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VERTEX: carbon cycling in the northeast Pacific

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Abstract—Particulate organic carbon fluxes were measured with free-floating particle traps at nine locations during VERTEX and related studies. Examination of these data indicated that there was relatively little spatial variability in open ocean fluxes. To obtain mean rates representative of the oligotrophic environment, flux data from six stations were combined and fitted to a normalized power function, $F = F_{100}\ (z/100)^{b}$; e.g. the open ocean composite C flux in mol $m^{-2}\ y^{-1} = 1.53\ (z/100)^{-0.858}$ with depth z in meters. It is shown that the vertical derivative of particulate fluxes may indicate solute regeneration rates, and accordingly regeneration rates for C, H and N were estimated. Oxygen utilization rates were also estimated under the assumption that 1.5, 1.0 and 0.25 moles of O_2 were used for each mole of N, C and H regenerated. Regeneration ratios of these elements were depth-dependent: i.e.

$$N{:}C{:}H{:}{-}O_2 =$$

$$1.0\ N{:}\ 6.2\ (z/100)^{0.130}\ C{:}\ 10.0\ (z/100)^{0.146}\ H{:}\ [1.5 + 6.2\ (z/100)^{0.130} + 2.5\ (z/100)^{0.146}] - O_2.$$

Comparisons of our rates with those in the literature indicate that trap-derived new productivities in the open Pacific (~1.5 mol C $m^{-2}\ y^{-1}$) are substantially less than those estimated from oxygen utilization rates in the Sargasso Sea (~4 mol C $m^{-2}\ y^{-1}$). A hypothesis is presented which attempts to explain this discrepancy on the basis of the lateral transport and decomposition of slow or non-sinking POC in the Sargasso Sea.

Data gathered during the VERTEX studies are also used for various global estimates. Open ocean primary productivities are estimated at 130 g C $m^{-2}\ y^{-1}$ which results in a global open ocean productivity of 42 Gt y^{-1}. Organic C removal fom the surface of the ocean via particulate sinking (new production) is on the order of 6 Gt y^{-1}. Fifty percent of this C is regenerated in the upper 300 m of the water column. The ratio of new production (measured with traps) to total primary production (measured via ^{14}C) is 0.14. It is concluded that the ^{14}C technique yields reasonable estimates of primary productivity provided that care is taken to prevent heavy metal contamination.

INTRODUCTION

THERE is a great deal of interest in oceanic organic C fluxes since several important problems hinge on their accurate measurement. For example, the organic matter passively sinking out of the euphotic zone represents new production (DUGDALE and GOERING, 1967; EPPLEY and PETERSON, 1979), and with the decomposition of this material, dissolved oxygen is consumed. Measurements of this oxygen consumption (JENKINS, 1982, 1984; JENKINS and GOLDMAN, 1985) suggest that open ocean new

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production is much higher (~4 mol C m^{-2} y^{-1}) than previously believed. This implies that the ^{14}C technique may seriously underestimate rates of primary production if *f* factors (ratios of new production to total production) typical of oligotrophic oceans are applied (EPPLEY and PETERSON, 1979). On the other hand, the use of the average *f* factors of PLATT and HARRISON (1985) indicates that primary productivities as currently measured are representative of those occurring *in situ*. Clearly, knowledge of total primary and new production is of fundamental importance in understanding the ocean, and additional information on this problem is needed.

C, H, N regeneration rates and the estimated oxygen utilization rates (OURs) resulting from them are also used in the calculation of initial (preformed) PO_4 and NO_3 concentrations. These initial nutrient concentrations together with potential temperatures and salinities enable the determination of the sources and flow patterns of deep ocean waters (BROECKER *et al.*, 1985). Recent studies of dissolved distributions along isopycnal surfaces indicate that P:N:C:$-O_2$ ratios, representing the decomposition of organic matter, are 1:16:122(±18):172 (TAKAHASHI *et al.*, 1985) rather than the classical Redfield ratio of 1:16:106:138 (REDFIELD *et al.*, 1963). Obviously, confirmation of these new ratios by independent techniques is needed, not only to gain further information on deep ocean circulation but also to better understand the formation and maintenance of oxygen minima systems in the ocean (e.g. WYRTKI, 1962).

All of the above are also important in understanding the global CO_2 cycle. One of the most fundamental estimates is that for the annual biogenic removal of carbon from the surface to the deep ocean in relation to global anthropogenic CO_2 input (5 Gt y^{-1}, SUNDQUIST, 1985). To provide these rates, organic C flux measurements are needed together with information on how fast this C is regenerated in various portions of the water column.

During the past 8 years we have compiled a large data set based on 123 individual C, H, N flux measurements made during VERTEX (Vertical Transport and Exchange) and related studies. The purpose of this paper is to present these data, interpret them, and discuss our findings in relation to the problems mentioned above.

METHODS AND MATERIALS

Free-floating particle interceptor traps (MULTITRAPS) were deployed at nine stations to measure the vertical fluxes of organic C, H and N associated with passively sinking particles. With the exception of one station off Peru, all were located in the northeast Pacific (Fig. 1). Other pertinent information about time and place of sampling is summarized in Table 1. Trap locations were monitored using the Argos satellite system.

The traps consist of PVC crossframes fitted with 12 identical polycarbonate cylinders filled with a high-density solution (seawater with added NaCl; ρ = 1.05 g cm^{-3}) containing 1% buffered formalin. All internal parts of the traps were thoroughly cleaned with an organic detergent (Micro)® and dilute quartz-distilled HNO_3. Further details about the traps have been published elsewhere (KNAUER *et al.*, 1979). These traps were tested in the Sediment Trap Intercomparison Experiment (STIE). Fluxes were comparable to those obtained by Honjo, Gardner, Soutar and Farrington using other trap systems (ANDERSON and BACON, 1981).

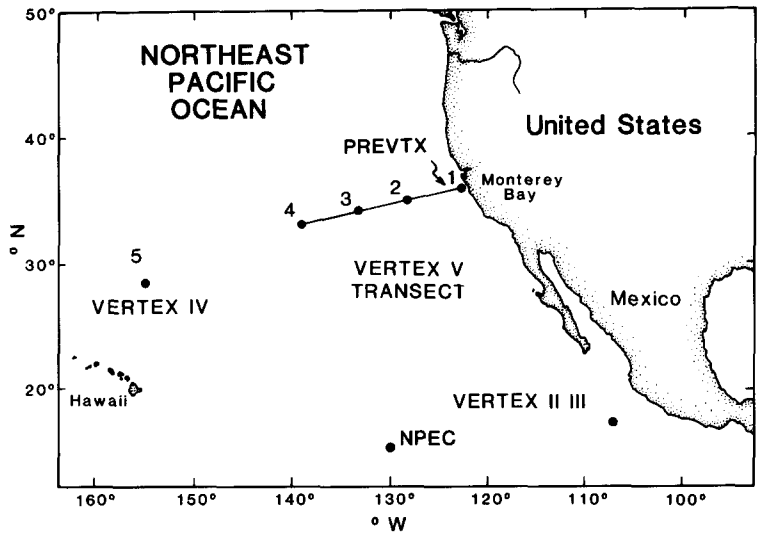


Fig. 1. VERTEX station locations.

Zooplankton swimmers (KNAUER *et al.*, 1984b) were meticulously picked out of all trapped particulates prior to drying and weighing. Carbon, hydrogen and nitrogen amounts in the trapped particulates were determined using a Hewlett-Packard model 185-B CHN analyzer after washing with dilute HCl to remove carbonate-C.

RESULTS

This paper is based on 123 flux estimates together with analyses of trapped particles for C, H, N for a total of 369 measurements; obviously, it is impractical to present all of these data in tabular form. As an alternative, we attempted the representation of these rate measurements with various interpolation functions.

The best fit was obtained with a log-log transformation and the use of a normalized power function of the form:

$$F = F_{100}\left(\frac{z}{100}\right)^{b}, \tag{1}$$

where F is the flux at any depth, F_{100} is the log-log intercept, z is depth in meters, and exponent b is the log-log slope. This form of the power function is a homogeneous equation and provides meaningful units to the intercept. It is emphasized that this equation is used as a statistical expedient, it is not based on *a priori* physics, but compared to other smoothing techniques tried by us, it produced excellent statistical results. For the 9 trap sets, r^2 ranged from 0.93 to 0.99 (see examples in Fig. 2), and in Table 1 we present the statistically determined intercepts, F_{100}, and exponents, b at each station. Equation (1) is then used as an interpolation function to estimate C, H, N fluxes between measured depths (100–2000 m) and to extrapolate to greater depths.

Table 1. Station locations, dates, number of days traps were set, primary productivities (mol C m^{-2} y^{-1}), *photic zone depths* (z *in* m), *trap depth ranges* (*in* m), *total number of flux measurements within depth range* (n), *and data deleted for various VERTEX* (1, 2, 4, 5, II *and* III) *and non-VERTEX* (*PREVTX, PERU and NPEC*) *stations*

	PREVTX	PERU	1	2	4	5	II	III	NPEC	OOC§
Latitude	37.0N	15.25S	36.1N	35.0N	33.3N	28.0N	18.0N	15.7N	14.0N	–
Longitude	123.0W	75.9W	122.6W	128.2W	139.2W	155.0W	108.0W	107.5W	130.0W	–
Month	Dec.	June	June	June	June	Jul./Aug.	Oct./Nov.	Nov.	Apr./May	–
Year	1978	1983	1984	1984	1984	1983	1981	1982	1980	–
Days set	6	8	13	22	22	34	22	21	26	–
Prim. prod.	20.9	22.2	34.7	9.88	7.45	12.2	23.1	14.3	–	–
Photic *z*	50	50	50	125	125	150	100	100	–	–
Trap *z* range	35	50	50	130	150	50	120	80	125	120
	1500	1000	2000	2000	2000	2000	1950	2000	900	2000
n	13	10	13	14	13	9	16	25	10	87
Data del.	1500†	100‡	50†	–	–	50†	–	80†	–	–
Carbon, hydrogen, nitrogen fluxes (mol m^{-2} y^{-1})										
C F_{100}	3.47	3.20	7.08	2.37	2.05	1.17	1.26	1.12	2.74	1.53
b	–0.863	–0.319	–0.833	–0.878	–0.968	–0.973	–0.805	–0.648	–0.862	–0.858
r^2	0.99	0.96	0.98	0.97	0.98	0.95	0.98	0.93	0.96	0.81
*n**	5	6	7	7	7	9	8	13	4	48
H F_{100}	6.84	4.84	14.8	4.39	2.75	1.48	2.23	1.88	5.43	2.53
b	–0.790	–0.305	–0.729	–0.850	–0.805	–0.883	–0.867	–0.680	–0.918	–0.842
r^2	0.97	0.99	0.93	0.97	0.97	0.91	0.97	0.93	0.97	0.75
N F_{100}	0.462	0.432	0.906	0.251	0.288	0.167	0.186	0.153	0.440	0.216
b	–1.082	–0.442	–0.914	–1.004	–1.145	–1.110	–0.874	–0.722	–1.007	–0.988
r^2	0.99	0.97	0.98	0.95	0.98	0.95	0.98	0.90	0.94	0.84
Carbon, hydrogen, nitrogen regeneration rates (mmol m^{-3} y^{-1})										
C R_{100}	30.1	10.2	58.9	20.9	19.1	11.3	10.2	7.23	23.6	13.1
b–1	–1.863	–1.319	–1.833	–1.878	–1.968	–1.973	–1.805	–1.648	–1.862	–1.858
H R_{100}	53.9	14.7	108	37.3	22.1	13.1	19.4	12.7	49.8	21.3
b–1	–1.790	–1.305	–1.729	–1.850	–1.805	–1.883	–1.867	–1.680	–1.918	–1.842
N R_{100}	5.00	1.91	8.29	2.52	3.29	1.85	1.62	1.10	4.43	2.13
b–1	–2.082	–1.442	–1.914	–2.004	–2.145	–2.110	–1.874	–1.722	–2.007	–1.988

Carbon, hydrogen, nitrogen flux data are expressed using the normalized power function, $F = F_{100}\,(z/100)^b$, where F is the flux at any depth, F_{100} is the flux at 100 m (the log-log intercept), z is the depth in meters, and the exponent, b, is the log-log slope. Values are also given for an open ocean composite (OOC; see text). C, H, N regeneration rates (R), based on the depth derivatives of the fluxes are also given again using the relationship mmol m^{-3} $y^{-1} = R_{100}\,(z/100)^{b-1}$; R_{100} is the regeneration rate at 100 m.

* *n* based on *means* of replicates for each depth used.

† Both replicates deleted at depths shown.

‡ One replicate deleted at depths shown.

§ Open ocean composite based on data from Stas 2, 4, 5, II, III, and NPEC.

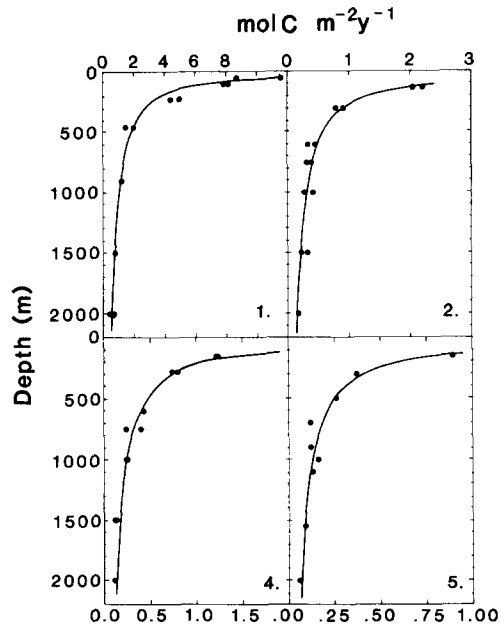


Fig. 2. Trap-derived C flux data fitted with the normalized power function, $F = F_{100}(z/100)^b$, using replicate measurements from Stas 1 ($r^2 = 0.96$; $n = 13$), 2 ($r^2 = 0.94$; $n = 14$), 4 ($r^2 = 0.97$; $n = 13$) and 5 ($r^2 = 0.95$; $n = 9$).

Carbon fluxes

The carbon fluxes for the 8 northeast Pacific stations are shown together in Fig. 3. As expected, the highest C fluxes were measured off Point Sur, California (Sta. 1), under intensive upwelling conditions, while the lowest were measured in the Central Pacific Gyre (Sta. 5) under summer maximum stratification conditions. It is interesting that the range between these two extremes is not large. For example, the 100 m flux at Sta. 1 (7.1 mol C m^{-2} y^{-1}) was only 6 times higher than that measured at Sta. 5 (100 m flux = 1.2 mol C m^{-2} y^{-1}), while at 2000 m, the fluxes at Sta. 1 were only 9 times as high (0.58 and 0.063 mol C m^{-2} y^{-1}) as those at Sta. 5. General onshore–offshore variability is also shown in a C flux section (Fig. 4) based on data from Stas 1, 2, 4, and 5.

It is also noteworthy that fluxes for the six offshore stations (2, 4, 5, II, III and NPEC) varied by a factor of only 2.4. For example, C fluxes at 100 m were: 2.37 for the western edge of the California Current (Sta. 2); 2.05 for the transition region between the California Current and Subtropical Gyre (Sta. 4); 1.17 for the Subtropical Gyre (Sta. 5); 1.26 and 1.12 for Stas II and III 400 km off Mexico and 2.74 for the North Pacific Equatorial Current (NPEC) all in units of mol C m^{-2} y^{-1}. Perhaps this is to be expected since all of these locations share common features: nearly complete nutrient depletion in the mixed layer and subsurface chlorophyll maxima located in or above the nutricline.

In recognition of the need for generalized estimates for the open ocean environment (e.g. carbon distributions and fluxes, see SUNDQUIST, 1985), and in view of the limited offshore spatial variability noted above, we combined the data from the six offshore

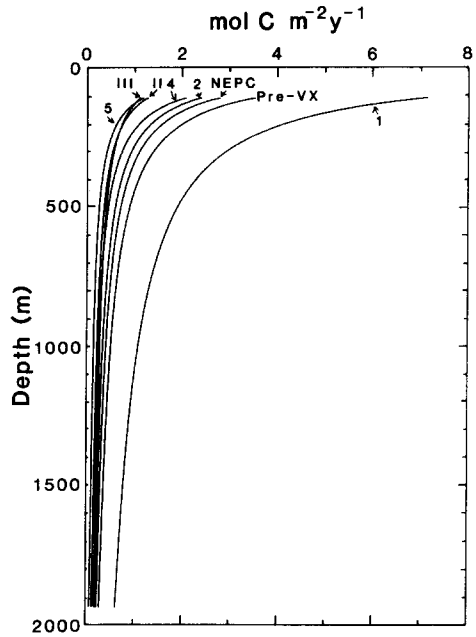


Fig. 3. Carbon fluxes for the eight northeast Pacific stations using the normalized power function, $F = F_{100}(z/100)^b$, and the coefficients in Table 1.

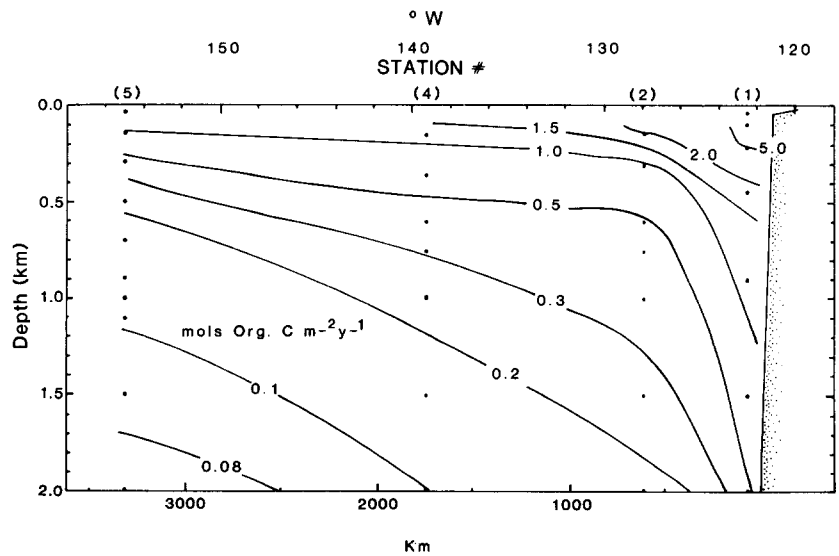


Fig. 4. A California–Hawaii C flux section. Station locations are shown in Fig. 1.

stations (2, 4, 5, II, III and NPEC) in order to derive what we refer to as open ocean composite (OOC) C, H, and N fluxes. Means of the replicates for a given station and depth were used to avoid weighting the data towards the stations with more replicates per depth (e.g. NEPC, n = 2–3) than those with few replicates per depth (e.g. Sta. 5, n = 1). Fitting these data with the power function from equation (1) resulted in estimated open ocean fluxes of mol C m^{-2} y^{-1} = 1.53 $(z/100)^{-0.858}$ (r^2 = 0.81); mol H m^{-2} y^{-1} = 2.53 $(z/100)^{-0.842}$ (r^2 = 0.75); and mol N m^{-2} y^{-1} = 0.216 $(z/100)^{-0.988}$ (r^2 = 0.84) with z in meters (see Fig. 5).

Carbon, hydrogen, nitrogen regeneration rates

One of our goals is to interpret C, H, N particle flux distributions in terms of solute concentration rates of change. In the following we show under what assumptions one can relate changes in particle fluxes to the regeneration of dissolved materials.

In the one-dimensional, steady-state, advection–diffusion equation that describes the distribution of a non-conservative solute, S, the local rate of change due to diffusion (with diffusivity, K_z), advection (with vertical velocity, w) and *in situ* production (R_s) can be considered to balance where lateral gradients are nil,

$$\frac{\partial}{\partial z} K_z \frac{\partial S}{\partial z} - \frac{\partial}{\partial z} (wS) + R_s = 0. \tag{2}$$

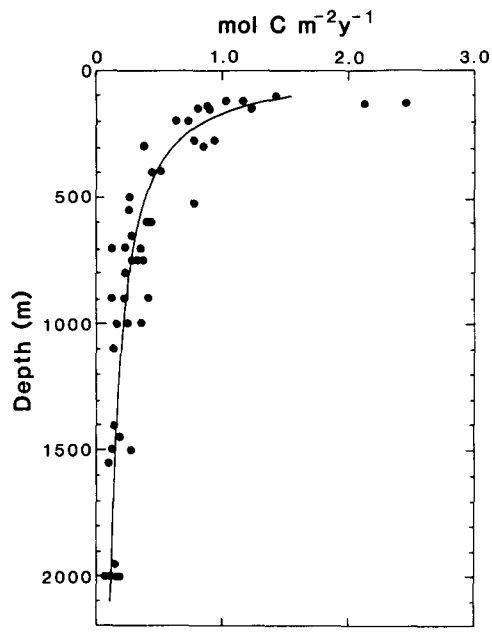


Fig. 5. Open ocean composite (OOC) fluxes for C using the means of replicates at various depths from Stas 2, 4, 5, II, III and NPEC: $F = 1.53(z/100)^{-0.858}$; $r^2 = 0.81$; $n = 48$.

The vertical integral of equation (2) evaluated between limits z_1 and z_2 (with z positive down) produces a local flux balance relation,

$$K_z \frac{\partial S}{\partial z}\bigg|_{z_2} - K_z \frac{\partial S}{\partial z}\bigg|_{z_1} - wS|_{z_2} + wS|_{z_1} + \int_{z_1}^{z_2} R_s \, dz = 0, \tag{3}$$

where the first two terms on the left represent the net diffusive flux of the solute across the vertical boundaries, the second two terms are the net advective flux, and the right-most term is the depth integral of solute production between z_1 and z_2. Equations such as this have been used extensively (e.g. WYRTKI, 1962) to model solute distributions.

If the non-conservative right-most solute term (equation 3) can be construed to represent the net *removal* rate of the solute by its incorporation into particles of concentration, P, then

$$-\int_{z_1}^{z_2} R_s \, dz = +\int_{z_1}^{z_2} R_P \, dz, \tag{4}$$

and the steady-state, vertical particle distribution is given by

$$\frac{\partial}{\partial z} K_z \frac{\partial P}{\partial z} - \frac{\partial}{\partial z}(wP) + R_P = 0. \tag{5}$$

In the case of particles, it may be argued that vertical particle diffusion is unimportant for large spatial scales, and hence that the vertical particle distribution represents a balance between particle advection (including their gravitational sinking) and *in situ* production. The vertical integral of equation (5) is

$$-wP|_{z_2} + wP|_{z_1} + \int_{z_1}^{z_2} R_P \, dz = 0,$$

and from (equation 4) it follows that

$$-wP|_{z_2} + wP|_{z_1} = \int_{z_1}^{z_2} R_s \, dz,$$

or equivalently that

$$-\frac{\partial}{\partial z}(wP) = R_s. \tag{6}$$

The left-hand term in equation (6) is the vertical divergence of the advective particle flux, where w is the net particle sinking speed and P is the particle concentration. Exact interpretation of both quantities is still elusive, but it may be assumed under some circumstances that the product wP is proportional to fluxes measured by particle traps. Therefore, one interpretation of the vertical gradient of measured particle fluxes is that they represent the local *in situ* solute regeneration rate. We will show that this interpretation is remarkably consistent with other inferences on the ratios of change of major nutrient elements in the sea, in spite of the obviously stringent assumptions imposed by ignoring lateral particle advection. When the normalized power function is

used to represent the vertical particle flux distributions, so that continuous derivatives can be obtained, and the aforementioned assumptions hold, then the *in situ* regeneration rates are

$$R_s \approx -\frac{\partial F}{\partial z} = R_{100}\left(\frac{z}{100}\right)^{b-1}, \tag{7}$$

where R_{100} is the regeneration rate at 100 m ($R_{100} = F_{100}\, b/100$). Representative C, H, N regeneration rates, R_s, are shown in Fig. 6 and Table 2 for OOC data.

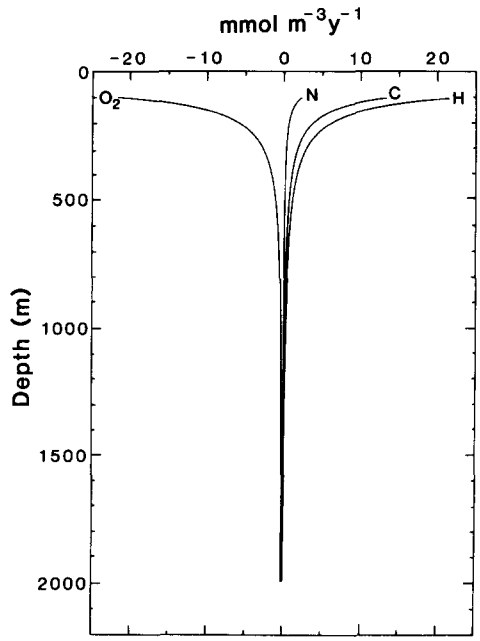


Fig. 6. Oxygen utilization rates and C, H, N regeneration rates vs depth using OOC data.

Table 2. Examples of C, H, N *fluxes and regeneration rates together with oxygen utilization rates for various depths using equation* (1) *and OOC coefficients from Table* 1

Depth (m)	C	H	N	C	H	N	O_2
	(mol m^{-2} y^{-1})			(mmol m^{-3} y^{-1})			
100	1.53	2.53	0.22	13.1	21.3	2.13	-21.6
500	0.38	0.65	0.044	0.66	1.10	0.087	-1.07
1000	0.21	0.36	0.022	0.18	0.31	0.022	-0.29
1500	0.15	0.26	0.015	0.86	0.15	0.010	-0.14
2000	0.12	0.20	0.011	0.050	0.086	0.006	-0.080
3000	0.083	0.14	0.008	0.024	0.041	0.002	-0.37
4000	0.065	0.11	0.006	0.014	0.024	0.001	-0.022
5000	0.053	0.094	0.005	0.009	0.016	0.001	-0.014

Oxygen utilization rates

These regeneration rates can also be used for the estimation of oxygen utilization rates (OURs). This usually involves use of the Redfield ratio of −138 moles of O_2 consumed for 106 moles of organic C decomposed ($-O_2$:C = 1.30; REDFIELD *et al.*, 1963). However, a recent study by TAKAHASHI *et al.* (1985), based on the distribution of dissolved species along isopycnal surfaces, suggests that the $-O_2$:C ratio is 172:122 ± 18, or in the range of −1.65:1 to −1.23:1. The uncertainty in the C value arises from a lack of understanding concerning the anthropogenic CO_2 dissolved in the water and the amount of oxygen used in the oxidation of organic H.

Since our trapped particulates were analysed for C, H and N, we can estimate oxygen consumption rates under the assumption that all of the regenerated C, N and H have been fully oxidized, and that: 1 mol of O_2 was required for the oxidation of each mol of C regenerated; 1.5 mol of O_2 were required for each mol of N (excluding amino H); and 0.25 mol of O_2 were required for each mol of organic H (including amino H) regenerated. Stoichiometrically, this would be $1C + 1O_2 \rightarrow 1CO_2$; $1N + 1.5O_2 \rightarrow 1NO_3$ and $4H + 1O_2 \rightarrow 2H_2O$.

Thus, oxygen consumption rates using our OOC data can be estimated as: mmol O_2 m^{-3} y^{-1} consumed = 1.0 (13.1 $(z/100)^{-1.858}$) + 0.25 (21.3 $(z/100)^{-1.842}$) + 1.5 (2.13 $(z/100)^{-1.988}$) for the mmol of C, H and N regenerated m^{-3} y^{-1}, respectively (Table 1). The oxygen consumed at 100 m by the oxidation of C, H, N in the above example is 13.1, 5.3, and 3.2 for a total of 21.6 mmol of O_2 consumed m^{-3} y^{-1}. This suggests that 60% of the oxygen is used for C, 25% for H, and 15% for N oxidation.

N:C:H: $-O_2$ *Ratios*

The apparent OURs mentioned above together with our CHN regeneration rate data from Table 1 can be used to estimate ratios for various depths in the water column. Again using our OOC data as an example, the depth-dependent N:C:H: $-O_2$ ratios become 2.12 $(z/100)^{-1.988}$ N:13.1 $(z/100)^{-1.858}$ C:21.3 $(z/100)^{-1.842}$ H:(1.0 C + 0.25 H + 1.5 N) $-O_2$. Ratios, normalized to N and P (assuming 16N:1P), are shown in Table 3.

Examples of these ratios based on CHN amounts regenerated into the water column and apparent oxygen consumed from the water column are shown in Table 4 at selected depths from our working range (100–2000 m) together with extrapolated ratios for

Table 3. Depth-dependent molar ratios based on OOC amounts regenerated into the water column, together with estimated oxygen consumption, (3/2N + 1C + H/4). Ratios for OOC trapped particles are also shown. Ratios are normalized to P *(assuming* N: = 16:1*) for ease of comparison with Redfield ratios*

	P	:	N	:	C	:	H	:	$-O_2$
Regeneration ratios									
Norm. to N	–	:	1	:	$6.2(z/100)^{0.130}$	:	$10.0(z/100)^{0.146}$	:	$[1.5 + 6.2(z/100)^{0.130} + 2.5(z/100)^{0.146}]$
Norm. to P	1	:	16	:	$99(z/100)^{0.130}$	:	$160(z/100)^{0.146}$	:	$[24 + 99(z/100)^{0.130} + 40(z/100)^{0.146}]$
Trapped particulate or flux ratios									
Norm. to N	–	:	1	:	$7.1(z/100)^{0.130}$	:	$11.7(z/100)^{0.146}$	:	$[1.5 + 7.1(z/100)^{0.130} + 2.9(z/100)^{0.146}]$
Norm. to P	1	:	16	:	$114(z/100)^{0.130}$	:	$187(z/100)^{0.146}$	:	$[24 + 114(z/100)^{0.130} + 47(z/100)^{0.146}]$

Table 4. Examples of various OOC molar ratios based on regeneration rates and fluxes using data from Table 3

Depth (m)	N:P	C:P	H:P	$-O_2$:P	C:N	H:N	$-O_2$:N	H:C	$-O_2$:C	$-O_2$:H
Regeneration ratios										
100	16	99	160	163	6.2	10.0	10.2	1.61	1.65	1.02
500	16	122	202	196	7.6	12.6	12.3	1.65	1.61	0.97
1000	16	134	224	214	8.4	14.0	13.4	1.67	1.60	0.95
1500	16	141	238	224	8.8	14.8	14.0	1.68	1.59	0.94
2000	16	146	248	232	9.2	15.5	14.5	1.69	1.59	0.94
3000	16	154	263	244	9.6	16.4	15.3	1.70	1.58	0.93
4000	16	160	274	252	10.0	17.1	15.8	1.71	1.58	0.92
5000	16	164	283	259	10.3	17.7	16.2	1.72	1.57	0.92
Trapped particle or flux ratios										
100	16	114	187	185	7.1	11.7	11.5	1.65	1.62	0.98
500	16	141	237	224	8.8	14.8	13.9	1.68	1.59	0.94
1000	16	154	262	244	9.6	16.4	15.1	1.70	1.58	0.92
1500	16	162	278	256	10.1	17.4	15.9	1.71	1.57	0.92
2000	16	168	290	265	10.5	18.1	16.5	1.71	1.57	0.91
3000	16	177	307	279	11.0	19.2	17.3	1.72	1.56	0.90
4000	16	184	320	289	11.5	20.0	17.9	1.73	1.56	0.89
5000	16	190	331	297	11.8	20.7	18.4	1.73	1.56	0.89

depths of 3000, 4000 and 5000 m. With the inclusion of O_2 consumed via H oxidation, our $-O_2$:C ratios are much higher than the Redfield value of 1.30 and in agreement with those of TAKAHASHI *et al.* (1985) when their low C value (122 – 18 = 104 C) is used (–172 O_2:104C = 1.65). It is also apparent that ranges in H:C ratios (1.61–1.72), $-O_2$:C ratios (1.65–1.57), and $-O_2$:H ratios (1.02–0.92) are relatively small over the 100–5000 m depth interval. However, since N and P are regenerated much more rapidly than C and H, the ranges in $-O_2$:N and $-O_2$:P ratios increase markedly from 10.2 to 16.2 and 163 to 259, respectively, in the 100–5000 m depth interval. In general, our northeast Pacific $-O_2$:P ratios are in fairly good agreement with those estimated by Takahashi *et al.* in the main thermoclines of the southern and northern Atlantic and Indian oceans (~200 vs their 172). However, in the deeper Pacific and Indian Ocean waters (~3500 m), the value of BROECKER *et al.* (1985) is less (183 vs ~250) than ours.

DISCUSSION

As stated in the Introduction, one of the most controversial and important problems facing oceanographers today is the questionable accuracy of primary productivity rates as measured by the ^{14}C technique. It is interesting to note that this questioning arose not from alternative direct C rate measurements but from measurements of dissolved oxygen in the water column; i.e. either as excess oxygen in the photic zone as a result of summer stratification (SCHULENBERGER and REID, 1981) or as apparent oxygen utilization beneath the photic zone (JENKINS, 1982; JENKINS and GOLDMAN, 1985). In the latter case, their estimates indicate that the amounts of C leaving the photic zone (*new* productivity) must be on the order of 4 mol C m^{-2} y^{-1} or 50 g C m^{-2} y^{-1} to fuel the OURs they measured. The latter value is typical of *total* primary productivity rates thought to be occurring in oligotrophic waters (e.g. RYTHER, 1969). Hence, this leads to the conclusion that: "^{14}C

assimilation techniques are in error and that oceanic 'oligotrophic' primary production is considerably higher than previously thought" (JENKINS, 1982).

The data presented here are of relevance to this problem in several ways. Our studies involve C rate measurements; both total productivity measured via the ^{14}C technique in the photic zone and new productivity assuming that the particulate organic C we collect in near-surface traps fell out of the overlying photic zone. The elemental ratios we measured are also of relevance since these are used to translate OURs into CO_2 regeneration rates, and vice versa. For this reason, we will discuss the elemental ratios first. Next we will compare the OURs we estimate with others from the literature and then we will relate our trap-derived C fluxes to the primary productivity rates we measured in the overlying surface waters. Finally, we will discuss these findings in relations to the global CO_2 cycle.

Depth-dependent elemental ratios

Conversion between OURs and CO_2 regeneration rates depends on the $-O_2$:C ratio used and until 1985, the Redfield ratio of $-138\ O_2:106$ C ($-1.30\ O_2:1.0$ C) was commonly used (REDFIELD *et al.*, 1963). However, in 1985 Takahashi *et al.* published new ratios of $-172\ O_2:122 \pm 18$ C (see Results) and here we present depth-dependent ratios estimated from our trap-derived N, C, H regeneration and $-O_2$ consumption rates (see Tables 3 and 4).

BROECKER *et al.* (1985) state that Δ AOU/$\Delta\ PO_4$ ratios are nearly constant with depth and geographic location (175 ± 6), based on the deconvolution of chemical data along various isopycnals in the Atlantic, Indian and Pacific oceans. With the use of our OOC and examination of individual station data, it appears that these ratios vary little with geographical location. However, our results clearly indicate that these ratios increase markedly with depth because of the more rapid regeneration of P and N in relation to C regeneration and O_2 consumption (e.g. $-O_2:P = 163$ at 100 m vs 259 at 5000 m).

Although we cannot state categorically that our ratios are more accurate than others, we can at least see if they make sense, by estimating some initial NO_3 concentrations. Initial (preformed) nitrate and phosphate concentrations are used in conjunction with potential temperature and salinity to deduce sources and flow patterns for deep ocean waters (BROECKER *et al.*, 1985). They are estimated by dividing AOU by the ratio of O_2 consumed to regenerated PO_4 or NO_3. This results in the amounts of regenerated P or N, and these values subtracted from the observed PO_4 and NO_3 concentrations are equal to the initial concentrations carried by these waters when they sank away from the surface. Thus, the choice of $-O_2$:N or $-O_2$:P ratio to be used becomes very important. To illustrate, we calculated initial NO_3 concentrations for GEOSECS Sta. 246 (Fig. 7) which is located on the equator north of the Samoan Passages (0°; 179°E; BROECKER *et al.*, 1982); here deep South Pacific water flows into the North Pacific. The use of our depth-dependent $-O_2$:N ratios indicates that the deep (>3000 m) waters have initial NO_3 concentrations of 24.7 ± 0.1 ($n = 16$) in comparison to 16.0 ± 0.8 and 20.1 ± 0.3 μmol NO_3° kg^{-1}, when the depth-independent ratios of Redfield *et al.* and Broecker *et al.* are used. Thus, the use of our constantly increasing $-O_2$:N ratios not only results in higher NO_3° values but also a lower standard deviation for the 16 GEOSECS 246 samples at depths between 3117 and 5403 m.

We also calculated initial NO_3 concentrations for southern component water using our depth-dependent $-O_2$:N ratios together with the data in Table 3 of BROECKER *et al.*

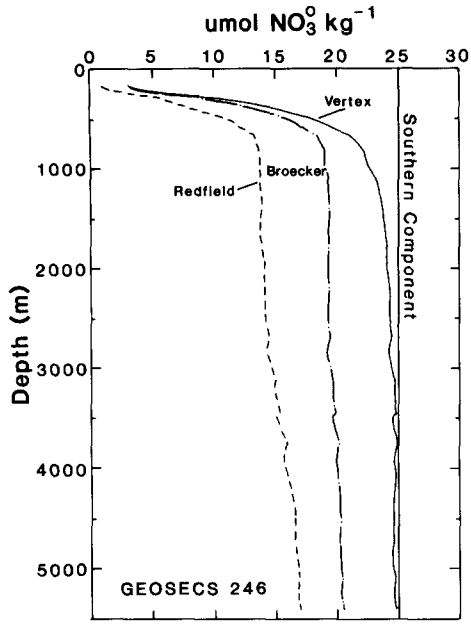


Fig. 7. Estimates of initial nitrate concentrations using depth-independent $-O_2$:N ratios of REDFIELD *et al.* (1963) and BROECKER *et al.* (1985) in comparison to those obtained with the depth-dependent $-O_2$:N ratios from this study.

(1985). This resulted in a mean of 25.1 ± 0.3 μmol NO_3^o kg^{-1} (n = 31). Dividing this value by a N:P ratio of 16:1 yields an initial PO_4 concentration of 1.57 μmol kg^{-1} which agrees well with the 1.58 μmol kg^{-1} value used by Broecker *et al.* in their Table 4. The use of our depth-dependent ratios indicates that the deep water north of the Samoan Passages (GEOSECS 246) consists of 98% southern component water type in comparison to 64 and 80% when the depth-independent ratios of Redfield *et al.* and Broecker *et al.* are used.

We believe that deep Samoan Passage water consisting of 98% southern component type makes sense and, consequently, gives us confidence that the higher $-O_2$:C:N ratios we report here are accurate. This information supports TAKAHASHI *et al.* (1985) and BROECKER *et al.* (1985) and indicates that converting from OURs to CO_2 regeneration rates and vice versa is best accomplished using an $-O_2$:C ratio of 1.6 (see Table 4) rather than the Redfield ratio of 1.3.

Oxygen utilization rates

Establishing an accurate $-O_2$:C ratio is important since we are primarily interested in C and there are few estimates of CO_2 regeneration, while there are many estimates of oxygen utilization made using a variety of techniques: advection–diffusion models (e.g. RILEY, 1951; WYRTKI, 1962; FIEDIERO and CRAIG, 1978), electron transport system (ETS) activity (PACKARD and WILLIAMS, 1981; GARFIELD *et al.*, 1983), AOU to $^3H/^3He$ age relationships (JENKINS, 1982), and the examination of long time series hydrographic records (JENKINS and GOLDMAN, 1985). Particle traps also have been used for OUR estimates. For example, KNAUER *et al.* (1979) estimated OURs based on C fluxes and

finite difference rates of change, while SUESS (1980) combined C flux data from several particle trap studies and developed a statistical model which enables estimates of OURs from the entire water column based on the primary productivity occurring at the surface.

As noted in Results, the normalized power function (equation 1) can be used over the entire water column, whereas the use of models such as $R = R_0 e^{-\alpha z}$ with constant α (WRYTKI, 1962) will fit in the upper waters where rates are decreasing rapidly or in the deeper waters where rates are decreasing slowly but not in both regions. For example, WYRTKI'S (1962) rates are fairly similar to ours in the upper water column (Table 5), but are far less when extrapolated to 5000 m (4×10^{-7} vs 0.014 mmol O_2 m^{-3} y^{-1}). FIEDIERO and CRAIG'S (1978) estimates based on the deep Pacific are also fairly similar to ours in the deeper waters (1000–5000 m, Fig. 8); however, extrapolating their rates to depths of 100 m results in unrealistically low values of 1.41 mmol O_2 m^{-3} y^{-1}.

All things considered, the agreement in OURs shown in Table 5 is reasonable provided that the first-order rate estimates are not extrapolated beyond the depths for which they were originally intended. With this caveat, the data in Table 5 could be considered to be in reasonable agreement and dismissed as such. However, recent studies by JENKINS (1982) and JENKINS and GOLDMAN (1985) indicate that knowledge of the accuracy of these various OUR estimates is very important since OURs are directly related to the magnitude of "new" primary productivity occurring in the overlying surface waters.

Our open ocean composite (OOC) oxygen utilization rates are compared with those of JENKINS (1982) and JENKINS and GOLDMAN (1985) in Fig. 8. It is apparent that our rates and theirs are fairly similar just beneath the photic zone. For example, our 100 m OUR (21.6) exceeds Jenkin's 100 m estimate of 16.8 mmol O_2 m^{-3} y^{-1}. However, at greater depths our OURs decrease much more rapidly than theirs. Our curve suggests the organic matter consumers (microorganisms and/or zooplankton) are utilizing most of the organic matter just beneath the photic zone where food quantities are greatest, food quality is best (more protein and readily oxidizable substrates), and temperature- and

Table 5. Comparisons of oxygen utilization rates based on data from sources in footnotes

Depth (m)	1	2	3	4	5	6
			(mmol m^{-3} y^{-1})			
100	21.6	9.9	16.8	–	6.4	17.4
150	10.1	8.3	14.5	24.1	3.8	12.2
200	5.9	7.0	12.5	17.4	2.6	9.5
250	3.9	5.9	10.8	11.6	2.0	7.9
300	2.8	4.9	9.3	13.8	1.6	6.7
350	2.1	4.1	8.0	8.9	1.2	5.9
400	1.6	3.5	6.9	6.7	1.1	5.2
500	1.1	2.4	5.2	–	0.8	4.3

1. OOC data from this study, using particle trap-derived CHN fluxes.
2. From WYRTKI (1962) based on RILEY'S (1951) three-dimensional advection diffusion calculations.
3. From JENKINS (1982) based on AOU-tritium/helium age.
4. From JENKINS and GOLDMAN (1985), based on long-term hydrographic records.
5. From GARFIELD *et al.* (1983), based on electron transport activity measurements in the North Atlantic.
6. Same as 5 for eastern Pacific, off Mexico.

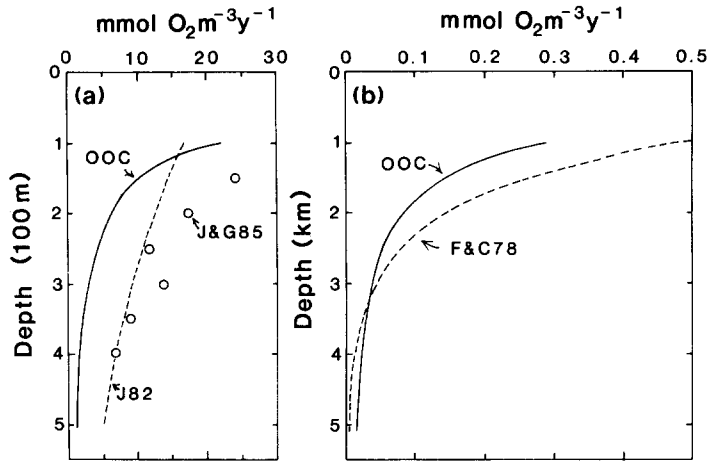


Fig. 8. Comparisons of OURs obtained in this study (OOC data) vs those of JENKINS (1982) and JENKINS and GOLDMAN (1985) for depths of 100–500 m (a) and vs those of FIADEIRO and CRAIG (1978) at depths of 1–5 km (b).

pressure-dependent metabolic rates are highest. Jenkins and Goldman's data suggest the same; however, the rates decrease much less rapidly with depth.

With our rapid decrease in OURs, sufficient organic matter is left to support deep ocean fluxes. For example, extrapolation of our C fluxes, upon which our OURs are based, to a depth of 3200 m gives a C flux of 0.078 mol C m^{-2} y^{-1}; this rate falls towards the high side of the range (0.033–0.084 mol m^{-2} y^{-1}) measured by DEUSER (1984, 1986) in a 6-year particle trap study in the same Sargasso Sea study area of Jenkins and Goldman.

The oxygen consumption we estimate for the entire water column is supported by a "new" productivity of 1.53 mol C m^{-2} y^{-1}. This is 14% of the average primary productivity we measured in the open ocean ($\bar{x}$ = 9.82 mol C m^{-2} y^{-1}; see below). This suggests that recent ^{14}C total productivity measurements do accurately represent those occurring *in situ* provided that care is taken to prevent heavy metal contamination (FITZWATER *et al.*, 1982).

Oxygen consumption in the slow sinking POC pool

Thus far we have dealt with these OURs from one point of view: that for large biogenic particles produced in the local euphotic zone that are decomposing and consuming oxygen as they rapidly sink away from the surface. However, work with radionuclides (KRISHNASWAMI *et al.*, 1976, 1981; BACON and ANDERSON, 1982; BACON, 1984; BACON *et al.*, 1985) indicates that two size classes for particles must be considered; i.e. not only the large rapidly sinking particles mentioned above, but also the fine particle size class that sinks slowly (<1 m d^{-1}) if at all (e.g. see Fig. 6 in BACON *et al.*, 1985). Thinking about the slow or non-sinking pool led to the development of the following hypothesis. Deep winter–spring mixing in the North Atlantic introduces nutrients into surface waters. A phytoplankton bloom develops; however, little of this material is removed by the zooplankton since temperature-dependent grazing rates are low and/or because the lag time for the zooplankton to reproduce and take advantage of this food source is too long.

The nutrients become depleted, phytoplankton growth stops and the population becomes senescent. Little of this material is grazed away since zooplankton do not feed on senescent phytoplankton (e.g. MARTIN, 1966). The waters sink away from the surface with the entrained slow or non-sinking phytoplankton. As a result, POC levels are high, but dissolved oxygen concentrations are still near saturation and the $^3H/^3He$ clock is zeroed.

These waters flow along isopycnals in the Sargasso Sea and as the entrained POC decomposes, oxygen is consumed. This oxygen consumption together with that from the fast sinking particles exiting the overlying local surface waters results in total oxygen consumption of 5–6 mol O_2 m^{-2} y^{-1}. The majority of this oxygen consumption results from the lateral transport of the non-sinking POC originally produced at the surface. This explains why the oxygen anomaly has decreasing amplitude and increasing time lag with depth, since the outcrop distances for the various isopycnals along which the entrained POC is transported, increase with depth (see Fig. 7 in JENKINS and GOLDMAN, 1985). It would also explain why JENKINS' (1982) OURs often do not decrease with depth (e.g. ~0.5 ml l^{-1} y^{-1} at 100 and 300 m; 0.2 ml l^{-1} y^{-1} at 200, 300 and 400 m, etc.; see Fig. 2 in JENKINS, 1982). This hypothesis does not conflict with the seasonal signal observed by DEUSER (1986) at 3200 m since this involves the rapidly sinking large particle pool. Elevated subsurface (100–500 m) POC levels similar to those in the North Atlantic would not be found in the North Pacific since isopycnals do not outcrop because of low salinities (e.g. REID, 1965).

If this hypothesis is correct, extraordinary amounts of new production (~4 mol C m^{-2} y^{-1}) would not have to occur in the oligotrophic waters of the Sargasso Sea and extraordinary amounts of fixed N (~0.6 mol N m^{-2} y^{-1}) would not have to be introduced into the photic zone to fuel this new production. The majority of the oxygen consumption would result from the decomposition of organic matter originally produced near the sea surface at normal productivity rates when nutrients were introduced into the surface waters via normal winter–spring mixing processes.

This hypothesis should be easy to test with seasonal observations of POC levels in the upper 1000 m at the Panulirus Station or at worst by comparison of spring and autumn POC sections from the Panulirus Station to appropriate isopycnal outcrop areas. This work together with AOU–$^3H/^3He$ aging, near-surface as well as deep particle trap studies along with other independent means of assessing new production and flux should lead to the resolution of the discrepancies that currently exist. Or paraphasing JENKINS (1984): resolving the conflicting results from different approaches to the same problem may give clues to actual biological and physical processes occurring in the ocean.

The global CO_2 cycle

The data presented herein are also of interest in relation to the global CO_2 cycle. SUNDQUIST (1985) recently summarized several global reservoirs and fluxes; our work is related to his estimates for global marine primary productivity and particulate organic C fluxes from the surface to the deep ocean.

In Table 1, we present primary productivity rates measured at the various VERTEX stations. Although this is a small data base, all of these measurements were made using methods in which heavy metal contamination was avoided (FITZWATER *et al.*, 1982). Because of the paucity of "clean" primary productivity rate data, we believe it worthwhile to make some initial global estimates from them. Although we do not have

enough measurements to reproduce the five productivity regimes reported by EPPLEY and PETERSON (1979), we can at least give some estimates for the three ocean provinces used by RYTHER (1969).

Our mean productivity rates for the open ocean, coastal zone and upwelling provinces (130, 250, and 420 g m^{-2} y^{-1}; Table 6) are higher by factors of 2.6, 2.5 and 1.4 than those summarized by Ryther (50, 100, 300 g m^{-2} y^{-1}) in 1969. This is not surprising since it is generally acknowledged that primary productivities reported before 1980 were too low. With these values we obtain a global productivity of ~50 Gt y^{-1}; 80% of this production occurs in the open ocean and 20% in coastal areas. This total is generally two times higher than the global estimates summarized in SUNDQUIST's (1985) Table 6, as well as those estimated by EPPLEY and PETERSON (1979).

The productivities and the C fluxes measured in this study can be used for the development of models that relate surface primary productivity to the particulate C flux at various depths (e.g. SUESS, 1980; EPPLEY and PETERSON, 1979; BETZER *et al.*, 1984). This is being done elsewhere for the VERTEX data set (PACE *et al.*, in preparation) and will not be repeated here. Nevertheless, our total global 100 m flux (7.4 Gt y^{-1}; see Table 6) is the same as that of Suess (7.3 Gt y^{-1}; see Table 7 in SUNDQUIST, 1985), and somewhat higher than those of BETZER *et al.* (1984; 2.4 Gt y^{-1}) and EPPLEY and PETERSON (1979; 3.5–4.7 Gt y^{-1}) as well as most of those based on a variety of steady-state ocean models (see Table 7 in SUNDQUIST, 1985).

Our OOC estimates indicate that ~6 Gt C are removed from the surface of the open ocean each year via the sinking of POC (Table 7). However, much of this C is regenerated before the particles sink very far into the ocean's interior. The depth derivative of our OOC flux function gives regeneration rates of 13.1 $(z/100)^{-1.858}$ mmol C m^{-3} y^{-1}. As a result of this rapid regeneration, 50% of the C removed from the surface is regenerated at depths of <300 m; 75% is regenerated by 500 m and 90% by 1500 m (see Table 7). This rapid regeneration in near-surface waters implies that the majority of the CO_2 removed from the upper 100 m will be available for exchange with the atmosphere on time scales of tens of years or less.

However, the values presented above are comparatively low if the new productivity value of JENKIN'S (1984; ~4 mol C m^{-2} y^{-1}) turns out to be more representative of the open ocean province. With the rates of JENKINS (1982) and JENKINS and GOLDMAN (1985),

Table 6. Estimates of total global primary productivity and global new productivity using data from Table 1 for the oceanic provinces originally used by RYTHER (1969)

Province	% of ocean	Area (10^{12} m^2)	Mean prod. (g m^{-2} y^{-1})	Global total prod. (Gt y^{-1})	New* prod. (g m^{-2} y^{-1})	Global new prod. (Gt y^{-1})	% of prim prod.	% of new prod.	f
Open ocean	90	326	130†	42	18‖	5.9	82	80	0.14
Coastal zone	9.9	36	250‡	9.0	42‡	1.5	18	20	0.17
Upwelling area	0.1	0.36	420§	0.15	85§	0.03	0.3	0.4	0.20
Total		362		51		7.4			

* New productivity = C flux at 100 m.
† Mean of Stas 2, 3, 4, 5 and III (from KNAUER *et al.*, 1984a).
‡ California coast non-upwelling (KNAUER and MARTIN, 1981).
§ California coast upwelling (Sta. 1).
‖ Open ocean composite from Table 1.

Table 7. Global open ocean organic C fluxes based on OOC flux data and open ocean area from RYTHER (1969) *at various depths in the water column. Cumulative percent regeneration is also shown*

Depth (m)	Organic C flux (Gt y^{-1})	Cumulative % regeneration
100	5.9	0
150	4.2	30
200	3.3	45
300	2.3	62
400	1.8	70
500	1.5	75
600	1.3	78
800	1.0	83
1000	0.83	86
1500	0.59	90
2000	0.46	92
3000	0.32	95
4000	0.25	96
5000	0.21	96

an equivalent of 15 Gt C would be removed from the upper 100 m of the global ocean each year. Obviously, the resolution of these differences is very important when attempting to understand the role of the oceanic biosphere in the global C cycle.

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