DOE/ER/61415--T1

DE-FG02-92ER61415

BURIAL, REMINERALIZATION AND UTILIZATION OF ORGANIC MATTER AT
THE SEA FLOOR UNDER A STRONG WESTERN BOUNDARY CURRENT

Final Report

For Period: 1 May 1992 - 30 April 1995

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24 August 1995

Prepared for:

The U. S. Department of Energy

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ABSTRACT

The overall goals of this project were to quantify the rates of organic carbon export from the southern mid-Atlantic Bight and to quantify the rates at which carbon is exchanged between the inorganic and organic pools within the bottom sediments. This information is necessary to constrain the role of the oceans in the control of carbon dioxide released to the atmosphere in association with energy production.

During this project, *in situ* benthic flux chamber incubations have been performed at six sites on the continental slope and rise adjacent to Cape Hatteras. Based on the analysis of the time-series samples recovered during each experiment, the sea floor exchange rates of the major biogenic elements, oxygen, carbon, nitrogen, phosphorus and silicon were calculated. From the estimated benthic flux rates and the ancillary pore water and sediment analyses, the deposition, remineralization and burial rates of organic carbon to the sea floor in this area was evaluated. This information has been incorporated into regional (attachment 1) and global (attachment 2) assessments of organic carbon fluxes to the deep sea.

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Attachment 1.

**PARTICLE FLUX AND ITS CONVERSION TO THE SEDIMENT RECORD:
COASTAL OCEAN UPWELLING SYSTEMS**

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Keywords: Burial, Deposition, Coastal Sedimentation

(1995) in *Upwelling in the Ocean: Modern Processes and Ancient Records*, C. P. Summerhayes, K.-C. Emeis, M. V. Angel, R. L. Smith and B. Zeitschel, eds. John Wiley and Sons, pp. 83-100.

Abstract

Despite previous research efforts, fundamental questions remain concerning the processes that determine the deposition and preservation of the major biogenic phases (organic carbon, CaCO_3 , and opal) in sediments underlying coastal upwelling systems. Our ability to interpret the accumulation of these phases in sediments is currently hindered by uncertainties in: 1. estimating the contributions of different sources of organic matter to a particular coastal location; 2. the processes and sea floor conditions controlling the remineralization and preservation of organic matter in sea floor sediments; 3. the linkages between sedimentary remineralization of organic matter and the dissolution of CaCO_3 ; and 4. the reactivity and solubility of biogenic opal at the sea floor. These fundamental limitations obviate accurate interpretation of variations in the sedimentary distributions of the major biogenic phases and paleoceanographic reconstructions will continue to depend on the interpretation of surrogate and tracer distributions until such uncertainties are resolved.

INTRODUCTION

Coastal upwelling systems play a uniquely important role in the biological productivity and chemical cycling within the oceans. While comprising a small portion of the sea surface area, continental margin regions account for greater than 50% of marine productivity and for more than 95% of the sedimentary accumulation of organic carbon (Romenkovich, 1984). Because of the large contribution of these systems to marine biogenic cycles, knowledge of variations in the biogenic production rates places important constraints on past oceanographic processes and conditions. Sediments underlying coastal upwelling systems display a variety of downcore patterns suggesting that they record the variability of the overlying upwelling.

However, relating the concentrations of biogenic phases preserved in sediments to overlying water column processes and rates of primary biological productivity is complicated by a variety of physical and diagenetic processes which may alter the primary deposition or may cause post-depositional changes in the materials. Accurate interpretation of the sediment record requires an understanding of the processes that control transport and input of biogenic materials to the sediments and transform the deposited materials into the permanent sediment record.

The sediments that comprise the geologic record under coastal upwelling systems represent the net balance between processes that supply particles to a location versus those that remove or alter them, either physically through resuspension and lateral and down-slope transport or diagenetically by degradation and dissolution (Figure 1). The relative importance of these different processes in controlling burial depends on a variety of factors. For example, river input depends on river discharge and rate of sea level rise. Shelf resuspension and subsequent lateral transport will depend on local winds. Depending on the water depth of deposition, resuspension and down-slope transport may also depend on wind strength as well as internal waves and subsurface currents. Many of these same factors also control or influence upwelling intensity and biological productivity. Thus, attempts to reconstruct past variations in biological productivity and upwelling must also assess to what extent these other processes have been altered and how they may influence the sediment signals.

In the following, we review the factors that determine the particle flux to the sea floor and control the characteristics of the sediments preserved in the geologic record. This includes a brief presentation of the sources of materials available for coastal sedimentation, followed by a discussion of the transport processes that control net deposition rates and the diagenetic processes which alter deposited materials. We focus the discussion on the major biogenic

phases; organic carbon (OC), CaCO_3 , and opal, as these potentially represent the most direct linkage between sediment characteristics and the processes that control biological productivity in the overlying coastal upwelling system.

SOURCES

In the continental margin, biogenic particles, especially OC, are supplied to bottom sediments from a variety of sources. These include particles produced in the overlying photic zone or in adjacent near-shore and estuarine waters, erosion from continental shelf sediments, and terrestrial sources, transported either by river discharge or atmospheric transport. OC derived from these various sources differs in isotopic and compositional characteristics. For example, characteristics used to identify sources include: stable carbon isotope ratios, C:N ratios of the organic matter, and specific molecular-level biomarker tracers such as plant wax n-alkanes (C₂₇, C₂₉, C₃₁) (Prahl, et al 1989) and lignin phenols (Hedges and Mann, 1979). The measureable differences in these characteristics provide the means by which the relative contributions of the different sources to a single location can be made.

Numerous examples have been reported where accumulating OC has been attributed to multiple sources in coastal systems. Anderson et al. (in press) suggest that as much as 1/2 of the OC accumulating on the upper slope of the Mid-Atlantic Bight (Eastern Seaboard of the U.S.) may be traced to old, refractory OC winnowed from the adjacent shelf sediments. They suggest that the deposition of a sizable fraction of old and refractory OC may account for the relatively old ^{14}C ages obtained from core top OC and for the relatively high burial efficiency (38%), despite bottom water O_2 concentrations of approximately 250 μM .

Based on plant wax n-alkane measurements of sediment trap and core samples, Lyle et al.

(1992) concluded that lateral inputs of terrestrial OC supplied approximately 20% of the OC accumulating in sediments adjacent to the Oregon coastal upwelling region. These authors report that the proportion of terrestrial OC did not vary dramatically along their entire study transect which extends from 100 km to greater than 600 km from the coast. These authors do report, however, that the proportions of marine and terrestrial OC varied between the last glacial maximum and the present suggesting temporal variations in the physical processes controlling these inputs.

Also based on plant wax n-alkane biomarkers measurements, Westerhausen et al. (1993) have concluded that more than 60% of the OC on the shelf off East Liberia is of terrestrial origin while the proportion drops to less than 20% on the adjacent slope. In general, these more recent studies based on plant wax biomarkers suggest that terrestrial systems contribute a larger proportion of the OC to marine systems than were reported in early studies based on $\delta^{13}\text{C}$ and lignin phenols (Hedges and Mann, 1979; Prah et al. in press). Recently, Prah et al. (in press) have directly compared the estimates obtained with the different biomarkers. Their results suggest that interpretations of biomarker distributions may be biased by uncertainties in the assigned endmember OC:biomarker values. Endmember values may vary regionally and thus, application of values determined in other locations may produce erroneous or at least inaccurate estimates.

The cause of the variations in endmember values is not firmly established but may be attributable to regional differences in types (and reactivity) of the OC, differences in soil degradation processes which alter the character of the residual material that is available to be transported to the coastal environment, and the effectiveness of hydraulic sorting of the material as it is transported away from the land surface (Gough et al. 1993). All of these

processes may act to alter the assigned OC:biomarker ratio and hence the estimated terrestrial contribution of the preserved OC. Without full resolution of the source of the materials accumulating in sediments under upwelling systems, accurate interpretations of past changes cannot be made. Future measurements of stable carbon isotopes on single compounds via isotope-ratioing GCMS may provide new constraints by which the above discrepancies are resolved.

RESUSPENSION AND LATERAL TRANSPORT

The influence of lateral transport of metabolizable OC has been demonstrated for a variety of margin systems (eg. Jahnke et al. 1990; Anderson et al. in press). The results of *in situ* benthic flux measurements at the indicated locations on the east and west coast margins of North America are presented in Figure 2. Results for each region are displayed relative to distance from the shelf break and water depth. At both margins, highest rates of deposition are not located at the nearshore, shallowest sites but rather at sites that are further offshore either on the slope or the continental rise. Because there is no evidence for comparable increases in the primary production in the water overlying the high deposition areas, it is clear that these patterns are maintained by lateral redistribution of OC. Models of particle flux suggest that greater than 1/2 of the total OC accumulating at the depocenter in the Atlantic and at the base of the continental slope in the Pacific are supplied by lateral transport.

It is important to note that while the depositional pattern extending offshore is similar on both margins, the depth distribution is significantly different. While highest rates of deposition on the east coast margin are observed on the upper slope depocenter at approximately 750 m water depth, highest rates of deposition on the west coast are observed at 3000-4000m, at the base of the continental slope. Because similarly high rates of deposition have been measured at the base of the West African continental slope (Jahnke et al. 1994, 1989), we suspect that

deep, high deposition regions are general features of eastern boundary current, upwelling systems.

Another example of lateral transport and redistribution of materials comes from recent submersible observations of the sea floor under the Peruvian upwelling region (Figure 3; Glenn, et al. 1994). Despite the well documented high rates of primary productivity that occur in this region, only small patches of modern sediments were encountered between 130 and 850 m along the two transects indicated in the figure. The vast majority of the upper slope was comprised of hard pavements and sand-sized phosphorite pellet beds. Thus, the biogenic particles produced in the highly productive waters overlying these locations have been laterally exported to other locations or remineralized (unlikely to be sufficiently complete to prevent at least some deposition at these shallow water depths). Thus, in this extreme example, the high productivity maintained in the surface waters by coastal upwelling is not reflected in the sea floor sediments directly below. Variations in lateral transport may, therefore, significantly alter the accumulation rates of organic materials and confound attempts to relate downcore variations in organic accumulations rates to variations in upwelling activity and surface waters primary productivities.

POST-DEPOSITIONAL PROCESSES

A variety of post-depositional processes act to alter biogenic materials deposited on the sea floor. The principal processes that we will consider include the degradation and oxidation of organic matter, the dissolution of CaCO_3 and the dissolution of opal.

Organic Matter Decomposition

In most sea floor environments, the majority of the deposited OC is lost from the sediment record by biologically-mediated remineralization processes. Relating sediment OC concentrations to particle input rates and water column conditions, therefore, requires a fundamental understanding of the recycling process. While the majority of the studies of benthic remineralization have focussed on the reactivity (or age) of the OC deposited and the oxidant availability (cf. Westrich and Berner, 1984; Emerson, 1985), numerous other processes may influence carbon preservation. While quantitative models of organic matter remineralization have been widely reported in the literature (cf. Boudreau and Ruddick, 1991), these models have generally focused either exclusively on anoxic or oxic to suboxic systems (Rabouille and Gaillard, 1992).

Sediments under coastal upwelling regions often exhibit a wide range of redox conditions and models capable of simulating the combined effects of oxic, suboxic and anoxic remineralization are required to accurately represent the fate of OC in these systems. An example of a model that considers the combined effects of oxygen, nitrate and sulfate reduction is schematically represented in Figure 4 (Jahnke et al. in prep.). The physical processes considered by this model include particle input, particle mixing, particle burial and solute pore water diffusion. Reactions include organic matter remineralization utilizing oxygen, nitrate and sulfate as the oxidants, and the oxidation of sulfide and ammonium. Pore water carbonate system is maintained in equilibrium and CaCO_3 is permitted to dissolve when under saturated conditions occur.

An example of the comparison between simulation results and benthic flux and pore water measurements is provided in Figure 5. In this example, the simple parameterizations indicated in Figure 4 are able to reproduce the observed fluxes and pore water distributions

reasonably well. While the overall agreement between this general simulation and observed results suggests that a reasonable description of the remineralization of OC in sediments may be achieved, several aspects and limitations of this type of modeling should be noted. First, these models are driven by the input and distribution of degradable OC and are not influenced by the input of recalcitrant OC. The majority of the OC buried, however, is non-degradable. Therefore, despite some success in representing sea floor diagenesis, these types of models generally do not provide an accurate means of linking total OC input with OC burial. Second, as will be discussed in the section concerning CaCO_3 , in selected locations, benthic flux chamber measurements indicate that the linkage between OC oxidation and CaCO_3 may not be well represented by this type of model. If supported by future results, fundamental changes in the parameterization of processes controlling sediment remineralization will be necessary to simulate the observations.

In addition to the uncertainties in the models of OC diagenesis, several other aspects of OC burial have provoked debate. The correlation between low bottom water oxygen concentrations and high OC concentrations have lead numerous previous researchers to conclude that oxic degradation is more efficient than anoxic degradation and that low oxygen conditions enhance OC preservation. This position has been challenged recently by Calvert et al (1992) who have suggested that high OC accumulation under low bottom water O_2 conditions are related more to higher rates of input rather than enhanced preservation. In general, laboratory studies (Westrich and Berner, 1984) and field pore water and benthic flux chamber measurements (Jahnke, 1990) support this position. On the other hand, widespread variation in OC in deep sea sediments that can not be ascribed to changes in input suggest that basin-wide factors such as bottom water oxygen concentration influence OC accumulation (Curry and Lohmann,

1990).

One possible means of reconciling these different views is displayed in Figure 6 where the fraction of OC remaining under hypothetical oxic and anoxic remineralization schemes are plotted vs the extent of reaction. In this scheme, the initial stages of degradation, representing the loss of the most labile fraction of the OC, are controlled by the reactivity of the OC itself, realatively independent of oxidant. In contrast, the remineralization rate of the relatively recalcitrant residual OC during the final stages of degradation may be significantly different under oxic and anoxic conditions. Thus, laboratory incubations, pore water measurements and *in situ* benthic flux chamber experiments, which are dominated by the degradation of relatively fresh OC, display no difference between oxic and anoxic conditions. Variations in OC in deep sea sediments, which represent the recalcitrant residual OC, may still be sensitive to bottom water oxygen.

With this conceptual model in mind, the role of bottom water oxygen in controlling burial can be estimated if the OC burial efficiency is known. If only a few percent of the deposited OC is preserved in the sediments, such as in the deep sea, preservation is very sensitive to minor changes in remineralization rate and oxygen availability may be an important factor controlling burial. On the other hand, if >10% of the deposited OC is preserved in sediments, it is unlikely that bottom water O₂ is a significant controlling factor.

Burial efficiency as a function of sedimentation rate is displayed in Figure 7. Because sedimentation rates under coastal upwelling systems generally exceed 5 cm/kyr, burial efficiencies are generally greater than 10%. Thus, it is unlikely that oxygen availability significantly influences OC accumulation under coastal upwelling systems.

Other factors have been suggested to influence the preservation of organic matter in marine sediments. Lee (1992) has suggested that while bacterial remineralization processes may be relatively insensitive to anoxia, the absence of grazing organisms under anoxic conditions alters the overall processing of sedimentary organic matter, particularly bacterial biomass carbon, and results in higher burial efficiencies. This would suggest that low oxygen conditions would yield a preferential preservation of organic matter. On the other hand, Mayer (1994) has demonstrated a positive correlation between sedimentary organic carbon content and sediment particle surface area. He suggests that this correlation is the result of adsorbed organic materials that are protected from further decomposition and enzymatic attack by occupying pores on the mineral surfaces that prevent enzymatic attack. The deposition of organic rich sediments then becomes a function of the hydrodynamic conditions that permit the accumulation of fine-grained, high surface-area sediments rather than a strict dependence on presence of bottom water oxygen or OC rain rate. Thus, the basic controls on OC preservation remain unknown and quantitative interpretations of OC variations in sediments in terms of changes in OC input rates must be viewed cautiously.

Dissolution of CaCO_3

Down-core variations in CaCO_3 content have been extensively used to infer changes in the surface water productivity and deep water chemistry and circulation. While the loss of sea floor CaCO_3 is controlled primarily by the saturation state of the overlying bottom water, diagenetic models (cf. Archer, 1991) suggest that the remineralization of OC in surface sediments may also drive dissolution and hence influence CaCO_3 preservation. This potential mechanism is calculated to be most important in regions of high organic input relative to CaCO_3 input. Such conditions exist under most coastal upwelling regions.

The magnitude of metabolically-driven CaCO_3 dissolution predicted by Archer (1991) is displayed in Figure 8. The accumulation rate of CaCO_3 estimated by the model at the saturation horizon is displayed versus the OC: CaCO_3 rain ratio. Because the results at the saturation horizon are presented, there is no inorganic dissolution and all of the predicted loss of CaCO_3 is attributable to dissolution driven by metabolically-released CO_2 . When there is no OC rain, there is no metabolic dissolution and 100% of the deposited CaCO_3 accumulates. On the other hand, when the OC: CaCO_3 ratio exceeds approximately 1.2, the calculation suggests that essentially all of the CaCO_3 dissolves at or above the saturation horizon. Thus, varying the OC: CaCO_3 rain ratio should have a significant influence on CaCO_3 accumulation and, by forcing changes in deep ocean carbonate alkalinity, atmospheric CO_2 (Archer and Maier-Reimer, 1994).

Recent attempts to directly verify these model predictions utilizing *in situ* benthic flux chambers have yielded ambiguous results (Figure 9). On the West African continental rise, a large drop in chamber oxygen is observed during the incubation, undoubtedly the result of high sea floor respiration rates. However, no concurrent increase in the chamber was observed in titration alkalinity or Ca^{2+} , suggesting little, if any, CaCO_3 dissolution. The results suggest that despite the high rain rate of organic carbon to the sediments in the region, significant metabolically-driven CaCO_3 dissolution was not observed. Conversely, on the continental rise adjacent to the upwelling region off central California and under the Gulf Stream adjacent to the North Carolina margin, the measured decrease in chamber oxygen is accompanied by increases in titration alkalinity and Ca^{2+} (Figure 9). Numerical simulations of the results indicate that the observed titration alkalinity benthic flux cannot be supported by inorganic dissolution alone and that at these locations metabolically-driven CaCO_3 dissolution is required.

These conflicting reports are at present unresolved and significantly confound attempts to relate changes in the CaCO_3 content of sediments to variations in surface water processes and sediment inputs.

Dissolution of Opal

Numerous researchers have attempted to use the abundance of opaline tests in sediments to infer past changes in surface productivity (cf. Calvert, 1966). Support for the use of opal as a proxy for primary productivity can be found in the overall correspondence between the opal content of modern sediments and overlying productivity and the correlation between OC flux and opal flux observed in sediment traps (Figure 10). Despite these observations, the relationship between the opal content of sediments and opal particle flux remain uncertain and the analysis of the opal distribution in sediments is generally restricted to variations in species composition. The primary reason for this is that in most locations, the majority of the deposited opal dissolves. The residual opal component in sediments is, therefore, very sensitive to even minor changes in opal dissolution rate. Despite previous research, the factors that control opal dissolution are not firmly established.

In an attempt to explain the large range of asymptotic pore water Si values observed in sediments, early research efforts attempted to explain pore water silica values as a dynamic balance between input and dissolution. The different asymptotic values were then explained as reflecting variations in the depth of dissolution, caused either by differences in dissolution kinetics or bioturbation rates.

More recently, Archer et al. (1993) have suggested that there is not a single solubility for biogenic opal but rather there is a systematic trend to higher solubilities at high surface water

production rates. Using such a model, they were able to reasonably reproduce the overall pattern of opal preservation and pore water Si asymptotic value for a variety of locations. Systematic variations in opal solubility may be caused by chemical variations, such as the aluminum content, which may be the result of changes in precipitation rate, temperature of formation, or species composition.

The implications of this model are that the asymptotic pore water Si value should be correlated with opal rain rate (Figure 11). The results reported by Archer et al. (1993) suggest that there is such a correlation. Most importantly, this model predicts that sedimentary opal content is insensitive to input rates and that there is, therefore, no basis for relating variations in sedimentary opal content to changes in surface productivity. Clearly, additional work is required if we are to properly interpret past variations in opal preservation.

CONCLUSIONS

Based on this review, we conclude that fundamental uncertainties remain concerning the relationships between the production, deposition and burial of the major biogenic phases in coastal upwelling systems. The limitations identified significantly hinder attempts to relate sediment distributions and characteristics with past upwelling conditions. The important concerns include uncertainties: 1. in the contribution of OC from multiple coastal sources; 2. in the processes and sediment characteristics that control OC preservation; 3. in the linkages between OC remineralization and CaCO_3 dissolution and 4. in the factors that control opal preservation. Many of the factors and characteristics that control the retention of these phases in sediments are also influenced by upwelling characteristics. Until resolved, the above concerns will hinder the interpretation of the sedimentary distribution of the major biogenic phases beneath coastal upwelling systems.

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FIGURE CAPTIONS

Figure 1. Schematic of the important processes that impact the characteristics of sediments accumulating at a hypothetical coastal upwelling location.

Figure 2. Deposition patterns of metabolizable OC determined by pore water and *in situ* benthic flux chamber incubations at the indicated locations and the eastern and western margins of North America. The Pacific results are from Jahnke et al. 1990; Atlantic results are from Anderson et al. in press (▽); Martin and McCorkle, 1993 (○); Smith, 1978 (●); and Jahnke, unpublished(▼).

Figure 3. Surface sediment type identified from direct observation and sampling via manned submersible at the indicated locations on the Peruvian continental slope and rise.

Figure 4. Schematic of combined oxygen, nitrate, and sulfate reduction model with CaCO_3 dissolution.

Figure 5. Examples of results of simulations performed with the model represented in Figure 4 (dashed lines) compared to field data from 3700 m water depth off central California. Linear regression and 95% confidence limit of benthic flux chamber results are indicated by thin solid lines.

Figure 6. Hypothetical degradation rates for OC under oxic and anoxic conditions as a function of the extent of reaction (P. van Cappellen, personal communication).

Figure 7. Organic carbon burial efficiency as a function of sedimentation rate. Re-drawn from Henrichs and Reeburgh, 1987.

Figure 8. The proportion of the CaCO_3 that accumulations at the saturation horizon as a function of the CaCO_3 :OC rain ratio as estimated by Archer, 1991.

Figure 9. Example benthic chamber results from the continental rise adjacent to the West African and California upwelling systems and the east coast U. S. margin. Results were obtained at the location indicated by the large solid square on each accompanying figure.

Figure 10. Correlation between opal and organic carbon flux measured in sediment traps. Redrawn from Milliman and Takahashi (in press).

Figure 11. Correlation between sedimentary opal content and pore water Si asymptotic value and opal rain rate.

FIG. 1

PACIFIC MARGIN

ATLANTIC MARGIN

Fig. 3

FIG. 4

Data Model Comparison - Site M

Fig. 5

FIG. 6

FIG. 7

FIG. 8

FIG. 10

FIG. 11

Attachment 2.

**THE GLOBAL OCEAN FLUX OF PARTICULATE ORGANIC CARBON:
AREAL DISTRIBUTION AND MAGNITUDE**

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Submitted to Global Biogeochemical Cycles - 24 August 1995

ABSTRACT

The distribution of the POC rain rate to the sea floor in the Atlantic, Pacific and Indian Ocean basins between 61°N and 61°S was calculated from a compilation of 136 individual sea floor oxygen flux estimates (for water depths ≥ 1000 m only). Rates were extrapolated between individual measurement sites using an empirically-derived correlation between the sea floor oxygen flux and the sediment organic carbon content, CaCO_3 content, and accumulation rate. The POC flux through the 1000 m depth horizon was then estimated from published correlations between sediment trap-determined fluxes and water depth.

Total oxygen utilization in the deep ocean is estimated to be 1.2×10^{14} mol O_2 yr^{-1} , a value that agrees well with previous estimates which were based on surface water primary productivity - sediment trap - depth relationships and with deep water respiration rates estimated from AOU- ^{14}C relationships. Based on the derived global ocean flux distribution, it is concluded that: 1. DOC inputs are not required to account for estimated deep water respiration rates; 2. the majority of the POC input to the deep ocean occurs within 30° of the equator; 3. the proportion of primary production that reaches the deep sea does not vary greatly with latitude; 4. gyre and continental margin regions contribute roughly equally to the deep POC flux with a relatively minor contribution from the equatorial divergence region; 5. of the estimated 7.2×10^{13} mol C yr^{-1} of POC that sinks below the 1000 m depth horizon, 45% (3.3×10^{13} mol C/yr) reaches the sea floor where it is oxidized; 6. when normalized to basin area, average deep flux rates in the Atlantic and Pacific are similar while highest rates are observed in the Indian Ocean; and 7. the results can be fully reconciled only if the benthic flux of DOC is significantly less than the benthic O_2 flux.

INTRODUCTION

One of the major goals of ocean science research is to quantify and develop an understanding of the processes that control marine biogeochemical cycles. Transfer of carbon and nutrient elements through the major marine reservoirs plays a key role in controlling atmospheric CO₂ concentrations and perhaps climate (Jahnke, 1990; and references therein). To accurately predict future atmospheric CO₂ levels and interpret past changes from the sedimentary record, it is necessary to determine on a global scale the rates at which carbon is cycled through the ocean and the factors that may alter the rates of transfer.

The present interpretations of flux patterns on a global basis are in all cases derived from surface properties which can be measured rapidly from ships, such as surface water primary production (Berger et al. 1988), remotely sensed properties from satellites such as surface water pigment concentrations (Longhurst et al. in press) or from models (Najjar et al. 1992). A variety of factors conspire to limit the accuracy of flux estimates to the deep ocean from these results. For example, these near surface properties are not direct measures of vertical flux but rather the flux must be estimated from them by some derived relationship or correlation. Also, only a very small proportion of the surface water primary production (on the order of 1%-3%) reaches the deep sea. Thus, models that estimate fluxes to the deep sea as the difference between production and remineralization in the upper water column are attempting to determine a small difference between two relatively large numbers and are therefore subject to large uncertainty. In addition certain processes, such as lateral transport from continental margins, are not reflected in the surface water characteristics and cannot be incorporated into these flux estimates.

In the following, direct measurements of oxygen fluxes across the sea floor are used to estimate the flux of particulate organic carbon (POC) to the deep sea. Since organic carbon deposited on the deep sea sediment is nearly quantitatively remineralized through oxic respiration (Bender and Heggie, 1984), benthic oxygen flux is quantitatively related to POC rain rate, assuming the sea floor exchange of dissolved organic carbon (DOC) is minor. This latter assumption is explicitly addressed in the discussion.

The advantages of assessing deep ocean fluxes through sea floor studies are that fluxes are measured directly, not estimated from static or non-fluxing quantities and are, therefore, not dependent on empirical correlations or model parameterizations. Also, sea floor fluxes display less temporal variability (cf. Smith et al. 1992; Martin and Bender, 1988; Sayles, 1994) than water column fluxes. This is primarily because sedimentary processes such as particle mixing, biological utilization, biomass production, and pore water molecular diffusion tend to dampen the response of benthic fluxes to short-term episodic depositional events. This greatly facilitates the establishment of mean flux rates. Additionally, remineralization is a destructive process. Thus, resuspension and lateral transport of particles do not confound estimates of primary vertical flux based on benthic chamber measurements. This is especially important near continental margins. While local topography must exert some influence on deposition and hence remineralization rates, in general, these local effects are much smaller than the large-scale regional trends (Jahnke et al. 1990).

In this manuscript, the sea floor distribution of benthic oxygen flux is calculated from a compilation of 136 individual estimates. Fluxes are extrapolated throughout the Atlantic, Pacific, and Indian Ocean basins from 61° N to 61° S utilizing an empirically-derived correlation

between sea floor oxygen flux and sedimentary organic carbon content, CaCO_3 content and accumulation. The distribution is then used to assess the rain of particulate organic carbon (POC) across the 1000 m depth horizon and to the sea floor. Because the primary goal is to assess the POC input to the deep ocean, only ocean areas with water depths > 1000 m are considered. Thus, continental shelf and upper slope regions are excluded from this discussion.

DATA SET

Locations of benthic flux estimates compiled into the present data-base are indicated in Figure

1. All parameter distributions discussed in the manuscript are digitized into two degree squares. Rather than indicate each of the individual locations for which there are results, Figure 1 shows those two degree regions for which there are estimates. While some of the squares represent a single measurement, many include multiple measurements within the region. This is not an exhaustive summary of all results; only a few of the pre-1980 data sets are included. Because of the methods employed, many benthic flux estimates prior to 1980 are judged to be unreliable.

Overall, 136 individual benthic flux estimates have been made within the regions indicated. Of these, 38 are based on direct *in situ* benthic flux chamber deployments and 26 are calculated from fine-scale (mm) pore water gradients determined either with *in situ* oxygen microelectrode measurements or shipboard whole core squeezing. The remaining flux estimates are based on cm-scale pore water gradient calculations.

Symbol colors represent different techniques for calculating benthic fluxes. Light blue are pre-1980 pore water gradients, generally of low vertical resolution (Grundmanis and Murray,

1982; Suess et al. 1980; Jahnke et al. 1982; Jahnke, unpublished results; Heggie et al. 1987; Patching et al. 1986; Froelich et al. 1979). Dark blue regions indicate locations where cm-scale pore water gradients have been measured more recently than 1980 (Emerson, unpublished results; Froelich, unpublished results, and Jahnke et al. 1989). These generally have higher resolution than the earlier results. Green represents regions where fine-scale interfacial measurements have been made either through shipboard whole core squeezer measurements (Bender et al. 1987; Martin et al. 1991) or *in situ* microelectrode oxygen measurements (Reimers et al. 1986; 1992; Hales et al. 1994; Archer et al. 1989). Pink regions represent locations where benthic fluxes have been estimated directly via *in situ* benthic flux chamber measurements (Smith, 1978; 1987; Berelson et al. 1990; Reimers et al. 1992; Jahnke, unpublished results; Pfannkuche, 1993; Sayles et al. 1994; Jahnke et al. 1994). For regions in which several different types of measurements have been made, the color corresponds to the technique of presumed greatest accuracy, ie light blue < dark blue < green < pink. Thus, while numerous *in situ* microelectrode O₂ profiles are available for the central California continental margin (Reimers et al. 1992) the region is indicated in pink because of the adjacent benthic flux chamber measurements.

It is important to note that the different measurements techniques are not concentrated in any particular type of setting but are reasonably distributed among the deep sea environments sampled. For example, *in situ* benthic flux chamber measurements have been made in the eastern, central and western equatorial Pacific, western equatorial Atlantic, eastern and western continental margin regions in the Atlantic and Pacific and a site as far north as 47° N in the Atlantic. Thus, while the sea floor is clearly undersampled and benthic flux estimates are not available for major regions, such as the Indian Ocean, the data that are available represent a

wide variety of deep ocean environments that may be representative of much of the sea floor.

GLOBAL FLUX DISTRIBUTIONS

To estimate fluxes on a basin-wide basis, an objective method is required by which the measured benthic fluxes can be extrapolated and interpolated to regions where there are no measurements. Jahnke and Jackson (1992) noted a relationship between benthic oxygen flux and organic carbon burial. Previous reported correlations between sediment oxygen consumption and sediment burial rate (Canfield, 1989) and organic carbon burial efficiency and sediment accumulation rate (Henrichs and Reeburgh, 1987) support this overall correlation. In essence, these correlations suggest that given the relatively narrow range of bottom water oxygen and sediment accumulation rates found in the deep ocean, the majority of variation in organic carbon burial efficiency is determined by input rates.

Jahnke and Jackson (1992) also reported, however, that different correlations were observed between the relatively high organic carbon input and low CaCO_3 regions indicative of much of the Pacific sea floor and continental margin regions and the high CaCO_3 low organic carbon input region of the Atlantic. Thus, Jahnke and Jackson (1992) were forced to use different relationships for the different regions they tried to assess.

These differences may be caused by consistently estimating burial at a more recent time horizon in the CaCO_3 -rich sediments than in the CaCO_3 -poor sediments. Organic carbon remineralization is generally envisioned as a continuous process, proceeding at a rate that decreases with time from deposition (Boudreau and Ruddick, 1991). Strictly speaking, burial can only be evaluated relative to a specific time horizon. Comparison between regions that

represent different time intervals will yield different results.

The relationships among organic carbon input, concentration and burial through a hypothetical mixed layer in the presence and absence of a dominate diluting phase such as CaCO_3 is illustrated in Figure 2. The left portion of the figure illustrates the situation where a small amount of non-reacting organic carbon is deposited in the presence or absence of a much larger amount of CaCO_3 . In this situation, the evaluation of organic carbon burial is independent of CaCO_3 accumulation because the organic carbon content and total sediment accumulation rate co-vary. In this simple situation it is clear why correlations with burial are used and not with organic carbon content which varies significantly with diluting phases.

The situation is more complex when the organic carbon is considered to be reactive (right side of the figure). In this case, the residence time of the organic carbon in the surface layer is significantly reduced with the addition of CaCO_3 . Thus, the age of the sediment at the depth horizon at which burial is evaluated is different with and without CaCO_3 . In the example provided in Figure 2, with no mixing, 500 years are required to accumulate 10 cm of sediment with CaCO_3 whereas 5000 years are required without CaCO_3 . If the reaction half-life of the decomposing organic matter is 500 years, a significantly different organic carbon burial rate is estimated for these two cases as shown in the figure. Note that the exact difference depends on the reactivity of the organic matter. Thus, relative to the low- CaCO_3 sites, organic carbon burial is systematically overestimated at the high- CaCO_3 sites because the average age of the sediments samples is younger. Thus, to compare results from all sites, the burial rate must be decreased at the high- CaCO_3 sites in proportion to the difference in sediment age.

As an approximate correction for the age difference, the accumulation rate of CaCO_3 has been

subtracted from the estimated total accumulation rate. Organic carbon burial is then estimated as the product of the organic carbon and the "CaCO₃-corrected accumulation rate." Note that this correction is empirical and provides only an approximate correction for the difference. As the reactivity of organic matter at the sea floor becomes better constrained, more accurate and mechanistically meaningful corrections will be formulated.

This correction has been applied to all regions except the equatorial Pacific (within 5° of the equator). Several authors have suggested that the equatorial Pacific sediments are not in steady state but are eroding in the present ocean due to an excess of dissolution over input of CaCO₃ (Keir, 1984; Broecker et al. 1991). Thus, at this location, the elevated CaCO₃ content may not indicate increased sediment accumulation and decreased residence time in the surface layer. Therefore, the CaCO₃ correction has not been applied to the results from this region.

The CaCO₃-corrected organic carbon burial rate and non-corrected rate for the equatorial Pacific correlates well with the benthic oxygen flux estimated from all *in situ* benthic flux chamber and microelectrode measurements and shipboard whole core squeezer pore water measurements (Figure 3). This analysis is restricted to these data sets because it is assumed that they represent the most accurate benthic flux estimates. Unlike the earlier correlation attempts, (Jahnke and Jackson, 1992) all results appear to follow a similar relationship. No significant difference is observed between the Pacific (closed symbols) and Atlantic results (open symbols). Thus, the CaCO₃-corrected organic carbon burial rate appears to provide a reasonable mechanism by which the measured benthic fluxes can be extrapolated throughout the ocean basins.

GLOBAL FLUX ESTIMATION

To extrapolate the benthic flux estimates at the locations indicated in Figure 1 throughout the ocean basins, an estimate of the global distribution of the CaCO_3 -corrected organic carbon burial rate is required. The latter is estimated from published sea floor distributions of organic carbon, bulk sediment accumulation rate and CaCO_3 sedimentary content. The organic carbon distribution was taken from Premuzic et al. (1982), digitized with 2° resolution (ie. into 2° by 2° regions; Figure 4).

The digitized representation of the sediment accumulation rate distribution compiled by Cwienk (1986) is presented in Figure 5. Two modifications have been made to the original distribution reported by Cwienk (1986). First, the accumulation rates in the western North Atlantic have been reduced as described in Jahnke and Jackson (1992) to values that are thought to better represent present-day conditions. Briefly, this region received significant quantities of sediment at the end of the last glacial. Sediment accumulation rates based on intervals extending back to the last glacial maximum would include accumulation associated with the end of the last glacial period and therefore overestimate present rates. Second, the accumulation rate along the southern portion of the East Pacific Rise has been modified so that the higher accumulation rate region occurs only at the southernmost portion, where it may be influenced by increased productivity associated with the polar front. This appears to better reflect present-day oceanographic conditions. From this distribution, the greater average accumulation rates in the Atlantic relative to the Pacific are clearly evident. In general, this is due to the CaCO_3 preservation in the Atlantic. Note that these two data sets did not include the Gulf of Mexico or the Caribbean Sea and, therefore, these regions are excluded from the compilation presented.

Multiplying the organic carbon and sediment accumulation rate distributions provides a global estimate for the organic carbon burial rate (Fig. 6). In general, the high accumulation rate regions are restricted to continental margin settings. On this scale, the average difference between the Atlantic and Pacific gyre regions is only slightly visible. Nevertheless, this difference is significant and as discussed before, is most probably an artifact of evaluating burial at a younger sediment horizon, on average, in the Atlantic than in the Pacific.

A digitized representation of the seafloor CaCO_3 distribution (Berger et al. 1976; Biscaye et al. 1976; Kolla et al. 1976) is provided in Figure 7. The CaCO_3 contribution to the sediment accumulation rate is subtracted at each location, except in the equatorial Pacific as described earlier. The global distribution of the sea floor oxygen flux (Figure 8) is then estimated from this factor and the relationship displayed in Figure 3.

In general, the distribution of the sea floor oxygen flux estimated in this manner follows the general features of the new production distribution estimated by Berger et al. (1988). On average, low fluxes are predicted for the central gyre regions of all of the ocean basins and high fluxes are predicted for continental boundary regions, especially those areas in which coastal upwelling occurs such as the Pacific margin of the Americas, the Atlantic African margin and the monsoonally influenced margins of the Arabian Sea. Elevated fluxes are also estimated for the equatorial Pacific region and the polar front region of the southern Atlantic and Indian Oceans.

To assess the validity of the estimated distribution presented in Figure 8, the estimated value and directly measured value for each sea floor flux estimate in the data base made since 1980 are plotted against each other (Figure 9). Note that this excludes the benthic flux chamber

results in the northwestern Atlantic reported by Smith (1978) which appear to be consistently lower than more recent measurements (Hales et al. 1994). To guide the comparison, the 1:1, 1:2 and 2:1 ratio lines are also indicated on the figure. Two conclusions can be drawn from this comparison. First, the overall magnitude of flux displayed in the global distribution (Fig. 8) agrees with the direct observations. That is, the predicted fluxes are evenly distributed about the 1:1 line; the extrapolation method employed does not consistently over-predict or under-predict the sea floor flux. The few instances where the results are outside of the 1:2 - 2:1 region are all cases where the measured value is greater than the predicted value and are from continental margin regions where the offshore gradient in the benthic flux is very steep. In these regions the two degree averaged flux cannot represent the true variability. Second, the predicted flux values are generally within a factor of two of the measured rates over the entire range of values (approx. a factor of 20) observed.

It is not possible to know in an absolute sense whether the distribution presented in Figure 8 is correct. This is especially true for large regions, such as the northwestern Pacific Ocean and Indian Ocean, in which there are no measurements. However, from the comparison provided in Figure 9, it is concluded that the global distribution presented is consistent, within a factor of two, with nearly all (63 out of 68) of the measurements of sea floor flux in the data base that have been made since 1980. Thus, Figure 8 provides a consistent summary of all the measurements since 1980 and a tangible framework in which to discuss what is presently known about global ocean fluxes and to pose future studies.

It is important to recognize that because the benthic O_2 flux must be supported by organic matter remineralization, the flux distribution provided in Figure 8 also represents the particu-

late organic matter rain to the sea floor, offset by the $O_2:C$ remineralization ratio. Assuming relatively slow remineralization rates in the deep water column (well supported by sediment trap studies), this distribution also approximates deep water column particulate fluxes. However, part of the distribution pattern displayed in Figure 8 is due to variations in water depth. Shallower sites will tend to have higher fluxes because there is less opportunity for remineralization in the water column. To facilitate the comparison of fluxes between regions, the rain rate of organic matter is also estimated at a constant depth horizon, chosen to be 1000 m. This depth is selected because it roughly corresponds to the base of the main thermocline. Fluxes across this boundary thereby approximate particulate exchange between the surface and deep ocean which facilitates the comparison to hydrographically determined parameters, such as apparent oxygen utilization, and is a key term in climate models.

The rain rate at 1000 m is estimated from the particulate vertical flux/water depth correlations compiled by Berger et al. (1988). A comparison between measured and calculated fluxes for the equation that yielded the "best-fit" for all of their data is provided in Figure 10A. The derived expression represents the measured fluxes to a reasonable approximation. The vertical distribution of fluxes (as a percentage of primary production) implied by the expression is displayed in Figure 10B. Note that the flux at 1000 m is only a little more than a factor of two greater than the flux at 4000 m. Thus, extrapolating sea floor fluxes with this expression to the 1000 m depth horizon alters the numerical value of the flux by only a factor of approximately 2 at most deep sea locations and less at shallower sites.

Based on the water depth and estimated sea floor flux in each two degree region, the relationship displayed in Figure 10 was used to estimate the flux at 1000 m (Figure 11). Note that

since this distribution is based on the sea floor oxygen flux distribution, it has been left in oxygen units. Thus, the distribution represents the POC flux across the 1000 m boundary expressed as the equivalent amount of O_2 required to remineralize the flux. To obtain the organic carbon flux, this distribution simply is multiplied by the C: O_2 ratio observed in the remineralization of the raining material (cf. Takahashi et al. 1985; Anderson and Sarmiento, 1994).

In general, the distribution of fluxes at 1000 m is very similar to that observed for the sea floor. The lowest values are observed in the central gyres; the highest values are along continental margins characterized by coastal upwelling systems. Again, relatively large fluxes are estimated for the equatorial Pacific and the polar front region of the southern Atlantic.

DISCUSSION

Previous workers have estimated the total organic carbon flux to the deep ocean or the total deep water oxygen utilization rate. Because these estimates are of total deep ocean metabolism, they include the combined effects of deep water and sedimentary respiration and are comparable to the 1000 m flux estimated here (Table 1). The value attributed to Berger et al. (1988) is based on their reported total primary production transposed into flux at 1000 m using their "best-fit" relationship and converted into O_2 units using a C: O_2 ratio of 0.6 (Takahashi et al. 1985). Other values were converted to basin totals assuming the volume of the Atlantic and Pacific below 1000 m is 2.47×10^{20} l and 5.29×10^{20} l, respectively. With the exception of the deep Atlantic oxygen utilization estimated by Hinga (1985), the values are very similar. The results summarized in the table indicate that the total flux represented in Figure 11 is clearly in line with rates of deep sea respiration estimated from hydrographic properties and

total global fluxes estimated from compilations of surface water primary production and sediment trap fluxes. The general agreement between these values supports the estimates of the total flux to the deep ocean estimated here and implies that on average, 45% of the organic matter remineralization in the deep ocean occurs in the sediments.

The latitudinal distribution of fluxes to the deep ocean and to the sea floor are displayed in Figure 12 A&B. In each figure, the flux has been integrated and cumulatively summed in 10 degree latitude bands extending away from the equator. Thus, the column in the 0 to 10° region represents the proportion of the total flux that occurs between 10°S and 10°N. The column in the 10 to 20° column is the proportion of the total flux that occurs between 20° S and 20° N.

It is clear from these histograms that in all ocean basins individually and combined, low latitude regions dominate the flux of particulate materials to the deep ocean. Globally, approximately 2/3 of the total particulate flux to the deep ocean and to the sea floor occurs within 30° of the Equator. This low latitude dominance follows very closely the latitudinal distribution of deep ocean area (Menard and Smith, 1966). Cumulatively summing the ocean area into the same 10° latitudinal regions as used in Figure 12 reveals that 60% of the ocean area is equatorward of 30° and 77% is equatorward of 40°. Less than 10% of the oceans regions depicted in Figures 8 and 11 are polarward of 50°. Thus, the dominance of the low latitude regions in the particulate flux to the deep ocean is a consequence of the larger area. Conversely, high latitude regions can be a major source of particulate organic matter to the deep ocean only if the area-normalized fluxes are larger by more than a factor of five than those observed in the low latitude regions. From the organic carbon, sediment accumulation rate and organic carbon

burial rate distributions depicted in Figures 4, 5, and 6, respectively, and the estimated flux distributions in Figures 8 and 11, there is no evidence to support significantly higher fluxes in high latitudes. Thus, it is concluded that low to mid-latitude regions dominate the particulate flux of biogenic materials to the deep sea.

The flux rate estimates are compared to basin-wide primary production estimates for the Atlantic north of 20°S (Table 2). The primary production estimates are based on remotely sensed pigment distributions and regionally applied correlations (Platt et al. 1991). Several important observations can be drawn from the comparison in this table. On average, a little more than 2% of the estimated primary production is transported below 1000 m and approximately 1% reaches the sea floor. These values are similar to but slightly smaller than the percentages implied by the primary production-depth relationship reported by Berger et al. (1988) and presented earlier in Figure 10. The relationship in Fig. 10 would imply that 3% of the primary production is transported below 1000 m and 1.5% reaches the sea floor. Uncertainties in the primary production and C:O₂ ratio of organic matter remineralization account for these differences.

It is also clear from Table 2 that the proportion of the primary production that is not remineralized in the upper 1000 m is similar in all of the regions studied. It had been suggested (Pomeroy et al. 1991) that due to colder temperatures, microbial remineralization in the upper water column of high latitude regions was less efficient at remineralizing the spring bloom and that the proportion of primary production transferred from the photic zone was, therefore, greater. No significant trend with latitude is observed in Table 2. It appears that high latitude regions do not export a higher proportion of the primary production to the deep ocean in parti-

culate form.

The distributions presented in Figures 8 and 11 can also be used to assess the relative roles of slope, rise, equatorial, and gyre regions in the flux of biogenic material to the deep sea. Various regions have been defined as depicted in Figure 13. The slope region is defined as the 2° square immediately adjacent to the continents. The rise region is defined as two 2° squares adjacent to the slopes. Although the actual distances vary with latitude, the slope region corresponds approximately to the 200 km nearest the continents. The rise region is approximately 200 to 600 km from the continents. It is important to remember that only seafloor areas that are deeper than 1000 m are considered in this manuscript. Thus continental shelves and the upper continental slope where extremely large fluxes are observed are not included here. The equatorial region is the sea floor between 5°N and 5°S , and the gyre region is all of the sea floor area not included in these other regions.

The 1000 m and sea floor flux distributions have been integrated for each of these regions (Table 3). The percentage of the total flux for each region is given in the parentheses. In the Atlantic, the sea floor along continental margins (slope + rise) accounts for 47% of the benthic oxygen flux while gyre regions account for 52%. The extremely low contribution by the equatorial Atlantic is in part due to the fact that the equatorial region south of western Africa was included in the slope and rise regions because of its proximity to the continent (see Figure 13).

Despite the much larger proportion of basin area occupied by the gyre region in the Pacific, the contribution of the gyre region to the carbon flux is similar to that observed for the Atlantic, approximately 50%. Although the equatorial region in the Pacific contributes a larger proportion of the organic carbon to the deep sea than in the Atlantic, it is still less than the other

regions. The Pacific margin regions contribute slightly greater than 1/3 of the total flux to the deep ocean. By contrast, the Indian Ocean gyres contribute somewhat less of the sea floor and 1000 m fluxes, and the margins assume a more important role in the deep fluxes.

In general, the gyre regions contribute approximately 50% of the flux, the equatorial regions contribute less than 10% and the margin regions (slope and rise) contribute 40% or more.

Note that since some of the major adjacent seas and basins, such as the Caribbean Sea and Gulf of Mexico, are not included in the present compilation, the role of margin systems may be underestimated. It is also important to note that some of the features that are clearly associated with equatorial or margin processes extend into the gyre regions as defined in Figure 13. For example, a significant portion of the high flux area occurring in the tropical eastern Pacific is outside of the 5°N - 5°S region. Although these high fluxes are undoubtedly driven by processes associated with the equator, they are included in the gyre regions in the table.

With these omissions in mind and the inherent uncertainties in the extrapolated distributions, it is concluded that margin and gyre systems contribute approximately equally to the organic carbon flux to the deep sea. Equatorial regions contribute less than 10%.

Despite the inherent uncertainties in estimating organic carbon burial rates due to time-scale differences discussed previously, the relative importance of each region for organic carbon burial can also be estimated. The distribution presented in Figure 6 represents a total deep ocean burial rate of 1.25×10^{12} moles C/yr, approximately 3% of the sea floor deposition rate. Recall that regions shallower than 1000 m have been excluded. Thus, high deposition areas associated with deltas, continental shelves and upper slope environments are not included in the total. For the deep ocean, 65% of the organic carbon burial occurs in the slope and rise

regions, the gyre regions account for 28% of the burial and only 7% occurs in the equatorial regions.

Based on his estimates of the oxygen utilization in the deep waters of the Atlantic and Pacific, Hinga (1985) suggested that the average per unit area new production in the Pacific was approximately 1.8 times greater than in the Atlantic. The results reported here do not support this conclusion. Dividing the 1000 m fluxes provided in Table 3 by the surface area of each ocean basin suggests that the area-normalized fluxes in the Atlantic and Pacific are the same, $0.30 \text{ mol/m}^2 \text{ yr}$ in oxygen-equivalent units. Area-normalized fluxes at 1000 m in the Indian Ocean are somewhat higher at $0.44 \text{ mol/m}^2 \text{ yr}$. These higher fluxes in the Indian Ocean undoubtedly are due to a combination of small basin area and large productivity associated with the monsoon-driven upwelling. Also, high productivities in the Sulu Basin are included in the Indian Ocean totals.

Finally, it is important to recognize that the loss of DOC from the sediments would represent a fraction of the total POC rain to the sea floor that is not included in the benthic estimates presented here. DOC fluxes have been reported to be significant, perhaps even equivalent in magnitude to the remineralization-driven fluxes (Martin and McCorkle, 1993; Burdige et al. 1992).

The distribution of fluxes presented here can be used to constrain the potential DOC flux from the sea floor. Two hypothetical situations exist. The DOC that originates from the sea floor may be unreactive, thereby contributing to the deep water DOC pool or it may be reactive, thereby contributing to deep water oxygen consumption via remineralization.

If the DOC is unreactive and its sea floor flux is equivalent to the O_2 flux, 5.4×10^{13} mol DOC-C/yr would be added to the deep water from the sediments. If the transit time between the Atlantic and Pacific is on the order of 670 years (Broecker and Peng, 1982) DOC concentrations should increase by $37 \mu\text{M}$, nearly doubling the deep water concentration (Williams and Druffel, 1987). No interocean difference of this magnitude is observed (Bauer et al. 1992). In fact, Bauer et al. (1992) report no significant difference in deep DOC concentration or ^{14}C age between the Atlantic and Pacific. From this they concluded that there was no evidence for a significant component of DOC that was cycled on time scales shorter than several thousand years. Thus, it appears that if benthic-derived DOC is unreactive, its sea floor flux must be much less than the O_2 flux.

As demonstrated in Table 1, the total input to the deep sea estimated here from benthic fluxes and particle flux-depth relationships agrees well with total deep water respiration rates derived from ^{14}C -AOU relationships, which are not subject to uncertainties related to sea floor DOC cycling. If benthic-derived DOC is remineralized in the deep sea, it will contribute to the oxygen consumption rate. If the DOC benthic flux and subsequent water column remineralization was equivalent to the benthic oxygen flux, sea floor and deep DOC remineralization would account for the consumption of 1.09×10^{14} mol O_2 /yr., 90% of the total deep respiration (See Table 3). This would imply that >90% of the POC sinking below the 1000 m depth horizon would reach the sea floor and not be remineralized in the water column. Sediment trap results (see Fig. 10) indicate a much larger POC loss in the deep water column. Thus, it again appears that the benthic DOC flux must be much less than the O_2 flux to reconcile the observations.

CONCLUSIONS

The distributions of sea floor and deep ocean POC fluxes have been estimated from benthic flux measurements and sediment characteristics. The derived distribution is consistent with nearly all post-1980 benthic flux estimates suggesting that it provides a reasonable summary of benthic sea floor flux. The benthic flux was extrapolated to the 1000 m depth horizon using published particle flux - depth relationships derived from sediment trap studies. Total deep water oxygen utilization estimated in this manner agrees well with deep water respiration rates based on apparent oxygen utilization - ^{14}C and primary production - sediment trap flux - depth relationships. Based on the calculated distributions, the following conclusions are made:

1. Total oxygen utilization in the deep ocean is $1.2 \times 10^{14} \text{ mol O}_2 \text{ yr}^{-1}$.
2. DOC inputs are not required to account for the estimated respiration rates.
3. The proportion of the primary production that reaches the deep sea does not appear to vary significantly with latitude.
4. Gyre and continental margin regions contribute roughly equally to the deep POC flux with a relatively minor contribution from equatorial regions.
5. 45% of the respiration in the deep sea occurs in the sediments.
6. The majority of the POC flux to the deep sea occurs within 30° of the equator.
7. Average fluxes per unit area are similar in the Atlantic and Pacific and slightly larger in the Indian Oceans.
8. The results can be fully reconciled only if the benthic flux of DOC is significantly less than the benthic O_2 flux. flux estimates to date. However, significant gaps in the data set exist and this distribution will be revised as more results become available.

ACKNOWLEDGEMENTS

I wish to thank M. Matheson for assistance in preparing the figures. This work was supported by Grants NSF OCE-9201896 and DOE DE-FG02-92ER61415.

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FIGURE CAPTIONS

Figure 1. Two degree by two degree regions (> 1000 m) for which sea floor benthic flux estimates have been determined and are included in the present data set. Different colors correspond to different methods of estimating the benthic flux; light blue - pre-1980 pore water measurements, dark blue - post-1980 pore water measurements, green - fine-scale pore water measurements (whole core squeezer or *in situ* oxygen microelectrode), pink - *in situ* benthic flux chamber. Where more than one measurement type has been made within a single region, the color has been assigned as described in the text.

Figure 2. Hypothetical schematic depicting the influence of CaCO_3 inputs on the evaluation of organic carbon burial rate.

Figure 3. Correlation between CaCO_3 corrected organic carbon burial rate and benthic oxygen flux. Open symbols represent Atlantic sites; closed symbols represent Pacific sites. The line is represented by the equation: $\text{O}_2 \text{ flux} = .143(\text{COB})^{.467}$ where COB = CaCO_3 -corrected organic carbon burial rate in $\text{mmol C/m}^2 \text{ yr}$.

Figure 4. Digitized sea floor distribution of organic carbon taken from Premuzic et al. (1982).

Figure 5. Digitized distribution of the sediment accumulation rate (modified as described in the text from Cwienk, 1986).

Figure 6. Organic carbon burial rate estimated by multiplying the organic carbon distribution (Fig. 4) by the sediment accumulation rate (Fig. 5).

Figure 7. Digitized sea floor CaCO_3 distribution taken from Berger et al. (1976), Biscaye et al.

(1976) and Kolla et al. (1976).

Figure 8. Distribution of sea floor oxygen flux.

Figure 9. Comparison of benthic oxygen fluxes measured directly against the values presented in Figure 8.

Figure 10. A) Comparison of the fluxes measured by sediment trap and the flux value estimated from the primary production of the overlying water and the indicated relationship with water depth. B) Depth variation in vertical particulate flux implied by the indicated expression (Berger et al. 1988).

Figure 11. The distribution of the vertical flux of organic matter (in oxygen units) at 1000 m water depth derived from the sea floor oxygen flux distribution, water depth and the expression reported by Berger et al. (1988) and displayed in Figure 10.

Figure 12. Cumulative histograms of the latitudinal distribution of the (A) sea floor flux (B) 1000 m flux (in oxygen equivalent units).

Figure 13. The specific regions integrated; resulting values are listed in Table 3.

Table 1. Comparison of basin-wide fluxes. Units are 10^{12} mol O₂/yr.

Deep Atlantic Oxygen Utilization

Broecker et al. (1991)	AOU- ¹⁴ C	30
Hinga (1985)	AOU- ¹⁴ C	17
This Study		32

Deep Pacific Oxygen Utilization

Hinga (1985)	AOU- ¹⁴ C	58
This Study		55

Global Deep Ocean Oxygen Utilization

Berger et al. (1988)	Productivity/Flux models	116
This Study		120

Table 2.

Comparison of Fluxes to Basin-Wide Primary Production* Estimates

(* From Platt et al. 1991)

LATITUDE	PP (10¹⁴ mol C/yr)	Sea Floor Remin ** (10¹² mol C/yr)	% of PP	Flux @ 1000m ** (10¹² mol C/yr)	% of PP
10°N - 10°S	2.08	2.04	.98	4.5	2.2
11°N - 37°N	2.67	2.59	.97	6.24	2.3
11°S - 20°S					
38°N - 50°N	.83	.84	1.01	1.68	2.0
50°N - 60°N	.41	.46	1.12	.78	1.9

** Converted to Organic Carbon using a C:O₂ ratio of 0.60
(TAKAHASHI et al. 1985)

Table 3.

FLUX SUMMARIES
 10^{12} mol O₂ yr.

	<u>Slope</u>	<u>Rise</u>	<u>Equator</u>	<u>Gyre</u>	<u>Total</u>
Atlantic					
Sea Floor	3.4 (23)	3.5 (24)	.53 (3)	7.4 (50)	14.8
1000m	5.3 (16)	7.7 (24)	1.3 (4)	18 (56)	32.3
Pacific					
Sea Floor	4.8 (20)	4.2 (17)	3.1 (13)	12 (50)	24.1
1000m	9.2 (17)	9.7 (18)	7.3 (13)	29 (53)	55.2
Indian					
Sea Floor	4.3 (28)	3.9 (25)	.9 (6)	6.3 (41)	15.4
1000m	7.1 (22)	8.6 (26)	2.0 (6)	15 (46)	32.7
Ocean					
Sea Floor	12.5 (23)	11.6 (21)	4.5 (8)	25.7 (47)	54.3
1000m	21.6 (18)	26.0 (22)	10.6 (9)	62.0 (52)	120.2

FIG. 1

Influence Of CaCO_3 On O.C. Burial - Remineralization Relationship (fluxes are in $\text{g/cm}^2 \text{ yr}$ except where indicated)

NON - REACTING		REACTING ($\tau_{1/2} = 500\text{yrs}$)	
No CaCO_3	CaCO_3	No CaCO_3	CaCO_3
0.1 O.C. 0.9 non - CaCO_3	0.1 O.C. 0.9 non - CaCO_3 9 CaCO_3	0.1 O.C. 0.9 non - CaCO_3	0.1 O.C. 0.9 non - CaCO_3 9 CaCO_3
10% O.C. 1 - Acc.	1% O.C. 10 - Acc.	~0.01% O.C. ~0.9 - Acc.	.5% O.C. 10 - Acc.
O.C. Burial = 0.1	O.C. Burial = 0.1	O.C. Burial = 0.09 $\text{mgC/cm}^2\text{kyr}$	O.C. Burial = 500 $\text{mgC/cm}^2\text{kyr}$

Fig. 2

Fig. 3

Fig. 4

FIG. 5

FIG. 6

FIG. 7

FIG. 8

F/6.9

FIG. 10

FIG 11

Sea Floor Flux

FIG. 12A

Rain at 1000m

FIG. 12B

FIG 13.