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## **The Global Last Glacial Maximum: the Eastern North Atlantic (marine sediments) and the Greenland Ice Sheet climatic signal**

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### **Abstract :**

This chapter summarises the environmental changes that accompanied the rapid growth of the Northern Hemisphere ice sheets c. 35–30 cal ka BP and their maximum extent c. 29–19 cal ka BP, with a special focus on the European continent and its surrounding margins.

**Keywords :** Last Glacial Maximum, ice sheets, palaeoenvironment, Europe

The Last Glacial Maximum (LGM) is the end stage of the last ice age that began about 116,000 years ago with the beginning of ice sheets building up. This first increase in the size of the ice caps corresponds to a decrease in the insolation received at 65°N and a synchronous decrease in temperatures in the North (Greenland) and South (Antarctica) Poles. Atmospheric CO<sub>2</sub> concentration decreases only several thousand years after this first decreased in insolation. The ice sheets continued to build up slowly over the next tens of thousands of years and the level of carbon dioxide also decreased progressively (notably by a greater dissolution of CO<sub>2</sub> in the ocean). This decrease of CO<sub>2</sub> reinforced the cooling initiated by astronomical parameters and amplified by the albedo increase as the result of the southern migration of the tundra and the growth of the ice sheets, and the associated atmospheric changes increasing the arrival of moisture to the northern hemisphere high latitudes (Crucifix and Loutre, 2002; Kageyama et al., 2004; Sanchez Goñi et al., 2005).

The LGM *sensu lato* occurred between 29,000 and 19,000 years before today (29-19 cal ka BP), although its definition is a matter of debate (e.g. Clark et al., 2009; Hughes & Gibbard, 2015; Mix et al., 2001). It is characterized by a maximum ice extent and hence a minimum sea level. LGM sea level is estimated about 120-130 m lower than today (Yokohama et al., 2000; Lambeck et al., 2014), thus implying that a significant portion of the continental shelves currently below sea level (including the shelf between France and the British Isles as well as the North Sea in Europe) were above sea level at the LGM. The level of CO<sub>2</sub> was at its lowest (185 ppm). The extension of sea ice in the northern hemisphere completely isolated Greenland in summer and winter. In the south, the sea ice around Antarctica extended further north than 50°S in winter, i.e., a northward extension of more than 10° than at present. Overall, the atmospheric temperature was about 6°C colder than today (−6.5 to −5.7 °C; Tierney et al., 2020). However, there are large regional disparities, especially the polar regions, where the

large ice sheets were located, cooled much more than the lower latitudes. These disparities are documented by numerous climate archives and reproduced by climate models. The extension of the ice pack towards the lower latitudes, particularly in the North Atlantic and around Antarctica, also lead to a decrease in air temperature of nearly 10°C in East Antarctica (Werner et al., 2018), more than 11°C in West Antarctica (Cuffey et al., 2016) and more than 20°C in the centre of the Greenland ice sheet (Dahl-Jensen et al., 2001; Buizert et al., 2018). The LGM is not only affected by colder temperature in the high latitudes but also by significant changes in the intensity of precipitation and in the location of monsoon zones (displaced further south than present day; Yang et al., 2015; An et al., 2015). Vegetation cover was significantly different from that at present (Prentice et al., 2000; Harrison and Bartlein, 2012).

In southwestern Europe, deep-sea and terrestrial pollen archives show that instead of the temperate forests of our interglacial period, *Pinus* and heathlands-steppes dominate this region (Fletcher and Sanchez Goñi, 2008; Naughton et al., 2007; Turon et al., 2003). Climatic conditions were colder and drier during the LGM in this region than they are today, explaining this glacial vegetation cover, but wetter than during the Heinrich Stadials (HS) bracketing it. The *Pinus* species characteristic of mountain forests, *P. nigra* and *P. sylvestris*, dominated southwestern Europe, a region characterised by Mediterranean vegetation today (Desprat et al., 2015).

Continental ice caps at the LGM essentially covered the northern part of the continents in the Northern Hemisphere (Elhers & Gibbard, 2004). Reconstructions of the ice sheet extent, topography and altitudes based on data and ice-sheet modeling (Peltier, 1994, 2004; Lambeck et al., 2000; Hughes et al., 2016) show that, in addition to the Greenland ice cap, a huge ice sheet (i.e. Laurentide Ice Sheet; LIS), with a thickness reaching 4 km in places, covered a large part of North America (Dyke et al., 2002). Northern Eurasia was also completely covered by the European Ice Sheet Complex (EISC; Hughes et al., 2016; Clark et al., 2012) that attained

its maximum extent (5.5 Mkm<sup>2</sup>) and volume (up to 3 km thick; ~24 m Sea Level Equivalent) at ~21 cal ka BP (Hughes et al., 2016). The latter was formed by the Fennoscandian and British-Irish Ice Sheets, and a recent compilation shows that its westernmost limit along the northwestern European margin was reached up to 6000-7000 years earlier (at ~27–26 cal ka BP) than the eastern limit on the Russian Plain (at ~20–19 cal ka BP) (Hughes et al., 2016; Patton et al., 2016).

Sediments from the northwestern European margin record the growth, as well as the fluctuations, of the western EISC. Depositional processes strongly change on the margin, especially when ice extended across the continental shelf, as shown along the British–Irish (Dunlop et al., 2010; O’Cofaigh et al., 2012) and Norwegian (Mangerud, 2004) shelves. Hence, glacial and glaciogenic sedimentation dominates north of ~50-55°N at the LGM (Weaver et al., 2000). The ice built trough-mouth fans (Laberg & Vorren, 1995; King et al., 1998) and left iceberg scour marks on the upper slope and shelf (Sacchetti et al., 2012; O’Cofaigh et al., 2012). European-sourced dropstones of ice-rafted origin are also widely recognized down to the abyssal plain, and their highly variable flux through the LGM emphasize the instability of the marine-terminating ice-streams (*e.g.* Irish Sea Ice Stream; Scourse et al., 2009; Peck et al., 2007). A major change in the sedimentation on the northwestern European margin also occurred when the Fennoscandian and British-Irish ice sheets merged into the North Sea (Sejrup et al., 2009; Clark et al., 2012), leading to the formation of a huge river system (2.5 x 10<sup>6</sup> km<sup>2</sup>; Patton et al., 2017), the so-called Fleuve Manche, or Channel River, that drained the western European continent and the southern limb of the EISC (Gibbard, 1988; Toucanne et al., 2010). Thus, the Channel River routed substantial amounts of meltwater to the North Atlantic from ~35-30 cal ka BP to ~17 cal ka BP (Zaragosi et al., 2001; Ménot et al., 2006; Toucanne et al., 2009) and inputs of glaciogenic sediment off the Channel River mouth increase substantially, as shown by the huge increase in turbidite flows in the northern Bay of Biscay 45-48°N (Zaragosi et al.,

2006). On this interval, Channel River sediments reveal that the evolution of western EISC margins was accompanied with substantial ice recession in the Baltic region (from Denmark to Poland; Toucanne et al., 2015). These events occurred both in the growth of the ice sheet to its LGM position (i.e., during HS 3 and HS 2) and during the last deglaciation (i.e. at ~22 cal ka BP, ~20-19 cal ka BP and from 18.2 to 16.7 cal ka BP that corresponds to the first part of HS 1). These meltwater discharges in the North Atlantic region could explain the concomitant weakening of Atlantic Meridional Overturning Circulation (AMOC) (e.g. Ivanovic et al., 2018) responsible for the marine-based ice stream purge cycle of the LIS, or so-called Heinrich events (HE) (Alvarez-Solas et al., 2010; Boswell et al., 2019). LIS-sourced IRD are recognized all along the European margin during the last glacial period (Grousset et al., 2000; Scourse et al., 2000; Peck et al., 2007; Hodell et al., 2017; de Abreu et al., 2003). Their arrival in the northeast Atlantic is coeval with a southward shift of the polar front, as far south as the Iberian Peninsula, as shown by planktic foraminifera (Eynaud et al., 2009) and alkenone-based sea surface temperature reconstructions (Bard et al., 2000; Rodrigues et al., 2017). Such a major cooling of the North Atlantic climate occurred for the last time at the end of the LGM, about 18-16 cal ka BP (i.e. HS 1 that is the marine equivalent of the Oldest Dryas; Naughton et al., 2007).

Although the Greenland water isotope records ( $\delta^{18}\text{O}$ ) are often considered as the best key archives for deciphering the climate evolution in the North Atlantic region over the last climatic cycle, it is worth noting that Greenland  $\delta^{18}\text{O}$  does not capture the full climatic variability (Buizert et al., 2018). As a prominent example, the Greenland  $\delta^{18}\text{O}$  is not able to capture the signature of the HE and HS: the HS do not exhibit lower  $\delta^{18}\text{O}$  than non-HS (Guillevic et al., 2014). We can thus wonder if the same is true for the LGM and the early EISC deglaciation and if the absence of any strong  $\delta^{18}\text{O}$  signal over the period encompassing 27 to 18 cal ka BP is the result of a lack of sensibility of the  $\delta^{18}\text{O}$  signal. The lack of sensitivity of the  $\delta^{18}\text{O}$  record to climatic variability during this period could be the result of the extremely large sea ice

extension around Greenland that would isolate Greenland from receiving moisture/precipitation from the North Atlantic. In particular, the  $\text{Ca}^{2+}$  concentration is an indicator of source strength and transport conditions from terrestrial sources (high latitude Asian desert for Greenland; Biscaye et al., 1997). The latter proxy suggests, for example, two massive reorganizations of the Northern Hemisphere atmospheric circulation during HS2 (see the d1 and d2 ion peaks in Rohling et al., 2003). They coincide (within age uncertainties) to the two-step retreat (by about 400 km) of the Irish Sea Ice Stream (Smedley et al., 2017; Chiverell et al., 2020), shown by the discrete pulses in Irish-sourced IRD recorded on the Porcupine Seabight (Scourse et al., 2000, 2009).

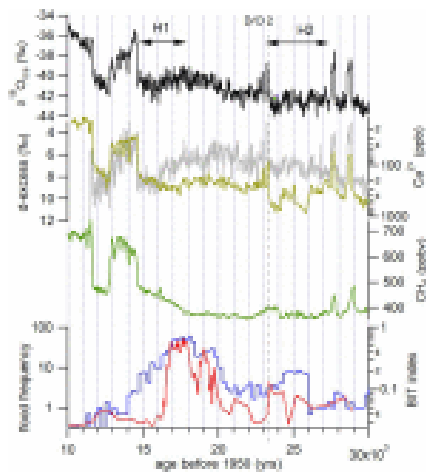
As for ice marginal fluctuations, the vegetation record, together with aeolian periglacial deposits (i.e. loess; Antoine et al., 2009; Moine et al., 2017), supports the idea of abrupt climate changes during the LGM. The vegetation record clearly identifies succession of a steppe-open forest landscapes from HS 2 to HS 1. Deep-sea pollen records from the western Iberian margin show that HS 2 and HS 1 were composed of two major phases. The beginning of each HS was marked on land by an important pine forest reduction and the expansion of heathers synchronous with sea-surface temperature cooling suggesting that these first phases were cold and wet. *Pinus* forest expansion characterising the second phase of each HS indicates a less cold episode associated, during HS1, with an increase of dryness as suggested by the development of semi-desert vegetation (Naughton et al., 2009, 2016). These vegetation changes suggest substantial shifts in the atmospheric circulation patterns. During the first phases of HS 2 and HS 1 the thermal front and polar jet stream was displaced as far south as 37°N, facilitating the transfer of moisture to western Iberia via the westerlies. In contrast, the thermal front and polar jet stream was displaced further north of 41°N during the second phase of HS 2 and HS 1 preventing the entering of moisture at these latitudes. The northward displacement of the

thermal front and jet stream during this second phase is also supported by climate simulations (Ziemen et al., 2019).

Knowledge of the state of the Earth at the LGM is an important benchmark for understanding the sensitivity of global environmental systems to change (Mix et al., 2001). For this reason, the LGM is a period widely targeted for simulations since the first earth system modelling (*e.g.* Aleya, 1972; Williams et al., 1974; Kutzbach and Guetter, 1986). The first and second phase of Palaeoclimate Modelling Intercomparison Project (PMIP: Braconnot et al., 2007a, b) focused on the LGM because a strong effort in paleoclimatic syntheses has been done (*e.g.* CLIMAP, 1981; EPILOG, Mix et al., 2001; MARGO Project Members, 2009; Bartlein et al., 2011; Schmittner et al., 2011; Braconnot et al., 2012) hence facilitating model evaluations. The LGM was hence chosen when paleoclimate experiments were included in the fifth phase of the Coupled Modelling Intercomparison Project (CMIP5: Taylor et al., 2012; Braconnot et al., 2012). As shown above, the LGM period in Europe is complex and continuing to explore this complexity is a unique opportunity to test model performance, and hence reduce the uncertainty about the magnitude of the ongoing climate changes (Braconnot et al., 2012).

**Figure 26.1-** Sequence of LGM and last deglaciation from ice core and marine records. First panel:  $\delta^{18}\text{O}$  of ice in the NorthGRIP ice core (NGRIP members 2004); second panel:  $\text{Ca}^{2+}$  and d-excess from the NorthGRIP ice core (Ruth et al., 2007; Landais et al., 2018); 3<sup>rd</sup> panel:  $\text{CH}_4$  from the WAIS-D ice core (Rhodes et al., 2015); 4<sup>th</sup> panel: Channel River discharge reconstructed from flood events per 250 years (blue line; Toucanne et al., 2015) and branched

and isoprenoid tetraether (BIT) index, a proxy for the relative fluvial input of terrestrial organic matter in the marine environment (red line; Ménot et al., 2006).



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