Marc Oliva · Daniel Nývlt ·
José M. Fernández-Fernández
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Foreword: Periglaciation Past and Present

It is about time we renew our recognition that history matters. During the last half of the twentieth century, periglacial geomorphology evolved from a discipline integral to the general preoccupation of the earth sciences with the history of Earth to a process-based and process-monitoring study of environments undergoing seasonal freezing and thawing (Church 2017). The field left behind largely descriptive work to develop quantitative measurements, experimental techniques and analytical statements based upon physical principles (e.g., Lachenbruch 1962; Seppälä 1982; Mackay 1997). The time scales of interest for landform development in the periglacial realm range from a few hours, in the formation of needle ice (Outcalt 1970), to several centuries for the development of pingos (Mackay 1998). The field now includes physically based consideration of the active layer and permafrost.

For periglacial geomorphology, the transition from descriptive to quantitative research began about 1960. It is well illustrated by P.J. Williams's work on slope movement (1959) and J.R. Mackay's (1963) memoir on the Mackenzie Delta. These publications illustrate the application of *immanent*, i.e., universal, principles to understanding of the periglacial environment: in Williams's case it was the use of soil mechanics and for Mackay, hydraulics. These principles are inherently ahistorical. In many cases, especially over times scales of a few decades, immanent approaches are useful, as demonstrated by successful infrastructure engineering. Nevertheless, there are reported circumstances where environmental history has left a legacy that modifies infrastructure performance, perhaps catastrophically (e.g., Skempton et al. 1991). Simpson (1963) referred to contingent aspects of a landscape as comprising the unique circumstances, or *configuration*, that modify the choice of appropriate immanent principles to be applied to specific problems. He was thinking in terms of the time scales associated with geology, but the argument is just as sound when considered in the context of geomorphology. Configuration is inherently historical for it is the legacy of past events.

Configuration is a key element in understanding the periglacial environment, for ground-ice conditions are always a product of site history and they control the spatial variation of terrain responses to surface disturbance, including climate change (Mackay 1970; Kokelj et al. 2017; Burn et al. 2021). Similarly, soil texture principally governs the frost susceptibility of a soil, its capacity for frost heave (Burt and Williams 1976) and hence the development of hummocks and sorted circles (Mackay 1980; Hallet et al. 1988).

In North America and Russia, periglacial studies have been dominated by research on the permafrost environment, while in China, periglacial research on the Qinghai-Tibet Plateau has become very active in the last 25 years. In Europe, periglacial studies have been in progress for over a century and are the most catholic, ranging from investigations of active periglacial processes, including freeze-thaw shattering, solifluction, cryoturbation and permafrost creep in Scandinavia and the high mountains of the continent (e.g., Delaloye et al. 2010; Deprez et al. 2020), through sedimentary research on Quaternary stratigraphy (e.g., Murton and Kolstrup 2003), to the critical legacy of periglacial environments on *in situ* engineering behaviour of soils and on other elements of geomorphology (Hutchinson 1991; Ballantyne and Harris 1994). Europe is in some ways the periglacial continent. It most certainly is for those who carry what T.S. Eliot called the *historical sense*, that is an awareness of the presence of the past in what we see and encounter today.

At present, permafrost science and engineering are preoccupied with questions concerning thawing of the ground. The thawing period will extend over centuries for even thin permafrost (Burn 1998) and take millennia where ice rich ground is tens or hundreds of metres thick. Nevertheless, near-surface permafrost may thaw this century for there is little prospect of short-term control over climate change with the global atmospheric concentration of carbon dioxide already higher than at any point in the Quaternary (Burn et al. 2021). Periglacial environments may ultimately persist where the microclimate permits, especially at high elevations. Seen in this light, the periglacial record of Europe indicates the future of the world's permafrost environments. The treatment that the editors have brought together in this book may have been assembled primarily with a continental audience in mind, but it is relevant to the global community and has done us a great service.

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Preface

Joined to the vast Asian continent during the Uralian orogeny, Europe has been considered a peninsula of Asia, a western appendix. Its comparatively reduced size does not imply the simplicity of the territory but rather the opposite, manifested in many faces. This continent encompasses due to a long geological history a great heterogeneity of landscapes, from the eastern extensive plains and the Anatolian Plateau shaking hands with Asia to the peripheral mountains of the Balkan, Italian, Iberian and Scandinavian peninsulas. This continental wide spectrum of landscapes is complemented by the singular characteristics of several surrounding islands and archipelagos that extend beyond the boundaries of the continent to the North Atlantic, the Arctic and the southern Mediterranean Sea.

European landscapes are amongst the most difficult ones to be studied as Europe is in essence a humanised continent resulting from its long history and the superposition of different civilizations and ways to rationalise and use the environment. The compartmentation of the territory is enhanced by the numerous mountain ranges distributed across the continent, and is manifested in different cultures, languages and nations that have shaped the western and eastern cultures. Indeed, Europe has been often referred as the 'old continent', being not only the cradle of the western civilization, or the origin of the colonisers that extended its boundaries and influence the rest of the world, but also the origin of a great part of the Philosophy, rationalism, empiricism and the knowledge that feeds Science as a whole and many of the Earth Science disciplines such as Geography, Geology, Meteorology, Climatology, or Physics and Chemistry of the Earth, amongst many others. In fact, the origin of the periglacial research must be found in Europe, where the tradition on the study of the legacy of the mountain glaciations and cold processes in general is quite long. Indeed, this continent includes some of the most studied mountain areas (i.e. European Alps) on Earth.

However, within the realm of the cold region processes in Europe, significantly more attention has probably been paid to the glacial footprints and the reconstruction of the past glacial evolution as a mean to reconstruct former climate evolution. One could conclude that it is logical given to the obvious relationship between glaciers and climate, its visual and spectacular character and the vast areas that were covered by the Pleistocene ice sheets and mountain glaciers not only in Europe. However, glacial processes are telling us only part of the general picture of the cold climate areas: the glaciated areas and when there are (were) glaciers. In other words, what happens when glaciers disappear, or where the cold climate with insufficient precipitations does (did) not allow the formation of glaciers? That is a great opportunity to open the treasure chest of the periglacial landforms, sediments and landscapes; they may be not as spectacular as the glacial ones, but have also a great potential to study the climatic alterations in the past and at present-day focusing on major sentinels such as rock glaciers, palsas, pingos, solifluction lobes, patterned ground and many other landforms linked to different processes within the periglacial domain, affected by permafrost, or seasonal frost conditions. In addition, a better knowledge of these landforms and processes is crucial in a continent essentially humanised, where periglacial dynamics may cause serious damages to extensive urbanised areas, dense networks of communication and transport infrastructures and extensive industrial areas, especially in the Northern European and high-altitude regions.

Therefore, with this humble collective work that you have now in your hands we aim to collect and present the vast knowledge on the European periglacial landscapes and processes that has been accumulated during the last century. And to do it in a very didactic way, with lots of pictures, diagrams and schematic figures in order to reach an audience beyond the academia. We do not only intend to synthesise all the previous works, but to better understand their interrelationships and to uncover the current knowledge gaps and pose them as new challenges and opportunities to recover new hidden information on the past climate evolution and track the impact of the current climatic change. And last and not least, we aim to highlight the valuable potential of the periglacial landscapes and features, and the urgent need of protecting and preserving them for the next generations. As global change advances, this may be one of the last opportunities to raise awareness in society.

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Introduction

Marc Oliva, José M. Fernández-Fernández, and Daniel Nývlt

Periglacial research in Europe has progressed substantially over the last decades in parallel to a better comprehension of the past and present-day cold-climate geomorphological processes that have shaped the current landscape of this continent. Advances on glacial geomorphology have been complemented with a better characterisation of the spatial distribution of periglacial landforms and deposits in Europe as well as of their palaeoenvironmental and palaeoclimatic significance. In this sense, the progress on dating techniques have allowed to refine the chronology of the development of periglacial features, which is crucial to understand the time scales, climate conditions and geomorphological processes behind their formation.

Europe includes some of the best studied areas with regards to their periglacial evolution, which is demonstrated by the large amount of studies on periglacial processes and landforms from the Mediterranean basin to the polar regions, and from isolated Atlantic islands to the vast steppes of Eurasia. Thus, the objective of this book is to update the vast amount of information produced over the last few decades on the periglacial landscapes of Europe, synthesising the most recent advances on periglacial research across the European mountains and the lowlands once reshaped by cryogenic processes. The book pursues a double purpose: educational, including a large number of figures that complement the text and make easier the interpretation of the different chapters; and academic, being a reference for the future studies on the periglacial dynamics across the Old Continent.

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With that aim, this book includes an initial chapter where Filipa Naughton and co-authors present a reconstruction of the climatic evolution since the penultimate glacial cycle to the present day using a wide range of marine records from the Atlantic Ocean and Mediterranean Sea and ice core records from Greenland Ice Sheet; the climatic background is of crucial importance to understand the climatic phases that have been (un)favourable for periglacial activity in the past, and thus, to better frame the palimpsest of the current distribution of periglacial landforms and deposits.

The next part of the book comprises 12 chapters divided in three sections following major latitudinal regions: Southern, Central and Northern Europe. Periglacial evidence in each of these studied regions is examined by the leading scientists, who have long investigated these areas. The chapters focused on the different areas follow a similar structure; first, the main geographical and climatic characteristics of the study areas are analysed, together with the history of periglacial research in the area. Subsequently, a longer section (the core of each chapter) describes the distribution and types of periglacial phenomena existing in the region, trying to distinguish between the active ones and those inherited from past conditions. Finally, each chapter finishes with a conclusive section highlighting the unique characteristics of the region within the context of European periglacial landscapes.

The first section follows a W-E transect, from Iberia to Anatolia. Firstly, Marc Oliva and co-authors present the main periglacial features existing in the Iberian Peninsula, which are mostly concentrated in mountain ranges. Subsequently, Adriano Ribolini describes the processes and landforms of periglacial origin distributed across the Italian Peninsula, outside the Alps. Manja Žebre and Emil Gachev present the periglacial phenomena of the Balkans, outside the Carpathian Mountains, whereas Attila Çiner and Akif Sarikaya do the same for the Anatolian Peninsula. Finally, Mauro Guglielmin summarises the periglacial landforms existing in several Mediterranean islands, which are mostly inherited from past colder periods.

Periglacial phenomena in central Europe is examined in four chapters. Andreas Kellerer-Pirklbauer and co-authors provide a detailed analysis of active and relict periglacial features across the Alps, one of the regions where periglacial research has been more profusely developed. Later, Piotr Migoń and Jarosław Waroszewski examine the periglacial landscapes in the Variscan ranges, where the geomorphic and sedimentary records show evidence of past widespread periglacial environments. Finally, Zofia Rączkowska presents the impact that periglacial morphogenesis has had on the mountain landscapes of the Carpathians. And finally, Barbara Woronko and Maciej Dąbski examine the implications that periglacial remodelling had on the landscape of the North European Plain during Quaternary cold stages.

Three chapters comprise the section on Northern Europe. Firstly, Colin Ballantyne and Julian Murton describe the rich variety of landforms and deposits that periglaciation left in Great Britain and Ireland, including minor active features. John Matthews and Atle Nesje examine the periglacial landscapes of Scandinavia, where active and relict landforms are widespread and have been a reference for periglacial research in Europe. Finally, José M. Fernández-Fernández and co-authors summarise how periglacial processes have shaped in the past and continue to shape nowadays the very dynamic landscapes of Iceland.

The book concludes with a chapter prepared by the editors that seeks to summarise the impact of periglacial dynamics in European landscapes, from Late Pleistocene times to present-day, including how human activities have transformed periglacial landscapes.

In short, this book does neither constitute a compendium of study cases nor a handbook of periglacial dynamics, but it is designed as an interlaced connection of concepts and examples across Europe with a rich and well-illustrated material. This approach will be useful for academicians working in similar topics as well as for non-academicians to better understand the landform features existing in their regions. Indeed, the orographic complexity of Europe, with unevenly distributed mountain ranges, peninsulas, bays, islands, etc. combined with Quaternary climate variability and advances and retreats of glaciers and ice sheets has resulted in the variety and uniqueness of periglacial landscapes in a continent, where these landscapes should be preserved for future generations.

Regional Setting

Quaternary Climate Variability and Periglacial Dynamics

Filipa Naughton, Maria Fernanda Sánchez Goñi, and Samuel Toucanne

1 Climatic Framework of Europe During the Last Glacial Cycle

The Last Glacial Cycle (LGC, ~116–15 ka) encompasses the most recent glacial period of a series of 10 glacial-interglacial cycles that occurred during the last ca. 1 million years. The cycles from 700 ka onwards, identified from changes in the marine oxygen isotopic ratio of calcite shells (δ^{18}Oc), are characterised by a dominant 100,000-year cyclicity originally triggered by changes in insolation (Shackleton and Opdyke 1973). During the last 800 ka, they are represented by 20 Marine Isotope Stages (MIS) from MIS 20 to MIS 1, where odd and even numbers broadly correspond to interglacial and glacial periods, respectively (Lisiecki and Raymo 2005; Railsback et al. 2015). After attaining a minimum of ice volume extent during Last Interglacial (MIS 5e), the LGC started at ~116 ka, the so-called last glacial inception (MIS 5e/5d transition). The ice volume progressively increased through MIS 5 and an additional decrease in summer insolation triggered the ice volume maxima, -80 m of relative sea level, centred at 65 ka during MIS 4 (Waelbroeck et al. 2002) (Fig. 1). The subsequent increase in summer insolation produced a partial deglaciation giving way to the MIS

The original version of this chapter was revised: The missing letter 'a' in section heading has been corrected. The correction to this chapter is available at https://doi.org/10.1007/978-3-031-14895-8_17

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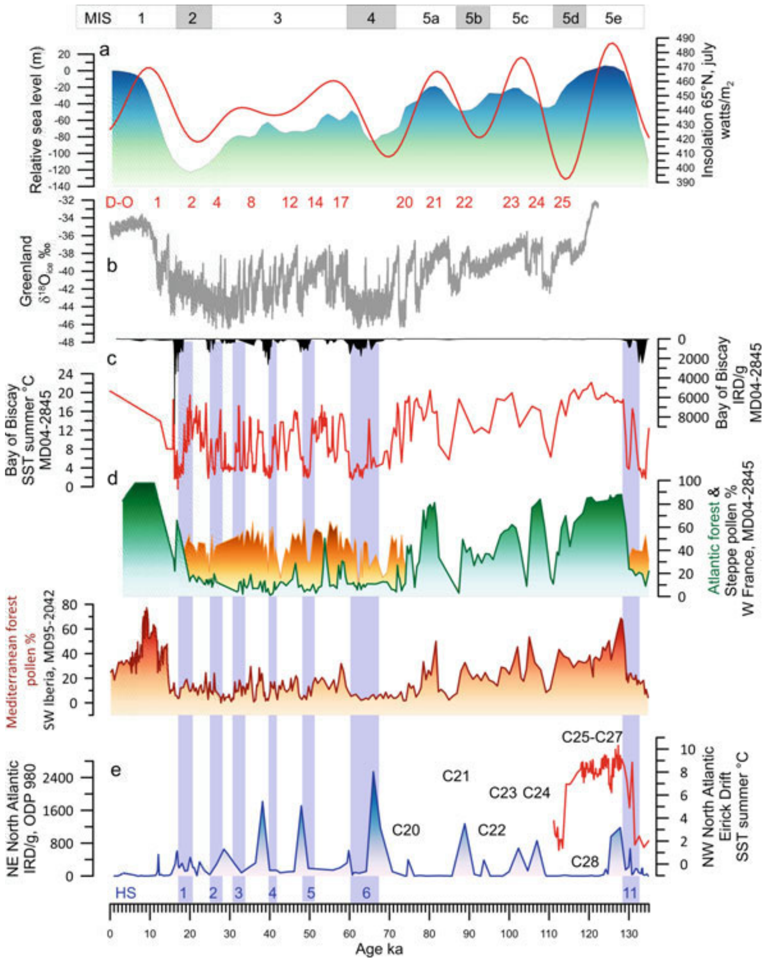

Fig. 1 Millennial scale palaeoenvironmental changes in Europe, Greenland and the North Atlantic during the last climatic cycle. a) July insolation at 65° N (red) (Berger and Loutre, 1991) and relative sea-level changes as an indicator for ice-volume changes (Waelbroeck et al. 2002), b) Greenland ice core $\delta^{18}O$ record (Rasmussen et al. 2014). Dansgaard-Oeschger (D-O) warming events (red), c) Foraminifera-based summer sea surface temperature (SST, red) and Ice Rafted Debris (IRD, black) records of core MD04-2845 from the Bay of Biscay (45°N) (Sánchez-Goñi et al. 2008), d) Atlantic temperate forest, steppe (orange, mainly Poaceae, Cyperaceae and Asteraceae) pollen percentage records from the Bay of Biscay (western France) and Mediterranean forest pollen percentage records from the SW Iberian margin (Sánchez-Goñi et al. 2008 and unpublished data), e) IRD concentration record from site ODP 980 (blue), NE North Atlantic. Grey intervals indicate the HS 6 to 1 and HS 11 and the cold events C20 to C24 (Bond and Lotti 1995; McManus et al. 1994). Summer SST from the NW North Atlantic indicating the cold events of the Last Interglacial C27 to C25 (Irvalı et al. 2016). MIS: Marine Isotope Stages. Cold events C19 and C18 are not detected in ODP site 980 (McManus et al. 1999) but recorded in the nearby site V29-191 (McManus et al. 1994). Cold event C28 follows HS 11 although the chronological uncertainties of both ODP 980 and the Erik Drift sites do not show it in the figure

3 intermediate ice volume period lasting from ~60 ka to 27 ka. The maximum global ice volume of the LGC was reached during MIS 2, 26–19.5 ka (Clark et al. 2009), and preceded the last deglaciation, composed of a series of climate shifts between 19.5 ka and 11.7 ka.

Geomorphological data and model simulations show that at ca. 115 ka ice sheets developed over Scandinavia and in Northern Eurasia, particularly during MIS 4 (73–60 ka), whilst large ice sheets close to Atlantic moisture sources, i.e. the Western European Ice Sheet (EIS), reached their maximum extent at the global Last Glacial Maximum (LGM, 26–19.5 ka) (Hughes et al. 2013; Batchelor et al. 2019; Palacios et al. 2022). Ice-sheets in Northern Eurasia were probably of similar size or even more extensive in MIS 5b (~85 ka) and MIS 5d (~115 ka) compared to MIS 4. Data and models also suggest that western Europe was more glaciated during the LGM than Eastern Europe and Northern Eurasia. Also, the partial marine-based nature of the EIS during the LGC seems to be marked by a higher susceptibility to rapid and frequent ice-sheet collapse (Batchelor et al. 2019). This high susceptibility is associated with the strong variability of the Greenland climate revealed by the analyses of water isotopes in deep ice cores from the centre of the Greenland Ice Sheet, known as the Dansgaard-Oeschger cycles (Dansgaard et al. 1984; Johnsen et al. 1992; Rasmussen et al. 2014) (Fig. 1). These cycles lasted 500–4000 years during MIS 3 and MIS 4, but were several millennia longer during MIS 5d-a (Rasmussen et al. 2014), supporting that climate was particularly variable during the middle part of the LGC (Bond et al. 1999). The abrupt transitions between the cold periods identified by low oxygen isotope ratios (δ^{18}Oice) to mild periods with high δ^{18}Oice are designed as Dansgaard-Oeschger (D-O) warming events. The LGC was punctuated by twenty-five D-O warming events in the atmosphere over Greenland, and were characterised by a high amplitude (7–16 °C) and rapid (within a few decades) warming event followed by a progressive decrease in temperature and a final abrupt cooling (Wolff et al. 2010). The D-O warming event and the progressive cooling phase form the Greenland Interstadial (GI), and the final cooling event leading to the cold phase form the Greenland Stadial (GS). The GI phases lasted between 100 and 2,600 years (Wolff et al. 2010; Rasmussen et al. 2014). A wide array of global climatic proxies such as CO_2 and CH_4 atmospheric concentrations from the air bubbles trapped in the Antarctic and Greenland ice cores have furthermore demonstrated the link between the initial observations in Greenland and the global expression of the D-O cycles (e.g. Brook et al. 1996). Marine sedimentary data also indicate that the D-O cycles were associated with 15–30 m of global sea-level increase during GI (Sierro et al. 2009).

Interestingly, the study of ice-rafted debris (IRD) found in North Atlantic deep-sea sedimentary sequences show that the LGC was also punctuated by six massive and repeated episodes of iceberg discharges, every ~7000 years, from the north-eastern Laurentide Ice Sheet and the Hudson Strait ice-stream. These Heinrich events (HE), as they are called, substantially cooled the surface of the North Atlantic (Heinrich 1988; Bond et al. 1993) (Fig. 1). A HE is defined as the event of the North American ice-rafted debris deposition resulting from the melting of the icebergs in the North Atlantic, preferentially between 45° N and 50° N, while the Heinrich Stadial

(HS) is the up to 3,000 years cold phase, in which occurred the HE (Sánchez Goñi and Harrison 2010). Moderate iceberg discharges from the Laurentide Ice Sheet, associated with the North Atlantic cold events C27 to C18 (Fig. 1), occurred during MIS 5 and MIS 4 (McManus et al. 1994; Shackleton et al. 2003; Oppo et al. 2006) but, paradoxically, those punctuating the MIS 5a/4 transition, C20 to C18 associated with GS 21 to GS 19, corresponded with warm sea surface temperatures (SST) in the mid-latitudes of the Western European margin (Sánchez Goñi et al. 2013). Weak iceberg discharges from the Icelandic, British-Irish, and Scandinavian Ice Sheets and with higher frequency than the HE also occurred between the HEs (Elliot et al. 2001) associated with the other GSs. Iceberg discharges have been simulated to last from 50–200 years (Roche et al. 2004) up to 1000–1500 years (Ziemen et al. 2019). Most of the HSs and non-HSs are related to cold sea surface conditions in the North Atlantic Ocean, and they are associated with decreases in the strength of the Atlantic Meridional Overturning Circulation (AMOC) (Henry et al. 2016; Lynch-Stieglitz, 2017). The AMOC is defined as the zonally integrated component of surface and deep currents in the Atlantic Ocean and it plays a crucial role in the Earth's climate by regulating the global transport of heat and freshwater. The tropical warm and saline surface waters flow northward densifying progressively and sink preferentially in the Norwegian Sea, between north of Iceland and south of Svalbard, and Irminger seas (Lozier et al. 2019). They release heat and humidity toward the atmosphere and form a southward flow of colder, deep waters that are part of the thermohaline circulation.

Iceberg discharges substantially slowdown the arrival of warm waters to Northern latitudes decreasing the deep-water formation and consequently cooling and drying the North Atlantic Ocean and the atmosphere (Ganopolski and Rahmstorf 2001). However, the fact that HEs occurred during a complex and longer period, i.e. the HS, and that lags the sea surface cooling (Barker et al. 2015) indicate that HE may be a consequence of AMOC weakening. This initial weakening, the origin of whose remains a subject of debate, may result from enhanced ice surface melting of the European and Pacific proximal ice settings (Boswell et al. 2019; Toucanne et al. 2022). The latter could cause, in turn, oceanic sub-surface sea warming, the ultimate destabilisation of the North Atlantic marine-terminating ice streams and the massive iceberg discharges corresponding to the HE (Álvarez-Solas and Ramstein 2011).

The expression of D-O cycles and HSs in European environments has been investigated by analysing speleothems, loess sequences and the pollen grains preserved in terrestrial and deep-sea cores, the latter allowing a direct comparison between D-O cycles and HSs. The general picture observed in Europe is a significant change in vegetation cover and soil characteristic between stadial and interstadial (Fletcher et al. 2010a). The stadials were characterised by much drier and colder conditions (steppe vegetation, loess development in continental Europe (Moine et al. 2017) and the interstadials were associated with forest expansion (Fig. 1). The Abrupt Climate change and Environmental Responses (ACER) pollen data synthesis (Sánchez Goñi et al. 2017) reveals the following features for the European vegetation during the succession of D-O events. Forest dominated Europe at latitudes lower than 40° N as well as north of 40° N during warmer and long interstadials (i.e., those associated with D-O events 12 and 14). In Western Europe pollen data reveal that the amplitude of Atlantic and Mediterranean forest expansions differs for any given D-O warming during the period encompassing MIS 4, 3 and 2 (73–14.7 ka). In the Western Mediterranean below 40° N, D-O 16–17 and D-O 8 and 7 were associated with strong expansion of forest cover, probably amplified by the high seasonality induced by the minima in precession at that time, contrasting with weak expansion of forest cover during D-O 14 and 12. The opposite pattern is revealed at the Atlantic sites, where the strongest forest development during D-O 12 and 14 was amplified by the maximum in obliquity affecting latitudes above 40° N (Sánchez Goñi et al. 2008). Some modelling simulations, in qualitative agreement with pollen data, indicate that the European vegetation responded quickly, by ~200 years, to the abrupt D-O warming events generated by the strengthening of the AMOC (Woillez et al. 2013). Associated with these vegetation changes, repeated inner-alpine ice fluctuations during D-O cycles of MIS 3 are also observed (Mayr et al. 2019; Martinez-Lamas et al. 2020).

As mentioned above, during the last interglacial-glacial transition (MIS 5a/4) the regional SST in the Western European margin remained relatively high, while those on the adjacent continent, identified by the replacement of temperate by boreal forest, were relatively cool and progressively decreasing. Superimposed on this long-term change, three cold-driven boreal forest regional expansions during GS 21 to 19 alternated with three temperate forest expansions during GI 21 to 19. In NW Iberia, cooling episodes were marked by the development of heathlands that alternated with oak forest increase during the warming events (Sánchez Goñi et al. 2013). During the short-term atmospheric cooling events SSTs, paradoxically, also increased or remained warm, at around 18 °C. Moderate North American iceberg discharges in the North Atlantic certainly deviated pinned the warm Gulf Stream and North Atlantic Current southwards to the European margin as described for example in Keffer et al. (1988) producing an increase in the land-sea thermal contrast and enhancing evaporation. Carried northward by the westerly winds, this moisture was the most likely source of the snowfall that formed the ice caps during MIS 4 (Sánchez Goñi et al. 2013). The cold-land warm-sea decoupling documented for the cooling events of the last glaciation in Western Europe from 42° N to 45° N is distinctive and promoted the regional development of boreal forest and heathlands. In contrast, the environments of Southern Iberia were dominated by semi-desert during the cold phases, which intervened between increases in Mediterranean woodland reflecting warm summers and humid winters.

Well-chronologically constrained deep-sea and terrestrial pollen records have shown furthermore a complex phase of sequences within the HSs and the contrasting regional impact of HEs on the European continent. Regarding the magnitude of the HE cooling, the NW Mediterranean borderlands (SE France and NE Iberia) experienced colder and drier conditions during HS 5 compared to HS 4 leading to a higher expansion of steppe vegetation. By contrast, during HS 4, the massive freshwater input in the North Atlantic may have led to the stratification of the Mediterranean water column and consequently limited upward mixing of cold water, resulting in regional atmospheric warming and wetting compared to HS 5 that allowed the maintenance of some temperate deciduous oak woodland stands (Sánchez Goñi et al. 2020). Similarly, HS 5 appears to be particularly cold and dry in NE Greece as shown by the dominance of steppe formations (Müller et al. 2011). In contrast, an area of relative ecological stability has been identified in SW Greece, where temperate tree populations survived throughout the LGC as the result of continued moisture availability and varied topography (Tzedakis et al. 2002). Regarding the complex structure of HSs, a wet to dry continental hydrological pattern has been identified in the Iberian Peninsula (Fletcher and Sánchez Goñi 2008; Naughton et al. 2009). In the north-Western Mediterranean region, a three-phase structure, wet-mild/dry-cold/wet-mild intervals, have been detected in the Gulf of Lyon and the adjacent landmasses (Sánchez Goñi et al. 2020). The multiphase structure of the HSs is also supported by both the Northern North Atlantic (Wary et al. 2018) and Greenland ice core records (Guillevic et al. 2014).

Despite the three last decades of research, the mechanisms at the origin of D-O cycles and HEs still remain elusive. A key role has been attributed to the variations of the AMOC as well as the Arctic sea ice. An important challenge in the modelling community is to be able to identify the original forcing, internal or external, of such a climatic variability and to make the link with the HSs (Landais et al. 2022). If the cause of the HSs remains enigmatic, the production of meltwater by mid-latitude proximal ice sheets from Europe and North America, could have triggered the initial AMOC weakening. This ice melting necessarily implies warming summer temperatures as cold conditions in the North Atlantic region and over Greenland are recorded at the same time. The mismatches between ice-core temperature oscillations in the polar regions and ice-margin fluctuations in mid-latitudes would, therefore, call for a paradigm shift in our understanding of past rapid climate changes (Toucanne et al. 2022).

2 Climate Changes During the Last Deglaciation

The last deglaciation was triggered by the increase of summer insolation at the Northern Hemisphere and consequent retreat of North American and Eurasian ice sheets, the increase in greenhouse gas concentrations and changes in other amplifying feedbacks that cause distinctive regional and global responses (Fig. 2) (e.g.; Clark et al. 2012; Shakun et al. 2015). Globally, the last deglaciation started at ~20–19 ka (ka: age in kiloanni) (e.g.; Denton et al. 2010; Carlson and Clark, 2012; Clark et al. 2012) and ended during the present day interglacial at ~6.8 ka (e.g.; Carlson et al. 2007, 2008). Multiple abrupt climate shifts were noticed in the North Atlantic, Greenland and Europe, superimposed on this long-term warming trend (Fig. 2) (e.g.; Mangerud et al. 1974; Dansgaard et al. 1982; Oeschger et al. 1984; Mix and Ruddiman, 1985; Alley et al. 1993; Alley and Clark, 1999). The most recognised abrupt climate episodes that punctuated the last deglaciation are the Heinrich Stadial 1 (HS 1, ~18–14.7 ka), the Bølling-Allerød interstadial (B-A, ~14.7–12.9 ka) and the Younger Dryas (YD, 12.9–11.7 ka) (Fig. 2).

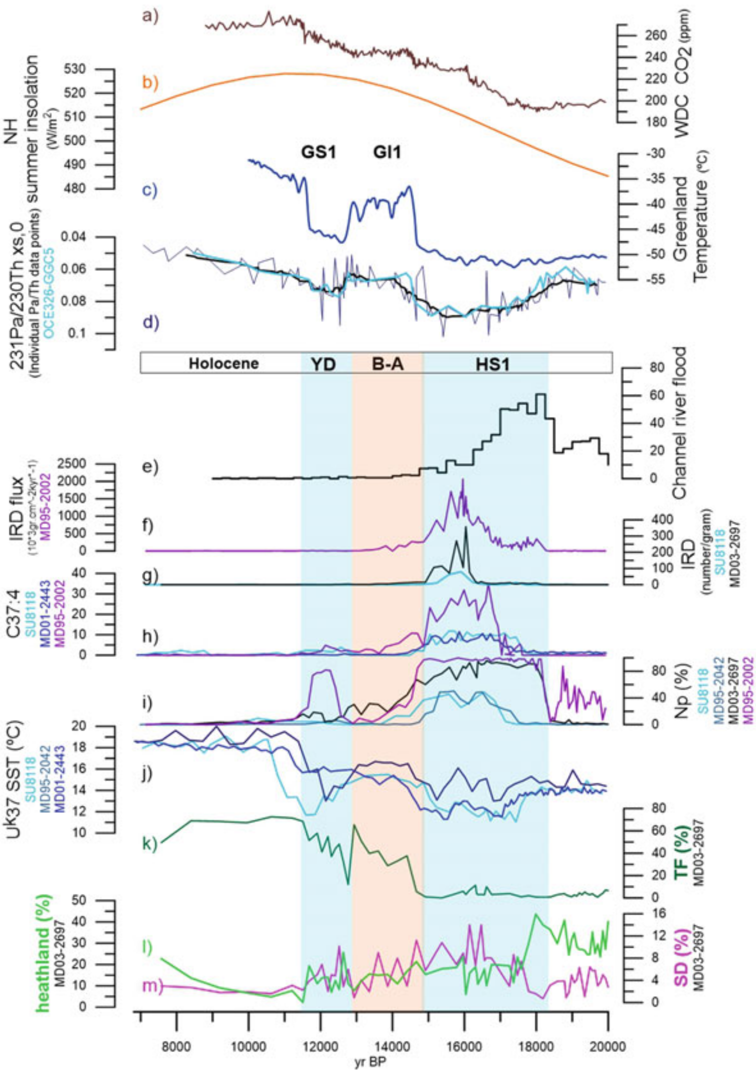

Fig. 2 a) Atmospheric CO_2 concentrations reconstructed from West Antarctic Ice Sheet Divide ice core (WDC) (Marcott et al. 2014); b) summer insolation at 65ºN (Berger and Loutre 1991); c) Greenland temperature (Buizert et al. 2014a, b); d) ex231Pa0/ex230Th0 from composite North Atlantic records (blue line) (black bold line: smoothed record) (Ng et al. 2018) and from the western North Atlantic (light blue) (OCE326-GGC5; McManus et al. 2004); e) Channel River Flood (number of flood events per 250 years) (Toucanne et al. 2015); f) IRD western French margin (Ménot et al. 2006) and g) in the western Iberian margin (Bard et al. 2000; Naughton et al. 2016); h) Meltwater discharges (Tetra-unsaturated alkenone $C_{37:4}$) from western Iberian and French margins (Bard et al. 2000; Ménot et al. 2006; Martrat et al. 2007); i) NP: planktic polar foraminifera *Neogloboquadrina pachyderma* abundances (%) from western Iberia and French margins (Salgueiro et al. 2014; Zaragosi et al. 2001); j) alkenone-derived SST records ($U^{k'}_{37}$ SST) from Iberian margin (Bard et al. 2000; Pailler and Bard, 2002; Martrat et al. 2007); k) TF: Temperate forest abundances (%) in NW Iberian margin record MD03-2697 (Naughton et al. 2016); l) Heathland; m) SD: Semi-desert plants. Light blue bands: HS 1 and YD; Light salmon band: B-A

2.1 Heinrich Stadial 1 (HS 1)

HS 1 is a long stadial episode (~18 to 14.7 ka) that includes two main phases: an early phase associated with increased surface melting of the Southern, land-terminating margins of the Laurentide (LIS) and Fennoscandian Ice Sheet (FIS) and a latest phase during which their marine-terminating ice-streams rapidly collapsed (e.g.; Barker et al. 2009, 2015; Sánchez Goñi and Harrison 2010; Toucanne et al. 2015; Boswell et al. 2019). The later corresponds to the so-called Heinrich event 1 (HE 1) (e.g.; Hemming, 2004). HE 1 was firstly identified in deep-sea cores located in the 'Ruddiman belt' (~43–53° N), by the anomalous presence of ice-rafted detritus (IRD), resulting from the collapse of Eurasian and Laurentide Ice Sheets and consequent debris-rich icebergs drifting and melting in the North Atlantic (e.g.; Heinrich 1988; Broecker et al. 1992; Grousset et al. 1993; Elliot et al. 1998; Hemming, 2004). Thin layers with IRD were also found outside the 'Ruddiman belt', at latitudes north of 53° N (e.g., Rasmussen et al. 1996; Elliot et al. 1998, 2001; Voelker et al. 1998; Van Kreveld et al. 2000) and South of 43° N (e.g.; Lebreiro et al. 1996; Bard et al. 2000; Chapman et al. 2000; de Abreu et al. 2003; Fletcher and Sánchez Goñi 2008; Naughton et al. 2009; 2016).

For long time climate modellers have been focused in exploring the impact of the melting icebergs in the North Atlantic region, i.e. during HE 1 (e.g.; Rahmstorf 1994; Seidov and Maslin 1999; Ganopolski and Rahmstorf 2001; Menviel et al. 2014). These simulations suggest that freshwater pulses in the North Atlantic and Nordic seas could easily force the AMOC to shut down, leading to the idea that iceberg calving (HE) could be the cause for Heinrich stadial (HS) conditions. However, others support evidence that increased iceberg calving could rather enhance and/or prolong cold stadial conditions (through a positive feedback on the AMOC in response to the addition of freshwater in the North Atlantic), suggesting that HE 1 was a consequence rather than the cause of meltwater pulses (McManus et al. 1999; Clark et al. 2007; Barker et al. 2015). Data and simulations further support the idea of a subsurface oceanic warming, under a reduced AMOC situation, as a possible trigger for the melting of the circum North Atlantic marine-terminating ice-streams and the subsequent HE 1 (Alvarez-Solas et al. 2010; 2013; Marcott et al. 2011; Bassis et al. 2017; He et al. 2020).

Most North Atlantic records show an extreme cooling (Fig. 2) contrasting with other data and climate simulations that indicate warm conditions at the onset of HS 1 (Fig. 2). In particular, SST profiles and the expansion of polar planktonic foraminifera *Neogloboquadrina pachyderma* reveal extreme cold conditions during the early phase of HS 1 in the North Atlantic and Mediterranean regions (e.g.; Bond and Lotti, 1995; Bard et al. 2000; Zaragosi et al. 2001; Hemming, 2004; Peck et al. 2006; Eynaud et al. 2007; 2009; Naughton et al. 2009; 2016; Penaud et al. 2009; Martrat et al. 2014). In contrast, Greenland ice core oxygen isotope $\delta^{18}O$ records do not show any cooling at the onset of HS 1 (He et al. 2021). Increased seasonality, with warm summers and severe cold winters have been proposed to explain this thermal contrasting signal (e.g.; Bromley et al. 2014a, b; Toucanne et al. 2015;

Boswell et al. 2019; Wittmeier et al. 2020; Fersi et al. 2021; He et al. 2021). During summer, warm atmospheric conditions would have favoured the surface melting of land-terminating margins and the resulted input of meltwater in the North Atlantic would have triggered the AMOC to almost shutdown (e.g.; Broecker 1994; McManus et al. 2004; Toucanne et al. 2015; Ng et al. 2018), favouring the expansion of sea ice during winter and leading to a cooling of the surface ocean (Denton et al. 2010).

Reduced temperate forest and a strong *Pinus* forest decline revealed by Iberian margin deep-sea pollen records testify the extreme cooling in Iberia at the onset of HS 1 (Fig. 2) (Roucoux et al. 2005; Fletcher and Sánchez Goñi 2008; Naughton et al. 2009; 2016). Similarly, speleothems from Southern France and Spain indicate atmospheric cold conditions in Western Europe during this early phase of HS 1 (Genty et al. 2006; Moreno et al. 2010). The cold signal might be likely the result of prolonged and extreme cold winters that result from the AMOC reduction and widespread winter sea-ice in the North Atlantic. This would lead to substantial changes in the atmospheric circulation patterns, as revealed by the complex spatial distribution of precipitation with contrasting wet or dry signals detected in the Iberian margin and speleothem records (e.g.; Genty et al. 2006; Naughton et al. 2009; 2016; Moreno et al. 2010; Pérez-Mejías et al. 2021).

The second phase of HS 1 (onset: ~16.2 ± 0.3 ka; Landais et al. 2018 and end: 14.7 ka Rasmussen et al. 2014), associated with the collapse of the LIS and EIS and marine terminating ice-streams, and consequent massive iceberg release in the North Atlantic, marks the initially defined HE 1 (e.g.; Hemming, 2004; Andrews and Voelker 2018). This phase is marked by large deposits of dolomite-rich IRD from LIS sources in the North Atlantic deep sea records, known as Heinrich layer 1 (e.g.; Bond et al. 1992; Broecker et al. 1992; Hemming 2004; Barker et al. 2009; Hodell et al. 2017; Andrews and Voelker 2018). The introduction of large quantities of meltwater, from the LIS and EIS marine-terminating ice-streams in the North Atlantic contribute to the sustained reduction of the AMOC and amplified North Atlantic cooling (Fig. 2) (Bond et al. 1992; Cortijo et al. 1997; McManus et al. 2004; Eynaud et al. 2007; Scourse et al. 2009; Ng et al. 2018). Cold winter SST persisted since the onset of HS 1, in the Iberian and French margins (Bard et al. 2000; Pailler and Bard, 2002; Naughton et al. 2009; 2016; Fersi et al. 2021). In Iberia, semi-desert plants expanded during the maximum deposition of IRD's and contracted towards the end of HS 1, suggesting a drying phase followed by moisture increase (Fig. 2) (e.g.; Fletcher and Sánchez Goñi 2008; Naughton et al. 2016). Several speleothem records from Northern Iberia and SW France show very cold and dry regional conditions during HE 1 (e.g.; Genty et al. 2006; Moreno et al. 2010).

2.2 *The Bølling-Allerød Interstadial (B-A)*

The B-A interstadial was the first warm and wet phase of the last deglaciation in the Northern Hemisphere that occurred after HS 1, i.e. between 14.7 ka and 12.9 ka (Fig. 2) (e.g.; Severinghaus and Brook, 1999; Hoek, 2009; Denton et al. 2010;

Rasmussen et al. 2014; Naughton et al. 2016). It was firstly noticed in two Danish records (the Bølling and the Allerød sites) based mainly on birch tree remains (Hartz and Milthers 1901; Iversen 1942, 1954). The presence of the alpine species *Dryas octopetala* in two distinct layers within the B-A revealed that this warm phase was however punctuated by the Older Dryas and the intra-Allerød abrupt cooling episodes (e.g.; Mangerud et al. 1974; Von Grafenstein et al. 1999; Brauer et al. 1999; Hoek, 2009). The B-A is marked by an abrupt temperature increase of about 4 to 10 ºC in the North Atlantic region, as revealed by SST estimates, the reduction of polar foraminifera *N. pachyderma* (Fig. 2) (e.g.; Bond and Lotti 1995; Bard et al. 2000; Cacho et al. 2001; Zaragosi et al. 2001; Pailler and Bard 2002; Martrat et al. 2007; 2014; Rodrigues et al. 2010; Salgueiro et al. 2014; Repschlager et al. 2015; Naughton et al. 2016) and the expansion of subtropical foraminiferal assemblages in the sea surface water masses of the mid-latitudes (e.g.; Chapman et al. 2000; de Abreu et al. 2003). An abrupt increase of 10 ºC, D-O warming event 1, followed by a gradual decrease in atmospheric temperatures are noticed in the Greenland ice cores (Buizert et al. 2014a, b). An increase of ~3 to 5 ºC in atmospheric summer temperatures is reconstructed for Northern and Southern Europe (Renssen and Isarin, 2001; Dormoy et al. 2009). Thus, the B-A is considered the terrestrial counterpart of the D-O 1 over Greenland (Fig. 2) (Rasmussen et al. 2014).

The B-A warming of the Northern Hemisphere was triggered by the increase of boreal summer insolation and CO_2 atmospheric concentrations since the last stages of HS 1, together with the strengthening of the AMOC (Fig. 2) (e.g.; Severinghaus and Brook 1999; Liu et al. 2009; Denton et al. 2010; Shakun et al. 2012; Zhang et al. 2017; Obase and Abe-Ouchi 2019). This warming favoured forest development in Europe (e.g.; Litt and Stebich, 1999; Hoek, 2009; Fletcher et al. 2010b; Naughton et al. 2016 and references therein), but its maximum expansion occurred latter during the Allerød, suggesting a deficit in moisture availability at the onset of the B-A (e.g.; Naughton et al. 2016 and references therein). This moisture availability deficit is also supported by Southern Europe speleothem records and pollen-based quantitative climate estimates from the Western Mediterranean region (Genty et al. 2006; Dormoy et al. 2009; Moreno et al. 2010). The displacement to the north of the polar front and jet stream (Eynaud et al. 2009; Naughton et al. 2016) with a more meandered configuration (Naughton et al. in press) would explain the deficit in precipitation at the onset of the Bølling in Europe (Naughton et al. 2016).

2.3 The Younger Dryas (YD)

The YD is the latest stadial event of the last deglaciation in the Northern Hemisphere, that occurred after the B-A and prior to the onset of the Holocene, i.e. between 12.9 ka and 11.7 ka, (Fig. 2) (e.g., Alley et al. 1993; Alley and Clark 1999; Tarasov and Peltier, 2005; Denton et al. 2010; Carlson, 2013). This stadial was firstly noticed in Scandinavian terrestrial records by the rapid development of the cold-tolerant *Dryas octopetala* (Iversen, 1954; Mangerud et al. 1974; Mangerud, 2021). The YD started at the same time as the Greenland Stadial-1 (GS-1) (e.g., Rasmussen et al. 2014; Naughton et al. 2019; Mangerud, 2021; Reinig et al. 2021), or was delayed about ~170/185 yr from the onset of GS-1 (e.g.; Brauer et al. 2008; Rach et al. 2014; Obreht et al. 2020) depending on the time response of certain climate indicators (e.g., Mangerud, 2021; Naughton et al. 2022).

Several hypotheses have been proposed to explain the causes of the YD. Some of them support that meltwater discharges from the LIS and/or FIS into the North Atlantic or Arctic Ocean (e.g.; Tarasov and Peltier 2005; Carlson et al. 2007; Murton et al. 2010; Muschitiello et al. 2015), caused by maxima of incoming solar radiance in the Northern Hemisphere, leading to substantial changes in oceanic and atmospheric circulation patterns (e.g.; Broecker 2003; McManus et al. 2004; Wünsch, 2006; Eisenman et al. 2009; Carlson, 2013; Ritz et al. 2013). Others support evidence of an extra-terrestrial (ET) impact event on or near the LIS that caused the above mentioned freshwater discharges (e.g., Firestone et al. 2007; Kennett et al. 2009; Sweatman 2021). Another proposal, based on numerical climate models, suggests that the YD results from combining processes such as AMOC slowdown, moderate negative radiative forcing and an altered atmospheric circulation (Renssen et al. 2015).

It is also known that both the AMOC slowdown and reduction of the North Atlantic Deep Water (NADW) formation, during the YD, have favoured the expansion of the winter sea ice and contributed for a cooling of 2 to 4 ºC and drying in the North Atlantic and over Europe (e.g.; Boyle and Keigwin 1987; Manabe and Stouffer 1997; Isarin et al. 1998; Rahmstorf 2002; McManus et al. 2004; Denton et al. 2005; Shakun and Carlson, 2010; Renssen et al. 2015; Ng et al. 2018), while of 5 to 9 ºC in Greenland ice records (Buizert et al. 2014a, b).

North Atlantic and European very high temporal resolution records have shown, however, that the impact of the YD climate was marked by a series of climate shifts with a complex hydroclimate spatial distribution in Europe in each phase (a, b, c and d) (Fig. 3) (e.g.; Magny and Begéot 2004; Brauer et al. 2008; Bakke et al. 2009; Lane et al. 2013; Rach et al. 2014; Baldini et al. 2015; Bartolomé et al. 2015; Naughton et al. 2019). They suggest that these climate shifts, with complex spatial moisture distribution across Europe, could not be explained by AMOC slowdown alone, but also by coupled changes in atmospheric circulation patterns. They particularly suggest that the combining changes of the AMOC strength and the North Atlantic sea-ice extent produced substantial latitudinal shifts of the polar jet stream and, therefore, changes in the position and strength of the westerlies across Western Europe that are in turn responsible for sea-land moisture transfer (e.g.; Renssen and Isarin 1997; Magny and Begéot 2004; Brauer et al. 2008; Bakke et al. 2009; Lane et al. 2013; Rach et al. 2014; Naughton et al. 2019; Rea et al. 2020).

The onset of the YD (~12.9–12.7 cal. ka BP; YDa) is marked by a progressive cooling and drying in the Eastern North Atlantic region (including the central and South-Western Europe) and over Greenland (Fig. 3) (e.g.; Alley et al. 1993; Brauer et al. 1999; Rasmussen et al. 2006; Bakke et al. 2009; Thornalley et al. 2011; Rach et al. 2014; Bartolomé et al. 2015; Muschitiello et al. 2015; Naughton et al. 2019). This cooling was triggered by the gradual slowdown of AMOC (Fig. 3) (McManus

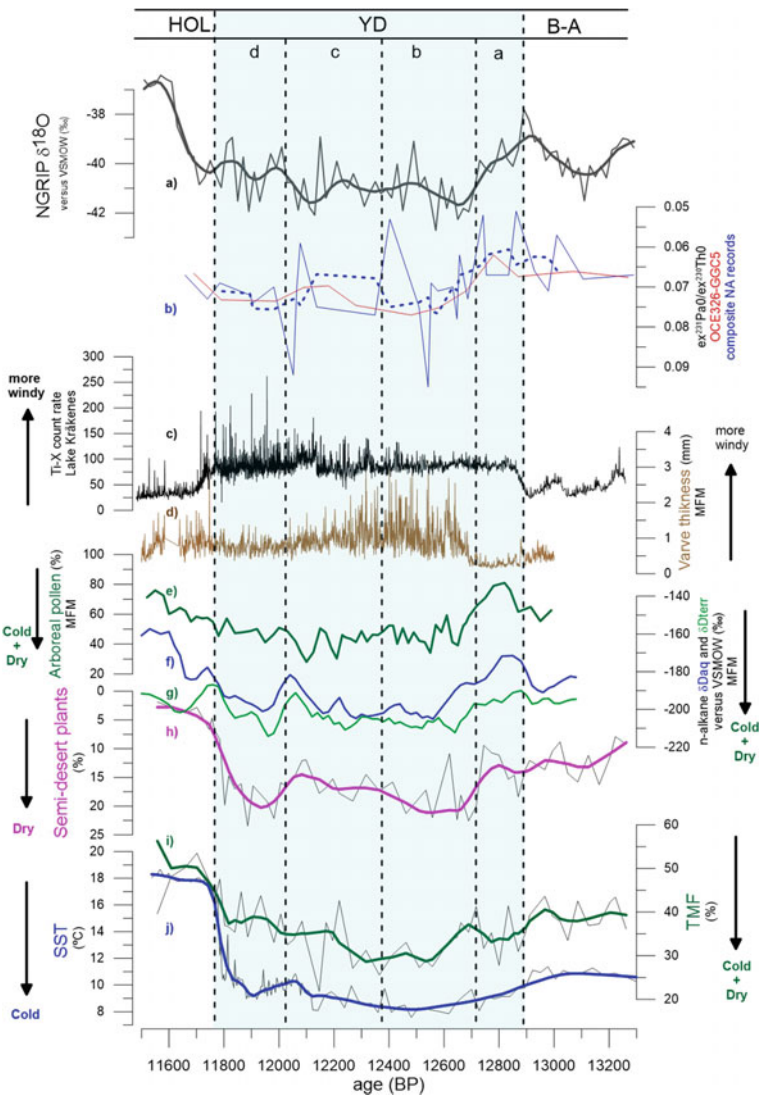

Fig. 3 a) North Grip (NGRIP) δ^{18}O versus VSMOW (in cal yr BP) (Rasmussen et al. 2006); b) ex231Pa0/ex230Th0 from core OCE326-GGC5 (red line) (McManus et al. 2004) and from a composite of North Atlantic records (thin grey line) (blue dashed line: smoothed record) (Ng et al. 2018); c) Ti count rate in Lake Kråkenes (Bakke et al. 2009); d) varve thickness and e) Arboreal pollen percentages at Lake Meerfelder Maar (MFM) (Brauer et al. 2008); f) δD values of the nC23 (aquatic plants, daq, blue line); g) δD values of the nC29 alkanes (higher terrestrial plants, dDterr, green line) at MFM (Rach et al. 2014); h) semi-desert plants percentages and i) Temperate Mediterranean forest (TMF) from western Iberian Peninsula (Naughton et al. 2019); j) SST from western Iberian margin (Naughton et al. 2019)

et al. 2004; Ng et al. 2018) and by the slight expansion of the sea ice in the North Atlantic (Cabedo-Sanz et al. 2013; Müller and Stein, 2014). The meridional thermal contrast in the North Atlantic was still reduced (Naughton et al. 2019) and the jet stream and the westerlies were relatively weakened in Western Europe (Fig. 3) (e.g. Rach et al. 2014; Bartolomé et al. 2015; Naughton et al. 2019).

The coldest phase of the YD (YDb) was detected in the Eastern North Atlantic, Western Europe and over Greenland during 12.7–12.4 cal. ka BP (Fig. 3) (e.g.; Brauer et al. 1999; Von Grafenstein et al. 1999; Rasmussen et al. 2006; Bakke et al. 2009; Dormoy et al. 2009; Fletcher et al. 2010b; Chabaud et al. 2014; Rach et al. 2014; Baldini et al. 2015; Bartolomé et al. 2015; Naughton et al. 2019). In contrast, the Western North Atlantic mid-latitudes were relatively warm (Carlson et al. 2008) making the meridional thermal contrast between the high- and mid- latitudes of the North Atlantic very high (Naughton et al. 2019). This thermal contrast was likely a response to the maximum slowdown of the AMOC (Fig. 3) (McManus et al. 2004; Ng et al. 2018) and consequent decrease in the northward heat transport, causing a relative warming in the Western North Atlantic mid-latitudes (Carlson et al. 2008) and very cold conditions in the North Atlantic high-latitudes and Eastern North Atlantic mid-latitudes (e.g.; Bakke et al. 2009; Naughton et al. 2019). This also favoured the decrease of the NADW formation and the southward extension of the sea ice (e.g.; Bakke et al. 2009; Thornalley et al. 2011; Cabedo-Sanz et al. 2013; Müller and Stein, 2014). As a consequence to the steep meridional thermal contrast, the polar jet stream strengthen and reached its southernmost position making the westerlies became stronger at the sea ice edge in Central Western Europe (Brauer et al. 2008; Bakke et al. 2009; Rach et al. 2014), which was demonstrated by climate simulations (Isarin et al. 1998). Although westerlies were not so intense at the mid-latitudes, the horizontal thermal contrast between the Western and Eastern North Atlantic would have favour some moisture delivery to Western Iberian Peninsula even if conditions in this region were mainly drier as a response to maxima of AMOC reduction (Naughton et al. 2019). This phase was followed by a progressive warming and increase in moisture conditions, between ~12.4 and 12.0 cal. ka BP (YDc), in South-Western and Central Europe and in the North Atlantic (Fig. 3) (e.g.; Bakke et al. 2009; Rach et al. 2014; Repschlager et al. 2015; Naughton et al. 2019). This warming was favoured by the slight AMOC re-invigorating (McManus et al. 2004; Ng et al. 2018), causing the sea ice and polar jet stream to retreat further north (e.g.; Thornalley et al. 2011; Cabedo-Sanz et al. 2013; Pearce et al. 2013; Müller and Stein, 2014; Gil et al. 2015), as revealed by the latitudinal switch of the westerlies from Central to Northern Europe (Fig. 3) (Bakke et al. 2009). Although the polar jet stream was located in Northern Europe, some moisture was delivered to Central and Southern Europe via the southern branch of the North Atlantic drift (Naughton et al. 2019).

Finally, the YD termination, or the YD-Holocene transition (YDd; 12.0–11.7 cal. ka BP) is marked by very unstable temperature and precipitation conditions in the North Atlantic and over Europe (Fig. 3) (Bakke et al. 2009; Rach et al. 2014; Naughton et al. 2019; Rea et al. 2020). These complex conditions seem likely to be the result of successive shifts in the mode of operation of the AMOC as part of its recovery at the onset of the Holocene (Naughton et al. 2019) and a more unsteady North Atlantic sea-ice cover in the North Atlantic and Nordic seas (Cabedo-Sanz et al. 2013; Pearce et al. 2013; Gil et al. 2015). This contributed to substantial changes in the strength and position of westerlies and a complex hydroclimatic scenario in the Central and Southern Europe (Bakke et al. 2009; Rach et al. 2014; Naughton et al. 2019; Rea et al. 2020).

2.4 The Holocene

The present-day interglacial, the Holocene (Gibbard et al. 2005), started at 11.7 cal. ka BP (Rasmussen et al. 2006; Steffensen et al. 2008; Walker et al. 2009) and was marked by an abrupt warming at the Younger Dryas (YD)-Holocene transition in a variety of records from Europe (e.g.; Mangerud et al. 1974; Bjorck et al. 1996; Litt et al. 2001; Moréllon et al. 2018 and references therein), North Atlantic mid-latitudes (e.g.; Naughton et al. 2007a; b; Fletcher et al. 2010b; Chabaud et al. 2014; Gomes et al. 2020) and in Greenland ice cores (Fig. 4) (Alley et al. 2003; Steffensen et al. 2008; Walker et al. 2009; Vinther et al. 2009). The Holocene interglacial started just after the peak maxima of Northern Hemisphere summer insolation (Fig. 4) (Berger and Loutre, 1991). Interestingly, the warmest conditions (also known as the Holocene Thermal Maximum: HTM) were attained, several millennia later than the peak maxima of Northern Hemisphere summer insolation, caused by the persistence of larger remaining ice sheets in the Northern Hemisphere (mainly LIS, but also Greenland Ice Sheet—GIS) until ~7 cal. ka BP (e.g.; Carlson et al. 2008; Renssen et al. 2009; Blaschek and Renssen, 2013; Bova et al. 2021). Data-model comparison shows that strong summer insolation during the Early Holocene favoured the disintegration of Northern Hemisphere ice sheets and consequent melting that cause a relative weak AMOC (Blaschek and Renssen 2013). This weak AMOC had a cooling effect in the Early Holocene, at least at a regional scale, compensating the relatively strong orbitally-forced boreal insolation. However, the timing of the HTM varied spatially as revealed by numerous terrestrial and marine proxy records and climate simulations (e.g.; Kaufman et al. 2004; 2020; Renssen et al. 2009; 2012). The HTM started earlier than 8 cal. ka BP in regions not strongly affected by the remnant of LIS, while after 8 cal. ka BP in other regions (Renssen et al. 2012). Furthermore, the warming was stronger in the polar regions than in the tropics, where the amplitude of warming was reduced during the HTM (Renssen et al. 2012). Some Southern Europe (e.g., Massif Central pollen-based temperature reconstructions, France) and North Atlantic mid-latitudes (Western Iberian margin SST) records show a progressive warming phase in the early part of the Holocene, between 11.7 ka and 10.5 cal. ka BP and that the HTM was detected earlier than 8 cal. ka BP, around 10.5 cal. ka BP (Rodrigues et al. 2010; Martin et al. 2020). SST in southern Iberian margin attain maximum values at around 10.5 ka BP, while pollen assemblages from the same records show maxima of forest development 1000 years later, i.e. by 9.5 cal. ka BP (e.g., Chabaud et al. 2014; Gomes et al. 2020). As forest, several other terrestrial climate indicators not

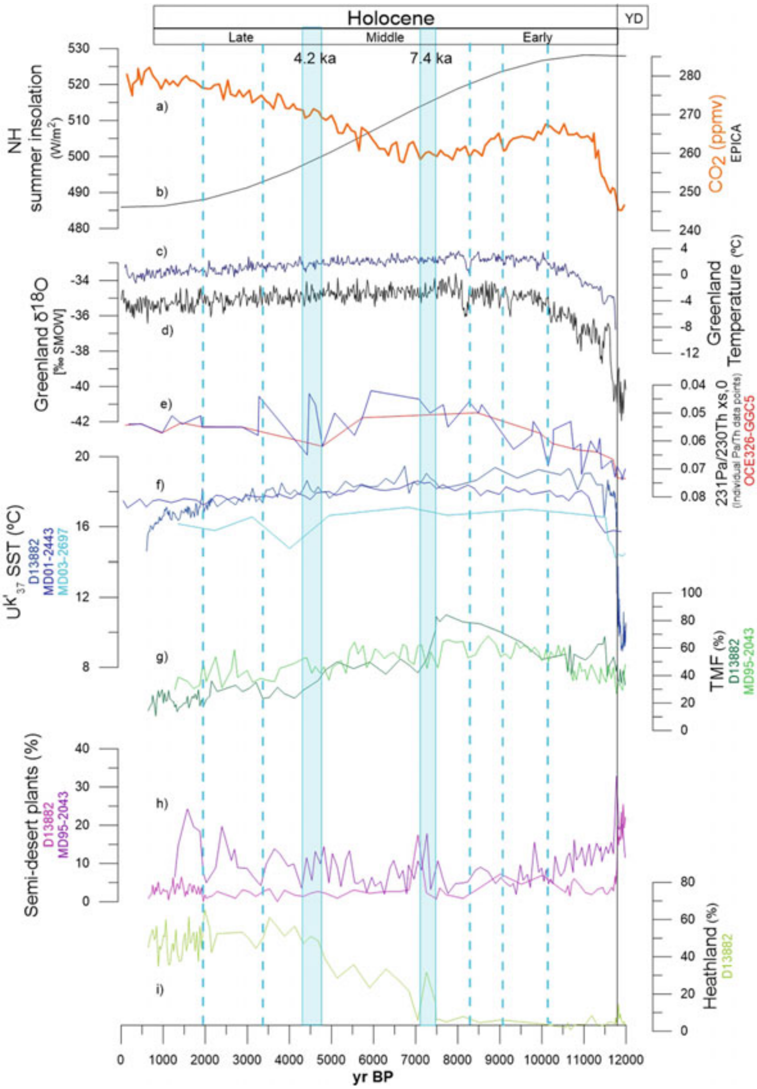

Fig. 4 a) atmospheric CO_2 concentrations reconstructed from EPICA DOME C ice core (EDC) (EPICA community members, 2004); b) summer insolation at 65ºN (Berger and Loutre, 1991); c) Greenland temperature (Vinther et al. 2009); d) Greenland ice core $\delta^{18}O$ record (Rasmussen et al. 2014); f) ex231Pa0/ex230Th0 from composite North Atlantic records (blue line) (black bold line: smoothed record) (Ng et al. 2018) and from the western North Atlantic (light blue) (OCE326-GGC5; McManus et al. 2004); e) alkenone-derived SST records ($U^{k'}_{37}$ SST) from Iberian margin (Martrat et al. 2007; Rodrigues et al. 2009; Naughton et al. 2016; Gomes et al. 2020); g) TMF: Temperate forest abundances (%); h) Semi-desert plants in i) Heathland Southwestern Iberian margin records (Rodrigues et al. 2009; Fletcher et al. 2013; Gomes et al. 2020); Light blue bands: millennial scale events that separates the Holocene Sub-series/Sub-epochs; Dashed light blue lines: other millennial scale events mentioned in the text