species, and contributed to the discoloration and defoliation of trees. Attention should be paid to such early warning signs, bearing in mind the irretrievable value of forest growth potential. Past development, nonetheless, guarantees that during the next 10 to 20 years, wood resources are plentiful and can be allocated among traditional forest industries, eventual novel technologies such as ethanol production (31), and nature protection and conservation purposes.

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Oceanic Uptake of Fossil Fuel CO₂: Carbon-13 Evidence

P. D. Quay,* B. Tilbrook, C. S. Wong

The δ^{13} C value of the dissolved inorganic carbon in the surface waters of the Pacific Ocean has decreased by about 0.4 per mil between 1970 and 1990. This decrease has resulted from the uptake of atmospheric CO₂ derived from fossil fuel combustion and deforestation. The net amounts of CO₂ taken up by the oceans and released from the biosphere between 1970 and 1990 have been determined from the changes in three measured values: the concentration of atmospheric CO₂, the δ^{13} C of atmospheric CO₂ and the δ^{13} C value of dissolved inorganic carbon in the ocean. The calculated average net oceanic CO₂ uptake is 2.1 gigatons of carbon per year. This amount implies that the ocean is the dominant net sink for anthropogenically produced CO₂ and that there has been no significant net CO₂ released from the biosphere during the last 20 years.

The fate of CO_2 loaded into the atmosphere by fossil fuel combustion and deforestation has been uncertain. Direct measurements of the increase in atmospheric CO_2 levels, when compared to the rates of fossil fuel combustion and cement production, indicate that about 57 percent of the CO_2 produced from these sources has accumulated in the atmosphere (1). The other 43 percent of the industrially derived CO_2 is in either the biosphere or ocean, the two carbon reservoirs that can exchange signif-

SCIENCE • VOL. 256 • 3 APRIL 1992

icant amounts of CO_2 on the time scale of the anthropogenic perturbation. Two recent studies of the atmospheric CO_2 budget came to opposite conclusions with regard to whether the ocean or biosphere has been the predominant CO_2 sink (1, 2).

Our uncertainty about the fate of anthropogenically produced CO_2 results from the lack of direct measurements of changes in the carbon content of the biosphere and ocean. As a result we have had to rely on model parameterizations to estimate oceanic and biospheric CO_2 uptake rates (1-3). Because the uncertainty in these parameterizations is substantial, the question of whether the ocean or biosphere is the major CO_2 sink has remained unanswered.

In this article, we use measurements of the decrease in the ${}^{13}C/{}^{12}C$ ratio of the total dissolved inorganic carbon (DIC) in

<sup>P. Quay is with the School of Oceanography, University of Washington, Seattle, WA 98195.
B. Tilbrook is with the Commonwealth Scientific and Industrial Research Organization, Division of Oceanography, Hobart, Tasmania, 7001 Australia.
C. S. Wong is with the Centre for Ocean Climate Chemistry, Institute of Ocean Sciences, Sidney, British Columbia, Canada V8L 4B2.</sup>

^{*}To whom correspondence should be addressed.



Fig. 1. The location of the cruise tracks for the HUDSON-70 cruise in May 1970 (solid diamonds) and the three NOAA cruises in February and March 1989 (circles), March and April 1990 (squares), and March and April 1991 (triangles).

the ocean since 1970 to determine the rate of oceanic uptake of anthropogenically produced CO_2 . The ¹³C/¹²C ratio, expressed as $\delta^{13}C$ (4), is a useful tracer of CO_2 derived from fossil fuel and deforestation sources. The tracer quality of the δ^{13} C value results from a difference between the δ^{13} C value of CO_2 in the atmosphere (about -8 per mil) and that of CO₂ released during the combustion of fossil fuels and biomass (about -27 per mil). The difference between the δ^{13} C values for plant material and atmospheric CO₂ results from the isotopic fractionation during photosynthetic fixation of CO_2 (5). Although the $\delta^{13}C$ values of terrestrial plants range from -27 per mil to -12 per mil depending on whether the plant uses a three-carbon or four-carbon intermediate to fix CO_2 (5), that is, C_3 or C₄ plants, the terrestrial biomass is overwhelmingly (>95 percent) C3 plant material (6).

The δ^{13} C value of atmospheric CO₂ is decreasing because CO₂ input from fossil fuel combustion and deforestation has a δ^{13} C value about 20 per mil lower than the δ^{13} C value of atmospheric CO₂. The δ^{13} C value of atmospheric CO_2 has decreased by 1.4 per mil during the last 300 years on the basis of measurements of air bubbles trapped in ice (7). Direct measurements show that the δ^{13} C value of atmospheric CO₂ decreased from -7.5 per mil in 1978 to -7.8per mil in 1988 (1). Similarly, the δ^{13} C value of DIC in the ocean should have decreased if anthropogenically produced CO2 has entered the oceans. A decrease of about 0.5 per mil from 1900 to 1970 for the surface layer of the tropical ocean has been inferred from measurements of the $\delta^{13}C$ value of aragonite in a sclerosponge (8). In order to quantify the amount of anthropogenic CO_2 taken up by the oceans we must determine the depth-integrated change in



Fig. 2. The decrease of the δ^{13} C values of dissolved inorganic carbon in the mixed layer of the Pacific Ocean since 1970. Symbols correspond to those in Fig. 1.

the oceanic δ^{13} C on an oceanwide scale.

On three National Oceanographic and Atmospheric Administration research cruises during the last 3 years we measured the δ^{13} C value of DIC in the Pacific Ocean from 60°S to 60°N (Fig. 1). Seawater samples, poisoned with $HgCl_2$, were collected and stored in prewashed and baked (450°C) ground glass-stoppered bottles. In the laboratory, we extracted the CO_2 from seawater using a helium stripping technique (9). The efficiency of the extraction was 100 \pm 0.5 percent based on gravimetrically prepared Na₂CO₃ standards. The overall precision of the δ^{13} C analysis was ±0.02 per mil based on a replicate analysis of standards and seawater samples.

To recognize changes over decadal time scales we compared our $\delta^{13}C$ data with similar measurements made in 1970 and 1971. δ^{13} C values for the mixed layer in the Pacific Ocean in 1970 were measured during the Canadian HUDSON-70 cruise along 150°W between 65°S and 55°N (10). Depth profiles were measured in 1970 at eight stations during the HUDSON-70 cruise, five stations during the SCAN-X cruise along 90° to 114°W, five stations during the Tow-7 cruise along 165° to 171°W, and in 1971 at one station during the Antipodes-15 cruise along 172°W (11). We estimated the uncertainty in the δ^{13} C data from 1970 at ±0.04 per mil on the basis of reported replicate analyses (11).

Oceanic δ^{13} C decrease. Comparison of the data shows that the average δ^{13} C value of DIC in the mixed layer of the Pacific Ocean decreased by 0.4 per mil during the last 20 years (Fig. 2), if we assume that the values obtained in 1970 are representative of the latitudinal variations over the entire Pacific Ocean. This assumption is supported by the similarity in the δ^{13} C values measured along 110°W and 170°W in 1989 and 1990, respectively. Data from the two NOAA cruises in the southern ocean along 110°W and 170°W show average decreases of 0.33 and 0.36 per mil, respectively, and data from the northern ocean cruise along 150°W in 1991 shows a 0.45 per mil decrease. The average basin-wide decrease in δ^{13} C values for the mixed layer, calculated by weighting the latitudinal trends according to ocean surface area, is 0.40 per mil.

With the $\delta^{13}C$ data available for the water column (Fig. 3), we can compare the change in the oceanic depth distribution of δ^{13} C values in the Pacific ocean since 1970 at seven sites (12). The average depthintegrated δ^{13} C value has decreased by 203 per mil meters, with a range of 60 to 305 per mil meters (Table 1); the units represent the integration of the area between the δ^{13} C depth profiles measured in 1970 and 1989 to 1991. A comparison of δ^{13} C values measured at depths below 2000 m in 1970 and 1989 to 1991 (Fig. 3) indicates that there is no systematic offset in the data sets. The average difference between 14 pairs of samples collected at comparable locations and depths below 2000 m was $-0.01 \pm$ 0.06 per mil.

The largest decreases in mixed layer and depth-integrated δ^{13} C values occurred in the subtropical gyres; smaller decreases occurred in the equatorial and subpolar oceans (Fig. 4 and Table 1). These δ^{13} C trends are consistent with the increase in mixed layer and depth-integrated ¹⁴C activity (Fig. 4 and Table 2) resulting from nuclear weapons testing (13). The similarity of these carbon isotope changes is not surprising because both the ¹³C and ¹⁴C perturbations are input to the ocean by way of CO_2 gas exchange and affected by the same basin-wide patterns of upwelling and downwelling within the upper portion (<1000 m) of the Pacific Ocean. The

oceanic ¹³C and ¹⁴C changes are not likely to be completely correlated because the atmospheric source functions differ. The magnitude of the oceanic ¹³C and ¹⁴C changes indicate where these anthropogenically derived carbon tracers have accumulated and not, necessarily, where they have entered the ocean. For example, the small changes in the mixed layer ¹³C and ¹⁴C south of 50°S (Fig. 4) do not imply that there has been no anthropogenic CO₂ or bomb ¹⁴C uptake in this region.

We estimate an oceanwide decrease in the depth-integrated $\delta^{13}C$ by normalizing our measurements at seven sites to the increase in depth-integrated ¹⁴C activity resulting from atmospheric nuclear testing (bomb ¹⁴C) measured at nearby stations during GEOSECS (Table 2). The relation between the decrease in δ^{13} C values and increase in bomb ¹⁴C activity (Fig. 5) is: ¹³C = -18.2 ¹⁴C - 55.7, where ¹³C is the depth-integrated decrease in δ^{13} C in per mil meters and ¹⁴C is the depth-integrated increase in ¹⁴C activity in 10⁹ atoms per square centimeter. The standard error (SE) of the slope and intercept is 5.0 and 45.2, respectively. From the oceanwide average bomb ¹⁴C burden of 8.4 × 10⁹ atoms of ¹⁴C per square centimeter measured during GEOSECS (13), we calculate that the average ocean-wide decrease in depthintegrated δ^{13} C values was 208 ± 45 per mil meters; the uncertainty is the standard



Fig. 3. The change in the depth distribution of the δ^{13} C of the dissolved inorganic carbon since 1970 in the Pacific Ocean. The station comparison at 0°N uses δ^{13} C measurements on samples collected in September 1991 at Station 282 of the WOCE P16C cruise. The uncertainty of the measurements is approximately ±0.04 per mil for the 1970 data and ±0.02 per mil for the 1989 to 1991 data.

SCIENCE • VOL. 256 • 3 APRIL 1992

error of the estimate derived from the regression analysis.

Time rates of change of atmospheric $^{12}CO_2$ and $^{13}CO_2$. In order to determine the amount of fossil fuel CO_2 taken up by the oceans from the change in the ${}^{13}C/{}^{12}C$ ratio of the DIC in the ocean between 1970 and 1990, we must examine the change in the ¹³C/¹²C ratio and concentration of atmospheric CO2 that occurred over this time interval. The change of atmospheric CO_2 concentrations between 1970 and 1990 depends on the time-integrated CO_2 releases from fossil fuel combustion and cement production $(S_{\rm ff})$ and the biosphere (S_{br}) , CO₂ uptake by the biosphere (S_{bu}) , and net \tilde{CO}_2 uptake by the ocean (S_{oc}) . This dependency is expressed as:

$$V(C_{t} - C_{o}) = (S_{ff} + S_{br} - S_{bu} - S_{oc})\Delta t$$
(1)

where V is the volume of the atmosphere $(3.9 \times 10^{18} \text{ m}^3)$, C is the atmospheric CO₂ concentration (grams of carbon per cubic meter), t refers to 1990, o refers to 1970, S is a CO₂ source or sink flux (grams of carbon per year), and Δt is 20 years. Because 99% of the carbon atoms exist as the ¹²C isotope, Eq. 1 represents essentially the time rate of change of the ¹²CO₂ species. The net biospheric CO₂ flux is represented by S_{br} - S_{bu}.

Because the net oceanic CO_2 uptake between 1970 and 1990 equals the oceanwide increase in the depth-integrated concentration of DIC, Eq. 1 can be expressed as: $V(C_t - C_o) = (S_{\rm ff} + S_{\rm br} - S_{\rm bu})\Delta t$

$$-A_{\rm oc} \int (DIC_{\rm t} - DIC_{\rm o}) dz \qquad (2)$$

where A_{oc} is the global ocean surface area (361 × 10¹² m²), DIC is the DIC¹² concentration (grams of carbon per cubic meter), and z is ocean depth.

The change of the concentration of atmospheric ${}^{13}CO_2$ between 1970 and 1990 can be expressed, with the use of a formulation similar to Eq. 2, as:

$$\begin{split} & \left[C_{t} \left({}^{13}\text{C} / {}^{12}\text{C} \right)_{at} - C_{o} \left({}^{13}\text{C} / {}^{12}\text{C} \right)_{ao} \right] \\ &= \left[S_{ff} \left({}^{13}\text{C} / {}^{12}\text{C} \right)_{ff} + S_{br} \left({}^{13}\text{C} / {}^{12}\text{C} \right)_{br} \right. \\ &- S_{bu} ({}^{13}\text{C} / {}^{12}\text{C})_{bu} \right] \Delta t \\ &- A_{oc} \int \left[DIC_{t} \left({}^{13}\text{C} / {}^{12}\text{C} \right)_{st} \right. \\ &- DIC_{o} \left({}^{13}\text{C} / {}^{12}\text{C} \right)_{so} \right] dz \end{split}$$
(3)

where $({}^{13}C/{}^{12}C)_{a}$, $({}^{13}C/{}^{12}C)_{ff}$, $({}^{13}C/{}^{12}C)_{br}$, and $({}^{13}C/{}^{12}C)_{bu}$ represent the isotopic compositions of CO₂ in the atmosphere, CO₂ released from fossil fuel combustion and cement production, and CO₂ released from and taken up by the biosphere, respectively. $({}^{13}C/{}^{12}C)_{s}$ represents the isotopic composition of oceanic DIC.

In order to describe the change in the atmospheric ${}^{13}\text{CO}_2$ concentration in terms

Fig. 4. The latitudinal trends in the changes in the δ^{13} C values (per mil) and nuclear-weapons produced Δ^{14} C (per mil) (*20*) in the mixed layer of the Pacific Ocean. The δ^{13} C values are determined from the data presented in Fig. 2 and the Δ^{14} C values are determined from GEOSECS data (*13*).



of the measured change in the depthintegrated ${}^{13}C/{}^{12}C$ ratio of the DIC, the change in the oceanic DIC 13 concentration is expressed as the sum of the changes in the oceanic DIC 12 concentration and the (${}^{13}C/{}^{12}C$) of the DIC, as follows:

$$\begin{aligned} \left[DIC_{t} ({}^{13}C/{}^{12}C)_{st} - DIC_{o} ({}^{13}C/{}^{12}C)_{so} \right] \\ &= ({}^{13}C/{}^{12}C)_{st} (DIC_{t} - DIC_{o}) \\ &+ DIC_{o} [({}^{13}C/{}^{12}C)_{st} - ({}^{13}C/{}^{12}C)_{so}] \end{aligned}$$
(4)

The change in atmospheric ${}^{13}\text{CO}_2$ concentration can be expressed as:

$$V[C_{t} ({}^{13}C/{}^{12}C)_{at} - C_{o} ({}^{13}C/{}^{12}C)_{ao}]$$

= $[S_{ff} ({}^{13}C/{}^{12}C)_{ff} + S_{br} ({}^{13}C/{}^{12}C)_{br}$
- $S_{bu} ({}^{13}C/{}^{12}C)_{bu}] \Delta t$
- $\{({}^{13}C/{}^{12}C)_{st}\}A_{oc} \int [DIC_{t} - DIC_{o}] dz$
- $\{DIC_{o}\} A_{oc} \int [({}^{13}C/{}^{12}C)_{st}]_{st}$
- $({}^{13}C/{}^{12}C)_{so}] dz$ (5)

where the $\{DIC_o\}$ and $\{({}^{13}C/{}^{12}C)_{st}\}$ terms represent values averaged over the integration depth.

The net oceanic CO₂ uptake rate (S_{oc}) can be expressed in terms of the measured changes in the concentration and $^{13}C/^{12}C$ ratio of atmospheric CO₂ and changes in the depth integrated $^{13}C/^{12}C$ ratio of the DIC by substituting for S_{br} and $A_{oc} \int [DIC_t - DIC_o] dz$ in Eq. 5 as follows:

$$\begin{split} S_{\rm oc} \left[\{ ({}^{13}{\rm C}/{}^{12}{\rm C})_{\rm st} \} &- ({}^{13}{\rm C}/{}^{12}{\rm C})_{\rm br} \right] \\ &= S_{\rm ff} \left[({}^{13}{\rm C}/{}^{12}{\rm C})_{\rm ff} - ({}^{13}{\rm C}/{}^{12}{\rm C})_{\rm br} \right] \\ &- S_{\rm bul} \left[({}^{13}{\rm C}/{}^{12}{\rm C})_{\rm bu} - ({}^{13}{\rm C}/{}^{12}{\rm C})_{\rm br} \right] \\ &+ (V/\Delta t)({\rm C}_{\rm t} - {\rm C}_{\rm o})({}^{13}{\rm C}/{}^{12}{\rm C})_{\rm br} \\ &- (V/\Delta t)[{\rm C}_{\rm t}({}^{13}{\rm C}/{}^{12}{\rm C})_{\rm at} - {\rm C}_{\rm o}({}^{13}{\rm C}/{}^{12}{\rm C})_{\rm ao}] \\ &- \{DI{\rm C}_{\rm o}\}({\rm A}_{\rm oc})/\Delta t \int [({}^{13}{\rm C}/{}^{12}{\rm C})_{\rm st} \\ &- ({}^{13}{\rm C}/{}^{12}{\rm C})_{\rm so}] dz \end{split}$$
(6)

Calculation of the rate of net oceanic CO_2 uptake requires estimates of the $\delta^{13}C$ values for the CO_2 released from fossil fuel combustion and cement production and taken up by and released from the bio-

sphere. The δ^{13} C of the CO₂ released by fossil fuel combustion and cement production during the last 20 years was -27.2 per mil (1). The CO₂ taken up currently by the biosphere has a δ^{13} C value about 20 per mil lower than the δ^{13} C of atmospheric CO₂ on the basis of isotopic fractionation during carbon fixation in, and predominance of, C₃ plants (5, 6). If we assume that the isotopic fractionation during carbon fixation has remained constant over time, then the δ^{13} C value of CO₂ taken up by the biosphere has decreased at the same rate as the δ^{13} C value of atmospheric CO₂.

Currently the $\delta^{13}C$ value of the CO_2 released from the biosphere is likely to be higher than that of the CO_2 taken up by the biosphere because of the lag time between photosynthetic carbon fixation and microbial degradation of woody plant material (14). Estimates of the net primary production rate for the terrestrial biosphere (S_{bu}) average about 60 gigatons (Gt) of carbon per year: About 40 Gt yr^{-1} enter a biospheric carbon pool with a short turnover time (<3 years) and about 20 Gt yr^{-1} enter into woody plant material with a decadal (about 30 years) turnover time (15). By coupling these biospheric carbon turnover times and flux rates with the measured change in the atmospheric δ^{13} C value, we estimate that the CO₂ currently released from the biosphere is, on average, about 0.2 per mil enriched relative to the CO₂ entering the biosphere.

Determining the rate of net oceanic CO_2 uptake in terms of the mass balances of atmospheric ${}^{12}CO_2$ and ${}^{13}CO_2$, as presented above, eliminates the need to specify both the rate and isotopic fractionation effect of air-sea CO_2 gas exchange and an oceanic mixing depth.

Rates of oceanic and biospheric net CO_2 uptake. Between January 1970 and 1989 the mean atmospheric CO_2 concentration increased from 324.0 to 351.0 ppmv (1). Detailed $\delta^{13}C$ measurements have been made only since 1978; between 1978 and 1988 the $\delta^{13}C$ decreased from -7.49 to -7.74 per mil (1). We used an empirical relation between the changes in these data and CO_2 concentration ($\delta^{13}C = -0.0147$)



Fig. 5. The correlation between the depthintegrated changes in δ^{13} C and bomb-produced ¹⁴C activity at the seven locations in the Pacific Ocean where both the ¹³C and ¹⁴C changes have been determined (see Tables 1 and 2). The linear regression, represented by the line, is defined by ¹³C = -18.2 ¹⁴C - 55.7, where ¹³C is in per mil meters and ¹⁴C is in 10⁹ atoms per square centimeter; the standard error of the Y estimate is ±45 per mil meters.

 $[CO_2] - 2.60$, where $\delta^{13}C$ is in per mil and the $\overline{\text{CO}}_2$ concentration is in parts per million; the SE of the slope and intercept is 0.0019 and 0.65, respectively) to estimate the atmospheric δ^{13} C in 1970 and 1990. The estimated δ^{13} C values were -7.36 per mil in 1970 and -7.76 per mil in 1990; for the latter value, we assumed that the atmospheric CO₂ concentration was 351.5 ppmv in January 1990, which was determined by a regression between CO_2 concentration and time derived from values measured between 1978 and 1988 (1). The oceanwide change in the depth-integrated ¹³C/ ¹²C ratio between 1970 and 1990 was -208 per mil meters based on our measurements in the Pacific Ocean and an oceanwide extrapolation with the use of the bomb $^{14}\mathrm{C}$ burden, as discussed above. In order to determine S_{oc} we assumed that the depthaveraged concentration of DIC in 1970, $\{DIC_{o}\}$, was 2 mmol liter⁻¹ and that the depth-averaged δ^{13} C value for the DIC in 1990, { $({}^{13}C/{}^{12}C)_{st}$ }, was 1.7 per mil on the basis of our measured values in the Pacific Ocean.

The amount of CO₂ released from fossil fuel combustion and cement production is estimated to have been 101.6 Gt of carbon between 1970 and 1990, on average 5.1 Gt yr^{-1} (16). During the same period, the atmospheric CO2 increase was 58 Gt of carbon or 2.9 Gt yr^{-1} , that is, 57 percent of the fossil fuel release. Our calculations indicate that the net oceanic uptake (S_{oc}) was 42 Gt of carbon C or 2.1 Gt yr^{-1} , that is, 41 percent of the industrial CO_2 release; we have assumed that the flux and time rate of change terms represented in Eq. 6 were constant between 1970 and 1990. If Eq. 1 properly represents all the major carbon reservoirs, then there has been little net CO_2 release from the biosphere, that is, (S_{br} $-S_{bu}$ = -0.1 Gt of carbon per year during the last 20 years. If deforestation was occur-

Table 1. Change in the δ^{13} C value of dissolved inorganic carbon in the mixed layer (in per mil) and integrated over depth in the Pacific Ocean (in per mil meters). Data for stations TOW-7, SCAN10, and ANTIPODES 15 are from Kroopnick (11). Lat., latitude; Long., longitude; HUD, HUDSON.

Station	Lat.	Long.	Date	Station	Lat.	Long.	Date	δ^{13} C change	
								Mixed layer	Integrated
HUD 297	29°N	150°W	5/70	HOTS	23°N	160°W	4/90	-0.42	-305
TOW 7-146	20°N	166°W	6/70	CGC91-19	21.6°N	150°W	3/91	-0.25	-231
HUD EQ	0°S	150°W	5/70	W16C-268	0°S	151⁰W	9/91	-0.39	-60
SCAN 10-38	6.5°S	107⁰W	1/70	RITS89-12	4°S	110⁰W	2/89	-0.46	-147
ANTI 15-06	16.5°S	170°W	8/71	CGC90-3	16.5°S	170°W	2/89	-0.41	-243
HUD 282	40°S	150°W	5/70	CGC90-14	37.5°S	170⁰W	3/90	-0.57	-216
HUD 280	50°S	150°W	5/70	RITS89-46	44°S	130°W	3/90	-0.24	-217

ring at 1.6 ± 0.8 Gt of carbon per year between 1970 and 1990 (16, 17), then the biosphere grew by almost an equivalent amount during this period. These results are similar to the conclusions reached a decade ago (18), but are not dependent on ocean ventilation models used for those and sub sequent calculations of oceanic CO2 uptake.

We evaluated the error of the calculated values of S_{oc} using a numerical bootstrap procedure (19). The uncertainties in the three measured time rates of change were ± 45 per mil meters for the depth-integrated ${}^{13}C/{}^{12}C$ ratio (equal to the standard error of the estimate derived from the regression analysis of the depth-integrated ¹³C and ¹⁴C changes, as discussed above), and ± 0.2 ppm and ± 0.03 per mil for the changes of the concentration and δ^{13} C value of atmospheric CO_2 , respectively (1). We estimated the uncertainties in S_{bu} , S_{ff} , $({}^{13}C/{}^{12}C)_{br}$, and $({}^{13}C/{}^{12}C)_{ff}$ to be ±30 and ±10 percent and ± 2 and ± 1 per mil, respectively. The difference in δ^{13} C values between CO₂ currently fixed into and released from woody plant material was taken as $-0.6 \pm$ 0.4 per mil. The estimated uncertainties in ${DIC_o}$ and ${({^{13}C/^{12}C})_{st}}$ were ±2 percent and ±0.3 per mil, respectively. On the basis of these uncertainties, we derived an error (± 1 standard deviation) of ± 15 Gt of carbon for S_{oc} and S_{br} . The error in S_{oc} and S_{br} depends on primarily (about 50 percent) the uncertainty in the magnitude of the depth-integrated change in the ${}^{13}C/{}^{12}C$ ratio of the DIC.

The uncertainty in S_{oc} depends mainly on how well the measured depth profiles of the change in δ^{13} C values from 1970 to 1990 represent the entire ocean. Thus, improved estimates of Soc will be possible as more station comparisons are made, especially when δ^{13} C data from the Indian and Atlantic oceans are obtained. With the data available, the seven profiles we have compared appear to be fairly representative of the global ocean. The average decrease in $\delta^{13}C$ values for the mixed layer at these seven sites was 0.39 per mil, close to the Pacific Ocean average of 0.40 per mil, which is derived from the data presented in Fig. 2. For the GEOSECS stations in close proximity to the stations where the $\delta^{13}C$ values were compared, the average bomb ¹⁴C burden of 8.1×10^9 atoms ¹⁴C per square centimeter is close to the oceanwide average of 8.4×10^9 atoms ¹⁴C per square centimeter (13). The mean depth-integrated change in $\delta^{13}C$ value at the seven stations was -202 per mil meters, close to the oceanwide mean -208 per mil meters value derived from the bomb ¹⁴C normalization procedure. Our measurements indicate that the average penetration depth of the change in δ^{13} C values over the last 20 years was 520 m, that is, the depth-integrated δ^{13} C decrease divided by the mixed layer decrease (208 per mil meters/0.4 per mil). This $\delta^{13}\dot{C}$ penetration depth is consistent with the average ocean-wide penetration depth of about 350 m determined for the bomb-produced ^{14}C (13) after consideration of the shorter time history (about 10 years) of the ¹⁴C perturbation at the time of measurement (1972 to 1978).

The measured time rates of change of atmospheric and oceanic $^{13}\mathrm{C}/^{12}\mathrm{C}$ ratios yield values for oceanic net CO₂ uptake

Table 2. The increase in ¹⁴C activity produced from testing of nuclear weapons measured at seven GEOSECS stations in the Pacific Ocean (13). These station locations correspond to the stations listed in Table 1. The mixed layer ¹⁴C changes are in per mil (20) and the depthintegrated changes are in 10⁸ atoms per square centimeter.

				¹⁴ C Change		
Sta- tion	Lat.	Long.	Date	Mixed layer	Depth inte- grated	
213	31.0°N	168.5°W	9/73	237	11.3	
227	25.0°N	170.1°E	11/73	235	11.1	
334	0.1°N	124.6°W	5/74	143	3.5	
331	4.6°S	125.1°W	5/74	111	2.4	
263	16.7⁰S	167.0°W	1/74	187	8.8	
303	38.6°S	170.1⁰W	3/74	160	10.6	
296	44.9°S	166.6°W	3/74	135	8.9	

SCIENCE • VOL. 256 • 3 APRIL 1992

rates that are not model-dependent. Until direct measurements can be made of the changes in the carbon content of the ocean and biosphere, continued oceanic and atmospheric δ^{13} C measurements may provide the best determination of the fate of fossil fuel CO₂.

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- 12. The depth-integrated change in δ^{13} C values was not calculated for the station pair at 10°S and 11°S, shown in Fig. 3, because $\delta^{13}C$ values in 1970 were not measured between 1 and 294 m, where most of the change in δ^{13} C values was likely to have occurred. However, this station pair does provide a useful comparison of deepwater $\delta^{13}C$ values measured in 1970 and 1990. In the station comparison at 0°S 150°W we used 813C values measured at station 268 during the WOCE P16 cruise in September 1991. A comparison of two stations at 54°N and 150°W (HUDSON 301 in 1970 and CGC91-58 in 1991) yielded a consistent offset in δ^{13} C values of about -0.2 per mil between 200 and 2500 m; we did not include this station in Table 1 because of this deep water offset.
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- 19. In the error analysis we used a random selection of

(Continued from p. 14)

what we know, or are troubled by, but don't know that we know (or don't admit that we are troubled by) because we don't have a satisfying explanation.

> David P. Barash Department of Psychology, University of Washington, Seattle, WA 98195

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The Search for Eve

Marcia Barinaga's article "'African Eve' backers beat a retreat" concerning the debate about modern human origins (Research News, 7 Feb., p. 686) contains many examples of the kind of thinking that still values for each measured parameter based on their respective means and standard deviations. A large number, for example, 1000, of calculations of the calculated parameter, S_{oc} , were made with the randomly selected values for each measured parameter. Then the mean and standard deviation of S_{oc} were determined from the 1000 calculated S_{oc} values; see B. Efron, *Biometrika* **68**, 589 (1981).

- 20. The ¹⁴C activity is expressed as $\Delta^{14}C = [(A_s/A_{ox}) 1]1000$ in per mil, where A_s is the activity of the sample and A_{ox} is the age-corrected activity of the oxalic acid standard after ¹³C normalization [M. Stuiver and H. A. Polach, *Radiocarbon* 19, 355 (1977)].
- 21. We thank the group at NOAA-Pacific Marine Environmental Laboratory, in particular D. Feely

plagues the study of human evolution. It

was clear by 1987 that new methods of

phylogenetic analysis for reconstructing hu-

man history would be necessary, and not

just new DNA sequences from more people

or a 4.0 version of some existing software

package that violated present knowledge of

the system (unequal sample sizes and vary-

from human fossils must now be accompa-

nied by better methods to absorb large data

sets for analysis of variance and clustering.

DNA sequences from actual fossils, not

reconstructed hypothetical ancestors, are

the key. Further arguments are tedious and

wasteful of limited resources and predispose

observers to equate evolutionary biology

Rebecca L. Cann

Molecular Biology,

Honolulu, HI 96822

Department of Genetics and

University of Hawaii at Manoa,

Breakthroughs in extraction of DNA

ing rates of evolution).

with street theater.

and P. Murphy, and M. Dudley for sample collections. J. Zhang helped with sample collection and extraction. S. Moe, M. Knox, and L. Lu helped with sample extraction, and D. Wilbur oversaw the mass spectrometer measurements. P. Kroopnick provided tabulated $\delta^{13}C$ data from SCAN and ANTIPODES cruises. T. Brazunias, U. Siegenthaler, S. Smith, and P. Tans provided helpful reviews. The U.S. work was supported by NOAA's Climate and Global Change Program (NA90AA-D-AC828) and the Canadian work was supported by the Department of Fisheries and Oceans and the Panel of Energy Research and Development (Project 48115).

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Corrections and Clarifications

- The News & Comment article "Wilson slashes spending for antismoking effort" by Marcia Barinaga (13 Mar., p. 1348) erroneously stated that the California legislature approved the shut-down of California's antismoking ad campaign. The legislature approved the removal of \$29 million from the smoking-initiative budget, but those funds do not influence the ad campaign. The ad campaign funding was removed by an action of the governor that did not have legislative approval.
- In Joseph Palca's News & Comment article "AAAS in Chicago: Dawn of a new era?" (21 Feb., p. 918), University of Miami philosopher Kenneth Goodman's name was inadvertently misspelled.
- The name of the associate vice chancellor for research at the University of Illinois, Tina Gunsalus, was inadvertently misspelled in the News & Comment article "What's left on the table" by David P. Hamilton (13 Mar., p. 1345).

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The AAAS-Newcomb Cleveland Prize is awarded to the author of an outstanding paper published in *Science*. The value of the prize is \$5000; the winner also receives a bronze medal. The current competition period began with the 7 June 1991 issue and ends with the issue of 29 May 1992.

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The award will be presented at the 1993 AAAS annual meeting. In cases of multiple authorship, the prize will be divided equally between or among the authors.