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VERTICAL ADVECTION — DIFFUSION RATES IN THE OCEANIC THERMOCLINE DETERMINED FROM ^{14}C DISTRIBUTIONS*

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ABSTRACT. The characteristics of a one-dimensional vertical advection-diffusion ocean mixing model were examined using temperature, salinity, and bomb ^{14}C measurements made during the GEOSECS program. Vertical advection (W) and eddy diffusion (K) rates for the main oceanic thermocline and CO_2 gas exchange rates (E) were determined from the depth distributions of salinity and bomb produced ^{14}C measured in the upper 1000m of the Atlantic and Pacific Oceans. In the Atlantic, the results suggest that vertical diffusion rates are lower in the equatorial region ($K = 0.6cm^2sec^{-1}$) than in the temperate region ($K = 1.6cm^2sec^{-1}$). Upwelling rates were calculated for stations located between about 30°N and 30°S and average 10m yr^{-1}, corresponding to an upward transport of about 10 Sverdrups. Model calculations of the gas exchange rate of CO_2 indicate a 2 to 3-fold decrease between temperate latitudes and the equatorial latitudes of the Atlantic. For many of the Pacific GEOSECS stations, the Δ^{14}C depth distribution is distinctly different than in the Atlantic, and cannot be used to calculate unique values of K and W that explain both the salinity and ^{14}C depth distributions.

INTRODUCTION

The design of a transport model for the upper 1000m of the ocean is important because such models are an integral part of global exchange models such as the one formulated by Oeschger, Siegenthaler, and Gugelmann (1975) to describe terrestrial CO_2 exchange. However, oceanographers have not resolved the question of the relative importance of vertical and horizontal transport within the thermocline of the ocean. Iselin (1939) first suggested that the characteristics of the North Atlantic thermocline were obtained where isotherms and isohalines reached the sea surface. Sverdrup, Johnson, and Fleming (1942) expanded on this idea noting the similarity of the temperature (T) and salinity (S) properties of Atlantic surface water and thermocline water. Stommel (1958) presented a different explanation of the thermocline formation, using a balance of upward vertical advection and downward vertical diffusion of heat. Theoretical models of thermocline formation, as reviewed by Veronis (1969), include those that require "large" vertical diffusion rates ($K > 1cm^2sec^{-1}$) and those in which horizontal transport processes dominate, thus requiring K values no larger than $0.1cm^2sec^{-1}$. As Worthington (1976) points out, it is difficult to obtain the physical data necessary

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to validate a model in which horizontal transport dominates because of the small spatial extent of the isopycnal surface exposures.

The oceanic distribution of transient tracers may help determine whether horizontal or vertical mixing processes dominate transport in the thermocline. Nuclear bomb test products, in particular, radiocarbon (^{14}C) and tritium (^{3}H), measured during the GEOSECS program (Östlund and Stuiver, 1980; Stuiver and Östlund, 1980) provide a quasi-synoptic look at the spatial distribution of transient oceanographic properties. The distribution of tritium in the North Atlantic and North Pacific show subsurface maxima which indicate the importance of horizontal (isopycnal) transport (Rooth and Östlund, 1972; Michel and Suess, 1975). In contrast to these results, Broecker, Peng, and Stuiver (1978) explain the major features of the distribution of bomb ^{14}C as the result of vertical advection and diffusion. Thus, even with the present knowledge of transient tracer distributions, causal transport mechanisms are still debated.

The purpose of this investigation is to determine if a one-dimensional (1-D) vertical advection-diffusion model can describe both steady state and transient properties of the surface mixed layer and thermocline. The model used for this purpose makes two assumptions: 1) bomb ^{14}C, T and S properties are dependent only on vertical transport parameterized as vertical advection (W) and vertical eddy diffusion (K), and 2) the mixing processes that determine the bomb ^{14}C depth distribution also determine the T, S depth distribution. These assumptions are questionable. As mentioned above, horizontal (isopycnal) advective and diffusive mixing must be responsible for subsurface maxima in bomb tracers. In addition, a specific physical mixing process may influence only particular properties of seawater (*ie,* temperature (T), salinity (S), ^{14}C, ^{3}H, etc). For example, outcropping of isopycnals at high latitudes during the winter can result in salinity and temperature changes which are then transported along isopycnals to the south. Because of the longer response time required to change the isotopic composition of dissolved inorganic carbon, outcropping events may not be a significant source for transport of bomb ^{14}C. Thus, ^{14}C versus Z profiles may be less affected by horizontal transport than S versus Z profiles. In this situation, the salinity and ^{14}C depth distributions could not be described by unique values of K and W. Unfortunately, quantifying horizontal transport of salt, heat, and $^{14}CO_2$ is difficult.

In this investigation we will use a vertical transport model (with its inherent assumptions) to describe the surface-mixed layer and thermocline. The model-calculated mixing parameters will be examined to determine if these calculations agree with independent observations (*ie,* compare the model calculated surface layer bomb ^{14}C time history with the bomb ^{14}C record measured in corals). These comparisons test the validity of a vertical transport model to describe both transient and steady-state properties.

PROCEDURES

The equation describing the steady-state vertical distribution of conservative elements in terms of vertical advection and diffusion rates (Craig, 1969) is listed below:

$$C_z = \frac{C_m \cdot \exp\left(\frac{-(Z_m - Z)}{2Z^*}\right) \cdot \sinh\left(\frac{Z}{2Z^*}\right) + C_o \cdot \exp\left(\frac{Z}{2Z^*}\right) \cdot \sinh\left(\frac{(Z_m - Z)}{2Z^*}\right)}{\sinh\left(\frac{Z_m}{2Z^*}\right)} \tag{1}$$

where C_m, C_o, and Z_m are boundary conditions of concentration and depth and $Z^* = K/W$.

From equation (1) it can be derived that the rate of change of one conservative element with respect to another (*ie,* dT/dS) will be constant

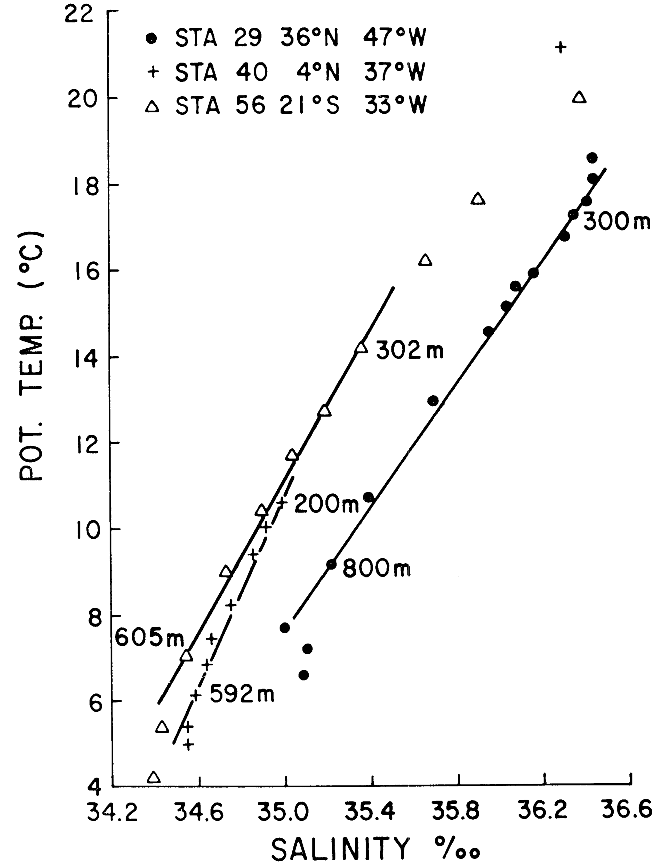

Fig 1. Temperature–salinity properties for three western Atlantic stations.

over the depth interval defined in equation 1. Thus, a linear section of the T–S plot implies that the vertical distribution of T versus Z and S versus Z can be described in terms of K and W, assuming a steady-state condition.

The GEOSECS stations at which ^{14}C samples were collected were examined for linear regions in the T–S plots. The T–S properties of three Atlantic stations are shown in figure 1. For those stations showing a linear T–S relationship (usually between 250m and 800m in temperate latitudes and 150m to 500m in equatorial latitudes) the value of K/W that results in the best fit to the observed S versus Z profile was determined by an iterative procedure. After the value of K/W is calculated, the values of K and W are determined using the bomb produced $^{14}CO_2$ versus Z distribution. The equation describing the time and depth dependence of bomb produced (excess) $^{14}CO_2$ on K and W is:

$$C_{z,t} = \frac{C_o}{2}\left[\mathrm{Erfc}\left(\frac{Z}{2\sqrt{Kt}} - \frac{w}{2}\sqrt{\frac{t}{K}}\right) + e^{wZ/K}\cdot\mathrm{Erfc}\left(\frac{Z}{2\sqrt{KT}} + \frac{w}{2}\sqrt{\frac{t}{K}}\right)\right] \qquad (2)$$

where C_0 is the mixed layer concentration of excess $^{14}CO_2$. For the calculation of K and W, C_0 was not kept constant with time. Instead, the mixed layer concentration of excess $^{14}CO_2$ (C_0) at each station is taken as zero until 1960 at which time it presumably increased linearly until 1969 when it remained constant at the specific value measured at that station during the GEOSECS program (1972-1974). The mathematical procedure to step the source function concentration (C_0) with time is discussed in Carslaw and Jaegar (1959), where the numerical approximation to erfc used to evaluate equation (2) is also found. Values of K and W are calculated for each depth using equation 2 and the excess $^{14}CO_2$ versus Z profile, which takes into account not only the measured changes in $\Delta^{14}C$ with depth but also the depth change in the concentration of dissolved inorganic carbon. Average K and W values for each station were calculated from these results.

The uncertainty in determining the "excess or bomb" ^{14}C measured during GEOSECS results from estimating the pre-bomb, steady-state ^{14}C versus Z distribution. The excess ^{14}C represents the difference between the present ^{14}C concentration and the ^{14}C concentration that existed before nuclear testing began. Because there are only a few direct measurements of pre-bomb ^{14}C values in the ocean (Broecker and others, 1960; Linick, 1978), the steady-state pre-bomb depth distribution of ^{14}C can only be estimated. A plot of pre-bomb $\Delta^{14}C$ values versus density (σ_t) measured on subsurface water (50 to 1200m) between 25N and 20S in the Atlantic (Broecker and others, 1960) can be expressed as a linear relationship (fig 2). To estimate the pre-bomb ^{14}C versus Z distribution for each station, a linear relationship between $\Delta^{14}C$ and σ_t is assumed to exist. By fixing the pre-bomb $\Delta^{14}C$ values for both the surface-mixed layer and a depth not yet influenced by bomb ^{14}C at the time of sample collection, we can then estimate the pre-bomb $\Delta^{14}C$ versus σ_t and, therefore, the pre-bomb $\Delta^{14}C$ versus Z profile at each station. The pre-bomb surface $\Delta^{14}C$ values were assumed to be –65‰ for the equatorial region, and –53‰ for the temperate region (Broecker and others, 1960). The maximum depth to which bomb-produced ^{14}C had penetrated was determined by the deepest penetration depth of tritium at that station (Östlund, Dorsey, and Brescher, 1976; Östlund and others, 1978). This depth increases from about 500m in the equatorial region to about 800m in the temperate latitudes except for Stations 27 and 29, where bomb tritium has penetrated to about 2500m. Integrated amounts of excess $^{14}CO_2$ at each station were calculated after estimation of the pre-bomb distribution.

Once K and W were determined for a station, the carbon dioxide gas exchange rate (E) was calculated using a vertical 1-D oceanographic mixing model. The model describes the ocean as a homogeneous surface-mixed layer of variable depth, a thermocline divided into 40 25m-deep boxes and a deep sea divided into 5 560m-deep boxes. This is similar to the model design of Oeschger, Siegenthaler, and Gugelmann (1975). Transfer of $^{14}CO_2$ between a well-mixed atmosphere and oceanic-mixed layer was parameterized in terms of E in moles of $CO_2m^{-2}yr^{-1}$. The source function for bomb ^{14}C was the measured atmospheric record of bomb produced ^{14}C (Nydal, Lövseth, and Gulliksen, 1979). With the

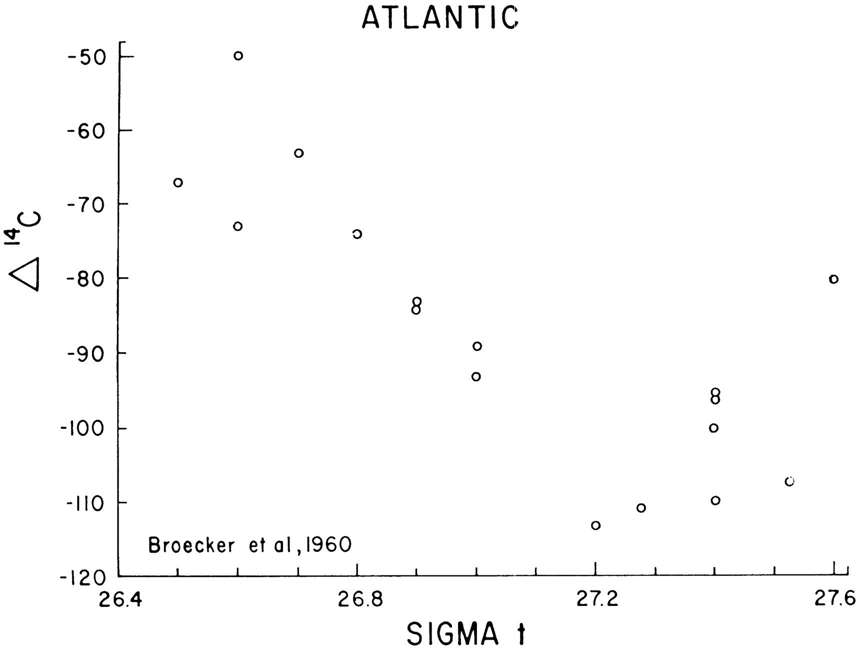

Fig 2. Pre-bomb $\Delta^{14}C$ values and density (σ_t) of samples collected at depths of 100m to 1000m in the Atlantic Ocean between 25°N and 20°S (Broecker and others, 1960).

previously determined values of K and W parameterizing $^{14}CO_2$ transfer between the mixed layer and thermocline, a unique value of E required to obtain the measured surface-mixed layer concentration of $C^{14}O_2$ at a particular GEOSECS station was determined using the time-dependent numerical model. For the model calculations of E, a volume of upwelling water, parameterized by a negative value of W, is assumed to horizontally advect out of the mixed layer with the bomb ^{14}C concentration of the mixed layer. A volume of downwelling water (positive W) is assumed to enter the mixed layer with a concentration of bomb ^{14}C equal to that of the mixed layer. A $\Delta^{14}C$ versus Z profile is calculated by the model for specific values of K, W, and E, as is a time history of bomb ^{14}C concentrations in the mixed layer (as discussed below).

RESULTS

The K/W values that result in the best fit to the salinity versus depth distributions for the GEOSECS stations in the Atlantic and Pacific are plotted by location in figures 3 and 4, respectively. The number in parentheses next to the K/W value represents a chi-square-

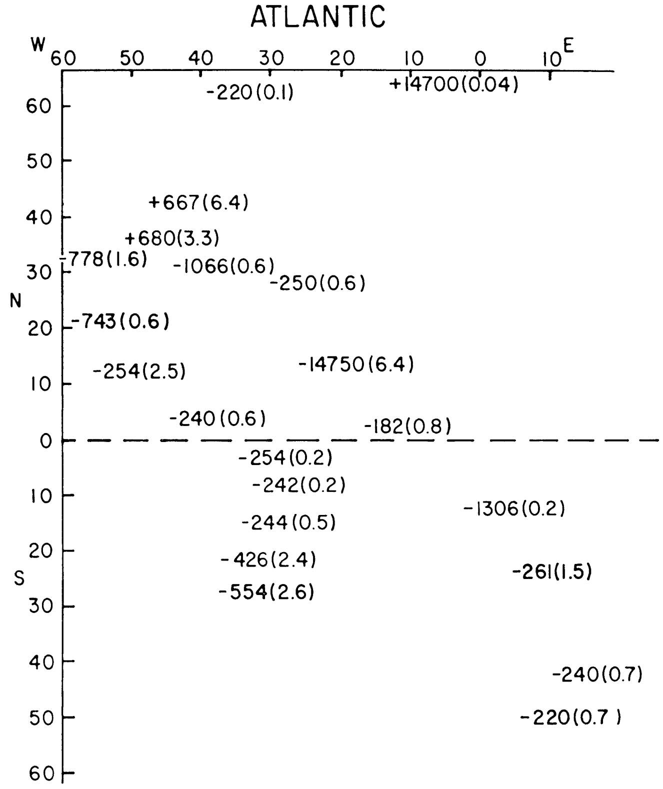

Fig 3. Values of K/W calculated from salinity vs depth distributions for several GEOSECS stations in the Atlantic Ocean.

fitting parameter, the smaller the number the better the fit of salinity versus depth calculated using equation 1 to the observed salinity distribution. Negative K/W values correspond to upwelling regions. In the equatorial West Atlantic (15°N-15°S) the K/W values are consistently about −245. This K/W value has also been determined for the eastern South Atlantic. Moving north and south from the equator the values of K/W become more negative which implies a relatively greater vertical eddy diffusion rate or smaller upwelling velocity compared to the equatorial region. North Atlantic stations 27 and 29, at 36°N and 42°N, show positive K/W values which imply a region of downwelling. The large K/W values (*ie*, +14,700, −14,750) are of little significance because the salinity versus depth profile is nearly linear. Also, the larger the K/W values, the less sensitive the K/W calculation is to the direction of advection because the eddy-mixing parameter (K) essentially produces the shape of the S versus Z profile. For the stations examined in the Pacific, linear T–S regions over the thermocline depth range only occurred in the southern ocean. Also, subsurface tritium maxima (at about 200 to 500m) were present in several North Pacific GEOSECS stations; these features cannot be explained by only vertical transport processes. The K/W values calculated for the South Pacific generally show a similar pattern as in the Atlantic, lower negative values of K/W in the

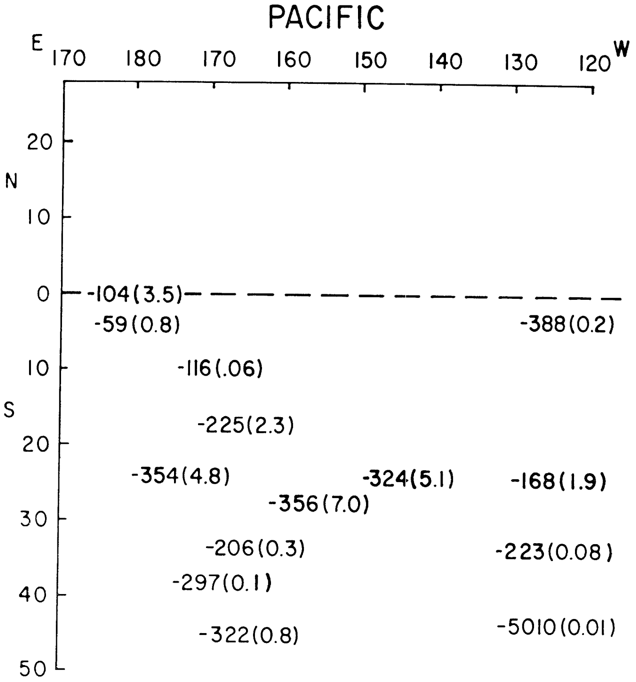

Fig 4. Values of K/W calculated from salinity vs depth distributions for several GEOSECS stations in the Pacific Ocean.

equatorial region, and larger negative values to the south. No positive K/W values were determined for the Pacific stations examined.

The values of K and W calculated using equation 2 and the $^{14}CO_2$ versus Z profiles for the stations in the Altantic are presented in table 1. These K and W values imply that the more negative K/W values determined by the salinity profiles outside the equatorial region are the result of higher vertical eddy diffusion rates rather than lower upwelling velocities. Within the equatorial region the average $K = 2274 m^2 yr^{-1}$ ($0.6 cm^2 sec^{-1}$) and $W = -9.2 m\ yr^{-1}$. Between 15° and 33° in both the North and South Atlantic, the average $K = 6125 m^2 yr^{-1}$ ($1.5 cm^2 sec^{-1}$) and $W = 9.4 m\ yr^{-1}$. The three stations in the East

TABLE 1

The rates of vertical diffusion (K), vertical advection (W), and CO_2 gas exchange (E) calculated for several GEOSECS stations in the Atlantic Ocean. The integrated amount of excess $^{14}CO_2$ at each station is also presented.

Station	Lat.Long.	$\Sigma C^{14}O_2$ $x10^{10}$ moles m^{-2}	K/W m	K $m^2 yr^{-1}$	W $m\ yr^{-1}$	E moles $m^{-2} yr^{-1}$
		W. Atlantic				
27	42N 42W	2.6	667	5931	8.9	21
29	36N 47W	2.8	680	6318	9.9	24
120	33N 57W	2.4	-778	6770	-8.7	31
117	31N 39W	2.3	-1066	9643	-9.1	31
33	21N 54W	1.7	-743	6320	-8.5	28
37	12N 51W	1.1	-254	3394	-13.4	15
40	4N 39W	0.8	-240	1158	-4.8	9
48	4S 29W	0.6	-254	776	-3.0	7
49	8S 28W	0.8	-242	2354	-9.7	9
54	15S 30W	1.1	-244	3692	-15.1	20
56	21S 33W	1.6	-426	5623	-13.2	25
58	27S 37W	1.7	-554	8245	-14.9	27
		E. Atlantic				
103	24S 8E	1.2	-261	4783	-18.3	16
107	12S 2E	0.4	-1306	917	-0.7	4
111	2N 14W	0.5	-182	2464	-13.5	7

TABLE 2

The rates of vertical diffusion (K), vertical advection (W), and CO_2 gas exchange (E) calculated for several GEOSECS stations in the Pacific Ocean. The integrated amount of excess $^{14}CO_2$ at each station is also presented.

Station	Lat.Long.	$\Sigma C^{14}O_2$ $x10^{10}$ moles m^{-2}	K/W m	K $m^2 yr^{-1}$	W $m\ yr^{-1}$	E moles m^{-2} yr^{-1}
		Pacific				
246	0° 179°E	0.77	-104			
251	4°S179°W	0.80	-59			
257	10°S170°W	1.1	-116			
263	17°S167°W	1.3	-225			
269	24°S174°W	1.8	-354			
310	27°S158°W	1.7	-356			
306	33°S167°W	1.3	-206			
303	38°S170°W	1.1	-297			
296	45°S167°W	1.1	-322			
322	43°S129°W	1.6	-5010	3714	-0.8	16
320	33°S128°W	1.3	-223	2606	-11.7	28
317	24°S128°W	1.7	-168	2770	-16.5	31
324	23°S146°W	1.5	-324	3503	-10.8	25
331	4°S125°W	0.24	-388	284	-0.7	5

Atlantic also show smaller K values in the equatorial region than in the southern temperate ocean. For the Atlantic, an average K value of $1.1 cm^2 sec^{-1}$ is calculated and the average upwelling velocity between 30°N and 30°S is $10 m\ yr^{-1}$, which corresponds to a volume transport of about 10 Sverdrups. The standard deviation/mean value for the K and W values calculated for each Atlantic station, presented in table 1, range between 10 percent and 100 percent with a mean of 45 percent.

Fourteen Pacific stations were examined. For only five of these stations, K and W values could be calculated that would explain both the salinity and $^{14}CO_2$ depth profiles. These five stations are located east of 150°W. The resulting K and W values are listed in table 2. Some insight into the apparent difference between the Pacific and Atlantic ^{14}C distribution can be obtained by plotting the $\Delta^{14}C$ versus Z profiles at equivalent latitudes in both oceans, as seen in figures 5 and 6. Although bomb ^{14}C in the equatorial stations of both oceans only appears above 500m depth, the shapes of the depth profiles in the Pacific and Atlantic are distinctly different. In the Pacific, surface ^{14}C concentrations generally extend to greater depths than in the Atlantic. The ratio K/W was calculated from the salinity versus depth distributions, as was done for the Atlantic stations. However, it is impossible to obtain K and W values (except Station 331) which describe the S versus Z distribution and can account for the distribution of ^{14}C in the upper 500m without overestimating the maximum penetration depth of bomb ^{14}C. This problem, which is apparent in the comparison between the model fits and observed ^{14}C versus Z profiles in figure 10,

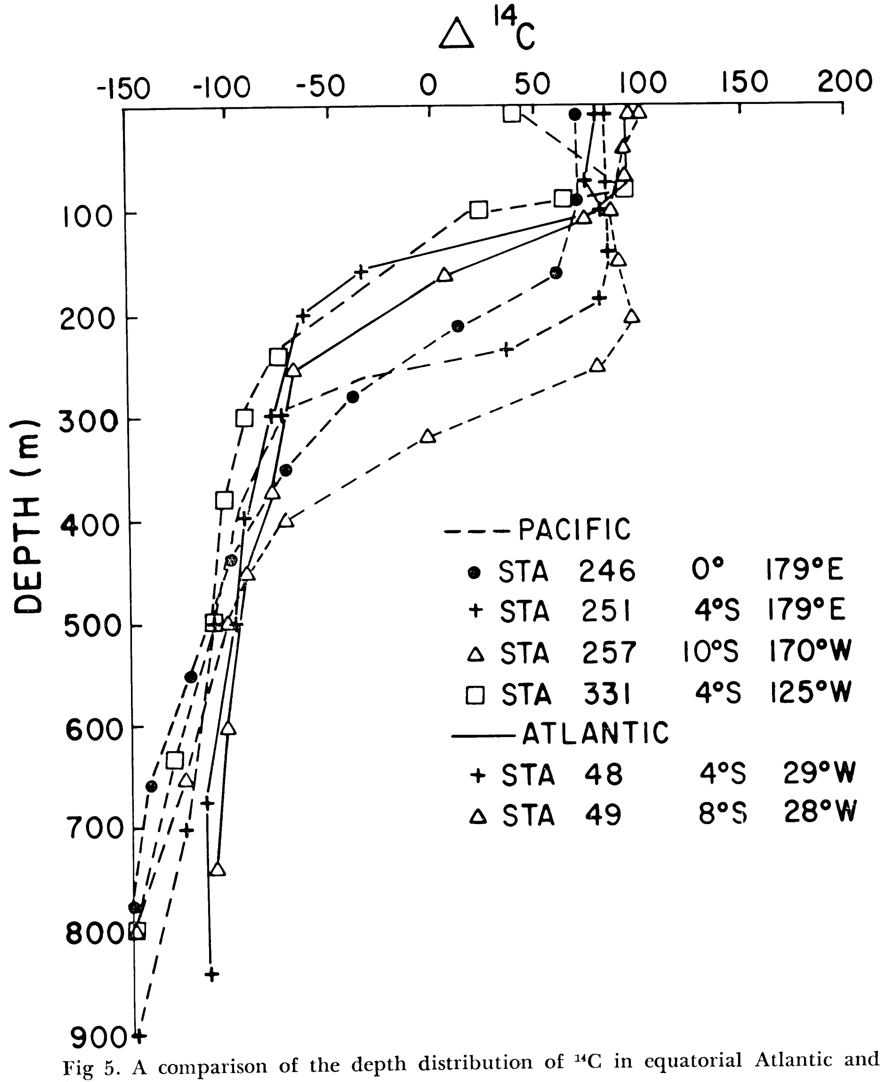

Fig 5. A comparison of the depth distribution of ^{14}C in equatorial Atlantic and Pacific Oceans.

does not occur in the Atlantic. Thus, the shape of the Pacific ^{14}C profile appears to be significantly influenced by horizontal $^{14}CO_2$ transport below the mixed layer, resulting in apparent deeper penetration of surface ^{14}C concentrations. These results suggest that a 1-D vertical transport model is inadequate to describe the $^{14}CO_2$, T, and S depth distributions in the Pacific.

Gas exchange rates of CO_2 and depth distributions of bomb ^{14}C were calculated using a numerical model of ocean-atmosphere exchange described above. The resulting values of E and the model fits to the observed $\Delta^{14}C$ versus Z distributions in the Atlantic are shown in table

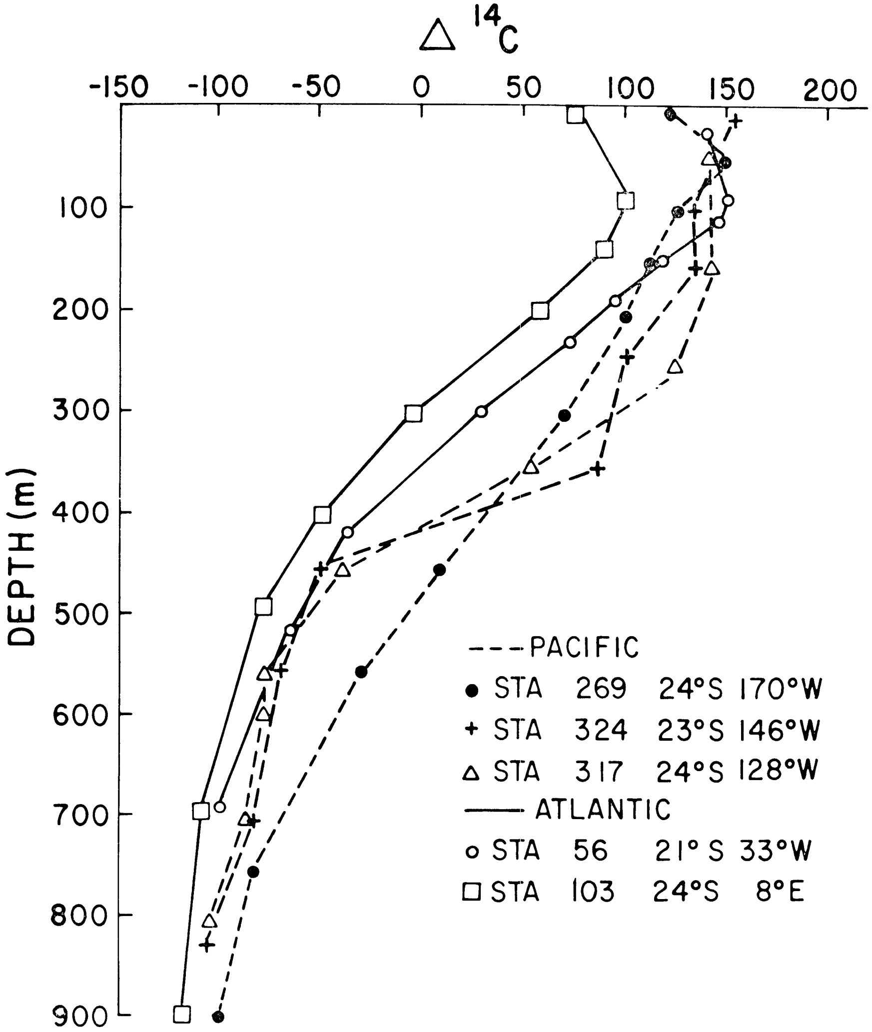

Fig 6. A comparison of the depth distribution of ^{14}C in the southern Atlantic and Pacific Oceans.

1 and figures 7, 8, and 9. The ^{14}C depth distribution for the stations in the North and South Atlantic show almost perfect symmetry about the equator (figs 7, 8). There are deeper penetration and higher mixed-layer concentrations of ^{14}C in temperate regions than in equatorial regions. Gas exchange rate calculations also show this symmetry. In the western Atlantic, the model calculated values of E range from 7 moles $m^{-2}yr^{-1}$ near the equator to 30 moles $m^{-2}yr^{-1}$ at 31°N. The average Atlantic

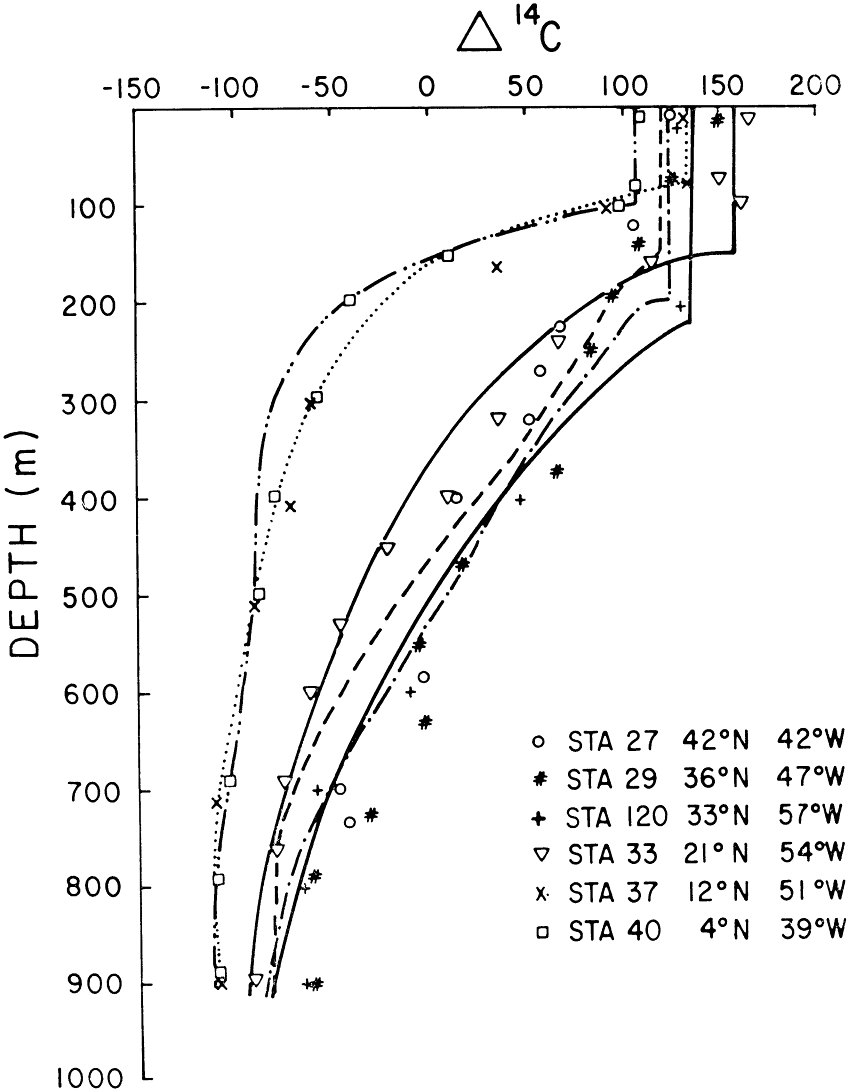

Fig 7. The measured Δ^{14}C vs Z distribution and the model prediction (using the parameters listed in table 1) for several GEOSECS stations in the northern west Atlantic Ocean.

value of E (weighted for surface area) is about 21 moles $m^{-2}yr^{-1}$. The total integrated amount of bomb ^{14}C in the Atlantic requires an average E value of 23 moles $m^{-2}yr^{-1}$ (Stuiver, in press).

For the Pacific, the model fits to the observed $\Delta^{14}C$ versus Z profiles are shown in figure 10. As mentioned above, except for the eastern equatorial stations (Station 331), the model calculations tend to overestimate the bomb ^{14}C concentrations below 500m. The model-calculated CO_2 gas exchange rates show the same pattern as in the Atlantic, higher in the temperate latitudes at an average of 25 moles $m^{-2}yr^{-1}$ and lower in the equatorial region where, for one station, E = 5 moles $m^{-2}yr^{-1}$ was calculated.

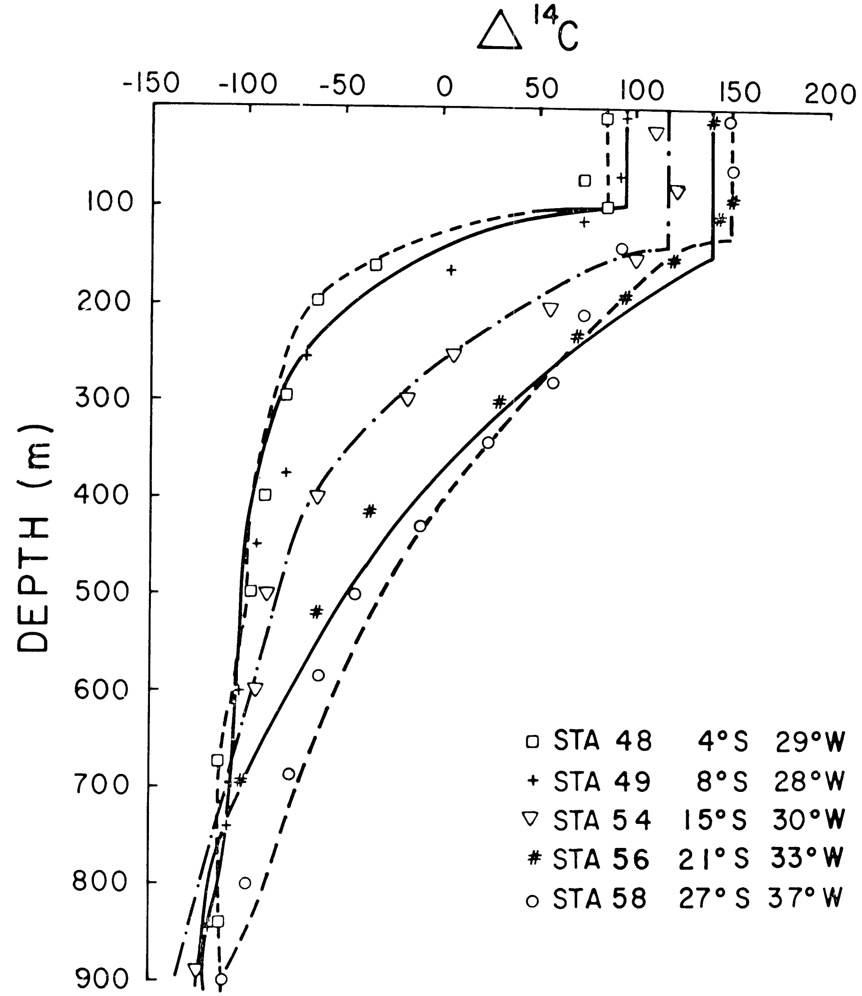

Fig 8. The measured $\Delta^{14}C$ vs Z distribution and the model predictions (using the parameters listed in table 1) for several GEOSECS stations in the southern west Atlantic Ocean.

DISCUSSION

The results indicate that a vertical transport model can explain both steady-state T–S properties and transient bomb ^{14}C distributions in the Atlantic Ocean between 42°N and 27°S. This is not the situation for the Pacific Ocean (specifically the western Pacific) where unique values of K and W could not be determined from the T–S and ^{14}C distributions for most of the stations examined. Although K, W, and E values were calculated for specific stations, only the general latitudinal trends shown by these parameters are discussed. In the Atlantic, the model calculations suggest that the shallow depth penetration and low-

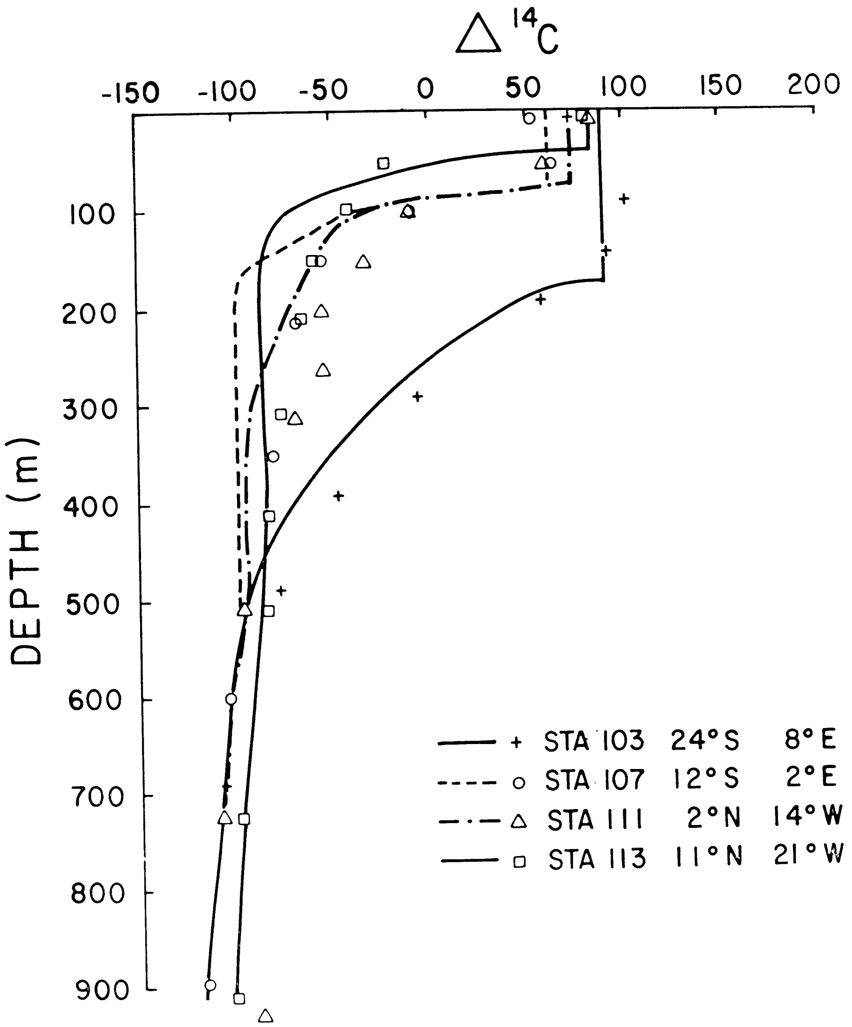

Fig 9. The measured $\Delta^{14}C$ vs Z distribution and the model predictions (using the parameters listed in table 1) for several GEOSECS stations in the eastern Atlantic Ocean.

standing crop of bomb ^{14}C in the equatorial region are the result of low vertical diffusion rates, low gas exchange rates, and upwelling of deeper ^{14}C-depleted water. In the temperate latitudes, north and south of the equator, the bomb ^{14}C distributions shows a deeper penetration and larger integrated amounts. These observations are reflected in the model calculations as higher vertical diffusion rates and gas exchange rates. Both upwelling and downwelling rates were determined for the temperate region although downwelling velocities were calculated for only two stations at 36°N and 42°N. The model calculations of W

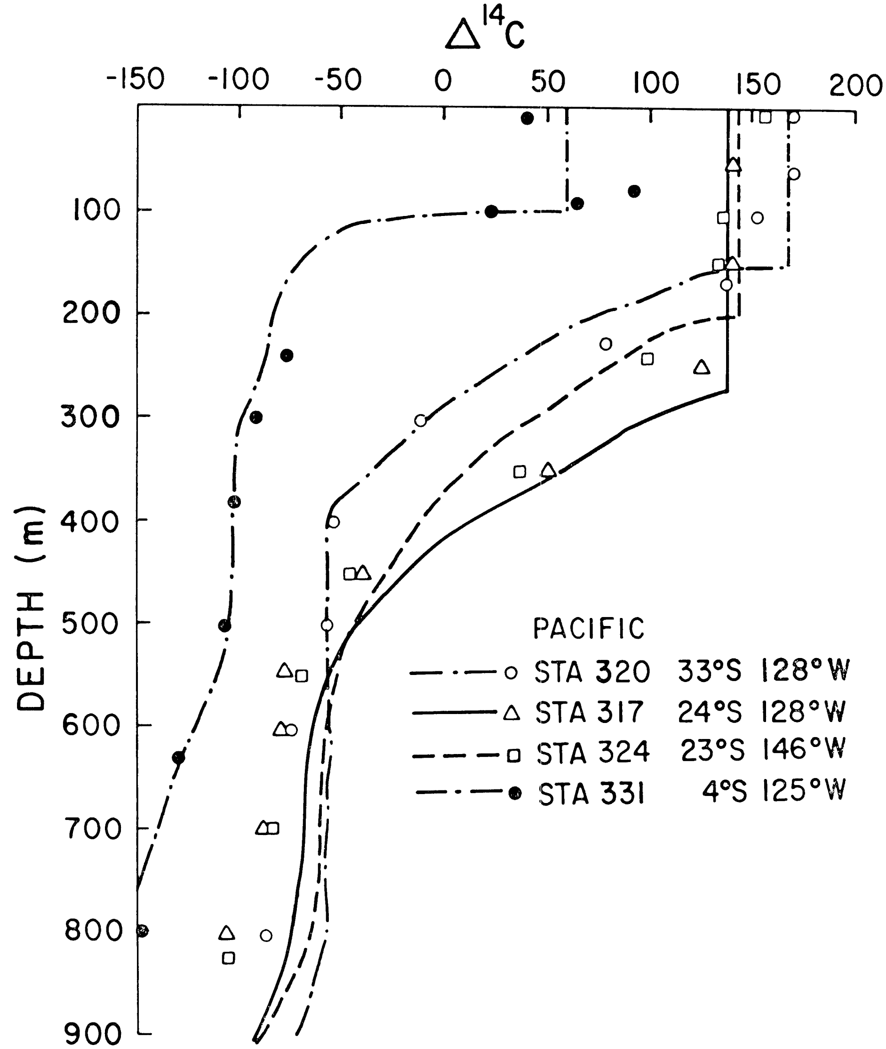

Fig 10. The measured $\Delta^{14}C$ vs Z distribution and the model predictions (using the parameters listed in table 2) for several GEOSECS stations in the southern Pacific Ocean.

become less sensitive to the direction of the vertical advective term as K increases in magnitude. This is because where $(K \cdot t)^{1/2} > W \cdot t$, t being time, most of the vertical transport results from diffusive mixing rather than advection. For example, although Station 120 and Station 27 have similar ^{14}C versus Z profiles (see fig 7), the model calculations imply that these profiles can be explained using upwelling and downwelling rates, respectively.

Once K, W, and E have been calculated and the bomb ^{14}C atmospheric input function is known, a time history of the bomb ^{14}C concentration in the surface-mixed layer can be calculated, using the ocean model discussed above. Comparing these model calculations to the observed ^{14}C time history recorded in corals provides one test of the validity of the model calculations. The measurements of Druffel and Linick (1978) on a coral collected off the coast of Florida (25°N, 81°W) are compared to the model predictions of surface ^{14}C time changes for four western Atlantic stations in figure 11. The model results of Stations 33 and 58 show the best agreement with the coral data. These stations are located at 21°N and 27°S, respectively. Their $\Delta^{14}C$ versus Z profiles are almost identical, as seen in figures 7 and 9; upwelling rates have been calculated for both. The rate of increase of mixed layer ^{14}C with time for the equatorial region (Station 49) and "downwelling" region (Station 29) is slower than observed for the coral data. Broecker, Peng,

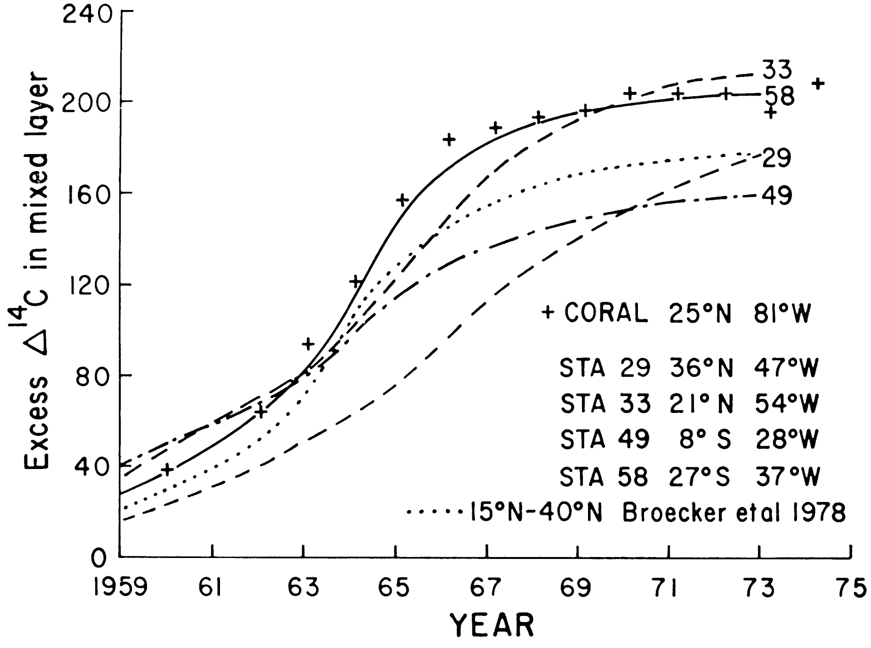

Fig 11. The model predictions of the time history of excess $\Delta^{14}C$ in the surface mixed layer of four western Atlantic GEOSECS stations and the excess $\Delta^{14}C$ vs time record measured in a coral from the Florida coast (25°N, 81°W) (Druffel and Linick, 1978).

and Stuiver (1978) estimates of this time history for the north temperate (downwelling) region is also shown in figure 11. These results suggest that the coral data is the most compatible with model predictions for stations where salinity and ^{14}C depth distributions have been described by upwelling velocities and relatively high rates of vertical diffusion and gas exchange.

The model calculations of gas exchange rates is oversimplified because horizontal transport of $^{14}CO_2$ in the surface ocean is neglected. At best, only the large-scale average values of E can be examined. In the Atlantic there is a calculated decrease from an average of 25 moles $m^{-2}yr^{-1}$ in the temperate latitudes (21°-42°N and S) to 10 moles $m^{-2}yr^{-1}$ in the equatorial region (15°N-15°S). This difference between temperate and equatorial values of E is much greater than the relative change of 25 percent indicated by the GEOSECS radon-222 results (Peng and others, 1979). However, direct comparison of gas exchange rates estimated by bomb $^{14}CO_2$ and radon is not straightforward. The radon method of estimating gas exchange rates integrates over the 6-day mean life of radon, whereas bomb ^{14}C estimates of gas exchange rates integrate over about a 10-year interval (*ie,* 1963 to 1973). Thus, physical processes with fluctuations on the order of a week or less (*ie,* wind speeds, currents) affect the radon concentration in the mixed layer and, thus, the gas exchange rate calculation. These short-term fluctuations will be averaged out over the time interval of bomb ^{14}C input into the oceans. Some of the discrepancy between the radon and bomb ^{14}C derived calculations of E reflect these different integration times. In addition, the radon results can underestimate the difference between temperate and equatorial regions because of the seasonality of the sample collection; most of the samples were collected during the summer. Equatorial meteorologic conditions should be less variable than at temperate latitudes, so that summer months would better represent equatorial rather than temperate gas exchange rates. The short "memory" of radon can miss the effects that episodic storms have on the magnitude of E. Unless enough radon profiles are collected to provide a statistically significant sampling of these storm events, the resulting value of E for temperate latitudes will be underestimated.

The latitudinal gradients in mixed layer bomb ^{14}C concentrations will result in horizontal transport of surface $^{14}CO_2$. Horizontal diffusive mixing transports ^{14}C from regions of high to low concentrations whereas horizontal advective transport will occur in the direction of the prevailing surface currents. If large scale diffusive transport of ^{14}C toward the equator is of the same order as the advective transport away from the equatorial region (balancing the volume of upwelling water), this will result in model underestimates of E for the temperate regions and overestimates of E in the relatively ^{14}C depleted equatorial region.

The model calculations of the temperate latitude mixed layer time history of bomb ^{14}C are compatible with the record of bomb ^{14}C concentrations measured in corals. This agreement supports the K, W, and

E values determined in this investigation. However, the model-calculated CO_2 gas exchange rates do not agree with the gas exchange rates estimated from GEOSECS radon measurements (Peng and others, 1979). If CO_2 exchange rates do not vary strongly with latitude, this would suggest that the T–S properties of the main thermocline of the equatorial Atlantic cannot be interpreted as simply derived from vertical mixing processes. If equatorial CO_2 gas exchange rates are greater than our calculations indicate, then higher upwelling rates would be required to balance the ^{14}C distribution, as calculated by Broecker, Peng, and Stuiver, (1978). Coral records of bomb ^{14}C in the equatorial region could resolve this discrepancy.

Any oceanographic model should be subjected to all tests that can be formulated from available chemical and physical data to determine if it is an adequate representation of oceanic transport processes. This investigation has examined the response of a 1-D vertical transport model which describes T–S properties and bomb ^{14}C distributions in the main thermocline of the oceans. Because of the assumptions inherent in a 1-D ocean model, the extrapolation of the transfer rates determined in this study to other dissolved species in the ocean may not be justified and should be made with caution.

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