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THE DISTRIBUTION OF RADIOCARBON IN THE GLACIAL OCEAN

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Abstract. Accelerator mass spectrometric radiocarbon measurements on benthic foraminifera shells, picked from samples on which concordant ages were obtained on the shells of two species of planktonic foraminifera, reveal that the age of deep water in the equatorial Atlantic during glacial time was 675±80 years (compared to today's age of 350 years) and that the age of deep water in the South China Sea was 1670±105 years (compared to today's value of 1600 years). These results demonstrate that the 1.3 to 1.5 times higher radiocarbon content of carbon in glacial surface waters of the Caribbean Sea reconstructed by Bard et al. [1990] was primarily the result of a higher global inventory of radiocarbon rather than a decrease in rate of mixing between surface and deep waters of the ocean. The results are also consistent with the conclusion by Boyle and Keigwin [1987] that the flow of North Atlantic Deep Water was considerably weakened during glacial time, allowing deep waters of Antarctic origin to push much further north into the Atlantic than they do today.

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INTRODUCTION

During peak glacial time 20,000 to 14,000 years ago, the Earth was a very different place. Not only was it colder and more ice covered, but it was dustier and poorer in greenhouse gases. About 14,000 years ago a change occurred which created conditions more akin to those of today. Broecker and Denton [1989] postulate that this change was brought about by a reorganization of the entire ocean-atmosphere system. If so, then it is important to learn as much as possible about the manner in which the system operated during glacial time. Fortunately, a wealth of information regarding the patterns and rates of large-scale circulation in the ocean is available in deep-sea sediments.

To date the most definitive information in this regard comes from cadmium to calcium ratios in shells of benthic foraminifera [Boyle and Keigwin, 1987; Boyle, 1988a]. These results reveal that during glacial time the pattern of circulation was much different than today's. The nutrient constituent maxima which in today's ocean lie at intermediate depth were shifted toward the bottom. The strong contrast between the nutrient content of deep water in the Atlantic and Pacific observed in today's ocean was smaller during glacial time.

Although not a recorder of any specific change in ocean operation, the lower CO_2 content of the glacial atmosphere can only be explained through a major change in the interaction between the sea's mixing

TABLE 1. Holocene Age Measurement Sets

Depth, cm	Planktonic* Age, Years		Benthic Age, Years	ΔBenth-Plank, Years
Vema 28-122, Caribbean Sea; 11° 56'N, 7° 41'W, 3623 m; Today's Deep Water Age, 300 Years				
1-3	G sacc	2930±120	3280±140	
	G rub	3040±130		
	Mean	2980±90	3280±140	300±165
Sonne 50-37KL, South China Sea; 18°54'N, 115°46'E, 2695 m; Today's Deep Water Age, 1600 Years				
8-13	G sacc	2040±70	3970±80	
	P obl	2430±60		
	Mean	2260±190	3970±80	1710±205
40-45	G sacc	5960±80	8270±100	
	P obl	6510±80		
	Mean	6235±275	8270±100	2035±290

*The species names are: Globigerinoides sacculifer, Globigerinoides ruber, Pulleniatina obliquiloculata, and Globoquadrina dutertrei.

TABLE 2. Glacial Age Measurement Sets Yielding Concordant Planktonic Ages

Depth, cm	Planktonic Age, Years		Benthic Age, Years	ΔBenth-Plank, Years
Knorr 110-82GGC Equatorial Atlantic Ceara Rise; 4° 20'N, 43° 29'W, 2816 m				
25-28	G sacc	14150±160	14930±200	
	G rub	13870±260		
	N dut	13860±140		
	Mean	13970±100	14930±200	960±225
30-33	G sacc	15100±250	16350±280	
	G sacc	15080±120		
	G rub	15450±260		
	G rub	15130±120		
	N dut	15170±260		
	N dut	14820±120		
	Mean	15050±75	16350±280	1300±290
35-38	G sacc	16090±320	16130±240	
	G rub	15870±290		
	N dut	16060±200		
	Mean	16020±145	16130±240	110±280
40-43	G sacc	16710±250	17870±370	
	G rub	17040±250		
	N dut	17610±280		
	Mean	17085±255	17870±370	785±450
45-48	G sacc	17780±360	17900±640	
	G rub	17430±430		
	N dut	17660±260		
	Mean	17650±190	17900±640	250±665
Knorr 110-66GGC Equatorial Atlantic Ceara Rise; 4° 34'N, 43° 23'W, 3547 m				
45-50	G sacc	16450±150	17030±150	
	N dut	16660±150		
	Mean	16555±105	17030±150	475±185

TABLE 2. (continued)

Depth, cm		Planktonic Age, Years	Benthic Age, Years	ΔBenth-Plank, Years
Knorr 110-66GGC Equatorial Atlantic Ceara Rise; 4° 34'N, 43° 23'W, 3547 m (continued)				
49-53	G sacc	16800±150	17690±150	
	N dut	17030±150		
	Mean	16945±115	17690±150	775±190
53-58	G sacc	18130±160	18630±180	
	N dut	18060±180		
	Mean	18100±120	18630±180	535±215
56-61	G sacc	18860±190	19540±220	
	N dut	18980±190		
	Mean	18920±135	19540±220	620±260
Knorr 110-50GGC, Equatorial Atlantic Ceara Rise; 4° 52'N, 43° 13'W, 3995 m				
20-23	G sacc	13850±130	14760±140	
	N dut	14500±200		
	Mean	14045±295	14760±140	715±325
22-27	G sacc	16900±170	17110±190	
	N dut	16740±210		
	Mean	16835±130	17110±190	275±230
27-32	G sacc	17350±170	18360±200	
	N dut	17540±160		
	Mean	17450±115	18360±200	910±230
31-36	G sacc	18210±190	19110±190	
	N dut	18620±200		
	Mean	18405±205	19110±190	705±280
35-40	G sacc	19470±180	20360±220	
	N dut	18980±160		
	Mean	19195±245	20360±220	1165±330
Vema 28-122, Caribbean Sea; 11° 56'N, 78° 41'W, 3623 m				
115-123	G sacc	17120±150	17300±150	
	G rub	16980±140		
	Mean	17045±100	17300±150	255±180
123-125	G sacc	17390±160	17610±180	
	G rub	17680±170		
	Mean	17525±145	17610±180	85±230
129-139	G sacc	17910±400	18530±420	
	G rub	18730±480		
	Mean	18245±405	18530±420	285±585
Sonne 50-37KL, South China Sea; 18° 54'N, 115° 46'E, 2695 m				
160-165	G sacc	15140±150	17100±200	
	P obl	15300±150		
	Mean	15220±105	17100±200	1880±225
175-180	G sacc	15910±110	17430±140	
	P obl	15890±120		
	Mean	15900±80	17430±140	1530±160
195-200	G sacc	17460±160	18940±160	
	P obl	17270±150		
	Mean	17360±110	18940±160	1580±195
205-210	G sacc	17660±180	19445±190	
	P obl	17225±190		
	Mean	17455±115	19445±190	1990±285

cycles and biological cycles [Sarmiento and Toggweiler, 1984; Knox and McElroy, 1984; Siegenthaler and Wenk, 1984; Boyle, 1988b; Broecker and Peng, 1989].

A property of importance to the understanding of today's rate of deep-sea ventilation, i.e., the $^{14}C/C$ difference between surface and deep water, can now be reconstructed for glacial time. Radiocarbon measurements by the accelerator mass spectrometric (AMS) method on shells of foraminifera handpicked from deep-sea sediments offer the possibility that this reconstruction can be accomplished for the last 20 thousand years or so [Broecker et al., 1984]. The idea is that the ratio of the $^{14}C/C$ for benthic foraminifera to that for planktonic foraminifera coexisting in deep-sea sediments does not change with time during the glacial period. Hence radiocarbon measurements on coexisting benthic and planktonic foraminifera shells provide a measure of the $^{14}C/C$ difference between surface water and water at the depth from which the core was taken. For convenience this difference is expressed as an age (i.e., the time required for the surface water $^{14}C/C$ ratio to decay to that for deep water). In today's tropical Pacific this age difference is about 1600 years, while in the tropical Atlantic it is about 350 years.

Using this strategy Andree et al. [1986] showed that over the course of Holocene time the age difference for waters at 2 km depth in the western Pacific was the same as today's to within the measurement uncertainty (i.e., ~200 years). Based on measurements on a core from the Ceara Rise, Broecker et al. [1988a] suggested that the age of deep water in the western tropical Atlantic was roughly twice as great during glacial time than it is today. Shackleton et al. [1988] and Broecker et al. [1988a] obtained results suggesting that the age of deep water in the tropical Pacific was somewhat greater during glacial time (~2000 years) than it is today (~1600 years). These latter results are however far from conclusive. The Shackleton et al. [1988] results come from an area where upwelling currently influences the $^{14}C/C$ ratio in the photic zone. Because of this, the planktonic results do not necessarily reflect conditions typical of the surface ocean. Further, they present only two results from peak stage 2 time (i.e., 20,000 to 14,000 years ago). Both suggest a lower age difference than those before 20,000 years ago.. Two of the three cores reported by Broecker et al. [1988a] come from areas of such low accumulation rate that the bioturbation-abundance couple [see Broecker et al., 1984] likely introduces biases in the age difference. The third core analyzed by Broecker et al. [1988a] showed a systematic (and unexplained) discordance between the ages obtained on two planktonic species, G. sacculifera and P. obliquiloculata.

Broecker [1989] pointed out that the smaller surface to deep $^{14}C/C$ ratio difference for the glacial Atlantic could be explained either by a slowdown of the Atlantic's conveyor or by an inversion of the Atlantic's circulation. However, with an inverted glacial circulation he was not able to account for Boyle's [1988a] observation that although the interocean nutrient content difference was smaller during glacial time, waters in the deep Atlantic still had a smaller nutrient content than those in the deep Pacific.

In this paper we present new radiocarbon results which broadly support the conclusions reached in previous papers and which overcome the uncertainties associated with the previous results from the tropical Pacific.

RESULTS

We use as a criterion for validity of each benthic-planktonic age difference measurement that agreement exist (at the 2σ level) between the ages obtained on two separate species of planktonic foraminifera. The logic behind this strategy is that neither the averaging of discordant planktonic results nor the arbitrary selection of one result over the other is likely to produce a reliable estimate of the benthic-planktonic age difference.

To date we have results which meet this criterion from five deep-sea cores, one from the South China Sea (representing 2.1-km depth water in the western tropical Pacific), one from the Caribbean (representing 1.8-km depth water from the western tropical Atlantic) and three from the Ceara Rise (representing 2.8-km to 4.0-km depth water in the western tropical Atlantic). The measurements representing Holocene time are listed in Table 1. Those representing glacial time are listed in Table 2. The age differences obtained for glacial time from those five cores are summarized in Table 3. A complete listing of results from these cores (including abundance data) have been published by Broecker et al. [1988b, 1990a].

For the tropical Pacific we see no discernable change in the surface to deep $^{14}C/C$ difference between glacial and Holocene time. While the mean age differences for the Holocene ocean (1820±165 years) and for the glacial ocean (1670±105 years) are slightly larger than that for today's ocean (1600 years), more benthic-planktonic measurements will have to be obtained before these small differences can be confirmed.

In the case of the Atlantic we think that we now have enough measurements in the 2.8-4.0 km depth range to say with some certainty that the age of North Atlantic Deep Water (NADW) in the western equatorial Atlantic was greater during glacial time than it is today. As shown in Figures 1 to 3 and summarized in Figure 4, the age of Atlantic deep water averaged 675±80 years during peak glacial time as opposed to 350 years today. As shown in Figures 1 to 3, the $^{13}C/C$ ratio for benthic forams was lower (by 0.8±0.2‰) during glacial time than during Holocene time at the sites of these cores.

The core from the Caribbean is representative of water at about 1.8 km depth at about 15°N in the

TABLE 3. Summary of Age Estimates for Glacial Deep Water (i.e., Benthic-Planktonic Age Differences)

Location	Core	Latitude	Longitude	Water depth, km	Number of Pairs	Mean Benth-Plank Age, Years	Today's Age, Years
Atlantic	V28-122	12°N	79°W	1.8*	3	195±140	300
Atlantic	K110-82	4°N	43°W	2.8	5	780±220	350
Atlantic	K110-66	5°N	43°W	3.5	4	600±105	350
Atlantic	K110-50	5°N	43°W	4.0	5	705±150	400
Pacific	S50-37	19°N	116°E	2.1	4	1670±105	1600

*Sill depth of Caribbean.

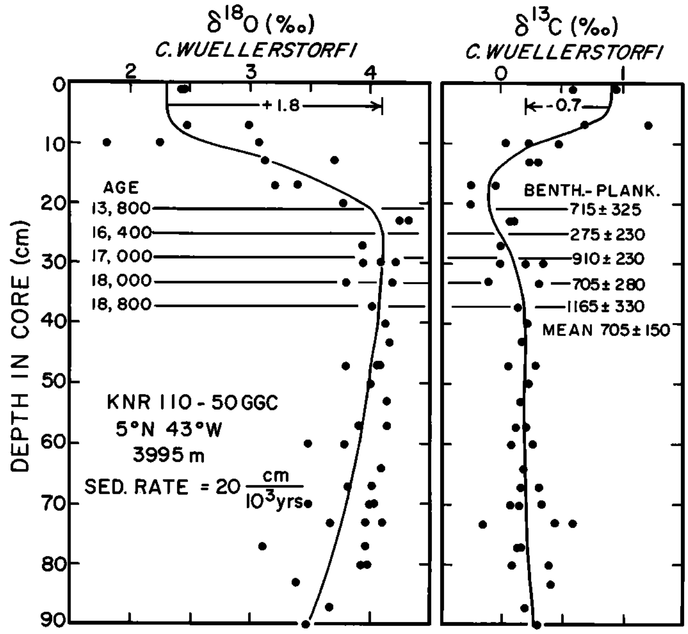

Fig. 1. Carbon and oxygen isotope composition of benthic foraminifera shells as a function of depth in western equatorial Atlantic core KNR110-50GGC [Curry et al., 1988]. Also shown are the levels at which planktonic-benthic age differences were determined. At the left the age of the horizon is given (planktonic mean - 400 years). At the right are the differences between the benthic age and the mean planktonic age.

western Atlantic. This is the depth and latitude of sill for this isolated basin. The Caribbean is currently flushed too rapidly for significant aging [Ribbat et al., 1976]. Hence, analysis on this core should be representative of Boyle's low cadmium content glacial intermediate depth water in the open Atlantic. Indeed the isotope record for benthic foraminifera from this core shows that the $^{13}C/C$ ratio was higher during glacial than during Holocene time (see Figure 5). As can be seen, the age differences for glacial age samples are slightly smaller than today's.

The relationship between the Holocene-glacial $\delta^{13}C$ change recorded in benthics and the planktonic-benthic age difference for glacial time is shown in Figure 6.

DISTRIBUTION OF RADIOCARBON IN THE GLACIAL OCEAN

An attempt at reconstructing the distribution of radiocarbon in the glacial ocean is shown in Figure 7. Based on measurements on corals recovered off Barbados by Fairbanks [1989], Bard et al. [1990] show that during peak glacial time a discrepancy of about 3200 years exists between the radiocarbon age and the radiothorium age scale. For example, a coral

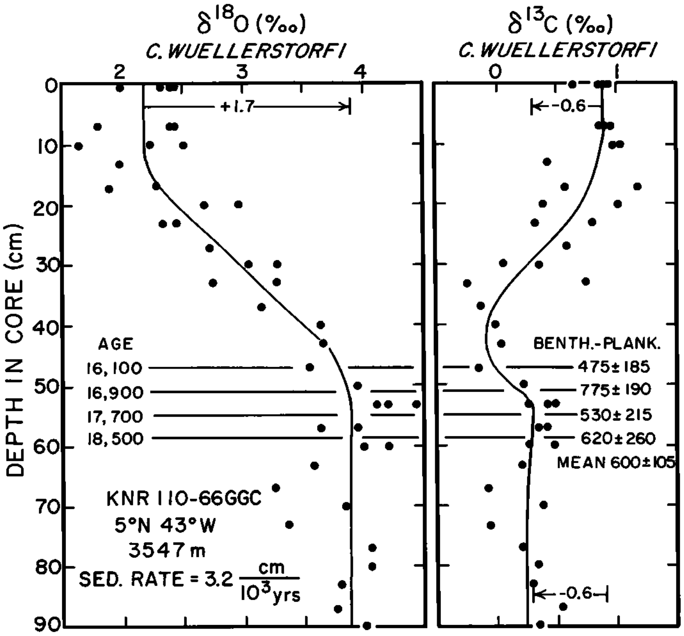

Fig. 2. Carbon and oxygen isotope composition of benthic foraminifera shells as a function of depth in western equatorial Atlantic core KNR110-66GGC [Curry et al., 1988]. Also shown are the levels at which planktonic-benthic age differences were determined. At the left the age of the horizon is given (planktonic mean - 400 years). At the right are the differences between the benthic age and the mean planktonic age.

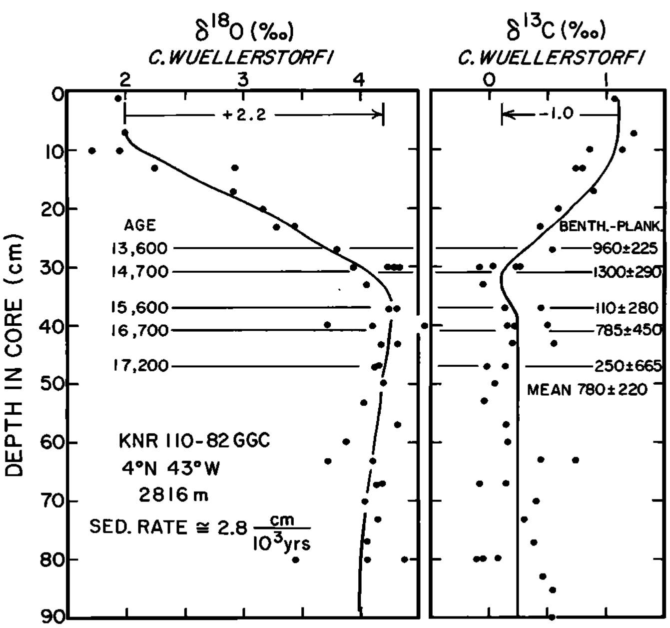

Fig. 3. Carbon and oxygen isotope composition of benthic foraminifera shells as a function of depth in western equatorial Atlantic core KNR110-82GGC [Curry et al., 1988]. Also shown are the levels at which planktonic-benthic age differences were determined. At the left the age of the horizon is given (planktonic mean - 400 years). At the right are the differences between the benthic age and the mean planktonic age.

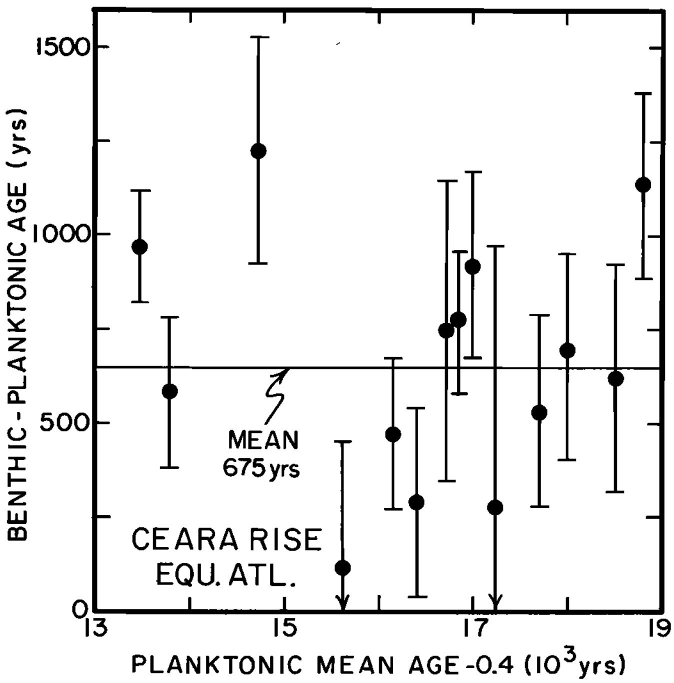

Fig. 4. Benthic-planktonic age differences versus horizon age for the three cores from the western basin of the equatorial Atlantic. The mean for these 14 determinations is 675±80 years.

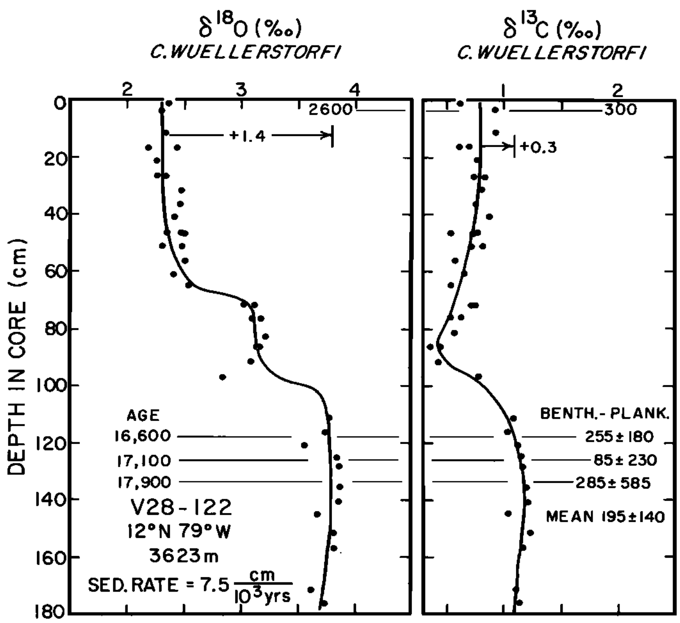

Fig. 5. Carbon and oxygen isotope records for core V28-122 from the Caribbean Sea [Oppo and Fairbanks, 1987]. Also shown are the levels at which benthic-planktonic age differences were determined. At the left are the horizon ages (i.e., mean planktonic age - 400 years). At the right are the age differences.

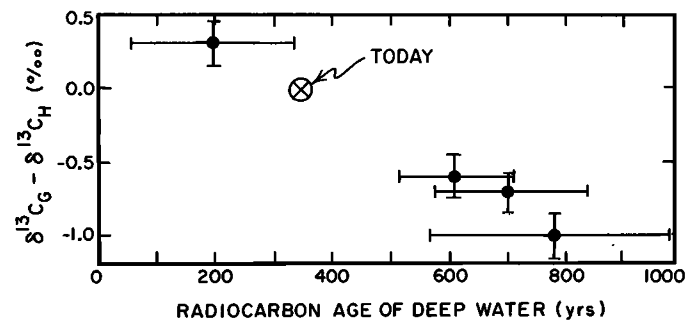

Fig. 6. Plot of the $\delta^{13}C$ difference between peak glacial and Holocene time against the mean age difference between planktonic and benthic foraminifera for the four Atlantic cores. The three for the Ceara Rise fall in the lower right quadrant, and the Caribbean core falls in the upper left.

yielding ^{14}C ages of 18,700 years yields a ^{230}Th age of 21,900 years. Bard et al. [1990] make a strong case that this difference reflects a 1.5 times higher Earth ^{14}C inventory during peak glacial time. The higher ^{14}C production rate required to maintain this inventory presumably reflects a lower Earth magnetic field. We accept this result and therefore place the atmospheric $^{14}C/C$ ratio for peak glacial time at 1.5 times the 1850 A.D. value.

At steady state the atmosphere to surface ocean $^{14}C/C$ ratio difference must be of the right magnitude in order that air-sea CO_2 exchange carries radiocarbon into the sea as fast as it decays within the sea. The delivery rate depends not only on the air-sea $^{14}C/C$ difference but also on the CO_2 partial pressure in the atmosphere and the wind velocity over the sea. For glacial time we know from ice core studies [Raynaud et al., 1988; Neftel et al., 1988] that the atmospheric CO_2 pressure was about 200 μatm as opposed to 280 in 1850 A.D. While the mean wind velocity over the ocean may have been different, we have no means to reconstruct even the sign of this difference. Hence we are forced to assume it to be similar to today's. While temperature influences both the CO_2 concentration in seawater and the diffusivity of CO_2 in seawater, these factors largely compensate one another with respect to CO_2 exchange [see Broecker and Peng, 1982]. Hence, the glacial to interglacial difference in sea surface temperature need not be considered. Hence

$$\left[1-\frac{^{14}C/C_{surf}}{^{14}C/C_{atm}}\right]_{glacial}\cong\frac{280}{200}\left[1-\frac{^{14}C/C_{surf}}{^{14}C/C_{atm}}\right]_{1850}$$

We are also forced to assume that the geographic pattern of $^{14}C/C$ ratios in the glacial ocean was similar to today's. In the 1850 A.D. tropical ocean the $^{14}C/C$ ratio for surface water was about 0.95 the atmospheric value [Broecker and Peng, 1982]. Based on the above assumptions the $^{14}C/C$ ratio in the glacial tropical surface ocean water is calculated to be 0.93 that in the glacial atmosphere. Note that this value is independent of the absolute ^{14}C inventory adopted.

We then use the measurements presented in this paper to establish the ratio of the $^{14}C/C$ for deep waters to that for surface waters in the tropical Atlantic and in the tropical Pacific. These ratios are used to fix the position of the tropical deep waters in Figure 7.

As concluded by Bard et al. [1990], the change in surface ocean $^{14}C/C$ ratio required by their $^{230}Th/^{234}U$ measurements on corals must be primarily the result of a change in the inventory of radiocarbon in the ocean-atmosphere system rather than a change in the distribution of radiocarbon among the reservoirs. The latter explanation would require that during peak glacial time the age of deep water in the ocean averaged about 3500 years.

IMPLICATIONS FOR GLACIAL CIRCULATION

In order to gain some insight as to how the deep-water radiocarbon ages depend on mixing rates, we employ our Pandora "something like the real ocean geochemical model" [Broecker and Peng, 1986, and 1987]. For this exercise we have converted the interbox transfer fluxes used previously into seven circulation loops (see Figure 8). In this way we can alter the magnitude of any one of the loops without having to concern ourselves with conserving water, for in this loop scheme the water fluxes are automatically balanced. In addition to the seven loops shown in Figure 8, an eighth connects the southern thermocline reservoir of the Atlantic (box 3) with the southern thermocline reservoir of the Pacific-Indian (box 7). We envision a rapid mixing between these reservoirs (via circumpolar flow) and therefore maintain a sufficiently high flow via this eighth loop to maintain nearly equal compositions in these two reservoirs. A seemingly curious feature of our loop structure is the branching that occurs in the upper ocean. A fraction, f for the Atlantic and g for the Pacific, traverses the mixed layer, and the remainder, (1-f) for the Atlantic and (1-g) for the Pacific, moves through the thermo-

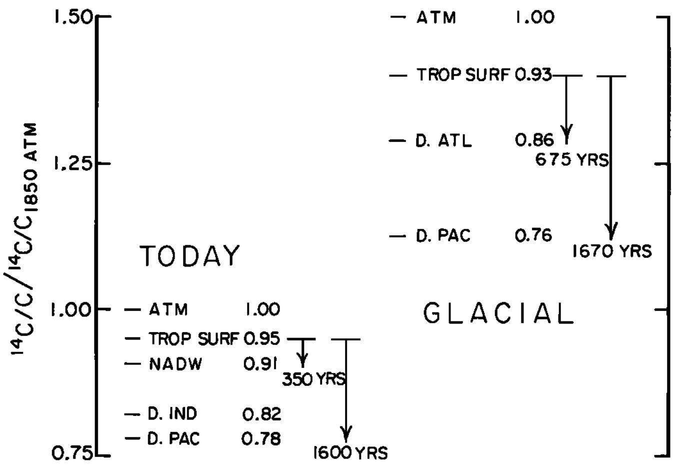

Fig. 7. Reconstruction of the radiocarbon content of various parts of the ocean-atmosphere system for glacial time. The estimate for tropical surface water is based on the ^{230}Th-^{14}C age comparisons on corals by Bard et al. [1990]. The difference in ^{14}C/C ratio for tropical surface water and the overlying atmosphere takes into account the lower CO_2 content of the glacial atmosphere. The deep Atlantic value is for the site of the Ceara Rise cores (i.e., equatorial western basin). The deep Pacific value is for the South China Sea. The numbers within the diagram are the ratios of the ^{14}C/C in the water of interest to that in the glacial atmosphere.

cline. This scheme proves to be an effective means of reproducing the observed properties of today's surface ocean. In Table 4 we list the fluxes adopted in order to replicate the Holocene ocean and the resulting distribution of properties.

Our first attempt to duplicate the observed glacial distribution of radiocarbon and nutrients involves a slowing down of the three loops (1, 2 and 3), which constitute the flow of NADW (see Figure 9). In order to maintain a nearly uniform ^{14}C/C difference between the surface Pacific-Indian box and the deep Pacific-Indian box, decreases in the strength of loop 1 are matched by corresponding increases in the strength of loop 6 (see Table 5). By cutting the strength of the NADW loops we bring about an increase in the surface to deep ^{14}C/C difference for the Atlantic Ocean (i.e., an increase in the deepwater age). By contrast, the phosphate content of deep water in the Atlantic does not change significantly with the NADW flux. To understand this, one must recall that the residence time of phosphorus in the ocean is many times longer than the oceanic mixing time. Hence, no significant gain or loss of this nutrient can occur during a single pass through the Atlantic. Since, at least in our model, only NADW is exported, the phosphate content of NADW must be equal to the average in the water entering the Atlantic. As in our model the mix of incoming water is nearly independent of the flux of water through the Atlantic, very little change in the phosphate content of Atlantic deep water occurs as the rate of NADW production changes. Thus, this simple exercise reveals an important truth: radiocarbon and nutrient distributions in the sea are not so tightly tied one to the other as one might expect. Hence, the information gained from radiocarbon measurements on foraminifera is not necessarily redundant to that obtained from cadmium, barium or ^{13}C/^{12}C measurements on foraminifera.

A second means of increasing the surface to deep radiocarbon difference in the model's Atlantic and one which should also increase the nutrient content of deep water in the Atlantic is to strengthen loop 4 (equivalent to the input of Antarctic Bottom Water into the Atlantic). However, as listed in Table 6, even a very large increase in this flux (i.e., from 4 to 20 sverdrups (Sv)) has only a small influence on both the age and nutrient content of deep Atlantic water.

This exercise in modeling has led us to do some thinking about what sets the nutrient content of Atlantic deep waters. As described in a separate paper (W. S. Broecker and T.-H. Peng, Factors controlling the distribution of phosphate in the deep ocean, submitted to Global Biogeochemical Cycles, 1990), while the sense of the deep Atlantic-deep Pacific nutrient difference is set by the existence of the global conveyor, the magnitude of the interocean difference depends on the fraction of the return water which comes via Gordon's [1986] Agulhas retroflection (poor in nutrients) as opposed to the fraction which is supplied by Antarctic surface, intermediate and bottom water (rich in nutrients). Hence, the important factor in determining the interocean nutrient is the pattern of return flow rather than the rate of exchange.

Another way to look at the problem is to consider the deep ocean as a mixture of two end-members, new deep water formed in the northern Atlantic and old deep water in the Pacific. Raymo et al. [1990]

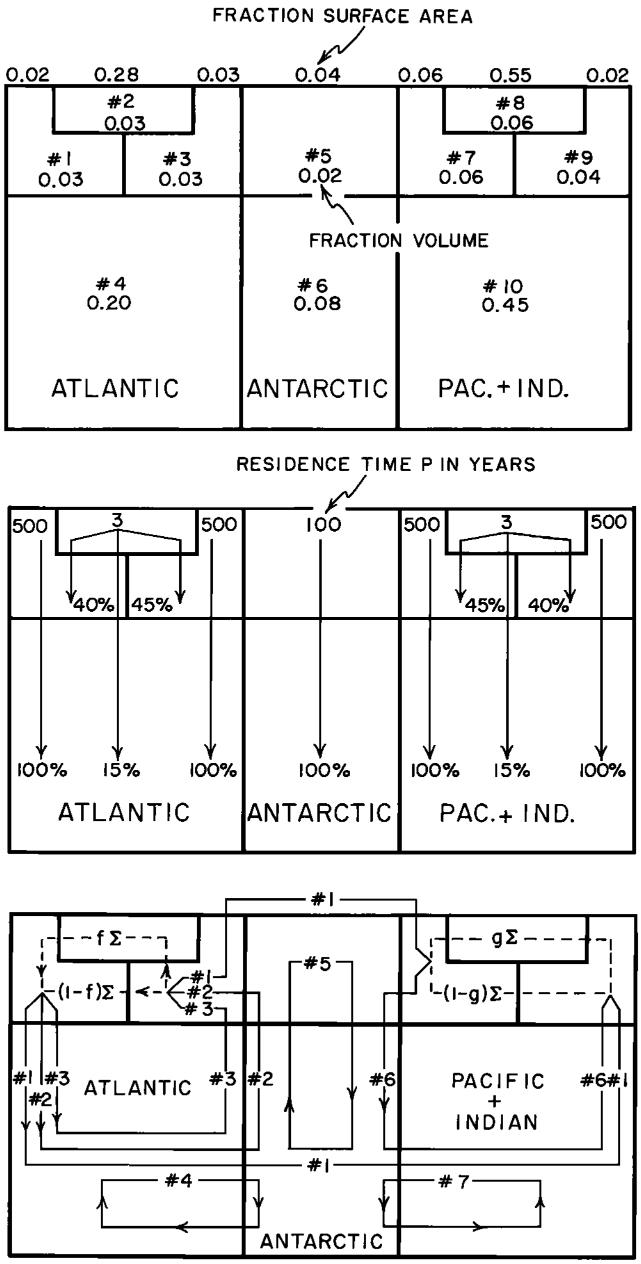

Fig. 8. A version of Pandora adopted for exploration of circulation patterns capable of reproducing the nutrient and radiocarbon distributions reconstructed for the glacial ocean. In the upper panel are given the volume and surface area fractions of the 10 reservoirs. In the middle panel are shown the residence times (in years) for phosphate in each of the reservoirs receiving sunlight. Also shown is the fate of the phosphorus atoms removed from surface water in particulate form. In the lower panel are shown the circulation loops.

have shown that over the last 3 million years the carbon isotope ratio in the deep western Atlantic has swung back and forth between the compositions of these end-members, giving the impression that the geometry of the zone of mixing has ranged from today's situation where NADW floods most of the deep Atlantic to a situation where Antarctic waters penetrate far up the Atlantic. This situation is not amenable to modeling with fixed geometry boxes. Rather, a scheme involving horizontal advection-diffusion is required. Deep water produced in the Atlantic is advected into a diffusive deep sea. In such an ocean the

TABLE 4. Pandora Holocene Standard Configuration

Parameter Assignment	Reservoir No.	Reservoir Identity	pCO_2, µatm	PO_4, µmol/kg	$CO_3^=$, µmol/kg	$\delta^{13}C$, ‰	$\Delta^{14}C$, ‰	Age Deepwater, Years
f=0.65	-	atmosphere	311	-	-	-6.4	-14	-
g=0.65	1	No. Atl. therm	282	0.86	119	2.2	-77	-
L1=6.0 Sv	2	Atl. surface	308	0.06	193	2.8	-57	-
L2=13.5 Sv	3	So. Atl. therm	338	1.40	108	2.1	-104	-
L3=10.5 Sv	4	Atl. deep	327	1.20	110	1.8	-111	500
L4=4.0 Sv	5	Ant. surface	303	1.39	108	2.2	-117	-
L5=3.0 Sv	6	Ant. deep	410	1.80	95	1.3	-146	-
L6=15.0 Sv	7	So. Pac.-Ind. therm	325	1.34	111	2.1	-104	-
L7=20.0 Sv	8	Pac.-Ind. surface	309	0.03	198	2.5	-54	-
L8=100 Sv	9	No. Pac. therm	369	1.92	125	1.8	-169	-
	10	Pac.-Ind. deep	576	2.58	82	0.6	-185	1240

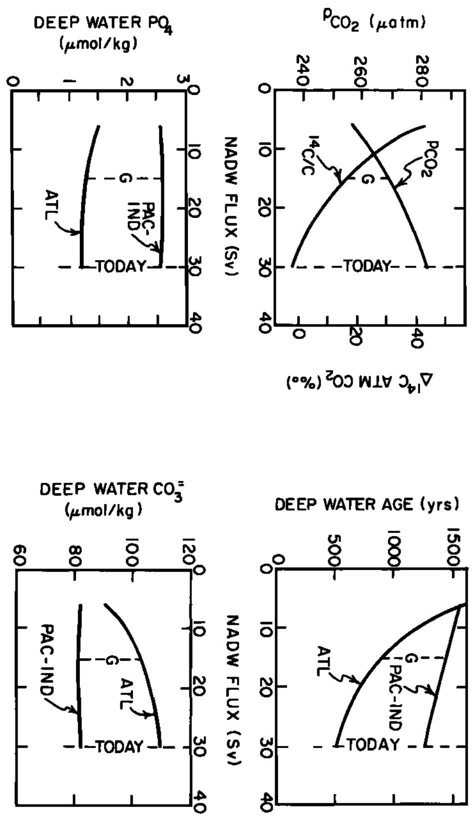

Fig. 9. Response of Pandora to changes in the flux of the combined NADW loops (i.e., 1, 2 and 3) from 30 Sv for the Holocene standard to values as low as 6 Sv. Note that in order to keep the age of deep water in the Indian-Pacific nearly constant, reductions in the strength of loop 1 are matched by increases in the strength of loop 6 (see Table 5). The dashed line labeled "G" marks the flux at which the age of deep water in the Atlantic is reduced by a factor of 2.

^{13}C and ^{14}C distributions in the sea would change sympathetically as the advective strength of NADW changed. As shown in Figure 10, a strong concordance exists in the trends for the two isotopes in today's ocean.

Could this same situation have applied in the glacial ocean? In an attempt to answer this question we combine the glacial to Holocene offset in ^{13}C measured in benthic forams with the ^{14}C results reported here. In order to remove the impact of any change in the ^{14}C inventory in the ocean-atmosphere system we reference the $^{14}C/C$ values for deep water to those for tropical surface water. The results of our reconstruction are shown in Figure 11. As can be seen, the reconstructed points for the Ceara Rise and for the South China Sea suggest that the slope of the ^{13}C to ^{14}C relationship remained unchanged. The 0.4‰ offset between the glacial and the Holocene trend lines is attributed to a change in the $^{13}C/^{12}C$ ratio for whole ocean carbon during glacial time resulting from lower forest biomass and lower soil humus inventories during glacial time [Shackleton, 1977; Curry et al., 1988]. Hence it is possible that the major difference between the glacial and Holocene oceans is that the strength of NADW was much reduced during glacial times, allowing radiocarbon deficient waters from the Antarctic to penetrate further into the Atlantic.

One further complication must be considered. The $^{18}O/^{16}O$ records for the Greenland ice cores suggest that millennium-long events punctuated the climate of the northern Atlantic region during much of glacial time [Dansgaard et al., 1982]. A likely cause for these oscillations is the turning "on" and "off" of deepwater production in the Atlantic [Broecker et al., 1985, 1990b]. If such alternations were occurring during the time period covered by our measurements (i.e., 13,000 to 19,000 years ago) then because of bioturbation we would obtain the average of the age of Atlantic deep water during the "on" and "off" parts

TABLE 5. Transports in Sverdrups Used in Calculations Yielding the Results Summarized in Figure 9

Water Route	Holocene Standard	NADW x0.8	NADW x0.6	NADW x0.4	NADW x0.2
Loop 1	6.0	4.8	3.6	2.4	1.2
Loop 2	13.5	10.8	8.1	5.4	2.7
Loop 3	10.5	8.4	6.3	4.2	2.1
Σ Loops 1,2,3	30.0	24.0	18.0	12.0	6.0
Loop 4	4.0	4.0	4.0	4.0	4.0
Loop 5	3.0	3.0	3.0	3.0	3.0
Loop 6	15.0	16.2	17.4	18.6	19.8
Loop 7	20.0	20.0	20.0	20.0	20.0
Loop 8	100.0	100.0	100.0	100.0	100.0

TABLE 6. Sensitivity to Flux of Antarctic Bottom Water (i.e., Loop 4) Into the Deep Atlantic

Flux Loop 4, Sverdrups	Atmospheric pCO_2, μatm	PO_4 Deep Atlantic, μmol/kg	PO_4 Deep Pacific and Indian, μmol/kg	Age Deep Atlantic, Years	Age Deep Pacific and Indian, Years
		NADW = Holocene Standard (i.e., 30 Sv)			
4	311	1.20	2.58	500	1240
8	311	1.26	2.55	540	1230
12	311	1.31	2.53	560	1225
16	311	1.34	2.52	580	1225
		NADW = 0.6 Holocene Standard (i.e., 18 Sv)			
4	302	1.23	2.60	770	1350
10	302	1.35	2.56	820	1340
20	301	1.45	2.52	880	1330
		NADW = 0.4 Holocene Standard (i.e., 12 Sv)			
4	296	1.31	2.60	1030	1420
10	295	1.44	2.55	1080	1420
20	295	1.53	2.51	1120	1410

of the cycle. For example, if the age during the "on" part of the cycle were 350 years as during Holocene time, then the age during the "off" part of the cycle would have to be considerably greater than 670 years. Indeed, since the time for a half cycle is about 1000 years, it is possible that a steady state of radiocarbon distribution was not achieved.

CONCLUSIONS

The results presented here suggest that the ratio of the radiocarbon age of deep water in the Pacific Ocean to the radiocarbon age of deep water in the Atlantic Ocean was smaller during glacial time than today. Taken together with the nutrient distribution reconstruction for glacial time, this result suggests that the mixing zone between North Atlantic Deep Water and waters in the Antarctic moved well northward from its present position at 30° to 40°S in the Atlantic.

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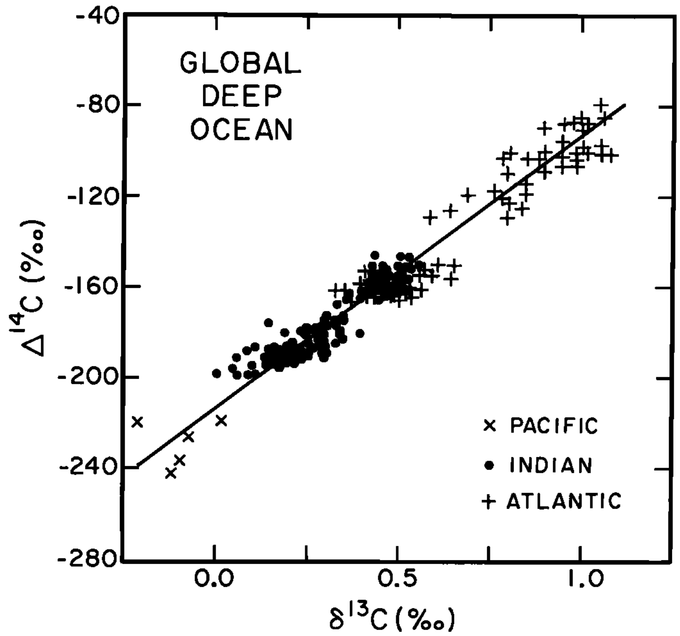

Fig. 10. Plot of $\Delta^{14}C$ against $\delta^{13}C$ for deep water (> 2 km) from throughout the world ocean. The ^{13}C and ^{14}C data for the Indian and Atlantic Oceans are from Ostlund et al. [1987]. Those for the Pacific are from [Kroopnick et al., 1970; Ostlund and Niskin, 1970].

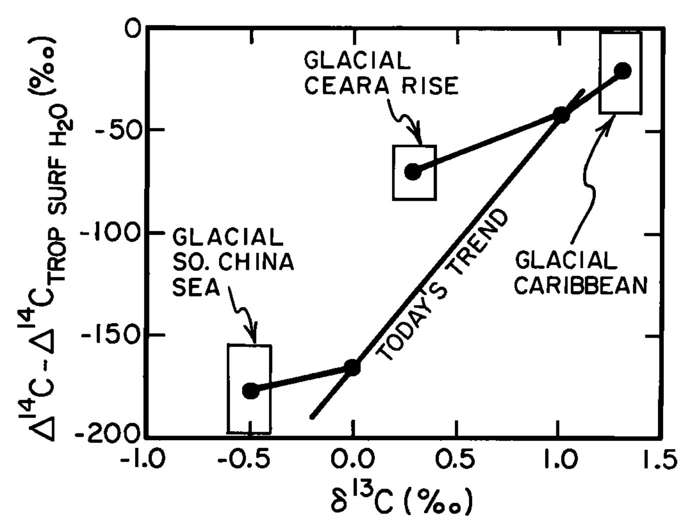

Fig. 11. Attempt to reconstruct the $^{14}C/C$ versus $^{13}C/C$ trend in the glacial deep ocean. The $\delta^{13}C$ offset for the deep Pacific is that observed in benthic foraminifera for the South China Sea [Oppo and Fairbanks, 1987]. That for the deep Atlantic is the 0.8‰ average observed for the three Ceara Rise cores [Curry et al., 1988]. In the case of ^{14}C the tropical surface ocean is used as a reference. This eliminates the influence of changes in the inventory of ^{14}C in the ocean-atmosphere system and changes in the air-surface $^{14}C/C$ difference. It is assumed that during glacial time no significant difference existed between the $\Delta^{14}C$ value for tropical surface waters in the Atlantic and those in the Pacific.

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