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Carbon budget for the mid-slope depocenter of the Middle Atlantic Bight

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Abstract—A mass budget was constructed for organic carbon on the upper slope of the Middle Atlantic Bight, a region thought to serve as a depocenter for fine-grained material exported from the adjacent shelf. Various components of the budget are internally consistent, and observed differences can be attributed to natural spatial variability or to the different time scales over which measurements were made. The flux of organic carbon to the sediments in the core of the depocenter zone, at a water depth of ~1000 m, was measured with sediment traps to be ~65 mg C m⁻² day⁻¹, of which 6–24 mg C m⁻² day⁻¹ is buried. Oxygen fluxes into the sediments, measured with incubation chambers attached to a free vehicle lander, correspond to total carbon remineralization rates of 49–70 mg C m⁻² day⁻¹. Carbon remineralization rates estimated from gradients of C_{org} within the mixed layer, and from gradients of dissolved ammonia and phosphate in pore waters, sum to only ~4–6 mg C m⁻² day⁻¹. Most of the C_{org} remineralization in slope sediments is mediated by bacteria and takes place within a few mm of the sediment–water interface.

Most of the C_{org} deposited on the upper slope sediments is supplied by lateral transport from other regions, but even if all of this material were derived from the adjacent shelf, it represents <2% of the mean annual shelf productivity. This value is further lowered by recognizing that as much as half of the C_{org} deposited on the slope is refractory, having originated by reworking from older deposits. Refractory C_{org} arrives at the sea bed with an average ¹⁴C age 600–900 years older than the pre-bomb ¹⁴C age of DIC in seawater, and has a mean life in the sediments with respect to biological remineralization of at least 1000 years. Labile carbon supplied to the slope, on the other hand, is rapidly and (virtually) completely remineralized, with a mean life of <~1 year. Carbon-14 ages of fine-grained carbonate and organic carbon present within the interstices of shelf sands are consistent with this material acting as a source for the old carbon supplied to the slope. Winnowing and export of reworked carbon may contribute to the often-described relationship between organic carbon preservation and accumulation rate of marine sediments.

INTRODUCTION

EARLY budgets for organic carbon on the continental shelf of the Middle Atlantic Bight (MAB) suggested that heterotrophic carbon consumption is decoupled from primary

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productivity during the winter and during spring bloom periods, with ~90% of the organic carbon (C_{org}) produced during the spring bloom and ~50% of the annual production of C_{org} being unconsumed (e.g. WALSH *et al.*, 1981, 1985; MALONE *et al.*, 1983). Shelf sediments in the MAB are generally sandy with a low C_{org} content, so it appeared that ungrazed C_{org} is not being buried on the shelf. It was postulated, therefore, that much of the ungrazed C_{org} is exported to the deep sea, either by cross-shelf transport extending beyond the shelf-slope break, or by along-shore transport with the mean southward flow of shelf water, eventually to be exported with the water at Cape Hatteras. Extrapolating this imbalance of shelf production and consumption globally (e.g. WALSH *et al.*, 1985) led to carbon fluxes that potentially represent a major route for the transfer of CO_2 from the atmosphere to the deep sea.

Several lines of evidence suggested that the upper slope might serve as a depocenter for particulate organic carbon exported from the shelf. In contrast to the shelf sands, upper slope sediments tend to be fine grained (BISCAYE and OLSEN, 1976) and have a high organic content (1–2% C_{org} ; Fig. 1). Low physical turbulence on the upper slope (CSANADY, 1990) provides a zone with ideal conditions for deposition of fine-grained suspended particles exported from the shelf (BISCAYE and ANDERSON, 1994). This evidence, in part, led to the hypothesis that the upper slope, centered at ~1000 m, serves as a depocenter for fine-grained particles exported from the shelf, including much of the C_{org} produced on the shelf but not consumed there (WALSH *et al.*, 1985).

Examining the fate of fine-grained particulate matter in shelf waters, including C_{org} and associated substances, has been a principal objective of the Shelf-Edge Exchange Processes (SEEP) program. The first field program (SEEP-I) was carried out in the northern part of the MAB (WALSH *et al.*, 1988), where fluxes of particulate organic carbon (POC) collected by sediment traps deployed along a transect normal to the slope suggested that very little ($\leq 1\%$) of the POC produced on the northern MAB shelf is exported to the adjacent slope (BISCAYE *et al.*, 1988). Contemporary studies indicated that earlier estimates of C_{org} consumption on the shelf had been low, and further indicated a much tighter coupling than had been thought previously between production and consumption of C_{org} (ROWE *et al.*, 1986; FALKOWSKI *et al.*, 1988).

Although SEEP-I studies demonstrated that most of the C_{org} produced on the shelf is also consumed there, the revised budgets still allowed for 10–20% of the annual production of C_{org} on the shelf to be exported (FALKOWSKI *et al.*, 1988). Extrapolating globally, using values from WALSH (1991), exporting 10–20% of shelf productivity to the deep sea represents a carbon flux of 0.5–1.0 gigatons per year. The potential magnitude of this flux warrants further examination of shelf export processes. It is unlikely that carbon export can be constrained by constructing carbon budgets for the shelf to a level of uncertainty better than the 10–20% reported by FALKOWSKI *et al.* (1988) because of the natural levels of spatial and temporal variability inherent in such systems. Better constraints, however, can be provided by examining directly the fluxes of exported carbon.

Part of the original SEEP hypothesis (WALSH *et al.*, 1988) was that the cross-shelf transport of fine-grained particulate matter, including POC, would increase in a southward direction in the MAB. Consequently, the SEEP-II field program was designed to study a region anticipated to exhibit maximum cross-shelf export of POC before the shelf waters themselves leave the coast at Cape Hatteras. The principal objective of this paper is to better constrain the export of shelf-derived POC by evaluating various components of

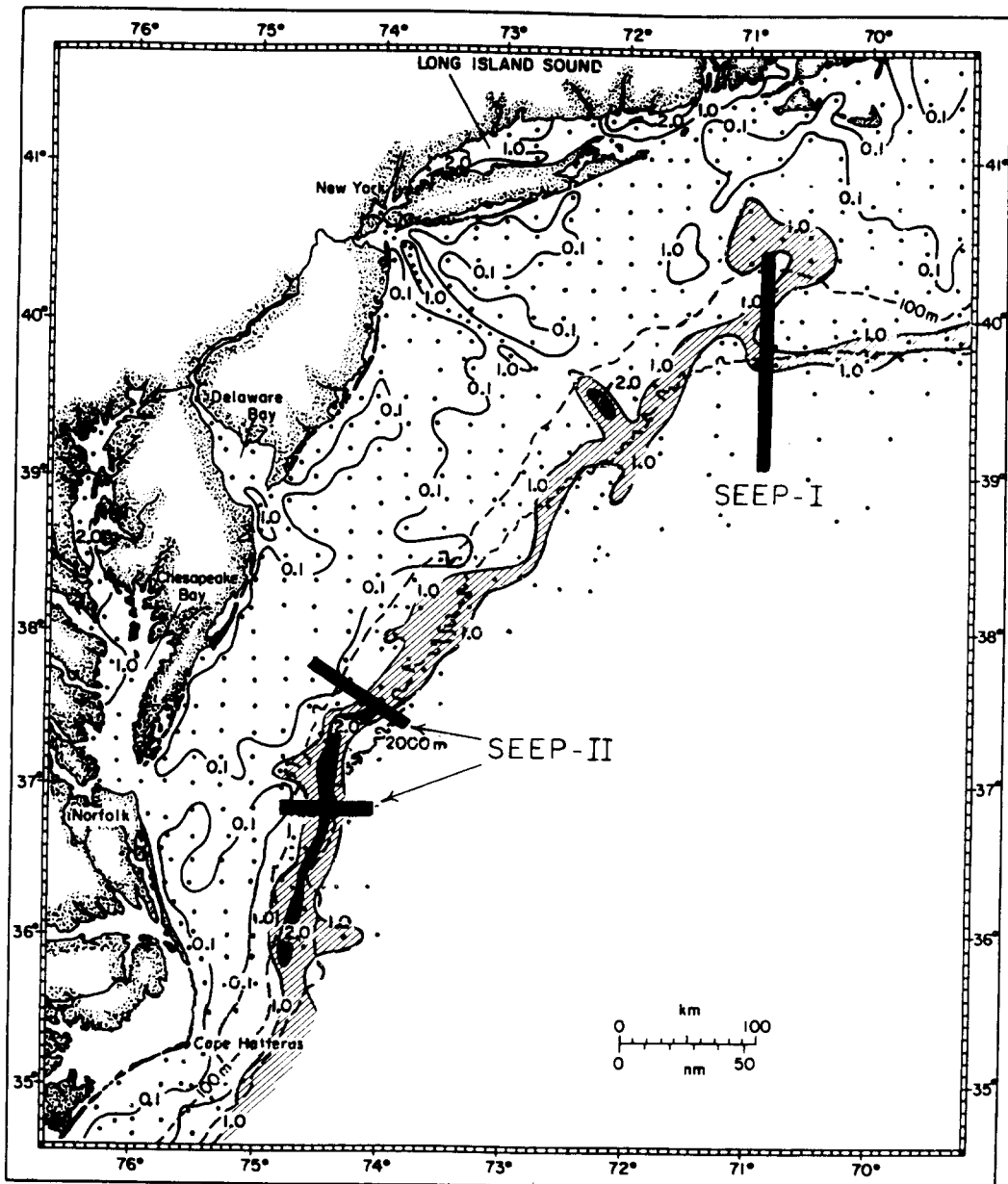


Fig. 1. Map of the Middle Atlantic Bight showing organic carbon content of surface sediments. The hatched area of $>1\%$ organic carbon roughly defines the zone referred to in the text as the depocenter. The SEEP-II study crossed the band of sediments exceeding 2% organic carbon (shaded area) along the two transects shown as solid lines. Adapted from WALSH *et al.* (1988).

the carbon budget for the depocenter region of the upper slope in the southern MAB. Distributions of C_{org} in the sediments, as well as the ^{14}C content of the carbon, will then be used to characterize the nature and the turnover rate of the carbon supplied to the depocenter.

STUDY AREA, SAMPLING AND ANALYTICAL METHODS

The SEEP-II field program was carried out in the southern part of the Middle Atlantic Bight off the coasts of Maryland and Virginia (Fig. 1). Here, the shelf is narrower than in the northern MAB, this being one of the factors postulated to contribute to the southward increase in the cross-shelf export of POC (BISCAYE *et al.*, 1994). Two transects of moored instruments were deployed to permit transport processes to be studied in a three-dimensional context (BISCAYE *et al.*, 1994). Coring (Table 1) and other benthic studies also were carried out along these transects.

Shelf sediments are generally sandy. Inner shelf sediments also contain a substantial gravel component, whereas outer shelf sands contain some fine-grained material within their interstices. Sand is also present in upper slope sediments down to a depth of ~500 m. Organic carbon contents achieve their maximum values, in excess of 2% by weight, at about the 1000-m isobath (Fig. 2, Appendix), consistent with the hypothesis that this is the locus of the depocenter zone. Shell fragments are abundant in upper-slope sediments, whereas forams constitute the predominant visible carbonates at greater depths. Maximum calcium carbonate contents are ~20% (Appendix).

Sediment collection and analysis

Detailed descriptions of the sediment traps deployed along both SEEP-II transects are given in BISCAYE and ANDERSON (1994). Briefly, traps were deployed at one location on the shelf, one at the shelf edge, and two over the slope on the northern transect. The southern transect was similarly instrumented, except that a trap was not deployed on the shelf. Traps were deployed three times over a total period of ~15 months. During each deployment, sample collection was divided into 10 segments, so that a complete record at each trap position consisted of 30 samples. Sodium azide was used as a poison/preservative during the first sediment trap deployment, whereas buffered formalin was used during the second and third deployments (see BISCAYE and ANDERSON, 1994, for rationale behind changing poisons). Swimmers were hand picked from each sample. Box cores and gravity cores (Table 1) were collected as described by ANDERSON *et al.* (1994). Additional samples of surface sediments (up to ~15 cm depth) were collected along both transects using a Shipek grab, which was used principally in sandy sediments where box coring generally fails to recover intact cores.

Box cores were subsampled with standard 4-inch i.d. plastic tubes (ANDERSON *et al.*, 1994) as well as with pressurized-barrel subcores designed by JAHNKE (1988) for the extraction of pore waters. Pore waters were collected as described by JAHNKE (1988), beginning within a few hours of core recovery. Pore waters were immediately frozen on board ship and returned to the lab (TAMU) for analysis by autoanalyzer. Separate aliquots of pore water were returned to LDEO unfrozen, but acidified, for measurement of dissolved iron by atomic absorption. Subcores from the box cores were carried back to the lab (LDEO) in an upright orientation to minimize disturbance. Subcores were kept refrigerated until they were extruded and sectioned at 1 to 2-cm intervals as described by ANDERSON *et al.* (1994). Wet and dry weights were obtained, from which porosities were calculated assuming a mineral density of 2.5 g cm^{-3} and a salinity of 35‰.

Additional subcores were collected from the box cores using 1-inch i.d. PVC tubes into which syringe plungers were inserted to act as miniature piston-coring devices. Pistons

Table 1. Locations of cores used in this study

Cruise/ core	Collection date (mm/dd/y)	Water depth (m)	Latitude (N)	Longitude (W)
<i>EN 179</i>				
BC5	6/19/88	384	37°37.81'	74°13.06'
BC1	6/16/88	417	37°37.97'	74°12.86'
BC2	6/17/88	892	37°37.48'	74°09.95'
BC3	6/17/88	1031	37°38.36'	74°08.57'
BC4	6/19/88	1318	37°31.74'	74°02.26'
BC7	6/20/88	1989	37°25.06'	73°49.41'
GC2	6/17/88	1034	37°37.86'	74°09.09'
<i>EN 187</i>				
BC2	10/26/88	60	37°45.13'	74°28.86'
BC1	10/25/88	265	37°38.06'	74°12.25'
BC3	10/26/88	410	37°38.11'	74°12.92'
BC4	10/27/88	512	37°37.33'	74°13.30'
BC10	10/29/88	580	36°52.45'	74°36.77'
BC8	10/29/88	1020	36°52.00'	74°34.48'
BC5	10/27/88	1045	37°37.05'	74°10.03'
BC11*	10/29/88	1125	37°02.30'	74°34.26'
BC9	10/29/88	1165	36°52.20'	74°33.84'
BC6	10/28/88	2000	37°24.01'	73°49.69'
GC2	10/27/88	770	37°37.33'	74°11.59'
GC3	10/28/88	1015	37°37.15'	74°09.76'
GC5	10/29/88	990	36°52.19'	74°34.65'

*Axis of Norfolk Canyon.

were held at a constant height above the sediment-water interface so that the subcores could be inserted without suffering from core compression that is otherwise inherent in subcoring with small-diameter tubes. Two or three of these subcores were collected from most of the box cores. Small subcores were capped and taped at sea so that samples were returned to the lab (LDEO) in the original subcoring tube. Sediments were extruded and cut into two sections (surface to ~15 cm and ~15 cm to the bottom), which were weighed without trimming, dried to a constant weight, and homogenized as described above.

Gravity cores, which typically recovered about 3 m of sediment, were cut into two sections at sea for transport back to the LDEO core facility where they were split, described, and stored under refrigeration. Sediments were taken from the working half of each gravity core for analysis of C_{org} , $CaCO_3$, and ^{14}C . These samples, too, were dried to a constant weight at 60°C, ground, and homogenized before aliquots were taken for various analyses.

Sediments collected by Shipek grab were transferred at sea into plastic jars (~600 cc) without attempting to preserve their *in situ* orientation. Grab samples were homogenized prior to analysis. For this study, 50 cc of wet sediment (measured by packing into a

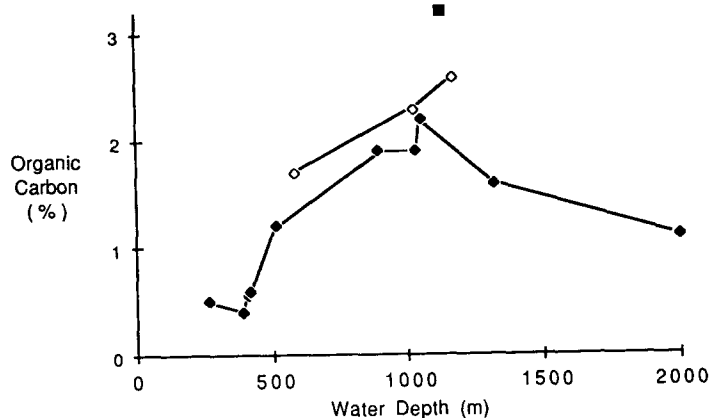


Fig. 2. Organic carbon content of surface sediments in the southern Middle Atlantic Bight plotted against water depth. Samples were collected by box coring as described in the text. Solid symbols represent samples collected along the northern SEEP-II transect; open symbols represent samples from the southern transect. The square indicates a sample collected from the Norfolk Canyon, which is located between the two SEEP-II transects.

graduated cylinder) were submitted to repeated resuspension in a 250 cc graduated cylinder. Sand was allowed to settle briefly (~1 min) after which the supernate, and the fines remaining in suspension, were decanted. This was repeated until the supernate remained clear following resuspension. The number of washings required to remove the fines increased from inner shelf to outer-shelf samples, but typically about four washings were employed. Fines were concentrated by centrifugation, dried at 60°C, weighed, ground and homogenized.

Calcium carbonate contents were determined by the vacuum gasometric technique of JONES and KATERIS (1983) for sediment samples from one gravity core (EN187 GC5) and for all box core samples except those from EN187 BC8. Organic carbon and nitrogen were determined for all sediment samples, after elimination of carbonate with sulfurous acid, following the method of VERARDO *et al.* (1990) using a Carlo Erba elemental analyzer. Total carbon contents of the samples from EN187 BC8 and from the other gravity cores were measured using the Carlo Erba, and carbonate contents were calculated from the difference between total carbon and organic carbon.

Samples from each of the box cores were analyzed for ^{210}Pb , and a subset of those samples was also analyzed for $^{239+240}\text{Pu}$. Sediments were dissolved in the presence of isotopic yield monitors, after which the radionuclides were purified and plated onto silver disks for counting by alpha spectrometry following the method of ANDERSON *et al.* (1988). In the case of ^{210}Pb , it is the decay product ^{210}Po that is actually counted.

Ten to 15 cm sections of gravity core material (~two-thirds of the working half of the core) were required to obtain sufficient carbon (≥ 1 g) for measurement of ^{14}C by conventional counting methods in the LDEO ^{14}C lab. The method involves the synthesis of benzene from CO_2 via formation of acetylene. ^{14}C in the benzene is measured by liquid scintillation counting. The carbonate fraction was analyzed for each sample. The organic fraction was analyzed for a selected subset of the samples. Carbonate carbon was obtained

by acidification. Organic carbon was obtained by burning samples in an oxygen atmosphere and trapping the CO₂.

Carbon-14 was measured in small samples by accelerator mass spectrometry (AMS). Small samples included fine-grained sediments separated from shelf sands, sediment trap samples, and sediments from one box core sampled at higher depth resolution than other cores. Samples from one gravity core (EN187 GC5) also were analyzed by AMS. It was sometimes necessary to composite material from multiple time-series sediment trap samples to obtain sufficient carbon (>1 mg each organic and carbonate carbon) to measure ¹⁴C. In compositing, amounts of material from each sample were taken in proportion to the mass flux collected during that sample interval to maintain the proper flux-weighted mean concentrations in the composites.

For those samples analyzed by AMS, carbon dioxide was evolved from CaCO₃ by acidification with phosphoric acid in an evacuated reactor vessel. Acidifications were performed on dried, ground bulk sediment for all but the samples from EN187 GC5. Sediments from GC5 were wet sieved through a 63-micron filter. The portion retained on the filter, including identifiable foraminifera and small mollusc (clam?) shell fragments, was acidified for determination of Ca¹⁴CO₃. The fine-grained (<63μ) material passing through the sieve was also acidified to remove carbonates prior to determination of ¹⁴C of the organic matter. Sediment residues after acidification were repeatedly washed with distilled water, vacuum dried and ground. Organic carbon was combusted to CO₂ at 900°C with cupric oxide wire in sealed, evacuated quartz tubes (BOUTTON *et al.*, 1983).

The CO₂ evolved from carbonates and organics was purified cryogenically, then reduced to graphite using the cobalt-catalyzed reduction method described in VOGEL *et al.* (1987). Graphite targets were run at the Center for Accelerator Mass Spectrometry at the Lawrence Livermore National Laboratory (DAVIS *et al.*, 1990). Accuracy and precision of the spectrometer system improved throughout the period during which samples for this project were run; errors were generally less than 2% of Modern. Carbon-14 results are reported as the deviation in parts per thousand of the ¹⁴C/¹²C ratio from that of an absolute standard (0.95 times the activity of oxalic acid in 1950), Δ¹⁴C (STUIVER and POLACH, 1977). Corrections for isotope fractionation were made assuming a δ¹³C value of 0 for carbonates, and using a representative value of -21‰ previously measured for organic carbon in MAB slope sediments (TANAKA *et al.*, 1991). Ages were calculated from the Δ¹⁴C values using the conventional ¹⁴C half-life (5567 years). Carbon-14 results are presented here as ages where this gives the reader a better intuitive feel for the time scales involved. Where ¹⁴C results are used to construct carbon budgets, they also are presented as per mill deviation from the modern standard (Δ¹⁴C) to simplify calculations when carbon components of different ages are added to, or subtracted from, one another.

Benthic oxygen fluxes

Oxygen fluxes into sediments were measured in June and October, 1988, using a free-vehicle lander equipped with two incubation chambers. Increases and decreases of oxygen and metabolites within the chambers were used to infer fluxes into or out of the sediments. Oxygen concentration was monitored continuously using polarographic electrodes and dissolved nutrients were measured in samples collected during the deployment using 60 cc hypodermic syringes. Additional details of the lander, and of its operation, are described by ROWE *et al.* (1994).

Bacterial abundance and production

Sediment samples for bacterial abundance and production estimates were collected in February, March, June and October 1988, and in March 1989. All samples collected in February, March and June, and one sample in October 1988 were obtained using a free-vehicle benthic lander. Thereafter, samples were collected in sandy shelf sediment using a van Veen grab and in muddy slope sediment using an Oceanic Instruments box corer.

Bacterial abundance and cell volumes in sediment were estimated by epifluorescence microscopy of diamidinophenylindole-stained (PORTER and FEIG, 1980), 0.25 cm^{-3} sub-samples, using homogenization in pyrophosphate to detach bacteria from sediment (VELJI and ALBRIGHT, 1986). Bacterial production estimates utilized here were obtained by shipboard incubations of surficial (3–4 mm layer) sediment with varying concentrations of tritiated thymidine, from 200–3200 nM. Production estimates are specifically for the surficial 3–4 mm of aerobic sediment from which samples were collected. Incubations were terminated with 2% formaldehyde. Radioactivity incorporated into the DNA fraction of TCA-precipitated macromolecules was converted to the equivalent number of cells produced using an empirically determined conversion factor, as described in KEMP (1994). Cell production was converted to the carbon equivalent assuming that the carbon content per cell was 0.1 g C cm^{-3} (NAGATA, 1986; NORLAND *et al.*, 1987). The selection of this conversion factor was further supported by a comparison of estimated benthic bacterial production to measured oxygen uptake in shipboard incubation of retrieved cores. Additional details are given in KEMP (1994).

RESULTS AND DISCUSSION

Overview of the carbon budget

Particulate material collected by time-series sediment traps provided the source term for the carbon budget for the upper-slope depocenter zone; i.e. the POC flux to the sediments. Carbon burial was calculated from the ^{14}C -derived sediment accumulation rate and the C_{org} content of sediments below the depth of the bioturbated mixed layer. Total organic carbon remineralization was calculated, assuming conditions of steady state, as the difference between supply and burial.

Several complementary approaches were utilized to evaluate C_{org} remineralization. Each has its inherent limitations, but internal consistency among the results lends confidence to the overall interpretation and carbon budget. Pore water profiles of ammonia and phosphate were modelled to estimate C_{org} remineralization within and below the mixed layer. Carbon gradients through the bioturbated mixed layer (to depths of 10–20 cm) were multiplied by the ^{210}Pb -derived sediment mixing rate to provide an independent estimate of C_{org} remineralization within the mixed layer. Rates of bacterial production in the upper 3–4 mm of sediment were used to estimate bacteria-mediated remineralization of C_{org} near the sediment–water interface. Fluxes of oxygen into the sediments provided a lower limit for total benthic C_{org} remineralization. Results from each approach will be examined in the context of complementary results to assess the reliability of the overall carbon budget.

Natural spatial and temporal variability makes it difficult to use any single value, or measurement, as representative of a particular component of the C_{org} budget for the entire depocenter region. Therefore, averages and ranges of values from a number of measure-

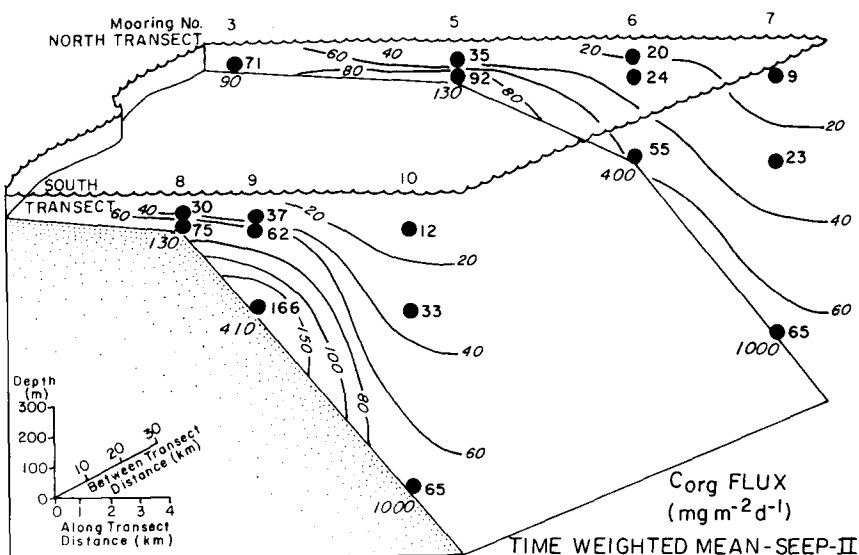


Fig. 3. Schematic representation of the outer shelf and upper slope of the MAB showing the relative positions of the moored sediment traps (solid dots). Locations of the north (Moorings 3, 5, 6 and 7) and south (Moorings 8–10) transects are shown in Fig. 1. The time-weighted mean annual flux of organic carbon collected at each trap position is shown by the value to the right of each dot, while the general pattern of carbon flux is illustrated by the contours (from BISCAYE and ANDERSON, 1994).

ments will be used in each case to construct an approximate average carbon budget for the depocenter region as a whole.

Carbon supply

Time-weighted mean annual fluxes of POC collected by the sediment traps provided the source term for the carbon budget. Two moorings on each SEEP-II transect were located within the upper-slope depocenter zone (Fig. 3): Moorings 6 (~400 m) and 7 (~1000 m) on the north transect and Moorings 9 (~400 m) and 10 (~1000 m) on the south transect. Time-series records and other details of the carbon fluxes are provided by BISCAYE and ANDERSON (1994). Here, it is sufficient to note that mean annual C_{org} fluxes in slope waters increased with increasing water depth (Fig. 3), in contrast to the situation in pelagic regions where POC fluxes decrease with depth due to net C_{org} consumption below the euphotic zone (see review by BISHOP, 1989). Fluxes of total mass and of ^{210}Pb also increased with depth (Fig. 4), and the ^{210}Pb flux increased at a rate much greater than could be supplied by decay of ^{226}Ra in the water column ($<0.5 \text{ dpm } ^{210}\text{Pb cm}^{-2} \text{ y}^{-1}$ per 1000 m of water column). Lateral supply of particulate material to the slope, or collection by the traps of locally resuspended sediments, or some combination of the two, must be invoked to explain these observations.

The degree to which fluxes collected by the sediment traps are representative of sediment accumulation on the slope over longer time scales (10–100 years) can be assessed by comparing mean annual fluxes of ^{210}Pb collected by near-bottom traps with fluxes required to sustain inventories of unsupported (xs) ^{210}Pb in the slope sediments. Inventor-

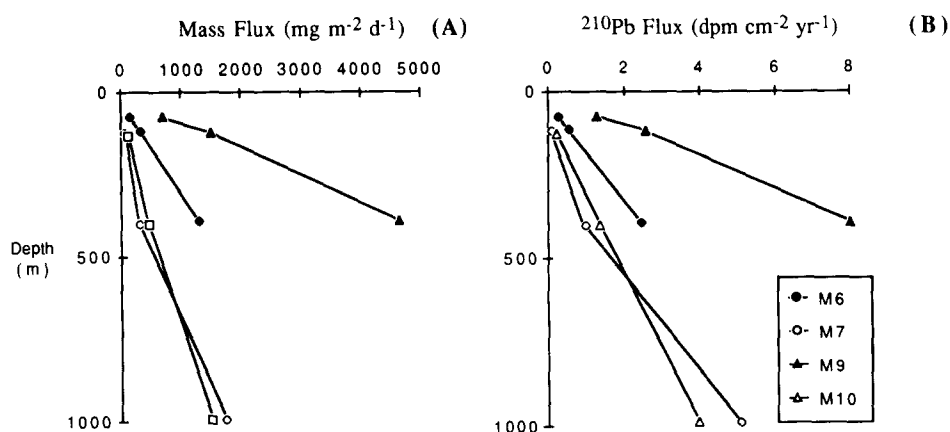


Fig. 4. Time-weighted mean mass fluxes (A) and fluxes of ^{210}Pb (B) collected by sediment traps deployed over the upper slope in the southern Middle Atlantic Bight. Symbols correspond to the moorings as defined in the key.

ies of $x_s^{210}\text{Pb}$ were calculated from results in the Appendix as well as from the $x_s^{210}\text{Pb}$ contents of the small-diameter subcores. Three or four measurements of $x_s^{210}\text{Pb}$ inventory were thus obtained from each box core. Multiple measurements were deemed necessary because of small-scale (tens of cm) spatial variability observed in previous studies (ANDERSON *et al.*, 1993 and references therein). Variability of as much as a factor of two was found within some of the SEEP-II box cores, consistent with earlier observations. By making many measurements of $x_s^{210}\text{Pb}$ inventories throughout the depocenter region, a more reliable average was obtained with which to compare the trap results.

Fluxes of ^{210}Pb collected by near-bottom sediment traps agreed well with the fluxes required to support inventories of $x_s^{210}\text{Pb}$ in nearby sediments for three out of the four moorings located in the depocenter region (Fig. 5). Only for Mooring 9, located at ~ 400 m on the south transect, did the ^{210}Pb flux collected by sediment traps appear to be too high. Fluxes of other constituents were also systematically high at this trap position throughout the field program (BISCAYE and ANDERSON, 1994). Whether this situation at Mooring 9 resulted from local resuspension of sediments, from collection by the traps of particles moving downslope past this site but not accumulating there, from a temporal anomaly of local particle flux, or from something else entirely, is not known at this time. Whatever the cause, the carbon flux collected in this trap will not be used to estimate the average rate of supply of POC to the depocenter.

Consistency among the ^{210}Pb fluxes collected by traps at the other three mooring sites and the fluxes required to support sediment inventories indicated that the fluxes collected by these traps are representative of the supply of particulate matter, and hence of POC, to the depocenter. The observed increase with depth in the POC flux was not caused by local sediment resuspension or by other problems that may bias the collection of fluxes by sediment traps. Mean annual POC fluxes collected by the near-bottom traps at the three representative moorings were 55 (Mooring 6), 65 (Mooring 7) and 65 (Mooring 10) $\text{mg C m}^{-2} \text{ day}^{-1}$. A value of $65 \text{ mg C m}^{-2} \text{ day}^{-1}$ will be taken to represent the depocenter as a whole because the larger value is more consistent with oxygen fluxes and microbial

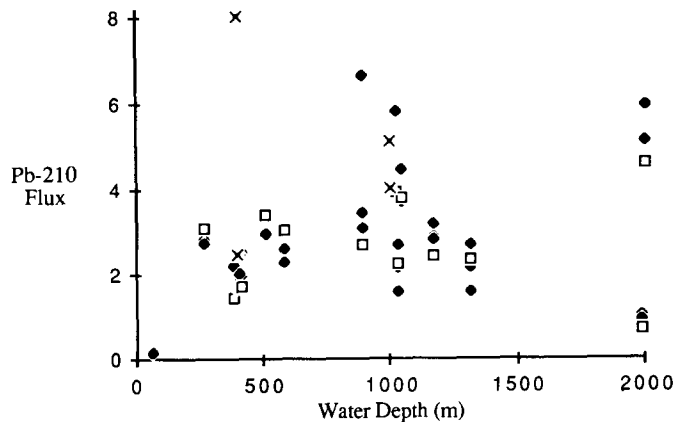


Fig. 5. Fluxes of ^{210}Pb ($\text{dpm cm}^{-2} \text{y}^{-1}$) plotted against water depth. Open squares and solid diamonds represent fluxes required to maintain measured inventories of unsupported ^{210}Pb in sediments. Open squares are from sectioned 10-cm i.d. subcores of box cores; solid diamonds are from 1-inch diameter subcores. Time-weighted mean fluxes of ^{210}Pb measured by sediment traps deployed ~ 10 m above bottom at the indicated water depths are shown as Xs. The high trap value at 400 m was obtained at Mooring 9 on the south transect.

rem mineralization rates (see below) and because it will provide a more conservative upper limit for the export of carbon from the shelf to the slope.

An independent estimate of the C_{org} flux to depocenter sediments can be obtained by multiplying the mean $C_{\text{org}}/x^{210}\text{Pb}$ ratio of trapped particles ($0.5\text{--}0.9 \text{ mg } C_{\text{org}}/\text{dpm } ^{210}\text{Pb}$; from data in BISCAYE and ANDERSON, 1994) by the ^{210}Pb flux required to support $x^{210}\text{Pb}$ inventories in the sediments. A representative flux of ^{210}Pb to the slope sediments is $\sim 3 \text{ dpm cm}^{-2} \text{y}^{-1}$ (Fig. 4), which leads to a range of C_{org} fluxes of $41\text{--}74 \text{ mg } C_{\text{org}} \text{ m}^{-2} \text{ day}^{-1}$ that is free of uncertainties related to biases in the collection efficiency of sediment traps. Consistency between these values, and the measured C_{org} fluxes of $55\text{--}65 \text{ mg } C_{\text{org}} \text{ m}^{-2} \text{ day}^{-1}$, again testifies to the reliability of the fluxes collected by the traps.

Carbon burial

Sediment accumulation rates were calculated from the change with depth in the ^{14}C age of the sediment. Accumulation rates were found to be $35\text{--}40 \text{ cm ky}^{-1}$ at ~ 1000 m water depth on the north transect [Fig. 6(b) and (c)] and slightly lower at 770 [$\sim 30 \text{ cm ky}^{-1}$; Fig. 6(a)]. Carbon-14 profiles on the south transect [Fig. 6(d)], obtained by combining detailed results from a box core (EN187 BC9; 1165 m) with results for sediments from a nearby gravity core (EN187 GC5; 990 m), are less easy to interpret. Sediment deposition at this site has clearly not been at steady state over the past several thousand years, as evidenced both by the change in the slopes of the ^{14}C profiles and by the downcore change in the offset of ages between carbonate and organic carbon. A reasonable accumulation rate of $\sim 60 \text{ cm ky}^{-1}$ is obtained from the C_{org} ages in the upper ~ 1.25 m. This will serve as an approximate upper limit for the average sediment accumulation in the depocenter zone.

Mass accumulation rates were calculated from the linear accumulation rates ($30\text{--}60 \text{ cm ky}^{-1}$) assuming an average porosity below the bioturbated layer of 70% (based on data

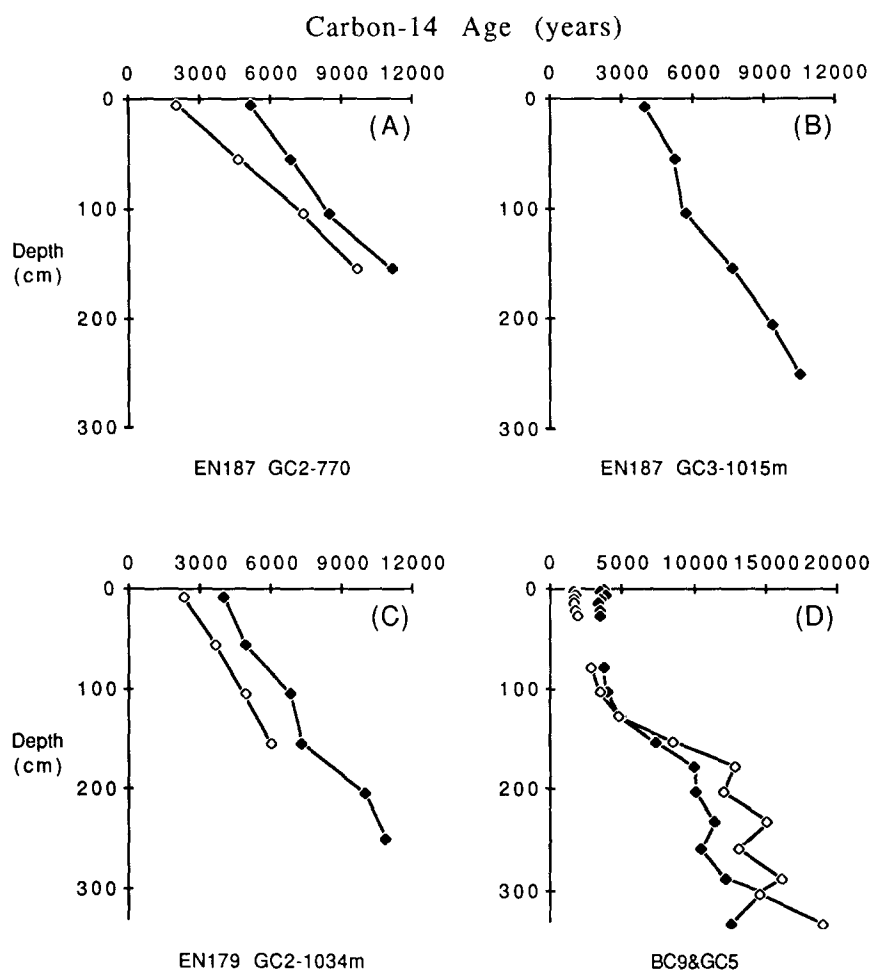


Fig. 6. ^{14}C ages of organic (open symbols) and carbonate (solid symbols) carbon in samples from cores collected in the main part of the depocenter of the southern Middle Atlantic Bight. The first three cores are from the northern SEEP-II transect; the composite profile from EN187 GC5 (990m) and EN187 BC9 (1165m) is from the south transect.

from gravity cores EN187 GC3, EN187 GC4, and EN179 GC2) to be $22.5\text{--}45\text{ g cm}^{-2}\text{ ky}^{-1}$ ($616\text{--}1233\text{ mg m}^{-2}\text{ day}^{-1}$). Time-weighted mean fluxes for the three representative traps were 1297 (Mooring 6), 1758 (Mooring 7) and 1522 (Mooring 10) $\text{mg m}^{-2}\text{ day}^{-1}$. Although fluxes collected by the traps were a little higher, the general consistency suggests that the trap fluxes provided a representative measure of particle deposition over time scales of millennia, just as the ^{210}Pb results indicated consistency over time scales of decades.

Rates of C_{org} burial in sediments were calculated from the mass accumulation rates and the C_{org} contents of sediments at depths below the mixed layer ($\sim 15\text{ cm}$, see below), although organic carbon distributions in these sediments exhibit little vertical gradient with depth (Fig. 7). Maximum C_{org} contents ($\sim 2\%$; except in the Norfolk Canyon) occur

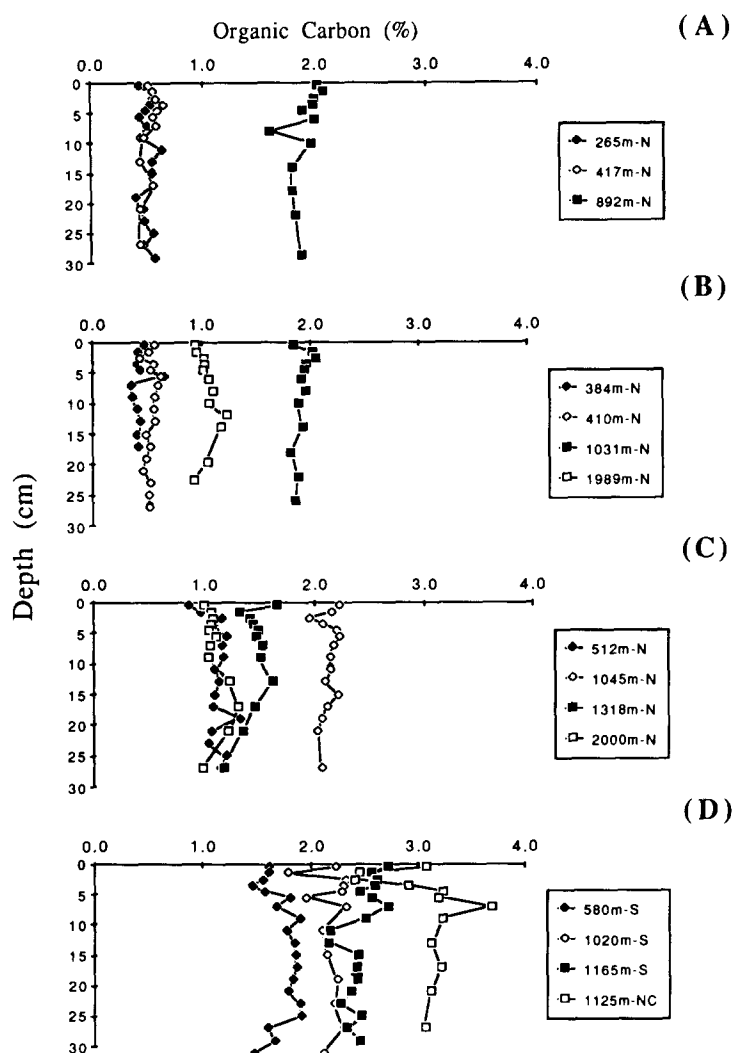


Fig. 7. Organic carbon profiles from box cores collected along the two SEEP-II transects in the southern Middle Atlantic Bight. Panels A–C represent cores from the northern SEEP-II transect; panel D contains three profiles from the south transect and one from the Norfolk Canyon.

near the 1000-m isobath, with values decreasing at both shallower and greater water depths (Figs 2 and 7). A full range of carbon accumulation rates can be estimated to be 6–24 $\text{mg C m}^{-2} \text{day}^{-1}$ assuming that sediments are buried at a rate of 616–1233 $\text{mg m}^{-2} \text{day}^{-1}$ with an organic carbon content of 1–2%. These flux estimates suggest that 10–40% of the POC flux to the sea bed is buried rather than consumed (remineralized). Subtracting carbon burial rates ($\sim 6\text{--}24 \text{ mg C m}^{-2} \text{day}^{-1}$) from POC fluxes to the slope sediments ($\sim 65 \text{ mg C m}^{-2} \text{day}^{-1}$) indicates that $\sim 41\text{--}59 \text{ mg C m}^{-2} \text{day}^{-1}$ must be remineralized if the system is at steady-state. Remineralization within the sediments and remineralization near the sediment–water interface will be addressed separately.

Carbon remineralization within the sediments

Remineralization of organic matter "at" the sediment–water interface, for example in surface fluff layers or in flocs of biogenic detritus, would release remineralized nutrients and CO₂ directly into bottom waters without significantly influencing their concentrations in the sediments. The sensitivity with which rates of C_{org} remineralization can be estimated from distributions of nutrients in pore waters or from C_{org} gradients in the sediments is further limited by the depth resolution of sampling procedures. Gradients in the upper few mm of sediments obviously cannot be determined by sampling at cm-depth intervals. For the present discussion, remineralization "within the sediments" refers to processes that regulate the concentrations of nutrients in pore waters and of C_{org} in the solid phases at depths of ~1 cm or greater.

Pore waters. Remineralization of organic matter within the sediments leads to the build up of inorganic forms of nitrogen and phosphorus in pore waters. Upward fluxes of dissolved N and P can be estimated from their concentration gradients if the sediment porosity is known and if some assumptions are made about the nature of diffusion through pore waters. If the composition of the organic matter undergoing remineralization is known, or if reasonable assumptions about C/N and C/P ratios can be made, then carbon remineralization rates can be estimated from the upward fluxes of remineralized nutrients (e.g. FROELICH *et al.*, 1979; EMERSON *et al.*, 1980; BENDER *et al.*, 1989; JAHNKE *et al.*, 1989).

Organic nitrogen remineralized within the sediments of the depocenter region is likely to be present in the pore waters as ammonia. Despite bottom water dissolved oxygen concentrations in excess of 200 micromolar (BEHRENS *et al.*, 1990), sediments become anoxic within a few mm of the sediment–water interface, as evidenced by the presence of dissolved ferrous iron in excess of 1 micromolar in all of the pore waters collected (Anderson, unpublished). Under these conditions, the stable form of remineralized organic nitrogen should be ammonia.

Concentrations of ammonia and phosphate increased with depth in the pore waters (Fig. 8), as expected, but not in a smooth, regular pattern, particularly for phosphate. All of the ammonia profiles exhibit similar concentration gradients in the upper 10 cm [Fig. 8(a)], so an approximate average flux (F_N) representative of all of the ammonia profiles was calculated from the concentration gradient, $\partial[\text{NH}_4^+]/\partial z$, indicated by the straight line in the figure as:

$$F_N = \phi^2 \cdot D_m \cdot (\partial[\text{NH}_4^+]/\partial z) \quad (1)$$

where ϕ is the porosity and D_m the molecular diffusion coefficient. Porosity was assumed to be uniform at 0.8, which is representative of the bioturbated layer (Appendix), and a value of D_m applicable at the bottom water temperatures of ~4°C of $1.0 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$ (LI and GREGORY, 1974) was used. Simplifying assumptions, such as constant porosity and uniform concentration gradient, are acceptable here because they suffice to demonstrate that the diffusive flux of ammonia is small, amounting to $0.92 \mu\text{mol cm}^{-2} \text{ y}^{-1}$. Assuming a C/N ratio of 7 for the metabolized organic matter, this corresponds to a carbon remineralization rate of $13.4 \mu\text{mol C cm}^{-2} \text{ y}^{-1}$, or $4.4 \text{ mg C m}^{-2} \text{ day}^{-1}$. Even allowing for a large error in this estimate, it is clear that only a small portion of the POC flux to the sea floor is remineralized within the sediments.

Irregularities in the phosphate profiles preclude precise calculation of concentration

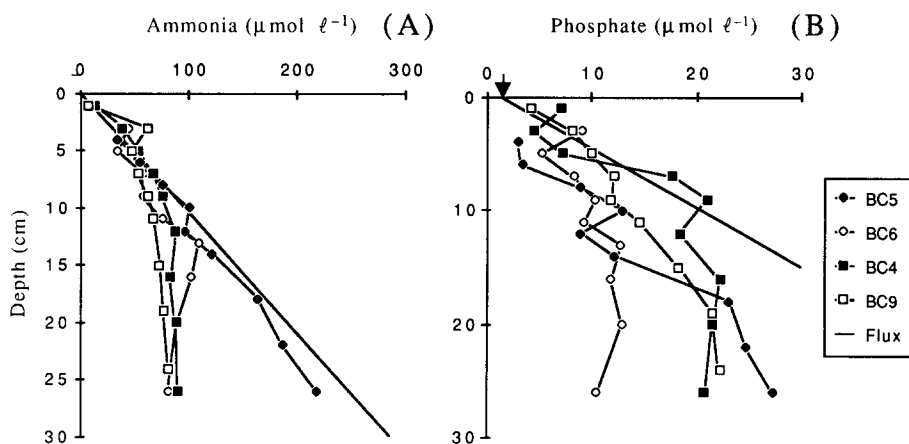


Fig. 8. Profiles of dissolved ammonia (A) and phosphate (B) in pore waters of four cores from the southern Middle Atlantic Bight. All cores are from EN187. BC9 is from the southern transect; the others are from the north transect. The solid lines indicate the concentration gradients used to calculate fluxes in the text. The arrow indicates the concentration of phosphate in the water column at a depth of ~ 1000 m.

gradients. The solid line in Fig. 8(B) is estimated to represent an upper limit for the phosphate gradient, from which a flux of $0.19 \mu\text{mol P cm}^{-2} \text{y}^{-1}$ is calculated following the approach described above for ammonia using a value for D_m of $5 \times 10^{-6} \text{ cm}^2 \text{ s}^{-1}$ (LI and GREGORY, 1974). Assuming a C/P ratio of 106:1 for remineralized organic matter, the corresponding carbon remineralization rate is $20.4 \mu\text{mol C cm}^{-2} \text{y}^{-1}$, or $6.7 \text{ mg C m}^{-2} \text{ day}^{-1}$. As this estimate was based on an upper limit for the phosphate gradient, a reasonable C_{org} remineralization rate within the sediments is taken to be $4\text{--}6 \text{ mg C m}^{-2} \text{ day}^{-1}$ based on the profiles of remineralized nutrients in pore waters.

Bioirrigation of sediments could lead to substantially larger fluxes of carbon and nutrients from the mixed layer than those computed here, where it is assumed that transport is limited to molecular diffusion. Appropriate measurements to estimate rates of bioirrigation were not taken, but other studies have indicated that bioirrigation has little influence on fluxes of dissolved species from slope sediments, in contrast to shelf sediments where bioirrigation may enhance fluxes as much as fourfold compared to those attributable to diffusion alone (ARCHER and DEVOL, 1992). Furthermore, rates of carbon remineralization estimated above are generally consistent with rates estimated using a sediment mixing model (see next section). Therefore, it seems reasonable, in this case, to neglect bioirrigation in calculating fluxes from profiles of nutrients in pore waters.

Carbon gradient-mixing model. Consumption within the sediments causes the C_{org} concentration to decrease with depth. Within the bioturbated mixed layer, random vertical movements of sediments in the presence of a concentration gradient results in a downward flux of sedimentary carbon. Under steady-state conditions, the downward flux of C_{org} driven by mixing balances the consumption of carbon in deeper sediments. Therefore, an independent estimate of the rate of C_{org} remineralization within the mixed layer can be obtained by evaluating the downward flux of C_{org} at the sediment-water interface (see EMERSON *et al.*, 1985 for details) as:

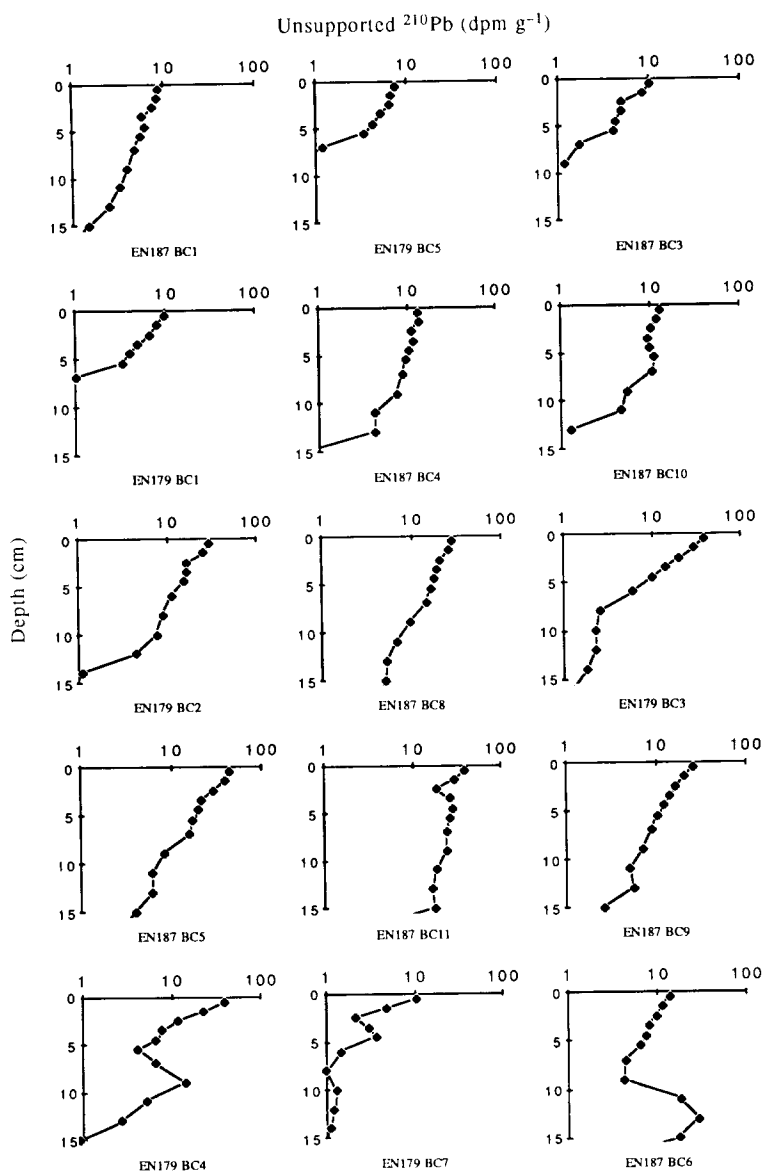


Fig. 9. Profiles of unsupported ^{210}Pb in box cores recovered from the continental slope of the southern Middle Atlantic Bight. Cores are arranged from shallowest water depth in the upper left to greatest water depth in the lower right. Locations and water depths are given in Table 1.

$$F_0 = D_B \cdot (\partial[C_{\text{org}}]/\partial z)_0 = \text{mixed layer consumption} \quad (2)$$

where D_B is the sediment mixing coefficient ($\text{cm}^2 \text{y}^{-1}$) and the subscript (0) indicates an evaluation of the flux using the C_{org} gradient at the sediment-water interface.

Mixing coefficients were estimated from profiles of unsupported ^{210}Pb (Fig. 9). Most of the profiles decrease with depth in the exponential pattern expected for downward mixing by a diffusion-like process. Some profiles, on the other hand, exhibit subsurface maxima

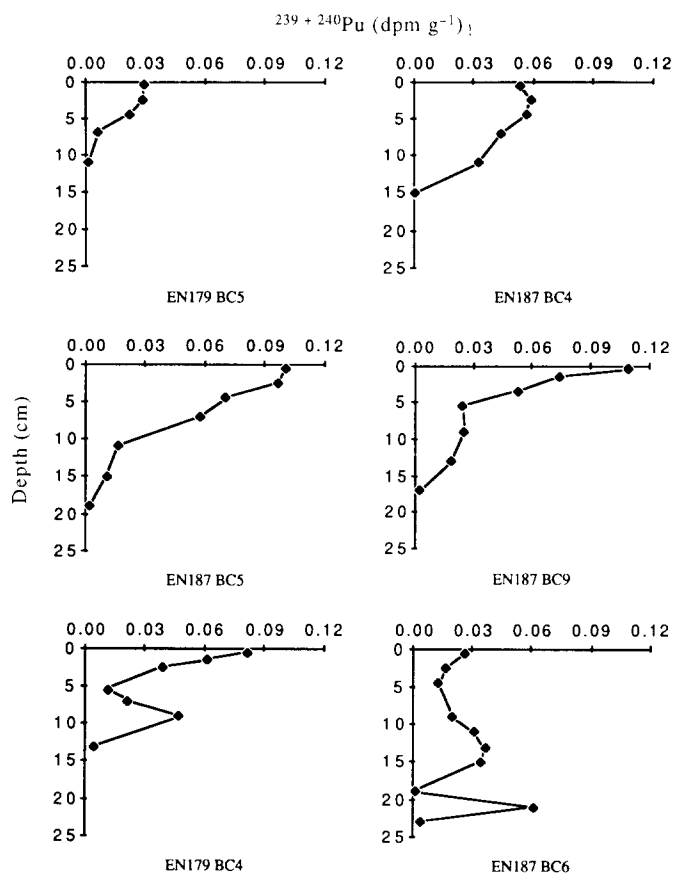


Fig. 10. Profiles of $^{239} + ^{240}\text{Pu}$ in box cores recovered from the continental slope of the southern Middle Atlantic Bight. Cores are arranged from shallowest water depth in the upper left to greatest water depth in the lower right. Locations and water depths are given in Table 1.

that also were noted in earlier studies of MAB sediments (ANDERSON *et al.*, 1988) and attributed to biological mixing in a non diffusion-like manner. Mixing coefficients were estimated from portions of ^{210}Pb profiles that could be fit by exponential gradients according to the relationship:

$$A(z) = A(0) \cdot \exp[-(\lambda/D_B)^{1/2} \cdot z] \quad (3)$$

where $A(z)$ and $A(0)$ represent the activity of $xs^{210}\text{Pb}$ at depth and at the sediment–water interface, respectively, and λ is the radioactive decay constant for ^{210}Pb .

Profiles of ^{210}Pb can be compared to those of another tracer having either a different half life or a different source function to verify that their distributions are, indeed, determined largely by mixing. Selected samples were analyzed for $^{239} + ^{240}\text{Pu}$, which has had a significant input only since the onset of atmospheric testing of thermonuclear weapons in the early 1950s. Similar depths of penetration of Pu and of $xs^{210}\text{Pb}$, as well as the similarity in the shapes of their profiles (compare Figs 9 and 10), including the subsurface maxima, confirm that the $xs^{210}\text{Pb}$ profiles are largely determined by sediment mixing. Implicit in the

use of equation (3) is the assumption that sediment accumulation is negligible over the time scales for which $x\text{s}^{210}\text{Pb}$ is detectable (~ 100 years for these sediments). Sediment accumulation rates are not quite negligible, however, if they are as high as $40\text{--}60\text{ cm ka}^{-1}$ as suggested above. By neglecting the effect of sediment accumulation, the values of D_B calculated using equation (3) will be too large. Because other factors limit the precision of carbon fluxes estimated by this approach (see below), however, more accurate estimates of sediment mixing rates were not required.

Organic carbon profiles exhibit virtually no gradient through the mixed layer in slope sediments of the MAB. This was seen in the SEEP-I region (ANDERSON *et al.*, 1988; ROWE *et al.*, 1988; VENKATESAN *et al.*, 1988) and is similarly evident in profiles from the SEEP-II region (Fig. 7). The absence of a C_{org} gradient, more than the accuracy and precision with which D_B has been estimated, limits the ability to derive carbon remineralization rates using this approach. Some limits can be placed on C_{org} gradients, however, and these, in turn, can be used to set limits on the rate of carbon remineralization in the mixed layer. Values of D_B derived from $x\text{s}^{210}\text{Pb}$ profiles, and upper limits for C_{org} gradients, are summarized in Table 2. Carbon remineralization rates derived from these parameters are $\leq 5\text{ mg C m}^{-2}\text{ day}^{-1}$ (Table 2), consistent with estimates derived from pore water profiles of ammonia and phosphate. The latter tend to be somewhat larger, but also include carbon remineralization below the depth of the mixed layer. While neither approach leads to a precise estimate of carbon remineralization rates, together they indicate that a value of about $4\text{--}6\text{ mg C m}^{-2}\text{ day}^{-1}$ is representative of carbon remineralization within the slope sediments.

Bacteria-mediated remineralization near the sediment–water interface

Estimated total C_{org} remineralization ($41\text{--}59\text{ mg C m}^{-2}\text{ day}^{-1}$) is about an order of magnitude greater than the estimated rate of C_{org} remineralization within the sediment

Table 2. Organic carbon consumption rates within the upper sediment mixed layer estimated by multiplying the organic carbon gradient by the ^{210}Pb -derived mixing coefficient

Core depth/transect		Mixing coefficient	Organic carbon gradient		Carbon consumption
		($\text{cm}^2\text{ y}$)	($\% \text{ cm}^{-1}$)	(mg cm^{-4})	($\text{mg m}^{-2}\text{ day}^{-1}$)
EN187 BC1	265 m N	2.80	<0.01	<0.05	<4
EN179 BC5	384 m N	1.20	<0.03	0.14	4.70
EN187 BC3	410 m N	0.48	<0.01	<0.05	<1
EN179 BC1	417 m N	0.67	<0.01	<0.05	<1
EN187 BC4	512 m N	3.82	<0.01	<0.05	<5
EN179 BC2	892 m N	1.12	0.02	0.10	3.10
EN179 BC3	1031 m N	0.27	0.015	0.075	0.55
EN187 BC5	1045 m N	1.14	0.01	0.05	1.56
EN179 BC4	1318 m N	0.15	<0.01	<0.05	<0.2
EN179 BC7	1989 m N	0.24	negative	negative	<1
EN187 BC6	2000 m N	0.98	<0.01	<0.05	<1.3
EN187 BC10	580 m S	3.50	negative	negative	<5
EN187 BC8	1020 m S	3.18	<0.01	<0.05	<2
EN187 BC9	1165 m S	1.30	0.02	0.10	<3.4
EN187 BC11	1125 m NC	10.80	<0.01	<0.05	<5

(4–6 mg C m⁻² day⁻¹), implying that most remineralization must occur at, or very near, the sediment–water interface. Rates of C_{org} remineralization near the sediment–water interface were estimated from bacterial production based on measurements of tritiated thymidine incorporation. Integrated total bacterial carbon production and production-based remineralization rates are presented in Table 3. Carbon production values for sediment bacteria are likely to underestimate total benthic bacteria production, since production in deeper, anaerobic sediments is not included. However, since previous studies have indicated that bacterial production decreases substantially and rapidly below the sediment surface (MORIARTY, 1986; FALLON *et al.*, 1983; KARL and NOVITSKY, 1988; NOVITSKY and KARL, 1986; KEMP, 1994), the resulting underestimation is likely to be a factor of 2 or less (KEMP, 1990).

Thymidine-based production rates rely on several critical assumptions, including the conversion from cell volume to carbon, and from thymidine incorporation to cell production. The conversion from cell volume remains debatable for water-column bacteria (NAGATA, 1986; LEE and FUHRMAN, 1987; BRATBAK and DUNDAS, 1984; BRATBAK, 1985) and undetermined for sediment bacteria. Of the six or more conversion factors reported in the literature, which range over roughly an order of magnitude, an intermediate value was chosen based on a comparison of independent estimates of bacterial production and oxygen uptake in shelf sediments (KEMP, 1994). However, this conversion factor value may result in overestimates or underestimates of true carbon production if other conversion factors would be more appropriate in slope sediments. The conversion from thymidine incorporation to cell production was based on an empirical procedure in which increases in cell counts in the absence of grazing are compared to concurrent measurements of thymidine incorporation (KEMP, 1994). Since five such comparisons were used to establish the conversion factor, this latter value is reasonably well constrained.

Carbon remineralization was estimated from production assuming that: (1) bacterial growth efficiency is *ca* 50%, such that carbon respired is equal to carbon fixed in biomass; (2) bacterial biomass is in steady state, because bacterial production is entirely consumed and remineralized by microbial and metazoan grazers; and therefore (3) bacteria-mediated carbon remineralization can be reasonably approximated by doubling the

Table 3. Integrated carbon production and estimated remineralization by bacteria in slope sediments in the SEEP-II region. Based on annual-average data in KEMP (1994)

Depth (m)	Production (mg C m ⁻² day ⁻¹)	Remineralization (mg C m ⁻² day ⁻¹)
400	67	135
1000	46	92
2000	35	70

Remineralization is estimated to equal two times the rate of bacterial carbon production assuming a steady-state bacterial population (i.e. grazing equals production). Consequently, "Remineralization" includes the activities of microbial grazers.

estimated bacterial carbon production. Bacterial production rates in slope sediments are approximately equal to the total POC flux to the sediments (compare Table 3, Fig. 3), while the estimated bacterial-mediated C_{org} remineralization exceeds the POC flux. While some uncertainty exists because of the assumptions and approximations discussed above, it is clear that bacteria-mediated C_{org} remineralization in the upper 3–4 mm of sediment is capable of accounting for the 41–59 $\text{mg C m}^{-2} \text{ day}^{-1}$ required to balance the difference between the flux of POC to the sediment, and the sum of carbon burial plus remineralization within the sediment.

Oxygen fluxes

Fluxes of oxygen into the sediments (Table 4) integrate the effects of heterotrophic respiration throughout the sediment column if other electron acceptors are not used or if the reduced forms of those electron acceptors are reoxidized within the sediments by reaction with molecular oxygen. Dissolved ferrous iron was easily detectable in all of the pore water samples collected, and H_2S was smelled when sectioning deeper portions of the subcores. Therefore, some burial of iron sulfides must occur. The contribution by sulfate reduction to total C_{org} remineralization is small, however, with an upper limit for C_{org} remineralization by sulfate reduction set by the upward ammonia fluxes described above.

With the assumptions and conditions described above, total C_{org} remineralization was estimated from measured oxygen fluxes (Table 4) assuming a Redfield stoichiometry for the molar ratio of O_2 consumed to CO_2 produced (138:106). Organic carbon remineralization rates derived from oxygen fluxes measured at water depths of $\sim 1000\text{m}$ (49–70 $\text{mg C m}^{-2} \text{ day}^{-1}$; Table 4) were close to the total POC flux collected by sediment traps at this depth (65 $\text{mg C m}^{-2} \text{ day}^{-1}$; Fig. 3), whereas remineralization rates derived from oxygen fluxes at $\sim 400\text{m}$ were less than the total POC flux.

In estimating remineralization rates using an $\text{O}_2:\text{CO}_2$ stoichiometry of 138:106, it is assumed that all of the remineralized organic N is returned to bottom waters as nitrate. Concentrations of combined N (nitrate plus nitrite) increased in the flux chambers during sediment incubations, but generally by less than the amount expected for quantitative oxidation of organic N to nitrate (Table 4), with the difference most likely attributable to denitrification. Assuming a stoichiometric ratio of CO_2 produced to NO_3^- -consumed of 1.25:1, denitrification leads to an additional carbon consumption amounting to $\sim 4\text{--}6 \text{ mg C m}^{-2} \text{ day}^{-1}$ in the depocenter region (Table 4), a small but non-negligible component of the total organic carbon budget.

Carbon remineralization rates estimated from oxygen fluxes are intermediate between bacteria production rates and the estimated rate of bacteria-mediated C_{org} remineralization at $\sim 1000 \text{ m}$, whereas remineralization rates derived from oxygen fluxes are relatively lower at depths of $\sim 400 \text{ m}$ and $\sim 2000 \text{ m}$ (compare Tables 3 and 4). Oxygen fluxes and bacterial production rates were each measured in multiple seasons, so the effects of temporal variability is reduced to some extent. Beyond that, whether the differences between POC fluxes, bacteria-mediated remineralization rates, and oxygen fluxes reflect spatial and temporal variability, or the different time scales of the different measurements (minutes–hours for bacterial production; hours–days for oxygen flux; ~ 1 year for POC flux), or analytical uncertainties, cannot be resolved at this time.

Despite these differences, however, the various rate estimates are internally consistent to within better than a factor of two, lending confidence to the overall carbon budget for

Table 4. Oxygen fluxes into, and combined N (nitrate plus nitrite) fluxes out of, upper slope sediments of the Middle Atlantic Bight. Computed rates of C_{org} remineralization associated with aerobic respiration and with denitrification, are also provided

Water depth (m)	Oxygen flux* (mMol $m^{-2} day^{-1}$)	Equiv. C flux† (mgC $m^{-2} day^{-1}$)	Combined N‡ (μM N $m^{-2} day^{-1}$)	Expected N‡ (μM N $m^{-2} day^{-1}$)	Denitrification§ (mgC $m^{-2} day^{-1}$)	Equiv. C flux (mgC $m^{-2} day^{-1}$)
130	1.53	14.1	-165.6	-167.9	2.3	0.03
420	1.84	17.0	NA			
436	1.51	13.9	-400.8	-165.5	-235.3	negative
750	7.61	70.1	NA			
792	6.32	58.3	-290.4	-694.0	403.6	6.05
880	5.34	49.2	NA			
1090	6.21	57.3	-362.4	-682.1	319.7	4.80
2010	1.40	12.9	-132.0	-153.6	21.6	0.32

NA = Data not available

*Oxygen and combined nitrogen fluxes were measured *in situ* using incubation chambers, except for the result from 750 m, which was obtained by shipboard incubation of a core.

†Equivalent carbon fluxes associated with aerobic respiration calculated assuming a ratio of O_2/CO_2 of 138/106 (see text).

‡Calculated from the Equivalent C Flux assuming a C/N molar ratio of 7:1 and the quantitative oxidation of organic N.

§Calculated as the difference between measured and expected N flux.

||Estimated rate of C_{org} remineralization via denitrification computed assuming an average ratio of 1.25 moles of CO_2 produced per mole of nitrate consumed.

the depocenter, which can be summarized as follows. Particulate organic carbon is deposited on the slope at a rate of about $65 \text{ mg C m}^{-2} \text{ day}^{-1}$, of which $6\text{--}24 \text{ mg C m}^{-2} \text{ day}^{-1}$ is buried and $\leq 5 \text{ mg C m}^{-2} \text{ day}^{-1}$ is remineralized within the sediments. The remainder of the POC flux is rapidly remineralized by microbial processes within a few mm of the sediment-water interface, where molecular oxygen is the principal electron acceptor for organic carbon oxidation.

Rate constants for organic carbon remineralization

Particulate organic carbon supplied to depocenter sediments consists of two general components, one that is extremely labile and almost completely consumed within the upper few mm of sediments, and one which is refractory. The absence of gradients in the C_{org} profiles (Fig. 7) indicates that little of the refractory component is remineralized on the time scale of the residence time of material in the mixed layer. Mixed layer depths range from about 15–20 cm, as indicated by the depth of penetration of $x\text{s}^{210}\text{Pb}$ (Fig. 9). With accumulation rates of $40\text{--}60 \text{ cm ka}^{-1}$, the residence time of sediment within the mixed layer ranges from 250–500 years. The mean life of the refractory component of the C_{org} , which includes virtually all of the C_{org} present in the mixed layer (Fig. 7), must then be at least several times greater than the 250–500 year residence time of sediments in the mixed layer.

Much of the particulate organic carbon supplied to depocenter sediments at $\sim 1000 \text{ m}$ is biologically labile and is rapidly remineralized, as indicated by the abrupt decrease in C_{org} content from $\sim 4\%$ in sediment trap samples at 1000 m (BISCAYE and ANDERSON, 1994) to $\sim 2\%$ in surface sediments (Fig. 7). Further constraints can be placed on the rate constants for remineralization of labile C_{org} using a numerical model for sediment mixing (see ANDERSON *et al.*, 1988 for details). Input parameters for the sediment mixing model in this illustration include a mass accumulation rate of $0.04 \text{ g cm}^{-2} \text{ y}^{-1}$ ($1096 \text{ mg m}^{-2} \text{ day}^{-1}$), a mixed layer depth of 15 cm, a sediment mixing rate of $1.0 \text{ cm}^2 \text{ y}^{-1}$, consistent with rates derived from $x\text{s}^{210}\text{Pb}$ profiles (Table 2), and a porosity of 0.8 which is representative of cores with negligible sand content (water depths $>500 \text{ m}$; Appendix). Sedimenting particles are assumed to have a C_{org} content of 4.5%, of which 2.5% is refractory and 2.0% labile. The amount of refractory C_{org} is arbitrary, as it is the gradient with depth of the labile component that establishes the rate constant for carbon remineralization.

Model runs were carried out using a range of rate constants, and outputs were compared with measured C_{org} profiles to constrain the rate constants for remineralization of labile C_{org} . Model outputs (Fig. 11) indicate that carbon enrichment in surface sediment would be detectable (a few tenths of a per cent increase in C_{org}) if the mean life of the labile C_{org} were >1 year. The absence of such an enrichment in measured profiles (Fig. 7) indicates that the mean life of the labile C_{org} must be <1 year, consistent with the results of laboratory studies of labile C_{org} remineralization rates (WESTRICH and BERNER, 1984). There exists, then, an extreme dichotomy between the labile C_{org} , with a mean life <1 year, and the refractory C_{org} which has a mean life much greater than 250–500 years. Organic matter with an intermediate mean life (e.g. 10–100 years), furthermore, does not contribute significantly to the total POC flux to depocenter sediments.

Nature and source of carbon supplied to the depocenter

Carbon-14 was used as a tracer to help elucidate the nature and origin of the refractory carbon supplied to upper slope sediments of the southern Middle Atlantic Bight. A

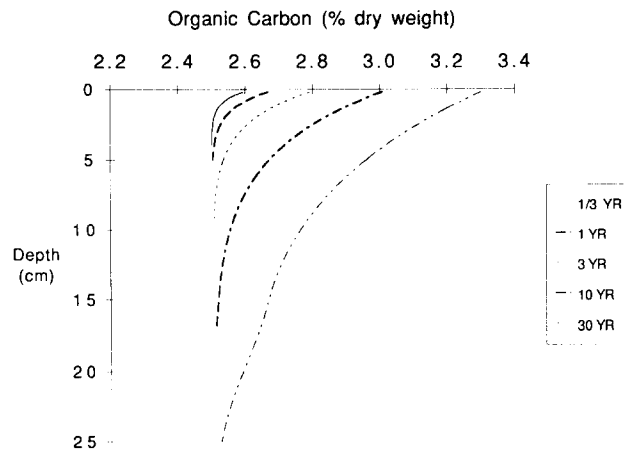


Fig. 11. Model-generated profiles of organic carbon in upper slope sediments using the mean annual POC flux measured by the near-bottom sediment trap at Mooring 10 as input, assuming that the particulate material is 2.5% by weight refractory carbon and 2% by weight labile carbon. The mean life of the labile carbon for each model run is shown in the key. Other input parameters include: a sediment mixing rate of $1.0 \text{ cm}^2 \text{ y}^{-1}$, which several ^{210}Pb profiles indicate is representative of the slope sediments; a mass accumulation rate of $0.04 \text{ g cm}^{-2} \text{ y}^{-1}$, and a porosity of 0.8.

detailed record of results obtained by AMS dating of samples from one box core showed an average age of 1560 years ($\Delta^{14}\text{C} = -177\text{‰}$) for C_{org} in surface sediments (Fig. 12). Aging occurs as carbon resides in the mixed layer, and it is necessary to correct for this to derive the age of the refractory C_{org} at the time it is supplied to the depocenter (NOZAKI *et al.*, 1977) using the relationship:

$$R_o = R_m \cdot [1 + \lambda \cdot (L/S)] \quad (4)$$

where R_m is the $^{14}\text{C}/^{12}\text{C}$ ratio of C_{org} in mixed layer sediments, R_o is the decay-corrected ratio, λ is the radioactive decay constant of ^{14}C , and L is the depth of the mixed layer (15

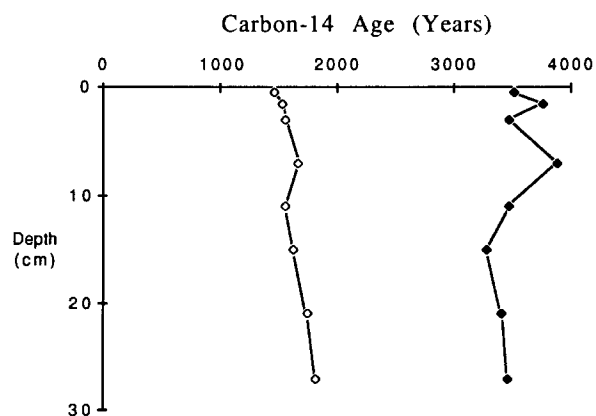


Fig. 12. Detailed ^{14}C profiles of organic (open symbols) and carbonate (solid symbols) carbon in samples from EN187 BC9 collected at a depth of 1165m on the southern transect.

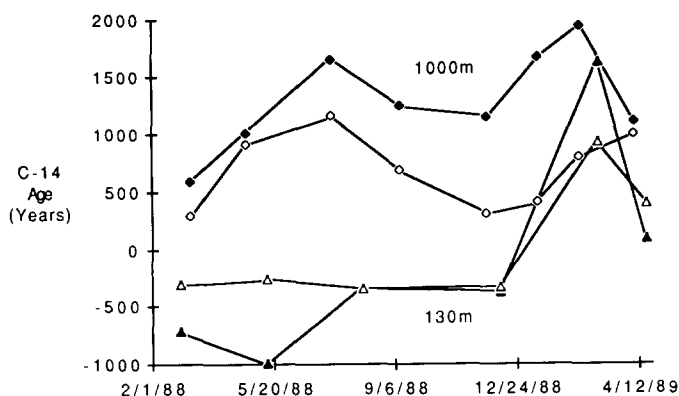


Fig. 13. Carbon-14 age of SEEP-II time-series sediment trap samples from two traps deployed on a mooring at the 1000-m isobath, within the depocenter region. Open symbols represent organic carbon, solid symbols represent CaCO_3 . Negative ages represent samples containing bomb radiocarbon. Ages have not been corrected for reservoir age of dissolved inorganic carbon in seawater. Triangles represent samples from a trap at ~ 130 m, diamonds represent samples from a trap placed 10 m above bottom. Samples were composited to obtain sufficient material to measure ^{14}C by accelerator mass spectrometry. Midpoints of the sampling intervals for shallow and deep traps do not necessarily coincide because it was necessary to composite more shallow samples than deep samples.

cm for EN187 BC9). For S of 60 cm ka^{-1} (based on results from EN187 GC5 collected near this location), R_o is calculated to be -151% , corresponding to an age of 1320 years, much older than the pre-bomb age of DIC in surface seawater (400–600 years). Carbonate carbon is systematically older than organic carbon in recent sediments of the region (Fig. 6), and the average ^{14}C of carbonate in surface sediments of EN187 BC9 (Fig. 12) exceeds the expected age of material with a pre-bomb modern source by more than 2000 years. Significant components of the fluxes of both organic and inorganic carbon to the depocenter, therefore, must be derived from reworking of older source material.

Carbon-14 profiles show no surface inflection indicative of accumulation of significant amounts of carbon tagged with bomb ^{14}C (Fig. 12). Evidently, modern C_{org} labelled with bomb ^{14}C in surface waters (see below) is rapidly and completely remineralized at the sediment–water interface. This is consistent with the conclusion that labile C_{org} has a mean life of <1 year (see above), but further suggests that so little “fresh” C_{org} escapes remineralization that the bomb signal, present in the ocean for the past ~ 30 years, has yet to leave its mark on depocenter sediments.

Carbon collected by the sediment trap deployed at a depth of ~ 130 m on Mooring 10 (~ 1000 m total water depth; Fig. 3) is clearly labelled with bomb- ^{14}C (Fig. 13). Organic material collected in the first four composite samples had a flux-weighted mean $\Delta^{14}\text{C}$ of $+35\%$ (i.e. a negative age of -280 years because of the presence of bomb ^{14}C). Particles collected by the near-bottom trap at this location, however, had a flux-weighted mean $\Delta^{14}\text{C}$ -org of -85% (700 years). Particulate organic carbon supplied by lateral transport to the depocenter, which is responsible for the increase with depth in C_{org} flux (Fig. 3), is on average, therefore, quite old.

In contrast to the first four samples from the 130-m trap at Mooring 10, the fifth sample contained much older carbon (Fig. 13). Both the organic and carbonate fractions of this

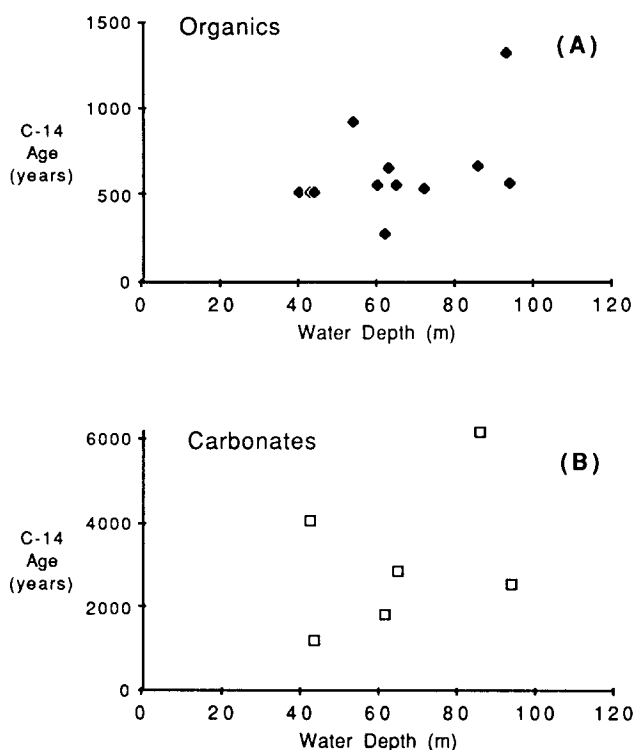


Fig. 14. Carbon-14 ages of fine-grained material separated from shelf sands collected along both SEEP-II transects plotted against water depth from which they were collected. Ages for organic and carbonate carbon are plotted on different scales.

sample had ages similar to those of the particles collected by the deep trap. The fifth sample from the shallow trap was collected during a period of maximum mass flux that immediately followed a massive resuspension event, in February 1989, on the adjacent shelf (BISCAYE and ANDERSON, 1994). Resuspended material must have been transported in surface waters out to the location of Mooring 10 (1000-m isobath) and in sufficient quantity to have dominated the flux of carbon collected at that time by the trap at 130 m.

The greater ages of CaCO_3 , compared to C_{org} , found in the 130-m trap following the resuspension event (Fig. 13), as well as the systematically greater ages of CaCO_3 throughout the record of the deep trap (Fig. 13), and in the sediments (Figs 6 and 12), provides a tracer with which to identify the source of the old reworked carbon deposited on the upper slope. Fine-grained material separated from shelf sands was analyzed to determine whether or not it has a ^{14}C content that is consistent with such material being the source of the old carbon supplied by lateral transport to the depocenter. Inner shelf sediments contain very little fine-grained material, but at depths >40 m, $1\text{--}10\text{ mg cm}^{-3}$ of fines could be separated from the sands in a settling tube, with amounts of recoverable fines generally increasing with increasing water depth. Carbon-14 results (Fig. 14) showed considerable scatter, but many of the organic fractions had ages of 500–700 years, whereas the carbonates were much older, perhaps reflecting a source from weathering of marine deposits left behind during Pleistocene high sea stands.

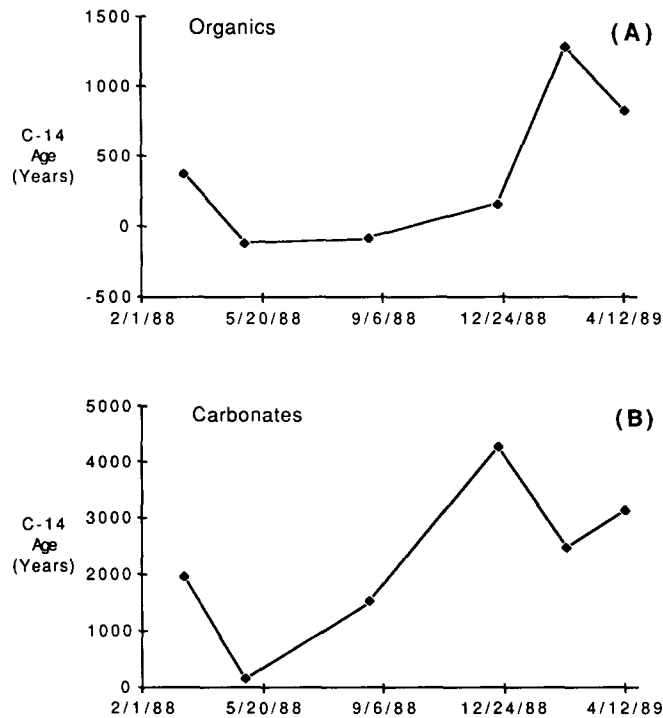


Fig. 15. Carbon-14 ages of particles collected by time-series sediment traps deployed 10 m above bottom in ~ 130 m water depth at the shelf edge along the southern SEEP-II transect (Mooring position 8 in Fig. 3). Ages for organic and carbonate carbon are plotted on different scales. The fifth sample in the sequence collected particles during the massive resuspension event that occurred on the shelf during February 1989.

Resuspended shelf sediments were also “sampled” by a sediment trap located ~ 10 m above bottom near the shelf edge (Mooring 8, ~ 130 m water depth; see position in Fig. 3). Traps at this location often collected old carbon, with carbonates being much older than the C_{org} (Fig. 15), just as in the shelf fines. During the massive resuspension event of February 1989, the trap at this location collected a flux of resuspended material that averaged more than $20 \text{ grams m}^{-2} \text{ day}^{-1}$ throughout a 17-day collection interval. Resuspended sediment collected by this trap included sand-sized material that would not have been transported as far as the depocenter, where the sediments are fine grained. Some physical sorting would have occurred, therefore, during the transport of this material out to the 1000-m isobath, and the sorting may have influenced the ages of the carbonate and organic phases exported to the slope. Nevertheless, the ages of the organic and carbonate fractions, and the difference between their ages, for both shelf fines and for the resuspended material, are consistent with this material acting as a source for the old carbon deposited on the upper slope.

A ^{14}C budget was constructed (Fig. 16) for the depocenter region (in this case, limited to the location of Mooring 10 at ~ 1000 m) to estimate the average age of POC supplied by lateral transport as well as the average age of remineralized C_{org} . Organic carbon produced in surface waters had an average $\Delta^{14}\text{C}$ of $+35\%$, based on those samples

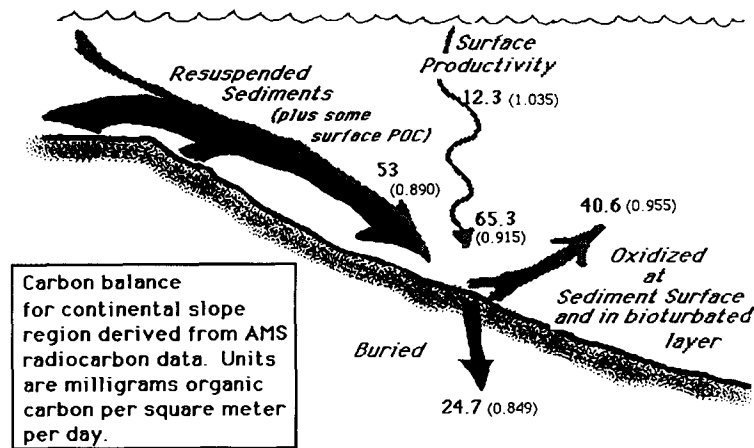


Fig. 16. Summary of the organic carbon budget at the location of SEEP-II Mooring 10 in the depocenter region ($\sim 1000\text{-m}$ isobath) of the southwestern Middle Atlantic Bight. Carbon fluxes, in $\text{mg C m}^{-2} \text{ day}^{-1}$, are shown in bold numbers. $\Delta^{14}\text{C}$ activities, in fraction modern, are given in parentheses. The fluxes of POC out of surface waters (12.3) and to the sediments (65.3) were measured with sediment traps. Burial was evaluated using accumulation rates derived from ^{14}C dating of EN187 GC5 and $\Delta^{14}\text{C}$ values measured in EN187 BC9. ^{14}C activities associated with these fluxes are measured values. Other fluxes, and their associated $\Delta^{14}\text{C}$ values, are estimated by difference.

collected by the 130-m trap at Mooring 10 which were not obviously influenced by the resuspension event. The time-weighted mean annual flux of POC collected by this trap was $12.3 \text{ mg C m}^{-2} \text{ day}^{-1}$. The near-bottom trap at this location ($\sim 1000 \text{ m}$) collected a time-weighted mean POC flux of $65.3 \text{ mg C m}^{-2} \text{ day}^{-1}$ with a mean $\Delta^{14}\text{C}$ of -85% (^{14}C age of 700 years). If the POC collected by the 130-m trap settled quantitatively and without alteration to 1000 m, then by difference the POC supplied by lateral transport amounted to a flux of $53 \text{ mg C m}^{-2} \text{ day}^{-1}$ and it must have had a mean $\Delta^{14}\text{C}$ of -110% (960 years).

Organic carbon supplied to the depocenter, as represented by the material collected by the trap at $\sim 1000 \text{ m}$, consists of both refractory and labile components. The flux and $\Delta^{14}\text{C}$ of the labile (remineralized) component can be estimated by difference if the corresponding values for refractory C_{org} are known. Refractory C_{org} was estimated (see above) to have a $\Delta^{14}\text{C}$ of -151% (1320 years) at the time of its deposition on the sea floor. Sediments accumulating at 60 cm ky^{-1} (EN187 GC5; see above) having 2% C_{org} and a porosity of 70% represent a C_{org} burial of $24.7 \text{ mg C m}^{-2} \text{ day}^{-1}$. If remineralization of refractory C_{org} is negligible between the time of its deposition on the sea floor and its burial below the mixed layer, then from the above information it is inferred that refractory C_{org} is supplied to the depocenter at a flux of $24.7 \text{ mg C m}^{-2} \text{ day}^{-1}$ with a mean $\Delta^{14}\text{C}$ of -151% . The difference between this and the total flux of POC to the sediments ($65.3 \text{ mg C m}^{-2} \text{ day}^{-1}$ with a mean $\Delta^{14}\text{C}$ of -85%) leads to an estimated rate of C_{org} remineralization of $40.6 \text{ mg C m}^{-2} \text{ day}^{-1}$ with a mean $\Delta^{14}\text{C}$ of -44.8% (370 years) for the remineralized C_{org} . At most, $12.3 \text{ mg C m}^{-2} \text{ day}^{-1}$ of the remineralized C_{org} is supplied by the flux of POC out of the overlying euphotic zone (based on the flux collected by the trap at 130 m), implying that $28.3 \text{ mg C m}^{-2} \text{ day}^{-1}$ ($40.6 - 12.3$) of remineralized C_{org} is supplied by lateral transport.

Labile C_{org} , therefore, constitutes about half (28.3 vs 53 $mg\ C\ m^{-2}\ day^{-1}$) of the total C_{org} supplied to the depocenter by lateral transport.

Each of the parameters in this budget has associated with it an inherent uncertainty with contributions both from analytical precision and from natural variability. Analytical precision of individual AMS ^{14}C measurements ranged between 1 and 3%, but there are not sufficient data to adequately constrain natural variability, which is much larger. Consequently, only best estimates for the calculated fluxes, rates and $\Delta^{14}C$ values are presented. Two noteworthy results are evident, however. First, the refractory C_{org} left behind after remineralization of the labile component is much older ($\Delta^{14}C = -151\text{‰}$; age ~ 1320 years) than the pre-bomb $\Delta^{14}C$ of DIC in surface waters, indicating that the refractory C_{org} originates by reworking of older material. Second, labile C_{org} remineralized at the depocenter ($\Delta^{14}C = -44.8\text{‰}$; age ~ 370 years) consists of a mixture of pre-bomb ($\Delta^{14}C = -50$ to -80‰ ; TANKA *et al.*, 1990) and bomb-labelled carbon ($+35\text{‰}$ at the time of the SEEP-II field program). It is not clear how C_{org} old enough to have a pre-bomb ^{14}C activity can be consumed in 1 year once it is deposited on the slope. One possibility, which is only speculation at this point, is that some of the labile carbon supplied to the depocenter by lateral transport represents secondary production of labile compounds by bacterial reworking of older carbon on the shelf. Bacterial production can recapture old DOC, as well as POC, and introduce it into the food web where it can be transformed into non-bacterial POC. Given the apparent old age of DOC (WILLIAMS and DRUFFEL, 1987), incorporation of a small component of recaptured DOC into the flux of exported POC could account for the pre-bomb age estimated for the labile component of exported POC.

These findings are consistent with, and extend, work carried out during SEEP-I, where it was also concluded that much of the C_{org} in slope sediments is refractory based on the observations that concentrations of C_{org} , as well as of specific organic compound classes, show little vertical gradient in the upper sediment column (VENKATESAN *et al.*, 1987, 1988; STEINBERG *et al.*, 1987; ANDERSON *et al.*, 1988; ROWE *et al.*, 1988). ROWE *et al.* (1988) compared standing stocks and turnover rates of benthic biomass with sediment inventories of C_{org} on the shelf and slope of the SEEP-I region, from which they concluded that most of the C_{org} in sandy shelf sediments is readily utilizable, but most of the C_{org} in slope sediments is "sufficiently depleted or altered that its value as nourishment has declined appreciably." Tracers of terrestrial C_{org} are present in slope sediments (VENKATESAN *et al.*, 1987), but the bulk of the C_{org} seems to be of marine origin based on its average $\delta^{13}C$ of -20 to -21‰ (TANAKA *et al.*, 1991). Organic C/N ratios (Fig. 17) are also consistent with a marine source for most of the sedimentary C_{org} . It is somewhat surprising, in fact, that the refractory material present in the slope sediments has a C/N ratio similar to that of the particles collected by the shallowest traps deployed at Moorings 7 and 10 (1000-m isobath), which had flux-weighted mean organic C/N ratios of 8.5 and 8.0 respectively (BISCAYE and ANDERSON, 1994). Refractory carbon, whatever its composition, does not appear to be very depleted in nitrogen compared to fresh POC settling out of the euphotic zone.

Carbon export from the shelf

Lateral supply of POC to the depocenter was estimated above to be 53 $mg\ C\ m^{-2}\ day^{-1}$ at the location of Mooring 10. Including the results from Mooring 7 at ~ 1000 m on the north transect (Fig. 3), where ^{14}C measurements were not made, increased this value

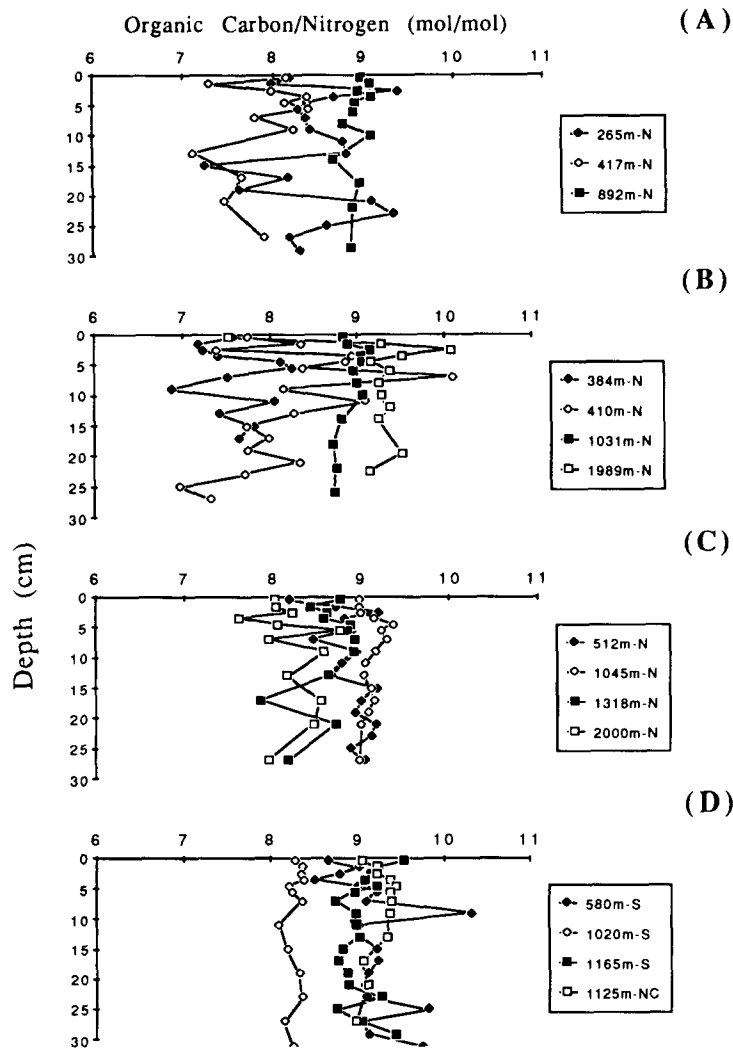


Fig. 17. Organic C/N ratio in sediments of the southern Middle Atlantic Bight. Panels A–C represent cores from the northern SEEP-II transect; panel D contains three profiles from the south transect and one from the Norfolk Canyon.

slightly to $55 \text{ mg C m}^{-2} \text{ day}^{-1}$. Lateral supply of POC to sediments at 400 m may have been greater than at 1000 m, as suggested by fluxes at Mooring 9 (Fig. 3) and by bacterial production rate measurements (Table 3), or less, as suggested by fluxes at Mooring 6 and by oxygen fluxes (Table 4). At water depths of $\sim 2000 \text{ m}$, bacterial production rates and oxygen fluxes both indicate lower POC fluxes to the sediments than at 1000 m, consistent with measurements in the SEEP-I region which showed POC fluxes between the 2000-m and 3000-m isobaths to be \sim half of those at $\sim 1000 \text{ m}$ (BISCAYE *et al.*, 1988). As the purpose here is to provide an upper limit for the export of POC from the MAB shelf, the lateral contribution to the POC flux at 1000 m ($55 \text{ mg C m}^{-2} \text{ day}^{-1}$) will be used to represent the lateral supply of POC to the entire upper slope region out to a water depth of 2000 m.

Particulate matter supplied to the depocenter may be carried from the north by along-shore currents as well as by transport off the shelf. It is not possible to estimate the supply of POC by along-shore transport, but an upper limit for shelf export is obtained by assuming that net along-shore transport is negligible.

Average annual primary production in shelf waters of the southern MAB is $\sim 800 \text{ mg C m}^{-2} \text{ day}^{-1}$ (P. Falkowski, personal communication). The width of the slope out to the 2000-m isobath in this region is about 1/3 to 1/5 of that of the shelf. Taking into account the smaller area of the slope, the lateral supply of POC to the depocenter ($55 \text{ mg C m}^{-2} \text{ day}^{-1}$) can represent no more than $\sim 2\%$ of the average annual production of POC on the shelf ($800 \text{ mg C m}^{-2} \text{ day}^{-1}$ times the areal concentration factor of 3 to 5). Because much of the total POC flux to the depocenter consists of refractory C_{org} reworked from older deposits, i.e. it is not recent phytoplankton detritus that has escaped consumption on the shelf (e.g. WALSH *et al.*, 1985), the upper limit of 2% of shelf production exported to the depocenter must be reduced further if applied strictly to carbon of recent origin. The carbon budget of FALKOWSKI *et al.* (1988) permitted 10–20% of shelf productivity to be exported. Only a small portion of this is deposited on the upper slope. If a much larger portion of the POC fixed on the shelf is exported to the deep sea, then it must be exported by another mechanism, such as: conversion to DOC, advection offshore with the shelf waters themselves at Cape Hatteras, or funnelling downslope through canyons that incise the MAB margin.

Supply of refractory organic carbon to sediments at other margins

Are results presented here, indicating a substantial flux of old, reworked, refractory carbon to slope sediments of the Middle Atlantic Bight, representative of conditions at other margin sites? Several lines of evidence point to an affirmative answer. PEDERSEN *et al.* (1992) described a similar situation where winnowing processes supplied degraded C_{org} to slope sediments on the Oman margin (Arabian Sea). CARPENTER (1987) published organic carbon profiles from the upper slope off the coast of Washington State which look remarkably similar to those reported here, with uniform carbon contents of 2–3% in the mixed layer (roughly the upper 10 cm). SMITH *et al.* (1987) found a uniform, but higher (6%), organic carbon content in the upper 10 cm of Santa Catalina Basin sediments, while MURRAY and KUIVILA (1990) reported nearly uniform organic carbon contents of between 2 and 3% for two cores from the slope off Monterey. A complete evaluation of the refractory and labile components of the POC flux to these margin sediments would require, in addition to the sediment carbon profiles, sediment trap samples to provide the source term for remineralization models. These data do not exist, to our knowledge, but the uniform and high carbon concentrations in mixed layer sediments at these eastern boundary sites suggests that a substantial amount of the POC flux to these sediments may also be refractory material.

Several studies of C_{org} remineralization in sediments of the borderland basins off southern California have found that an unusually large fraction of the POC flux to the sediments survives remineralization to be buried. BERELSON *et al.* (1987) found that $>60\%$ of the POC flux to sediments of San Pedro and San Nicholas Basins is buried. BENDER *et al.* (1989) estimated a C_{org} burial efficiency of 36% in San Clemente Basin. JAHNKE (1990) reported that 43% of the C_{org} arriving at the floor of Santa Monica Basin is buried. Furthermore, JAHNKE (1990) determined that, much like the situation in the MAB, the

POC flux contains a minimum of two distinctly different components of organic matter, one remineralized rapidly with a half-life <6.5 years, and one remineralized on a much longer time scale (~1700 years).

Organic carbon concentrations in marine sediments are determined by the rate of C_{org} supply, dilution by other phases, and the efficiency of C_{org} preservation (or burial). Active debate has occurred in recent years over conditions which regulate the last of these factors: C_{org} preservation. Principal candidates include bottom-water oxygen content, carbon flux and bulk sediment accumulation rate (e.g. DEMAISON and MOORE, 1980; EMERSON, 1985; HENRICHs and REEBURGH, 1987; EMERSON and HEDGES, 1988; PEDERSEN and CALVERT, 1990; BETTS and HOLLAND, 1991; CALVERT and PEDERSEN, 1992). Attempts to link C_{org} enrichments in sediments to conditions of reduced oxygen content of bottom waters (e.g. DEMAISON and MOORE, 1980; EMERSON, 1985) have met with rebuttals demonstrating that C_{org} oxidation rates and C_{org} burial efficiencies generally show no correlation with bottom water oxygen levels (e.g. HENRICHs and REEBURGH, 1987; JAHNKE, 1990; CALVERT and PEDERSEN, 1992; REIMERS *et al.*, 1992). A number of investigators have shown good correlations between C_{org} preservation efficiency and sediment accumulation rate (see BETTS and HOLLAND, 1991, for a recent review), and various hypotheses have been put forward to explain the means by which increasing the sediment accumulation rate sequesters, or protects, otherwise metabolizable C_{org} .

Results presented here show clearly that the quality (lability and age) of the C_{org} supplied to the sediments is a critical factor contributing to the high C_{org} preservation efficiency in slope sediments. Ocean margin sediments provide the main body of evidence showing the strongest correlation between C_{org} preservation efficiency and sediment accumulation rate (rates ranging from 1 to 100 cm ky^{-1} ; see Fig. 4 in BETTS and HOLLAND, 1991). Rates of supply to slope sediments of refractory C_{org} and of bulk solids both depend on the strength and character of winnowing processes, and this mutual dependency may be responsible, at least in part, for the good correlation between the apparent C_{org} preservation efficiency (reflecting the source of reworked, refractory C_{org}) and sediment accumulation rate.

CONCLUSIONS

Most of the carbon flux to upper-slope sediments in the Middle Atlantic Bight is supplied by lateral transport rather than by settling from the overlying euphotic zone. However, even if all of this carbon is exported from the adjacent shelf, it represents the fate of no more than 2% of the mean annual productivity of the adjacent shelf.

Much of the organic carbon supplied to the slope is refractory (i.e. not easily metabolizable). It has a mean life in the sediments with respect to biological remineralization of at least 1000 years, and it arrives at the slope depocenter with an average ^{14}C age of about 600–900 years pre-bomb. Admixture of a small amount (~15–20%) of dead (e.g. >40,000 years) carbon with modern with bomb ^{14}C -tagged carbon would result in a similar average age; however, the absence of gradients in either percentage of C_{org} or in its ^{14}C activity in the upper-sediment column suggests that this carbon is uniformly old and refractory. Upper-slope sediments accumulate with a high C_{org} content (~2% in the southern MAB) because the carbon deposited on the slope is refractory, rather than because of an unusually large rate of carbon supply.

Most of the labile C_{org} flux to upper slope sediments is also supplied by lateral transport.

Labile C_{org} is rapidly remineralized, with a mean life with respect to biological consumption of no more than ~ 1 year, yet part of the remineralized C_{org} seems to have a pre-bomb ^{14}C activity. It is not clear how C_{org} with a pre-bomb ^{14}C activity can be consumed in 1 year once it is deposited on the slope. One possibility is that much of the labile carbon supplied by lateral transport represents secondary production by bacterial reworking of older carbon on the shelf.

Carbonate carbon is systematically older than coexisting organic carbon in sediments and in particulate matter collected by near-bottom sediment traps throughout the shelf-slope system of the southern Middle Atlantic Bight. Carbon-14 ages of the organic and carbonate fractions of fine-grained material present within sandy shelf sediments are consistent with this material acting as a source of old carbon supplied to the slope. Old carbon resuspended from the shelf was collected by a sediment trap deployed in shallow water (~ 130 m) over the upper slope depocenter (1000-m isobath) following one large resuspension event. A trap at the shelf-edge collected old carbon at other times as well, however, suggesting that major "events" are not necessary to induce export of old particulate carbon from the shelf.

A growing body of evidence suggests that the winnowing, transport, and deposition of reworked and refractory C_{org} is a principal factor contributing to the high C_{org} contents, and to the high apparent C_{org} preservation efficiencies, typical of ocean margin sediments. A lack of appreciation for the contribution of reworked C_{org} to slope sediments has added confusion and controversy to the study of the factors, processes, and conditions which regulate the preservation efficiency of C_{org} in marine sediments.

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