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Dissolved inorganic radiocarbon in the North Pacific Ocean and Sargasso Sea

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Abstract

We present radiocarbon measurements of dissolved inorganic carbon (DIC) in depth profiles from reoccupations of our central North Pacific Ocean and Sargasso Sea (SS) sites. From 1985 to 1999, an increase in $\Delta^{14}C$ values of 6–17\% was measured between 1150 and 2400 m depth in the North Central Pacific. Natural changes in deep circulation are likely responsible for variability of $\Delta^{14}C$ signatures in the deep ocean, though we cannot rule out the presence of bomb $^{14}C$ at this depth range. Bomb $^{14}C$ had increased in the deep SS from 1989 to 2000; this is the result of southward transport of North Atlantic Deep Water (NADW) to this site.

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Keywords: Dissolved inorganic carbon; Radiocarbon; Bomb radiocarbon; Deep circulation

1. Introduction

Oceanic dissolved inorganic carbon (DIC) is the largest pool of exchangeable carbon on Earth (36,000 GtC). It is important to know the timescales of mixing and ventilation of deep ocean waters because most of the excess CO$_2$ produced by fossil fuel and biomass burning will eventually be stored in the oceans. Early radiocarbon measurements in seawater revealed several hundred-year transit times of deep water masses in the world’s oceans (Bien et al., 1963; Broecker et al., 1960; Stuiver et al., 1983). Global data sets amassed during programs such as the Geochemical Ocean Sections Study (GEOSECS) and the World Ocean Circulation Experiment (WOCE) have been used for quantitative studies of the oceanic carbon cycle (Key, 1996; Stuiver et al., 1983).

Since the production of bomb $^{14}C$ in the late 1950s and early 1960s, it has been possible to measure short-term exchange of carbon, e.g. the transfer of CO$_2$ across the air–sea interface. The level of bomb $^{14}C$ in a given carbon pool is a reflection of the turnover time of the carbon with respect to exchange with the atmosphere. By the 1970s, bomb $^{14}C$ was detected to depths of several
hundred meters in the non-polar oceans (Ostlund and Stuiver, 1980; Stuiver and Ostlund, 1980). The deep northern North Atlantic contained bomb
14C throughout the entire water column, introduced during North Atlantic Deep Water (NADW) formation. By 1991, penetration of bomb 
14C extended to 1000 m depth in the Pacific (Key, 1997; Stuiver et al., 1996).

The Δ14C data reported here show an increase in values in the deep waters of both the North-central Pacific (NCP) and the Sargasso Sea (SS) (Druffel et al., 1992). Change in circulation is the likely mechanism responsible for the higher Δ14C values in the upper part of the deep NCP. Remineralization of bomb-laden particulate organic matter from the surface is unlikely to have caused a measurable increase in the Δ14C values of DIC in the deep North Pacific or North Atlantic.

2. Methods

Water samples were collected from the NCP site (31°N, 159°W, bottom depth 5820 m) on the Avon cruise from 26 May to 12 June 1999 and from the SS site (31°50’N, 63°30’W, bottom depth at 4500 m) on the SarC cruise from 14 to 29 June 2000. The NCP site is 1000 km north of Hawaii and the SS site is 100 km southeast of Bermuda. Radiocarbon and δ13C measurements of suspended particulate organic carbon collected during these cruises were reported previously (Druffel et al., 2003), and radiocarbon and abundance data of dissolved organic carbon will be reported separately (Bauer et al., in preparation; Loh et al., 2004). The NCP and SS sites were occupied earlier, from 6 June to 3 July 1987 (Eve cruise) and from 29 May to 20 June 1989 (Hydros-6 cruise), respectively, and isotope results were reported separately (Druffel et al., 1992).

Seawater samples were collected in 1999 and 2000 using 12-L or 30-L Go-flo bottles deployed on a hydrowire. Samples for DIC Δ14C and δ13C analyses were filtered through glass fiber filters (1 μm effective pore size) directly into 1-L glass containers and poisoned with a saturated solution of mercuric chloride. Separate samples were filtered and poisoned in the same manner for alkalinity and total CO2 ([DIC]) analyses. For the DIC Δ14C and δ13C analyses, the samples were acidified and sparged of CO2 gas according to published techniques (McNichol et al., 1994). Carbon dioxide was converted to graphite using hydrogen gas and cobalt metal at 550 °C (Vogel et al., 1987).

The Δ14C measurements from the Avon 1999 cruise were made at the National Ocean Sciences Accelerator Mass Spectrometry Facility (NOSAMS) of the Woods Hole Oceanographic Institution (WHOI), and those from the SarC 2000 cruise were made at the W.M. Keck Carbon Cycle AMS Laboratory at the University of California, Irvine (UCI). The DIC Δ14C values were reported according to standard techniques (Stuiver and Polach, 1977) and have a total uncertainty determined from replicate seawater analyses of ±3.9‰. The δ15N measurements were performed at NOSAMS with a total uncertainty of ±0.1‰. Alkalinity and [DIC] measurements were obtained by closed vessel titration (Bradshaw et al., 1981; Brewer et al., 1986) in the laboratory of D. McCorkle (WHOI). Alkalinity and [DIC] measurements were determined using a nonlinear curve fitting approach (DOE, 1994) and standardized using certified reference materials obtained from Andrew Dickson (Scripps Institution of Oceanography). The standard deviation of pairs of replicate analyses of culture water was 4 μeq/kg for alkalinity and 6 μmol/kg for [DIC]. The alkalinity results were consistently high by 25–40 μeq/kg compared with other cruises, which we attribute to the unusually long storage times of the samples prior to analysis (>9 mos); for this reason, the alkalinity results were not reported.

3. Results

The DIC Δ14C values of samples collected from the 1999 NCP site ranged from 95‰ to −243‰ (at 3 m depth) to −243‰ (at 3221 m) (Fig. 1a). Values remained low (average −239.0 ± 2.9‰, n = 4) between 2400 and 3600 m, and increased to an average of −215 ± 6‰ (n = 5) between 4200 and 5770 m (50 m above bottom). Values of [DIC] were low in the upper 100 m (2050 ± 6 μmol/kg, n = 4), reached a maximum at 1300 m (2382 μmol/kg) and decreased slightly in deep waters (Table 1).

The DIC Δ14C values obtained for samples from the SS site in 2000 (Fig. 2) decreased from 81‰ at 3 m to a low of −72‰ at 1005 m (Antarctic Intermediate Water (AAIW)); values increased to a secondary maximum (−32‰) at 1513 m, and decreased to −84‰ at 50 m above bottom. The [DIC] values were low in the upper 50 m (2099–2107 μmol/kg), highest in AAIW (2203–2221 μmol/kg between 850 and
1300 m) and slightly lower in NADW (2182–2201 μmol/kg at 1500–3600 m) (Table 2).

The DIC δ13C values in the NCP were highest in surface waters (1.08% at 3 m), decreased to a minimum of −0.84% at 1141 m, and increased with depth to 5770 m (0.22%) (Fig. 3a). At the SS site, the DIC δ13C values were highest in surface water (1.09% at 20 m), decreased to a minimum (−0.21%) at 1005 m, and increased to 0.7–0.9% in deep water (Fig. 3b).

4. Discussion

In the following discussion, we compare the isotopic measurements of samples reported for the recent cruises to the NCP and SS sites with those obtained from earlier cruises.

4.1. North Central Pacific

Surface Δ14C values in June 1999 were about 40–50% lower than those measured in samples from the Eve cruise in June 1987 (Druffel et al., 1992) (Fig. 1a). This reduction reflects the decrease of bomb 14C in DIC in subtropical surface waters since the early 1970s (Druffel, 1987). In contrast, Δ14C values between 600 and 2400 m depth were significantly higher in 1999 than those in 1987 (Fig. 1a, b). The average Δ14C difference at depths between 600 and 900 m (38 ± 21‰, n = 3) was greater than that between 1150 and 2400 m (8 ± 1‰, n = 4). Comparison of our DIC Δ14C data with profiles obtained during GEOSECS in September 1973 (Stn 204, 31°N, 150°W) (Ostlund and Stuiver, 1980) and WOCE (March 1991, Stn 31, 30°N, 152°W) (Key, 1996; Stuiver et al., 1996) is also plotted with our data in Fig. 1a, b. These sites are located 860 and 680 km east of our NCP site, respectively, and some spatial variability of Δ14C may be expected. In the upper 100 m, Δ14C values were highest during 1973 (146–178‰) and lowest during the 1999 cruise (77–95‰). Between 400 and 2400 m, the 1999 values are higher than Δ14C values from similar depths for any of the previous cruises (Fig. 1a, b). Additionally, the depth of the Δ14C minimum appears to have deepened with time, from about 1700–2400 m in 1973, to 2200–2400 m in 1987–1991, to 3200 m in 1999 (Fig. 1a, b).

To more clearly illustrate the change of Δ14C with time, Δ14C measurements from nine depth ranges (50–2400 m) are plotted versus time of collection (Fig. 4). Data from an earlier cruise (Alcyone October 1985) to the NCP site (Druffel et al., 1989) are also included. At ~50 m depth, Δ14C
values decrease from 1973 to 1999, and at 150 m a maximum in 1985 is apparent. At 300 m and deeper, $\Delta^{14}C$ values generally increase with time. Least squares fits of the data from 900, 1150, 1800 and 2400 m depth (Fig. 4 inset) reveal increases of 35\%, 17\%, 14\% and 12\%, respectively, during the 26-year period. These increases in $\Delta^{14}C$ are > 3 times the 1-$\sigma$ uncertainty of our measurements (3.9\%), and thus statistically significant. Could this increase be due to penetration of bomb $^{14}C$ below the main thermocline?

From data compiled as a part of the Global Ocean Data Analysis Project, Key et al. (2004) showed that bomb radiocarbon had penetrated no deeper than 1000 m in the North Pacific by 1991. They found a weak bomb $^{14}C$ signal at 1000 m only in the region of intermediate water formation in the far northwest Pacific. The data presented here (Fig. 4) suggest that, by 1999, $\Delta^{14}C$ values were higher in the North central Pacific basin, as deep as 2400 m.

Is it possible that the 12–17\% increase in the deep ocean DIC $\Delta^{14}C$ values represents natural variability? The ranges of $\Delta^{14}C$ values taken over a 2-year period at 1600 and 2500 m depth at Stn M in the NE Pacific were 16\% ($n = 7$) and 18\% ($n = 6$), respectively (Masiello et al., 1998). The ranges of $\Delta^{14}C$ values measured in samples from our cruises to the NCP site at 1800 and 2400 m in 1985 and 1987 were 8\% ($n = 2$) and 4\% ($n = 4$), respectively. The Stn M site is a more productive, coastal site, and comparison with the midgyre NCP site may not be valid. Roussenov et al. (2004) used an isopycnic circulation model to show that DIC $\Delta^{14}C$ values in the deep Pacific are controlled by lateral transport of bottom water from the south and balance between advection–diffusion and decay of $^{14}C$ in the vertical. They found that a strengthening of diapycnic mixing causes an overall increase in $\Delta^{14}C$ values over the North Pacific basin similar to the increase that we see at the NCP site. Thus, it is possible that shifts in deep circulation could cause

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changes in $\Delta^{14}C$ of the order that we observe at 1150–2400 m depth. Is it possible that the increase in $\Delta^{14}C$ is due to input of bomb $^{14}C$? The two possible sources of bomb $^{14}C$ would be from remineralization of surface-derived particulate organic carbon (POC) to DIC, and physical mixing of bomb-laden upper waters into the deeper layers of the ocean. First, we calculate the change in deep DIC $\Delta^{14}C$ expected from the input of remineralized bomb-laden POC from surface waters. The amount of DIC in the water column from 1000 to 2000 m with 1 m$^2$ area is 2.93 x $10^4$ gC/m$^2$, assuming an average [DIC] of 2370 m mol/kg, and an average density of 1030 kg/m$^3$. The potential change of $\Delta^{14}C$ by remineralization of POC using the average global flux rate of sinking POC at 1000 m depth of 2.8 gC/m$^2$ (Martin et al. (1987) for 26 years = 73 gC/m$^2$), with a $\Delta^{14}C$ of 89%, is only 0.75%, which is much smaller than the observed 12–17% difference. This calculation assumes only vertical transport of remineralized POC and does not include lateral transport of material. As a note, we could neither use the potential alkalinity method (Rubin and Key, 2002) to estimate the actual bomb $^{14}C$ concentration (alkalinity measurements were too high), nor could we use Broecker’s silica method (Broecker et al.,

![Fig. 2. $\Delta^{14}C$ values of DIC from the SS site (31.5°N, 63.5°W) during the SarC (June 2000) and Hydros (June 1989) cruises. Also shown are the $\Delta^{14}C$ profiles presented previously during GEOSECS (1973, Stn 120, 33°16'N, 56°33'W) (Stuiver and Oslund, 1980) and TTO in 1981 (Stn 237, 33°8'N, 56°30'W) cruises (Oslund, 1981).](image)

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<td>3650</td>
<td>–78.1</td>
<td>0.86</td>
<td>2201</td>
<td>34.898</td>
<td></td>
</tr>
<tr>
<td>4899</td>
<td>151</td>
<td>4320</td>
<td>–84.1</td>
<td>0.80</td>
<td>2212</td>
<td>34.913</td>
<td></td>
</tr>
</tbody>
</table>
1995) because silicate was not measured in our water samples. Second, it is difficult to attribute physical mixing of upper waters into the deep waters as a source of bomb $^{14}$C without adopting unrealistically high cross isopycnal mixing rates. Thus, we conclude that the increase in the $\Delta^{14}$C signature between 1150 and 2400 m depth does not reflect bomb $^{14}$C addition to the deep Pacific. Horizontal variability in $\Delta^{14}$C of up to 14‰ is observed below 1000 m at WOCE stations along 152°W line and 134°W line near our NCP station in 1991 (Stuiver et al., 1996; Key et al., 2004). Therefore, we believe that the increase in the $\Delta^{14}$C values with time represents changes in circulation and mixing in the deep waters, causing shifts in the baseline $\Delta^{14}$C signature.

Important processes contributing to $\delta^{13}$C of DIC are (1) mixing, (2) transfer of CO$_2$ between air and sea, (3) remineralization of organic matter to CO$_2$, and (4) removal of CO$_2$ during photosynthesis in the surface water. At the NCP site, DIC $\delta^{13}$C values in the upper 1100 m were lower in 1999 than during any of the earlier cruises (Fig. 3a). This depletion is evidence of the $^{13}$C Suess Effect, which is the input of $^{13}$C-depleted CO$_2$ from fossil fuels and terrestrial biomass to the main thermocline (McNichol and Druffel, 1992; Quay et al., 1992). The $\delta^{13}$C values in the surface decreased by 0.8‰ from 1973 (Kroopnick, 1985) to 1999 and at 1100 m by about 0.3‰ (Fig. 3a). The decrease of $\delta^{13}$C in the surface of the North Pacific was about 0.02‰ yr$^{-1}$ from 1970 to 1990 (Quay et al., 1992). This is similar to our results ($0.8{\pm}0.02$‰/26 yr = $0.03{\pm}0.01$‰ yr$^{-1}$) from 1973 to 1999, and is consistent with an increased rate as anthropogenic CO$_2$ input increases with time.

4.2. Sargasso Sea

The $\Delta^{14}$C value of SS surface water sampled in 2000 was about 40‰ lower than that obtained 11 years earlier at this site (Fig. 2). In contrast, at 1800 and 2200 m, $\Delta^{14}$C values were significantly higher (by 19±4‰) during 2000 than those obtained in 1989, and the same or slightly higher (7±3‰, n = 4) below 2200 m depth (Fig. 2).

A comparison of our data in 2000 with a profile obtained during GEOSECS in March 1973 (Stn 120, 33°16’N, 56°33’W) (Stuiver and Ostlund, 1980) and one depth (1994 m) from TTO in October 1981 (Stn 237, 33°8’N, 56°30’W) (Ostlund, 1981) are also shown in Fig. 2 (both located 670 km to the east of our SS site). In the upper 200 m, $\Delta^{14}$C values were highest during 1973 (128–130‰) and lowest in 2000 (57–81‰). At 1800, 2700 and 4400 m, the $\Delta^{14}$C values in 2000 were significantly higher than those from similar depths for the previous cruises (Fig. 5). Similar to the NCP, surface $\Delta^{14}$C values decreased from 1973 to 2000. At 600 m and deeper, $\Delta^{14}$C values increased with time. Least squares fits of the data from 1800, 2700, 3600 and 4400 m revealed increases of 25‰, 15‰, 9‰ and 16‰, respectively, over this 27-year period. These $\Delta^{14}$C increases show the progressive penetration of bomb $^{14}$C with time into the deep water column of the SS.

The primary source of bomb $^{14}$C in the deep SS is from NADW that is produced in the Norwegian and Greenland Seas. Previous studies have shown that tritium and chlorofluorocarbons were...
incorporated into NADW and transported south to
the subtropics via deep thermohaline circulation
(Ostlund and Rooth, 1990; Smethie et al., 1986). As
shown above for the NCP site, remineralization of
surface-derived POC to DIC is an unlikely source of
bomb $^{14}$C to the deep SS, unless POC fluxes and
dissolution rates are seriously underestimated.

At the SS site, DIC $^{13}$C values in the upper 1500 m were lower during 2000 than those ob-
tained from Stn 31 (27°0’N 53°32’W) during the
1972 GEOSECS cruise (Kroopnick, 1985) (Fig. 3b)
(no $^{13}$C measurements were available from Stn
120). From 1972 to 2000, the $^{13}$C values had
decreased by about 0.8‰ in the surface and by
about 0.5‰ between 200 and 700 m. At 1000 m, the
2000 value appears anomalously low ($–0.2‰$),
though there is no obvious reason to discount this
measurement. As in the NCP, the decrease of $^{13}$C
values indicates the presence of $^{13}$C-depleted at-
mospheric CO$_2$ from anthropogenic sources. Kortzin-
ger et al. (2003) reported that the mean $^{13}$C
decrease in the upper 1000 m of the North Atlantic
from 1950 to 1993 was 0.026 ± 0.002‰ y$^{-1}$. Data
from Fig. 3b show a similar decrease from 1972 to
2000 of 0.015–0.026‰ y$^{-1}$ in the upper 700 m.

5. Conclusion

Bomb $^{14}$C has continued to increase in the deep
waters of the SS from 1973 to 2000. What is
surprising is the increase in $^{14}$C to a depth of
~2400 m in the NCP, because anthropogenic tracers
(e.g., bomb $^{14}$C, tritium and CFCs) have been found only to a depth of about 1000 m in the North Pacific by the 1990s (Key et al., 2004). It does not seem feasible that remineralization of POC from the surface could provide a sufficient source of bomb $^{14}$C to the deep North Pacific. Instead, we conclude that changes in deep water circulation patterns offer a plausible explanation of the increase in $\Delta^{14}$C that we observe in the deep waters of the North Pacific Ocean.

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