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DECADE TIME SCALE VARIABILITY OF VENTILATION IN THE NORTH ATLANTIC: HIGH-PRECISION MEASUREMENTS OF BOMB RADIOCARBON IN BANDED CORALS

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Abstract. The first high-precision radiocarbon measurements for the upper ocean are presented for banded corals from two sites in the North Atlantic Ocean. The striking dissimilarities between the post-1950 records at Bermuda in the Sargasso Sea and the Florida Straits in the Gulf Stream illustrate the different mixing processes in the upper ocean at each site. Convective overturn associated with 18° degree water formation during late winter in the northern Sargasso Sea facilitates storage of considerable quantities of bomb radiocarbon at depth, which accounts for the damping of the $\Delta^{14}C$ signal at Bermuda during the 1960's. A multibox isopycnal mixing model is used to estimate the ventilation rate of the upper 700 m of the water column in the Sargasso Sea from 1950 to 1983. An inverse model is used; that is, the water mass renewal rate was calculated for the post-bomb period in order to satisfy the bomb radiocarbon time history in the corals. Sea water radiocarbon measurements made during the GEOSECS (1972-1973) and Transient Tracers in the Ocean (1980-1981) surveys are used to constrain the subsurface radiocarbon values calculated by the model. Results show that the rate of water mass renewal in the Sargasso Sea was high during 1963-1964, decreased during the late 1960s, and remained low during most of the 1970s. The ^{14}C-derived record of water mass renewal precedes by about 4 years that derived from isopycnal salinity in the Sargasso Sea [Jenkins, 1982], illustrating that the coral ^{14}C record is controlled to a large extent by changes in ocean circulation rather than by atmospheric exchange of CO_2.

Introduction and Background

One of the few positive outcomes of the nuclear weapons testing era of the 1950s and early 1960s was the production of bomb radiocarbon and tritium, which offers geochemists the opportunity to study ocean circulation on relatively short time scales. Numerous studies have been conducted to determine mixing rates in the upper ocean (for example, see Michel and Suess [1975], Jenkins [1980], Broecker et al. [1985]) and flow rates of major ocean currents [Fine, 1985].

The majority of studies reported to date are based on synoptic observations of transient tracers taken over a short period of time. This "snapshot" approach to oceanic tracers (e.g., the Geochemical Ocean Sections (GEOSECS) and Transient Tracers in the Ocean (TTO) surveys) provides resolution on an ocean-wide scale, however, temporal variations which occur on time scales longer than the period of observations are not detected. Time histories of tracer distributions can be obtained only by regular reoccupations of the same ocean stations or by extraction of these data from unaltered integrators of the tracer signals, such as banded corals. Nonetheless, study of variability on both spatial and temporal scales is needed to define the ocean-atmosphere coupling so critical for understanding climate. This approach is also important for quantifying anthropogenic perturbations on climate such as those associated with the greenhouse gas carbon dioxide.

Until recently, little attention has been paid to variability of oceanic parameters other than those on very long (glacial-interglacial) or very short (diurnal or seasonal) time scales. Decade and century time scale variations have been reported in air temperature measurements taken on land [Jones et al., 1982] and at sea [Folland et al., 1984] during the past 100 years. Venrick et al. [1987] have linked a significant increase in total chlorophyll a during the past 30 years in the central North Pacific water column to a long-term increase in winter winds, coincident with a decrease in sea surface temperature. Brewer et al. [1983] noticed a significant freshening of the North Atlantic from 1972 to 1981, which they attributed to short-term climatic forcing. Decade time scale variability of water mass renewal in the upper 600 m of the Sargasso Sea was reported by Jenkins [1982] based on a 27-year record of salinity and oxygen observed at the Panulirus station (station "S") southeast of Bermuda.

The ocean is not impervious to changes in primary production, circulation, and water chemistry on decade time scales. Efforts to understand and apply these long-term variations of the ocean-atmosphere system to geochemical models, especially in connection with global climate, must accompany studies of real-time transient tracer distributions.

Demonstration of decade time scale variability in water mass renewal rate in the Sargasso Sea is presented here using high-precision bomb radiocarbon records obtained from annually banded corals. Owing to the gaseous nature of the bomb radiocarbon transient in the atmosphere, its input to and hence distribution in the ocean differ from those of other tracers, for instance, salinity, oxygen, or tritium. Independent determination of the water mass renewal rate record in the Sargasso Sea is made using an isopycnal exchange model to represent mixing in the upper 700 m of the water column and is constrained using GEOSECS and TTO depth profiles taken during 1973 and 1980.

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Buddemeier et al. [1974] were the first to measure radiocarbon in coral skeletons as a method for determining the $^{14}C/^{12}C$ ratio in the dissolved inorganic carbon (DIC) in past ocean waters. Subsequently, numerous investigators have exploited this technique to study the input of fossil fuel CO_2 and bomb radiocarbon into the oceans [Nozaki et al., 1978; Druffel and Linick, 1978; Druffel and Suess, 1983; Toggweiler, 1983; Konishi et al., 1981; Druffel, 1987]. I have taken this approach one step further and obtained high-precision (± 2°/₀₀) radiocarbon measurements of annual coral bands, so that small differences (>8°/₀₀) in the bomb radiocarbon time histories from different locations can be determined accurately.

Approach

Radiocarbon measurements from banded corals are assumed to represent annually averaged radiocarbon levels in surface water DIC from the area surrounding the coral reef. This assumption extends from general agreement between coral [Druffel and Linick, 1978; Druffel, 1987] and seawater results. Fairbanks and Dodge [1979] demonstrated that $^{18}O/^{16}O$ ratios in corals from three North Atlantic locations were proportional to sea surface temperature (SST), provided salinity and water composition were constant. A biweekly $^{18}O/^{16}O$ record from a 3-year section of one of the Bermuda corals is used to verify the assumption of constant aragonite accretion during the year. The reproducibility of $\Delta^{14}C$ obtained from different coral heads on the same reef is demonstrated; this takes into account mapping and cutting errors, which can be significant when the growth rate is low and the seasonal change of $\Delta^{14}C$ is large.

Bermuda is located in the middle of the Sargasso Sea and is assumed to be influenced by waters representative of the gyre as a whole [Worthington, 1976; Jenkins, 1980]. Sea level at Bermuda has not changed dramatically over the past few decades [NOAA, 1983]. Even though this midgyre location would be the least sensitive to gyre movement, the sea level data suggest that the gyre placement has not varied significantly.

It is assumed that the ocean mixes predominantly along isopycnal surfaces [Iselin, 1936; Montgomery, 1938]. Jenkins [1980] showed that tracer distributions in the Sargasso Sea are consistent with along-isopycnal exchange as the major mixing process in the main thermocline (upper 1000 m) and that vertical diffusivities are low (0.1 cm^2/s or less) in the North Atlantic subtropical gyre.

An isopycnal transport model is used here to represent mixing of water and radiocarbon in the upper waters of the Sargasso Sea. Radiocarbon records in banded corals from Bermuda are used to represent the time history of ^{14}C in a well-mixed subtropical gyre surface box. The radiocarbon time history reported here for Florida is taken to represent the lateral input from the Gulf Stream which feeds the surface layers of the Sargasso Sea. Water mass renewal rate (W_i, in reciprocal years) is defined as the exchange of water between the surface box and an isopycnal subsurface box i. Assuming that exchange of CO_2 between the surface box and the atmosphere is a function of wind speed, W_i is calculated on an annual basis from 1950 to 1983 in order to match the Bermuda record. GEOSECS and TTO radiocarbon depth profiles are used as constraints to fit the model calculated radiocarbon in the isopycnal boxes. Various sensitivity analyses are performed to test the reliability of the resultant W_i time history.

Methods

Cores from living colonies of Diploria strigosa and Diploria labyrinthiformis were collected from 11 m depth at two sites off Bermuda: (1) Sam Hall's Bay (32°21'N, 64°41'W), 0.3 km from shore off south Bermuda (samples S-a and S-b), and (2) North Rock (32°29'N, 64°46'W), 10 km north of Bermuda (N-a and N-b). North Rock is at the northern edge of a shallow (<10 m depth) reef flat that extends south to the island of Bermuda; thus waters on this reef flat may be closer to equilibrium with respect to gas exchange with the atmosphere than waters off the south coast. Coral cores were drilled from living heads (1-1.5 m diameter) using an hydraulically driven coring device, equipped with a diamond bit on the end of a 0.7 m core barrel (with 10 cm diameter). The device was powered by a gasoline engine operated aboard the R/V Culver (Bermuda Biological Station). Five small cores (TR3, 7 cm diameter) were collected using a hand-held hydraulic drill from "The Rocks" reef, off Islamorada in the Florida Keys.

The cores were flushed with fresh water for several hours to remove polyp material from the living surface. They were dried and slabbed (8-9 mm thick) along the corallite growth axes. Annual bands were mapped using X-radiography and high-contrast film. Methods for sectioning the coral were outlined by Griffin and Druffel [1985]. Approximately 10-20% of each annual band was lost owing to the thickness of the band saw blade. However, this does not affect the annual averaged $\Delta^{14}C$ value, since we estimate that the portion removed was from a period of intermediate $\Delta^{14}C$ on the scale of the seasonal signal [Broecker and Peng, 1980]. Eight seasonal bands, of approximately 3 month growth increments each, were sanded from one of the Florida cores for the years 1973 and 1974. A seawater sample was collected in March 1983 from "The Rocks" area, stripped of DIC according to Linick [1980], and measured for ^{14}C according to the methods described below.

Each sample was converted to acetylene gas via a lithium carbide intermediate [Griffin and Druffel, 1985]. Samples were counted in quartz gas proportional beta counters for five 2-day periods to obtain errors based on counting statistics of ± 2.2°/₀₀. Results are reported as $\Delta^{14}C$ in per mil (°/₀₀) according to the convention of Stuiver and Polach [1977]. All measurements were corrected for isotope fractionation by measuring $\delta^{13}C$ (relative to PBD-1) on reburned acetylene.

Stable carbon and oxygen isotope ratios were measured on 250-μm-wide subannual samples

TABLE 1. Stable Oxygen and Carbon Isotope Measurements in Subannual Coral Bands From Bermuda

Year	$\delta^{18}O$	$\delta^{13}C$
75.05	-3.28	-0.37
75.01	-3.41	-0.72
74.97	-3.52	-1.09
74.93	-3.52	-0.63
74.89	-3.49	-1.07
74.85	-3.51	-1.19
74.81	-3.51	-0.96
74.77	-3.48	-0.97
74.73	-3.69	-1.21
74.69	-3.73	-1.00
74.65	-3.84	-0.62
74.61	-4.01	-0.90
74.57	-4.29	-1.10
74.53	-3.90	-0.94
74.49	-4.01	-1.10
74.45	-3.90	-1.01
74.41	-3.66	-0.81
74.37	-3.51	-0.67
74.33	-3.64	-0.87
74.29	-3.69	-0.60
74.25	-3.42	-0.61
74.21	-3.94	-0.83
74.17	-3.73	-1.20
74.13	-3.62	-1.05
74.09	-3.90	-1.36
74.00	-3.83	-1.14
73.94	-3.85	-0.51
73.89	-3.92	-1.05
73.84	-4.11	-0.86
73.80	-4.15	-1.19
73.73	-4.05	-1.27
73.68	-3.94	-0.82
73.63	-3.86	-0.94
73.58	-4.00	-0.80
73.52	-3.72	-0.56
73.47	-3.66	-0.64
73.42	-3.57	-0.51
73.37	-3.44	-0.37
73.31	-3.61	-0.28
73.26	-3.73	-0.40
73.21	-3.42	-0.33
73.16	-3.52	-0.52
73.10	-3.52	-0.54
73.05	-3.58	-0.85
73.01	-3.60	-0.79
72.96	-3.53	-0.89
72.92	-3.57	-0.71
72.88	-3.65	-0.99
72.83	-3.94	-0.67
72.79	-3.91	-0.63
72.75	-3.96	-0.88
72.70	-3.92	-0.89
72.66	-3.91	-0.84
72.62	-3.88	-0.64
72.57	-3.82	-0.23
72.53	-3.44	-0.25
72.48	-3.47	-0.20
72.44	-3.65	-0.50
72.40	-3.39	-0.70
72.35	-3.75	-0.54
72.31	-3.37	-0.55
72.27	-3.44	-0.78
72.22	-3.60	-0.58
72.18	-3.41	-0.99
72.14	-3.73	-0.76
72.09	-3.62	-1.03
72.05	-3.56	-0.56

(approximately 15-day sampling) from N-a (D. strigosa) ground with a Dremel tool, and on the annual samples cut for the radiocarbon analyses. The aragonite was prepared and $\delta^{18}O$ and $\delta^{13}C$ measured according to methods reported by Druffel [1985]. A V.G. Micromass 602E mass spectrometer was used and the precision obtained for each was ±0.07°/₀₀.

Results and Discussion

Stable Isotopes

Stable oxygen and carbon isotope data from annual and biweekly sections of northern Bermuda coral (N-a) (Table 1), with accompanying sea surface temperature and salinity records, are presented in Figure 1. The seasonal $\delta^{18}O$ record shows a periodic, annual variation over the 3-year time span studied (Figure 1a). The seasonal variation ranges from 0.6°/₀₀ in 1972 to 0.9°/₀₀ in 1974. Although the observed seasonal change is in the expected direction (high $\delta^{18}O$ during late winter), it is less than half of that expected (2.0°/₀₀) from the change in annual SST (Figure 1c, 1.7°/₀₀) and in surface salinity (Figure 1d, +0.3-0.4°/₀₀) [Fairbanks and Dodge, 1979]. The $\delta^{18}O$ record is damped due to a combination of the slow growth rate of the Diploria corals (3-6 mm/yr) and the uneven topography of the calcio-blastic layer, both of which contribute to sampling of material accreted over several months in a single 250-µm-wide increment.

The seasonal $\delta^{13}C$ record (Figure 1b) appears to covary with the $\delta^{18}O$ record. This does not agree with Fairbanks and Dodge [1979] who reported inverse correlation between the two isotopes in Montastrea annularis from Bermuda. I suspect this bears on McConnaughey's [1986] observation that slow growing species from depth (11 m) contain an isotopic composition that is closer to the equilibrium value than faster-growing specimens (M. annularis) from shallower depths, the isotopic composition of which is controlled to a greater extent by photosynthesis of the symbiotic algae within the coral polyp.

The large corallite size of Diploria lends itself to severe damping of the seasonal signal, hence rendering it of little value for extracting records of biweekly isotope composition. However, the overall shape of the $\delta^{18}O$ signal appears to be without hiatuses, which indicates a constant growth rate throughout the year, a primary concern of this study. Thus on an annual basis, Bermudian Diploria appears to be an

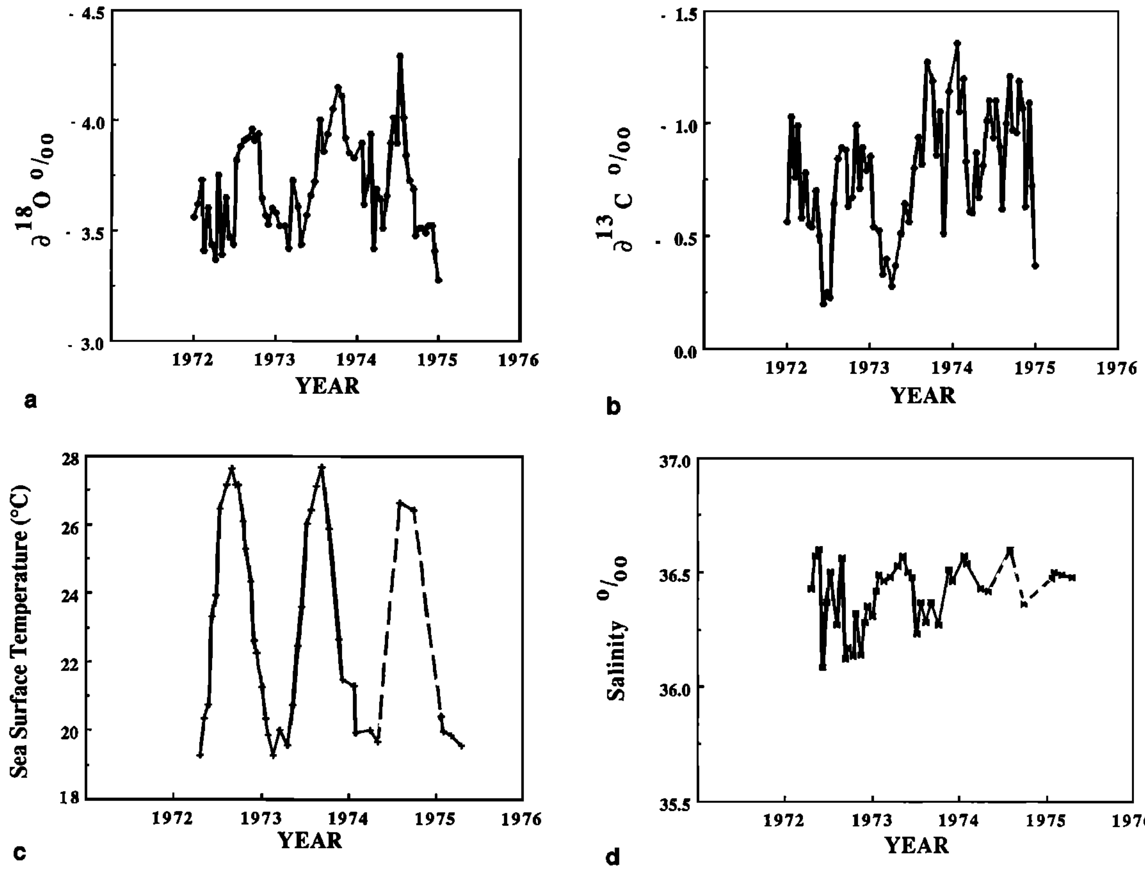

Fig. 1. (a) Stable oxygen and (b) stable carbon isotope ratios measured in biweekly sections of northern Bermuda coral (N-a) from 1972 to 1975. (c) Sea surface temperature in degrees celsius and (d) surface salinity (per mil) taken from the R/V Panulirus near 32°10'N, 64°30'W on a biweekly basis are also shown for this time period.

adequate integrator of the past chemical and isotopic changes in sea water.

In order to determine whether long-term trends in stable isotopes are possible integrators of various sea water properties, $\delta^{18}O$ and $\delta^{13}C$ were measured in annual coral bands from the same N-a coral (Figure 2, Table 2). A slight decrease of 0.1‰ from 1950 to 1983 (Figure 2a) coincides with a slight warming trend in SST of the expected magnitude ($d\delta^{18}O/dSST$ = 0.22‰ per 1°C [Epstein et al., 1953]) (Figure 2c), although neither change is statistically significant. A shift in average $\delta^{18}O$ toward higher values from 1958-1963 to 1964-1968 coincides with an average decrease of 0.15°C in SST and a 0.13‰ rise in salinity (Figures 2c and 2d). These SST and salinity changes will each cause a rise in $\delta^{18}O$, which when combined, approximately equal that observed in the data. There is a linear relationship between $\delta^{18}O$ and salinity ($r = 0.49$, $N = 27$), that is significant at the 99% confidence level ($\alpha = 0.99$). This correlation may illustrate a link between $\delta^{18}O$ in the coral and climate-affected parameters, for example, the latent heat of evaporation which controls salinity in the upper waters of the subtropical gyre.

The $\delta^{13}C$ record appears to increase from an average of -0.4‰ in the 1950s and early 1960s to about -0.25‰ by 1972, and then decreases thereafter to an average of -0.75‰. Although correlations between $\delta^{13}C$ and other parameters ($\delta^{18}O$, salinity, SST) are not significant, co-variance with the water mass renewal rate record presented later in this paper suggests that $\delta^{13}C$ records in banded corals may be potential integrators of ventilation rate in the upper ocean.

Radiocarbon

Radiocarbon results from southern and northern Bermuda corals are shown in Figure 3 and listed in Table 2. Individual analyses from two coral colonies in the north (N-a, N-b) and combined analyses from two in the south (S-a/b) are shown. The line represents the weighted average (with respect to $\pm 1\sigma$ error) of the $\Delta^{14}C$ results.

Five prebomb $\Delta^{14}C$ results from 1950 to 1953 average -48.3‰ (±4.9‰, standard deviation), which compares favorably with Nozaki et al.'s [1978] coral result from North Rock of -52 ± 8‰ for the same time period. Bomb radiocarbon is clearly present in bands younger than 1958 and is perhaps present as early as 1955, as has been observed in equatorial Pacific coral [Druffel, 1987].

Agreement between $\Delta^{14}C$ results for a given year from the three different coral heads is within two σ counting error (4-5‰), except

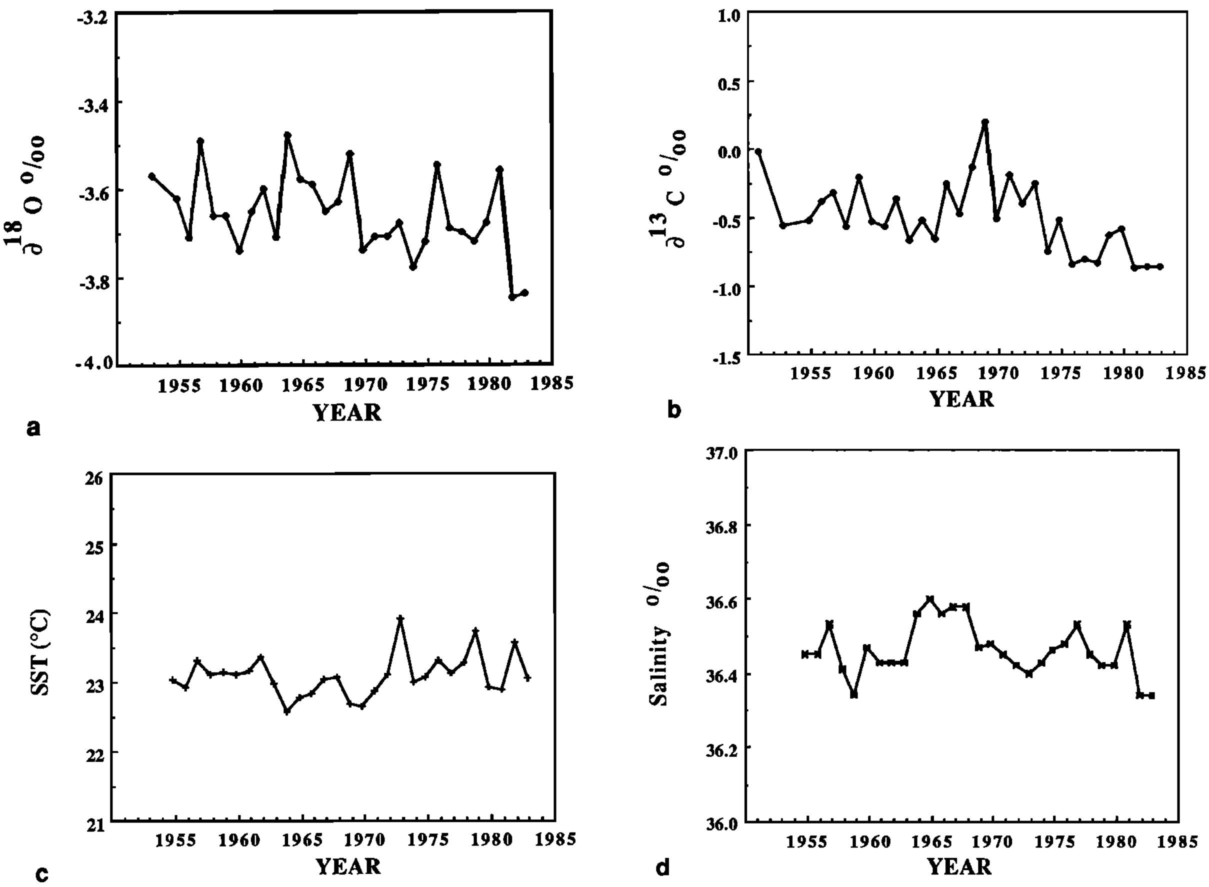

Fig. 2. (a) Stable oxygen and (b) stable carbon isotope ratios measured in annual sections of northern Bermuda coral (N-a) from 1950 to 1983. (c) Annually averaged SST and (d) annually averaged salinity at the Panulirus station for this time period.

for the period 1962 to 1969 when the spread in results is as much as 8 to 9 σ (20°/oo). I suspect that this disparity is due to the error in sectioning the coral bands during a period when the seasonal $\Delta^{14}C$ variation, caused by the large ^{14}C gradient between air and sea, was the greatest.

A regular variation of 10-15°/oo is apparent in the post-1971 $\Delta^{14}C$ results. This is in phase with a variation of 0.25°/oo in the $\delta^{18}O$ data but is offset by 1 year with the $\Delta^{14}C$ record (e.g., $\delta^{18}O(t+1)\ \alpha\text{-}\Delta^{14}C(t)$) (Figure 3b). Post-1981 stable isotope results have been eliminated from this comparison due to the interference of organic matter in the analyses, which is ubiquitous in the 2 youngest bands. A least squares fit of $\Delta^{14}C(t)$ versus $\delta^{18}O(t+1)$ reveals the relationship $\delta^{18}O(t+1) = -0.0103*\Delta^{14}C(t)-2.11$ ($r = -0.68$, number of points $N = 9$) and is statistically significant at the 95% confidence level ($\alpha = 0.95$). This suggests that low SST and/or high salinity accompany periods of low $\Delta^{14}C$ in the Sargasso Sea.

A similar correlation between $\delta^{13}C$ and $\Delta^{14}C$ appears to also be delayed by 1 year (Figure 3c). A least squares analyses reveals the relationship $\delta^{13}C(t+1) = 0.0252*\Delta^{14}C(t)-4.57$ ($r = 0.89$, $N = 8$) and is statistically significant at the 95% confidence level. Whether the $\delta^{13}C$ record in *Diploria* is controlled by water mass changes or a measure of primary productivity is addressed later in the paper.

The radiocarbon results obtained from Florida corals are shown in Figure 4a and listed in Table 3. The points represent the high-precision analyses, and the line is the weighted average of these and results obtained earlier by Druffel and Linick [1978] and Druffel and Suess [1983]. The new results confirm the previously reported bomb radiocarbon time history; a clear downward trend of values from 1975 to 1983 is apparent. $\Delta^{14}C$ of water DIC (117.0°/oo) collected in March 1983 agrees within error with the coral $\Delta^{14}C$ result for 1983 (Table 3). Radiocarbon results from seasonal bands during 1973 and 1974 are shown in Figure 4b. There is a 25-35°/oo range in the $\Delta^{14}C$ values, but the record is too short to discern a regular seasonal cycle. This seasonal range is similar to that found by Broecker and Peng [1980] for surface North Atlantic waters during GEOSECS (1972-1973).

A comparison between the Bermuda and Florida radiocarbon records (Figure 5) is divided into three zones. First, pre-1958 $\Delta^{14}C$ results are lower at Florida by 15°/oo, predominantly due to the input of ^{14}C-depleted equatorial waters to the Gulf Stream precursor [Iselin, 1936]. Second, 1959-1972 results show a 0.5- to 2-year lag in

TABLE 2. Radiocarbon and Stable Isotope Results in Annual Coral Bands From Bermuda

		$\Delta^{14}C$ in Individual Corals[a]			Average[b]		
WHOI Nos.	Year	N-b	N-a	S-a/b	$\Delta^{14}C$	$\delta^{13}C$	$\delta^{18}O$
276	1950.8			-45.6	-45.6	0.02	-3.22
599	1951.3[c]		-46.4		-46.4		
278	1951.8			-54.6	-54.8		
277	1952.8			-52.4	-52.4	-0.56	-3.57
353	1953.3[c]		-43.1		-43.1		
	1954.8			-40.2	-40.2	-0.52	-3.62
	1955.8			-46.3	-46.3	-0.38	-3.71
348	1956.8			-46.6	-46.6	-0.32	-3.49
533, 435	1957.8		-49.3	-40.2	-44.8	-0.57	-3.66
343, 350	1958.8		-37.0	-46.5	-41.8	-0.21	-3.66
534, 535	1959.8		-24.1	-27.1	-25.6	-0.53	-3.74
339, 347	1960.8		-15.7	-21.7	-18.7	-0.57	-3.65
346, 373	1961.8		-12.1	0.8	- 5.6	-0.36	-3.60
340, 349	1962.8		9.1	-13.2	- 2.1	-0.67	-3.71
407, 344, 375	1963.8	37.3	34.5	17.1	25.8	-0.52	-3.48
352, 377	1964.8		68.3	56.8	62.3	-0.66	-3.58
409, 342, 376	1965.8	73.3	78.8	80.5	77.5	-0.25	-3.59
467, 468	1966.8		102.2	89.5	95.8	-0.47	-3.65
408, 341, 374	1967.8	119.0	106.8	105.8	110.5	-0.14	-3.63
473, 464	1968.8		108.6	125.0	116.8	-0.19	-3.52
472, 470	1969.8		125.9	129.6	127.8	-0.51	-3.74
345, 273	1970.8		138.2	137.7	138.0	-0.19	-3.71
275	1971.8			145.6	145.6	-0.40	-3.71
274	1972.8			153.6	153.6	-0.25	-3.68
354	1973.8		157.6		157.6	-0.75	-3.78
296	1974.8		146.5		146.5	-0.52	-3.72
294, 662	1975.8		150.7		150.7	-0.84	-3.55
291	1976.8		150.8		150.8	-0.80	-3.69
295	1977.8		158.8	159.6	159.2	-0.83	-3.70
351, 297	1978.8	158.3	151.5		154.9	-0.63	-3.72
292, 661	1979.8		146.3	153.2	149.8	-0.58	-3.68
285, 283, 271	1980.8	143.3	144.4	146.7	144.8	-0.87	-3.56
272	1981.8			148.7	148.7	-0.86	-3.85
284, 286, 270	1982.8	144.5	138.1	137.6	140.1	-0.86	-3.84
541	1983.3			127.9	127.9		

[a]Precision averages ± 2.5°/₀₀.
[b]Precision averages ± 2.2°/₀₀.
[c]Analyses of 2 consecutive annual coral bands.

the Bermuda $\Delta^{14}C$ results with respect to the Florida record. This is due mainly to the mixing down and subsequent storage of bomb radiocarbon in the upper few hundred meters of the water column in the Sargasso Sea during mode or 18° water formation in late winter. This results in dilution of the bomb radiocarbon signal in the surface layer. The Gulf Stream, in contrast, has no deep convective mixing and thus concentrates the bomb radiocarbon in a relatively shallow mixed layer (100 m); as a result, the levels rise quickly. Third, post-1975 results display the predominance of equatorial water input to the Florida Straits (after the air-sea ^{14}C gradient had decreased), causing radiocarbon to fall once again to levels below those at Bermuda.

Water Mass Renewal: Model Description

The offset of the Bermuda and Florida bomb radiocarbon records may be due mainly to spatial circulation differences in the North Atlantic. However, recent evidence suggests that the ventilation rate has varied on an annual basis off Bermuda [Jenkins, 1982]. Also, there is a significant inverse correlation between SST and wind speed at Bermuda (Figure 6) from 1954 to 1983, indicating that changes in climate vary similarly with changes in upper ocean character (and possibly circulation) during the past few decades. These factors, which implicate nonsteady state conditions with respect to mixing on an annual basis, must be considered when interpreting the Bermuda and Florida ^{14}C records.

In order to use bomb radiocarbon to quantify water mass renewal rate in the Sargasso Sea, a model is constructed that uses transport along isopycnals as its major mixing mode. This model attempts to reproduce the actual mixing processes that occur in the upper ocean, unlike vertical diffusion models which have been used in the past to quantify the distribution of bomb radiocarbon.

A schematic of the multibox model is shown in Figure 7. The surface boxes B, GS, and S represent total $^{14}CO_2$ concentration in Sargasso Sea

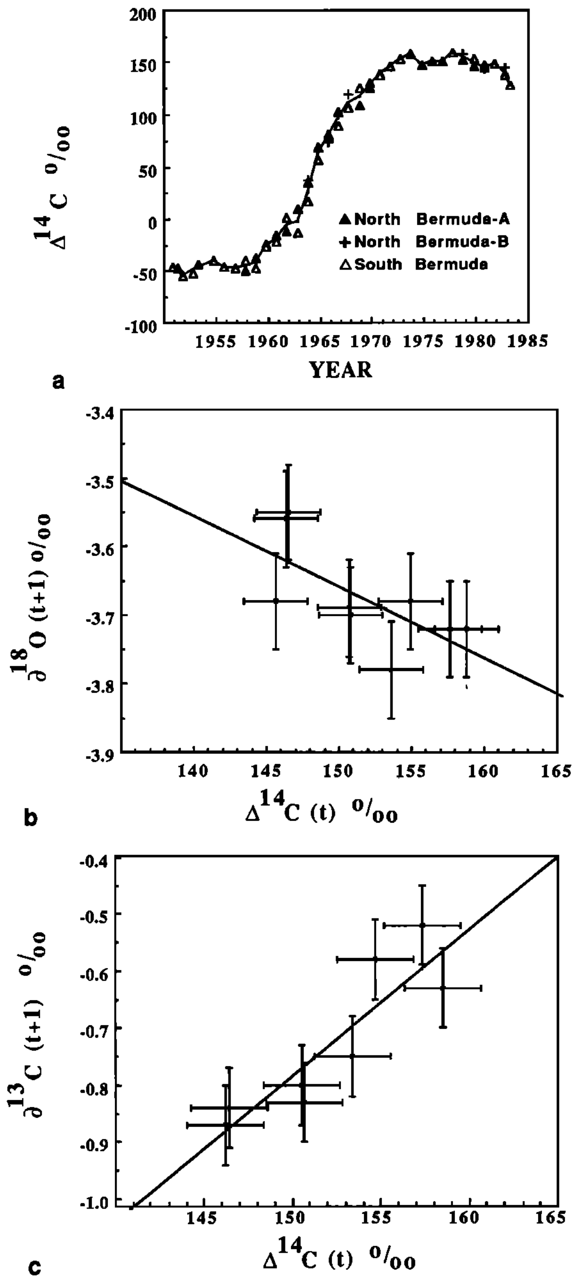

Fig. 3. (a) High precision $\Delta^{14}C$ in annual bands from three corals in the Bermuda area (b) $\Delta^{14}C(t)$ versus $\delta^{18}O(t+1)$ in annual coral bands from 1972 to 1981. Cross-correlation techniques were used to determine that this relationship is significant to the 95% confidence level. (c) $\Delta^{14}C(t)$ versus $\delta^{13}C(t+1)$ in annual coral bands from 1972 to 1981.

surface water at Bermuda, B(t), Gulf Stream surface water as recorded at Florida, GS(t), and slope water entrained in the Sargasso Sea, S(t) (N. Tanaka et al., manuscript in preparation, 1988), respectively. The annually averaged ^{14}C time history in atmospheric CO_2, A(t), is shown in Figure 8 [Levin et al., 1985; Cain and Suess, 1976]. The seven subsurface boxes contain $^{14}CO_2$ concentrations D1(t) through D7(t) which are homogenously mixed along surfaces of constant density (0.1 σ_θ units wide) from 26.4 to 27.0. Deeper density surfaces are assumed to be ventilated exclusively at higher latitudes. Mixing between the surface box B and the subsurface boxes occurs during instantaneous events, attempting to mimic late winter mixing that takes place in the northern half of the Sargasso Sea.

The concentration of $^{14}CO_2$ in box B is affected by input from boxes GS, S, A, and Di. On the basis of the imbalance between the total upward and downward Ekman pumping in the North Atlantic as reported by Sarmiento [1983], horizontal input of Gulf Stream and slope waters (with a 17:3 volume ratio) into B is assumed to be 15% of the volume of B per year, with an equal volume lost out of the sides of B each year to conserve mass. As will be shown later, the model results are relatively insensitive to the value of this parameter. The mixed layer depth, Z, of B is 135 m, the average winter mixed layer at 32°N in the North Atlantic [Levitus, 1982].

The CO_2 gas exchange rate I (in moles/m^2/year), is calculated according to

$$I = Vp * 0.019 \text{ mM} * 365 \text{ d/year} \quad (1)$$

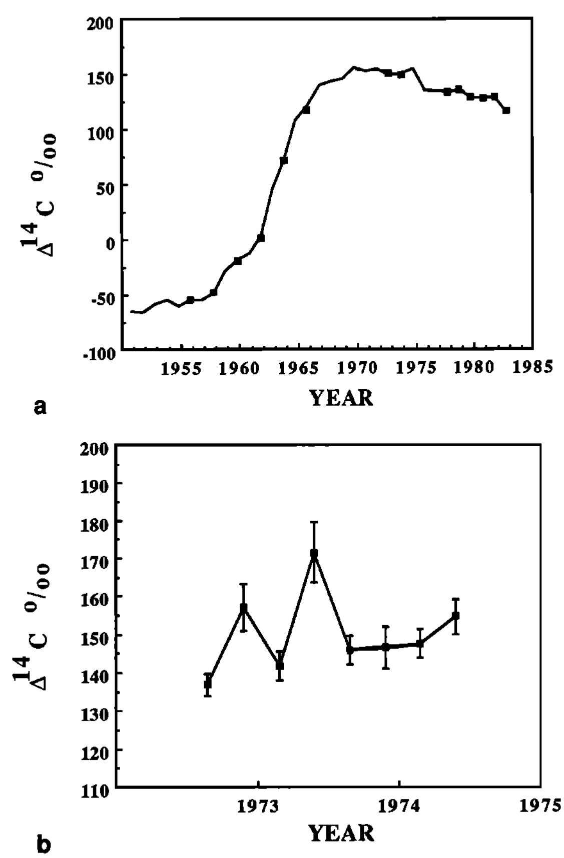

Fig. 4. (a) High-precision $\Delta^{14}C$ measurements in annual coral bands from "The Rocks," Florida (squares) and weighted average $\Delta^{14}C$ trend (line) from the high-precision results and the results obtained earlier by Druffel and Linick [1978] and Druffel and Suess [1983] (b) $\Delta^{14}C$ measurements from seasonal bands during 1973 and 1974.

TABLE 3. Radiocarbon Results in Annual and Quarter-Annual Coral Samples From "The Rocks" Reef off Southern Florida

Year	$\Delta^{14}C$ TR1,2[a]	$\Delta^{14}C$ TR3[b]	$\Delta^{14}C$ Average	Increment, Years	Seawater $\Delta^{14}C$[b]
1951.0	-65.0		-65.0	1	
1952.0	-66.0		-66.0	1	
1953.0	-58.0		-58.0	1	
1954.0	-55.0		-55.0	1	
1955.0	-60.0		-60.0	1	
1956.0		-54.4	-54.4	1	
1957.0	-54.0		-54.0	1	
1958.0	-54.0	-47.2	-48.6	1	
1958.5	-28.0		-28.0	2	
1960.0	-11.0	-19.7	-18.0	1	
1961.0	-12.0		-12.0	1	
1962.0	14.0	0.7	3.4	1	
1963.0	45.0		45.0	1	
1964.0	73.0	71.5	71.8	1	
1965.0	108.0		108.0	1	
1966.0	134.0	117.0	120.0	1	
1967.0	140.0		140.0	1	
1968.0	143.0		143.0	1	
1969.0	146.0		146.0	1	
1970.0	156.0		156.0	1	
1971.0	152.0		152.0	1	
1972.0	155.0		155.0	1	
1973.0	144.0	149.6	148.5	1	
1974.0	155.0	148.6	150.0	1	
1975.0	154.0		154.0	1	
1976.0	135.0		135.0	1	
1977.0	134.0		134.0	1	
1978.0	132.0	132.6	132.5	1	
1979.0	135.0[b]	135.4	135.3	1	
1980.0	128.0[b]	128.3	128.3	1	
1981.0	127.0[b]	127.3	127.2	1	
1982.0	128.0[b]	128.4	128.3	1	
1983.0	115.0[b]	115.4	115.3	1	
1983.2					117.0
1972.65		136.8	136.8	0.25	
1972.90		157.1	157.1	0.25	
1973.15		141.8	141.8	0.25	
1973.40		171.5	171.5	0.25	
1973.65		145.7	145.7	0.25	
1973.90		146.6	146.6	0.25	
1974.15		147.4	147.4	0.25	
1974.40		154.4	154.5	0.25	

TR1 and 2 are separate cores from the same coral colony collected in 1975 and 1978, respectively. TR3 is a suite of five cores from smaller, individual coral colonies.

[a]Precision averages ±3.0 to 6.0‰ [Druffel and Linick, 1978; Druffel and Suess, 1983].

[b]Precision averages ±2.5‰.

where the gas exchange piston velocity, Vp (in meters per day), is a function of wind speed WS (for WS > 4 m/s) according to the fit by Jenkins [1988] of the data presented by Roether [1986]:

$$Vp = WS * 0.9995 - 3.47 \qquad (2)$$

Annually averaged scalar wind speeds WS from the Bermuda Naval Air Station were used. This WS record agreed within 12% of that reported by Bunker [1975] for Marsden square 115 obtained from ship observations. For simplicity, a steady state is assumed over the course of the model run (1950-1983), which means there is no net flux of CO_2 into the ocean over time. This assumption does not change the outcome of the model results.

The input of $^{14}CO_2$ to the ocean is a function of the differences between both the partial pressure of CO_2 (pCO_2) and the $^{14}C/^{12}C$ ratios in the atmosphere (a) and those in the surface waters (s) [Druffel, 1987], according to equation (3):

$$F(t) = \frac{A(t)/0.983 - B(t)*(pCO_2s/pCO_2a)}{[A(t)/0.983 - B(t)]} \qquad (3)$$

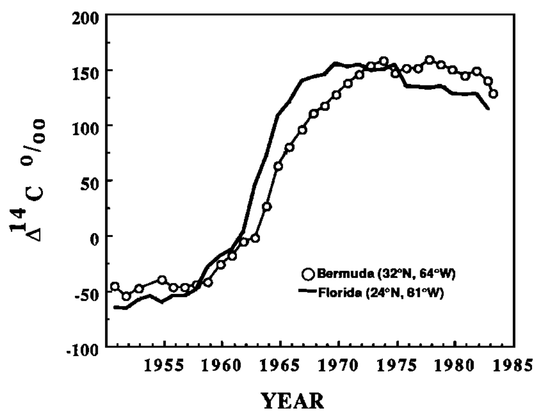

Fig. 5. Comparison of $\Delta^{14}C$ trends in Bermuda and Florida corals.

For example, if $pCO_2a = pCO_2s$, then $F(t) = 1.0$, which means that the net transfer of $^{14}CO_2$ to the surface ocean is a function of the $^{14}C/^{12}C$ gradient exclusively. According to Broecker et al. [1985], the annually averaged difference in partial pressure of CO_2 between air and sea is about 16 ppm between 10°N and 40°N in the North Atlantic, thus F(t) ranges from 1.05 to 1.50 throughout the postbomb period. Since the maximum seasonal input of CO_2 to Sargasso Sea surface waters during February [Brewer, 1986] does not occur simultaneously with the seasonal maximum in tropospheric ^{14}C noticed during the 1960s during June-July [Nydal et al., 1979], it is reasonable to assume that $^{14}CO_2$ input during this period is adequately represented using annual $\Delta^{14}C$ averages for A(t) (Figure 8).

The $^{14}CO_2$ concentration in box B at time $t+\Delta t$ is equal to B(t) plus the $^{14}CO_2$ transfer (per liter) into and out of box B:

$$\begin{aligned} B(t+\Delta t) = {} & B(t)+0.15*\Delta t[0.85*GS(t) \\ & + 0.15*S(t)-B(t)]+\Delta t*[k_{-1}*A(t) \\ & - k_1*B(t)]+\Sigma D_i(t)*W_1(t)*\Sigma S_i*\Delta t \\ & - W_1(t)*\Sigma S_i*B(t)*\Delta t \end{aligned} \tag{4}$$

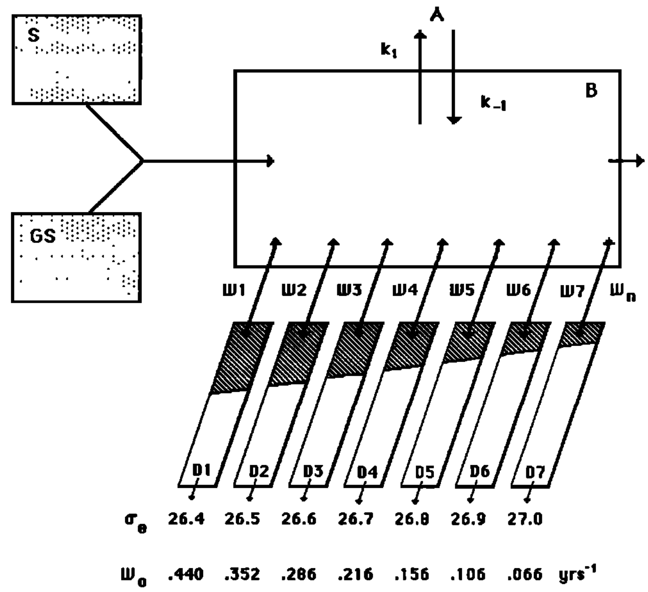

Fig. 7. Schematic of the multibox model used to calculate water mass renewal rate in the Sargasso Sea (see text for details).

where

$$\Delta t = 0.1 \text{ year}$$

$$\Sigma D_i(t) = [D_1(t)+S_2*D_2(t)+S_3*D_3(t)+S_4*D_4(t) + S_5*D_5(t)+S_6*D_6(t)+S_7*D_7(t)]*[\Sigma S_i]^{-1}$$

$$S_i = W_i/W_1(t)$$

$$\Sigma S_i = 1.00 + 0.800 + 0.650 + 0.490 + 0.355 + 0.240 + 0.150 = 3.68$$

$$k_{-1} = F(t)*AFAC/0.983 \ (\text{year}^{-1})$$

$$k_1 = AFAC = I/(Z*\Sigma CO_2) \ (\text{year}^{-1})$$

Z depth of box B (meters)

ΣCO_2 total CO_2 concentration in box B (moles/m^3)

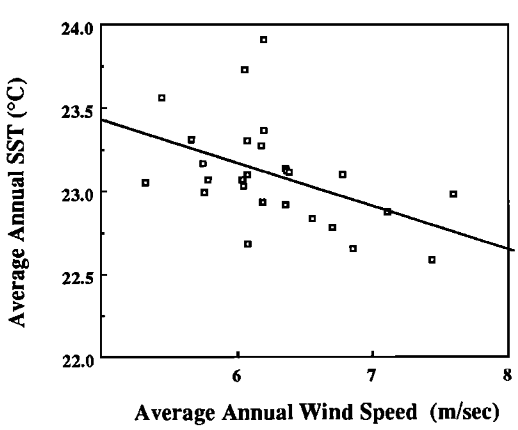

Fig. 6. Correlation between annually averaged SST (see Figure 1c) at Panulirus and annually averaged wind speed at Bermuda Naval air station ($r = 0.45$, $N = 28$, $\alpha = 0.98$).

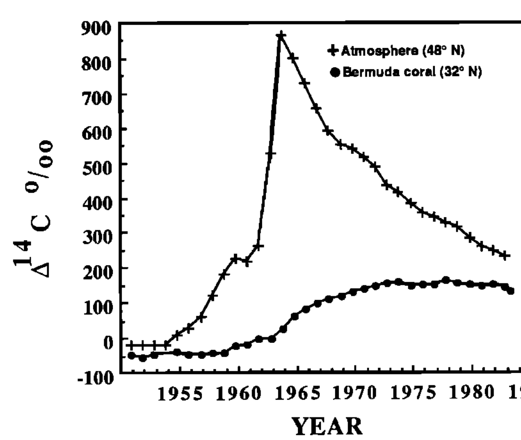

Fig. 8. Time histories of annually averaged bomb radiocarbon in atmospheric CO_2 of the northern hemisphere [Levin et al., 1985; Cain and Suess, 1976] and in Bermuda corals.

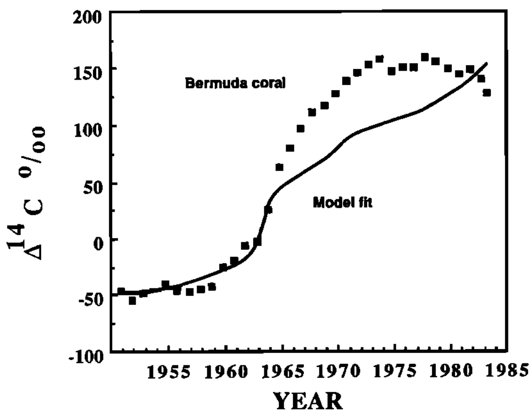

Fig. 9. Results of the forward model calculation (line) using a constant W_1 value of 0.44 yr^{-1}. Bermuda coral results are shown for comparison.

A 0.1-year time interval was chosen on the basis of stability requirements. The relative change in the ventilation rates on each isopycnal, S_i, are assigned in accordance with those observed by Jenkins [1982]. For example, when W_1 (26.4 σ_θ) increased by a factor of 2, then W_4 (26.7 σ_θ) increased by only 40%.

Model Results

Equation (4) is solved for $B(t+\Delta t)$ using the constant annual value for $W_1(t)$ of 0.44 yr^{-1} in order to satisfy the prebomb, steady state $\Delta^{14}C$ value of -48.3°/oo in the Bermuda corals. Results of the forward calculation are shown in Figure 9. The line shows the model-calculated surface $\Delta^{14}C$ record, which deviates from the observed Bermuda record (squares) for the period > 1960. Thus constant $W_1(t)$ of 0.44 yr^{-1} is too high during the early 1960s and the late 1960s through the 1970s; there was an apparent excess of $^{14}CO_2$ in box B during these periods. The assumption of constant W_1 is not consistent with the high precision coral record.

It would be most informative to do the inverse calculation, that is, to calculate the W_1 record needed to reproduce the Bermuda coral $\Delta^{14}C$ exactly. Equation (4) is rearranged to solve for the water mass renewal rate, $W_1(t)$, of water in the uppermost subsurface box with respect to exchange with B:

$$W_1(t) = \frac{B(t+\Delta t) - B(t) - X(t)}{[\Sigma D_i * \Sigma S_i - B(t) * \Sigma S_i] * \Delta t} \tag{5}$$

where

$$X(t) = 0.15*\Delta t[0.85*GS(t)+0.15*S(t)-B(t)] + k_{-1}*A(t)*\Delta t - k_1*B(t)*\Delta t$$

The inverse model is run using a 0.1-year time interval. The subsurface $\Delta^{14}C$ values are calculated using equation (6):

$$D_i(t+\Delta t) = D_i(t) - W_i*D_i(t)*\Delta t + B(t)*W_i*\Delta t \tag{6}$$

and are compared below with GEOSECS and TTO depth profiles for the periods 1973 and 1981, respectively. The slopes of the model calculated depth profiles are controlled by the S_i values, where $S_i = W_i/W_1$.

The resultant time history of $W_1(t)$ for the time period 1950 to 1983 is presented in Figure 10. Values decrease from 1959 to 1962 owing to the quick rise of $\Delta^{14}C$ at Bermuda (Figure 3a). A subsequent slowing of the rate of increase of $\Delta^{14}C$ causes the $W_1(t)$ record to recover to normal values (0.3 to 0.5 yr^{-1}) from 1962 to 1964. Low values (about 0.1 yr^{-1}) are obtained for the late 1960s through 1978 to satisfy the high $\Delta^{14}C$ values. A subsequent recovery to higher W_1 values is encountered after 1981, in keeping with the crossover of the predicted and observed $\Delta^{14}C$ records shown in Figure 9. $W_1(t)$ during the 1950s is highly prone to small errors in the $\Delta^{14}C$ record ($\pm$2°/oo); thus pre-1958 $W_1(t)$ values are known only to $\pm$50%. Also, the early period in the $W_1(t)$ record is artificially high because of the finite mixing time of the isopycnal layers within the gyre.

An important test of the model's ability to distribute ^{14}C into the main thermocline is the agreement between the $\Delta^{14}C$ values calculated for the subsurface boxes (D_1-D_7) to those in the GEOSECS and TTO data sets. The S_i values were initially assigned based on the slope of the $\Delta^{14}C$ depth profiles of the GEOSECS and TTO data. Figure 11 shows an adequate fit of the model results during 1973 (heavy line) to data from the GEOSECS depth profiles (stations 120 and 29). However, by 1981, the model-calculated values are not as high as data from the TTO depth profiles (test stations 2, 3, and 4, and station 11). The model appears unable to pump down enough $^{14}CO_2$ to match the amount present in the main thermocline by the early 1980s, unless unrealistically large input functions of atmospheric ^{14}C are assigned. The excess ^{14}C probably comes from ventilation processes on the northern boundary of the Sargasso Sea (not accounted for by the model) where most of the 18° (mode) water is formed.

Mean sea level at Bermuda [NOAA, 1983] directly correlates with depth of the 26.2 σ_θ surface [Jenkins and Goldman, 1985] for 1961-1980 ($r = 0.68$, $N = 15$, $\alpha = 0.99$). There was a deepen-

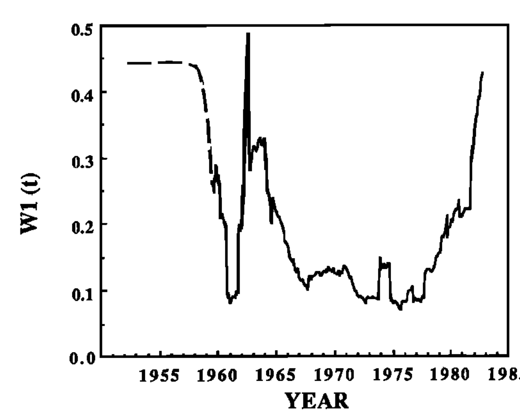

Fig. 10. Time history of $W_1(t)$ for the time period 1950 to 1983 calculated using the multibox model (see text for details).

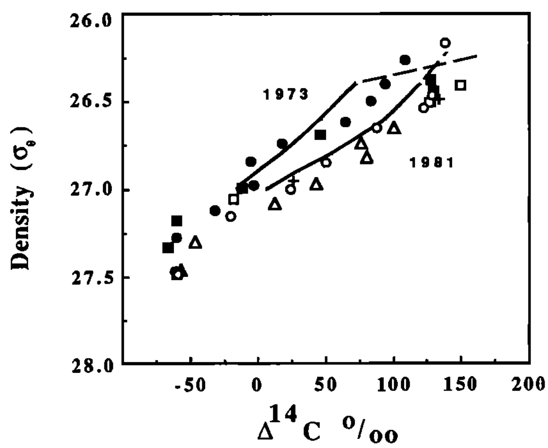

Fig. 11. Model-calculated $\Delta^{14}C$ versus σ_θ (lines) for 1973 and 1981. $\Delta^{14}C$ results of seawater measurements made during GEOSECS in 1973 (solid circles, station 29; solid squares, station 120) and during TTO in 1981 (open circles, station T3; open triangles, station T4; open squares, station 11; pluses, station T2) are shown for comparison.

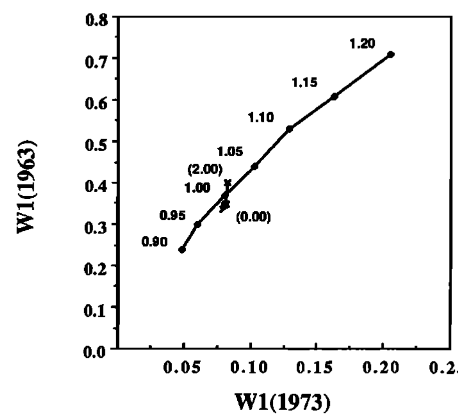

Fig. 12. Calculated values of $W_1(t)$ during 1973 versus those during 1963, using changes in wind speed (0.9 to 1.2 times the actual values) and in annual rate of lateral replacement of waters from the Gulf Stream/slope water (0 to 2 times the actual values) into box B. $W_1(1973)$ and $W_1(1963)$ represent minimum and maximum values in the $W_1(t)$ record (see Figure 10).

ing of isopycnals (26.2-26.4 σ_θ) in the Sargasso Sea from 1972 to 1979 [Jenkins and Goldman, 1985]. This coincides with very high surface $\Delta^{14}C$ (Figure 9) and the lowest $W_1(t)$ values (Figure 10). It appears that less bomb radiocarbon penetrated the subsurface isopycnals during this period.

In order to determine the significance of the $W_1(t)$ record, we submit the model to a suite of sensitivity analyses which consist of unreasonably large changes in several parameters. First, if we interject a random noise to the $\Delta^{14}C$ record in an effort to increase the error in the measurements to ± 7‰, the resultant $W_1(t)$ record is noisier, but the same general features are retained. Thus high-precision $\Delta^{14}C$ analyses are desirable to detect an accurate $W_1(t)$ record. Second, the effect of the large spread of individual $\Delta^{14}C$ results on $W_1(t)$ during 1962-1969 was tested by running the inverse model (equation (5)) using $\Delta^{14}C$ values that were both ± 10‰ during this period. This changed the amplitude of the W_1 maximum in the early 1960s by a factor of $\pm 30\%$, but the W_1 values were still a factor of 2 to 4 times higher during the early 1960s than those during the late 1960s and early 1970s. Third, elimination or doubling of the Gulf Stream/Slope water input to B changes the maximum and minimum $W_1(t)$ during 1963 and 1973, respectively, by less than 10% (Figure 12). Fourth, 10-20% changes in WS result in 30-90% changes in the $W_1(t)$ values, although the ratio of $W_1(1963)/W_1(1973)$ remains relatively constant (Figure 12). Fifth, substitution of the Florida coral record for $\Delta^{14}C$ in B(t) reveals a $W_1(t)$ record with relatively low values (~0.15 yr^{-1}) throughout the 1960-1980 period. Although the mixing processes are different for the Florida location (no mode water formation), this result suggests that the fringes of the Sargasso Sea have also been affected by a decrease in ventilation.

A more stringent test of the model is to assume that $W_1(t)$ is constant during the period of observation and to calculate changes of various parameters necessary to accomodate this assumption. First, the CO_2 exchange rate I was calculated assuming a constant $W_1(t)$:

$$I = [(B(t+\Delta t)-B(t)*(1-W_i(t)*\Sigma S_i)-\Sigma D_i(t)*\Sigma S_i) - X]*Z*\Sigma CO_2)/\Delta t*[F(t)*A-B(t)] \quad (7)$$

Results of this calculation reveal that annually averaged I values would have had to have been <5 $mol/m^2/yr$ both before 1959 and after 1978 and >40 $mol/m^2/yr$ between 1965 and 1973. Values of <5 and >40 $mol/m^2/yr$ are outside of the range of estimates determined using ^{222}Rn [Smethie et al., 1985; Peng et al., 1979] and ocean-wide natural and bomb radiocarbon distributions [Broecker et al., 1985]. Although it is difficult to envision decade time scale variation in the exchange rate of CO_2 of the order of 4-5 times, it cannot be completely ruled out. As I is a function of WS, this would mean that annually averaged wind speed

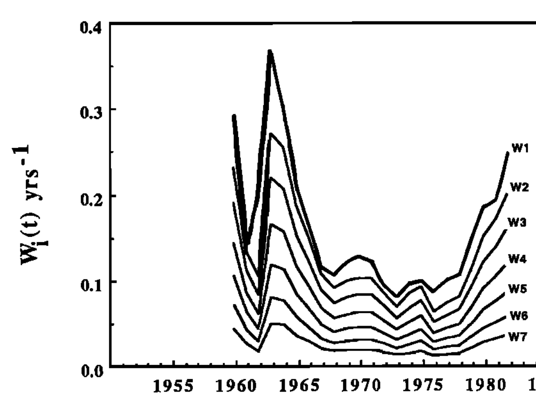

Fig. 13. Annually averaged rates of water mass renewal $W_i(t)$ on all seven isopycnal surfaces as used in the model.

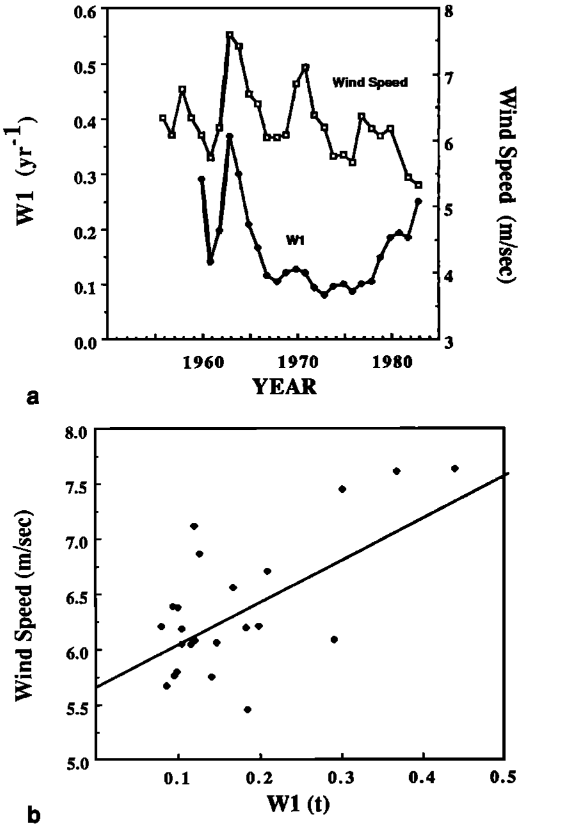

Fig. 14. a) Annually averaged $W_1(t)$ (solid diamonds) and average scalar wind speed (open squares) (from monthly averaged scalar wind speeds) at Bermuda Naval Air Station (W. Spitzer, personnal communication, 1988). (b) Linear regression of $W_1(t)$ versus WS ($r = 0.73$, $N = 23$, $\alpha = 0.95$).

would have had to vary by nearly a factor of 2 over the course of the model run, a situation for which there is no basis in the recorded wind data.

It appears that the $W_1(t)$ record withstands the scrutiny of these sensitivity analyses and that a decrease in W_1 during the late 1960s and 1970s is apparent within the bounds of the model. Annually averaged $W_1(t)$ values for all of the isopycnals are shown in Figure 13. Although W_2-W_7 are not independently derived, they are shown to illustrate that the renewal rate of the densest water mass (27.0 σ_θ) is ventilated, according to assumptions made in the model, at a rate approximately 6 times slower than the shallowest subsurface water mass (26.4), with turnover times of the order of 20 to 60 years.

There is a significant positive correlation ($r = 0.68$, $N = 22$, $\alpha = 0.999$) between our annually averaged $W_1(t)$ record and wind speed (Figure 14a). This does not imply that wind speed is necessarily the direct initiator of increased water mass renewal; however, the correlation may indicate common cause. The correlation is still significant between WS and a $W_1(t)$ record determined using constant wind speed ($R = 0.36$, $N = 22$, $\alpha = 0.90$). This suggests that the ^{14}C record in the Bermuda corals is not simply a function of the CO_2 exchange rate.

This decoupling implies that ^{14}C is primarily a water mass tracer, sensitive to changes in the renewal rate between surface and deeper isopycnal levels. Systematic variations of salinity on surfaces of constant density, caused by changes in latent heat of evaporation and significant correlations with oxygen, led Jenkins [1982] to calculate that water mass renewal or the rate of ventilation in the Sargasso Sea had changed by a factor of 2 over the past 3 decades. Helium-3 distributions during two time periods were consistent with this record. The changes in isopycnal salinity with time (>0.10°/₀₀) were equal to or greater than the changes observed in isopycnal salinity throughout the entire gyre [Bainbridge, 1981]. This observation, coupled with changes in climatology, suggest that drifting or flopping of the gyre with time was not the major cause of the changes in salinity at the Panulirus station and that changes in the latent heat of evaporation are at least partially responsible for the change of isopycnal salinity. A similar look at the $\Delta^{14}C$ gradients on isopycnals (26.2-27.0 σ_θ) from 1980 to 1981 [Ostlund, 1981; Top, 1984] reveals a range of less than 25°/₀₀ over a 1600-km span of the gyre.

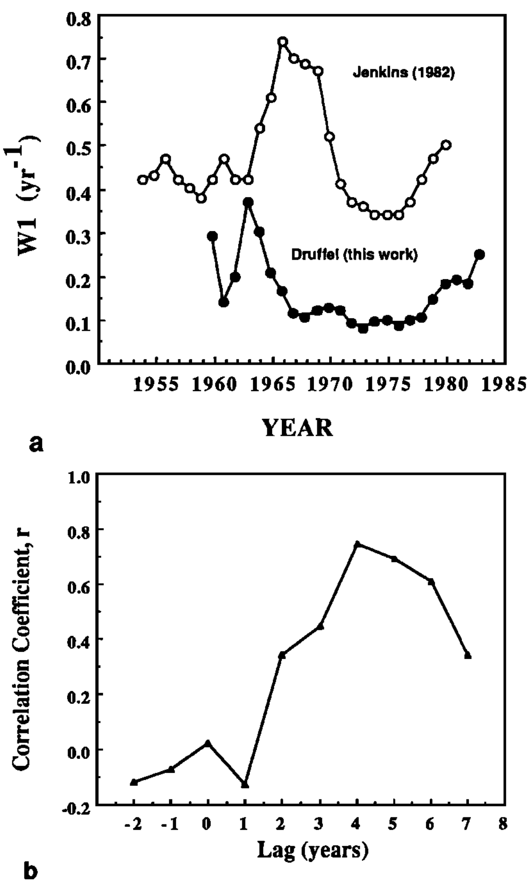

Fig. 15. a) Comparison of $W_1(t)$ record calculated using bomb radiocarbon and salinity/AOU [Jenkins, 1982]. b) Correlation coefficients of $W_1(t)$ reported here and that reported by Jenkins [1982] versus time lag in years of the Jenkins data. The confidence levels for t+4, t+5 and t+6 are $\alpha = 0.999$ ($N = 17$), 0.995 ($N = 16$) and 0.98 ($N = 15$), respectively.

TABLE 4. Lagged Linear Correlation Coefficients (r) for $W_1(t)$ Determined From ^{14}C Model (This Work) versus $W_1(t)$ Determined From Isopycnal Salinity/AOU [Jenkins, 1982] (see Figure 15b)

Lag		Number of		Confidence
$W_1(t)$ (This Work)	$W_1(t)$ [Jenkins, 1982]	Points (N)	r	Level α
t	t	22	0.02	
t	t+1	21	−0.12	
t	t+2	20	0.34	
t	t+3	19	0.44	0.90
t	t+4	17	0.75	0.999
t	t+5	16	0.69	0.995
t	t+6	15	0.61	0.98
t	t+7	14	0.34	
t	t−1	21	−0.07	
t	t−2	20	−0.12	

Both water mass renewal rate records (Figure 15a) show a $W_1(t)$ maximum in the early to mid 1960s and the correlation coefficient is highest ($r = 0.75$, $N = 17$) when there is a lag of 4 years in Jenkins' isopycnal salinity record (Figure 15b, Table 4). The 4-year lag between the ^{14}C- and salinity-derived records is greater than the one year lag between the $\delta^{18}O$ and $\Delta^{14}C$ records observed in the Bermuda coral during the period 1972-1981 (Figure 3b). This suggests that the processes controlling radiocarbon precede those controlling salinity in the surface by 1 year and at 26.4 σ_θ by 4 years.

The lag of the salinity/AOU record implies that the rise in salinity on density surfaces below the mixed layer (>26.3 σ_θ) is not realized for years (of the order of the circulation time of the gyre) after the rise in latent heat flux out of the surface ocean, as was originally projected by Jenkins [1982]. The salinity and $\delta^{18}O$ signals in the surface ocean are apparently controlled by atmospheric forcing, but transfer of this signal to subsurface isopycnals (>26.3 σ_θ) takes on the order of years. Radiocarbon, in contrast, is controlled by mixing among water masses, and to a much lesser extent by changes in the CO_2 exchange rate between air and sea.

Presumably, if $W_1(t)$ has changed with time, the $\delta^{13}C$ of ΣCO_2 in surface waters may also have changed in accordance with observed depth variations of $\delta^{13}C$ in DIC [Kroopnick, 1985]. There is a significant direct correlation between the annual coral $\delta^{13}C$ record and $W_1(t)$ when $\delta^{13}C$ is lagged by 6 years ($r = 0.74$, $N = 16$, $\alpha = 0.998$), although any correlation must be approached with caution in view of the growth rate changes in the coral specimen. Barring any change in the internal fractionation of C isotopes by the coral with time, a direct correlation between $\delta^{13}C$ (t+6) and $W_1(t)$ could represent a link to primary production. That is, there is more $^{12}CO_2$ stripped from DIC during photosynthesis, leaving the DIC pool enriched in ^{13}C. The lag time in the correlation indicates that there may be a long response time, though 6 years seem unreasonably long for an increase in primary productivity following a period of high water mass renewal rate.

Comparison of W_1 values for the North Atlantic obtained from three separate studies is presented in Figure 16: (1) this work, ^{14}C, 1973; (2) Jenkins [1982], salinity/AOU, 1972; and (3) Sarmiento [1983], tritium inventories, 1972-1973. At 26.7 σ_θ and below, the agreement is within 50%, which is acceptable considering the range of tracers used and the variations in the geographical locations studied; Sarmiento [1983] used tritium inventories that were averaged over the entire North Atlantic region. Above the 26.7 σ_θ surface, the salinity/AOU-derived W_1 values are higher than those derived from bomb radiocarbon or from tritium inventories. It has been suggested by Doney and Jenkins [1988] that tritium underestimates the ventilation of the upper ocean with respect to other tracers (e.g., ^3He), due to incomplete resetting of boundary conditions during winter mixing. Radiocarbon has a similar boundary condition to tritium, since it exchanges very slowly with the atmosphere.

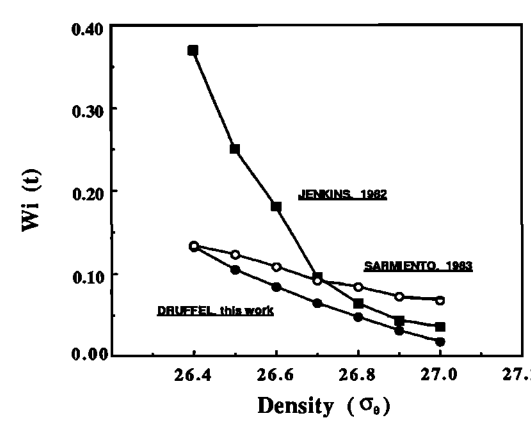

Fig. 16. Comparison of $W_1(t)$ on σ_θ surfaces 26.4 to 27.0 during 1972 to 1973 as calculated by Jenkins [1982], Druffel (this work), and Sarmiento [1983].

Implications of Changing Water Mass Renewal Rate

Changes in water mass renewal rate influence several fields of geochemistry. First, if there are changes in the rate at which subsurface waters are brought to the surface, presumably the rate at which preformed nutrients enter the surface ocean also changes. As the net flux of nutrients into the surface from below is equal to the net flux out of the euphotic zone (new production), one can estimate that new production varied by more than a factor of 2 during the last few decades, assuming a one-to-one correlation between preformed nutrient supply and new production. This has important implications for global estimates of carbon fluxes from the surface into the main thermocline.

Changes in the rate of water mass renewal will also affect models of excess CO_2 uptake by the ocean, cycling of trace elements that are scavenged from the surface waters, and long-term climate models addressing such phenomena as El Nino/Southern Oscillation.

Conclusions

1. High-precision bomb radiocarbon measurements made in annual bands from different Bermudian coral colonies are statistically the same, except during the 1960s when there was high seasonal variability of $\Delta^{14}C$.

2. High-precision radiocarbon trends from Bermuda and Florida are used in conjunction with an inverse model to calculate water mass renewal rates for the North Atlantic subtropical gyre. The rate of water mass renewal in the Sargasso Sea was especially low during the late 1960s and early 1970s to mid-1970s, coincident with lower annually averaged wind speed.

3. The lag between the radiocarbon-derived W_1 record and Jenkins isopycnal salinity/AOU-derived record indicates an inherent time shift between different tracers of water mass renewal rate and the forcing functions thereof (e.g., wind speed).

4. Changes in the rate of water mass renewal have important implications for present theories of new production, input of fossil fuel CO_2 to the oceans, and determination of ocean circulation in general.

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References

Bainbridge, A. E., GEOSECS Atlantic Expedition, vol. 1, Hydrographic Data 1972-1973, 120 pp., U.S. Government Printing Office, Washington, D. C., 1981.

Brewer, P. G. Upper oceans Global Ocean Flux Study, U.S. GOFS Study 7, 35 pp., U.S. GOFS Planning Office, Woods Hole Oceanogr. Inst., Woods Hole, Mass., 1986.

Brewer, P. G., W. S. Broecker, W. J. Jenkins, P. B. Rhines, C. G. Rooth, J. H. Swift, T. Takahashi, and R. T. Williams, A climatic freshening of the deep Atlantic north of 50°N over the past 20 years. Science, 222, 1237-1239, 1983.

Broecker, W. S., T.-H. Peng, H. G. Ostlund, and M. Stuiver, The distribution of bomb radiocarbon in the ocean, J. Geophys. Res., 90(C4), 6953-6970, 1985.

Buddemeier, R. W., J. E. Maragos, and D. W. Knutson, Radiographic studies of reef coral exoskeletons: rates and patterns of coral growth, J. Exp. Mar. Biol. Ecol., 14, 179-200, 1974.

Bunker, A. F, Energy Exchange at the Surface of the western North Atlantic Ocean, Tech. Rep. 75-3, Woods Hole Oceanogr. Inst., Woods Hole, Mass., 1975.

Cain, W. F. and H. E. Suess, Carbon-14 in tree rings, J. Geophys. Res., 81, 3688-3694, 1976,.

Doney, S. C., and W. J. Jenkins, The dependence of ventilation rates on transient tracer boundary conditions, J. Mar. Res., 46, 1988.

Druffel, E. M., and T. W. Linick, Radiocarbon in annual coral rings of Florida, Geophys. Res. Lett., 5(11), 913-916, 1978.

Druffel, E. M., and H. E. Suess, On the radiocarbon record in banded corals: exchange parameters and net transport of $^{14}CO_2$ between atmosphere and surface ocean, J. Geophys. Res., 88(C2), 1271-1280, 1983.

Druffel, E. R. M., Detection of El Nino and decade time scale variations of sea surface temperature from banded coral records: Implications for the carbon dioxide cycle, in The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present, Geophys. Monogr., Ser., vol. 32, edited by E. T. Sundquist and W. S. Broecker, pp. 111-122, AGU, Washington, D. C., 1985.

Druffel, E. R. M., Bomb radiocarbon in the Pacific: Annual and seasonal time scale variations, J. Mar. Res., 45, 667-698, 1987.

Epstein, S., H. A. Buchsbaum, H.A. Lowenston, and H. C. Urey, Revised carbonate-water isotopic temperature scale, Geol. Soc. Am. Bull., 64, 1315-1326, 1953.

Fairbanks, R.G., and R. E. Dodge, Annual periodicity of the $^{18}O/^{16}O$ and $^{13}C/^{12}C$ ratios in the coral Montastrea annularis. Geochim. Cosmochim. Acta, 43, 1009-1020, 1979.

Fine, R. A., Direct evidence using tritium data

for throughflow from the Pacific into the Indian Ocean, Nature, 315, 478-480, 1985.
Folland, C. K., D. E. Parker, and F. E. Kates, Worldwide marine temperature fluctuations 1856-1981. Nature, 310, 670-673, 1984.
Griffin, S., and E. R. M. Druffel, Woods Hole Oceanographic Institution Radiocarbon Laboratory: Sample treatment and gas preparation, Radiocarbon, 27(1), 43-51, 1985.
Iselin, C. O. D., A study of the circulation of the western North Atlantic, Pap. Phys. Oceanogr. Meteorol., 4, 101, 1936.
Jenkins, W. J., Tritium and He-3 in the Sargasso, J. Mar. Res., 38, 533-569, 1980.
Jenkins, W. J., On the climate of a subtropical ocean gyre: Decade time-scale variations in water mass renewal in the Sargasso Sea, J. Mar. Res., 40, 265-290, 1982.
Jenkins, W. J., The nitrate flux into the euphotic zone near Bermuda, Nature, 331, 521-523, 1988.
Jenkins, W. J., and J. C. Goldman, Seasonal oxygen cycling and primary production in the Sargasso Sea, J. Mar. Res., 43, 465-491, 1985.
Jones, P. D., T. M. L. Wigley, and P. M. Kelly, Variation in surface air, 1, Northern hemisphere, 1881-1980, Mon. Weather Rev., 110, 59-70, 1982.
Konishi, K., T. Tanaka, and M. Sakanoue, Secular variation of radiocarbon concentration in seawater: Schlerochronological approach, Proc. Int. Coral Reef Symp., 4(1), 181-185, 1981.
Kroopnick, P, The distribution of ^{13}C of ΣCO_2 in the worlds oceans, Deep Sea Res., 32(1), 57-84, 1985.
Levin, I., et al., Twenty-five years of tropospheric ^{14}C observations in Central Europe, Radiocarbon, 27, 1-19, 1985.
Levitus, S, Climatological atlas of the world ocean, NOAA Prof. Pap. 13, U.S. Government Printing Office, Washington, D. C., 1982.
Linick, T. W., Bomb-produced carbon-14 in the surface water of the Pacific Ocean, Radiocarbon, 22, 599-606, 1980.
McConnaughey, T. A., Oxygen and carbon isotope disequilibria in Galapagos corals: Isotopic thermometry and calcification physiology, Ph.D. thesis, Univ. of Wash., Seattle, 1986.
Michel, R. L., H. E. and Suess, Bomb-tritium in the Pacific Ocean, J. Geophys. Res., 80, 4139-4152, 1975.
Montgomery, R. B., Circulation in the upper layers of the southern North Atlantic deduced with use of isentropic analysis, Pap. Phys. Oceanogr. Meteorol., 6, 55, 1938.
NOAA, Sea level variations for the United States 1855-1980, 130 pp., U.S. Department of Commerce, Rockville, Md, 1983.
Nozaki, Y., D. M. Rye, K. K. Turekian, and R. E. Dodge, C13 and C14 variations in a Bermuda coral, Geophys. Res. Lett., 5, 825-828, 1978.
Nydal, R., K. Loveseth, and S. Gulliksen, A survey of radiocarbon variation in nature since the Test Ban Treaty, in Radiocarbon Dating, edited by R. Berger and H. E. Suess, pp. 313-323, University of California Press, Berkeley, 1979.
Ostlund, H. G, TTO test cruise--Radiocarbon and tritium results, Data Rel., 81-23, 19 pp., Tritium Lab., Univ. of Miami, Miami, Fla.,1981.
Peng, T.-H., W. S. Broecker, G. G. Mathieu, and Y. H. Li, Radon evasion rate in the Atlantic and Pacific oceans as determined during the GEOSECS program, J. Geophys. Res., 84, 2471-2486, 1979.
Roether, W., Field measurements of gas exchange, in Dynamic Processes in the Chemistry of the Upper Ocean, edited by J. D. Burton, pp. 117-128, Plenum, New York, 1986.
Sarmiento, J. L, A tritium box model of the North Atlantic thermocline, J. Phys. Oceanogr., 13, 1269-1274, 1983.
Smethie, W. M., Jr., T. Takahashi, and D. W. Chipman, Gas exchange and CO_2 flux in the tropical Atlantic Ocean determined from ^{222}Rn and pCO_2 measurements, J. Geophys. Res., 90(C4), 7005-7022, 1985.
Stuiver, M., and H. Polach, Discussion of ^{14}C data, Radiocarbon, 19, 355-363, 1977.
Toggweiler, J. R, A six zone regionalized model for bomb radiotracers and CO_2 in the upper kilometer of the Pacific Ocean, Ph.D. thesis, Columbia Univ., New York, 1983.
Top, Z., TTO/NAS C14 and tritium, odd and ends, Data Rel., 84-34, 53 pp., Tritium Lab., Univ. of Miami, Miami, Fla., 1984.
Venrick, E. L., J. A. McGowan, D. R. Cayan, and T. L. Hayward, Climate and chlorophyll a: Long-term trends in the central North Pacific Ocean, Science, 238, 70-72, 1987.
Worthington, L. V., On the North Atlantic circulation, in The John Hopkins Oceanographic Studies, Johns Hopkins University Press, Baltimore, Md., 1976.

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