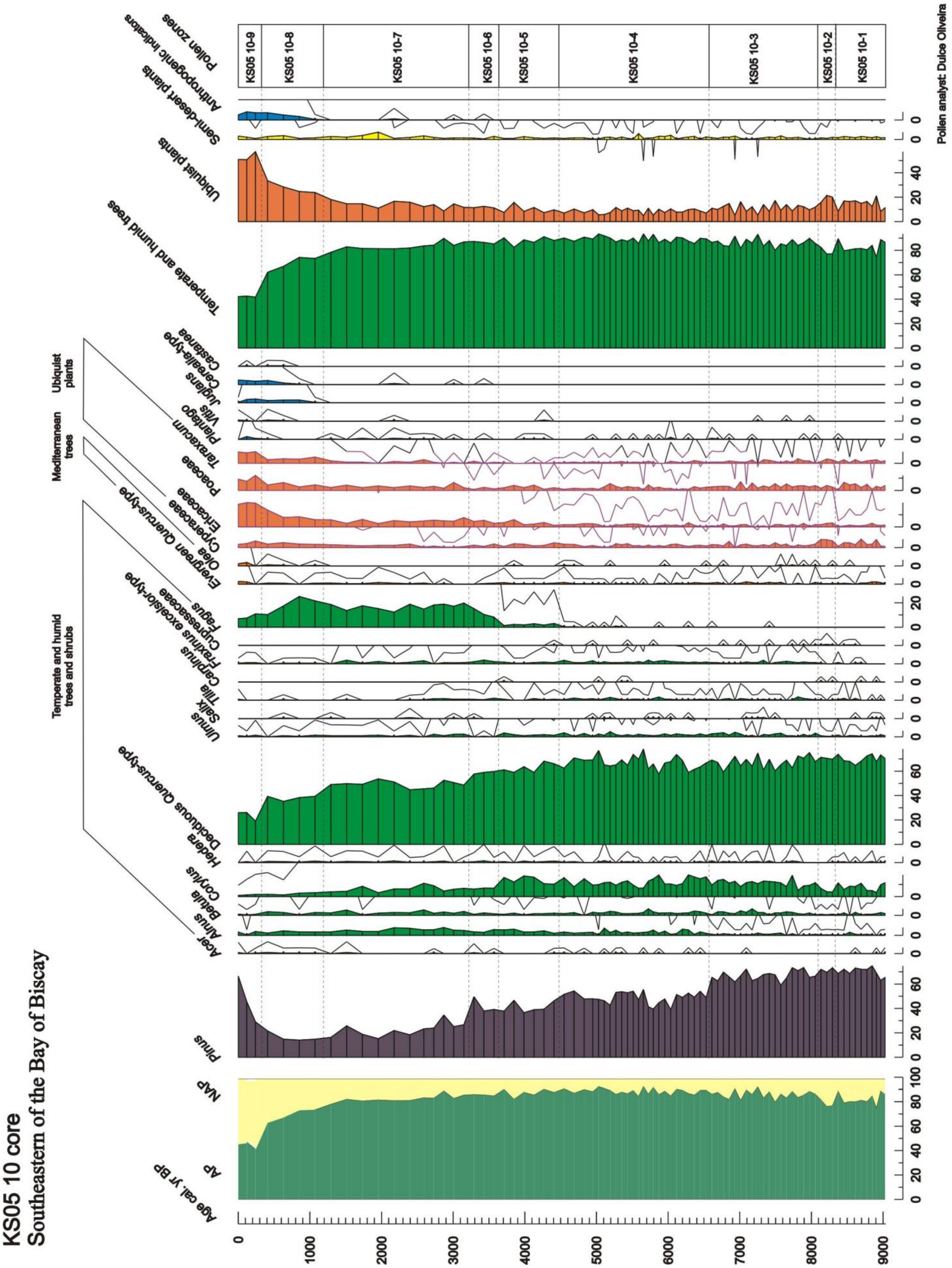

Fig. 4. Synthetic pollen percentage diagram against calibrated ages (cal. yr BP) of KS05 10 record, with a curve at × 10 beyond the principal curve. Dots indicate percentages of less 0.5%.

Pollen zones/Age (yr cal. BP)	**Pollen signature**	**Vegetation formations**
<u>KS05 10 – 9</u> **400 cal. yr BP to the present**	- Reduction of temperate and humid trees (62 to 42% with an important decline of deciduous oak forest and *Fagus*. - Strong increase of *Pinus* (22 to 67%) and herbaceous plants (33 to 51%), particularly Cyperaceae, Ericaceae, Poaceae and *Taraxacum.* - Continuous presence of anthropogenic pollen indicators (Cereal-type, *Juglans*, *Vitis*, *Plantago* and *Castanea*). - Semi-desert group < 2%.	Open pine and mixed deciduous oak-beech forest with heathland and grassland
<u>KS05 10 – 8</u> **1300-400** **cal. yr BP**	- Steadily decrease of *Pinus* (average ~16%), deciduous *Quercus* (49 to 39%) and *Corylus* (average < 3%). Maximum percentages of *Fagus* between ~1070 and 850 cal. yr BP and subsequent gradual decline (25 to 10%). - Strong expansion of ubiquist plants (18 to 33%). Cyperaceae, Ericaceae, Poaceae and *Taraxacum* are the most common plants taxa. - Appearance and development of cultivated species after ~1070 cal. yr BP. Occurrence of a continuous curve of *Juglans* and cereal-type, and presence of *Castanea* and *Vitis.* - Semi-desert group < 2%, with a weak peak between ca. 630 and 410 cal. yr BP.	Mixed deciduous oak-beech and pine forest with ubiquist plants
<u>KS05 10 – 7</u> **3200–1300** **cal. yr BP**	- Strong decline of *Pinus* (50 to 6%) and gradual contraction of deciduous *Quercus* (58 to 49%) and *Corylus*. Consolidation phase of *Fagus*, with relatively constant high percentages (average ~17%). Small increase in *Betula* and *Alnus*. - Gradual rise of ubiquist plants (12 to 18%), mainly Ericaceae and Poaceae. Cyperaceae continuous curve after ca. 2395 cal. yr BP. Sporadic appearances of cereal-type pollen. - Semi-desert group < 1%, with a weak peak between ca. 1955 and 1735 cal. yr BP.	*Pinus* and mixed deciduous oak-beech forest with hazel, alder, birch and heathland
<u>KS05 10 – 6</u> **3700-3200** **cal. yr BP**	- Pine and deciduous *Quercus* forest (average 59%) with *Fagus* and *Corylus*. Important reduction of *Corylus* (16 to 6%) contemporaneous with the *Fagus* expansion phase (1 to 15%). Weak peak of *Fraxinus excelsior*-type. - Nearly constant percentages of ubiquist plants (average ~12%). First occurrence of cereal-type pollen. - Semi-desert group < 1%.	*Pinus* and mixed deciduous oak-beech-hazel forest with *Alnus*, *Fraxinus excelsior*-type and heathland

<u>KS05 10 – 5</u> **4550 -3700** **cal. yr BP**	- Important reduction of *Pinus* (52 to 38%), gradual contraction of deciduous *Quercus* forest and slight expansion of *Corylus* (average ~14%). Small increase of *Alnus* values and *Fagus* phase of arrival associated with its continuous presence (~2%). - Practically steady percentages of ubiquist plants (average ~11%), with a slight increase of Ericaceae, Poaceae and Cyperaceae. - Semi-desert group < 1%.	*Pinus* and mixed deciduous oak-hazel forest with alder, beech and ubiquist plants
<u>KS05 10 – 4</u> **6600 to 4550** **cal. yr BP**	- Distinctive decline of *Pinus* (average 49%). Mixed forest of deciduous *Quercus* (average 67%) with *Corylus* (average 12%) and maximum occurrences of other temperate and humid trees (average 10%), namely *Acer, Betula, Hedera, Ulmus, Salix, Tilia* and *Fraxinus excelsior*-type. Continuous presence of *Alnus*. Some appearances of *Carpinus,* Cupressaceae and *Fagus* pollen grains. - Relatively constant values of herbaceous plants (average ~9%). - Semi-desert group < 2%.	Establishment of a *Pinus* and rich mixed deciduous oak-hazel forest with *Alnus, Betula, Ulmus* and *Fraxinus excelsior*-type
<u>KS05 10 – 3</u> **8100 to 6600** **cal. yr BP**	- Small contraction of deciduous *Quercus* (~ 67%) along with a gradual reduction of pine (average 67%). Expansion and subsequent maintenance of *Corylus* trees (average 11%). Expansion of *Betula* and *Ulmus* and small increase of *Fraxinus excelsior*-type and *Tilia.* Presence of some Cupressaceae grains and first appearance of *Fagus* and *Vitis.* - Slight reduction of ubiquist plants (11%). - Semi-desert group < 1%.	*Pinus* and deciduous oak forest with hazel and some groves of birch and elm
<u>KS05 10 - 2</u> **8300 to 8100** **cal. yr BP**	- Important reduction of temperate and humid trees, marked by a drastic reduction of *Corylus* (average 5%). Appearances of *Carpinus* and Cupressaceae pollen grains. - Increase of ubiquist taxa (average 16%), reflecting a strong expansion of Cyperaceae. Semi-desert group < 2%.	*Pinus* and deciduous oak forest with hazel. Relatively open forest with heathland and grassland
<u>KS05 10 - 1</u> **9030 to 8300** **cal. yr BP**	- *Pinus* (63-75%) and deciduous *Quercus* forest (average 70%) with *Corylus* (3-11%) and small percentages of others temperate and humid trees (average 5%). Occurrences of *Carpinus* and Cupressaceae. - Relatively high percentages of ubiquist taxa (average 14%), essentially Cyperaceae, Poaceae, and *Taraxacum*. - Semi-desert group < 2%.	*Pinus* and deciduous oak forest with hazel groves. Relatively open forest with heathland and grassland

Table 3. Description of pollen zones from the well-dated KS05 10 sedimentary sequence.

The first pollen zone, **KS05 10-1** (ca. 9030-8300 cal. yr BP), is marked by a pine and deciduous oak forest with hazel groves and ubiquist plants (mainly Cyperaceae, Poaceae, *Taraxacum* and Ericaceae) indicating an open temperate forest with heathlands and grasslands (Fig. 4). There are also some scattered pockets of *Alnus*, *Betula*, *Hedera*, *Ulmus* and *Tilia*. This vegetation assemblage suggests relatively warm/cool and wet conditions in southwestern France/northern Iberian Peninsula.

The **KS05 10-2** pollen zone (ca. 8300 to 8100 cal. yr BP) is characterized by an episode of hazel contraction accompanied by a significant increase of heaths and herbs, mostly Cyperaceae, a reduction of temperate and humid trees and the presence of some *Carpinus* pollen grains (Fig. 4 and 5). The reduction of temperate and humid trees and, in particular of *Corylus*, suggests a relatively cooling while the slight expansion of heaths and Cyperaceae and *Carpinus* appearances may reveal an increase in moisture conditions in the southwestern France/northern Spain (see explanations below). This variation in the vegetation composition and climate can be associated with the well known 8.2 ka event.

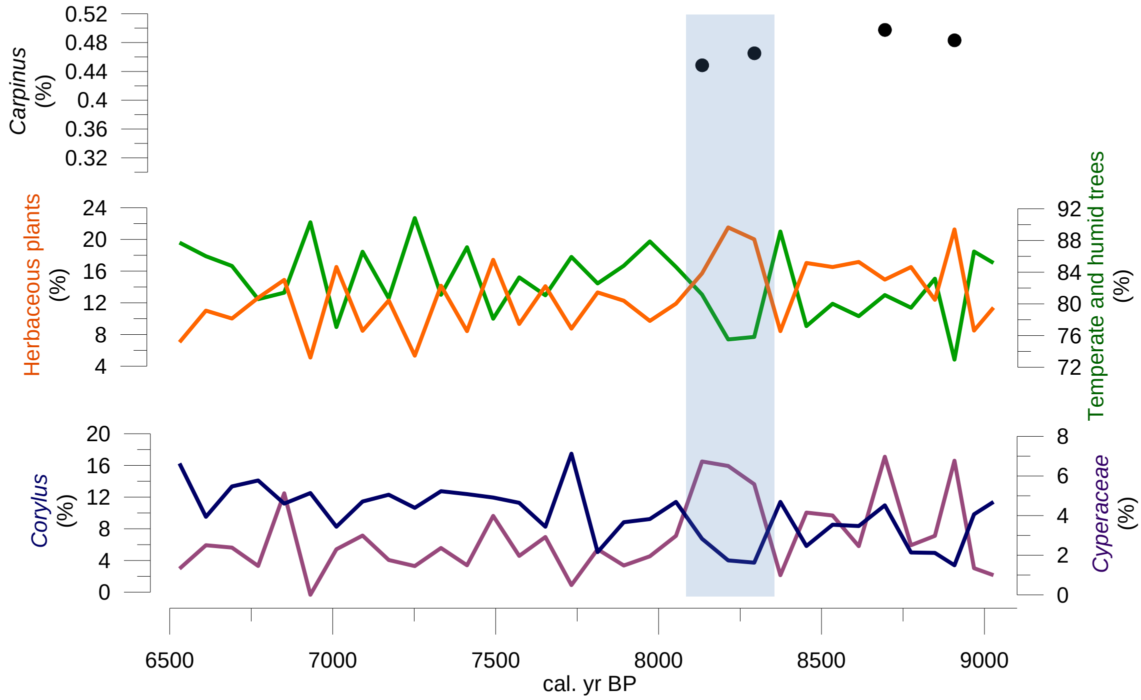

Fig. 5. KS05 10 pollen record with *Corylus*, Cyperaceae, ubiquist plant group, temperate and humid trees and *Carpinus* pollen percentage curves from 6500 to 9000 cal. yr BP.

Temperate and humid trees contraction during the 8.2 ka event, associated with a temperature drop of ~0.5-1.5ºC (e.g. Seppä et al., 2007) were detected in many European records below 61ºN (e.g. Seppä et al., 2007) as far south as the southern Iberian Peninsula (e.g. Fletcher et al., 2007; Combourieu Nebout et al., 2009). Evidences of slightly declining of temperate and humid trees and decrease of arboreal pollen around 8.2 ka cal. BP have been detected in northern Iberian Peninsula and in particular the Cantabrian region (Penido Vello and Puerto de los Tornos-1, Muñoz Sobrino et al., 2005; Pozo do Carballal, Muñoz Sobrino et al., 1997; La Roya , Allen et al., 1996; Lagoa de Lucenza, Muñoz Sobrino et al., 2001 and Lago Enol, Moreno et al., 2011), the central-western Spanish Pyrenees (El Portalet; González-Sampériz et al., 2006) and the northwestern Iberian Peninsula (Naughton et al., 2007b).

The particular reduction of hazel trees in the forest cover, as a response to the 8.2 ka event, has also been detected in several regions as in northern Europe (Raigastvere, Viitna, Rõuge, Ruila in Estonia and Lake Flarken in Sweden; Seppä and Poska, 2004; Veski et al., 2004; Seppä et al., 2005; 2007), southern-central Europe (Soppensee and Bibersee in Switzerland and Schleinsee in Germany; Tinner and Lotter, 2001; 2006) and in northwestern France (Naughton et al., 2007a).

Cyperaceae expansion, around 8.2 ka cal. BP has also been identified in others pollen diagrams from northern Spain, namely: El Portalet in central-western Spanish Pyrenees (González-Sampériz et al., 2006) and Puerto de los Tornos-I in eastern Cantabrian (Muñoz Sobrino et al., 2009). However, there is no agreement on the climate conditions associated to this change. González-Sampériz et al. (2006) suggest that this episode was relatively dry while this work and Muñoz Sobrino et al. (2009) points to a wet phase during the 8.2 ka event. Indeed, Cyperaceae expansion is normally associated with wet conditions (Dupont et al., 1989; Mighall et al., 2006; Muñoz Rodrigues et al., 2007) and can be used as a proxy for determine past changes in moisture (Mighall et al., 2006).

Since *Carpinus* is considered as a moisture-demanding taxa (de Nascimento et al., 2009; Pons et al., 1992; Sanchez Goñi et al., 1999), its presence in the KS05 10 pollen record during the 8.2 ka event further confirms the wet conditions in the area. Nonetheless, only few zones of northern Spain detect the occasional presence of *Carpinus* in their pollen records dispersed along the Holocene period such as in the

Cantabrian–Atlantic region including Lago Marinho, Chan do Lamoso, La Piedra (Ramil-Rego et al., 1998), Urtiaga (Sánchez Goñi, 1992) and Bidasoa basin (Sánchez Goñi, 1996). Some authors suggest that the sporadic appearances of *Carpinus* are very locally confined (Aizpuru and Catalan, 1984; Peñalba, 1994) while others point to the existence of Pleistocene refugia (Ramil-Rego et al., 1998; Postigo-Mijarra et al., 2010). In fact some Iberian and France pollen records from the last glacial (Reille and Lowe, 1993; Burjachs and Julia, 1994) and previous interglacial periods (Sánchez Goñi et al., 1999; 2005) show sporadic occurrences of this taxa suggesting that these areas could have sustained so far undetected refugia (Rodríguez-Sánchez et al., 2010).

These wet conditions within the 8.2 ka event have been also detected in other European records such as in the: West-Central-Europe (high lake-levels, Magny et al., 2003; Magny, 2007), northwestern France (vegetation and quantitative climate reconstructions, Naughton et al., 2007a), Austrian Central Alps (sedimentological and biological analysis, Schmidt et al., 2006), Denmark (macrofossil data and pollen analysis, Hede et al., 2010), Swiss Alps (sedimentological, palynological and macrofossil record, Haas et al., 1998), and in Switzerland and Germany (pollen and charcoal analysis, Tinner and Lotter, 2001; 2006).

A minor reduction of deciduous *Quercus* and gradual contraction of pine associated with the expansion of other temperate and humid trees including *Corylus*, the small spread of *Betula*, *Ulmus*, *Tilia* and *Fraxinus excelsior*-type and reduction of ubiquist plants suggesting that the woodland formation became more closed, marks the **KS05 10-3** pollen zone (ca. 8100-6600 cal. yr BP) (Fig. 4). This suggests a slight warming and the prevalence of moisture conditions when compared with the previous pollen zone. The first appearance of *Fagus*, around 7410 cal. yr BP, is detected in the **KS05 10-3** pollen zone. The detection of *Fagus* pollen prior to its widespread in mid- to late- Holocene is widely recognized in northern Spain and Pyrenees (e.g. Mallarach et al., 1986; Costa Tenorio et al. 1990; Montserrat, 1992; Ramil Rego, 1992; Maldonado, 1994; Penãlba, 1994; Uzquiano, 1995; Rodríguez Guitián et al., 1996; Sánchez Goñi, 1996; Costa et al., 1998; Sánchez Goñi and Hannon, 1999; Ramil-Rego et al., 2000; Muñoz Sobrino, 2001; Gómez-Orellana et al., 2007; Muñoz Sobrino et al., 2009).

The next pollen zone, **KS05 10-4** (ca. 6600-4550 cal. yr BP) is marked by a mixed deciduous *Quercus* forest with *Corylus, Alnus, Betula, Ulmus, Tilia* and *Fraxinus excelsior*-type and a clear decline of pine. Occurrences of *Acer*, *Hedera*, *Salix* and *Fagus* became more regular (Fig. 4). This vegetation cover indicates a stable period of rich and mixed *Quercus* forest, suggesting milder (reduced seasonality) and humid climatic conditions.

The KS05 10 record shows the beginning of a continuous presence of *Fagus* (phase of arrival, **KS05 10-5**; 4550-3700 cal. yr BP) followed by its expansion phase (**KS05 10-6**; 3700-3300 cal. yr BP), a consolidation phase (**KS05 10-7**; ca. 3300-1300 cal. yr BP) and its maximum development (**KS05 10-8**; 1300-400 cal. yr BP). Several pollen diagrams of the northern Spain, exhibit the same pattern of *Fagus* development (e.g. Lagos Altamirano, 1990; Peñalba et al., 1994; von Engelbrechten, 1998; Sánchez Goñi and Hannon, 1999). Also, the timing of beech expansion in KS05 10 record is in agreement with that of northern Spain pollen data (Muñoz Sobrino et al., 2009) and of other Europe records (e.g. Tinner and Lotter 2006; Giesecke et al., 2007; Naughton et al., 2007a). Pollen zones **KS05 10-5 to KS05 10-8** are also marked by the gradual contraction of pine and oak and the steady increase of ubiquist plants. The beginning of beech expansion contemporaneous with the onset of oak forest decline as been also noticed in several Holocene sequences from southwestern France (Reille and Lowe, 1993; Reille and Andrieu, 1995; Reille et al., 2000) and northern Spain (Allen et al., 1996; Muñoz Sobrino et al., 2009). Moreover, previous studies on past interglacials in western Iberian Peninsula and in particular, on the possible analogue of the present day interglacial such as the MIS (Marine Isotopic Stage) 11 shown the presence of *Fagus* in the end phase of VIGO interglacial, associated with a decrease of northern hemisphere summer insolation (Desprat et al., 2005; 2007). The same pattern of *Fagus* development is detected during the PONTEVEDRA interglacial within the MIS 9 (Desprat et al., 2009). The expansion of *Fagus* is probably driven by a long term gradual cooling, reduction of seasonality and increase of moisture conditions during the present interglacial. Similar climatic conditions have been detected by several works on Iberian Peninsula (Laguna de la Roya; Allen et al., 1996) and in northwestern France (Naughton et al., 2007a). Tinner and Lotter (2006) also suggested that *Fagus* expanded naturally across southern Central Europe in response to gradual climatic changes, favoured by more humid summer conditions and less extreme seasonality.

The beginning of the continuous presence of anthropogenic indicators such as *Juglans*, *Cerealia*, *Castanea* and *Vitis* is evident in **KS05 10-8** pollen zone (1300-400 cal. yr BP) reflecting the widespread human activity in south-western France and northern of Spain since ca. 1100 cal. yr BP. This is in agreement to what have been proposed by Allen et al. (1996). Palynological data from a peat bog in València d'Àneu (Lleida, NE Iberian Peninsula; Pèlachs et al., 2009) and from Queixa Sierra (northwestern Iberian Peninsula; Santos et al., 2004) also shows that human impact became stronger and reaches its maximum activity in this last millennium. The maximum values of *Fagus* between ca. 1070 and 850 cal. yr BP might be also related to human actions by the opening of the forest cover. Since *Fagus* compete for slightly opened ground, and exclude other trees when it forms a closed canopy (García Antón et al., 1997; Negral et al., 1997; Pott, 1997), human activities appear to have provided the suitable conditions for its expansion during this interval. This final peak of *Fagus,* was also detected in the eastern Cantabria at around 1250 cal. yr BP by Muñoz Sobrino et al. (2009) and in Central Europe by Tinner and Lotter (2006).

In the last pollen zone, **KS05 10-9** (last ca. 400 cal. yr BP) (Fig. 4) there is a strong increase of pine, heathlands and herbaceous plants, mainly Cyperaceae, Ericaceae, Poaceae and *Taraxacum*. In contrast, temperate and humid trees exhibit a strong decline, principally oak and beech woodlands. The substantial reduction of all trees, except pine, combined with the elevated frequencies of ubiquist plants suggests an open environment probably resulting of more intense human activity. Evidences for anthropogenic deforestation exist from at least the 15th century in the Basque country. At this time the forests were a key factor in the development of the region because of the demand of wood for the shipbuilding, as well as for warfare, and the industrial use of the abundant iron ore together with charcoal (Rodríguez, 2004). Moreover, the continuous presence of cereal-type and *Juglans* and the occurrences of *Castanea* and *Vitis* confirm the effect of human disturbance on vegetation. It should be also noted the increase in *Plantago* and *Olea* values, that reflect as well agriculture/arboriculture activities (e.g. Carrión et al., 2010, and references therein). *Pinus* re-expansion, which began at ca. 300 yr BP ago, reflects the reforestations policies that were implemented with the main aim of preventing the extinction or degradation of the forests (Valdès and Gil Sánchez, 2001; Desprat et al., 2003; Rodríguez, 2004).

4.2. Primary causes of Holocene climate variability in the northern Iberian Peninsula/southwestern France

4.2.1 Holocene long term climatic changes

A general trend of pine and oak trees decline and the increase of herbaceous plants reflect a long term cooling in southwestern France/northern Spain between 9000 and 1000 cal. yr BP (Fig. 6). This small-amplitude long-term pine and oak trees contraction pattern follows the general reduction of mid-latitude summer insolation and also mimics the general decreasing trend observed in the $\delta^{18}O$-isotope composition of the NorthGRIP ice-core (Johnsen et al., 2001; NGRIP Members, 2004; Rasmussen et al., 2006) and of southwestern Iberian margin Sea Surface Temperature (SST) (Rodrigues et al., 2009), until at least 1000 cal. yr BP (Fig. 6). The reduction in the temperate forest coeval with the general decrease in mid-latitude summer insolation has also been detected in the northwestern France (Naughton et al., 2007a) during the Holocene, as well as in north and southwestern Iberian Peninsula during previous interglacial periods (MIS 5, 7, 9 and 11; Desprat et al., 2005; 2007; 2009; Sánchez Goñi et al., 2005). All these evidences point to an orbital induced long term cooling, which has been confirmed by climate simulations (e.g. Crucifix et al., 2002; Weber and Oerlemans, 2003; Renssen et al., 2005).

The gradual replacement of the oak forest by beech in southwestern France/northern Spain reflects not only the long term cooling pattern but also the reduction of seasonality and the increase of moisture conditions in mid to late Holocene (Fig. 6). The reduction of seasonality was probably triggered by the gradual increase of the precession (Fig. 6) as revealed by the pollen quantitative climate estimates from a northwestern France shelf core (Naughton et al., 2007a).

These findings support the hypothesis of Magri (1995) and Naughton et al. (2007a) that Holocene forest decline was mainly controlled by natural processes rather than human impact as suggested by Iriarte (1997; 2009). However, human disturbance on vegetation seems to be superimposed since ca. 1000 cal. yr BP.

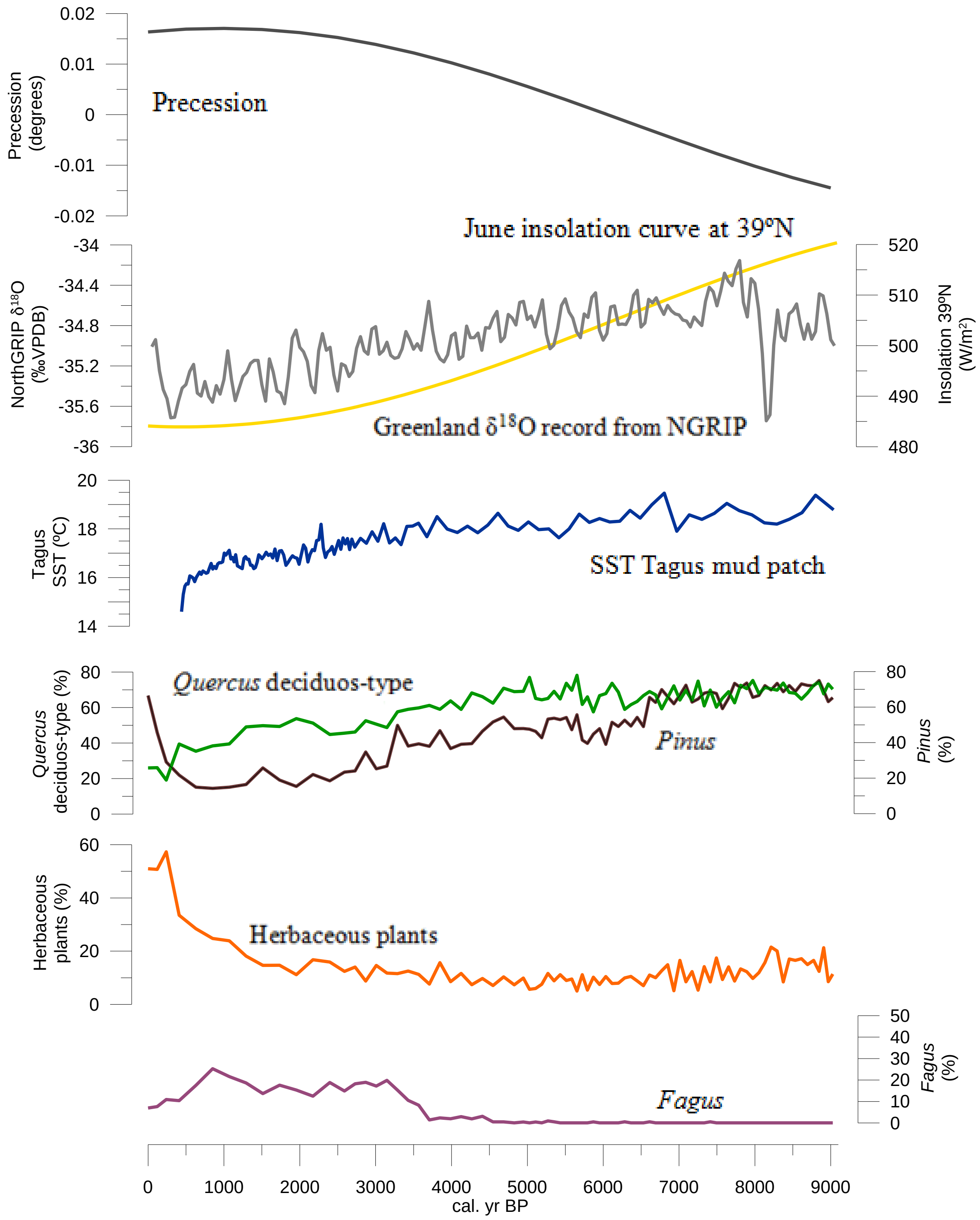

Fig. 6. Correlation between vegetation changes, June insolation curve at 39ºN and precessional signal (after Berger, 1978), $\delta^{18}O$-isotope composition of the NorthGRIP ice-core (Johnsen et al., 2001; NGRIP Members, 2004; Rasmussen et al., 2006) and Sea surface temperature (SST) of Tagus mud patch (Rodrigues et al., 2009) during the last ca. 9000 cal. yr BP. Herbaceous plants association include: Ericaceae, *Calluna*, *Ulex*-type, *Anthemis*-type, *Aster*-type, *Centaurea cyanus*-type, *Centaurea nigra*-type, *Taraxacum*, Apiaceae, Brassicaceae, Caryophyllaceae, Liliaceae, *Asphodelus*, *Mercurialis*-type, *Plantago*, Plumbaginaceae, Poaceae, Ranunculaceae, *Rumex*, Saxifragaceae, *Scabiosa*, Boraginaceae, Campanulaceae, *Cerealia*-type, Cyperaceae, *Euphorbia*, Fabaceae, *Filipendula*, *Galium*-type, Gentianaceae, *Geranium*, *Pedicularis*, *Polygonum aviculare*-type, *Rosaceae*, *Thalictrum*, *Helianthemum*, *Urtica*, Valerianaceae, *Armeria*-type, Eleagnaceae, Polygonaceae and *Potentilla*-type.

4.2.2 Sub-orbital climate variability

Superimposed on the orbitally induced Holocene long-term cooling, sub-orbital climatic oscillations were detected in southwestern France/northern Iberian Peninsula during the last ca. 9000 cal. yr BP.

The 8.2 ka event

The most extreme short-lived Holocene episode detected between 8.3 and 8.1 ka in KS05 10 pollen record is marked by the contraction of temperate trees including hazel, the synchronous expansion of herbaceous plants including cyperaceous plants and the presence of hornbeam (*Carpinus*), suggesting relatively cool and wet conditions. This episode is associated with the well-known 8.2 ka event (Fig. 7).

These vegetation changes are contemporaneous with the decrease of the $\delta^{18}O$-isotope composition recorded in the NorthGRIP record (Johnsen et al., 2001; NGRIP Members, 2004; Rasmussen et al., 2006) and with the southwestern Iberian margin cooling of about 1-2 °C (Rodrigues et al., 2009) (Fig. 7). This event has been also detected in several other records such as: in the mid- and high-latitudes of the North Atlantic (e.g. Marchal et al., 2002; Knudsen et al., 2004; Ellison et al., 2006), in northern and central Europe (e.g. von Grafenstein et al., 1998; Nesje and Dahl, 2001; Tinner and Lotter, 2001; 2006; Hammarlund et al., 2003, Magny et al., 2003; Veski et al., 2004; Seppä et al., 2007), in southern Europe (including Iberian Peninsula and France) (Muñoz Sobrino et al., 2005; 2009; González-Sampériz et al., 2006; Fletcher et al., 2007; Naughton et al. 2007a; b; Combourieu Nebout et al., 2009; Moreno et al., 2011) and other Greenland ice cores records (GRIP, Dye-3, Renland, NorthGRIP and GISP2 ice cores) (Alley et al., 1997; Johnsen et al., 2001; NGRIP Members, 2004; Rasmussen et al., 2006; Thomas et al., 2007).

This cooling was triggered by a succession of events which began with the catastrophic freshwater influxes from the proglacial Laurentide lakes (lakes Agassiz and Ojibway) into the Hudson Bay at ca. 8470 cal. yr BP (Fig. 7) (e.g. Barber et al., 1999; Teller et al., 2002; Clarke et al., 2004). The introduction of large amounts of freshwater into the North Atlantic (Alley et al., 1997; Clark et al., 2001) affected the thermohaline circulation (THC) as showed by model simulations (e.g. Wiersma and Renssen, 2006; Wiersma et al., 2006) and reduced the formation of North Atlantic Deep Water

(NADW), causing a weakening in the conveyor belt (e.g. Barber et al., 1999; Rahmstorf, 2002; Renssen et al., 2001; LeGrande et al., 2006; Wiersma and Renssen, 2006; Wiersma et al., 2006; Li et al., 2009). Consequently, the ocean's heat transport was affected (decreased) and the North Atlantic and Europe underwent a significant cooling. Paleoclimatic records from the North Atlantic also reveal changes in the deep ocean circulation (Fig.7) (Oppo et al., 2003; Hall et al., 2004; Piotrowski et al., 2004; Ellison et al., 2006; Flesche Kleiven et al., 2008).

Other hypotheses have been invoked for explain this cooling episode. A reduction in solar irradiance (Bond et al., 2001; van Geel et al., 2003) could have modified the behaviour of the oceanic circulation (Weber et al., 2004; Renssen et al., 2006). Also, and based on western-central Europe lake levels data Magny and Bégeot (2004) suggest that the solar activity might have played an essential role in amplifying (or damping) the potential effects on climate of freshwater outbursts events. Nevertheless, the fact that this event is more prominent in the North Atlantic region; that took place after following outbursts flooding episodes; and that the existing similarities between reconstructed anomaly patterns and patterns expected following a North Atlantic freshening appears to favour the freshwater pulse mechanism as the primary cause of the 8.2 event (Alley and Ágústsdóttir, 2005). In addition, corroborative modeling studies also stresses that low solar output can be hardly considered as a main triggering factor for abrupt climate events (Ganopolsky et al., 2001; Goosse et al., 2002; Bauer et al., 2003).

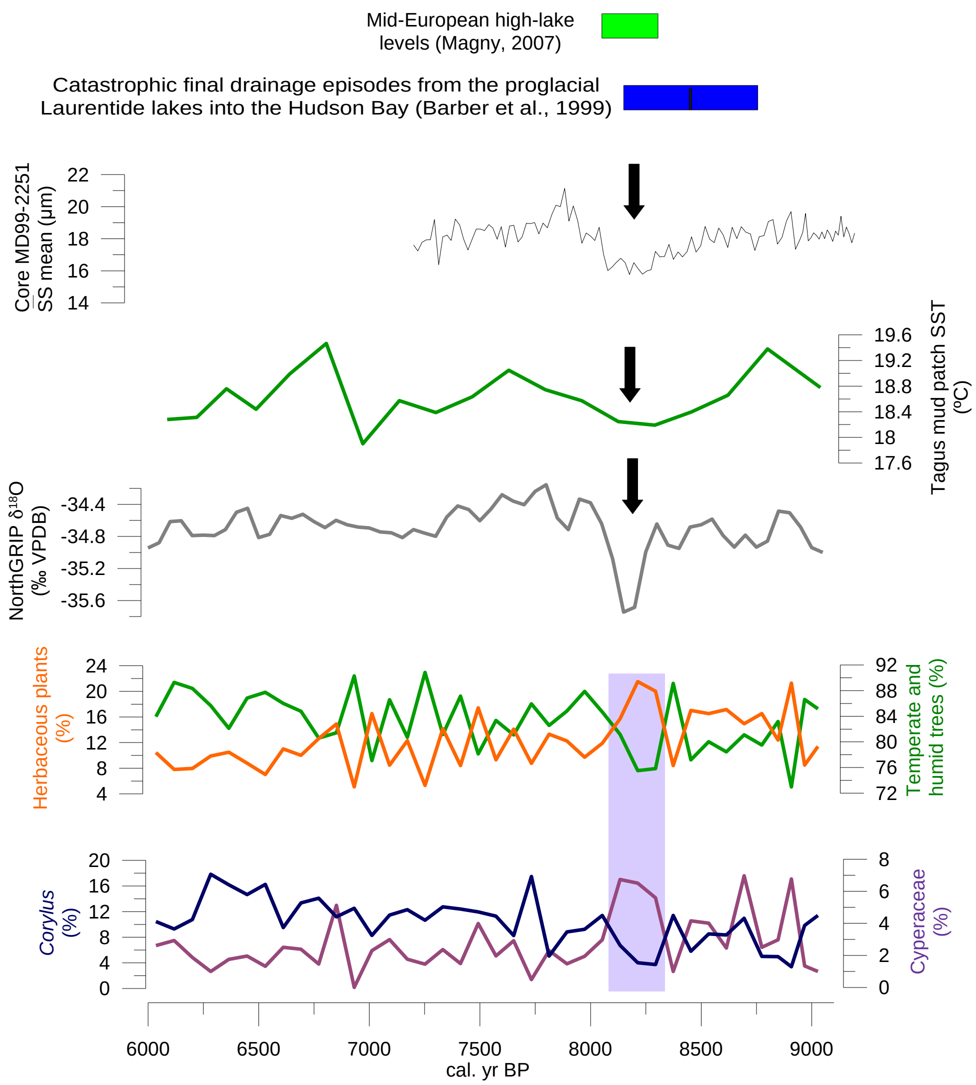

Fig. 7. Correlation between KS05 10 vegetation changes, δ^{18}O-isotope composition of the NorthGRIP ice-core (Johnsen et al., 2001; NGRIP Members, 2004; Rasmussen et al., 2006), Sea Surface Temperature in the Tagus mud patch (Rodrigues et al., 2009), $\overline{SS}$ mean grain size (paleocurrent flow speed proxy, higher mean indicates stronger flow of the depositing current and vice versa; Ellison et al., 2006), the catastrophic final drainage episodes from the proglacial Laurentide lakes into the Hudson Bay at ca. 8470 cal. yr BP (error range of 8160–8740 cal. yr BP; Barber et al., 1999) and episodes of higher lake level in west-central Europe (Magny, 2007), during the 8.2 ka cooling event.

However, the reduction of the termohaline circulation can not explain the wet conditions detected in southwestern France/northern Spain. Other mechanism must have been involved. Magny et al. (2003) suggested that the North Atlantic cooling might have induced a stronger thermal gradient between high and low latitudes and an increasing cyclonic activity over the European mid-latitudes. In turn, these changes in the atmospheric circulation would have resulted in a southward displacement of the Atlantic Westerly Jet stream, which determines the latitudinal extension of the Hadley cell and the mid-latitude storm tracks. As a result, mid-latitudes, between ca. 50º and 43º, underwent wetter conditions in response to the cooling of the 8.2 ka event, while the northern (e.g. Scandinavia) and southern (e.g. southern Iberia) Europe were marked by dry conditions (Fig 8.).

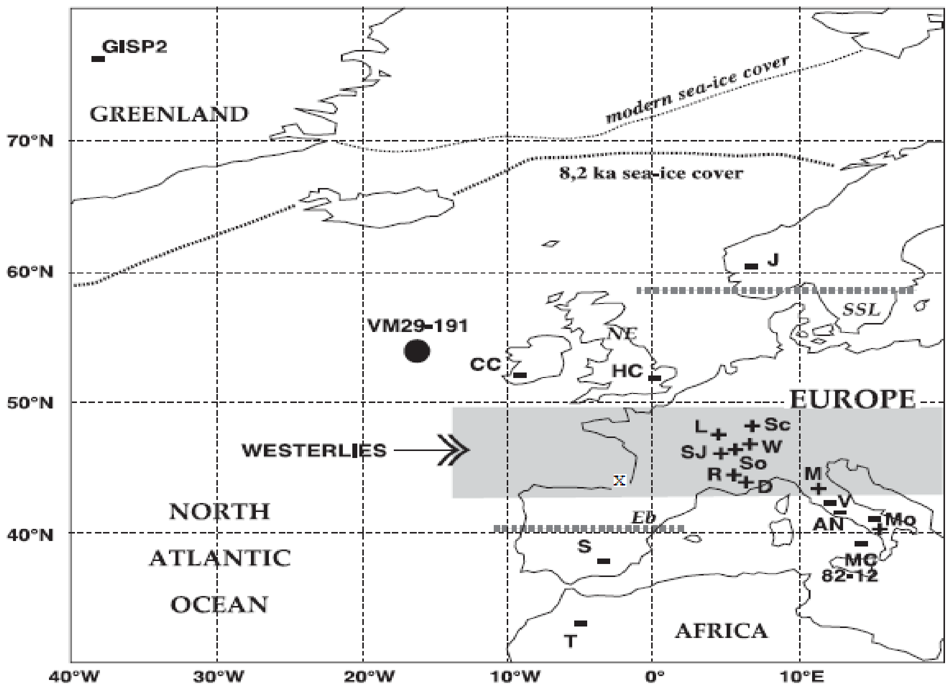

Fig. 8. Comparison of hydrological signals related to the 8.2 ka cold event in Europe (from Magny et al., 2003). Shaded area represents the mid-European zone with wetter conditions. Lake-level maxima is marked with (+) and minima with (-) during the 8.2 ka event. The extension of sea-ice cover during the 8.2 cal. yr BP (Renssen et al., 2001). Dashed lines correspond to possible northern and southern limits of the mid-European wetter zone during the Holocene cooling phases weaker than the 8.2 ka event. X marks the location of KS05 10 sequence. For reference sites included in the Fig. during the 8.2 ka event see Magny et al. (2003).

We propose that the mentioned large scale atmospheric mechanism associated with more storms crossing the North Atlantic region mid-latitudes might be related with the absence of blocking episodes, more specifically a prevalence of strong zonal flow. This situation could explain the observed wet conditions in southwestern France/northern Spain during the 8.2 ka event. The blocking episodes correspond to an atmospheric setting of predominantly meridional circulation that favours the formation of strong slow-moving or stationary anticyclones (blocking events) over the European mid-latitudes (Treidl et al., 1981). Blocking events are related with the blocking of the mid-latitude westerly Jet stream, which is usually located over the Atlantic at around 50ºN latitude. These atmospheric events are characterised by enhanced meridional component of the mid-troposphere circulation, clearly visible up and downstream of the British Isles, and by a configuration that usually involves a branching of the Jet stream into two divergent branches (Fig. 9a) (Rex 1950a,b; Treidl et al. 1981, Trigo et al., 2004b). Conversely, a situation of strong zonal flow (non-blocking events) is characterised by a strengthening of the zonal circulation (Fig. 9b) (Trigo et al., 2004b).

In the Euro-Atlantic sector, Trigo et al. (2004b) through a 40-year (1958–97) consistent dataset from NCEP/NCAR showed that the precipitation rate variability (Fig.11) is controlled by synoptic low pressure mid-latitude systems that induce precipitation, often known as mid-latitude cyclones. A simple proxy variable to depict these disturbances is given by the surface vorticity field that is related to the favoured locations of associated cyclones (Fig.10). In the studied region, during blocking situations there is a lack cyclones centres crossing the Northern Europe, including the western France, and a small number travelling in the northern Iberian Peninsula (Fig.10a), and drier than normal conditions in all western Europe, including the northern Iberian Peninsula and southwestern France (Fig. 11a) (Trigo et al., 2004b). Conversely, months dominated by strong zonal flow (i.e. no blocking events) are characterised by a significant increase of cyclones (Fig.10b) that lead to wetter than normal anomalies in western Europe, including the northern Iberian Peninsula and southwestern France (Fig.11b) (Trigo et al., 2004b). The anomalies of the precipitation rate associated with non-blocking events in the study region during the period 1958–1997, ranges from ~0.3 to 1.5 mm/day (northern Iberian Peninsula and western France, respectively) for the 90-day winter period (Trigo et al., 2004b).

Since individual blocking episodes can persist for several weeks, a small number of blocking episodes might influence the climate characteristics of one winter (Stein, 2000). In this sense, we can hypothesize that past oscillations in the predominance and spatial distribution of blocking episodes (and their non-blocking counterpart; strong zonal flow) could explain changes in the precipitation rate of the studied region (prevalence of strong zonal flow situation) during the 8.2 ka event, as well as support the hydrological tripartite division of Europe (dry-wet-dry) during this event described by Magny et al. (2003) and Magny (2007).

In agreement with the hypothesis of moister conditions during the 8.2 ka event in the studied region, higher precipitation rate can be appropriately related to increased cloudiness and the consequent decreasing of incoming sunlight. Since *Corylus* is a light-demanding tree (Bradshaw and Hannon, 2004), this cloudiness effect may have contributed to the significant hazel decrease by affecting negatively populations growth and/or pollen productivity.

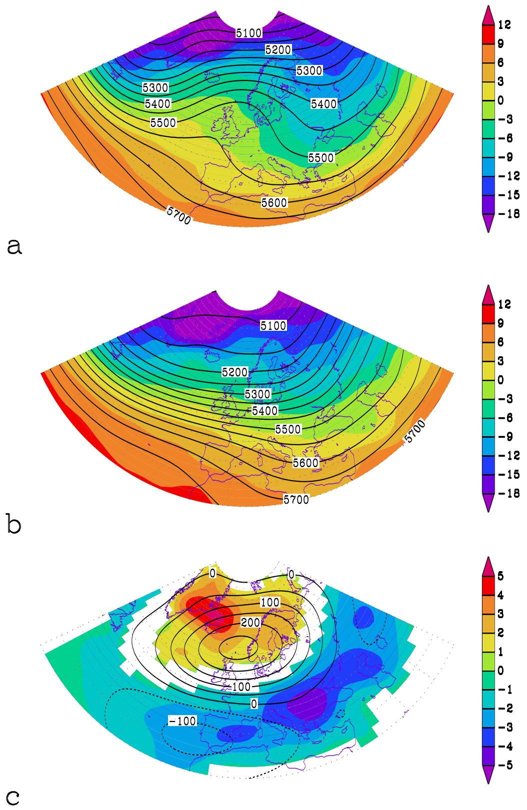

Fig. 9. The mean 500 hPa geopotential height (gpm) anomalies for all winter a) blocking, and, b) non-blocking (strong zonal flow) episodes, with a minimum duration of 10 days (from Trigo et al., 2004b). The shading shows the corresponding 850 hPa temperature field (ºC); c) Differences between the mean 500 hPa geopotential height (gpm) composites and the corresponding 850 hPa (ºC) temperature composites (represented only if significant at the 1% level).

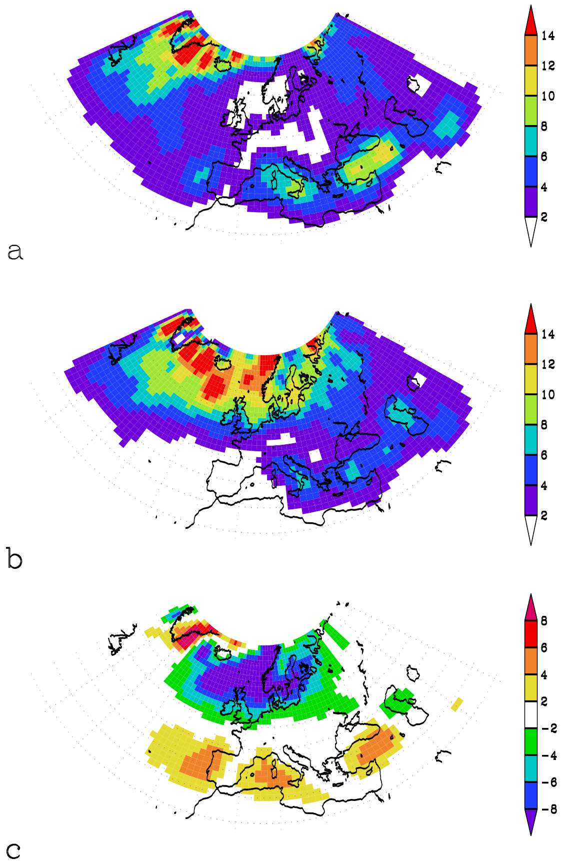

Fig. 10. Number of cyclones per winter, detected per 5º x 5º area normalised for 50ºN, for a) blocking, b) non-blocking, and c) their difference (from Trigo et al., 2004b).

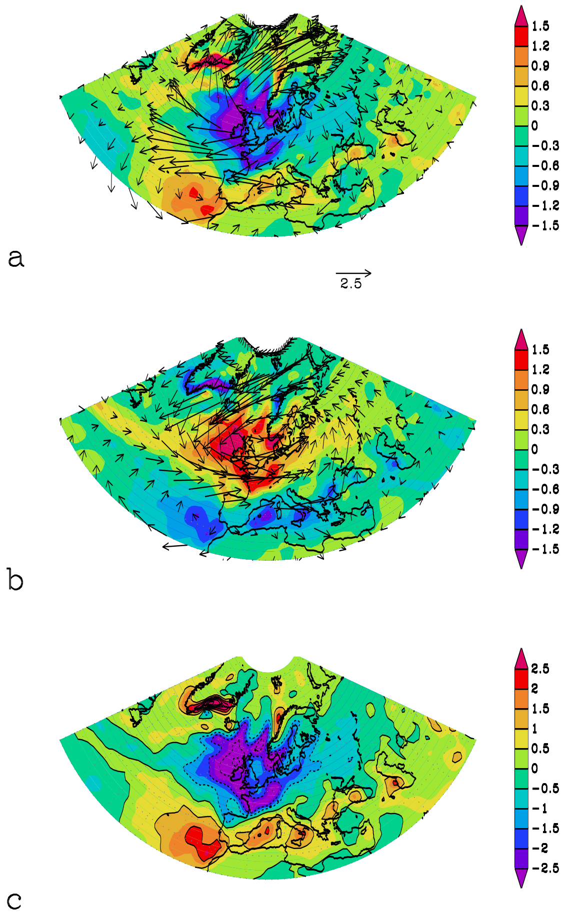

Fig. 11. Anomalies of the precipitation rate (mm/day) for winter composites of a) blocking episodes, b) non-blocking episodes, and c) their difference (represented only if significant at the 5% level). The arrows show the respective anomaly of the 2.5 m wind field (ms^{-1}) (adapted from Trigo et al., 2004b).

Discrete millennial scale climatic events after the 8.2 ka event

After the strong short-term climate anomaly at ca. 8.2 ka, millennial scale climatic events became less perceptible. The application of standard spectral analysis such as the widely used Fast Fourier analysis can lead to misleading results because the power spectrum associated with the major periodicities can vary with time. Here, we decided to apply a more robust technique that can depict changes of periodicities trough time, i.e. wavelet analyses revealing that *Corylus* percentages exhibits prominent

concentrations of variance centered at 512 years (Fig. 12) from 9000 to 3000 cal. yr BP. This climatic cyclicity might be related to that of ~500-yr recognized in several the North Atlantic paleoclimate records (e.g. Chapman and Shackleton, 2000; Bond et al., 2001; Yu and Ito, 2002) and over Greenland (Stuiver and Braziunas, 1993; Stuiver et al., 1995) for the mid-late Holocene. If so, these findings could suggest that *Corylus* trees in northern Iberian Peninsula/southwestern France may have responded to the same climate forcing mechanisms of the mentioned studies, such as changes in solar irradiance (e.g. Bond et al., 2001) and perturbation of the North Atlantic circulation (e.g. Chapman and Shackleton, 2000). Nevertheless, additional palaeoclimate records with sufficient resolution and/or chronological control are necessary to investigate this connection.

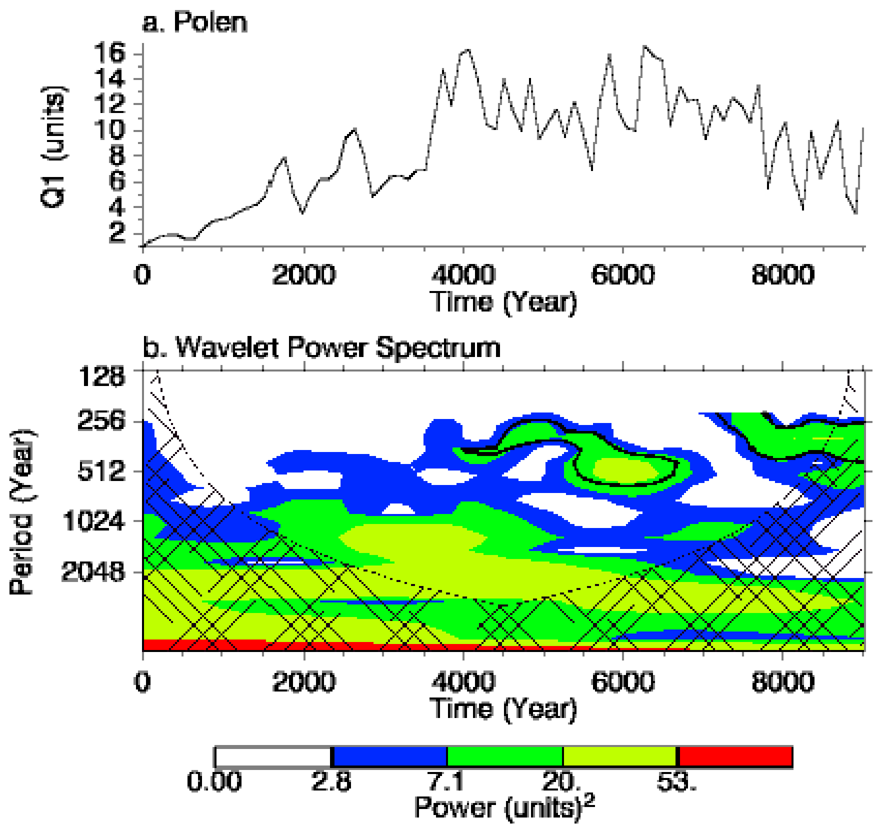

Fig. 12. Wavelet analysis of the hazel percentages for the last ca. 9000 yr. Wavelet power spectra illustrate the change in concentration of spectral power with time in *Corylus* values. Black line defines power spectrum significant at 90% red noise spectrum. Dashed lines define the cone of influence where the spectrum has no significance at all. Analysis undertaken using interactive software available at http://paos.colorado.edu/research/wavelets/.

5. CONCLUDING REMARKS AND FUTURE RESEARCH

High-resolution pollen analysis performed on the KS05 10 shelf core reflects the paleoclimatic evolution and vegetation's history over the southwestern France/northern Spain for the last ca. 9000 years.

The major findings of this study can be summarized as follows:

- The gradual decline of some temperate trees, namely pine and deciduous *Quercus*, together with the steady increase of herbaceous plants follows the general long-term cooling trend driven by mid latitude summer insolation until at least 1000 cal. yr BP. The gradual decrease of deciduous *Quercus* and steady increase of *Fagus* during the mid-Holocene might reflect besides the long term cooling, an increase in moisture conditions and a decrease of seasonality;

- The contraction of *Corylus* together with synchronous expansion of herbaceous plants including Cyperaceae and the presence of *Carpinus*, suggesting cool and wet conditions was detected between 8.3 and 8.1 ka and is associated with the well known 8.2 ka event;

- The particular cool conditions associated with the 8.2 ka event were triggered by a succession of events: a) catastrophic drainage episodes of the proglacial Laurentide lakes b) freshwater input in the North Atlantic region and c) consequent perturbation of the thermohaline circulation which preclude the transfer of warm and wet conditions to the southwestern France/northern Spain;

- The particular wet conditions associated with the 8.2 ka event were triggered by the North Atlantic cooling which produce an increase of the thermal gradient between high and low latitudes favouring the southward displacement of the Atlantic Westerly Jet in conjunction with the prevalence of strong zonal flow in the studied region;

- *Corylus* (light-demanding tree) decline can also be explained by the increase of clouds (high precipitation) which may have contributed to a reduction of incoming sunlight;

- After ca. 9000 cal. yr BP, *Corylus* changes occurred every 500 years until 3000 cal. yr BP;
- The climate variability over the last millennium was possibly masked by human disturbance.

This study highlights the need for Holocene multi-proxy records with high temporal resolution and robust chronology from the North Atlantic region and Europe, which will: a) provide an accurate reconstruction of the vegetation's response to the global orbital induced long-term cooling and to millennial-scale climate variability of the North-Atlantic, including oceanic and atmospheric dynamics; and b) allow the identification of the nature, timing and causes of climate oscillations and to determine how widespread, systematic and abrupt they may have been. In this sense, the results of this master thesis will be integrated with the existing high resolution data from several key geographic sites of the Atlantic realm included in the CLIMHOL project "Holocene climatic variability in the North Atlantic and adjacent landmasses: land-sea direct correlation" (reference PTDC/AAC-CLI/100157/2008).

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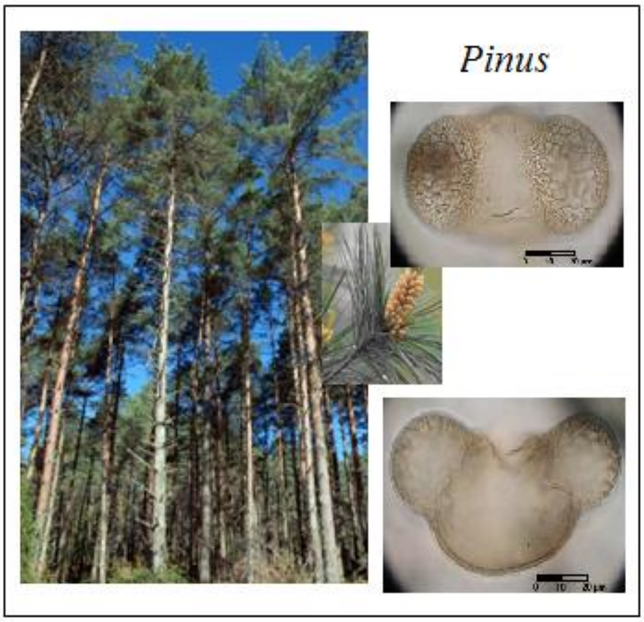
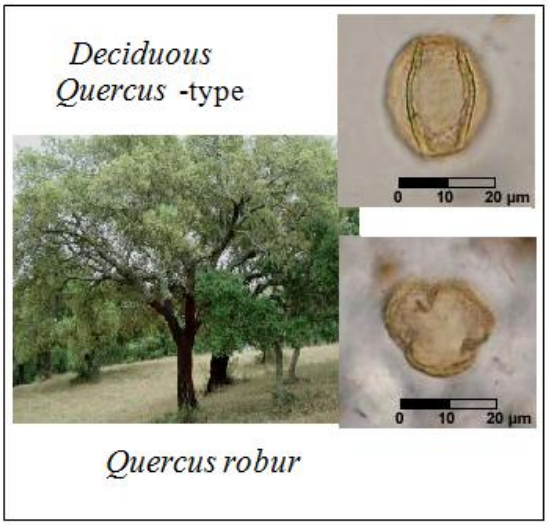
Appendix 1

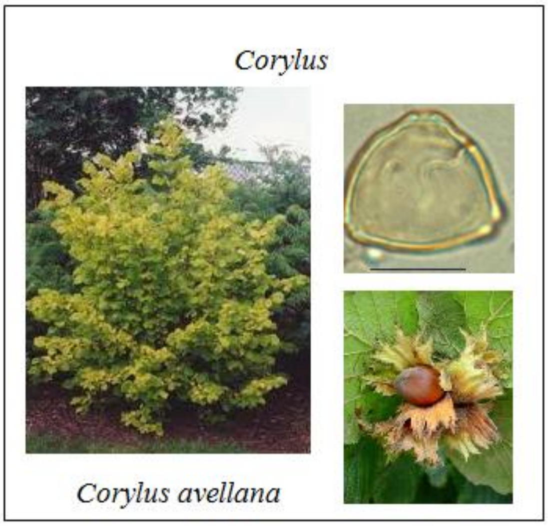
Images of the pollen grains presented in KS05 10 synthetic pollen diagram.

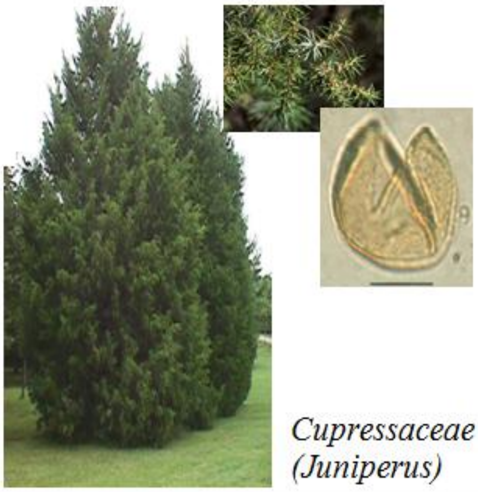
Pinus



Temperate and humid trees







Cupressaceae
(Juniperus)

Fagus sylvatica

Tilia
Tília tomentosa

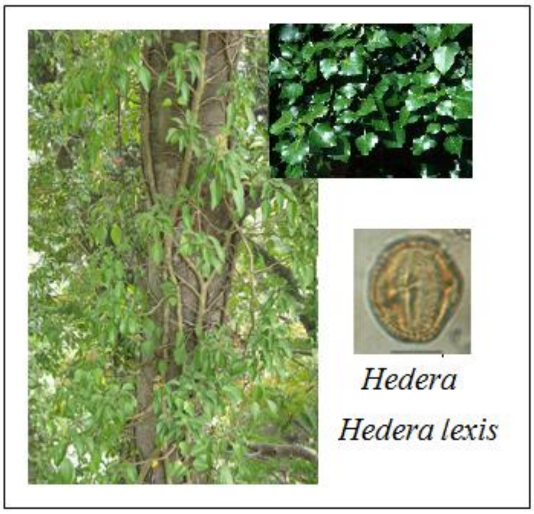
Salix

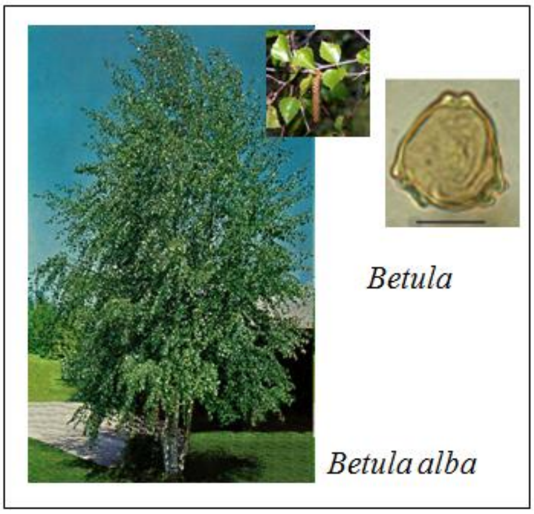
Alnus

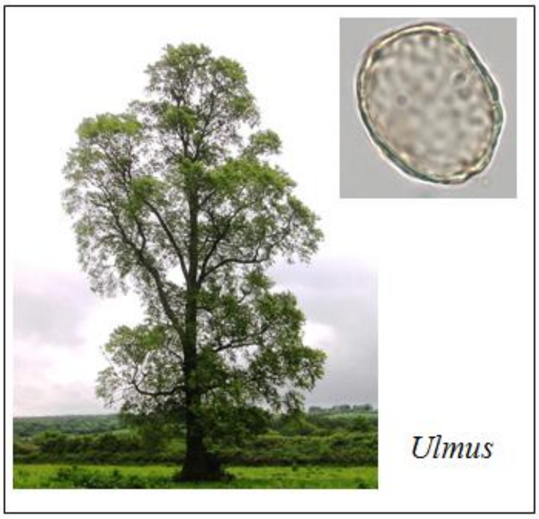
Alnus glutinosa

Acer

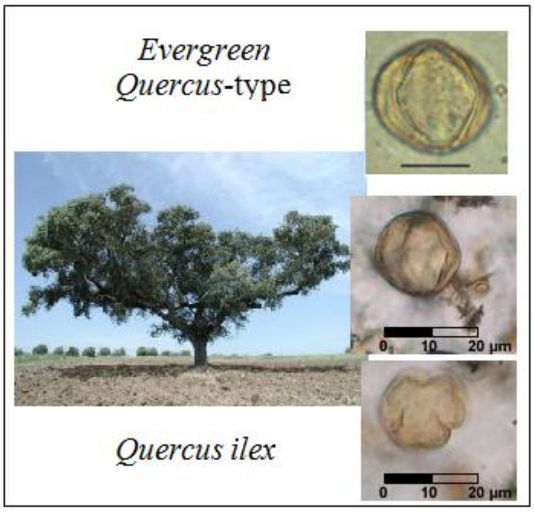
Acer campestre

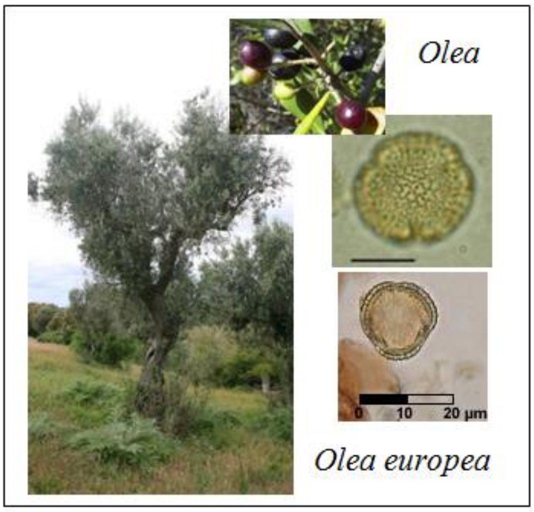




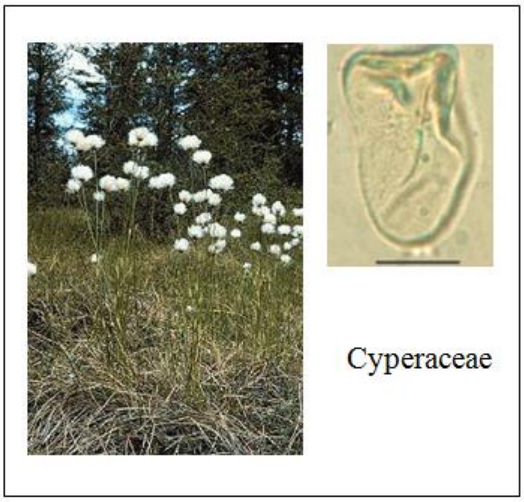


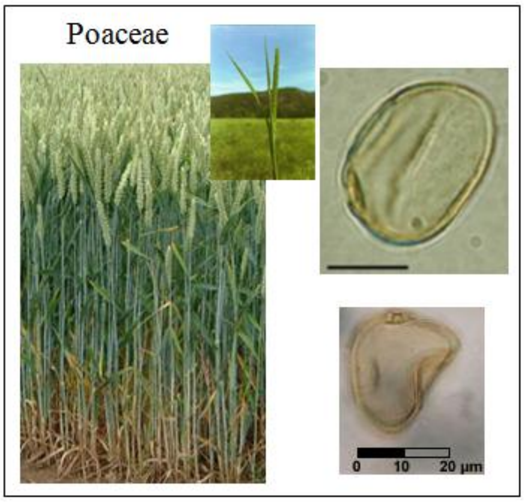
Mediterranean trees

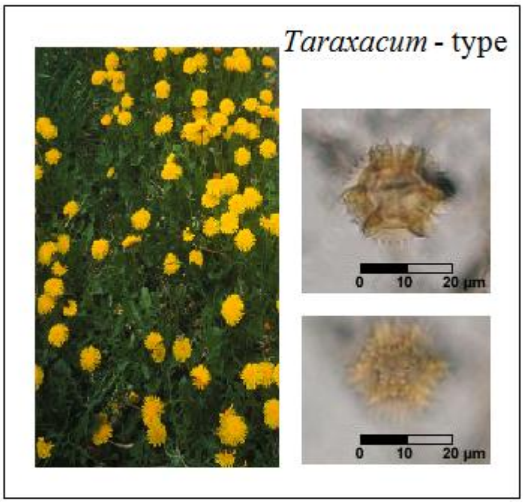




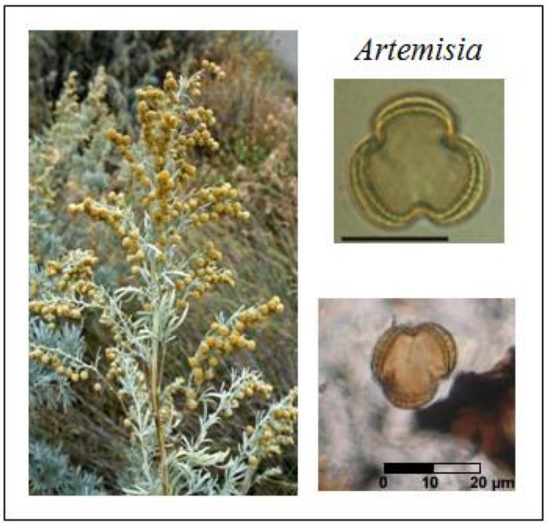
Ubiquist plants

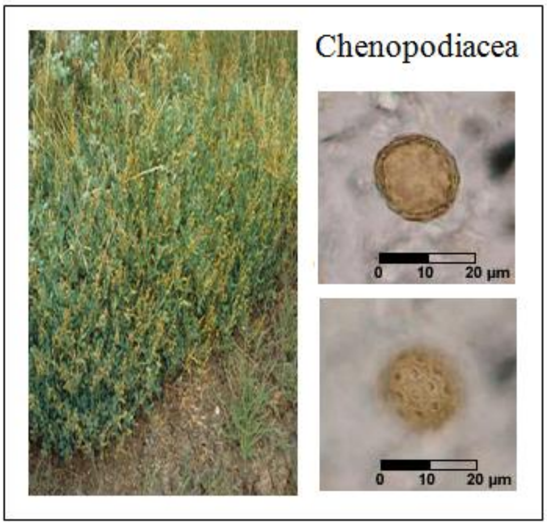


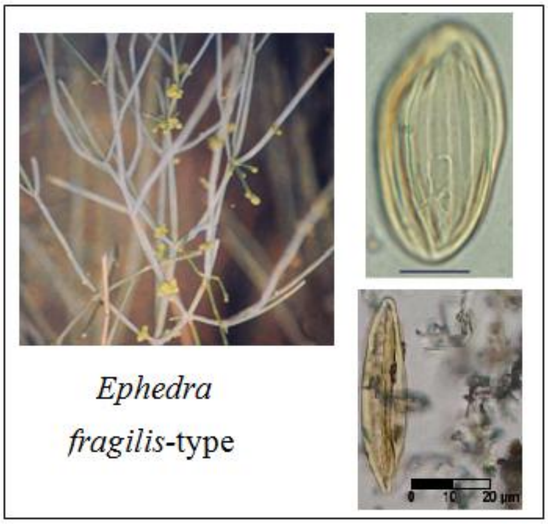




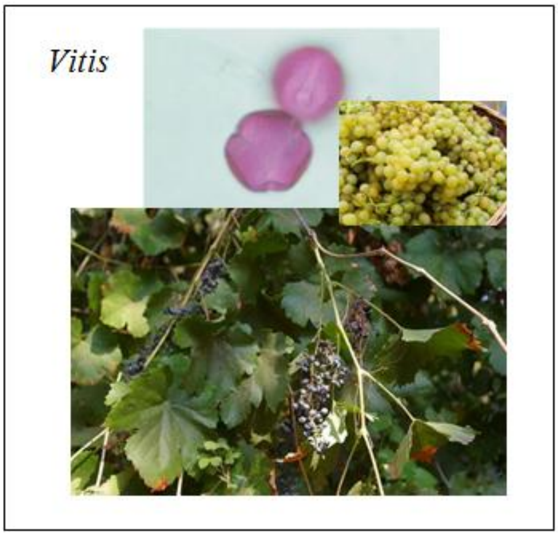
Semi-desert plants

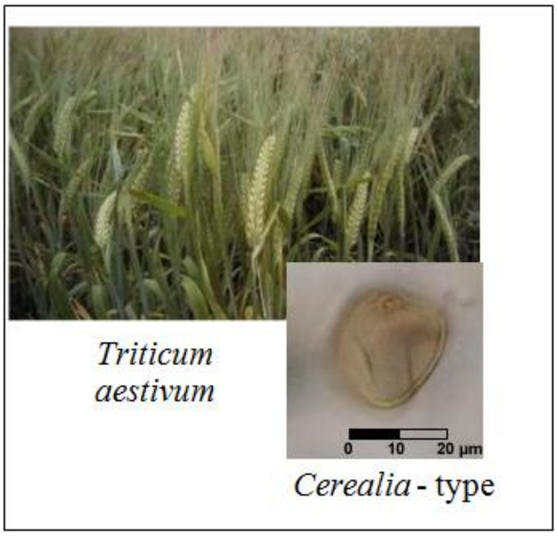


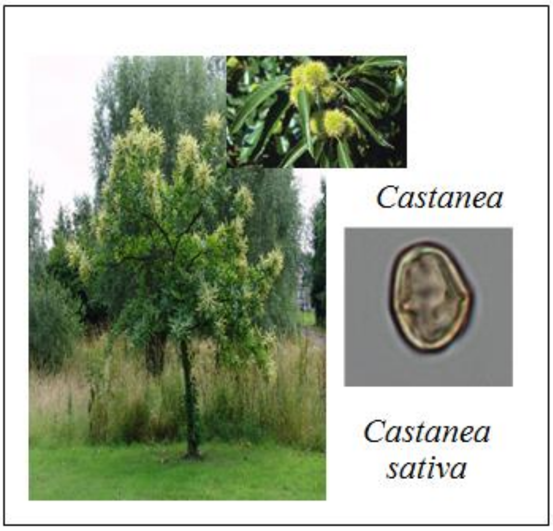




Anthropogenic pollen indicators







Plantago