Terrigenous fluxes at the Celtic margin during the last glacial cycle

Gérard Auffret, Sébastien Zaragosi, Bernard Dennielou, Elsa Cortijo, David Van Rooij, Francis Grousset, Claude Pujol, Frédérique Eynaud, Martin Siegert

Abstract

The sedimentary sections of three cores from the Celtic margin provide high-resolution records of the terrigenous fluxes during the last glacial cycle. A total of 21 14C AMS dates allow us to define age models with a resolution better than 100 yr during critical periods such as Heinrich events 1 and 2. Maximum sedimentary fluxes occurred at the Meriadzek Terrace site during the Last Glacial Maximum (LGM). Detailed X-ray imagery of core MD95-2002 from the Meriadzek Terrace shows no sedimentary structures suggestive of either deposition from high-density turbidity currents or significant erosion. Two paroxysmal terrigenous flux episodes have been identified. The first occurred after the deposition of Heinrich event 2 Canadian ice-rafted debris (IRD) and includes IRD from European sources. We suggest that the second represents an episode of deposition from turbid plumes, which precedes IRD deposition associated with Heinrich event 1. At the end of marine isotopic stage 2 (MIS 2) and the beginning of MIS 1 the highest fluxes are recorded on the Whittard Ridge where they correspond to deposition from turbidity current overflows. Canadian icebergs have rafted debris at the Celtic margin during Heinrich events 1, 2, 4 and 5. The high-resolution records of Heinrich events 1 and 2 show that in both cases the arrival of the Canadian icebergs was preceded by a European ice rafting precursor event, which took place about 1–1.5 kyr before. Two rafting episodes of European IRD also occurred immediately after Heinrich event 2 and just before Heinrich event 1. The terrigenous fluxes recorded in core MD95-2002 during the LGM are the highest reported at hemipelagic sites from the northwestern European margin. The magnitude of the Canadian IRD fluxes at Meriadzek Terrace is similar to those from oceanic sites.

1. Introduction

Continental margins are sites of preferential deposition of terrigenous sediments eroded from
lands. The evolution of the terrigenous fluxes at the northwestern European margin and more specifically the comparison between those fluxes at glaciated and non-glaciated margins was the main objective of the ENAMII project (Table 1). Here we evaluate terrigenous fluxes at the Celtic margin, which constitutes the non-glaciated end member. The terrigenous flux is the product of land erosion exported to the ocean essentially through transport by rivers, and eventually by icebergs, sea ice and winds. Whilst the river inputs have been permanent with possible increase related to change in the atmospheric water budget and ice sheet melting, the input of ice-rafted debris (IRD) from icebergs has taken place mainly during a series of discrete events: the ‘Heinrich events’ (Heinrich, 1988; Andrew and Tedesco, 1992; Huon and Jantschik, 1993; Bond and Lotti, 1995; Dowdeswell et al., 1995; Auffret et al., 1996a; Chi and Mienert, 1996; Gwiazda et al., 1996a,b; among others).

The magnitude of the Quaternary terrigenous flux to the Celtic margin is essentially controlled by two factors: the climate regime and sea level variability. These factors control e.g. the weathering of soils, the transport capacity of rivers, ice sheet dynamics, the annual coverage of sea ice, and the prevailing wind/current regimes. All of which bear heavily on the fate of soils, their erosion and the transport of the eroded material to the deep ocean.

Most previous studies in the northeastern Atlantic (Grousset, 1977; Thiede, 1980; Auffret, 1985; Balsam et al., 1987; Cremer et al., 1992; Haflidason et al., 1998) have dealt with average, long-term flux variability. These studies demonstrated the important role played by the sea level and the climate regime on the accumulation rate of terrigenous and biogenic materials. However the recognition of the high degree of climatic variability that prevailed during the last glacial period (Heinrich events, Dansgaard–Oeschger cycles) has recently led to a focus on determining the origin and consequences of these ‘short-term events’. It is now well recognised that these events are very important in terms of sediment accumulation budgets (Cortijo et al., 1995; Manighetti and McCave, 1995; Baas et al., 1997; Abrantes et al., 1998; Hall and McCave, 1998).

In order to establish a high-resolution chronology for the last glacial cycle it is necessary to obtain cores from locations with sedimentation rates of at least 10–20 cm kyr⁻¹. The Celtic margin, facing the continental shelf of the Western Approaches, has been exposed to high terrigenous fluxes from the Celtic Sea delta during periods of low sea levels (Reynaud et al., 1999; Lericolais, 1997; Loncaric et al., 1998).

This study is based on IMAGES core MD95-2002 recovered from the Meriadzek Terrace, at a water depth of 2174 m (Auffret and Sichler, 1982), and two shorter cores: NORESTLANTE (N) KS12 from the Goban Spur and MODENAM (M) KS03 from the Whittard Ridge. Together these cores provide a record of sediment fluxes to the Celtic deep sea fan.

In order to allow correlation with the Greenland ice core GISP2 all stratigraphic references have been converted to calendar years, using the calibration curve of Stuiver et al. (1998) up to 20.7-kyr ¹⁴C ages and Bard (1998) between 20.7- and 36-kyr ¹⁴C ages.

2. Regional setting

During the Quaternary terrigenous material has been transported from the northwestern European continent to the deep sea along two main sedimentary pathways (Fig. 1). First, from the North Sea to the North Sea deep sea fan; this northern route has been heavily affected by erosion and accumulation processes related to the Fennoscandian ice sheet (Haflidason et al., 1998; Nygård et al., 2002). Second, from the English Channel to the Celtic deep sea fan (Auffret et al., 2000); this route has not been significantly affected by direct glacial process (Fig. 2), except at its northwestern boundary, in the vicinity of the Melville bank, where the maximum extension of the British ice sheet has been located (Scource et al., 1990). Shore lines have also been affected by large amplitude migrations (Lambeck, 1995) particularly at times of lowest sea levels, when the Channel River drained a large area including the
Rhine, Maas and Thames basins (Gibbard, 1988) and the Celtic Sea delta was developed (Lericolais, 1997).

Ice sheet expansion has also been of crucial importance. When the British and Fennoscandian ice sheets coalesced (Sejrup et al., 1994) a huge ice dam extended over the present North Sea and diverted fresh water and sediments from the North European rivers. Gibbard (1988) postulated recurrent developments of large pro-glacial lakes south of the ice barrier. The rupture of the embankments of these lakes probably resulted in major flooding events. Indeed, an extensive palaeovalley network has been described in the eastern English Channel (Larsonneur et al., 1982). No evidence of such a network exists in the western English Channel, with the exception of a local trough of possible fluvial origin (Lericolais, 1997). An incised palaeovalley network reappears near the shelf break (Bourillet and Loubrieu, 1995). Some of the buried valleys have superimposed sandbanks, which could be former channels of the Channel River delta (Berne et al., 1998; Reynaud et al., 1999). In some cases a direct con-

Table 1
Location of studied and reference cores

<table>
<thead>
<tr>
<th>Cores</th>
<th>Latitude (N)</th>
<th>Longitude (W)</th>
<th>Water depth (m)</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>ENAM93-21</td>
<td>62°44.01'</td>
<td>3°59.90'</td>
<td>1020</td>
<td>Abrantes et al. (1998)</td>
</tr>
<tr>
<td>91-1-89-03</td>
<td>60°38.30'</td>
<td>3°43.40'</td>
<td>300</td>
<td>Hallidason et al. (1995, 1998)</td>
</tr>
<tr>
<td>0DP-609</td>
<td>49°53.00'</td>
<td>24°14.00'</td>
<td>3884</td>
<td>Bond et al. (1992)</td>
</tr>
<tr>
<td>OM-5-K</td>
<td>48°50.07'</td>
<td>12°39.95'</td>
<td>2333</td>
<td>Hall and McCave (1998)</td>
</tr>
<tr>
<td>NKS12</td>
<td>49°10.29'</td>
<td>12°35.73'</td>
<td>1190</td>
<td>this work</td>
</tr>
<tr>
<td>T88-9P</td>
<td>48°23.03'</td>
<td>25°05.10'</td>
<td>3193</td>
<td>van Kreveld-Alfane et al. (1996)</td>
</tr>
<tr>
<td>MD95-2002</td>
<td>47°27.12'</td>
<td>8°27.03'</td>
<td>2174</td>
<td>this work</td>
</tr>
<tr>
<td>ESCAMP-02</td>
<td>60°38.30'</td>
<td>3°43.40'</td>
<td>2192</td>
<td>Loncaric et al. (1998)</td>
</tr>
<tr>
<td>MKS03</td>
<td>46°54.63'</td>
<td>10°03.60'</td>
<td>4540</td>
<td>this work</td>
</tr>
<tr>
<td>PO28-1</td>
<td>41°29.30'</td>
<td>9°43.30'</td>
<td>2160</td>
<td>Abrantes et al. (1998)</td>
</tr>
</tbody>
</table>

Fig. 1. Location of studied cores and reference cores, average current flow lines at 150 m below sea floor for the LGM (continuous line) and MWE (dashed line) after Seidov et al. (1996). The grey area corresponds to the Ruddiman belt (Ruddiman, 1977) extended to the southeastern Atlantic following Zahn et al. (1997). Open circles figure the possible route of icebergs originating from the Labrador Sea during Heinrich events.
The connection between the incised valleys and the upper course of canyons located on the Celtic margin has been demonstrated (Bourillet and Loubrieu, 1995). To the northwest Lambeck (1995) has reconstructed the evolution of the landscape around the present day Irish Sea. Small rivers flowing from a pro-glacial lake between 23 and 17 kyr could also have contributed sediments to the Celtic margin. There is however no evidence of an incised palaeovalley networks within the sea floor of the southern Irish Sea (Jean-François Bourillet, personal communication).

On the slope two canyon networks merge respectively, at the northwest within the Whittard canyon and at the southeast within the Shamrock canyon. Finally these networks join between the Trevelyan marginal plateau and the Whittard Ridge, from where the radiating pattern of the Celtic deep sea fan is developed (Auffret et al., 2000; Zaragosi et al., 2000).

The geological evolution of the fan has been reconstructed by Droz et al. (1999) from the seismic stratigraphy and correlation with the nearby DSDP Hole 400. Since its initiation during the early Miocene the fan has developed in three phases. During the last phase, encompassing the Pleistocene, the Whittard canyon drainage system, directly influenced by the development of the British ice sheet, appears as the major contributing source to the deep sea fan.
3. Materials and methods

3.1. Lithology, stratigraphy and palaeoenvironmental indicators

We have used core MD95-2002 (Bassinot and Labeyrie, 1996) as a reference core for the establishment of a high-resolution chronology for the last glacial cycle. As noticed in a number of other cores retrieved with the Calypso corer ‘over-sampling or stretching’ prevailed in the upper part of this core (Széreméti et al., in preparation). Based on the lithology and stratigraphy of the nearby core ESCAMP-02 (Loncaric et al., 1998) we have applied a depth correction to the first 12 m of the MD95-2002 (see Appendix 1). Our depth scale adjustment is supported by a significant improvement in the fit of a synthetic seismogram derived from MD95-2002 to the 3.5-kHz data. Such a depth correction is vital for the evaluation of the sediment fluxes in MD95-2002. Cores NKS12 (7 m) and MKS03 (5 m) are conventional Kullenberg piston cores collected by R/V Jean Charcot and R/V Le Noroit respectively. The age models are based on 16 AMS 14C datings for core MD95-2002, and five AMS 14C datings for MKS03. The age model of core NKS12 has been obtained by cross-core correlation with MD95-2002 based on their gamma-density logs.

Bulk densities for core MD95-2002 have been measured on board R/V Marion Dufresne by gamma ray attenuation with the GEOTEK Multi-Sensor Core Logger (MSCL). Gamma ray attenuation was calibrated with 6-mm-thick aluminium plates inside a liner (Bassinot and Labeyrie, 1996). Wet and dry bulk densities were also measured on discrete subsamples by wet and dry weighting a known volume of sediment, sampled with a syringe. Comparison of the results shows that the gamma-density is typically ~ 0.21 g cm\(^{-3}\) higher than the wet bulk density. In core MD95-
2002 the relation between wet bulk densities \( (T_{\text{wet}}) \) and log densities is:

\[
\text{wet bulk density} = (0.8247 \cdot \text{gamma} - \text{density}) + 0.0948
\]  

(1)

Dry bulk density \( (T_{\text{dry}}) \) has been calculated through relation (2), assuming a grain density of 2.65 g cm\(^{-3}\) and an interstitial water density of 1.024 g cm\(^{-3}\):

\[
T_{\text{dry}} = 2.65 \cdot (10.24 - T_{\text{wet}})/(10.24 - 2.65)
\]  

(2)

Bulk densities for core MKS03 and NKS12 were also derived from gamma-density measurements performed with the GEOTEK MSCL.

All cores were visually described and X-ray images obtained for MD95-2002 and MKS03 with the SCOPIX system (Migeon et al., 1999). Sub-samples were then obtained at intervals ranging from 1 to 2 cm for MD95-2002 and 2 to 10 cm for MKS03 and NKS12. Carbonate content measurements (Bernard calcimeter, precision \( \pm 2\% \)) and limited grain size analyses with the laser microgranulometer (Coulter counter LS130; size range \( 0.4 \mu m \) to 1 mm) were made on the total sediment fraction. Following which about 15 g of sediment was dried and wet sieved in order to separate the coarser than 150-\( \mu m \) size fraction. Qualitative and quantitative analysis of the lithic grains was then performed on this size fraction in cores MD95-2002 and NKS12. In the latter, magnetic grains were extracted with a hand magnet for examination with an electron-scanning microscope.

In the same size fraction (coarser than 150 \( \mu m \)) of cores MD95-2002 and MKS03 qualitative and quantitative analyses of the microfauna were performed. More than 300 planktonic foraminifera, separated with a micro-splitter, were identified and counted, specimen were also hand-picked for AMS \(^{14}\)C dating and oxygen isotope analysis.

Stable isotopes \( \delta^{18}O \) measurements were performed using a Finnigan MAT 251 mass spectrometer coupled with an automated carbonate preparation device. Five to six specimens of \( \text{Globigerina bulloides} \) or \( \text{Neogloboquadrina pachyderma} \) were used for each measurement and the external reproducibility is 0.5\% for \( \delta^{18}O \). All values are given versus PDB after calibration via the NBS 19 standard. The August sea surface temperatures (SST) have been estimated from core MD95-2002 using the MAT 556 transfer function of Pflaumann et al. (1996). Sea surface salinity (SSS) anomalies were obtained following the methods developed by Duplessy et al. (1991) and Shackleton (1974).

The dinocyst assemblages have been studied using standard protocol (de Vernal et al., 1996) summarised in Eynaud et al. (2000). From these results, the annual duration of the sea ice coverage has been evaluated using the transfer function established by Guiot (1990) and de Vernal et al. (1993, 1998).

3.2. Isotopic (Sr–Nd) study of the coarse lithic fraction

The \( \delta^{18}O \) of the lithic (non-calcareous) coarse fraction from 24 samples of core MD95-2002 has been analysed, in order to trace the province source of the grains (Table 2). Approximately 15 g of bulk sediment was washed with tap water and dried. The sediment was then washed over a 150-\( \mu m \) mesh sieve and the coarse fraction dried at 60°C. The carbonates were leached using 0.6 N HCl and the remaining ‘coarse lithic’ fraction digested using HF+HClO\(_4\). Finally, Sr and Nd were chemically separated through ionic chromatographic columns and the isotopic ratios measured using a mass spectrometer (Grousset et al., 1988; Grousset and Biscaye, 1989).

3.3. Chronological framework

In order to link the history of terrigenous fluxes at the Celtic margin to the record of air temperatures over Greenland inferred from the ice core oxygen isotope signal (Grootes and Stuiver, 1997) one needs to rely on a high-resolution age model with ages expressed in calendar years. A correction of 400 yr for reservoir age has been applied to the AMS ages and the corrected ages translated in calendar ages following Stuiver et al. (1998) for corrected \(^{14}\)C ages younger than 20.7 kyr and Bard (1998) for ages between 20.7 and 36 kyr.
The age model for core MD95-2002 is based on 16 AMS $^{14}$C samples and five additional stratigraphic control points (Table 3). Within the range of AMS $^{14}$C ages the additional pointers are the AMS age of a foram event in core ESCAMP-02 (Loncaric et al., 1998) and the average age of Heinrich layer 3. Beyond the range of AMS $^{14}$C ages, we also used the average ages of Heinrich layers 5 and 6, as well as the age of the dominance of the foraminifera species $G$. truncatulinoides sinistral within isotopic stage 5 in the same area (Pujol and Turon, 1986). The reference ages for Heinrich layers 3, 5 and 6 are from Elliot et al. (1998). The isotopic stratigraphy is based on the correlation of the $\delta^{18}$O record with the SPECMAP time scale (Martinson et al., 1987).

Ages between the stratigraphic references have been calculated by linear interpolation. Ages calculated before the youngest reference have been extrapolated with the sedimentation rate derived from the two youngest references. Ages calculated after the oldest reference have been extrapolated from the sedimentation rates derived from the two oldest references. The age model for core MKS03 (Table 4) is based on five AMS $^{14}$C age datings and the change in coiling ratio of $G$. hirsuta and $G$. truncatulinoides at 11.1 kyr (Pujol, 1980). The age model for core NKS12 (Table 5) is based on 11 stratigraphic markers including the average ages of Heinrich events 1–6.

### 3.4. Calculation of the terrigenous fluxes

Bulk sediment accumulation rates (total flux) ARb in g cm$^{-2}$ kyr$^{-1}$ were calculated with Eq. 3:

$$\text{ARb} = \text{LSR} - T_{\text{dry}}$$

(3)

**LSR**: linear sedimentation rate in cm kyr$^{-1}$; $T_{\text{dry}}$: dry bulk density in g cm$^{-3}$.

Accumulation rates for non-carbonate (ARresidual) expressed in g cm$^{-2}$ kyr$^{-1}$ were calculated with Eq. 4:

$$\text{ARresidual} = \text{ARb} \times (1 - \text{carbonate content})$$

(4)
Since opaline silica is very scarce and since there is no evidence of diagenetic neoformation, the non-carbonate fraction can be considered to represent the terrigenous component.

Accumulation (ARk) of lithic grains larger than 150 μm (LLG) expressed in # cm\(^{-2}\) yr\(^{-1}\) was calculated with Eq. 5:

\[
ARk = Cc \cdot ARb
\]  
(5)

With Cc (large lithic grain (LLG) contents) in # grains (g dry sediment)\(^{-1}\).

4. Results

4.1. Lithology

From the visual examination of the cores, X-ray radiography, calcium carbonate and LLG contents, four sediment types have been distinguished in the studied cores (Fig. 3–5).

4.1.1. Pelagic and hemipelagic sediments

Pelagic sediments consist in light grey nannofossil ooze (CaCO\(_3\) > 60%), hemipelagic sediments consist mainly in bioturbated light brownish grey marly nannofossil ooze (30% < CaCO\(_3\) < 60%), both types have been deposited at the three sites during marine isotopic stages (MIS) 5 and 1.

4.1.2. Terrigenous sediments

The terrigenous sediments are represented by calcareous mud, with carbonate contents ranging between 10 and 30%.

Particular facies have been distinguished within this subgroup.

4.1.2.1. Mud exhibiting black colours banding

The banding consists of black laminae a few millimetres thick at intervals of a few centimetres. The black colour is attributed to the presence of hydrotroilite (iron sulphide) which is oxidised rapidly when exposed to air. In core MD95-2002 this facies is present in sediment deposited at...
the end of MIS 3 and a large part of MIS 2. This type of banding has been observed in several deep environments (Caralp et al., 1981; Cremer, 1982; Nelson et al., 1992; Normark and Damuth, 1997).

4.1.2.2. Mud exhibiting clayey laminae. The millimetre scale laminations (Plate 1) are only visible in the X-ray images of core MD95-2002. Their mean grain size is very small, ranging from 5 to 6.5 μm. The X-ray images occasionally show gravel size drop stones included in the laminated sediments (Plate 1).

4.1.2.3. LLG rich mud. The proportion of LLG exceeds 500 grains g⁻¹ in decimetre thick layers of cores MD95-2002 and NKS12 (Fig. 3 and 4). These sediments are bioturbated and the LLG are clearly visible in X-ray images (Plate 1).

4.1.2.4. IRD rich mud. Six sediment layers prac-

Fig. 3. Lithology of core MD95-2002, the position of samples in which LLG were counted is indicated by dots.
tically barren of foraminifera (Plate 1) are interbedded within the LLG rich mud of cores MD95-2002 and NKS12. Four of them are also characterised by a high magnetic susceptibility and a high density. The characteristics of these centimetres thick layers and their ages allow us to correlate them with Heinrich layers 1–6.

By means of textural analyses of bulk and decalcified sediments we have evaluated the percentage of total carbonate in the size fraction coarser than 150 μm of layers HL1 to HL5. We have then evaluated the abundance of the detrital carbonate as the difference between total carbonate and the estimated percentage of biogenic carbonate in this size fraction. This evaluation confirms the results of Loncaric et al. (1998) with only trace amounts of detrital carbonates for Heinrich layer 1 and up to 9% of the IRD in Heinrich layer 2 (Table 2).

The examination and chemical analyses of magnetic grains from Heinrich layer 2 in core NKS12 have allowed us to identify a set of minerals including feldspars, amphiboles, pyroxenes and olivines. These observations confirm that the high magnetic susceptibility of the Heinrich layers is related to the high proportion of plagioclase feldspars with titanomagnetite inclusions (Robinson et al., 1995). The X-ray images generally show clear lower and upper boundary and most often homogeneous internal facies within the layers. In core MD95-2002 however, a layered structure is observed in Heinrich layer 4 (Plate 1).

4.1.3. Turbiditic mud

This facies is restricted to MIS 1 and 2 of core MKS03 from the Whittard Ridge (Plate 1). It consists in alternating silt and clay laminae. These sediments are organised in sequences of alternating silty and clayey intervals. They are separated
by sharp boundaries and their thickness varies from 1 to 20 cm. The number of silt laminae in a sequence varies from one to 15; sometimes the base laminae show cross-stratifications. Within each sequence there is an upward decrease in thickness of individual silt laminae (thinning up). The mean grain size varies from 5 μm (clayey intervals) to 50 μm (silt laminae). This facies is interpreted (Zaragosi et al., 2000) as fine-grained turbidites: Td and Te divisions of the classic Bouma turbidite sequence (Bouma, 1962).

4.2. Isotopic (Sr–Nd) analysis of coarse lithic fractions

Results of the εNd(o) analyses of the lithic (non-calcareous) coarse fraction are given in Table 2.

(1) A European origin is characterised by radiogenic εNd(o) (−15 to −42); the LLG are derived from young formations (< 1 Ga). European origin material is identified in three deposits:
   – the ‘background glacial deposits’
   – HL3 and HL6
   – LLG rich mud immediately above and below HL1, HL2, HL4 and HL5

(2) A Canadian origin is characterised by un-radiogenic εNd(o) (−42 to +10); LLG are derived from old shields (2–3.5 Ga). Moreover, probably due to the plagioclase feldspar abundance, these samples are also characterised by a strong europium anomaly. The Canadian origin is identified only in HL1, HL2, HL4 and HL5.

(3) An Icelandic origin is characterised by very radiogenic εNd(o) (0 to +10); LLG are derived from very young ‘mantle-derived’ rocks (< 0.3 Ga). The Icelandic LLG are a ubiquitous component in all Heinrich layers (in particular in the ‘atypical’ Heinrich layers 3 and 6).
4.3. Quantitative evaluation of the terrigenous fluxes

At Meriadzek Terrace seven episodes can be distinguished within the terrigenous flux history of the last glacial cycle (Fig. 6). The first episode (120–71 kyr) corresponds to MIS 5. It is characterised by low terrigenous inputs at Meriadzek (5 g cm\(^{-2}\) kyr\(^{-1}\)) and Goban Spur (1 g cm\(^{-2}\) kyr\(^{-1}\)). During this time period the sea level was at about the same height as today, the air temperature over Greenland and the SST in the Bay of Biscay were relatively warm.

Episode 2 corresponds to the period between 71 and 46 kyr BP (Fig. 6) and is characterised by moderate terrigenous flux (30–20 g cm\(^{-2}\) kyr\(^{-1}\)) at Meriadzek. During this time the sea level was on average about 40 m below present.

Episode 3 spans the interval between 46 kyr and Heinrich event 2 (23.8 kyr). Terrigenous fluxes are still moderate with an average of 30 g cm\(^{-2}\) kyr\(^{-1}\) at Meriadzek Terrace and 21 g cm\(^{-2}\) kyr\(^{-1}\) at Goban Spur. The sea level was decreasing and the climatic conditions worsened during this episode. Some of the interstadial Dansgaard–Oeschger oscillations (interstadials 8, 4, 3, 2) shown in the variation of the air temperature over Greenland (Fig. 6) are also well expressed in the Meriadzek SST.

Episode 4 (23.8–18 kyr) corresponds to the first part of the Last Glacial Maximum (LGM), defined as the period between Heinrich 2 and 1 (Schneider et al., 2000). In core MD95-2002 (Figs. 6, 7), terrigenous fluxes were relatively high (40–50 g cm\(^{-2}\) kyr\(^{-1}\)) during Heinrich event 2. Maximum terrigenous fluxes occurred after the event, between 23.6 and 23.4 kyr. At this time the fluxes increased from 50 to 140 g cm\(^{-2}\) kyr\(^{-1}\). This increase occurred during a period of very low SSS and decreasing sea ice coverage. Fluxes then decreased again to 50 g cm\(^{-2}\) kyr\(^{-1}\) while the sea level decreased to 100 and 120 m (Fairbanks, 1989). The Irish ice sheet model (Siegert and Dowdeswell, 2002) suggests that the calving of
icebergs and incipient melting began respectively at 25 and 20 kyr.

Episode 5 (18–16.7 kyr) is characterised at Meriadzek Terrace by a paroxysmal increase (150 g cm\(^{-2}\) kyr\(^{-1}\)) which occurred in about 100 yr (Figs. 6, 8). This increase occurred simultaneously with a pronounced decrease of the SSS and peak occurrences of L. machaerophorum, a dinocyst of estuarine affinity (Morzadec-Kerfourn, 1977), and Pediastrum sp., a fresh water algae. These paroxysmal fluxes ended precisely before the arrival of Canadian icebergs in the Bay of Biscay.

Episode 6 (16.7–11 kyr) represents the deglaciation (Fig. 6). The terrigenous fluxes at Meriadzek (18 g cm\(^{-2}\) kyr\(^{-1}\)) and Goban (9 g cm\(^{-2}\) kyr\(^{-1}\)) decreased drastically though the sea level was still relatively low. During Heinrich event 1 terrigenous fluxes were relatively low (10 g cm\(^{-2}\) kyr\(^{-1}\)) as generally noticed in the Northeastern Atlantic (Table 6). On the Whittard Ridge, deposition from turbidity current overflows prevailed which account for high average terrigenous sediment fluxes of about 45 g cm\(^{-2}\) kyr\(^{-1}\).

During Episode 7 (11 kyr to Present) fluxes became everywhere very low as during MIS 5.

In core NKS12 (Fig. 6) fluxes are generally lower than at Meriadzek Terrace, particularly during the LGM. They are about the same intensity during the Holocene. In core MKS03 (Fig. 6) from the Whittard Ridge, the fluxes at the beginning of MIS1 (45 g cm\(^{-2}\) kyr\(^{-1}\)) are higher than those
4.4. LLG and IRD fluxes

From the result of the lithological, geochemical and micropalaeontological studies we have identified Heinrich layers 1–6 in MD95-2002 and NKS12 (Figs. 3, 4 and 9). As criteria we have used the magnetic susceptibility for HL1, 2, 4 and 5 and the abundance of LLG and the dominance of *Neogloboquadrina pachyderma* (s) for HL3 and HL6.

In order to avoid a misinterpretation of our results from the European margin, we have chosen to interpret LLG as IRD only when a Canadian origin has been clearly established or when the LLG rich layers were stratigraphically correlable with Heinrich layers 3 and 6. In the following we also use the term IRD events (IRDE) instead of Heinrich events because the rafting of IRD at the Celtic margin may correspond only to part of

Fig. 7. (a) Terrigenous flux in core MD95-2002 from 26 to 22 kyr; (b) magnetic susceptibility; (c) sea level; (d) August SST; (e) SSS anomaly. The positions of 14C analysis are shown by stars.
the duration of the Heinrich events established in other regions of the North Atlantic.

4.4.1. Ice rafting event 6 (IRDE6: 65–62 kyr)

Event 6 (Fig. 9) has been recognised in core MD95-2002 and occurred before the GISP2 ice core interstadial 18. It corresponds to relatively high magnetic susceptibility level and the beginning of a substantial negative SSS anomaly. The event has been preceded by an LLG input, which may represent a European precursor, also identified at Goban Spur.

4.4.2. Ice rafting event 5 (IRDE5: 45.5–43.8 kyr)

At Meriadzek Terrace and Goban Spur event 5 (Fig. 9) is characterised by a moderate input of IRD at 45.5 kyr (20 grains cm$^{-2}$ yr$^{-1}$). At the time of ice rafting in the Bay of Biscay, SST was about 3°C (Fig. 6), a SSS anomaly prevailed and the ice cover lasted about 1 month yr$^{-1}$. It is followed at 43.8 kyr by a second input of LLG of European origin.

4.4.3. Ice rafting event 4 (IRDE4: 41–39 kyr)

At Meriadzek Terrace, event 4 is characterised
by two inputs of Canadian IRD (Fig. 9). This is indicated by two peaks of magnetic susceptibility, both levels bearing a Canadian Nd signature. The first arrival (about 60 grains cm$^{-2}$ yr$^{-1}$) occurred at about 40 kyr. During the event, SST (Fig. 6) was about 2°C, a well defined SSS anomaly prevailed and the ice cover duration decreased from nearly 10 months to 1 month. At Goban Spur IRD fluxes are maximum reaching 150 grains cm$^{-2}$ yr$^{-1}$, detrital carbonate is common.

4.4.4. Ice rafting event 3 (IRDE3: 32.5–29.6 kyr)

At Meriadzek Terrace four pulses of IRD of moderate intensity (20 grains cm$^{-2}$ yr$^{-1}$) are recorded (Fig. 9). During this time period, SST (Fig. 6) was about 2°C, a SSS minimum prevailed and the annual ice cover persisted for about 10 months. At Goban Spur the IRD fluxes were also of moderate intensity (15 grains cm$^{-2}$ yr$^{-1}$).

4.4.5. Ice rafting event 2 (IRDE2: 23.8–23.6 kyr)

At Meriadzek the maximum Canadian IRD flux, at 23.7 kyr, reached 150 grains cm$^{-2}$ yr$^{-1}$, and includes a significant detrital carbonate component. It is immediately followed by a peak of LLG of similar intensity at 23.4 kyr (Fig. 10). Both SST and SSS were at a minimum and the annual duration of the sea ice coverage reached a maximum at the end of the event. Noteworthy is the fact that the LLG fluxes showed a first maximum (20 grains cm$^{-2}$ yr$^{-1}$) at 25.1 kyr, i.e. 1.4 kyr before the arrival of the Canadian icebergs. At Goban Spur, the maximum IRD flux reached 75 grains cm$^{-2}$ yr$^{-1}$ and detrital carbonate was common in the lithic fraction.

4.4.6. Ice rafting event 1 (IRDE1: 16.7–16.4 kyr)

At Meriadzek Terrace a first peak in the LLG fluxes occurred at 17.7 kyr (Fig. 11). The deposition of the Canadian IRD characterised by their Nd/Sr ratio occurred between 16.7 and 16.4 kyr, the total fluxes reaching 40 grains cm$^{-2}$ yr$^{-1}$. It was preceded by a maximum flux (200 grains cm$^{-2}$ yr$^{-1}$) of LLG from European origin. SST and SSS reached their minimum values at the end of IRDE1 when sea ice coverage was in the order of 6–10 months yr$^{-1}$. It is followed by a last peak of LLG from European origin (45 grains cm$^{-2}$ yr$^{-1}$). At Goban Spur the IRD flux was of about the same intensity.

5. Discussion

5.1. Evolution of terrigenous fluxes and IRDE events

The long-term evolution of the terrigenous fluxes (Figs. 6 and 12) is related to the climatic and sea level changes that occurred during the last glacial cycle (Cremer et al., 1992; Abrantes et al., 1998; Haflidason et al., 1998). However the flux evolution also shows steep transitions between the various episodes. The low fluxes of Episode 1 (120–71 kyr) are related to a relatively mild climate and high sea levels favouring the accumu-
tion of sediment over the continental shelf. Episode 2 (71–46 kyr BP) was characterised by moderate fluxes at Goban Spur and Meriadzek Terrace. These fluxes were higher at Meriadzek Terrace during the beginning of the episode, which also corresponds to a period of decreasing sea level (Labeyrie et al., 1987).

During Episode 3 (46–23.8 kyr) terrigenous fluxes increased at Meriadzek Terrace and Goban Spur (Figs. 6 and 12). It is during this period that the Fennoscandian ice sheet (Fig. 12) began to expand in size (Siegert and Dowdeswell, 2002). A blockage of the North Sea by this ice sheet occurred between 34 and 27 kyr. It is likely that in such a situation all the waters and sediment loads from the northern European plumes (Rhine, Maas, Thames..) would have been considerably enlarged the Channel River. According to Mulder and Syvitski (1996) the drainage basin of the Channel River, including the Elbe, could have reached an area larger then 2 000 000 km². According to Van Vliet-Lanoé and Guillocheau (1995) an important soil erosion phase also occurred at about 33 kyr at the latitude of France.

Fig. 9. (a) LLG fluxes in core MD95-2002; (b) magnetic susceptibility; (c) $\varepsilon_{\text{Nd}}$ of the non-carbonate lithic fraction coarser than 150 $\mu$m; (d) SSS anomaly; (e) ice $\delta^{18}$O from GISP2.
and Belgium. The first occurrences of the black banding facies appear at Meriadzek Terrace during this period, this facies is common in area facing high fresh water discharges (Normark and Damuth, 1997). It is specifically located within the Bay of Biscay with an extension off Cap Sines on the Iberian Margin (Thiede, 1971) and has previously been correlated with the continental
The increasing terrigenous fluxes of Episode 4 (23.8–18 kyr) correspond to an improvement of the climate as shown by the relatively high SST (Figs. 6, 7 and 12). As already suggested by de Vernal and Hillaire Marcel (2000) this result is contradictory with the CLIMAP reconstitution...
A period of paroxysmal fluxes occurred at 23.5 kyr, immediately after the arrival of the Canadian icebergs (Fig. 7). This is coincident with the period of European IRD rafting characterised by very cold SST and low SSS. In addition to fluvial inputs these paroxysmal fluxes most probably include a significant contribution from European icebergs, although for the LGM (CLIMAP, 1981). A period of paroxysmal fluxes occurred at 23.5 kyr, immediately after the arrival of the Canadian icebergs (Fig. 7). This is coincident with the period of European
we are presently unable to quantify this specific flux. The Irish ice sheet model indicates that the calving of icebergs and incipient melting began respectively at 25 and 20 kyr. This suggests that the SSS evolution at Meriadzek Terrace is largely related to the in situ melting of European and Canadian icebergs. A new phase of soil erosion also occurred at about 21 kyr on the continent (Van Vliet-Lanoe and Guillocheau, 1995). The black banding facies already noticed during Episode 3 became more common, which is contradictory with its earlier correlation with the Würm 2–3 continental interstadial. During this period, both high sediment fluxes and surface water stratification may have contributed to a decrease of the oxygen content and the instauration of anoxic conditions at the sediment interface.

During Episode 5 (18–16.7 kyr) terrigenous fluxes became very high at Meriadzek Terrace (Figs. 8 and 12) and the sea level was decreased to 100 and 120 m below present levels (Labeyrie et al., 1987; Fairbanks, 1989). The sediments deposited during this period of paroxysmal fluxes (between 18 and 17.2 kyr) display a characteristic laminated facies including LLG. The fine laminations suggest a sedimentation process producing a rapid accumulation, such as deposition from turbid plumes. The origin of such plumes, cascading (as postulated for the interpretation of a similar facies in the Gulf of St-Lawrence by Hesse and Khodabasksh, 1998), hyperpycnal flows or iceberg melting are not clear to us. The dinocyst assemblages from this interval effectively indicate fresh water as well as estuarine inputs (Eynaud, 1999). A steep decrease of the SSS occurred precisely at the onset of drastic fluxes (Fig. 8). However as illustrated in Fig. 12, this meltwater ‘event’ (MWE) coincided with the period of maximum calving of the Irish ice sheet and most of the fresh water may represent the product of in situ iceberg melting rather than a dilution by fluvial water.

According to McCabe and Clark (1998) the Irish ice sheet margin retreated markedly from 21 to 18 kyr, as a consequence glacial materials have been exposed to erosion. An important phase of shelf sediment erosion is also indicated by the abundance of reworked dinocysts (Eynaud, 1999), such erosion is in agreement with the evolution of the Celtic banks as described by Berné et al. (1998). Our interpretation is that the flux event recorded at Meriadzek before Heinrich event 1 resulted from the conjunction of a number of factors: an early phase of deglaciation (Zaragosi et al., 2001), a very low sea level allowing the maximum seaward extension of the Celtic Sea delta, the erosion of older delta deposits possible cascading processes and ice rafting from European icebergs.

During Episode 6 (16.7–11 kyr) the fluxes decreased drastically though the sea level was still relatively low (Fig. 6). On the Whittard Ridge deposition from turbidity current overflows prevailed which accounts for the relatively high terrigenous sediment flux.

In summary, the comparison of the fluxes at Meriadzek Terrace and Goban Spur (Fig. 6) shows that they were higher at former, particularly after 30 kyr. This corresponds to the time at which the North Sea was blocked by ice and as a consequence most fluvial inputs were diverted to the Channel River. The compilation of flux data from the northeastern Atlantic (Table 6) shows that the terrigenous fluxes recorded at Meriadzek Terrace during the LGM are the highest recorded at the northeastern Atlantic margins. These very high fluxes are mainly related to the conjunction of fluvial inputs from the Celtic Sea delta and ice rafting from European icebergs. Following Heinrich event 1, maximum terrigenous fluxes are recorded in the Norwegian Sea, where the effects of the deglaciation occurred later. Importantly, the terrigenous fluxes at Meriadzek during Heinrich event 1 (Table 6) were of the same magnitude as recorded in core T88-90 from the oceanic domain at about the same latitude (van Kreveld-Alfane et al., 1996). This suggests that during this event the western European margin did not receive a high flux of sediment due to the blockage of the transport system within frozen rivers. Finally, the fluxes recorded at Meriadzek Terrace for the late Holocene are about the same value as those evaluated by Van Weering et al. (1998).

5.2. Ice rafting events (IRDE)

Whilst it is quite clear that LLG consisting of
continental crust debris present in sediments from the Ruddiman belt (Ruddiman, 1977; Grousset et al., 1993; Robinson et al., 1995; Revel et al., 1996) have been deposited from drifting icebergs, the source of lithic grains on the European continental margin is not so obvious because a number of processes could provide alternative means for the erosion and transport of fine-grained sands in the size range from 0.1 to 1 mm to the deep sea. Gravity processes such as low-density turbidity currents or hyperpycnal or density flows are possible processes, as well as deposition from melting sea ices that could have drifted offshore. According to Mulder and Syvitski (1996) the occurrence of hyperpycnal flows should have decreased during the LGM, as riverine inputs from the low sea stage should have been large and their bed load diluted. The main problem remains the distinction between grains rafted from sea ices, fluvial ices and grains rafted from icebergs. There is to our knowledge no direct mean to distinguish between these different sources, however a high ratio of lithic to the total grain in the coarser than 150-μm fraction and the isotopic δ18O anomaly associated classically to the melting of icebergs are two criteria often associated to the IRD identification.

Based on the size distribution of sediment, the scarcity of foraminifera and nannofossils and the SSS salinity anomaly generally associated, we assume that most of the LLG grains deposited at Meriadzek Terrace between 25.5 and 24.5 kyr (European precursor), 23.6 and 23.4 kyr (European phase following IRDE2), 18 and 16.6 kyr (European precursor and European event preceding IRDE1) and 16.4 and 16 kyr (European phase following IRDE1) can effectively be considered as IRD. This assumption is in agreement with the occurrence of such LLG rich levels from Goban Spur (Scource et al., 2000) to the Iberian margin (Snoeckx et al., 1999; Grousset et al., 2000), interpreted as IRD.

The detailed timing of the IRD fluxes for IRDE2 and 1 allow us to compare their respective scenario. At the Celtic margin (Fig. 10) the ice rafting peak of Canadian IRD during event 2 (at 23.7 kyr) was preceded by a peak of European IRD (European precursor 2: EP2) occurring some 1.4 kyr earlier. The Canadian IRD deposition constitutes the first half of the 400-yr-long period of major IRD fluxes, which also ended by the deposition of European IRD of the same magnitude (European event 2: EE2). As pointed out in Grousset et al. (2000) the European precursor coincides precisely with a dust event in the GISP2 ice core (Mayewsky et al., 1994). Event 1 is also preceded by a European precursor event (EP1), the peak of which at 17.7 kyr predates of about 1 kyr the deposition of Canadian IRD (Fig. 11). The event itself is immediately preceded by a European event of greater magnitude (EE1). What is common between the two events is the occurrence of a European precursor ice rafting episode before the rafting of Canadian IRD at the Celtic margin. However their scenario differs in the succession of the phases, in the case of Heinrich event 2 the Canadian rafting episode leads a major European event, whilst in the case of Heinrich event 1, it is the European event which leads the Canadian influx.

We have also to consider the scenario of the events established at Meriadzek within the chronological framework established for Heinrich events 2 and 1 in the northeastern Atlantic. Here again we refer to these events as the period at which IRD from Canadian origin have been rafted in each site, the criteria for such an origin being either detrital carbonates or the Nd/Sr signature and the high magnetic susceptibility. There are relatively few sites in the northeastern Atlantic were a Canadian origin has been established with a detailed time scale. At DSDP site 609 (Bond et al., 1992) the deposition of IRD from Canadian origin ended at 24.8 kyr, i.e. 1.0 kyr before the onset of the event at Meriadzek. However IRD (from unknown origin) rafting at this site still prevailed during the deposition of Canadian IRD at Meriadzek. At Goban Spur a precursor event consisting in the deposition of IRD containing Cretaceous chalk eroded from the continental shelf by a tongue of the ice sheet (Scource et al., 1990, 2000) also occurred, as at Meriadzek, about 1 kyr before the arrival of Canadian icebergs. The peak deposition of dolomite from Canadian origin occurred at about the same time as the beginning of IRE 2 at Meriadzek and IRD deposition...
ended at the same time as they did at Meriadzek (Scourse et al., 2000). The land data (McCabe and Clark, 1998) indicate that after a major retreat from 21 to 18 kyr the Irish ice sheet readvances notably during Heinrich event 1.

In terms of IRD fluxes there are also relatively few published data expressed in number of grains cm\(^{-2}\) yr\(^{-1}\) (Table 7). The comparison between our data from Meriadzek and Goban Spur indicates that the fluxes were larger at Meriadzek Terrace during event 2. At the DSDP site 609 (Bond et al., 1992) where the fluxes are well constrained, the rates for event 1 and event 2 are respectively 11 and 20 grains cm\(^{-2}\) yr\(^{-1}\). However the comparison with data from the southeast Atlantic (van Kreveld-Alfane et al., 1996) shows a remarkable correspondence with fluxes recorded at Meriadzek, with the exception of Heinrich event 1, during which they are higher at the oceanic site.

### 5.3. Terrigenous input by gravity currents

Most of the sediments from core MKS03 sampled on the Whittard Ridge have been deposited from currents generated by gravity processes.

From stage 2 to the early–late Holocene transition, the overflow deposits give evidence of the activity of turbidity current flows in the slope canyons. During this period the broad Celtic Sea delta developed, in this configuration a wide size spectrum of material was available for transport to the deep sea. These conditions are comparable to those of passive margin fans located downstream large rivers as the Amazon and the Mississippi.

At the end of MIS 2 and during the deglaciation turbiditic supplies continue to be deposited. This period corresponds to the rapid deglaciation phase when meltwater discharges and sediment loads were higher than during the maximum low-stand when part of the source area was glaciated. This configuration is similar to the Mississippi Fan one before the Younger Dryas (Broecker et al., 1989; Twichell et al., 1991). Associated to the meltwater discharge, the erosion of the delta induced by the sea level rise (Lericolais, 1997) would have allowed significant sediment supply to the deep sea. Thus, the turbiditic overflow deposits located on the Whittard Ridge contain particles originated from: (1) the North European palaeo-rivers during MIS 2; (2) the meltwater discharges during the European deglaciation; (3) the erosion of the Channel River delta during the sea level rise. The late Holocene is marked by the interruption of turbiditic overflow deposition on the Whittard Ridge, which implies the disappearance of the active delta environments on the shelf.

### 5.4. Evolution of terrigenous fluxes and environmental changes during the last 40 kyr

During the last 40 kyr there is a generally good synchronicity between the evolution of SST in the Gulf of Biscay and the air temperature over Greenland (Fig. 13). Interstadials 1B, 2, 3, 4 and 8 identified in the GISP2 ice core correspond to warm SST in the Gulf of Biscay. We also note that the terrigenous flux increases during the lowering phase of the sea level.

The SSS evolution in the Bay is closely related to the calving of icebergs from the Irish ice sheet, however very low salinities followed the arrival of the Canadian icebergs. The abundance of the

---

Table 7

<table>
<thead>
<tr>
<th>Core</th>
<th>HE1</th>
<th>HE2</th>
<th>HE3</th>
<th>HE4</th>
<th>HE5</th>
<th>HE6</th>
</tr>
</thead>
<tbody>
<tr>
<td>ENAM93-21</td>
<td>40</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ODP 609</td>
<td>11</td>
<td>20</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>NKS12</td>
<td>40</td>
<td>80</td>
<td>15</td>
<td>150</td>
<td>20</td>
<td></td>
</tr>
<tr>
<td>T88-9P</td>
<td>200</td>
<td>140</td>
<td>–</td>
<td>75</td>
<td>50</td>
<td>25</td>
</tr>
<tr>
<td>MD95-2002</td>
<td>40</td>
<td>140</td>
<td>15</td>
<td>75</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>PO21-1</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

---

fresh water algae noticed during the period of maximum fluxes at about 17 kyr appears to result from the conjunction of the very low sea level and an enhancement of the fluvial water inputs related to incipient melting of the ice sheets and possibly more humid climate.

Three periods of rafting from Canadian icebergs have to be noticed, they correspond respectively to Heinrich events 4, 2 and 1. For events 2 and 1, which we have studied at very high resolution, they represent a time period of about 300 yr. In both cases, a European precursor event has been identified, which happened about 1200 yr before the arrival of the Canadian icebergs. Both precursor events occurred during a cold phase of the Dansgaard–Oeschger cycles.
According to the oceanic circulation model established by Seidov et al. (1996) icebergs from Canadian origin could reach the European margin, only under the climatic and hydrologic conditions of the LGM (Fig. 1). During MWE like those that occurred in the Norwegian Sea between 16.9 and 15.7 kyr (starting just before the arrival of the Canadian icebergs at Meriadzek at 16.6 kyr) the current regime displayed by the model would not allow Canadian iceberg to reach the European margin. The southward flowing currents predicted by the model could however transport icebergs from the Fennoscandian ice sheet down to the Iberian margin. The short periods during which Canadian icebergs entered the Bay of Biscay correspond precisely with the period of coolest air temperatures over Greenland. We postulate that under these extreme conditions, either slower melting and/or an enhancement of the westerlies at the latitude of the Gulf have allowed the Canadian icebergs to reach the Celtic margin. One can also notice that the duration of the ice coverage increases considerably during both events which is evidence of the drastic cooling affecting western Europe.

6. Conclusions

Sea level evolution and climatic conditions have been the main control of the terrigenous fluxes at the Celtic margin during the last glacial cycle. The maximum terrigenous fluxes of the LGM have prevailed during the lowering stage, hence sea level per se is not the control, it is the conjunction of climate regime, soil vulnerability and river erosional power which have controlled the magnitude of the terrigenous fluxes. Paroxysmal flux ‘events’ have been recorded. The first one followed the arrival of Canadian icebergs of Heinrich event 2. The second one started simultaneously with a European IRD precursor of Heinrich event 1. The first event could be mainly related to erosion following a phase of soil defrost. The second one may result essentially from the reworking of the Celtic Sea delta deposits. Both of them include most probably a component of ice-rafted materials from European sources, the contribution of which cannot presently be evaluated. Turbidity current activity has been prevailing within the slope canyons until the beginning of MIS 1.

The arrivals of the Canadian icebergs of Heinrich events 2 and 1 are well recorded at the Celtic margin as short periods of about 300 yr during which the coolest air temperature occurred in Greenland (summit) and either stronger wind or reduced iceberg melting (or both) prevailed. Both events occurred 1.4 and 1 kyr, respectively, after a European ice rafting precursor.

Acknowledgements

This work has been funded by IFREMER, the European Union MAST programme ENAM2, and the French INSU programme VARIANTE. Core MD95-101-2002 has been retrieved by the R/V Marion-Dufresne within the context of the IMAGES programme. S.Z. has been financially supported by an EPSHOM grant during his thesis at the University of Bordeaux. We wish to thank Mr. Yvon Balut for his assistance at sea, Drs Jurgen Mienert and Laurent Labeyrie for their continuous support during the course of the ENAM programmes. Most of the $^{14}$C analyses have been performed at Gif/Yvette and we want to particularly acknowledge the collaboration of Martine Paterne, the staffs from the Radiocarbone LSCE and UMS-2004 Gif-Tandédron, as well as the engineers and technicians from the LSCE stable isotope laboratory. The comments and critics provided by Drs Franck Bassinot, Haflidi Haflidason, Laurence Vidal and Ian Hall have also helped to improve substantially the article.

Appendix 1

Depth correction for core MD95-2002

The correlation of the core MD95-2002 with its short twin core ESSCAMP KS02 (Loncaric et al., 1998) showed an elongation of the MD95-2002 upper 12 m. The elongation was also confirmed through the synthetic seismogram computed from
the physical properties of the core, which appear to be several milliseconds thicker than the 3.5-kHz record. Strong reflectors related to thick and dense IRD layers enabled such a correlation. The disturbance apparently does not affect the chemical, the lithological and the physical property records such as carbonate content, δ¹⁸O, strontium and neodymium isotopic ratios, ¹⁴C measurement, IRD content, foraminifera content and the gamma-density. Such disturbance has been noticed already in the first 10 m of several long cores from the cruise MD101 of R/V Marion-Dufresne.

This is a major disturbance when intending to calculate sedimentary fluxes, as sedimentation rates are one of the main inputs for this calculation. We then corrected the depth of the core MD95-2002, using the core ESSCAMP KS02 as a reference. We did not intend to make a very 'refined' correction but only to identify the disturbed length and to calculate a new depth scale using the simplest correction as possible. With the help of AnalySerie software (Paillard et al., 1996) we graphically identified 27 correlation pointers on the base of carbonate content, gamma-density and magnetic susceptibility variations (Tab 2.8/1). The pointers draw a broken line that visualises the disturbed upper part of core MD95-2002 with a slope of about 2, and the undisturbed lower part with a slope of about 1. A linear regression within each of these groups gives two equations with a slope of 2.27 in the disturbed part and 1.16 in the undisturbed part. The equations are (Fig. 13):

\[ D_{\text{MD95-2002\_disturbed}} = b_1 + a_1 D_{\text{ESSCAMP KS02}} \]  
\[ D_{\text{MD95-2002\_undisturbed}} = b_2 + a_2 D_{\text{ESSCAMP KS02}} \]

with \( b_1 = -2.0610 \); \( a_1 = 2.2709 \)

\[ D_{\text{MD95-2002\_disturbed}} = b_2 + a_2 D_{\text{ESSCAMP KS02}} \]

with \( b_2 = 4.9193 \); \( a_2 = 1.1608 \)

It is then possible with Eqs. A1 and A2 to calculate a 'pseudo' undisturbed depth for the disturbed part:

\[ D_{\text{MD95-2002\_undisturbed}} = B + AD_{\text{MD95-2002\_disturbed}} \]  

\( B \) is the excess length in the whole core, due to the disturbance.

\[ B = b_2 - \left( \frac{a_2 b_1}{a_1} \right) = 5.97 \text{ m} \]

\[ A = \frac{a_2}{a_1} = 0.5112 \]

Eq. A3 allows the calculation of the limit depth of the disturbance in core MD95-2002 (Eq. A4):

\[ D_{\text{limit\_of\_disturbance}} = \frac{B}{1-A} = \frac{(a_1 b_2 - a_2 b_1)}{a_1 - a_2} = 12.22 \text{ m} \]  

(A4)

It is then possible to correct the whole depth in core MD95-2002:

From the top of MD95-2002 to the limit depth of disturbance (12.22 m) included (Eq. A4):

\[ D_{\text{MD95-2002\_corrected\_2002\_disturbed}} = AD_{\text{MD95-2002\_disturbed}} \]

and from the limit depth of disturbance (excluded) to the bottom of the core MD95-2002 in subtracting the value ‘\( B \)’ found in Eq. A3:

\[ D_{\text{MD95-2002\_corrected\_2002\_undisturbed}} = D_{\text{MD95-2002\_undisturbed}} - B \]  

(A6)

Eq. A2, for undisturbed sediment, shows that sedimentation rates are 16% higher in core MD95-2002 than in core ESSCAMP KS02. This is realistic, the two cores being about 7 km apart. This is also in agreement with the 3.5-kHz profile that
shows slightly thicker series at the core MD95-2002 site. Eq. A1 shows that the thickness of the disturbed sediment is twice what it should be for undisturbed sediment.

References


McCabe, M., Clark, P.U., 1998. Ice sheet variability around
Morzadec-Kerfourn, M.-T., 1977. Les kystes de dino£agelle ¤s
Pujol, C., Turon, J.L., 1986. Comparaison des cycles clima-
Reynaud, J.Y., Tessier, B., Berne, S., Chamley, H., Debatist,
Reynaud, J.Y., Tessier, B., Berne, S., Chamley, H., Debatis,
Schneider, R., Bard, E., Mix, A.C., 2000. Last ice age global ocean and land surface temperatures: The EPILOG initia-
PAGES Newsl. 8, 19–21.
Scource, J.D., Austin, W.D.N., Buteman, R.M., Catt, J.A., Evans, C.D.V., Robinson, J.E., Young, J.R., 1990. Sedimento-
Shackleton, N.J., 1974. Attainment of isotopic equilibrium be-
tween ocean water and benthonic foraminifera genus Uviger-
stance 392, 373^377.
the North Atlantic Ocean during the last deglaciation. Na-
Manighetti, B., McCave, I.N., 1995. Depositional £uxes, pa-
leoproductivity and ice rafting in the NE Atlantic over the past 30 Kyr. Paleoceanography 10, 579–592.
Gow, A.J., Grootes, P.M., Meese, D.A., Ram, M., Taylor,
L., 1996. SIM- MAX, a modern analogue technique to deduce Atlantic sea
Hughen, K.A., Kromer, B., McCormac, G., van der Plicht,
tion, 24000–0 cal BP. Radiocarbon 40, 1041–1083.
gel ice Sheet: results from numerical ice-sheet modelling. Mar. Geol. 188, S0025-3227(02)00277-3.
European contribution of ice-rafted sand to Heinrich layers
tion, 24000–0 cal BP. Radiocarbon 40, 1041–1083.
Siegert, M.J., Dowdeswell, J.A., 2002. Late Weichselian ice-
berg, surface-melt and sediment production from the Eur-
asiatic ice Sheet: results from numerical ice-sheet modelling. Mar. Geol. 188, S0025-3227(02)00277-3.
European contribution of ice-rafted sand to Heinrich layers
tion, 24000–0 cal BP. Radiocarbon 40, 1041–1083.
Siegert, M.J., Dowdeswell, J.A., 2002. Late Weichselian ice-
berg, surface-melt and sediment production from the Eur-
asiatic ice Sheet: results from numerical ice-sheet modelling. Mar. Geol. 188, S0025-3227(02)00277-3.
European contribution of ice-rafted sand to Heinrich layers