Chapter 8

Early reactivation of European rivers during the last deglaciation

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Abstract

During the Last Glacial Maximum, the sea-level low stand combined with the large extent of the Fennoscandian and British ice sheets led to the funneling of European continental runoff, resulting in the largest river system that ever drained the European continent. Here we show an abrupt and early reactivation of the European hydrological cycle at the onset of the last deglaciation, leading to intense discharge of the Channel River into the Bay of Biscay. This freshwater influx, probably combined with inputs from proglacial or ice-dammed lakes, dramatically affected the hydrology of the region, both on land and in the ocean.

Despite the recognized sensitivity of oceanic circulation to changes in the freshwater budget at high latitudes (Broecker et al., 1989; Stocker and Wright, 1991; Manabe and Stouffer, 1997; Ganopolski and Rahmstorf, 2001), river runoff studies have so far mainly been focused on low-latitude palaeorecords (Adegbie et al., 2003; Jennerjahn et al., 2004; Schefuß et al., 2005). Furthermore, except for a few continental archives such as speleothems and wetlands that reflect local conditions, little is known about hydrological and water drainage changes in Europe during the last deglaciation. During the Last Glacial Maximum (LGM), a large ice sheet (known as the Fennoscandian ice sheet) was established on the Eurasian continent. Both the sea-level low stand and the extent of the ice sheet deeply influenced the drainage basins of European rivers that flowed into the Channel River, thus generating one of the largest rivers ever to have extended across the European continent (Fig. 8.1). Because this river transported much of the meltwaters coming from the European glaciers as well as from the Fennoscandian and British-Irish ice sheets (Scourse et al., 2000; Antoine et al., 2003), its runoff is expected to have reacted strongly to the retreat and growth of the Eurasian ice sheets and of the alpine glaciers. Therefore, a record of the activity of this palaeoriver could provide a detailed account of the effect of European deglaciation on the hydrological cycle.

Core MD952002 (47°27’N, 8°32’W, 2174 m water depth) was recovered on the northwestern slope of the Bay of Biscay in the direct axis of the English Channel during the IMAGES 101 cruise of the research vessel Marion Dufresne (Fig. 8.1). The chronology of this core is based on calibrated ^{14}C ages (see supporting material for methods at the end of this chapter). This core covers a critical period including the last deglaciation, as well as abrupt climatic changes such as Heinrich events 1 and 2 (H1 and H2), which are clearly identified by two discrete peaks in the abundance of lithic grains and the magnetic susceptibility at 16 and 24 thousand years before the present (kyr B.P.) (Fig. 8.2B) (Zaragosi et al., 2001). Total organic carbon (TOC) content varies between 0.2 and 1.2%, with minima during both H1 and H2 events (green circles in Fig. 8.2C). The $C_{37:4}$ alkenone, a biomarker derived from haptophyte algae and thought to be a proxy for low-salinity water associated with icebergs (Bard et al., 2000), is absent during the Holocene but exhibits high values between 11 and 18 kyr B.P., with a prominent maximum reached during H1, corresponding to 30% $C_{37:4}$ among the total of C_{37} alkenones (black diamonds in Fig. 8.2C). This is a characteristic feature for H1 in this area and has been related to the advection of low-salinity water associated with icebergs (Bard et al., 2000).

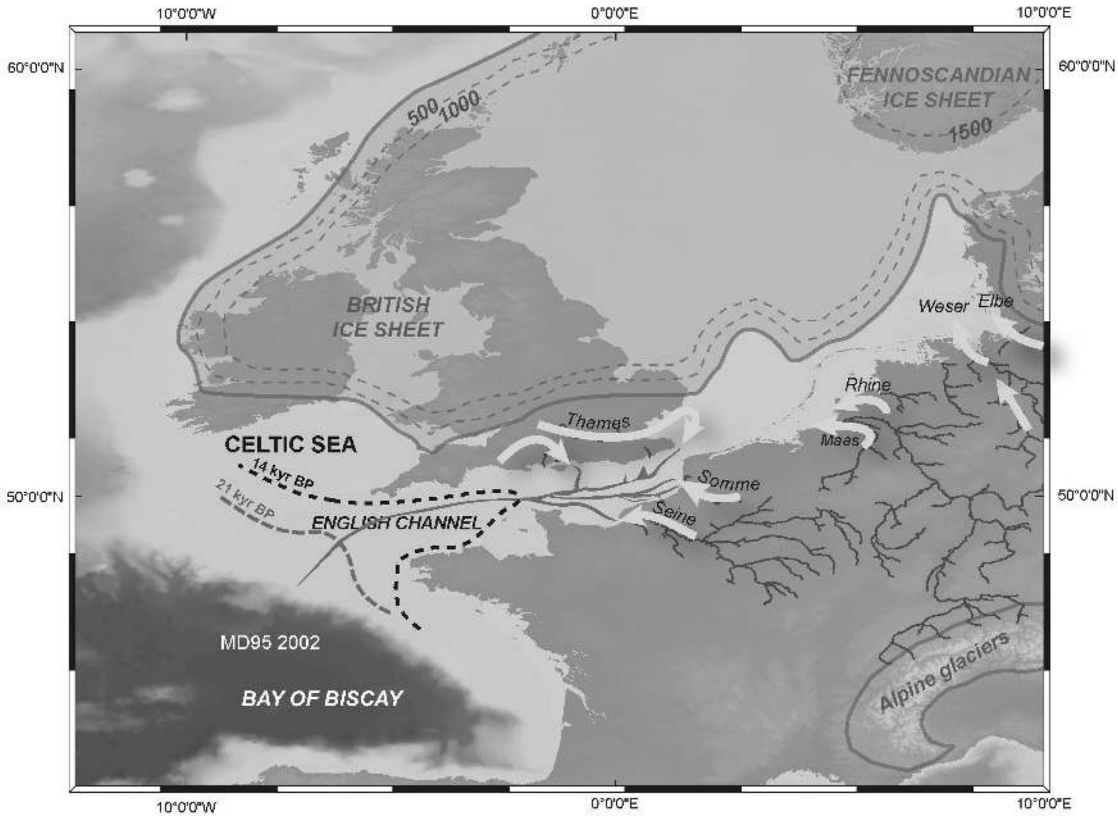
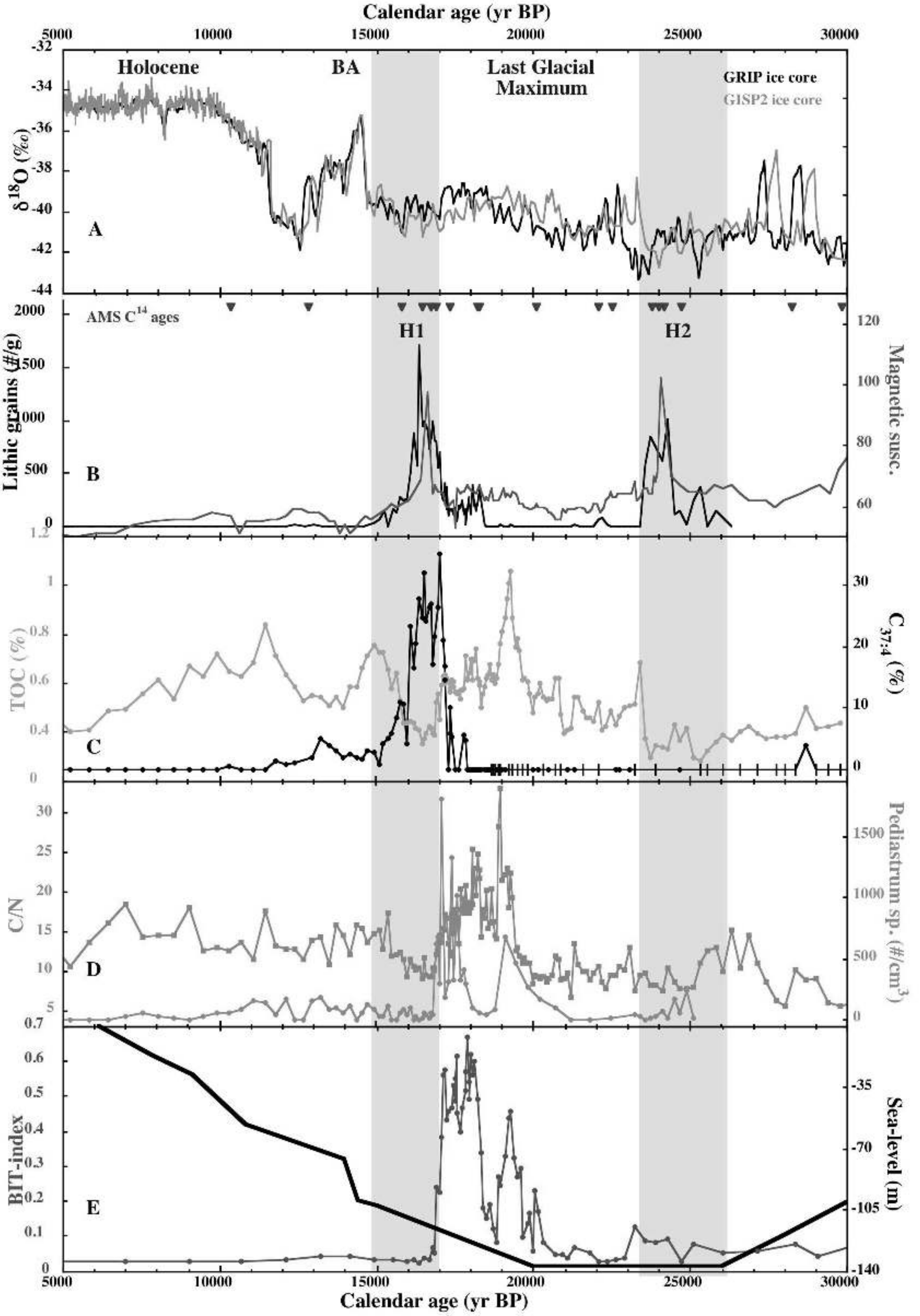


Figure 8.1: The palaeoenvironment of the LGM on the Eurasian continent was radically different from today. The Fennoscandian ice sheet was established on the northern part of Europe, extending west into the Norwegian Sea, south across the north German Plain into Poland, and eastward into North Poland and Russia (Bowen et al., 2002; Mangerud et al., 2004). A smaller dome was installed on the British Isles (Bowen et al., 2002). Recent geomorphological evidence indicates that the British-Irish ice sheet (BIS) and Fennoscandian ice sheet coalesced, and a huge ice dam extended over the present-day North Sea (Mangerud et al., 2004; Svendsen et al., 2004). The Alps were almost entirely covered by an ice dome formed by valley glaciers (Denton and Hughes, 1981). The maximum extent of ice sheets at the LGM is illustrated by the blue contours. A final ice-age sea-level lowstand led to emersion of the channel between England and France, with the coastlines at 14 and 21 kyr B.P. illustrated by the dashed lines (after Lambeck, 1997). A palaeoriver, known as the Channel River (in orange), extended across the emerged continental margin (Bourillet et al., 2003). It drained most of the major rivers in northwestern Europe, that is, the Rhine, Maas, Seine, Solent and Thames (yellow arrows on the map). In addition to these rivers, the Irish Sea drained a large part of the BIS meltwaters (McCabe and Clark, 1998). Furthermore, damming by the Fennoscandian ice sheet favored the development of southward-flowing meltwater valleys and ice-margin spillways running westward. These spillways collected proglacial waters from rivers even farther east than the Elbe basin and allowed drainage to the Channel River (Marks, 2002; Mangerud et al., 2004). Core MD952002 (red dot) was taken at a water depth of 2174 m in the axis of the English Channel, close to the LGM position of the Channel River outlet. See page 167 for color figure.


Calendar age (yr BP)
Holocene
BA
Last Glacial Maximum
GRIP ice core
GISP2 ice core
δ18O (‰)
A
AMS C14 ages
H1
H2
Lithic grains (#/g)
Magnetic susc.
B
TOC (%)
C37:4 (%)
C
C/N
Pediastrum sp. (#/cm3)
D
BIT-index
Sea-level (m)
E

We applied the branched and isoprenoid tetraether (BIT) index to reconstruct terrestrial organic matter fluvially transported to the ocean (Hopmans et al., 2004). This proxy uses the relative abundance of membrane lipids (i.e., non-isoprenoid glycerol dialkyl glycerol tetraethers) (Sinninghe Damsté et al., 2000) derived from anaerobic bacteria thriving in soils and peats (Weijers et al., 2006a), compared with crenarchaeol, a structurally related isoprenoid molecule characteristic of ubiquitous marine planktonic and lacustrine crenarchaeaota (Sinninghe Damsté et al., 2002). BIT-values for suspended particulate matter in river waters are typically >0.9 (Herfort et al., 2006). A survey of Holocene sediments showed that the BIT index can be directly correlated to the relative amount of fluvial terrestrial organic matter input. BIT index values of <0.1 are typical for open marine settings receiving only small amounts of terrestrial organic matter, while values >0.4 are typical for river fans and fjord systems (Hopmans et al., 2004). The BIT index values throughout the core MD952002 remain below 0.1, except for two well-defined peaks with values as high as 0.7 centered at 19.5 kyr B.P. and between 19 and 17 kyr B.P. (red circles in Fig. 8.2E). The two maxima in the BIT-index profile, therefore, reveal periods during which large amounts of terrestrial organic matter must have been transported to this site in the Bay of Biscay. These maxima are consistent with those obtained from the abundance of remains of freshwater algae (*Pediastrum sp.*) (blue curve on Fig. 8.2D) (Zaragosi et al., 2001). The total organic carbon-to-nitrogen ratio (C/N) varies between 5 and 30, with minimum values typical of the marine environment end-member during H1 and H2 events, and higher values in the intervening period, also indicative of a larger terrestrial contribution (Fig. 8.2D).

Figure 8.2 (opposite page)**:** Deglaciation-Holocene records of the past activity of the Channel River as a function of palaeoclimatic changes. The chronology is based on calibrated ^{14}C ages measured on planktonic foraminifera [shown as triangles in (B)] (see supporting material at the end of this chapter). Climatic events are abbreviated as follows: B-A, Bølling-Allerød; H1, Heinrich 1; and H2, Heinrich 2. The grey areas underline the H1 and H2 events. (**A**) $\delta^{18}O$ GRIP (black line) (Johnsen et al., 2001) and $\delta^{18}O$ GISP2 (light blue line) (Stuiver and Grootes, 2000) records reflecting Greenland air temperatures. (**B**) Black line shows the counting of grains identified as ice-rafted debris (IRD) per 10 g for the size fraction coarser than 150 μm, and the grey curve shows the magnetic susceptibility (MS) record measured on board Marion Dufresne (Auffret et al., 2002). (**C**) Green circles represent the total organic carbon contents, and the black diamonds the percentage of $C_{37:4}$ among C_{37} alkenones, i.e., $\%C_{37:4} = 100 \times [C_{37:4}]/[C_{37:2} + C_{37:3} + C_{37:4}]$. Beyond 19 kyr B.P., the relative percentage of $C_{37:4}$ could not be quantified because alkenones are very scarce in the sediments corresponding to the last glacial period (black ticks in Fig. 8.2C). (**D**) Orange squares indicate the total organic carbon-to-nitrogen ratios (C/N), and the blue symbols show the abundance of freshwater algae, *Pediastrum sp.* (counts from Zaragosi et al., 2001). (**E**) The BIT index is defined as follows: BIT = (I+II+III)/[(I+II+III)+(IV)] (the roman numbers refer to the glycerol dialkyl glycerol tetraethers in Figure 5.1, Hopmans et al., 2004), and is represented here by red dots. The black curve shows the sea-level curve (Lambeck et al., 2002). See page 168 for color figure.

During H2 and the LGM, cold and dry conditions prevailed on the European continent (Allen et al., 1999). At that time, ice sheets reached their maximum extent (Fig. 8.1) and sedimentation at the Channel River outlet was typical of a marine environment with low values of the BIT index and TOC as well as low C/N ratios (Fig. 8.2, C to E). At the end of the LGM, between 21 and 17 kyr B.P., an early warming is observed in the Greenland air-temperature record (Fig. 8.2A). This temperature increase is also clearly detected in several North Atlantic records (e.g., Jones and Keigwin, 1988; Bard et al., 2000; Pailler and Bard, 2002; Alley et al., 2002) as well as in continental reconstructions inferred from pollen in lacustrine and peat sequences over Europe (e.g., Sanchez Goñi et al., 2000; Combourieu Nebout et al., 2002). This warming was accompanied by enhanced precipitation, as is also evident from the pollen assemblages. Despite this climatic warming, soils remained partly frozen and hence impermeable (Renssen and Vandenberghe, 2003). Furthermore, the vegetation cover was scarce and spatially discontinuous, mainly composed of peat with only few woody species (Tzedakis et al., 2002). This situation led to the development of large fluvial systems, intense soil erosion, and enhanced river discharge. This transient period is coincident with an abrupt maximum in the BIT index in core MD952002 (Fig. 8.2E), indicative of an early and drastic reactivation of European rivers.

Associated with this reactivation of the hydrological cycle, meltwaters might well have played a role in the runoff increase. In fact, when the Fennoscandian ice sheet started to retreat from its maximum position at ~22 kyr B.P., a new series of short-lived glacial lakes formed at its southern margin, more particularly in the Polish basins and German lowlands (Marks, 2002; Bowen et al., 2002). Because of the position of the ice margin, the meltwaters first drained through the southern Peribaltic area toward central Poland, to the Elbe River, and then to the Channel River (Mangerud et al., 2004). The retreat was not continuous, and readvances of the Fennoscandian ice sheet have been recognized based on geomorphological and lithostratigraphic evidences as well as by cosmonuclide and thermoluminescence dating (Marks, 2002; Bowen et al., 2002). Two major deglaciation phases have been reported in Poland during the low sea-level stand: the Poznan and the Pomeranian phases at 22.0 and 18.6 kyr B.P. (Marks, 2002). Similar pulsations have also been described for the British-Irish ice sheet over the same period (Bowen et al., 2002). On the southeastern sector, the Scandinavian ice sheet begins to retreat around 19 kyr B.P. after a phase of maximum extent at 20.9 kyr B.P. (Rinterknecht et al., 2006). A massive and early breakdown of the LGM system of ice domes in the Alps is reported to occur simultaneously (e.g., Ivy-Ochs et al., 2006).

A peculiar geographic setting reinforced the effect of this increased water runoff from the European continent, leading to increased discharge of the River Channel into the Bay of Biscay. In fact, the low sea level during the LGM (Fig. 8.2E) means that the river mouth was located very close to the core location [the dashed line in Fig. 8.1 represents the palaeocoastline at 21 kyr B.P., reconstructed after Lambeck (1997)]. Moreover, due to the topography of the catchment basin, the Channel River drained a large area with inputs from the Rhine, Seine, Maas and Thames basins (Antoine et al., 2003). This topographic funneling effect was reinforced by the location of the British-Irish ice sheet, which reached its maximum extent at 16.7 kyr B.P. (McCabe and Clark, 1998). A simultaneous readvance of the Scandinavian ice sheet is recorded on the southeastern sector (Rinterknecht et al., 2006).

The onset of the H1 event, at 17 kyr B.P., is characterized by the sudden drop in the BIT index (Fig. 8.2E). As already observed for Heinrich events (Pailler and Bard, 2002), the biological productivity is low, but the BIT index indicates a predominant marine origin for the sedimentary organic matter (Fig. 8.2). Consistently, the C/N record exhibits a clear minimum over this time interval. Furthermore, a prominent $C_{37:4}$ alkenone peak is synchronous with the maximum abundance in lithic grains. A similar maximum of $C_{37:4}$ linked to H1 has already been described at other sites (Rosell-Melé, 1998; Bard et al., 2000). The fall in BIT index is clearly simultaneous with the rise in lithic grain abundance and percentage of $C_{37:4}$ alkenone, indicating that this switch was probably due to the impact on marine hydrology of icebergs coming from the Fennoscandian and the Laurentide ice sheets (Grousset et al., 2000). In parallel, the return to dry and cold conditions on the continent during H1 probably led to a regime with less fluvial runoff.

There was no recurrence of high BIT index values when warmer and wetter conditions returned during the Bølling-Allerød and the Holocene period (Fig. 8.2E). This is probably due to the sea-level rise of about 60 m compared with the LGM lowstand, which caused a northward displacement of the river mouth by about 300 km and thus a more attenuated influence of the Channel River at the core site (Fig. 8.1). Furthermore, due to the position of the Fennoscandian ice margins during the Bølling-Allerød, the meltwaters of the Peribaltic area drained into the southern part of the Baltic Basin and no longer through Poland and the Elbe Basin (Mangerud et al., 2004).

The abrupt runoff event that occurred at the onset of the last deglaciation on the European continent is unique in magnitude and timing and reflects an early reactivation of the European hydrological cycle leading to an intense discharge of terrestrial organic matter on the Celtic Margin and Bay of Biscay. The intensity of this event is due to a peculiar combination of topographic and palaeoclimatic factors: the large extent of the Fennoscandian and the British ice sheets, which coalesced over the North Sea, forced the drainage of rivers into the Channel River, thus creating one of the largest river systems ever existing on the European continent. The reactivation of European river runoff has also probably been fuelled by proglacial lakes developing at the southern margin of the ice sheets. Indeed, high abundances of remains of freshwater algae are found simultaneous with the large increase in BIT index (Fig. 8.2).

Interestingly, although the freshening of the surface waters starting at 21.5 kyr B.P. is progressive and parallel to the temperature increase after the LGM, the return to fully marine conditions is sharp and occurs in about a century at the start of the H1 event. As a result of sea-level rise, after 17 kyr B.P., conditions never became suitable again for recording events of the Channel River of such a magnitude in the Bay of Biscay.

Our results reveal large changes in the magnitude of the discharge of cold fresh water into the North Atlantic during the last deglaciation. This situation is similar to that reconstructed for the Laurentide ice sheet meltwater outflow which probably affected the meridional overturning circulation (Broecker et al., 1989; Manabe and Stouffer, 1997). Modeling experiments could help to evaluate the effect of the European river reactivation on the millennium-scale climatic events that punctuated the last deglaciation.

Acknowledgments
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Supporting Material

Analytical methods

After freeze-drying and grinding, sediments were analyzed at CEREGE for their nitrogen and organic carbon contents with a Fisons NA-1500 Elemental Analyzer [Carlo Erba NA-1500 Elemental Analyzer; see Pailler and Bard (2002) for details]. For lipid analysis, 1 to 5 g of sediment was extracted for biomarkers by the accelerated solvent extraction method (ASE 200 system, Dionex, California, USA) at 120°C and 100 bars with dichloromethane/methanol (9:1 v/v). The total lipid extract was analyzed at CEREGE for alkenone concentrations by gas chromatography (GC8000 Series Fisons) with flame ionization detection (GC-FID) [using analytical conditions similar to Sonzogni et al. (1997)]. % $C_{37:4}$ expresses the percentage of the tetra-unsaturated C_{37} alkenone among the total of C_{37} alkenones. Identification of alkenones is based on GC mass spectrometry (GC-MS; GC8000 MD800 Fisons) and quantities are based on the chromatographic peak areas. The total lipid extract was subsequently separated into polar and apolar fractions using a column packed with Al_2O_3 using hexane/dichloromethane (9:1, v/v) and dichloromethane/methanol (1:1, v/v) as eluents, respectively. The polar fraction was then filtered through a 0.45-µm, 4-mm diameter PTFE filter prior to injection. Glycerol dialkyl glycerol tetraethers were then identified and quantified at NIOZ by high-performance liquid chromatography/atmospheric pressure chemical ionization mass spectrometry using a HPLC/MS 1100 Series as described by Hopmans et al. (2004).

The age/depth scale is based on tie points shown as triangles on Figure 8.2A, which rely on ^{14}C ages (Auffret et al., 2002; Zaragosi et al., 2006) measured by accelerator mass spectrometry on monospecific samples of planktonic foraminifera *Neogloboquadrina pachyderma* (s.) or *Globigerina bulloïdes*. We calibrated these ^{14}C ages by using the Calib 5.0 radiocarbon calibration program (Stuiver and Reimer, 1993) with the Marine04 curve (Hughen et al., 2004), and an extension (Bard et al., 1998; Bard et al., 2004) for the three oldest ^{14}C ages. The age/depth model is then derived by a fifth order polynomial. The chronology is therefore fully independent of the GRIP and GISP2 records for the time span between 0 and 30 kyr B.P..