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Sedimentary Processes and the Creation of the Stratigraphic Record in the Late Quaternary North Atlantic Ocean [and Discussion]

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Sedimentary processes and the creation of the stratigraphic record in the Late Quaternary North Atlantic Ocean

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SUMMARY

The primary difficulty in the interpretation of the stratigraphic record is that a multiplicity of sedimentary processes, some producing similar effects, are responsible for it. We seek to unravel the effects of the more important processes through analysis of sedimentary properties. The effects to be unravelled are those relating to *pelagic input* (vertical flux) due to organic productivity, wind-blown dust, ice-rafting and volcanic ash; to *horizontal flux* in turbidity currents, debris flows, and nepheloid layers caused by the reworking of sea-bed sediments by internal waves and bottom currents; and to *degradation* of the record by dissolution, oxidation and mixing of components. Contrasting regions of the North Atlantic are used to show the effects of bottom currents, ice-rafting, wind and productivity on sediments. Applications to estimates of changes in bottom currents, productivity and carbon sequestration in the N.E. Atlantic over the past 30 ka are given.

1. INTRODUCTION

A feature of recent interpretations of stratigraphic records of environmental change recorded in marine sediments and ice caps is very rapid inferred changes. This has grown from, but not replaced the more gradualistic change, suggested by the great success of attributing environmental change to orbitally induced insolation changes (Hays *et al.* 1976). Very rapid, catastrophic events such as volcanic eruptions and turbidity current have long been recognized but seen as superimposed on gradual changes taking thousands of years. Recent work documenting significant temperature changes over a few tens of years recorded in ice (Dansgaard & Oeschger 1989), and huge changes in ice flux modelled as lasting a few hundred years (Bond *et al.* 1992; Dowdeswell *et al.* 1995), indicate an increasingly catastrophic structure to the stratigraphic record. Even in laminated, apparently varved, sequences, events of ENSO origin occurring every three to seven years may actually be responsible for the laminae (Christensen *et al.* 1994), whereas in other cases the laminae are indeed annual (Koide *et al.* 1972). We have to confront the magnitude–frequency structure of events that yield the stratigraphic record, and realize that resolution of the temporal structure of brief events requires stratigraphic completeness (*sensu* Anders *et al.* 1987) at short timescales of 10–100 years. Is the significance of what we observe simply a function of the temporal resolution we can obtain? To this the answer is clearly 'no', the magnitude of events is of paramount importance. But rapid events of large magnitude, particularly if of uncertain cause have excited more than geological interest recently. A satisfactory record is one in which the magnitude and frequency of events can be inferred so that their significance can be ascertained. To attain annual resolution a sedimentation rate of at least 0.3 mm a^{-1} or 30 cm ka^{-1} would be needed.

Such resolution would only be obtained if the sediment remained undisturbed after deposition. Unfortunately 99 % of marine sediments are mixed by burrowing organisms. Even in rapidly deposited sediments this degrades the resolution potential to a few hundred years if the benthic biota are not affected by events. These problems and possibilities are discussed below.

2. CONTROLS AND EFFECTS OF VERTICAL FLUX

Because of rapid particle sinking (McCave 1975), geologically integrated vertical flux depends on the rate of arrival at the sea surface (wind-blown dust deposition flux, iceberg sediment release rate) or net, new productivity.

Products of vertical flux have been captured by many sediment traps. It was initially surmised that faecal pellets would comprise a large part of the flux (Honjo 1976), but early sediment trap results showed that a large amount of apparently unaggregated fine sediment accompanied the less abundant faecal pellets (Honjo 1978). Radiochemical measurements show that short half-life particle-reactive elements (e.g. ^{234}Th, $t_{\frac{1}{2}} = 24d$) are being rapidly taken down to the sea bed

(DeMaster *et al.* 1985). This probably involves scavenging by colloidal particles, Brownian aggregation, and collection of the small aggregates by large rapidly sinking ones (Honeyman & Santschi 1989; McCave 1984; Hill & Nowell 1990). The result is that the particle population delivered to the sea bed is not appreciably sorted by the processes involved in vertical flux. An unsorted mixture delivered to the sea surface (e.g. ice-rafted detritus, IRD) should arrive at the sea bed with most of its fine particles. Terrigenous sand and most foraminifera have settling speeds > 1 cm s^{-1} and reach the bed in less than a week. Measurements of aggregate sinking speed and time-series sediment traps indicate delivery to abyssal depths in a month or less at 100–200 m d^{-1}. On these timescales the lateral displacement from the point of input or origin of sinking detritus does not exceed a few tens of kilometres, except in the rare cases of unidirectional current velocity structure from surface to bottom when it could be a few hundred. Vertical flux should faithfully deliver a record of biological productivity, carbonate and silica skeletal mineralization, IRD flux, volcanic eruption magnitude, and wind strength. It is other processes that degrade it so drastically.

3. CONTROLS AND EFFECTS OF LATERAL FLUX AND RESUSPENSION

(*a*) *Turbidity currents and debris flows*

The supply to the Atlantic basins differs greatly and ranges from rare but huge flows of the order of 100 km^3 each on the Madeira Abyssal Plain (Weaver *et al.* 1992), to the frequent small flows of 1–10 km^3 on the Tagus and Horseshoe Abyssal Plains off Portugal (Lebreiro 1995), for example. It is often assumed that turbidity currents are more frequent at low stands of sea level, yet the available data suggest all of the following: triggering at times of sea-level change, triggering by 23 ka climatic changes, and uniform high frequency from glacial to Holocene (Schönfeld & Kudrass 1991; Weaver *et al.* 1992; Weltje & de Boer 1993). The mass failures yielding large turbidity currents and debris flows involve failure of ten to several tens of metres of sediment on the slope (Embley & Jacobi 1977), thereby mixing the products of several glacial–interglacial cycles (Weaver & Thomson 1993). Basal erosion by turbidity currents will also produce a mixture of sediment ages. Gravity flows do not, therefore, deliver sediment bearing a pure signature characteristic of conditions in the source area at the time they were triggered. The record in turbidites is thus of events and their dynamics, but with scrambled information from the continent, the sea bed in the area of initial deposition and the path over which the current has flowed. There are very few records taken from debris flows but the faunal work of Jansen *et al.* (1987) on cores from the Storegga Slide and Simm *et al.* (1991) from the Saharan Slide shows them to contain a mixture of Pleistocene and sometimes older shelf, slope and deep water faunal assemblages.

The influence of these mass flows on the stratigraphic record is widespread and usually obvious and avoidable when seeking palaeo-environmental records. The potential influence of the fine tails of turbidity currents entrained in deep current systems on continental margins is more insidious because it is largely unrecognized. A particular marker that demonstrates the effect, however, is the red/pink sediment colour due to clay-sized haematite under the Western Boundary Undercurrent (WBUC) occurring in sediments deposited during isotope stages 2 and 6, times of the lowest sea level derived from turbidites on the Laurentian Fan (Hollister & Heezen 1972). If the material were fine coccoliths deposited in a warmer period, analysis of their organic biomarkers would appear to show an anomalous warm pulse in a cold period.

(*b*) *Resuspension and deposition*

Rates of sedimentation in the North Atlantic arising from vertical flux, away from continental margins, are typically 1–2 cm ka^{-1} during interglacials and 3–4 cm ka^{-1} during glacials, averaging about 2 cm ka^{-1} over much of the area since the last glacial maximum (Balsam & McCoy 1987). However, under deep current systems rates of deposition of 10–20 cm ka^{-1} are not unusual. Clearly there is a substantial addition of material that has been eroded from somewhere and deposited in places by deep currents.

(i) *Critical erosion conditions*

Well-sorted fine sediments behave in a non-cohesive manner down to about 10 μm diameter. Below this size, particles become cohesive partly because clay minerals enter the compositional spectrum and van der Waals forces become significant in particle adhesion (Russel 1980), even for quartz (figure 1). The sea bed generally has a loose layer of aggregates produced by biological action, weakly bound to the substrate. On the Nova Scotian Rise turbidity, hydrographic and current meter data indicate intermittently strong currents and sediment resuspension (Hollister & McCave 1984) with a range of measured eroding and depositing velocities consistent with the laboratory values of erosion and deposition stresses. Experimental data indicate a critical erosion shear velocity (u_{*c}) of $\sim$ 0.70 cm s^{-1} for 16 μm silt, which is close to the 0.68 cm s^{-1} deduced for the HEBBLE site from the data of Gross *et al.* (1988), and is produced by a geostrophic current speed, U_g, of 15–20 cm s^{-1} (using $u_* = U_g/30\text{-}U_g/22$).

(ii) *Critical deposition conditions*

Measurements of critical deposition stress, τ_d, by Self *et al.* (1989) give the approximate relation $\tau_d \approx 10^3 d$ for quartz-density solids in water (in S.I. units), thus deposition of 10 μm particles occurs at stresses $<$ 0.010 Pa (given by U_g = 7–10 cm s^{-1}). These values, are rather less than those of McCave & Swift (1976) based on a non-cohesive erosion stress ($\sim$ 0.045 Pa, U_g = 15–20 cm s^{-1}) or than 0.015–0.03 Pa (U_g = 8–23 cm s^{-1}) via the analytical expression of Dade *et al.* (1992). This range of depositional shear stresses

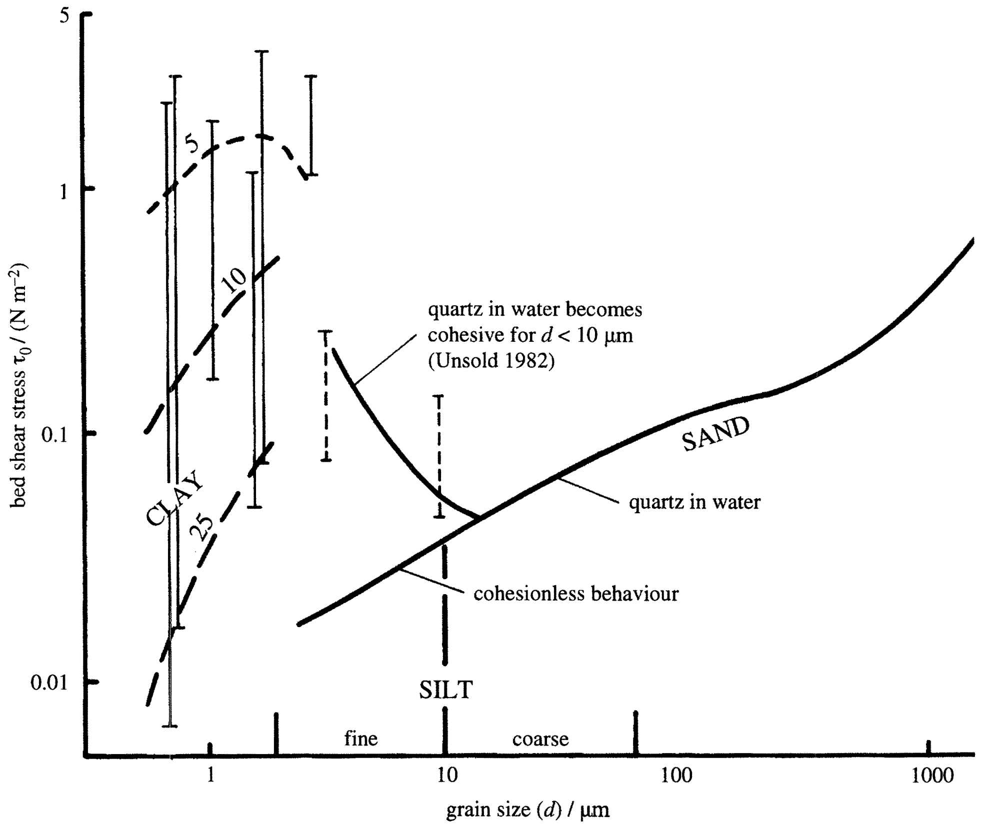

Figure 1. Critical bed shear stress τ_0 for erosion of quartz showing the onset of a cohesion effect at about 10 µm, after Unsold (1982). The data in the clay range were taken from various authors and are contoured for voids ratio.

(0.01–0.045 Pa) is a little less than that required to break up clay flocs, namely 0.04–0.16 Pa (U_g = 13–38 cm s^{-1}) (Hunt 1986). Deposition of small aggregates can thus take place under geostrophic current speeds less than 7–15 cm s^{-1} but at higher speeds they are likely to be broken up and transported downstream. Primary deposition of much larger aggregates (marine snow) on the sea floor also occurs. These are subject to degradation and, in current-swept areas, resuspension and deposition of their surviving components. Only in very low stress environments will the aggregates contributing to vertical flux be unaffected by physical reworking. As they constitute the food supply for the benthos they are rapidly processed biologically.

(iii) *Sorting*

The processes of erosion and deposition lead to sediment sorting which occurs by processes of aggregate break-up and particle selection according to settling velocity and stress. Sorting takes place through differing rates of sediment transport, so that an originally unsorted mixture is converted downstream into narrower distributions. During deposition particle populations are sorted by some grains and aggregates being trapped in the viscous sublayer, while others of smaller settling velocity w_s, in general of the order $w_s < 0.64\, u_*$ (Allen 1971) are not, and are transported further downcurrent. There is a substantial region of overlap between the trapped and rejected populations. The controlling variables are the critical erosion stress (τ_e), the critical suspension stress (τ_s) which may be greater than the erosion stress for fine non-cohesive silts according to Dade *et al.* (1992), and the critical deposition stress (τ_d). In general $\tau_d < \tau_e < \tau_s$ giving a bedload region for fine sediment between τ_e and τ_s, (Dade *et al.* 1992) in which silt ripples are found (Mantz 1978; Hollister & McCave 1984).

In regions where the currents frequently exceed about 20 cm s^{-1} the fine components are removed, leaving a lag of sand. This is most frequently foraminiferal sand with a particle density of 1.5 g cm^{-3} and a critical erosion shear stress of 0.067–0.085 Pa for 200–300 µm sand (Miller & Komar 1977), (U_g of 24–27 cm s^{-1}). Although the foram sand has the erodibility of very coarse silt, it has the settling velocity of very fine sand and is not transported far, being largely confined to bedload. To move quartz sand of this size would take currents of U_g = 35–37 cm s^{-1}. The speeds required to move foraminiferal sand are generally not exceeded frequently enough to yield a well sorted sand: a notable exception is sand dune valley in Panama Basin (Lonsdale & Malfait 1974) and sand dunes have also been detected on the lower flanks of Hatton Drift (McCave *et al.* 1980). Medium quartz sand is moved so rarely in the deep sea that it may be regarded as immobile.

The significance of this is that the deposits of lateral flux driven by deep sea currents are in the silt–clay size range and are sorted to a greater or lesser extent depending on currents, whereas those resulting from vertical flux comprise both fines, foraminiferal sand, and during glacials, quartz sand and gravel. Only the sand and gravel sized material may be unambiguously related to the properties of the water column overlying it. In current-affected areas estimates of productivity, for example inferred from foraminiferal properties and from fine sediment properties (e.g. coccolith carbonate, organic carbon), will probably differ considerably.

(*c*) *The nature of the currents that move deep-sea sediment*

Several idealized cases of current-time series may be imagined (figure 2). In case A the flow speed is constantly above the critical erosion threshold for a certain size, resulting in removal of that material and a residue of coarser sediment (a situation with high mean kinetic energy KE_M). A current that never reaches critical (case B) will produce no effect, even though it has a faster mean than a variable current (case C) which occasionally exceeds critical. This is the common case of a current with an underlying mean and an added source of variability. Finally, case D represents the situation of zero (or very small) mean current and large variability (high eddy kinetic energy KE_E)such that currents in several directions are able to move sediment but with zero net transport. The important cases are clearly C and D. Both are capable of producing sorted sediment, but only case C is likely to yield the net flux of sediment and high accumulation rate typical of 'contourite' sediments (deposited under deep thermohaline contour-following currents).

The origins of the mean flow are mainly deep geostrophic currents driven by temperature and salinity differences (Warren 1981). There is also a strong influence of the magnitude of the slope (α) along which geostrophic currents flow (where isotypically are nearly parallel with the bed, $U_g = g'\alpha/f$ in which f is coriolis frequency and g' is reduced gravity). A steeper bed slope supports a locally steeper isopycnal gradient and faster current, thus on steep scarps very fast currents are inferred geostrophically and measured by current meters. In some regions of significant eddy activity a deep mean flow may be driven by eddies, and Hogg *et al.* (1986) suggest that the deep flow over the Nova Scotian Rise is driven in this way. The idealized time series D of figure 2 is thus somewhat unrealistic in that regions of high eddy energy are likely to have a mean flow driven by eddies.

In recent years the pronounced effects of eddy energy on the sea bed have been recognized as 'benthic storms' (Gardner & Sullivan 1981; Hollister & McCave 1984; Gross *et al.* 1988). In these, long period variability of the deep flow shows short episodes of intense sediment resuspension. The eddy motions are coherent throughout the water column from top to bottom. This means that regions of high abyssal KE_E underlie high surface KE_E (Schmitz 1984) which is mapped by satellites using the variability of sea surface elevation or slope (Shum *et al.* 1990). One can identify those oceanic regions where strong mean flow and those where strong eddy variability is likely to occur (Dickson 1983). In some regions both signals are found and these will present difficulties in interpretation of sediment. To infer flow speed related to palaeocirculation, regions of low KE_E and high KE_M should be sampled. The satellite maps of KE_E suggest regions where the Pleistocene record may at best be ambiguous.

4. DEGRADATION OF THE RECORD

The previous section clearly documents the physical processes that degrade the record while augmenting its depositional flux. Particular aspects of degradation are dissolution and remineralization, winnowing and focusing, bioturbative mixing and pumping, and the impact of fluctuating sedimentation rates on biological mixing.

(*a*) *Biological mixing and pumping*

The most serious degradation of the sediment record's temporal resolution occurs through biological mixing by burrowing animals for which the parameter $(D_B/SL) = G$ expresses the balance between mixing and accumulation, where D_B = biological diffusivity, S = linear sedimentation rate and L = mixed layer depth (Guinasso & Schink 1975). Small values of $G <$ 0.1, are argued by Nittrouer & Sternberg (1981) to characterize laminated sediment while $G > 10$ denotes complete mixing. D_B for BOFS box cores using ^{210}Pb are 0.21–0.66 $cm^2\ a^{-1}$ in unwinnowed areas (Thomson *et al.* 1993*a*). The lead is mainly attached to the fine particles and records their mixing behaviour. Wheatcroft's (1992) experiments show that the diffusivity of 300 μm sand is about a tenth of that for 10 μm silt particles. Larger particles should thus show less stratigraphic dispersal than fine ones.

A new problem is that coexisting fine carbonate (coccoliths) and foraminifera may have significantly different radiocarbon ages, the coarser material being older. This emerges most clearly from recent work by Thomson *et al.* (1995) who show offsets of 1.1 and 0.76 ka, apparently inversely proportional to sedimentation rate, but the effect had been seen earlier by Jones *et al.* (1989) and Paull *et al.* (1991) (figure 3). Having foraminifera older than associated fine carbonate is the reverse of what would be expected from homogeneous bioturbation with a lower diffusivity for the sand. It is also the reverse of what would be expected if the fine carbonate contained a redeposited component transported laterally to the site of deposition. As age models are usually based on AMS ^{14}C dated foraminifera, this offset makes the interpretation of historical changes recorded in the fine fraction rather difficult. At present the most plausible explanation for this behaviour is that the forams are 'pumped' upwards by burrowing meiofauna whose body diameter is comparable to or less than that of the foraminifera, as proposed by McCave (1988) to explain aspects of gravel distribution in sediments. It was argued that gravel was pumped upwards for $G \geqslant 10$ (with G based on radioactive tracers on the fine fraction). The values of G for BOFS box cores 4C and 8C (stations 11881 and 11886) are 9 and 14, in the right range for pumping to occur.

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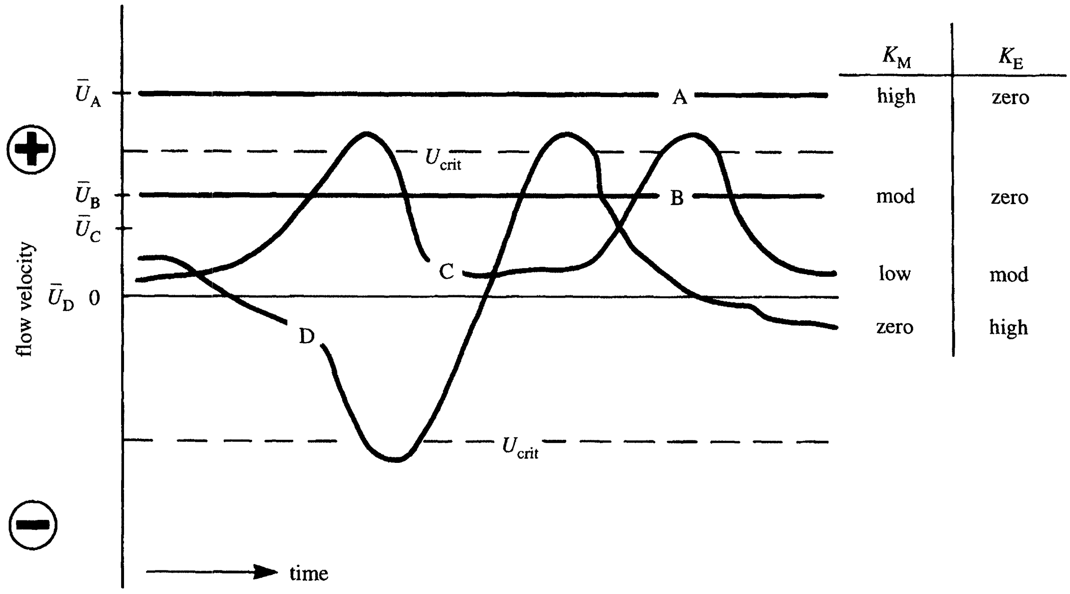

Figure 2. Idealized time series of currents. The time axis could represent a scale of days or months. See text for discussion of cases.

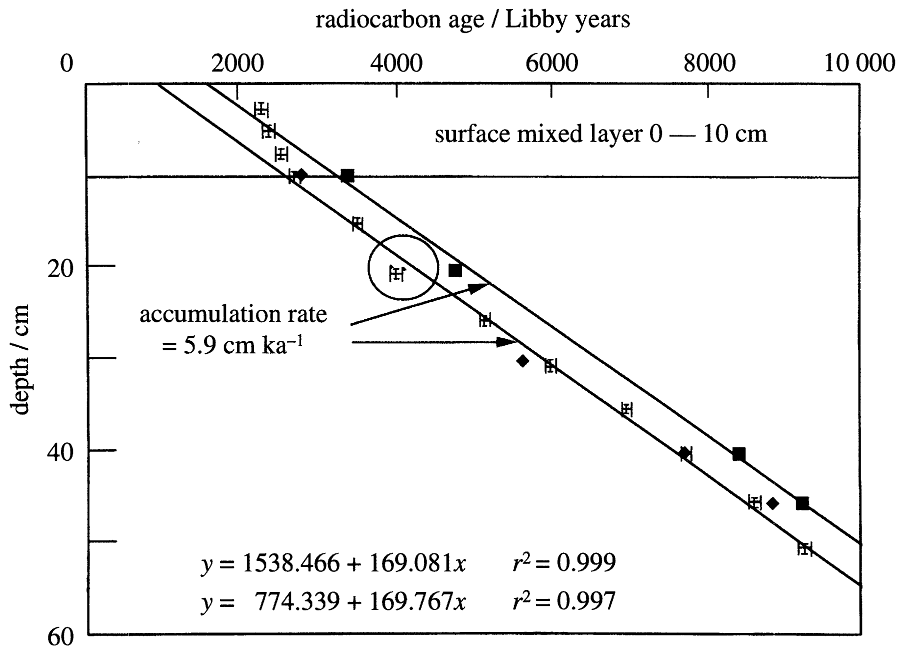

Figure 3. Radiocarbon age and depth profile for box core BOFS 8C taken at Discovery Station 11886. Points with error bars are bulk sediment, black diamonds are < 5 μm fraction, squares are AMS ages of mixed pelagic foraminifera > 150 μm and circled triangles are AMS ages of the 5–20 fraction. From Thomson *et al.* (1995).

(*b*) *Interruptions to mixing*

Bioturbation and sedimentation models can take account of small changes in sedimentation rate. However, the record contains horizons where there are abrupt large rises in sedimentation rate, the most obvious example being turbidites where 10 cm may be deposited in a matter of an hour or so (figure 4). It has been observed by Wetzel (1983) that the uppermost layer of the sediment, a fossil sea bed containing fine scale burrows ('*Mycellia*' and *Protopalaeodictyon*), may be preserved beneath fine-grained turbidites (figure 4*a*). Normally, this layer is worked over by larger organisms as it descends, and the fine burrows are destroyed. Similar preservation of fine burrows is also found under thick volcanic ash layers. A feature of deposition in the N.E. Atlantic over the past 70 ka has been the rapid deposition of material from massive pulses of icebergs released from the Laurentian and maybe other ice sheets, the so-called Heinrich layers (Heinrich 1988). These layers also overlie the preserved uppermost tier characterized by filamentous burrows (figure 4*c*), testifying to rapid deposition. Modelled deposition rates suggest durations of 100–500 years for these events (Dowdeswell *et al.* 1995). These layers are sufficiently rapidly deposited that they are not colonized by burrowers until the end of the event, when a burrowed top to the layers appears (figure 4*d*). A

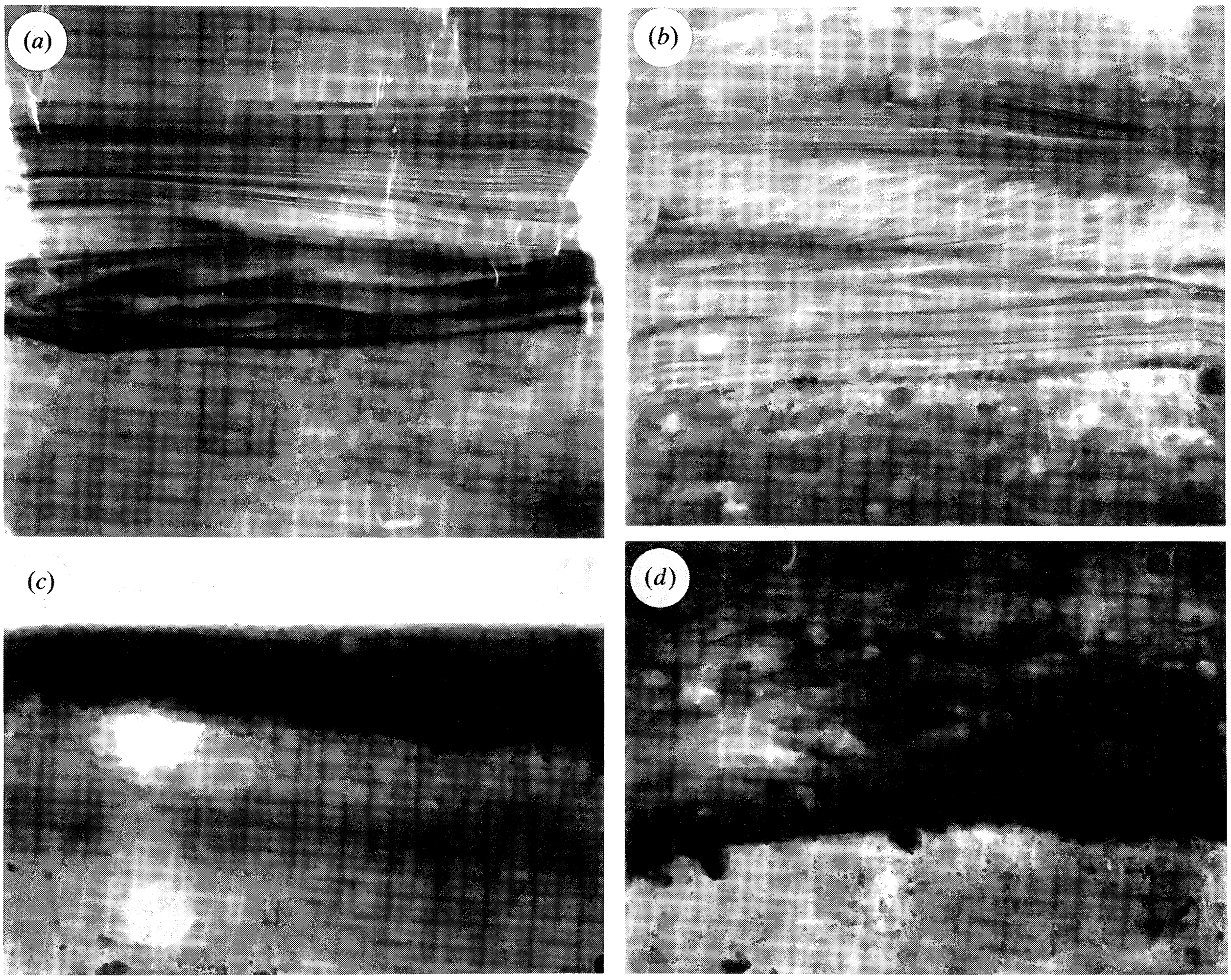

Figure 4. Photographic prints of X-radiographs of cores. Turbidites in cores (*a*) BOFS 15K, and (*b*) BOFS 2K. Possible preservation of fine burrows beneath the turbidites is visible in (*a*) and (*b*), and onset of burrowing can be seen in the top of (*b*) which also shows superb climbing ripple structure. (*c*) Heinrich layers with preserved fine burrow structures beneath BOFS 8K (H-2). (*d*) Burrows in the top of the layer BOFS 5K (H-2) which do not disturb its lower interface.

further consequence of this rapid deposition is the enhanced preservation of organic matter beneath the layers (Manighetti & McCave 1995*a*), (figure 5). Age modelling in sequences with pulses of rapid deposition must take account of the discontinuities in mixing which the pulses may induce, resulting in apparently wide age spreads for short events (Manighetti *et al.* 1995).

(*c*) *Winnowing and focusing*

The redistribution of sediment after deposition is shown by several physical and chemical indicators. On the East Thulean Rise (50°–52° N, 21°–23° W) three cores BOFS 5K, 6K and 7K, contain a topographically controlled glacial flux of fine materal (Manighetti & McCave 1995). Core 7K is on top of a hill, 6K is on a plateau and 5K is in a depression. All three show virtually the same (immobile) sand and gravel flux (figures 6 and 7). However the silt and clay flux is very low at 7K and high at 5K. None of these sites lie in the path of a significant deep current yet the pelagic input appears to show fines winnowed from 7K and focused at 5K. This is borne out by the radiochemical studies of Thomson *et al.* (1993*a*,*b*) who record 73% of the ^{210}Pb flux predicted from water-column decay at the site of 7K and up to 127% at 5K with similar trend for $^{230}Th_{xs}$. Both sedimentological and radiochemical data indicate focusing of material on current-affected sediment drifts in the deep passage just to the north of East Thulean Rise. In core BOFS 8K the glacial sand:mud ratio is 0.21 whereas in BOFS 6K it is 0.58 and in the Holocene the difference is more extreme, values being 0.20–0.71 respectively, testifying to increased Holocene currents. These are quite modest increments compared with what must be the situation on some drifts. The mean Atlantic sedimentation rate of 2 cm ka^{-1} (Balsam & McCoy 1987) is enhanced on some drifts, for example the Bahama Outer Ridge, to 20–25 cm ka^{-1} (Flood 1978). This degree of focusing must result from substantial winnowing of the continental margin turbidite fans, debris flows and ice rafted detritus. Lateral sorting along the western North Atlantic transport path results in the Greater Antilles

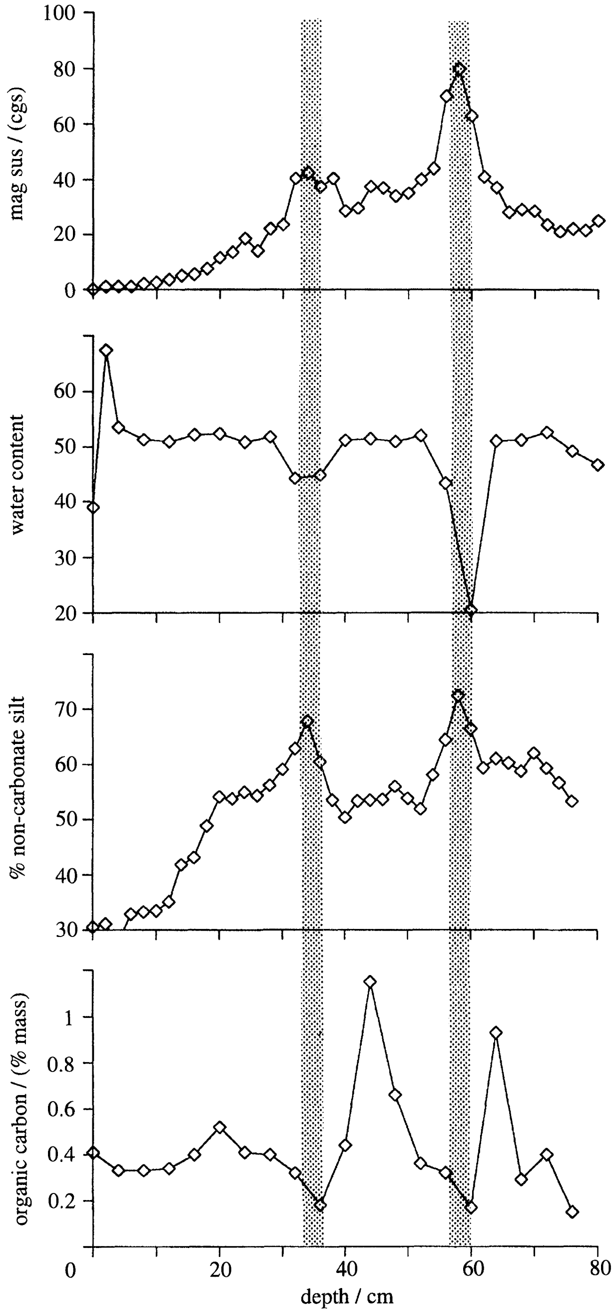

Figure 5. Magnetic susceptibility, water content, non-carbonate silt and organic carbon content with depth in core 6K. The shaded bands mark Heinrich Events 1 and 2. Note the position of organic carbon peaks in 6K just below the peaks in other properties indicative of ice-rafting pulses. (From Manighetti & McCave 1995*a*.)

Drift at 20° N, at the end of the transport path being almost entirely (> 90 %) fine silt and clay (Tucholke 1975).

5. THE PROBLEM OF OVERPRINTING

Unfortunately most sedimentary records bear the signature of several processes, organic productivity, wind strength, current strength and ice-rafting. Near to continents there is also variable supply and influence of rivers, turbidity currents and debris flows. The problem of inferring changes in the Quaternary sedimentation process is somewhat easier in areas unaffected by ice-rafting though that restricts us to areas south of 35°–40° N in the North Atlantic. It exemplifies one approach to the problem which is conditional sampling, others are the use of compositionally distinctive material, the examination of joint probabilities spectrally and the subtraction of one signature to reveal another.

(*a*) *Conditional sampling for wind-blown dust*

Aeolian dust is not distinctive in composition or size distribution, thus it cannot be extracted from deep sea sediment if it is mixed with material introduced by currents or ice. Many studies of aeolian sediment have been made off the Sahara (e.g. Parkin 1974; Sarnthein *et al.* 1981). On this margin there is also sediment introduced by turbidity currents, and deep currents (Jacobi & Hayes 1982; Lonsdale 1982). To infer aeolian history, material must therefore be selected to: (i) avoid current influences (avoid sediment with current laminae, and known zones of deep currents); (ii) have a carbonate-free accumulation rate which decreases monotonically along the wind path and is not too high; (iii) avoid any sediment with evidence of episodic deposition seen in X-radiographs as this may be due to turbidity currents. Good sites may be on elevated knolls and hills on the flanks of the Mid-Atlantic Ridge. At suitable sites terrigenous sediment size parameters can be related to wind strength (Parkin 1974).

(*b*) *Compositionally distinct productivity fluxes*

Size analysis of the total sediment followed by removal of the carbonate (and opal) and re-analysis for size allows one to estimate the size spectra of carbonate and non-carbonate (dominantly terrigenous) sediment (McCave *et al.* 1995). Analysis of organic carbon content permits estimates of productivity based on organic carbon flux and percentage. In the glacial the terrigenous fraction is assumed to be dominated by ice-rafted sediment because we have no way of estimating the aeolian contribution. In core BOFS 6K, regarded as neutral in terms of focusing, the organic carbon content increases up-core and, by the method of Berger & Herguera (1992), indicates an increase of 60 % in the productivity from glacial to Holocene (Manighetti & McCave 1995*a*). However the more elaborate procedures of Sarnthein *et al.* (1992) give about the same glacial and Holocene export productivity of 7 gC m^{-2} a^{-1}. Size analysis, carbonate content and dating give the flux of < 10 μm carbonate, more or less equivalent to late Quaternary coccolith flux (core 6K at 2865 m is well above the calcite lysocline). The coccolith flux also increases by 60 % (from 0.33–0.53 g cm^{-2} ka^{-1}) from glacial to Holocene supporting the idea of increased productivity. The carbonate–carbon burial rate is 0.98 gC m^{-2} a^{-1} for the Holocene and 0.71 gC m^{-2} a^{-1} for the glacial (of which some is reworked). The carbonate dominates the sequestration of carbon in the sea bed as the burial values for organic carbon are 0.1 gC m^{-2} a^{-1} for the glacial and 0.04 for the Holocene. This carbon burial

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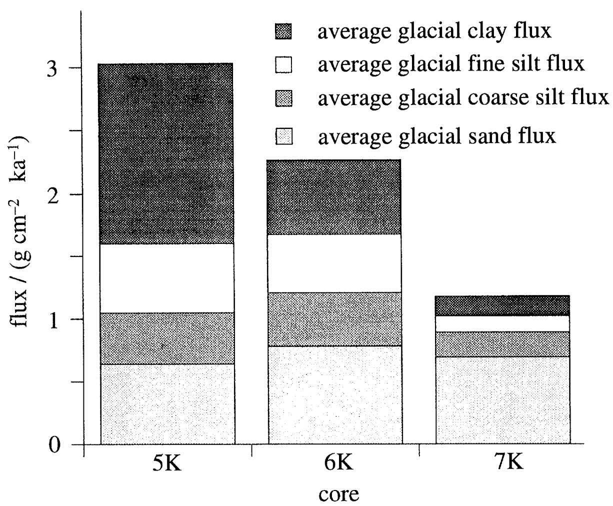

Figure 6. Average glacial flux (20–30 Ka) of non-carbonate material in different size classes in three cores from the East Thulean Rise. Lower flux in 7K (ridge-top location, see figure 7) and higher flux in 5K (base of slope location) is a result of redistribution of the fine fraction, while the coarse faction (> 63 μm) remains unaffected. Core 6K has experienced neither net winnowing nor focusing due to its position on a gentle local high. (From Manighetti & McCave 1995*a*.)

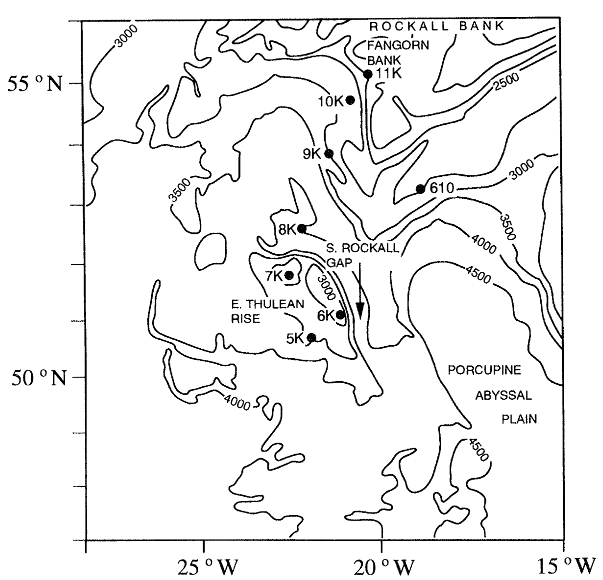

Figure 7. Location of Kasten cores taken in 1989 from *RRS Discovery*.

rate, of order 1 gC m^{-2} a^{-1}, is a rather small fraction of the export productivity for the region. If applied to the 2.6×10^6 km^2 of the N.E. Atlantic from 45°–63° W and 20°–30° N, the annual storage in the sea bed is 2.6 MtC a^{-1}.

(*c*) *Comparison with a standard to estimate flow speed changes*

One method of getting at the problem of examining properties of current-focused sediments in the presence of direct input by vertical flux is to subtract the signature due to vertical flux. This has been done by Manighetti & McCave (1995*b*) for the current-controlled sediments in South Rockall Gap (figure 7). The flux in nearby core BOFS 6K is considered to represent unmodified pelagic input. Following the arguments presented above on sorting, the mean size of the silt coarser than 10 μm (the 'sortable silt' of McCave *et al.* 1995) is used as an index of current speed. The size of the unmodified pelagic material is subtracted from the record of size in current-affected cores BOFS 8K and 10K (figure 8). Positive deviations are interpreted as stronger currents removing finer material and allowing deposition of coarser silt.

Noteworthy features of the time-series for cores 8K and 10K which presently lie beneath Lower Deep Water (LDW of Schmitz & McCartney 1993) from the south and Lower N. Atlantic Deep Water (LNADW) respectively are the in-phase fluctuations in flow speed from the Last Glacial Maximum (LGM) through to Younger Dryas. The low flow speeds prior to the LGM are probably due to small production of NADW and a large volume of slow-flowing LDW bathing both sites. The striking rise in flow speed of either LDW or LDW and NADW together between the LGM and start of the glacial termination IA indicates a significant increase in the vigour of the North Atlantic Circulation, possibly caused by an increase in NADW production.

A comparison which adds some weight to the inference of increasing flow speed after the LGM is the simple grain size record of Haskell *et al.* (1991) at 2950–3818 m on Blake Outer Ridge (BOR) (figure 8). This shows similar features – low glacial flow, pre IA peak, IA sharp decline and Bölling-Alleröd speed up – as on BOFS 8K and 10K. There can be no provenance effect linking the records, and no input function subtraction was necessary on the BOR stack because the area is south of glacial influence, yet it shows the same pattern. This is strong evidence for current speed control of the deposition patterns.

(*d*) *Joint occurrence of properties*

It was noticed by Wang & McCave (1990) that the coarseness of terrigenous and biogenic silt varied in the same way: both were coarser during mid-Pleistocene glacial periods but were not exactly in phase with ice-rafted input shown by the percentage of terrigenous sand. As the two silt components are not linked by any source function, their covariance was ascribed to an environmental variable, namely current speed. This approach was taken further by Robinson & McCave (1995) who examined covariances of both percentage composition and flux of components as well as their grain size. It is argued that covariance of the percentage of terrigenous and carbonate silt > 10 μm in diameter provides an index of current strength. Cross-spectral comparison of the time series for the two percentages plotted in standard deviation units show large amplitude and coherency at the principal Milankovitch frequencies. The two time series are then summed, which gives a strong signal where the two are in phase, and spectrally compared with the ice volume model of Shackleton *et al.* (1990) (figure 9). Again there is strong coherent amplitude at the principal Milankovitch periods (125, 41, 23 ka). It might be argued that the amount of coarse terrigenous silt is due

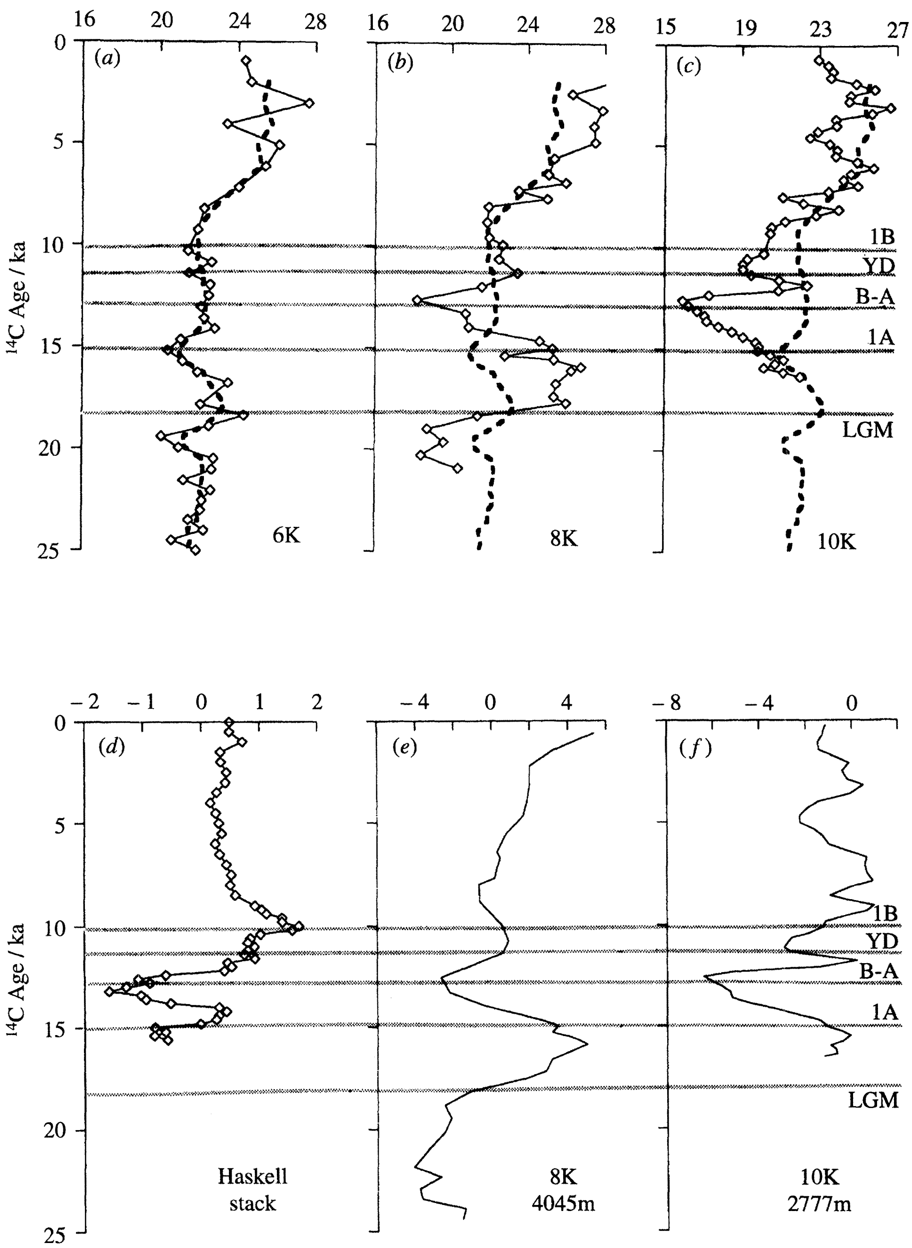

Figure 8. (*a*) Mean size of the sortable silt fraction (10–63 μm non-carbonate) at BOFS 6K over the past 30 ka. The smoothed profile is plotted with each of the other core records to show deviation from a standard (non current affected) input. (*b*) Core 8K shows a marked difference between the LGM and the Younger Dryas and striking differences are also evident in (*c*), the profile of core 10K. These differences are attributed to bottom currents. (*d,e*) Sortable silt mean in cores 8K and 10K expressed as deviation, from the 6K input function (from Manighetti & McCave 1995*b*). (*f*) Mean size of the 6–63 μm non-carbonate fraction against ^{14}C age averaged for three cores (CH84–7P, -10P and -11P) from Blake Outer Ridge using data from Haskell *et al.* (1991). Although the start of Termination IA marked here is slightly younger than that observed in BOFS core 5K (Manighetti *et al.* 1995), the plot resembles strongly those of cores 8K and 10.

to ice-rafted delivery and the amount of coarse carbonate silt is related to abundance of forams. A refinement is therefore to subtract the time series of terrigenous sand from silt and that of foram abundance from carbonate silt, yielding non-IRD terrigenous silt and non-foram carbonate silt time series. These also show peak amplitude and coherency at 125, 41, 23 and 19 ka periods, and the sum of these time series is also coherent with the ice volume model at Milankovitch periods. Finally, the mean size of the coarse (> 10 μm) silt is strongly coherent with the percentage of terrigenous silt at Milankovitch periods. The size and compositional properties of the different components of the system are strongly orbitally forced. All the indices suggest that bottom currents were stronger during mid-Pleistocene glacial stages than interglacials at DSDP Site 610 on Feni Drift at 2500 m depth. This is a somewhat surprizing result in light of detailed late Pleistocene results of Manighetti & McCave (1995*b*) and present views of NADW reduction and increase in LDW volume (but not flow speed) in this area during the last glacial maximum. However site 610 lies on Feni Drift in Rockall Trough and it may have been constructed during glacials by what Zimmerman (1982) termed the 'Feni glacial current' of overflow water flowing down Rockall Trough which delivered a

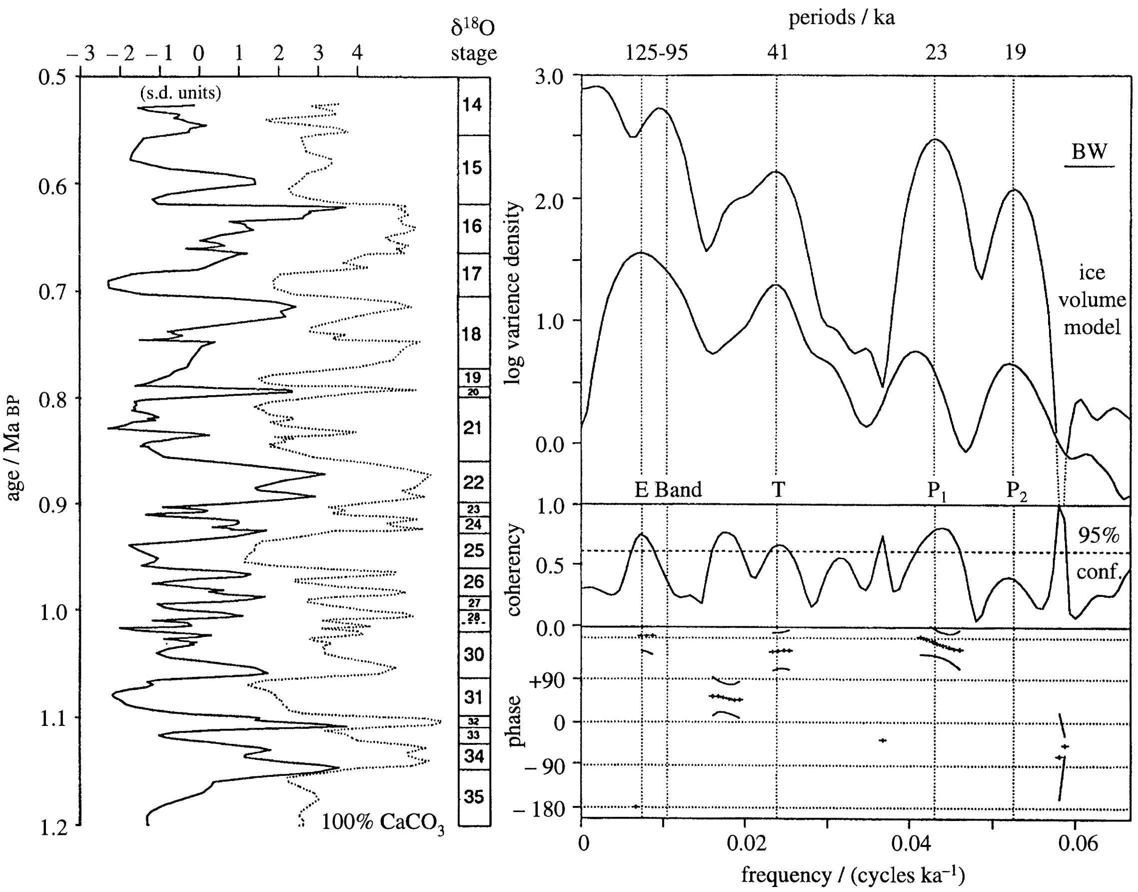

Figure 9. Contourite coarse silt enrichment index (sum of the time series of percent > 10 μm in the biogenic and lithogenic subfractions, each subfraction being scaled-up to 100 %) (data in standard deviation units around a zeroed mean): its relation to variations in % non-carbonate and δ^{18}0 Stages (left), and cross-spectral comparisons with ice volume model (right). From Robinson & McCave (1994).

rich clay load compared with the Holocene. The spectral comparison method implicitly classifies changes as gradual and not as the abrupt shifts seen in the late Pleistocene to Holocene transition.

I am most grateful to Barbara Manighetti, John Thomson, Simon Robinson, Brian Haskell and their co-authors on whose data a significant part of this paper is based. Much of this data collection was made possible by NERC grants for BOFS. I thank Dudley Simmons for his work on figure 4. Cambridge Earth Sciences Contribution No. 4082.

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Discussion

P.M. Holligan (Plymouth Marine Laboratory, U.K.). Professor McCave inferred that the fine (< 63 μm) carbonate fraction from glacial and interglacial sediments in the N.E. Atlantic is a measure of relative biological productivity (coccoliths, small foraminifera, etc). The results of McIntyre *et al.* (1972) and my own unpublished observations on cores from the Rockall Trough and from northern sites on the JGOFS 20° W line indicate that coccolithophore production was extremely low during the glacial period. The calcite is largely very fine crytalline material, with few recognizable biogenic particles. Is it possible that this material is of secondary origin, perhaps transported from the continental shelf to the ocean when sea level is relatively low?

Reference

McIntyre, A., Ruddiman, W.F. & Jantzen, R. 1972 Southward penetrations of the North Atlantic Polar Front: faunal and floral evidence of large-scale surface water mass movements over the last 225000 years. Deep Sea Res. 19, 61–77.

I.N. McCave. Actually I used the < 10 μm carbonate fraction. The recent results of Dr P.P.E. Weaver (personal communication) on cores from the N.E. Atlantic suggest that no more than 6% by number of the coccoliths are reworked Cretaceous species during the glacial. However, these are much larger than the late Pleistocene varieties and may contribute over 30% of the $CaCO_3$ (they are > 10 μm and late Pleistocene ones are < 5 μm in the main), so that I think our use of < 10 μm carbonate has some claim to be related to late glacial and Holocene coccoliths. However, this is not the case in Heinrich layers which do contain a large percentage of fine crystalline carbonate, probably from the lower Palaeozoic (Andrews & Tedesco 1992). One must be careful in sampling to establish what type of deposit is being interpreted.

Reference

Andrews, J.T. & Tedesco, K. 1992 Detrital carbonate-rich sediments, northwestern Labrador Sea: implications for ice-sheet dynamics and iceberg rafting (Heinrich) events in the North Atlantic. *Geology* **20** 1087–1090.

F.A. Street-Perrot (*Environmental Change Unit, University of Oxford, U.K.*). Is it correct that the impact on bioturbation of the mudblanket deposited during a Heinrich event results in a perturbation of the ^{14}C chronology such that it is only possible to get wide bracketing dates for these events using the radiocarbon mehtod?

I.N. McCave. Yes, as long as the sediment above and below the event layer is bioturbated and no mixing occurs across it, as can be seen in my X-radiographs. In that case the layer below will look like a preserved seabed mixed layer with an age of a few thousand years, and the layer above the event will not have older material burrowed in to offset the younger components mixed down. In ODP core 609 Bond *et al.* (1993, *Nature* **365**, 143–147) give AMS dates for events H-1 and H-2 with spreads of about 2400 and 3600 years, much shorter than the duration modelled by Dowdeswell *et al.* (1995).

L. Labeyrie. (*CFR, Domaine du CNRS*, 91198 *Gif/Yvette, France*). About the duration of the Heinrich Layers: Roger Francois used ^{230}Th fluxes within, below, above the events, and estimated dilution by the ice-rafted detritus, yielding approximately 500 years duration for the events.

I.N. McCave. That duration is at the upper end of the age spread, modelled by Dowdeswell *et al.* (1995) to be 50–500 years.

M. Sarnthein. Did Professor McCave consider the small equivalent diameter of forams forming ripples (that are actually silt ripples)?

I.N. McCave Yes I did. The critical erosion condition (which I used) is defined in the non-dimensional framework of Shields (θ-Re_b) or Yalin ($\theta - \Xi$) (Miller *et al.* 1977), applied to foraminifera with density $\rho_s = 1500$ km m^{-3} by Miller & Komar (1977). In this, forams of 250 μm behave like coarse quartz silt on erosion but have settling velocity of fine quartz sand and thus undergo very little suspension transport (τ_{ocrit} 250 μm quartz = 0.20 Pa, forams = 0.067 Pa at abyssal 2 °C). (Definitions $\theta = \tau_o/\Delta\rho g d$, $Re_b = u{*}d/\nu$, $\Xi = (\Delta\rho g d^3/\rho\nu^2)$ in which τ_o = bed shear stress applied by a current, $\Delta\rho = \rho_s - \rho$ where ρ, ρ_s are densities of water and sediment, shear velocity $u{*} = (\tau_o/\rho)^{\frac{1}{2}}$, d is sediment grain diameter, ν is kinematic viscosity, and g is acceleration due to gravity.)

Reference

Miller, M.C., McCave, I.N. & Komar, P.D. 1977 *Sedimentology*, **24**, 507–527.

M. Sarnthein. Professor McCave showed a different variability of coccolith and C_{org} accumulation rates. Could this difference be the result of a significant change in the rain ratio between carbonate particles and particulate organ carbon?

I.N. McCave. Yes it could. What I showed was that the coccolith flux increased by 60 percent and the productivity estimated by Berger & Herguera's method also increased by 60% whereas the estimate via the equation of Sarnthein *et al.* showed no change. The suggestion that there was an increase in productivity is corroborated by data on benthic foraminiferal populations (Thomas 1995). However, I accept that the exact agreement of two estimates based on carbonate flux and carbon percentage is probably fortuitous.

Reference

Thomas, E., Booth, L., Maslin, M. & Shackleton, N.J. 1995 Northeastern Atlantic benthic foraminifera during the last 40 ka. *Paleoceanography*, **10**. (In the press.)

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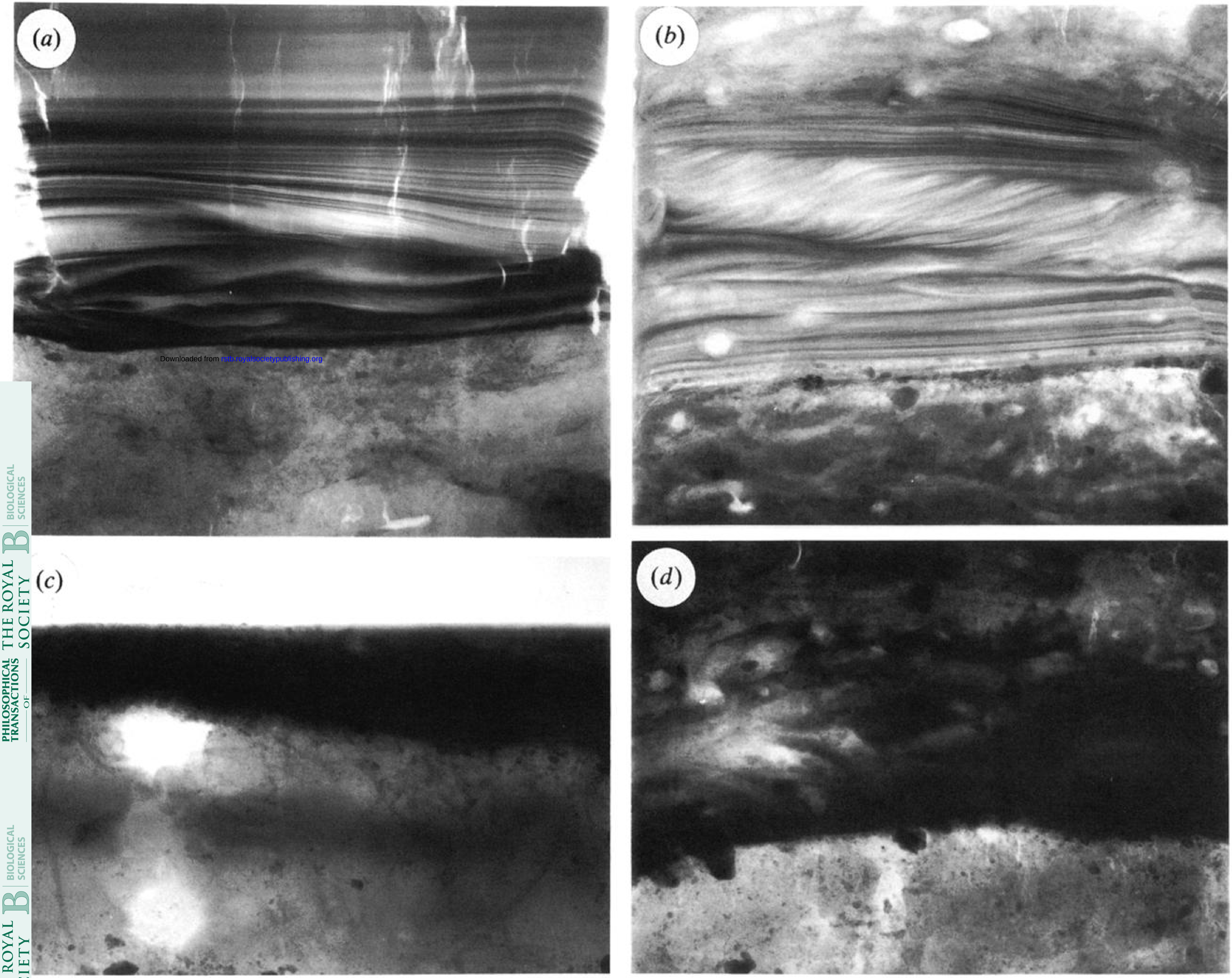

Figure 4. Photographic prints of X-radiographs of cores. Turbidites in cores (*a*) BOFS 15K, and (*b*) BOFS 2K. Possible preservation of fine burrows beneath the turbidites is visible in (*a*) and (*b*), and onset of burrowing can be seen in the top of (*b*) which also shows superb climbing ripple structure. (*c*) Heinrich layers with preserved fine burrow structures beneath BOFS 8K (H-2). (*d*) Burrows in the top of the layer BOFS 5K (H-2) which do not disturb its lower interface.

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