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The Benthic Geochemical Record of Late Holocene Carbon Flux in the Northeast Atlantic [and Discussion]

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The benthic geochemical record of late Holocene carbon flux in the northeast Atlantic

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SUMMARY

This study centered around a transect of high-resolution (multi)cores from the 20° W meridian, 60–18° N in the northeast Atlantic. It spans a range of primary productivity zones, and was used to quantify and examine the vertical flux of organic carbon from the euphotic zone (50 m deep) to the sediment–water interface, through the sediment mixed layer, to burial in late Holocene sediment. The disequilibrium between members of the natural uranium decay series (^{226}Ra, ^{210}Pb and ^{210}Po) – which track the biogenic flux through scavenging of the particle-reactive nuclides – was employed. Together with experimentally and observationally derived factors, these data were used to convert nuclide flux to organic carbon flux resulting in an estimate of the water column flux of organic carbon. At the sediment–water interface micro-oxygen electrodes were used to quantify the consumption of organic carbon by aerobic respiration. It was noted that the estimated organic carbon flux is strongly dependent on the intensity of bioturbation. The late Holocene organic carbon burial flux was calculated for selected cores from measured organic carbon profiles and sediment accumulation rates over approximately the last 10000 years. This combined approach reveals a strong spatial and temporal signal in the flux of organic carbon through the deep sea in the northeast Atlantic, and provides additional insight into the fate of carbon in this area of the ocean.

1. INTRODUCTION

Recent evidence for the rapid transport of biogenic detritus to the deep-sea floor of the northeast Atlantic (during seasonal phytoplankton blooms) has rekindled interest in the geochemical dynamics of the benthic interface. Uncertainties in our knowledge of the spatial and temporal flux of carbon to, and through, the biologically active sediment–water interface is an important factor in models of the global carbon cycle. Despite a lack of detailed knowledge of these processes at the benthic interface, palaeoceanographers still aspire to unravel the history of primary and new production in the world's ocean from the remaining vestiges of the vertical carbon flux.

The North Atlantic is characterized by a strong seasonal variation in phytoplankton biomass which exerts a considerable impact on the oceanic carbon cycle (see Ducklow & Harris 1993). In 1989 and 1990 several international expeditions, under the framework of the Joint Global Ocean Flux Study (JGOFS) conducted a series of process studies along the 20° W meridian, centered on 47° N. These studies evaluated the impact of community structure and metabolism on carbon and nutrient cycling, CO_2 draw-down, and the vertical flux of carbon from the euphotic zone to the seafloor. Concurrently, deep-sea sediment cores were taken from the transect to examine the benthic record of modern fluxes and the palaeoceanographic record. The primary aim of these studies was to quantify the standing stock of organic carbon in the different trophic levels and ecosystems in the northeast Atlantic and the resulting flux to the seafloor.

Because of the nature of the techniques and observations used to examine these processes, all that is possible is a 'snapshot' of the intra- and interannual variability in organic carbon flux. Under the detailed JGOFS time-series protocols, shipboard primary productivity incubations are performed at 24 h intervals throughout the spring and early summer period. In contrast, the deep-sea sediment record integrates carbon fluxes over several centuries. In addition, the range in latitude covered along the 20° W transect spans oligotrophic to mesotrophic–eutrophic conditions occurring in the central north Atlantic gyre and the northwest African upwelling regime, respectively. Our objective is to couple the range of flux measurements (made during the North Atlantic Bloom Experiment (NABE) over a range of timescales) with the longitudinal productivity gradient. In this way, the observations completed during the NABE can be placed in a wider temporal and spatial framework, creating an improved knowledge of the flow of organic carbon through the north Atlantic.

In this paper we have used a variety of techniques to examine the sub-euphotic zone, benthic and Holocene burial flux of organic carbon along the 20° W transect. Comparisons were made between the sedimentary oxygen consumption rate, benthic carbon flux and bioturbation rate to evaluate the role of the benthos in converting the incoming carbon flux to the burial (palaeoceanographic) carbon flux. Despite the large

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vertical and horizontal scales examined in this study, it is apparent that the magnitude of primary productivity, and resulting export flux, plays a significant role in determining the benthic flux and burial of organic matter in the northeast Atlantic.

2. RESULTS AND DISCUSSION

All the primary data used in this study can be found in the British Oceanographic Data Centre, Bidston and is published on CD-ROM (Lowry *et al.* 1994). The sites selected occur at approximately 5° intervals from 60–18° N in a variety of water depths (see table 1). Radiochemical, sedimentary dissolved oxygen and organic carbon measurements were made at Edinburgh using previously published methods (Brand & Shimmield 1992; Ritchie & Shimmield 1992). The modelling of carbon flux from these measurements is given below. Ancillary data on ^{234}Th-based carbon flux (Buesseler *et al.* 1992), floating sediment trap (Martin *et al.* 1993) and moored sediment trap (Newton *et al.* 1995) carbon fluxes, and sediment accumulation rates (Manighetti 1995; Thomson *et al.* 1993*b*) were used.

(*a*) *Estimation of export carbon flux from ^{210}Po–^{210}Pb disequilibria*

Dissolved and particulate ^{210}Po and ^{210}Pb were combined to give total activities in disintegrations per minute (dpm) per 100 litres seawater. These were compared to the activity of the ^{210}Pb grandparent, ^{226}Ra, estimated from the dissolved silicate distribution using the relation given by Rhein *et al.* (1987). Due to the particle-reactive nature of both ^{210}Pb and ^{210}Po, specific activities in seawater are often less than their progenitor, ^{226}Ra (except near the surface where atmospheric input of ^{210}Pb is significant). In addition, the bio-reactive ^{210}Po is often deficient relative to ^{210}Pb in the euphotic zone due to phyto- and zooplankton uptake, but increases in the dissolved phase with depth due to subthermocline regeneration. Thus, the ^{210}Po (half-life 138.4 days) residence time in the euphotic zone is several months, whereas for ^{210}Pb (half-life 22.3 years) it is 2–3 years (Bacon *et al.* 1976). The extent of the observed disequilibrium between ^{210}Po and ^{210}Pb in the mixed layer can be used to estimate removal rate of the daughter nuclide from the water column by sinking particles. A simple irreversible scavenging model is used which includes both the removal of the dissolved nuclide onto particles, and their subsequent sinking from the mixed layer; because most of the ^{210}Po is in the dissolved phase, particulate sinking is the rate-dominating step (Kadko 1993). Using Bacon *et al.*'s 1976 procedure, inventories of ^{226}Ra, ^{210}Pb and ^{210}Po were calculated for a 50 m mixed layer at each station along the transect, based on samples from three depths within the mixed layer. (Although there is a small latitudinal variation in mixed layer thickness, the mean depth is used here). Stations north of 47° N were sampled in August–September 1989; those south of 47° N in September–October 1990. At this time, the mixed layer depths had stabilized to a large extent with moderate stratification of the water column. For the bio-reactive ^{210}Po, at steady state, the production from ^{210}Pb decay is balanced by loss from radioactive decay and scavenging:

$$\lambda_{Po} A_{Pb} = \lambda_{Po} A_{Po} + J_{we}, \quad (1)$$

where A_{Po} is the integrated activity in the mixed layer (dpm cm^{-2}), λ_{Po} is the ^{210}Po decay constant (1.83 a^{-1}) and J_{we} is the ^{210}Po scavenging rate by particles in the water column.

The conversion of ^{210}Po scavenging flux to organic carbon flux requires knowledge of the ^{210}Po: organic carbon ratio in sinking material immediately beneath the mixed layer. As this ratio was not available we have used experimentally derived concentration factors of radionuclides in phytoplankton (Fisher *et al.* 1988) for our export production estimates:

$$P_n = F/(4DC_{dm}), \quad (2)$$

where P_n is the amount of new (export) production in g C m^{-2} a^{-1}; F is the radionuclide flux in

Table 1. *Locations and statistics of stations used in this paper*

(Sediment cores were collected using the SMBA Multi-corer. Water samples for ^{210}Po-^{210}Pb measurements were collected using 20 l samples collected by Go-Flo bottles on the CTD. Station numbers refer to BOFS station number list and may be accessed on the BOFS database (Lowry *et al.* 1994).)

BOFS station code no.	latitude	longitude	date of collection	water depth / m	primary productivity[a] / gC m^{-2} a^{-1}
15M	59°04.9′	20°05.8′	04/08/89	2790	105
11M	55°11.3′	20°21.0′	01/08/89	2080	85
5M	50°40.4′	21°51.6′	27/07/89	3560	60
1M	47°09.6′	22°30.2′	20/07/89	3945	55
22M	32°52.8′	29°12.9′	30/09/90	3780	45
23M	32°31.7′	20°23.4′	2/10/90	4650	45
26M	24°28.7′	19°53.5′	7/10/90	3660	60
28M	24°34.2′	22°49.3′	10/10/90	4855	60
29M	20°32.3′	21°06.9′	13/10/90	4000	125
31M	18°59.6′	20°09.2′	17/10/90	3295	135

[a] Berger (1989).

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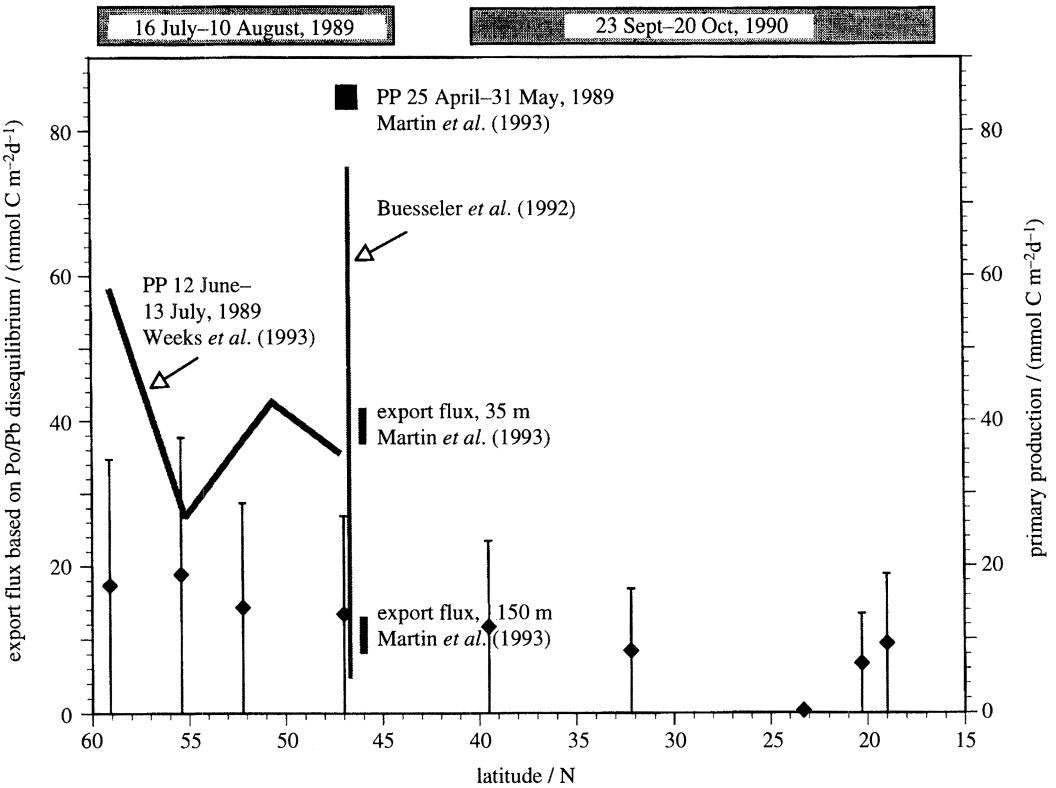

Figure 1. The export flux of organic carbon (mmol C m^{-2} d^{-1}) estimated from the ^{210}Po-^{210}Pb scavenging model, and a nuclide:carbon conversion factor (Fisher *et al.* 1988) plotted against latitude along the 20° W transect. These measurements were made in the late summer of 1989 and 1990 (see legend at top of graph). The primary production at the NABE site (47° N) at the time of sediment trap deployment (86 mmol m^{-2} d^{-1}) is marked by the filled square, whereas the latitudinal range in primary production between 47–60° N measured the following month by Weeks *et al.* (1993) is marked by the solid line. At 47° N the range in ^{234}Th -estimated organic carbon fluxes from the NABE experiment (Buesseler *et al.* 1992) and the carbon flux collected in floating sediment traps at 35 m and 150 m depth (Martin *et al.* 1993) is shown (all vertical lines indicate the range in carbon flux).

10^{-3} dpm m^{-2} a^{-1} (equivalent to J_{wc} above); D is the dissolved radionuclide concentration (dpm l^{-1} of surface water; C_{dm} is the dry mass concentration factor of a radionuclide in phytoplankton (dpm kg^{-1}/dpm l^{-1}, with the factor 4 being the dry mass:carbon ratio in cells). For ^{210}Po the experimentally derived mean C_{dm} in phytoplankton is 2.5×10^5.

The result of these calculations for each station along the transect is plotted in figure 1 as the organic carbon export flux (mmol C m^{-2} d^{-1}). The use of the concentration factor, C_{dm}, introduces considerable uncertainty into the estimation of carbon flux, particularly as the transect spans a range of biogeochemical provinces. However, Fisher *et al.* (1988) point out that the estimated concentration factors for the radionuclides in planktonic debris is within a factor of two of the measurements in zooplankton faecal pellets. This uncertainty, together with the error associated with the measurement of nuclide inventory within the mixed layer, is shown by the vertical error bar in figure 2. These flux estimates apply to conditions found in the late summer (August–September–October), after the main bloom event. No primary productivity measurements are available from the same cruise, however figure 1 shows how the primary productivity had decreased from 86 mmol C m^{-2} d^{-1} in late April–May 1989 (Martin *et al.* 1993) at 47° N to 36 mmol C m^{-2} d^{-1} in early July 1989 (Weeks *et al.* 1993). It is likely that the primary productivity was somewhat lower at the time of the radionuclide sampling at 47° N (early August, 1989). Estimates of the sinking flux of organic carbon at 35 m and 150 m depth at 47° N in 1989 (April–June) have been obtained by Martin *et al.* (1993). The flux at 150 m depth (*ca.* 10 mmol C m^{-2} d^{-1}) is similar to that estimated from the radionuclide scavenging model. At the southern extremity of the transect, upwelling processes advect dissolved ^{210}Po and ^{210}Pb into the upper mixed layer. Because ^{210}Po is regenerated more rapidly that ^{210}Pb in thermocline and sub-thermocline waters, the estimate of vertical nuclide flux from equation 1 has a large error. A detailed analysis of the impact of upwelling and aeolian dust input on ^{210}Pb and ^{210}Po is in preparation (G. Shimmield & G. Ritchie, unpublished data) Despite the uncertainties associated with the radionuclide:organic carbon conversion factor, and the application of a simple steady-state model, we believe that the approach used here provides a useful assessment of the magnitude of the organic carbon export flux at 50 m depth in the late summer for the northeast Atlantic.

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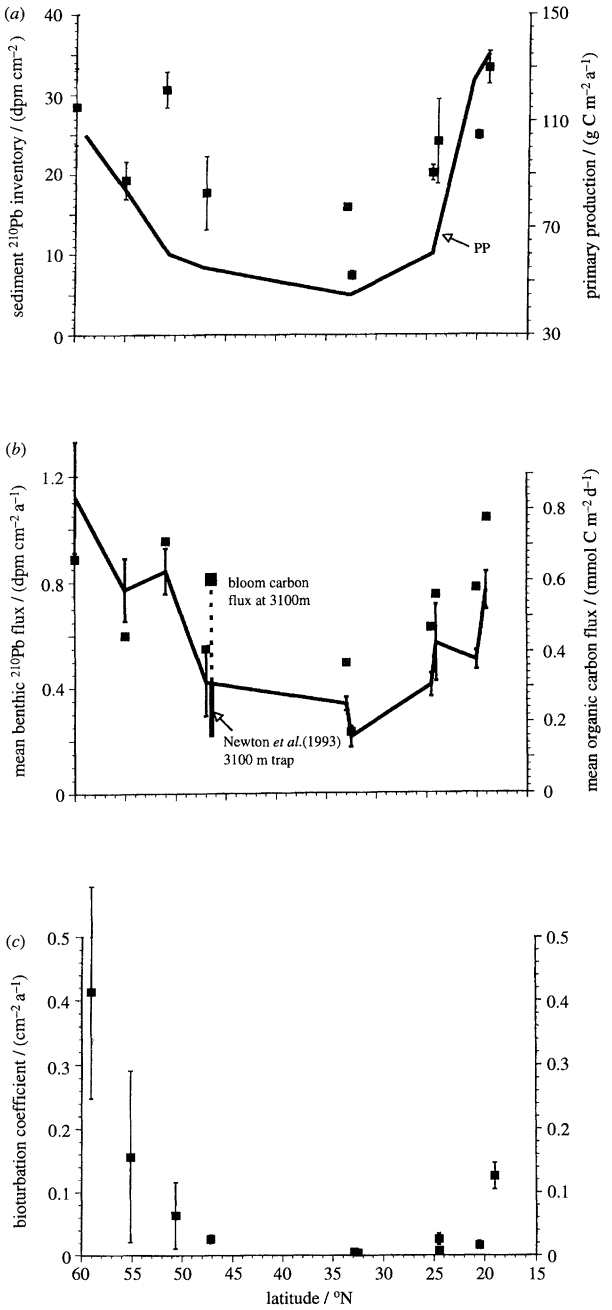

Figure 2. (*a*) Sedimentary excess ^{210}Pb inventories (dpm cm^{-2}) and primary production (Berger *et al.* 1988) along the 20° W transect. Vertical range bars on the inventories represent the inter-core variation from a single multi-core at each particular site. Counting statistics and analytical uncertainties are within the dimensions of the symbol. (*b*) Mean excess ^{210}Pb flux (filled squares: dpm cm^{-2} a^{-1}) and estimated benthic carbon flux (solid line: mmol C m^{-2} d^{-1}) along the 20 °W transect. These flux estimates are derived from the sedimentary ^{210}Pb inventory and represent the average flux over the last 100 years. Vertical range bars on the benthic carbon flux profile represent the inter-core variation from a single multi-core at each particular site. The benthic carbon flux estimate has an uncertainty of ±20%. Also shown is the measured POC flux at the 3100 m trap at 47 °N (Newton *et al.* 1995). The solid vertical line is the non-bloom annual range, with the peak bloom flux connected by the dashed line. (*c*) Calculated bioturbation rates (black squares: see text for details) along the 20° W transect. The vertical range bars on the bioturbation coefficient represent the inter-core variation from a single multi-core at each particular site.

(*b*) *Sedimentary ^{210}Pb inventories and the estimation of benthic carbon flux*

Although the upper water column ^{210}Po:^{210}Pb disequilibrium provides information on vertical fluxes on the weekly to monthly timescale, the quantity of ^{210}Pb removed in the deep ocean, relative to it's progenitor, ^{226}Ra, can be used to examine fluxes on the 100-year timescale. With a half-life of 22.3 years, the inventory of excess ^{210}Pb, and the mean flux required to support it, provides information on scavenging processes averaged over 4–5 half-lives.

If the principal transport mechanism for excess ^{210}Pb to the sediment in the open ocean is the vertical biogenic flux, it is likely that there is a relation between scavenging efficiency and organic carbon flux. This relation was explored by Moore & Dymond (1988) who showed that there is indeed a correlation between ^{210}Pb removal and carbon flux in the Pacific Ocean. They developed the following algorithm using sediment trap data:

$$Flux\ C_{org}\ (\mu g\ cm^{-2}\ y^{-1}) = 8.5 + 430(F/P), \quad (3)$$

where F/P is the ratio of measured ^{210}Pb flux to ^{210}Pb production estimated from the standing crop of its progenitor, ^{226}Ra (including atmospheric ^{210}Pb deposition). Using this relation they were able to accurately predict the organic carbon flux in the Sargasso Sea to within 12%.

With a half-life of 22 years, excess ^{210}Pb in the sediment has an average lifetime of 100 years. By measuring the inventory of excess ^{210}Pb in the sediment, the mean benthic flux of ^{210}Pb (J_b, dpm cm^{-2} a^{-1})over the last century may be calculated:

$$J = \lambda_{210} I, \quad (4)$$

where the inventory I (dpm cm^{-2}) has been estimated from:

$$I = \Sigma(\rho_i A_i z_i) \quad (5)$$

(ρ_i, is the sediment slice dry bulk density or dry material/unit wet sediment, g cm^{-3}; A_i is the slice activity in dpm g^{-1}; and z_i is the slice thickness in cm). In figure 2*a* the variation with latitude of the excess ^{210}Pb inventory is shown. Clearly, sediment redistribution processes can be important in affecting the measured inventory (Thomson *et al.* 1993), but we have taken some care in excluding any cores that show sedimentological attributes of sediment focusing or winnowing. At some locations several cores from the same multi-core array were analysed. This gives an indication of the intra-site variability in inventory; it is shown by the vertical range bars in figure 2*a*. The error associated with counting statistics and analytical uncertainty is always less than this (see Brand & Shimmield 1991).

The mean benthic flux of excess ^{210}Pb, J_b, has been normalized to the water column production flux from ^{226}Ra (to correct for water depth effect) and used in equation 5. The benthic carbon flux, is substituted for F and the ^{226}Ra production flux for P. The resulting estimate of organic carbon flux to the sediment, averaged over the last century, is termed the 'benthic flux'. Using the Moore & Dymond (1988) relation,

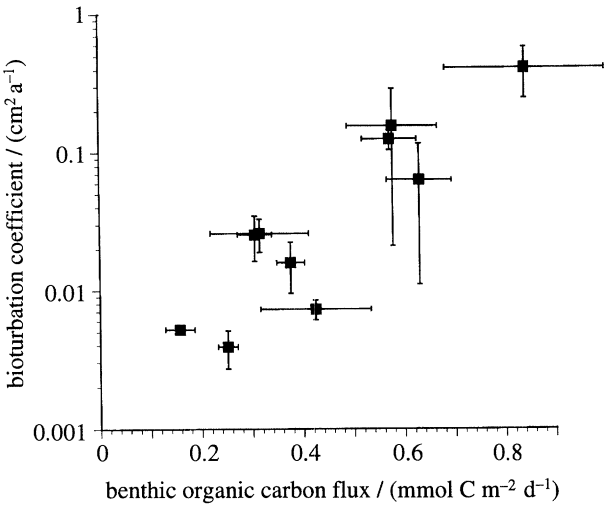

Figure 3. The positive relation between bioturbation coefficient ($cm^{-2}\ a^{-1}$) and benthic organic carbon flux (mmol C $m^{-2}\ d^{-1}$; estimated from the mean excess ^{210}Pb flux and the relationship of Moore & Dymond (1988)), suggesting that the activity of the benthos increases in proportion to the food supply. Range bars represent the intercore variation from a single multi-core at each particular site. Note the bioturbation coefficient is plotted on a log scale.

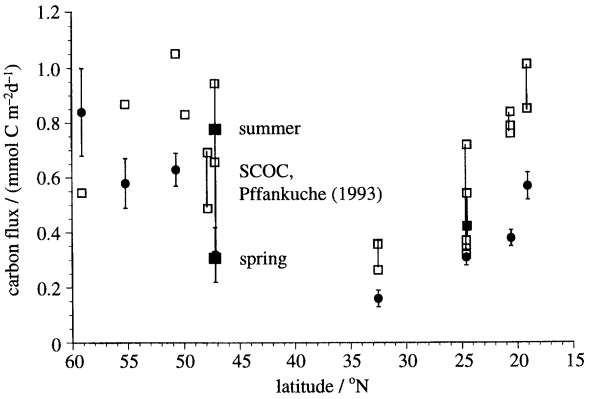

Figure 4. Oxygen consumption converted to organic carbon flux (see text for details) and estimated benthic organic carbon flux (both in mmol C $m^{-2}\ d^{-1}$) for the 20° W transect. Note that the oxygen consumption flux nearly always exceeds the benthic organic carbon flux (estimated from ^{210}Pb), at any given site. Vertical lines connecting the oxygen consumption fluxes indicate the range obtained from several cores on the same multi-core drop; open squares refer to oxygen consumption; filled circles refer to benthic carbon flux.

together with the radionuclide counting and analytical error propagation, introduces an uncertainty of $\pm 20\,\%$, whereas the variation in inventory between cores from the same multicore drop may be larger than this (Brand & Shimmield 1991; Thomson *et al.* 1993). In figure 2*b* the mean excess ^{210}Pb flux and derived benthic organic carbon flux (mmol C $m^{-2}\ d^{-1}$) are plotted with latitude to display the clear primary productivity dependence. To assess the appropriateness of this estimate, the sediment trap data from Newton *et al.* (1995), is shown for the 47° N site. The year-round organic carbon flux average for the 3100 m trap is very similar to the long-term organic carbon flux average estimated from the ^{210}Pb flux model, with the peak bloom detritus flux being somewhat greater.

One consequence of the organic carbon benthic flux variation is changes in faunal benthic activity. As well as recording the sedimentary inventory, ^{210}Pb also tracks particle mixing processes. We have employed the standard approximation of the benthic mixing process to a dispersion process (Fick's second law of diffusion) and calculated bioturbation rate constants for all the cores (see figure 2*c*). Along the transect there is a clear trend from lumpy, 'conveyor belt' mixing in the north, to more 'diffusive-like' mixing in the south (Shimmield & Brand 1995). The simple mixing model applied here is not sufficient to resolve the depth dependant detail, but serves to indicate the relation between carbon flux and bioturbation rate (see figure 3). We believe this is one of the clearest examples of mixing rate dependence on food supply.

(*c*) *Interstitial dissolved oxygen consumption and carbon flux*

The benthic sediment–water interface is a dynamic environment where substantial remineralization of organic matter takes place. This remineralization takes place with a range of oxidants that are consumed progressively in order of their free energy yield. Bottom water oxygen content is an important quantity in determining the extent of organic matter degradation in many pelagic settings (Jahnke *et al.* 1989). Recently, Pfannkuche (1993) showed that the oxygen demand sediment community oxygen consumption (SCOC) at a single site may vary seasonally according to the supply of organic matter from the overlying water column. We measured dissolved oxygen profiles, using microelectrodes at millimetre scale resolution, to estimate the rate of oxygen consumption at the benthic interface. Several profiles from a single multi-corer tube, and several tubes from the same core, were analysed at *in situ* temperatures and atmospheric pressure on board ship. This methodology may impose an artefact on the measurements, (Jahnke *et al.* 1989). However, we have adopted the procedure of Jahnke *et al.* to estimate the oxygen consumption gradient over the upper 7 mm in an effort to avoid the major contamination effects of oxygen diffusion from the interface preceding and during sampling, (4–5 mm in 3 h), and 2 mm of physical disturbance. The resulting oxygen gradient has been calculated using Fick's first law of diffusion applied to sediments:

$$F = -\phi_o D_o (dC/dZ) \qquad (6)$$

where F is the oxygen flux ($\mu mol\ m^{-2}\ d^{-1}$), ϕ_o is the porosity at the interface (volume pore water–volume bulk sediment; $cm^3\ cm^{-3}$), D_o is the bulk sediment diffusion coefficient for oxygen at the sediment–water interface = 0.85 $cm^2\ d^{-1}$, (dC/dZ) is the oxygen concentration gradient ($\mu mol\ cm^{-3}\ cm^{-1}$) over the upper 7 mm. The resulting oxygen flux calculated has been converted to organic carbon flux using the Redfield relation (138 moles O_2 consumed for every 106 moles of organic carbon).

In figure 4 the calculated flux of organic carbon consumed by oxic degradation has been plotted against latitude, in conjunction with the benthic organic

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Table 2. *Average and range of Holocene organic carbon accumulation rates in selected cores from the BOFS transect*

(Sedimentation rates for the late Holocene are based on age models derived by Manighetti *et al.* (1995) based on oxygen isotope stratigraphy and mineral magnetics.)

BOFS station code no.	mixed layer depth	average holocene sedimentation rate	average organic carbon accumulation rate
	cm	cm k^{-1}	μmol m^{-2} d^{-1}
11M	10	2.3	13.5±2.1
5M	10	4.3	20.1±0.1
26M	2.5	1.5	21.4±2.6
28M	4	2.1	18.6±0.1
29M	4	2.5	25.7±0.9
31M	8	2.6	42.6±0.4

[a] Manighetti *et al.* (1995).

carbon flux estimated from the mean ^{210}Pb flux. This has been compared to Pfannkuche's (1993) measurements at 47° N of scoc using *in situ* measurements. Overall the oxygen consumption carbon fluxes overestimate the ^{210}Pb-derived benthic flux at each site by as much as 100%; however, this is the same as the range found between spring and summer by Pfannkuche. Given that the oxygen electrode work was carried out after the spring bloom, the high dissolved oxygen fluxes measured here may be a result of rapid degradation of labile, recently sedimented phytodetritus. The moored sediment trap at 47° N recorded a second major phytodetritus flux event between mid-August and mid-September 1989 (the time of the sampling cruise). For sites north of 47° N, it is probable that the fluxes measured here can be regarded as almost a maximum, certainly higher than the interannual long-term average estimated from the ^{210}Pb inventories.

(*d*) *The Holocene burial flux*

The final burial flux of organic carbon is the net result of the sum of all aerobic (and sometimes anaerobic) respiration reactions occurring in the water column and upper sediment. This is the record that palaeoceanographers seek to unravel in an attempt to back-estimate previous variations in oceanic new production (Müller & Suess 1979; Sarnthein & Winn 1990). In table 2 the annual burial flux estimated as an average for the last 10 000 years is shown. The burial flux is calculated as:

$$F_b = \rho \cdot C \cdot S \qquad (7)$$

where is the burial flux of organic carbon; ρ is the sediment dry bulk density; C is the organic carbon concentration; and S is the linear sediment accumulation rate estimated from oxygen isotope profiles measured on foraminifera, ash deposits and radiocarbon dating (Manighetti *et al.* 1995).

The considerable burial diagenesis of the organic matter reduces the carbon burial flux by an order of magnitude, although (see table 2) there is still a suggestion that increased organic carbon burial fluxes occur southwards in the direction of increased surface ocean productivity.

3. CONCLUSIONS

The results described above present a coherent picture of export flux, benthic flux and oxygen respiration of organic carbon; it reflects the surface productivity in the northeast Atlantic. ^{210}Po and ^{210}Pb can be considered as viable and valuable tracers of organic carbon flux through the marine system despite the high degree of advection of both water masses and near-bottom particulate material common in the north Atlantic. At the benthic interface, the seasonal arrival of labile phytodetritus stimulates benthic activity and bacterial degradation promotes active oxygen consumption. Both processes display a spatial trend reflecting primary production and export flux gradients along the 20° W meridian. The data presented here support other published work (Pffankuche 1993) suggesting that remineralization processes occurring at the seafloor are strongly seasonal in this region. Only a small proportion of the benthic carbon flux survives the remineralization process at the sediment–water interface to become buried in the late Holocene sediment.

It is now clear that deep-water and benthic processes are closely linked to upper water column biology. An integrated approach using a range of geochemical tracers is required to understand the flow of organic carbon from the atmosphere to its burial in deep-sea sediments. As our ability to track these important temporal and spatial fluxes of carbon improves, we will be in a far better position to read the long-term historical record laid down in the palaeoceanography of the sediments.

This work was made possible by the co-operation of numerous colleagues within the BOFS programme: Hilary Kennedy, Phil Newton, Tim Fileman, Maureen Conte, Simon Wakefield, John Thomson, Roy Lowry & Polly Machin deserve special mention. Thorough reviews by Maureen Conte and Christophe Rabouille helped clarify inconsistencies in the original text. This work was supported by NERC grants GST 02/300 and GST 02/427.

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Discussion

C. Rabouille (*CFR,CNRS-CEA, Gif-sur-Yvette, France*). Thinking of what Dr M. Conte has showed about the rapid inclusion of organic tracers in the sediment down to a few centimetres, would it not be possible to track newly deposited particles by $^{210}Po/^{210}Pb$ disequilibrium in the sediment?

G.B. Shimmield. In principle, rapidly deposited bloom phytodetritus should display measurable $^{210}Po/^{210}Pb$ disequilibrium. In practice, sediments will need to be recorded and analysed quickly due to the 138 d half life of the ^{210}Po, and sampling will need to be carried out at the mm depth scale on multi-core samples with intact sediment–water interface.

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