Marine Geology
January 2008, Volume 247, Issues 1-2, Pages 84-103
http://dx.doi.org/10.1016/j.margeo.2007.08.006

Activity of the turbidite levees of the Celtic–Armorican margin (Bay of Biscay) during the last 30,000 years: Imprints of the last European deglaciation and Heinrich events

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Abstract:

High-resolution sedimentological and micropaleontological studies of several deep-sea cores retrieved from the levees of the Celtic and Armorican turbidite systems (Bay of Biscay — North Atlantic Ocean) allow the detection of the major oscillations of the British–Irish Ice Sheet (BIIS) and 'Fleuve Manche' palaeoriver discharges over the last 30,000 years, which were mainly triggered by climate changes.

Between 30 and 20 cal ka, the turbiditic activity on the Celtic–Armorican margin was weak, contrasting with previous stratigraphic models which predicted a substantial increase of sediment supply during low sea-level stands. This low turbidite deposit frequency was most likely the result of a weak activity of the 'Fleuve Manche' palaeoriver and/or of a reduced seaward transfer of sediments from the shelf to the margin. However, two episodes of turbiditic activity increase were detected in the Celtic–Armorican margin, during Heinrich events (HE) 3 and 2. This strengthening of the turbiditic activity was triggered by the meltwater releases from European ice sheets and glaciers favouring the seaward transfer of subglacial material, at least via 'Fleuve Manche' palaeoriver.

At around 20 cal ka, a significant increase of turbidite deposit frequency occurred as a response to the onset of the last deglaciation. The retreat of the European ice sheets and glaciers induced a substantial increase of the 'Fleuve Manche' palaeoriver discharges and seaward transfer of continentally-derived material into the Armorican turbidite system. The intensification of the turbiditic activity on the Celtic system was directly sustained by the widespread transport of subglacial sediments from the British–Irish Ice Sheet (BIIS) to the Celtic Sea via the Irish Sea Basin. A sudden reduction of turbiditic activity in the Armorican system, between ca. 19 and 18.3 cal ka, could have been triggered by the first well known abrupt sea-level rise ('meltwater pulse', at around 19 cal ka) favouring the trapping of sediment in the 'Fleuve Manche' palaeoriver valleys and the decrease of the seaward transfer of continentally-derived material.

The maximum of turbiditic activity strengthening in the Celtic–Armorican margin, between ca. 18.3 and 17 cal ka, was induced by the decay of European ice sheets and glaciers producing the most extreme episode of the 'Fleuve Manche' palaeoriver runoff and a great seaward transfer of subglacial material into the Bay of Biscay. Between ca. 17.5 and 16 cal ka, the turbiditic activity significantly decreased in both Celtic and Armorican turbidite systems in response to a global re-advance of glaciers and ice sheets in Europe. The last episode of ice sheet retreat, between ca. 16 and 14 cal ka, is well expressed in the Celtic system by a new increase of the turbiditic activity. The major episode of sea-level rise at around 14 cal ka (‘Meltwater Pulse 1A’), precluding the seaward transfer of sediments, induced the end of turbiditic activity in both the Celtic and the Armorican system.

Although two main phases of global sea-level rise seem to have had an effect on the Celtic–Armorican margin, this work proposes the BIIS retreat and associated riverine discharges as the main trigger mechanisms of the turbiditic activity in this region during the last 30,000 years.

Keywords: Bay of Biscay; British–Irish Ice Sheet; ‘Fleuve Manche’; palaeoriver; last deglaciation; LGM; Heinrich events; turbidites

1. INTRODUCTION

It is widely acknowledged that climate change and resulting sea level oscillations affect in some way the sedimentary processes operating along continental margins and in particular fine grained turbidite systems (Stow et al., 1985). This is the case of non-glaciated margins located at mid- to low latitudes of the eastern North Atlantic (south of 26°N), far way from glaciers (e.g. Weaver et al., 2000). Inversely, the eastern North Atlantic margin (north of 56° N) and adjacent submarine fans have been particularly affected by ice-sheet oscillations during the last part of the full-glacial period (e.g. Dowdeswell et al., 2002; Elverhoi et al., 1998). The effectiveness of ice sheets for sustained glaciated margins is recorded in the Bear Island Fan (western Barents Sea – 75°N). The Bear Island Fan has a similar area and volume to the low-latitude fluvially-derived Amazon and Mississippi turbidite systems but a smallest drainage basin (Dowdeswell et al., 2002) suggesting that the adjacent glaciers have a great ability to erode their substrate. Recent surging glaciers (e.g. Gilbert et al., 2002) also show the close connection between sediment supply and ice-sheet oscillations in the high-latitude continental margins.

The Celtic and Armorican turbidite systems (Bay of Biscay – 46°N) are located at the transition zone between the eastern North Atlantic glaciated and non-glaciated margins. Weaver and Benetti (2006) have suggested that deep-sea sedimentation in this region is mostly like influenced by sea level changes. However, previous studies on continuous hemipelagic sequences suggest that the Celtic - Armorican margin was affected by the British-Irish Ice Sheet (BIIS) oscillations and in particular during an extreme episode of meltwater discharge via the “Fleuve Manche” palaeoriver at around 18 cal ka (Eynaud et al., 2007; Mojtahid et al., 2005; Zaragosi et al., 2001b). A recent multi-proxy study on three turbidite levees from northern Celtic - Armorican margin also suggests that the BIIS oscillations have had an impact on the deep-sea clastic sedimentation during the last deglaciation and can provide important information about palaeoenvironmental changes at a high resolution time-scale (Zaragosi et al., 2006).

However, none of these studies have showed how sedimentary processes operating along this continental margin have been affected by the successive BIIS oscillations occurring between the final stages of the last glacial and the last glacial-interglacial transition (LGIT). The aim of this study is therefore to investigate the relationship between gravity processes in the Celtic and Armorican turbidite systems and the BIIS oscillations for the last 30,000 years. Towards this aim we have performed a high-resolution sedimentological and micropaleontological study from five long piston cores (MD04-2836, MD04-2837, MD03-2690, MD03-2688 and MD03-2695) retrieved in turbidite levees of the Celtic – Armorican margin. In particular, we have estimated the frequency of turbidite deposits which allow quantification of the continental sediment supply removing the problems inherent to local sedimentation rate and/or of coring deformations (Skinner and McCave, 2003).

2. GEOLOGICAL AND ENVIRONMENTAL SETTINGS

The Celtic – Armorican margin is a passive margin composed of two medium-sized deep-sea clastic systems: the Celtic and the Armorican turbidite systems (Droz et al., 1999; Le Suavé, 2000; Zaragosi et al., 2001a; Zaragosi et al., 2000). The Celtic and Armorican turbidite systems are located in the northern and central part of the Bay of Biscay abyssal plain respectively (Figure 1), and have been active since the Early Miocene (Droz et al., 1999; Mansor, 2004). Each system covers about 30,000 km² in water depths ranging from 4100 m to 4900 m. The turbidite systems are sustained by more than thirty deep canyons capturing continentally-derived sediments. These canyons converge down to five submarine drainage basins (Bourillet et al., 2003) (Figure 1):

- The 'Grande Sole' extends from the Goban to the Brenot spurs. The Whittard channel-levee system (Figure 2) is located basinwards of this catchment area;
- The 'Petite Sole' extends from the Brenot to the Berthois spurs and nourishes the Shamrock channel-levee system (Figure 2);
- The 'La Chapelle' is located between the Berthois and the Delesse spurs and connects the Blackmud and Guilcher channel-levee systems;
- The 'Ouest Bretagne' is located between the Delesse and the Bourcart spurs linking downstream with the Crozon channel-levee system;
- The 'Sud Bretagne', located between the Bourcart and the Folin spurs, is linked downstream to the Audierne channel-levee system (Figure 2).

During the last glacial period, the Celtic and Armorican turbidite systems seems to have been particularly influenced by the British-Irish Ice Sheet (BIIS) oscillations and 'Fleuve Manche' palaeoriver discharges (e.g. Bourillet et al., 2003). It is widely known that the 'Fleuve Manche' palaeoriver activity started during the last glacial period, favoured by the lowering of the sea-level stand and by episodes of the BIIS meltwater discharge (Mojtahid et al., 2005; Zaragosi et al., 2001b) which covered Great Britain and Ireland during the last glacial period (e.g. Bowen et al., 2002). The ‘Fleuve Manche’ palaeoriver had a large catchment area, including the continental palaeodrainage system of major West European palaeorivers such as the Rhine, Meuse, Seine, Somme, Thames and Solent (Bourillet et al., 2003; Lericolais, 1997) (Figure 1).

Besides the ‘Fleuve Manche’ palaeoriver, the Irish Sea Basin seems to have played an important role in sediment supply from the continent to the Celtic - Armorican margin. It is known that the Irish Sea ice stream protruded in the southern Irish Sea and Celtic Sea although its extent is still a matter of debate. Recent simulations (Boulton and Hagdorn, 2006) and geological field studies (Evans and O'Cofaigh, 2003; Hiemstra et al., 2006; O'Cofaigh and Evans, 2007) seem to confirm that the southern limit of this ice stream reached the Isles of Scilly, as previously suggested by Scourse et al. (1991; 1990) (Figure 1).

Many studies from marine deep-sea cores have showed that the BIIS was very sensitive to abrupt climatic changes (e.g. Knutz et al., 2007; Peck et al., 2006) and in particular during the last deglaciation (Eynaud et al., 2007; Mojtahid et al., 2005; Zaragosi et al., 2006; Zaragosi et al., 2001b). This hypothesis has been confirmed by several continental studies which reveal a progressive but complex decay of the BIIS during the last deglaciation (McCabe and Clark, 1998; McCabe et al., 2007b).

3. MATERIALS AND METHODS

Five long piston cores (Table 1) were retrieved by using the 'Calypso' corer in the Celtic – Armorican margin during the MD133-SEDICAR (Bourillet and Turon, 2003) and the MD141-ALIENOR (Turon and Bourillet, 2004) oceanographic cruises on board the R/V Marion Dufresne (IPEV). Cores MD03-2688, MD03-2690 and MD03-2695 were recovered in the Crozon, Guilcher and Audierne turbidite levees (Armorican turbidite system) while cores MD04-2836 and MD04-2837 were collected in the Whittard and Shamrock turbiditic levees (Celtic turbidite system) (Figure 1 and 2). Previous studies on this region have shown that some of these turbidite levees are mainly composed of a complex sedimentological succession of turbiditic sequences alternating with ice-rafted laminae and hemipelagic layers (Zaragosi et al., 2006).

3.1. Chronostratigraphy

The age models for cores MD03-2688, MD03-2690, MD03-2695 and MD04-2836 have been determined based on foraminiferal stratigraphy, AMS dating and by using additional control points from the reference core MD95-2002 (Table 2). The age model of core MD95-2002 was based on 20 ^{14}C AMS ages spanning the last 30 ka (Table 2) (Auffret et al., 2002; Grousset et al., 2000; Zaragosi et al., 2006; Zaragosi et al., 2001b).

Cores were sub-sampled with a sample spacing of 5 to 20 cm for micropaleontological analysis along the hemipalegic layers. These hemipelagic layers are not contaminated by reworked material and represent intervals of continuous sedimentation. The subsamples were then dried, weighed and washed through a 150 µm mesh sieve. At least 300 polar foraminifera *Neogloboquadrina pachyderma* (s) were counted jointly with a number of other planktonic species in order to determine the relative abundances (%) of this polar species. Previous studies on this region have shown the suitable use of *N. pachyderma* (s) to reconstruct drastic sea-surface changes which are stratigraphically contemporaneous with major climatic events (Mojtahid et al., 2005; Peck et al., 2007; Zaragosi et al., 2001b).

Thirty four accelerator mass spectrometer (AMS) ^{14}C dates were obtained from cores MD03-2688, MD03-2690, MD03-2695 and MD04-2836 (Table 2).

3.2. Sedimentological analyses

The sedimentological analyses of the Celtic – Armorican deep-sea cores consists firstly of visual description and X-ray analysis obtained with a SCOPIX image processing tool (Migeon et al., 1999). Additionally, grain-size analysis were performed using a Malvern™ Supersizer 'S'. Finally, microscopical observations of about ten thin-sections (10 cm long) of impregnated sediments selected from well-preserved and representative sedimentary facies were performed using a fully automated Leica™ DM6000B Digital Microscope. The last method has been recently detailed in Zaragosi et al. (2006).

In order to understand the activity of the Celtic and Armorican turbidite systems, we have detected and quantified the number of turbiditic deposits in the Whittard, Guilcher, Crozon and Audierne turbidite levees. For this, we firstly observed several thin sections of impregnated sediments representing distinctive alternated facies of ice-rafted, turbiditic deposits and hemipelagic layers (Figure 3). Secondly, we have determined the criteria to distinguish each facies *via* microscope and X-ray imagery. Finally, we applied these criteria to distinguish each facies in all cores using X-ray imagery. Indeed, microscopic observation of IRD laminae reveals heterogeneous and scattered angular lithic grains within fine-bioturbated clay while fine (mm-thick) and slightly dark layers are observed in the X-ray imagery (Figure 3). Turbiditic deposits are generally thicker (mm-thick to cm-thick) than IRD laminae and present usually sharply eroded basal contacts. The progressive transition from very dense (dark) contacts to a slightly lighter (grey) top of sequences, visible on X-ray imagery, is associated with the typical fining-up trend of turbiditic deposits (Bouma, 1962; Stow and Piper, 1984) (Figure 3).

Each turbiditic deposit of cores MD04-2836, MD03-2690, MD03-2688 and MD03-2695 has been counted using X-ray imagery. Turbidites have not been counted in core MD04-2837 because this record presents important disturbances linked to coring stretching. Following this, we have quantified the turbidite deposit frequency on the Whittard, Guilcher, Crozon and Audierne levees per 1000 years. We assumed that this quantification represents the minimum value of turbidite frequency because of possible erosive losses and/or non-deposit events (i.e. by-pass).

4. RESULTS

4.1. Chronological framework

It is usually difficult to reconstruct an accurate stratigraphy in turbidite levees because these environments are mainly composed of reworked sedimentary material. Therefore, we have used the abundance peaks of *N. pachyderma* (s) determined in well preserved hemipelagic material as primary tool to establish the age model of cores MD04-2836, MD04-2837, MD03-2688, MD03-2690 and MD03-2695. This method allows the detection of several paleoclimatic events during the end of the Marine Isotopic Stages (MIS) 3, MIS 2 and MIS 1: Heinrich events (HE 3, HE 2 and HE 1), Last Glacial Maximum (LGM), Greenland Interstadial 1 (GIS1) / Bölling-Alleröd (B-A), Younger Dryas (YD) and the Holocene (Figure 4).

The maximum expansion of the polar foraminifera *N. pachyderma* (s) in the Celtic–Armorican margin between ca. 18.3 and 16 cal ka (Figure 4), suggesting extremely cold sea surface waters, is contemporaneous with the presence of ice rafted detritus (IRD) in the reference core MD95-2002 (Zaragosi et al., 2001b). Although IRD are detected between ca. 18.3 and 16 cal ka, their maximum expression occurred within the interval 17-16 cal ka. The age limits of this cold episode are synchronous with those proposed by Elliot et al. (2001) for Heinrich (HE) 1 event elsewhere in the North Atlantic region. The other episodes of *N. pachyderma* (s) maximum expansion (~90-100%) occurring at ca. 23.5-26 cal ka and at ca. 30-32 cal ka are also synchronous with the age limits of HE 2 and HE 3, respectively (Elliot et al., 2001). We assume therefore that these cooling events detected in the Celtic-Armorican margin are most likely the result of the impact of Heinrich events. Previous works on the eastern North Atlantic (e.g. Bond et al., 1992; Eynaud et al., 2007) have shown a sea surface cooling episode preceding the maximal arrival of IRD. Other records from the mid-latitudes of the North Atlantic region have shown the same complex pattern in both marine and terrestrial environments, which have been associated to the well known Heinrich events (e.g. Bard et al., 2000; Chapman et al., 2000; Naughton et al., 2007; Naughton et al., submitted).

The Younger Dryas cold period is also defined by the increase of the *N. pachyderma* (s). However, an intriguing sedimentary hiatus is observed at around this period in cores MD03-2688 and MD03-2695 (Figure 4). The planktonic foraminiferal assemblages show that the Early Holocene and Bølling-Allerød periods are well recorded in both cores (Duprat, *comm. pers.*).

Furthermore, radiocarbon results have confirmed that core-to-core correlations, based on abrupt increases in abundances of *N. pachyderma* (s), represent the temporal limits of the cold episodes that punctuated the final part of the last glacial period (Table 2 and Figure 4).

The age model of core MD03-2688 indicates that this core covers the last ca. 34 cal ka (~29.1 ^{14}C ka); core MD03-2695 extends back to ca. 33.5 cal ka (~28.7 ^{14}C ka) and core MD03-2690 to ca. 26 cal ka (~22 ^{14}C ka); and core MD04-2836 spans ca. 20.4 cal ka (~17.3 ^{14}C ka) (Figure 4 and Table 2).

4.2. Evolution of sedimentary conditions

The detailed sedimentological analysis (visual description, X-ray imagery, grain-size measurements and thin section analysis) of the studied cores has allowed the identification of six lithofacies (Figure 5, 7 and 8). These lithofacies represent the evolution of the sedimentary conditions on the Whittard, Blackmud, Guilcher, Crozon and Audierne levees during the last 30,000 years (Figure 5):

Lithofacies 1, between 0 and 8 cal ka (Mid- and Late-Holocene), is constituted by homogeneous, structureless marly ooze containing a temperate foraminiferal assemblage

(*Globigerinoides ruber, Globigerina bulloides, Globorotalia hirsuta, Globorotalia truncatulinoides, Orbulina universa*) (Figure 5). This lithofacies forming the modern deep-sea Bay of Biscay seafloor has been interpreted on the turbidite levees as a pelagic to hemipelagic drape deposits without significant turbidite supplies from the continental shelf (Zaragosi et al., 2006).

Lithofacies 2, between ca. 8 and 15.5 cal ka, consists of homogeneous structureless clay interbedded with some centimetre-scale silt to very fine sand layers (Figure 5). Sedimentation rates range from 270 to 370 $cm.ka^{-1}$ and reach 770 $cm.ka^{-1}$ in core MD04-2836 (Figure 6). Beds display a sharply erosive basal contact and are normally graded, with a basal grain-size median ranging from 20 to 80 µm in core MD04-2836, 40 to 80 µm in core MD03-2690 and 60 to 140 µm in core MD03-2688. According to Stow and Piper (1984), these beds represent silt-mud turbidites deposited from the overflow of turbidity currents while homogeneous clay are interpreted as hemipelagic deposits.

Lithofacies 3, between ca. 15.5 and 17 cal ka, ca. 23.5 and 26 cal ka and ca. 30.5 and 32 cal ka., shows a monospecifism of the polar foraminifera *N. pachyderma* (s) and contains frequent thinning- and fining-upward sequences of very fine sand and silt deposits with erosive basal contacts (Figure 5). These sequences are interpreted as fine-grained turbidites. Turbidite layers are thin (1 to 10 cm) and their basal grain-size ranges from 40 to160 µm in core MD04-2836, 50 to 140 µm in core MD03-2690, 30 to110 µm in core MD03-2688 and 15 to 100 µm in core MD032695. Numerous IRD-rich millimetre-scale clay layers are also interbedded with the turbidite sequences. Sedimentation rates range from 110 to 500 $cm.ka^{-1}$ in cores MD04-2836 and MD03-2690 respectively (Figure 6). Lithozone 3 reveals periods of important turbidite deposits associated with numerous ice-rafting events on the sedimentary levees of the Celtic – Armorican margin.

Lithofacies 4, between ca. 17 and 18.3 cal ka, is an ultra-laminated sediment composed of IRD-rich millimetre-scale clay layers and fine fining-upward silty laminae with sharp basal contacts (Figure 5 and 7). Some silty to very fine sandy deposits are also observed and show thin cross-rippled laminations. These laminations are interpreted to be of turbiditic origin. Their basal grain-size ranges from 40 to 140 µm in cores MD04-2836 and MD03-2690, 20 to 120 µm in core MD03-2688 and 20 to 180 µm in core MD03-2695. Load casts and flame structures are commonly present at the lower contacts of the turbidites. A monospecifism of *N. pachyderma* (s) is also described in lithofacies 4. Sedimentation rates are extremely high (>600 $cm.ka^{-1}$) and reach up to 950 $cm.ka^{-1}$ in core MD03-2695 (Figure 6). Lithofacies 4 reveals a high sediment supply period produced by very frequent turbidity currents in the channel-levee systems and numerous ice-rafting events.

Lithofacies 5, between ca. 18.3 cal ka and ca. 20 cal ka, is characterized by homogeneous structureless clay interbedded with some fining-upward millimetre- to centimetre-scale silt to sand deposits with erosive basal contacts (Figure 5). These sequences are interpreted as fine-grained turbidites. Some parts of the top of lithozone 5 are laminated and represent a transition zone between the ultra-laminated lithofacies 4 and the base of the lithofacies 5 which is mostly composed of scattered centimetre-scale turbidites. The grain-size of the base of turbidite beds ranges from 40 to 160 µm in core MD04-2836, 50 to 190 µm in core MD03-2690, 60 to 240 µm in core MD03-2688 and 50 to 230 µm in core MD03-2695. Mean sedimentation rates range from 290 to 375 $cm.ka^{-1}$ except in core MD04-2836 where it reaches 545 $cm.ka^{-1}$ (Figure 6). Lithofacies 5 shows hemipelagic sedimentation interbedded with some turbidites deposits that are more massive and more spaced in its basal part, thus defining a transition sedimentary facies between lithofacies 4 and lithofacies 6.

Lithofacies 6 was deposited between ca. 20 and 34 cal ka, except during ca. 23.5 – 26 cal ka and ca. 30.5 – 32 cal ka periods which corresponds to lithofacies 3. Lithozone 6 is dominated by massive, fining-upward silt to sand deposits, interpreted as turbidites (Figure 5 and 8). Grain-size appears to be similar to that characterising lithofacies 5. However, turbidites of lithofacies 6 are thicker (centimetre to decimetre-scale) than turbidites of lithofacies 5. Sedimentation rates are moderate to low with values ranging from 15 to 100 $cm.ka^{-1}$ (Figure 6). Lithofacies 6 reveals a period of rare but massive turbidity current activity.

4.3. Turbidite deposit frequency

The frequency of the turbiditic deposits ($turb.ka^{-1}$) has been estimated from cores MD04-2836, MD03-2690, MD03-2688 and MD03-2695 for the last 30 ka (Figure 6, 9 10 and 11). Three main periods of turbiditic activity are observed:

a) From ca. 33 to 20 cal ka, there is a general low turbiditic activity in the Guilcher, Crozon and Audierne channel levee systems. The turbidite deposit frequency ranges from 0 to 40 turbidites per thousand years ($turb.ka^{-1}$). A moderate frequency of turbiditic deposits occurred during HE 3 and HE 2 (30 to 40 $turb.ka^{-1}$) while low turbiditic activity (max. 15 $turb.ka^{-1}$) is associated with the end of MIS 3 and the early- and mid-LGM (Figure 6 and 9).

b) Between ca. 20 to 17 cal ka, there is a general huge increase in the frequency of the turbidite deposits (75 $turb.ka^{-1}$ in core MD03-2688 and 230 $turb.ka^{-1}$ in core MD04-2836) (Figure 6, 9 and 10). A higher resolution study of the frequency of the turbidite deposits (number of turbidites per 250 years), in core MD03-2688, shows a sudden episode of turbidite deposit frequency decrease between ca. 19 and 18.3 cal ka (Figure 10). The turbiditic activity reached a maximum intensity between ca. 18.3 to 17 cal ka in all cores independently of the time resolution used to calculate those frequencies (Figure 6, 9 and 10).

c) From ca. 17 to 16 cal ka, there is a sharp decrease of the turbiditic activity in all cores. The turbidite deposit frequency reached 60 to 120 $turb.ka^{-1}$ on the Guilcher, Crozon and

Audierne levees and only 25 turb.ka^{-1} on the Whittard levee between 17 and 16 cal ka (Figure 6, 9, 10 and 11).

d) From ca. 16 to 0 cal ka, although there is a gradual decrease of the turbiditic activity in most areas (Figure 6, 9, 10 and 11), Whittard records shows a significant re-activation of gravity processes at the beginning of this interval. Indeed, turbidite deposit frequency of core MD04-2836 reaches 130 turb.ka^{-1} between ca. 16 and 14 cal ka while attaining only 25 turb.ka^{-1} between ca. 17 – 16 cal ka and 8 turb.ka^{-1} between ca. 14 – 13 cal ka (Figure 11).

5. DISCUSSION

Implications of the BIIS and 'Fleuve Manche' palaeoriver activities in the Celtic – Armorican margin during the last 30 ka

The high-resolution sedimentological and micropaleontological study of several marine deep-sea cores retrieved on the turbidite levees of the Whittard, Blackmud, Guilcher, Crozon and Audierne channel-levee systems allows the detection of the major BIIS oscillations and 'Fleuve Manche' palaeoriver discharges during the last 30 ka. The turbidite deposit frequency estimated in MD04-2836, MD03-2690, MD03-2688 and MD03-2695 deep-sea cores reflects important oscillations of sediments supply into the Celtic and Armorican turbidite systems between 30 ka and 14 ka BP (Figure 6 and 9).

The last glacial period is marked by the long-term increase of the global ice volume, contemporaneous with the global sea-level fall (Chappell, 2002; Lambeck et al., 2002). The last sea-level lowstand, contemporaneous with the final stages of the global ice expansion occurred between ca. 30 and ca. 20 cal. ka (Lambeck et al., 2002). However, during this interval, several millennial-scale climate oscillations have been observed in both Greenland and North Atlantic records (Bond et al., 1993; Dansgaard et al., 1993) producing substantial sea-level changes (Siddall et al., 2003). In this work, we define the LGM as a period of relatively stable climate that occurred between HE 2 and HE 1 following the EPILOG (Mix et al., 2001) and MARGO (Kucera et al., 2005) suggestions.

It is commonly accepted that sea-level lowstand conditions favoured the seaward sediment transfer from the continent to the deep-sea turbidite systems (e.g. Posamentier and Vail, 1988). Therefore, we should expect to detect a maximum of the turbidite frequency in the Celtic and Armorican turbidite systems synchronous with the last lowest sea-level stand. Our data shows, on the contrary, that there is a general weak sediment supply to the Celtic and

Armorican turbidite systems between ca. 30 and ca. 20 cal. ka (Figure 6 and 9). The low turbidite deposit frequency within ca. 30 and 20 cal. ka in the Celtic - Armorican margin can probably be due to weak runoff rates of the ‘Fleuve Manche’ palaeoriver and/or to low seaward sediment transfer which was probably blocked on the shelf as a response to the deposition of sand banks in the Celtic sea (Reynaud et al., 1999). Nonetheless, the presence of some turbidite sequences in this region is most likely the result of sediment seaward transfer from the delta located in the ‘Fleuve Manche’ palaeoriver mouth (Lericolais, 1997; Zaragosi et al., 2001a). Two episodes of turbidite frequency increase contemporaneous with HE 3 and HE 2 punctuated the general low turbidite activity period (Figure 6, 9 and 10). This increase of turbiditic activity during HE 3 and HE 2 was likely the result of an increase of seaward transfer of subglacial sediment as a response to meltwater releases from surrounded ice sheets and glaciers, confirming what has been previously proposed by climate simulations (e.g. Clarke et al., 1999; Forsström and Greve, 2004) as well as by sedimentological studies on the eastern Canadian margin (e.g. Hesse et al., 2004; Rashid et al., 2003).

A significant increase of sediment supply shown by the increase of the turbidite deposit frequency is observed since ca. 20 cal ka in cores MD04-2836 (i.e. in the Celtic turbidite system), MD03-2690, MD03-2688 and MD03-2695 (i.e. in the Armorican turbidite system) (Figure 6 and 9). The quantity of sediment supply into the Celtic turbidite system is higher than that of the Armorican turbidite system.

The ‘Grande Sole’ drainage basin was connected to the Celtic Sea (Figure 1) funnelling substantial volume of sediment directly released by the BIIS and the Irish Sea ice stream (Bowen et al., 1986; Eyles and McCabe, 1989) into the Celtic turbidite system via the Irish Sea Basin. The ‘Cooley Point Interstadial’, starting at ca. 20 cal ka (~16.7 ^{14}C ka BP, (McCabe and Clark, 1998)), characterises the beginning of the BIIS deglaciation and induced the widespread transport of subglacial sediments to the south-east Irish ice-margin as previously suggested by several continental records (Bowen et al., 2002; McCabe et al., 2005). The high turbiditic activity in the Whittard channel synchronous with the major episode of the Irish Sea ice stream retreat (Figure 11) suggests that the Irish Sea Basin was probably affected by fully marine conditions, favouring the direct seaward transfer of sediments from the BIIS. These fully marine conditions were attained because the isostatic depression of the Irish Sea Basin vastly exceeded the eustatic lowering as suggested by Clark et al. (2004) and McCabe et al. (2007a; 2007b).

Although the turbidite deposit frequency in the Armorican turbidite system is lower than that recorded in the Celtic system, an increase of the turbiditic activity has been also detected in cores MD03-2690, MD03-2688 and MD03-2695 at around 20 cal ka (Figure 6, 9 and 10). This increase of the turbidite deposit frequency in the Armorican turbidite system suggests the strengthening of the 'Fleuve Manche' palaeoriver discharges at around 20 cal ka. Seismic records from this region show the presence of Neogene fluvial palaeovalleys in the present-day shelf (Figure 1), between the 'Fleuve Manche' palaeoriver and canyons of the Armorican margin (Bourillet et al., 2003). This suggests that these sub-environments were directly connected in the past, favouring the great seaward transfer of sediments from the 'Fleuve Manche' palaeoriver via the numerous canyons which composed the 'La Chapelle', 'Ouest Bretagne' and 'Sud Bretagne' drainage basins (Figure 1 and 2). Furthermore, previous studies on this region detected an increase of *Pediastrum sp*. concentration (freshwater alga) (Zaragosi et al., 2001b) and of BIT-index (Branched and Isoprenoid Tetraether) (Ménot et al., 2006) reflecting the introduction of high quantities of fluvial terrestrial organic material in the Armorican margin contemporaneous with the increase of turbiditic activity in this area (Figure 10). Additionally, recent simulations have shown that tides and tidal currents of the Celtic - Armorican shelf also contributed to the seaward transfer of continental material between 20 and 10 ka (Uehara et al., 2006).

The strengthening of the ‘Fleuve Manche’ palaeoriver discharges was probably induced by the retreat of the BIIS, the European glaciers and the south-western part of the Fennoscandian ice sheet. The well known episodes of glacier decay in Scandinavia (Rinterknecht et al., 2006), Poland (Marks, 2002) and in the European Alps (Hinderer, 2001; Ivy-Ochs et al., 2004) at around ca. 20 cal ka corroborate our hypothesis.

The retreat and melting of the European ice sheets and glaciers at ca. 20 cal ka contributed to an abrupt sea-level rise, known as a ‘meltwater pulse’ at around 19 cal ka (19-ka MWP) (Clark et al., 2004; Yokoyama et al., 2000). This abrupt episode lasted 500 years and sea level rise amounted to over 15 m (Yokoyama et al., 2000) favouring the trapping of sediments in the ‘Fleuve Manche’ palaeoriver valleys. Synchronously, the decrease of both BIT-index and *Pediastrum sp*. concentration in the neighbouring core (MD95-2002) (Ménot et al., 2006; Zaragosi et al., 2001b) (Figure 10) suggests a decrease of continentally-derived material to the Armorican margin, also supporting the idea of reduced ‘Fleuve Manche’ discharges in the northern part of the Bay of Biscay.

Following this, the observed abrupt increase of sediment supply in the Celtic and Armorican turbidite systems at ca. 18.3 cal ka (Figure 6, 9, 10 and 11) was most likely the result of a seaward sediment transfer increase from the south-east Irish ice-margin and an intensification of the ‘Fleuve Manche’ palaeoriver runoff, respectively. In the Armorican turbidite system, the highest turbidite deposit frequency is synchronous with the maximal arrival of continental material as demonstrated by BIT-index (Ménot et al., 2006) and *Pediastrum sp*. concentration (Zaragosi et al., 2001b) (Figure 10). This suggests that the ‘Fleuve Manche’ discharges increased drastically at around 18.3 cal ka confirming what has been previously proposed by Zaragosi et al. (2001b) and Ménot et al. (2006). This episode of high riverine discharges, occurring at ca. 18.3 cal ka, was clearly more intense than that characterizing the beginning of the deglaciation (ca. 20 cal ka) (Figure 6, 9 and 10) and was also likely the result of the European glacier retreat. Several studies have shown that important environmental changes leading to a substantial retreat of the BIIS occurred in the north and north-western UK margin (Knutz et al., 2002a; Knutz et al., 2002b; Wilson et al., 2002), contemporaneous with the maximum decay of the Fennoscandian ice sheet at ca. 18.3 cal ka (Dahlgren and Vorren, 2003; Lekens et al., 2005; Nygard et al., 2004). The deposition of one ultra-laminated facies in the Celtic – Armorican margin between ca. 18.3 cal ka and 17 cal ka (lithofacies 4 – Figure 5) reveals that a significant environmental change has had an impact in northern and southern part of the glacially-influenced European margin. This facies has been recognized as marine ‘varves’ resulting from episodic cycles between meltwater discharges and iceberg calving (Zaragosi et al., 2006).

Between ca. 17.5 and 16 cal ka, there was a decrease of the turbiditic activity in the Celtic turbidite system (Figure 6, 9 and 11) which was contemporaneous with the two main general re-advance phases of the BIIS: the ‘Clogher Head’ and ‘Killard Point’ stadials (after 15 and until ~13 ^{14}C ka BP), detected in the northern Irish Sea Basin by McCabe et al. (2007b). This suggests that sub-glacial material transfer from the BIIS to the ‘Grande Sole’ drainage basin was most likely reduced during re-advance episodes of the BIIS. Theses episodes were synchronous with the re-advance of the Fennoscandian ice-sheet and central European glaciers (Alps, Jura) (Buoncristiani and Campy, 2004; Everest et al., 2006; Ivy-Ochs et al., 2006; Knies et al., 2007). The rapid decrease of the turbidite deposit frequency in the Armorican turbidite system (Figure 6, 9 and 10) reveals a substantial decrease of the ‘Fleuve Manche’ palaeoriver runoff, in response to the episodic ‘pause’ of the last deglaciation.

The resumption of the last deglaciation is particularly well expressed in the Celtic turbidite system record during the well known Bölling-Alleröd episode. Indeed, the last major decay of the BIIS, associated with the ‘Stagnation Zone Retreat’ and the ‘Rough Island Interstadial’ episodes detected in the northern Irish Sea Basin (McCabe et al., 2005; McCabe et al., 2007b), induced a relatively huge increase of the turbiditic activity in the Celtic turbidite system, between ca. 16 and 14 cal ka (Figure 11). Besides the last stages of the BIIS decay, the Celtic - Armorican margin was also affected by a rapid and sustained rise of the global sea-level from 16 to 12.5 cal ka (Lambeck et al., 2002). Indeed, the cessation of turbiditic activity on the Celtic - Armorican margin occurred during or after the episode of maximum sea-level rise (Figure 10), known as the ‘Meltwater Pulse 1A’ (MWP 1A - ca. 14 cal ka) (Fairbanks, 1989) contributing to the disappearance of the ‘Fleuve Manche’ palaeoriver. Moreover, the increase of dry conditions in Europe during the Younger Dryas at around 13 cal ka (e.g. Watts, 1980) also decreased the seaward transfer of fluvially-derived sediment onto the Celtic – Armorican margin.

The comparison of the turbiditic activity between the Celtic turbidite system and the Laurentian Fan, which extends at the outlet of the Laurentide Ice Sheet (LIS) (Skene and Piper, 2003), reveals a similar sedimentary pattern over the last deglaciation. Two main phases of turbidite deposition occurred at the end of the LGM and after ca.16 cal ka, bracketing a huge reduction of sediment supply at around 16.5 cal ka. Despite a short time lag between the BIIS and the LIS oscillations over the last deglaciation (McCabe and Clark, 1998), the similarity of both turbiditic records from the Laurentian Fan and the Celtic system suggests that seaward transfer of glacially-derived material to the deep-sea North Atlantic have been clearly forced by the combined effect of global climate changes and amphi-North Atlantic ice sheets oscillations for at least the last 20,000 years.

6. CONCLUSIONS

The high-resolution sedimentological (including turbidite frequency analysis) and micropaleontological studies performed in the Celtic - Armorican margin document the evolution of the turbidite systems in this region over the last 30,000 years. Changes in the frequency of turbidite deposits in the Celtic - Armorican margin were mainly triggered by the British – Irish Ice Sheet (BIIS) and European glaciers oscillations (advance and retreat episodes). The retreat of the BIIS and European glaciers favoured the transfer of continentally-derived material *via* the Irish Sea ice stream and the 'Fleuve Manche' palaeoriver into the Celtic and the Armorican systems respectively. Inversely, the BIIS and European glaciers advances preclude the introduction of large amounts of meltwater into the 'Fleuve Manche' palaeoriver, reducing drastically the seaward transfer of sediments in the Bay of Biscay. This evidence, contrasting with stratigraphic models which predict that turbidite systems are mainly controlled by sea level changes, confirms that glacially-influenced turbidite systems are largely controlled by ice sheets and glaciers oscillations. However, the synchronicity between turbidite deposit frequency reduction and the abrupt meltwater pulse episode (19-ka MWP) suggests that this drastic sea-level rise would have favoured the trapping of sediments in the 'Fleuve Manche' palaeoriver. Similarly, after the last stage of the BIIS decay a second sudden episode of sea-level rise (MWP 1A) contributed to the end of the 'Fleuve Manche' palaeoriver discharges and consequent turbiditic activity in the Celtic - Armorican margin.

ACKNOWLEDGEMENTS

The authors warmly thank G. Chabaud, G. Floch, R. Kerbrat, B. Martin, M. Rovere, J. Saint Paul and O. Ther for their technical support and J. Duprat and A. Van Toer for their useful assistance to the biostratigraphic approach. We thank also G. Ménot for data of the BIT-index; P. De Deckker, M. Gaudin and M.F. Sánchez Goñi for valuable comments and language improvement; and the crew and scientific teams of MD133/SEDICAR and MD141/ALIENOR cruises on the ‘R/V Marion Dufresne’ (IPEV) for the recovery of the long-piston cores. We acknowledge financial support by the French Programme ‘GDR MARGES’ and ‘RELIEFS DE LA TERRE’, the ‘ARTEMIS’ ^{14}C AMS French Project and the ANR ‘IDEGLACE’. We finally acknowledge A.M. McCabe, J.D. Scourse and Editor D.J.W. Piper for their helpful comments which greatly improved this paper. This is an UMR 5805 'EPOC' (Bordeaux 1 University - CNRS) contribution n°1637.

FIGURE CAPTION

Figure 1: Physiography of the Celtic - Armorican margin (north-western Europe) during the Last Glacial Maximum (LGM). (1) extent of the British – Irish Ice Sheet (BIIS) (Bowen et al., 2002); (2) southern extent of the Irish Sea ice stream proposed by Scourse and Furze (2001); (3) extent of the European Alps glacier; (4) 'Fleuve Manche' palaeoriver (Bourillet et al., 2003); (5) Celtic sand banks (Reynaud et al., 1999); fluvial palaeovalleys (blue dashed lines) (Larsonneur et al., 1982); submarine drainage basins: (A) 'Grande Sole', (B) 'Petite Sole', (C) 'la Chapelle', (D) 'Ouest Bretagne', (E) 'Sud Bretagne' (Bourillet et al., 2003); (6) Celtic turbidite system (Droz et al., 1999; Zaragosi et al., 2000); (7) Armorican turbidite system (Zaragosi et al., 2001a). Red circles indicate the core locations.

Figure 2: Shaded morphologic map of the Celtic – Armorican margin. White circles and associated numbers indicate core locations.

Figure 3: Recognition of turbiditic and ice-rafted laminae on X-ray imagery and on the microscope using thin-sections of impregnated sediment.

Figure 4: Abundance (%) of foraminifera *N. pachyderma* (s.) in cores MD04-2836, MD04-2837, MD03-2690 (Zaragosi et al., 2006), MD03-2688, MD03-2695 (this study) and MD95-2002 (Zaragosi et al., 2001b). Black triangles indicate the depth of samples used for AMS dating. Blue shading corresponds to cold periods. Dashed red lines represent core-to-core correlation using the limits of cold episodes. Hol: Holocene, YD: Younger Dryas, BA: Bölling-Alleröd, HE: Heinrich events (1 to 3), LGM: Last Glacial Maximum.

Figure 5: Examples of some representative X-rayed slabs of lithofacies 1 to 6.

Figure 6: Evolution of sedimentation rates (continuous black line - $cm.ka^{-1}$) and of turbidite deposit frequency (continuous red line – $turb.ka^{-1}$) in cores MD04-2836, MD03-2690, MD03-2688 and MD03-2695. Although sedimentation rates must be considered with precaution because of frequent oversampling in Calypso piston cores (Skinner and McCave, 2003), the resulting curves parallel those of the turbidite deposit frequency suggesting that sedimentation rates can be considered as fairly valid.

Figure 7: Example of a thin-section of impregnated sediment (left) from lithofacies 4 in core MD03-2690, X-ray (middle) and grain-size measurements (right). D50 (continuous line) = grain-size at which 50% of sample is finer; D90 (dashed line) = grain-size at which 90% of sample is finer. Open circles represent the chosen samples used for grain-size analysis. Black layers represent turbidite deposits in the X-ray imagery.

Figure 8: Example of an X-rayed slab (left) from lithofacies 6 in core MD03-2695 and grain-size measurements (right). D50 (continuous line) = grain-size at which 50% of sample is finer; D90 (dashed line) = grain-size at which 90% of sample is finer. Open circles represent the chosen samples used for grain-size analysis. Black layers represent turbidite deposits in the X-ray imagery.

Figure 9: Evolution of the turbidite deposit frequency (histograms – turb. ka^{-1}) using time slices (1 ka) of the age models of cores MD04-2836, MD03-2690, MD03-2688 and MD03-2695. The continuous red line represents the frequency of turbidite deposits (turb. ka^{-1}) calculated by using two consecutive control-points.

Figure 10: Comparison between: (A) sedimentologic data of core MD03-2690 and (B, C) palaeoclimatic records of core MD95-2002 between 27 and 12 cal. ka. (A) Histogram represents the turbidite deposit frequency with a 250-year resolution (turb./0.25 ka). Red curves show the relative sea-level evolution (m.) described by Fairbanks et al. (1989) (continuous line) and by Waelbroeck et al. (2002) (dashed line). Blue line represents the rate of sea-level rise (mm.a^{-1}) and the 'Meltwater Pulse 1A' (MWP 1A - ca. 14 cal ka) (Fairbanks, 1989). (B) BIT-index (Ménot et al., 2006) and *Pediastrum* sp. abundances (#.cm^{-3}) (Eynaud, 1999; Zaragosi et al., 2001b). (C) Abundances of IRD > 150 µm (#/g.) and *N. pachyderma* (s.) (%) (Zaragosi et al., 2001b).

Figure 11: Comparison between turbidite deposit frequency in the Celtic turbidite system (core MD04-2836) (this study) and the BIIS oscillations in the Irish Sea Basin (McCabe et al., 2007b). Continuous red line show the frequency of turbidite deposits (turb.ka^{-1}) between two consecutive control-points of the age-model of core MD04-2836 while the histogram show the turbidite deposit frequency using a time slicing of 1 ka. 1: 'Cooley Point Interstadial' (from ≥16.7 to ≤15 ^{14}C ka BP); 2: 'Clogher Head Stadial' (from ≥15 to ≤14.2 ^{14}C ka BP); 3: 'Linns Interstadial' (~14.2 ^{14}C ka BP); 4: 'Killard Point Stadial' (from ≥14.2 to ~13.0 ^{14}C ka BP); 5: 'Rough Island Interstadial' (after ~13.0 ^{14}C ka BP) (McCabe et al., 2007b). Note the close relationship between the main retreat periods of the BIIS and enhanced turbiditic activity in the Whittard channel.

TABLE CAPTION

Table 1: Key parameters of cores discussed in this study including core number, geographic position, water depth and oceanographic missions.

Table 2: Radiocarbon ages of cores MD04-2836, MD03-2688, MD03-2690 and MD03-2695 and of the neighbouring core MD95-2002. Radiocarbon ages of this study were performed at the ‘Laboratoire de Mesure du Carbone 14’ in Saclay (‘SacA’). Radiocarbon dates have been corrected for a marine reservoir effect of 400 years and calibrated to calendar years using CALIB Rev 5.0 / Marine04 data set (Hughen et al., 2004; Stuiver and Reimer, 1993; Stuiver et al., 2005) up to 21.78 ^{14}C ka and Bard et al. (1998) thereafter.

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Table1

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Core number	Latitude	Longitude	Depth (m)	Cruise	Year	institute
MD952002	47° 27.12' N	08° 32.03' W	2.174	MD 105 IMAGE 1	1995	IFREMER
MD032688	46° 48.03' N	07° 02.93' W	4.385	MD 133 SEDICAR	2003	IFREMER
MD032690	47° 01.25' N	07° 44.99' W	4.340	MD 133 SEDICAR	2003	IFREMER
MD032695	47° 43,14' N	06° 12,68' W	4.375	MD 133 SEDICAR	2003	IFREMER
MD042836	47° 16.57' N	10° 07.69' W	4.362	MD 141 ALIENOR	2004	IFREMER
MD042837	47° 31.99' N	09° 44.01' W	4.176	MD 141 ALIENOR	2004	IFREMER

Table2

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Core number	Depth (cm)	Material	Laboratory Number	Corrected ^{14}C age (yr BP)	Calendar age (cal yr BP)	Data origin
MD95-2002	0	*G. Bulloides*	LSCE-99360	1660 +/- 70	1624	Zaragosi et al., 2001
MD95-2002	140	*G. Bulloides*	LSCE-99361	9080 +/- 90	10329	Zaragosi et al., 2001
MD95-2002	240	*N. pachyderma* s.	LSCE-99362	10790 +/- 100	12809	Zaragosi et al., 2001
MD95-2002	420	*N. pachyderma* s.	LSCE-99363	13330 +/- 130	15798	Zaragosi et al., 2001
MD95-2002	454	*N. pachyderma* s.	LSCE-99364	13800 +/- 110	16426	Zaragosi et al., 2001
MD95-2002	463	*N. pachyderma* s.	LSCE-99365	14020 +/- 120	16709	Zaragosi et al., 2001
MD95-2002	510	*N. pachyderma* s.	LSCE-99366	14170 +/- 130	16897	Zaragosi et al., 2001
MD95-2002	550	*N. pachyderma* s.	SacA-003242	14430 +/- 70	17327	Zaragosi et al., 2006
MD95-2002	580	*N. pachyderma* s.	Beta-141702	14410 +/- 200	17332	Zaragosi et al., 2001
MD95-2002	869	*N. pachyderma* s.	SacA-003243	14900 +/- 70	18241	Zaragosi et al., 2006
MD95-2002	875	*N. pachyderma* s.	SacA-003244	14880 +/- 160	18224	Zaragosi et al., 2006
MD95-2002	1320	*G. Bulloides*	SacA-003245	18450 +/- 90	22062	Zaragosi et al., 2006
MD95-2002	1340	*G. Bulloides*	SacA-003246	19030 +/- 100	22514	Zaragosi et al., 2006
MD95-2002	1390	*G. Bulloides*	SacA-003247	20220 +/- 80	24690	Zaragosi et al., 2006
MD95-2002	1424	*N. pachyderma* s.	Beta-123696	19840 +/- 60	23777	Grousset et al., 2000
MD95-2002	1453	*N. pachyderma* s.	Beta-123698	20030 +/- 80	23984	Grousset et al., 2000
MD95-2002	1464	*N. pachyderma* s.	Beta-123699	20200 +/- 80	24174	Grousset et al., 2000
MD95-2002	1534	*N. pachyderma* s.	Beta-123697	21850 +/- 70	25734	Grousset et al., 2000
MD95-2002	1610	*N. pachyderma* s.	Beta-99367	24010 +/- 250	28222	Auffret et al., 2002
MD95-2002	1664	*N. pachyderma* s.	Beta-99368	25420 +/- 230	29830	Auffret et al., 2002
MD04-2836	100.5			9275	10700	correlation MD95-2002
MD04-2836	150.5	*N. pachyderma s.*	SacA-003248	10730+/-50	12788	Zaragosi et al., 2006
MD04-2836	411.5			11900	13938	correlation MD95-2002
MD04-2836	1354.5	*N. pachyderma s.*	SacA-003249	12840+/-120	15159	Zaragosi et al., 2006
MD04-2836	1656.5	*N. pachyderma s.*	SacA-003253	13480+/-60	16017	Zaragosi et al., 2006
MD04-2836	1761.5	*N. pachyderma s.*	SacA-003254	14210+/-70	16956	Zaragosi et al., 2006
MD04-2836	2131,5	*N. pachyderma s.*	SacA-005971	14650+/-50	17727	this paper
MD04-2836	2534.5			15091	18396	correlation MD95-2002
MD04-2836	3525.5	*N. pachyderma s.*	SacA-003256	17090+/-80	20209	Zaragosi et al., 2006
MD03-2688	157	*G. bulloides*	SacA-004927	8495 +/- 35	9541	this paper
MD03-2688	480	*N. pachyderma s.*	SacA-004928	12580 +/- 90	14751	this paper
MD03-2688	1084	*N. pachyderma s.*	SacA-004929	14200 +/- 70	16941	this paper
MD03-2688	1704	*N. pachyderma s.*	SacA-004930	14650 +/- 110	17699	this paper
MD03-2688	1955			15091	18396	correlation MD95-2002
MD03-2688	2422	*G. bulloides*	SacA-004931	16930 +/- 80	20057	this paper
MD03-2688	2695			20220	23722	correlation MD95-2002
MD03-2688	2910	*N. pachyderma s.*	SacA-004932	21570 +/- 110	25410	this paper
MD03-2688	3136	*N. pachyderma s.*	SacA-004933	24890 +/- 140	29227	this paper
MD03-2688	3520	*G. bulloides*	SacA-004793	29160 +/- 180	34038	this paper
MD03-2690	151	*G. Bulloides*	SacA-001894	8730 +/- 60	9900	Zaragosi et al., 2006
MD03-2690	245	*G. Bulloides*	SacA-003233	9450 +/- 60	10774	Zaragosi et al., 2006
MD03-2690	425			11900	13938	correlation MD95-2002
MD03-2690	626	*G. Bulloides*	SacA-003234	12620 +/- 60	14863	Zaragosi et al., 2006
MD03-2690	692	*N. pachyderma* s.	SacA-003235	12770 +/- 70	15074	Zaragosi et al., 2006
MD03-2690	1094	*N. pachyderma* s.	SacA-003236	13840 +/- 70	16483	Zaragosi et al., 2006
MD03-2690	1213	*N. pachyderma* s.	SacA-003237	14030 +/- 70	16715	Zaragosi et al., 2006
MD03-2690	1885	*N. pachyderma* s.	SacA-003238	14650 +/- 70	17717	Zaragosi et al., 2006
MD03-2690	2233	*N. pachyderma* s.	SacA-003239	14960 +/- 70	18287	Zaragosi et al., 2006
MD03-2690	2276	*N. pachyderma* s.	Poz. Rad. Lab.	15080 +/- 70	18392	Zaragosi et al., 2006
MD03-2690	2923	*G. Bulloides*	SacA-005972	16990 +/- 110	20115	this paper
MD03-2690	3156	*G. Bulloides*	SacA-003240	18850 +/- 100	22378	Zaragosi et al., 2006
MD03-2690	3376	*N. pachyderma* s.	SacA-003241	20560 +/- 70	24600	Zaragosi et al., 2006
MD03-2690	3576	*N. pachyderma* s.	Poz. Rad. Lab.	21880 +/- 120	25769	Zaragosi et al., 2006
MD03-2695	242			13463	15970	correlation MD95-2002
MD03-2695	878	*N. pachyderma s.*	SacA-005609	14640+/-60	17703	this paper
MD03-2695	1187.5	*N. pachyderma s.*	SacA-005610	14830+/- 60	18030	this paper
MD03-2695	1347	*N. pachyderma s.*	SacA-005611	14990+/- 60	18305	this paper
MD03-2695	1420			15091	18396	correlation MD95-2002
MD03-2695	1991	*N. pachyderma s.*	SacA-005612	17130+/- 70	20248	this paper
MD03-2695	2255	*N. pachyderma s.*	SacA-005613	20300+/- 100	24284	this paper
MD03-2695	2393			22,028.2	26032	correlation MD95-2002
MD03-2695	2444	*N. pachyderma s.*	SacA-005614	25600+/- 150	30034.4	this paper
MD03-2695	2600			27,400	32068	Elliot et al., 2001 (HE3)
MD03-2695	2758	*N. pachyderma s.*	SacA-005616	28710+/- 210	33536.2	this paper

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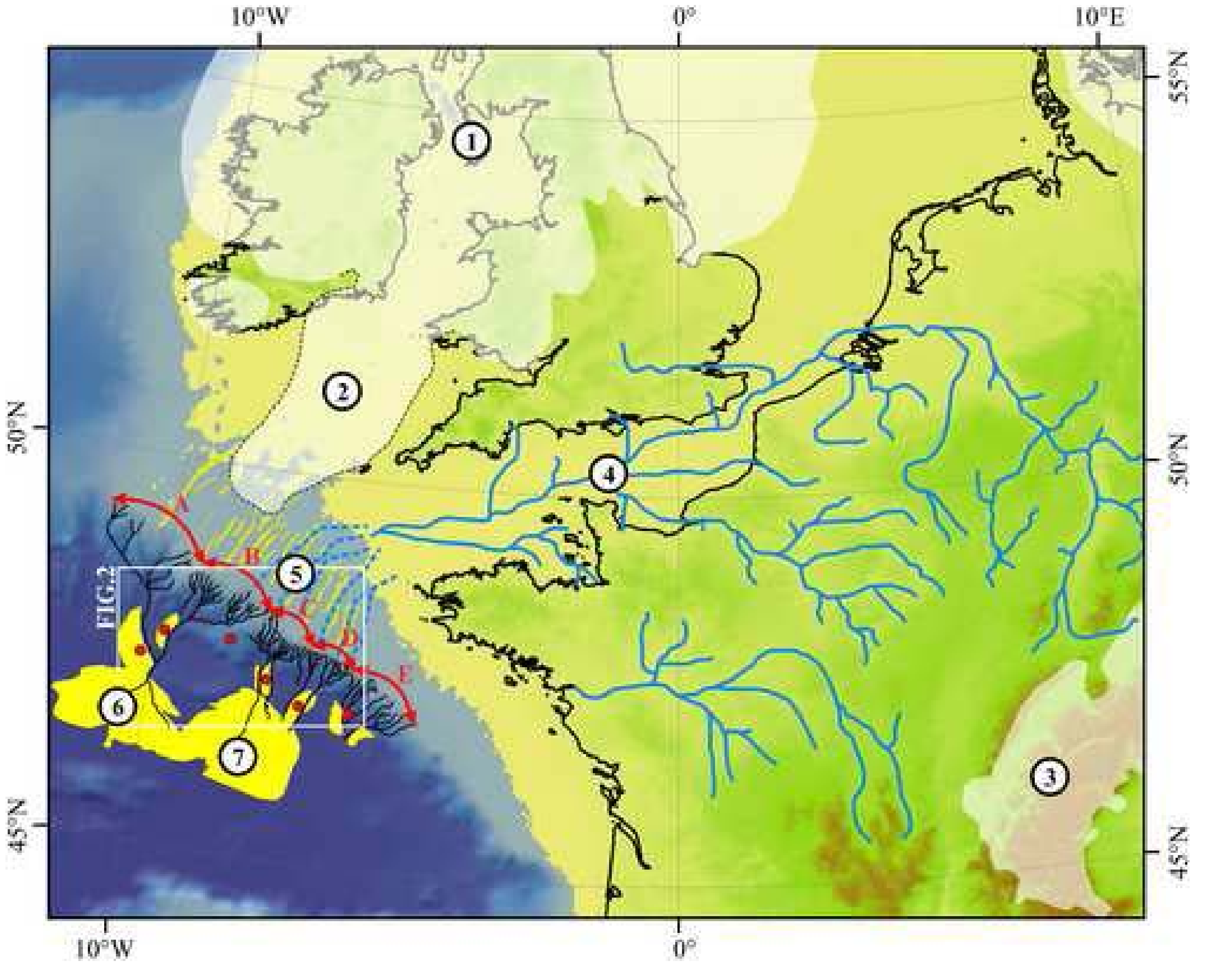

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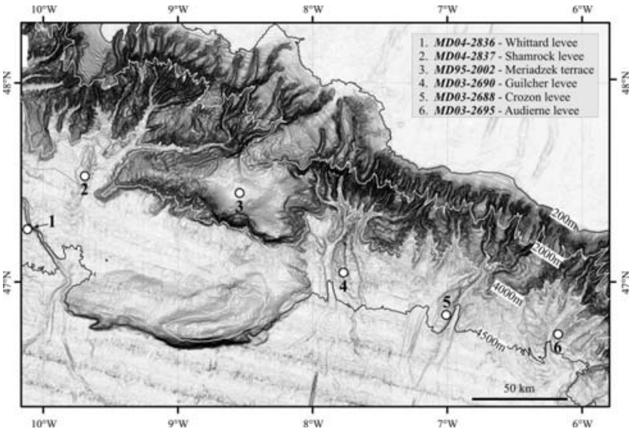

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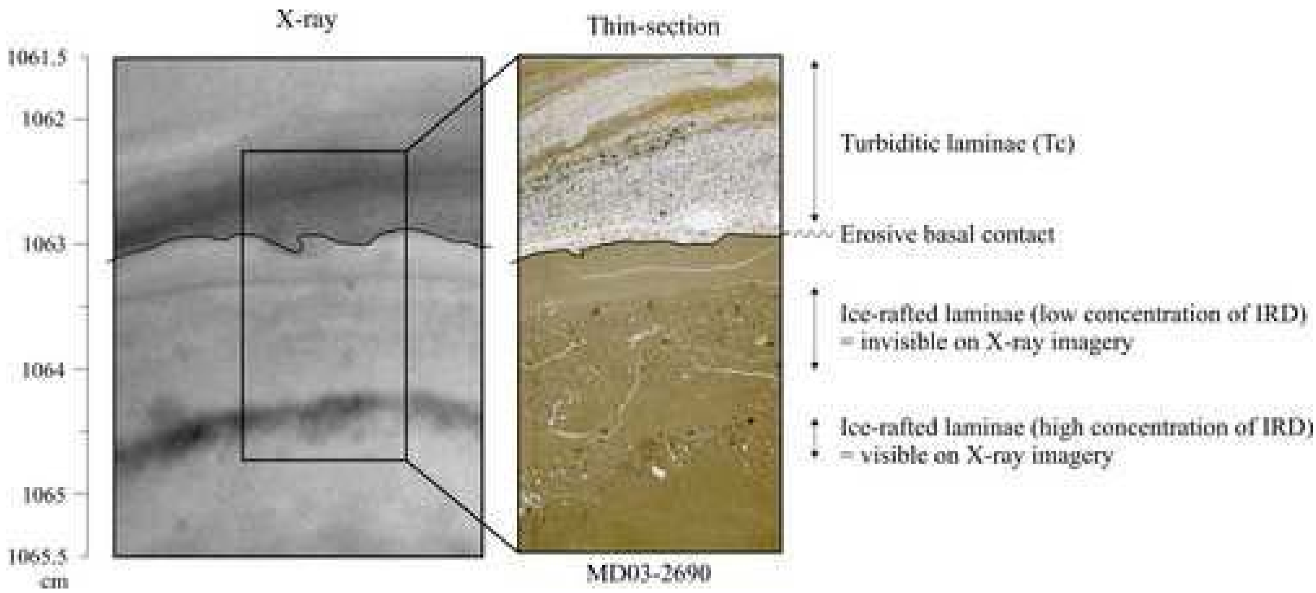

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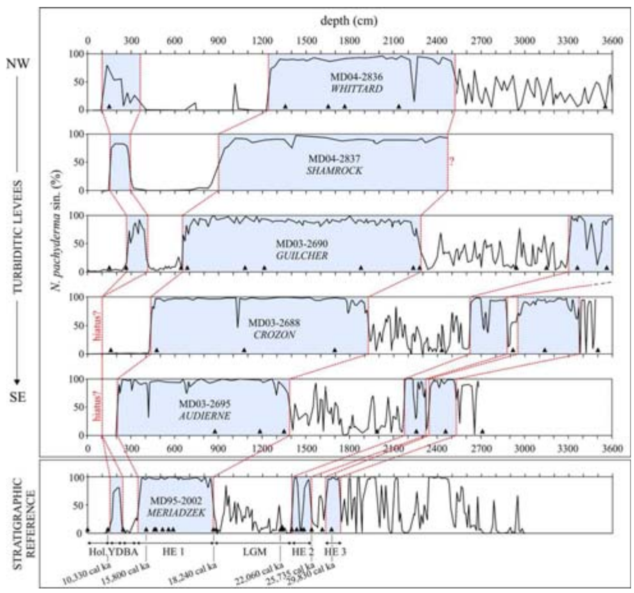

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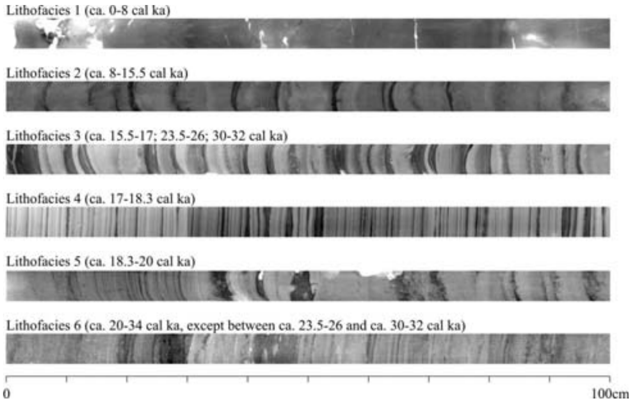

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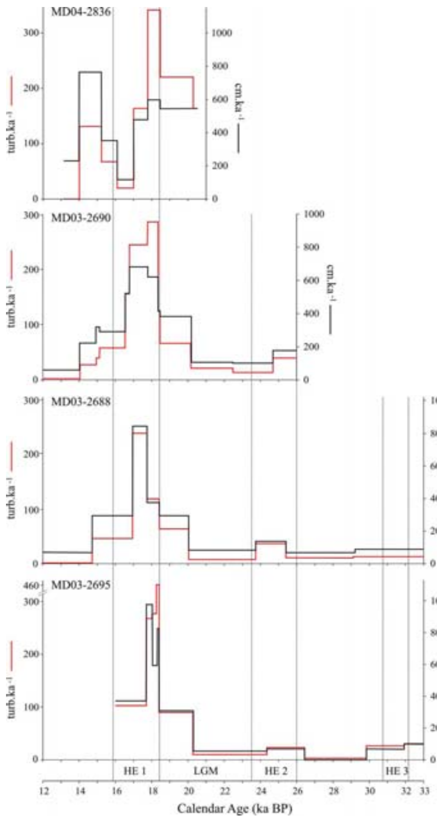

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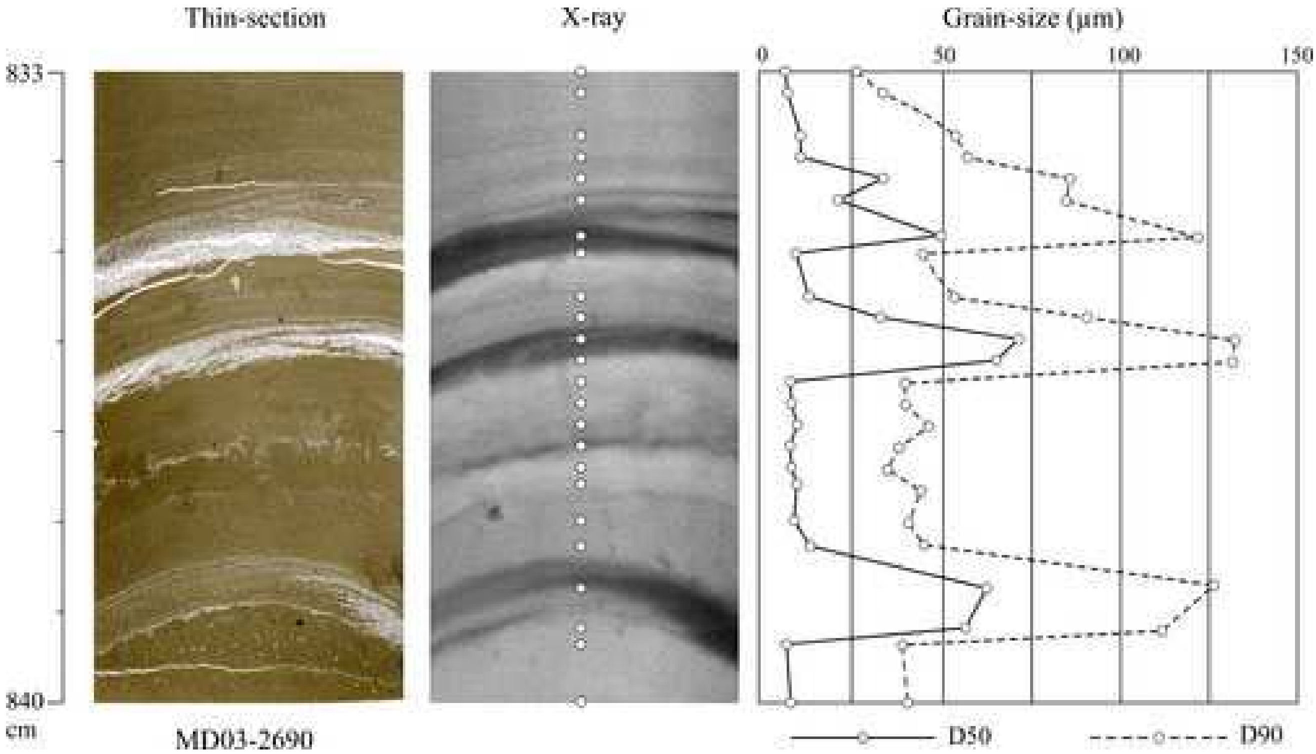

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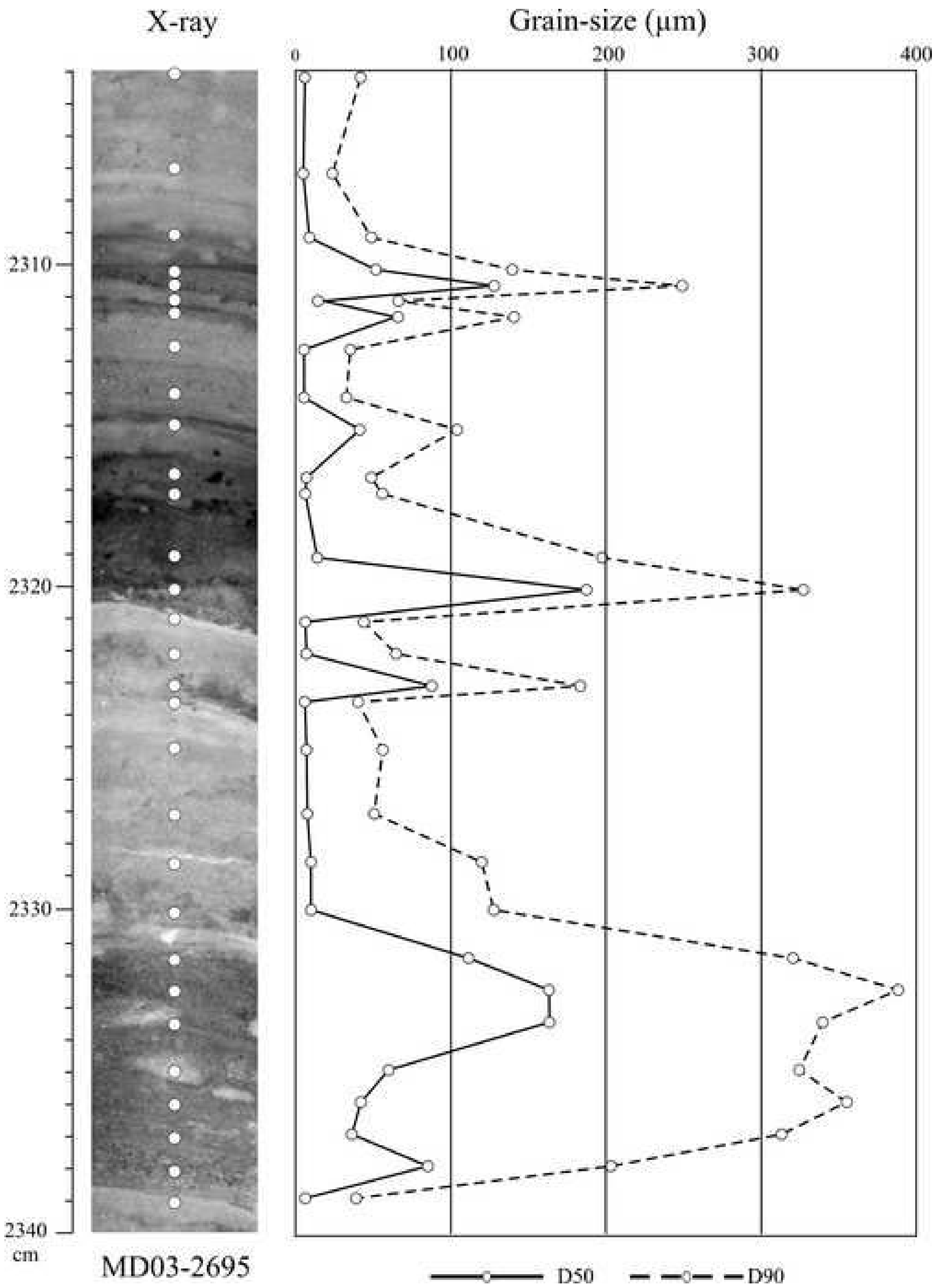

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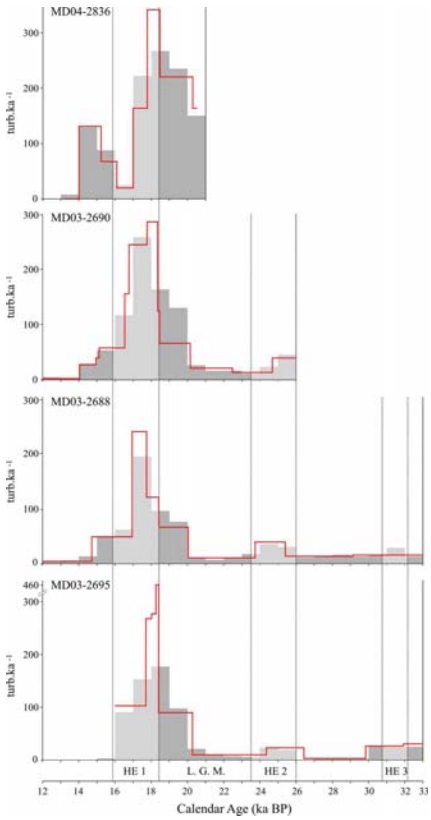

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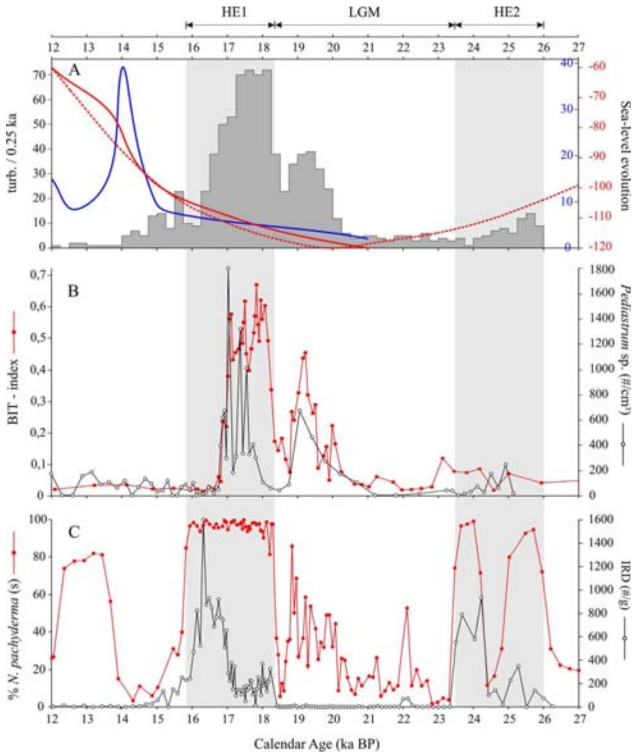

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