Natural Radiocarbon Distribution in the Deep Ocean

Katsumi MATSUMOTO\textsuperscript{1} and Robert M. KEY\textsuperscript{2}

\textsuperscript{1}Geological Survey of Japan, AIST, 1-1-1 Higashi, Tsukuba, Ibaraki 305-8567, Japan
\textsuperscript{2}Atmospheric and Oceanic Sciences Program, Princeton University, Princeton, NJ 08540, U.S.A.

Abstract. We present for the first time objectively mapped natural radiocarbon (\textsuperscript{14}C) abundance in the global deep ocean using mostly new measurements from the World Ocean Circulation Experiment (WOCE). Compared to earlier data from the Geochemical Ocean Sections Study, the new data number almost seven times more. Also, the WOCE data collected during the 1990s have sufficient spatial coverage to make objective mapping meaningful. The new natural $\Delta^{14}$C maps clearly show the deep path of the well-known global thermohaline circulation. We compare the new maps with simulation results from three versions of the Princeton Ocean Biogeochemistry Model, as a means to illustrate how such maps can be used to constrain ocean carbon cycle models. We show that the three Princeton models, which are not necessarily atypical of currently available coarse resolution ocean general circulation models, have a wide range of behavior in simulating the deep natural $^{14}$C distribution. Some deviate significantly from the observed in both amplitude and spatial pattern. The same models make significantly different centennial time scale projections of future anthropogenic carbon uptake by the ocean. This highlights the importance of using the new natural $^{14}$C data to properly evaluate the performance of ocean carbon cycle models in ventilating the deep ocean.

Keywords: radiocarbon, ocean general circulation model, World Ocean Circulation Experiment (WOCE), ocean ventilation, North Atlantic Deep Water, Antarctic Bottom Water, Circumpolar Deep Water, Pacific Deep Water, thermohaline circulation

1. INTRODUCTION

Deep ocean circulation is difficult to characterize by observations. The ocean covers nearly 70\% of the Earth’s surface area and the average depth of 3700 m make the deep ocean virtually impossible for synoptic sampling by man-powered efforts. Despite the difficulty, scant temperature observations were used to infer correctly nearly two centuries ago that the deep ocean is part of a global overturning circulation. As noted by Warren (1981), the earliest recorded subsurface temperature measurements in the open ocean were made in 1751 by Henry Ellis, captain of a British slave-trading vessel. These measurements, made in the subtropical North Atlantic, revealed that much of the deep water is cold, and warm water is confined to a thin layer near the surface. Count Rumford, who experimentally discovered convection in liquids, correctly understood that the
temperatures observed by Ellis indicated that the cold deep water in the subtropics must derive ultimately from the polar surface waters and that there must be an overturning circulation system that carries deep water equatorward to close the loop (Rumford, 1800). The modern version of this circulation system is of course the thermohaline circulation, also known as the “conveyor belt” circulation (Broecker, 1991).

A description of the conveyor belt was made possible by the radiocarbon (\(^{14}\text{C}\)) measurements from the Geochemical Ocean Sections Study (GEOSECS) in the 1970s (Ostlund and Stuiver, 1980; Stuiver and Ostlund, 1980, 1983). During GEOSECS the Atlantic, Pacific, and Indian Ocean basins were each sampled by two north-south transects over the course of separate two-year periods. Natural radiocarbon, which is produced only in the atmosphere, is very useful in understanding the deep ocean circulation, because any natural \(^{14}\text{C}\) atoms found in the ocean interior must have come by means of air-sea gas exchange and subsequent transport within the ocean. Because \(^{14}\text{C}\) decays with a known half-life of 5730 years, its abundance in the deep ocean is a direct measure of how much \(^{14}\text{C}\) is supplied to the deep ocean from the surface by ventilation. The GEOSECS \(^{14}\text{C}\) data showed that the deep ocean ventilates on centennial time scale (Broecker and Peng, 1982). This means that the GEOSECS data obtained from different years in a decade can be considered synoptic for the deep ocean.

A primary motivation for studying deep circulation is that the deep ocean is a major component of the global carbon cycle. The world ocean has approximately 38,000 Pg-C (10\(^{15}\) grams C), which dwarfs the atmospheric and terrestrial biospheric reservoirs that respectively have about 730 and 2200 Pg-C (Houghton et al., 2001). Since most of the oceanic carbon resides in the deep ocean, even a small change in its carbon budget can significantly impact the atmospheric budget and hence the global climate. Under natural conditions, the chances of such an event may seem remote, because the deep ocean is a slow component compared to the atmosphere, upper ocean, and terrestrial biosphere.

However, measurements from polar ice cores provide abundant evidence for multiple, abrupt climate changes in the last glacial cycle (Dansgaard et al., 1993), some of which may have involved changes in the deep ocean (Broecker, 1998, 2003). Abrupt climate change may be a real possibility today, when human activities that modify the physical environment are increasing globally (Alley et al., 2003; Broecker, 1997).

Another important reason for accurately characterizing the deep ocean is the need to validate ocean carbon cycle models. These models are used frequently to predict the response of the ocean to increasing atmospheric \(\text{CO}_2\). Projections of future carbon uptake by the ocean (Houghton et al., 2001) inevitably involve the deep ocean. We would be hard pressed to place confidence in projections from any model that does not reproduce the modern deep ocean behavior reasonably well.

In this work, we use the new measurements from the World Ocean Circulation Experiment (WOCE) conducted during the decade of 1990s to present for the first time objectively gridded global maps of deep natural \(^{14}\text{C}\). Objective gridding is
possible with the 15,000 WOCE measurements (e.g., Key et al., 1996, 2002) except in the Atlantic where the data set is supplemented with results from the late 1980s. In contrast, there are approximately 2,200 GEOSECS $^{14}$C data, which despite the global coverage lack sufficient spatial resolution to make objective mapping meaningful. In addition, we compare the observed deep $^{14}$C distribution with results from three selected Princeton Ocean Biogeochemistry Models. The three models, which are part of a larger family of Princeton models, serve to illustrate a wide range in model behavior (Gnanadesikan et al., submitted; Matsumoto et al., in press).

2. DATA AND MODELS

2.1 Data

During WOCE two analytical methods were used for $^{14}$C. The first, known as the large volume technique (Key, 1991; Key et al., 1991; Stuiver et al., 1974), was identical to that used for GEOSECS and required a 250 liter water sample. Inorganic carbon was extracted from the sample at sea and returned to shore based laboratories for measurement by beta-decay counting. The second, the small volume technique (Key, 1996; McNichol et al., 1994), only required a 250 milliliter water sample. These water samples were returned to a shore base laboratory and the $^{14}$C content determined by actually counting atoms using an accelerator-mass spectrometer combination (AMS). The reproducibility of both methods is about 4‰ (Elder et al., 1998). Large volume sampling during WOCE was restricted to Pacific Ocean deep waters. Once the measurements were completed, the results were subjected to stringent quality control procedures and subsequently merged with other results.

During the 1950s and 1960s the United States and the former Soviet Union carried out a significant number of large atmospheric nuclear weapon tests. Radiocarbon produced by these tests is commonly referred to as “bomb-produced radiocarbon” or simply “bomb $^{14}$C”. In the atmosphere bomb-$^{14}$C reacts with oxygen to form $^{14}$CO$_2$ which subsequently enters the ocean primarily by air-sea gas exchange. This bomb $^{14}$C contamination raised surface ocean radiocarbon values from approximately –60‰ to as high as 250‰ in some locations. Over time the bomb $^{14}$C contamination is mixed into the ocean interior. Here we are focusing on distribution of natural $^{14}$C (also referred to as background $^{14}$C or pre-bomb $^{14}$C) therefore the natural and bomb components which contribute to the measured values in the upper ocean must be separated. We have used the separation method described by Rubin and Key (2002), which is based on the strong linear correlation between natural $^{14}$C and potential alkalinity (PALK). The potential alkalinity separation method is outlined by the following equations:

\[
\text{Measured } ^{14}\text{C} = \text{Natural } ^{14}\text{C} + \text{Bomb } ^{14}\text{C},
\]

\[
\text{Natural } ^{14}\text{C} = -0.59 - 0.962(\text{PALK} - 2320),
\]
During the 1990s the bomb-produced radiocarbon signal had not penetrated significantly below 1000 m except in the North Atlantic through the formation of North Atlantic Deep Water and immediately adjacent to Antarctica wherever Antarctic Bottom Water is formed (Weddell Sea, Ross Sea, Adele Coast, etc.). Ostlund and Rooth (1990) presented a dramatic demonstration of the continuing penetration of the bomb signal into the Northwest Atlantic by comparing GEOSECS data to measurements taken in the 1980s as part of the Transient Tracers in the Ocean (TTO) expedition. In order to eliminate this contamination, all of the data and figures in this work are natural $^{14}$C derived from the equations above.

Production of the mapped distribution on a depth surface is a two step procedure. At each sampling station the value at the depth of the surface is estimated by interpolation. The maximum distance over which interpolation is allowed varies with the depth of the surface. That is, the measurements must be sufficiently close to the surface or the interpolation is disallowed. Next, the values on the surface are interpolated horizontally using an objective analysis scheme (Sarmiento et al., 1982). For surfaces 3500 m and shallower, the correlation length scales used for the mapping are 1550 km east to west and 740 km north to south. Below 3500 m both were set to 740 km. The relatively sparse data combined with long correlation length scales results in maps with significantly smoothed features. That is, all extrema are reduced and features such as boundary currents are broadened.

2.2 Models

All three Princeton models used here are based on the same MOM3 (GFDL Modular Ocean Model) code and have identical model resolution (3.5° latitude × 4.5° longitude) and boundary conditions (Gnanadesikan et al., 2002). The models have a prescribed depth profile of vertical eddy diffusion, represented by a diffusion coefficient $K_v$. Lateral mixing is oriented along isopycnal surfaces, using of the so-called GM parameterization or the eddy-induced transport velocity parameterization of Gent et al. (1995). Along-isopycnal mixing coefficient $A_I$ represents the lateral eddy diffusion. These models are distinctive largely on the basis of their $K_v$ and $A_I$.

The Low-Low (i.e., low $K_v$ and low $A_I$) model is nominally our “standard” model, from which other models are derived. The depth-dependent $K_v$ is 0.15 cm$^2$/sec in the upper ocean and 1.3 cm$^2$/sec in the deep ocean. There is a hyperbolic tangent transition at 2500 m to connect the two $K_v$ values (Bryan and Lewis, 1979). Everywhere $A_I$ is fixed at 1000 m$^2$/sec. In the High-High model, both the upper ocean $K_v$ and global $A_I$ are increased from the Low-Low model values to 0.6 cm$^2$/sec and 2000 m$^2$/sec respectively. The particular combination of $K_v$ and $A_I$ in the High-High model largely maintains the general shape and depth of the pycnocline of the standard model but changes the advective pathway of the deep ocean (Gnanadesikan, 1999). The High-High model returns the deep
water to the surface by upwelling more in the low latitudes, whereas the return path occurs predominantly in the Southern Ocean in the standard model. As we show below, the deep natural $\Delta^{14}C$ distributions predicted by these two models are not satisfactory. Partly in an effort to rectify the situation, the third model was developed (Gnanadesikan et al., submitted). In the Med-Low model, the upper ocean $K_v$ has an intermediate value of $0.3 \text{ cm}^2/\text{sec}$ with an inflection depth of 2000 m and the low $\alpha_i$ value of 1000 m $^2/\text{sec}$. However, south of 55°S, $K_v$ throughout the water column is set to 1.0 cm$^2$/sec. This results in a more vigorous vertical mixing in the Southern Ocean. Additionally, the Medium-Low model is forced to have convection during three winter months at four grid boxes in the Ross and Weddell Seas. This improves the temperature and salinity properties in the deep Southern Ocean.

These circulation models are coupled to the common “Abiotic” biogeochemistry code from the second phase of Ocean Carbon Cycle Intercomparison Project (OCMIP) (Orr et al., 2001). The code standardizes air-sea gas exchange, seawater CO$_2$ system, and surface fresh water flux (see HOW-TO documents on the OCMIP web site at www.ipsl.jussieu.fr/OCMIP). State variables include temperature, salinity, CO$_2$, and $^{14}$CO$_2$. Surface alkalinity is scaled to surface salinity. Natural $^{14}$C results are taken after the models have achieved equilibrium.

3. DEEP NATURAL RADIOCARBON

An objectively gridded global map of natural radiocarbon abundance at 3500 m water depth shows, as expected, the highest (i.e., the least negative) $\Delta^{14}C$ in the North Atlantic (Fig. 1). The highest $\Delta^{14}C$ near Labrador Sea is about than $–50‰$, which is roughly the preindustrial surface ocean $\Delta^{14}C$ value, as indicated by $^{14}$C measurements on samples of sea water, corals, mollusks, and planktonic foraminifera (Bard, 1988). In conventional $^{14}$C age, $–50‰$ is approximately 400 years and reflects the apparent age or the reservoir age of the surface ocean. If the air-sea gas exchange were infinitely fast, the preindustrial surface $\Delta^{14}C$ would be identical to the preindustrial atmospheric $\Delta^{14}C$ of 0‰. That the deep North Atlantic at 3500 m has natural $\Delta^{14}C$ that is close to the preindustrial surface value is evidence for the deep vertical mixing in the Labrador Sea and Greenland-Iceland-Norwegian Sea that is associated with the formation of North Atlantic Deep Water (NADW). On this 3500 m level, the deep Northeast Pacific has the most negative $\Delta^{14}C$ of approximately $–250‰$, indicating the terminus of the conveyor belt circulation. The Southern Ocean has an intermediate $\Delta^{14}C$ of about $–160‰$ at this level.

The long time scale associated with the deep ocean circulation are readily appreciated when $^{14}$C abundance is converted to $^{14}$C age (Fig. 2). The difference of about 1000 years between the Atlantic and Pacific is the basis for characterizing the global overturning circulation as having millennial time scale. At this slow rate of circulation, sharp tracer gradients are generally eroded by diffusive mixing. This and the 1550 km east-west and 740 km north-south correlation
Fig. 1. Objectively mapped natural $^{14}$C abundance on the 3500 m water depth level. Radiocarbon abundance is expressed in $\Delta^{14}$C (‰). See Subsection 2.1 for data source and mapping methodology. Contours are 25‰ apart. Dots indicate station locations.

length scales used in the objective mapping account for the smoothness in the maps of deep natural $\Delta^{14}$C and $^{14}$C age (Figs. 1 and 2).

We also produced a “bottom” map of natural $\Delta^{14}$C shown in Fig. 3. In making this figure, rather than vertically interpolating the values at each station, the deepest datum from each station was used with the constraints that the sample is within 250 m of the seafloor and the seafloor is at least 3500 m. These values were then objectively horizontally mapped. As in Fig. 1, the general pattern of the deep limb of the global thermohaline circulation is depicted in this bottom map. However unlike the map on 3500 m where there is relatively little topographic barriers, Fig. 3 shows the paths of the bottom and abyssal waters as they are guided by the bottom topography. Circumpolar Deep Water, and to a topographically limited extent Antarctic Bottom Water, generally move northward adjacent to the bottom filling both the Indian and Pacific Oceans. In the Pacific the primary core of the northward flow is adjacent to the island arch system near the date-line. Careful examination of the abyssal Pacific flow implies that this northward flowing core bifurcates north of the equator with the majority following a general clockwise flow along the west and north edges of the basin. The other tongue is difficult to follow, but it appears to flow northeastward passing south of the Hawaiian Islands. In the abyssal Indian Ocean the northward flow of Circumpolar Deep Water is strongly controlled by topography. The oldest abyssal water found in the Indian Ocean is in the Central Indian Basin which is filled through gaps in the 80 East Ridge at both the north and south ends of the basin.
Ocean general circulation models generally do not simulate the deep natural radiocarbon distribution well (Gnanadesikan et al., submitted; Matsumoto et al., in press). For direct comparison with the observed horizontal map of natural $\Delta^{14}C$ at the 3500 m level (Fig. 1), we show the same for the three Princeton models in Fig. 2. Objectively mapped conventional $^{14}C$ age of natural radiocarbon on the 3500 m level. This figure does not correlate exactly with natural $\Delta^{14}C$ (Fig. 1), because $^{14}C$ age is a non-linear function of $^{14}C$ abundance. Also, in making this figure, $\Delta^{14}C$ from Fig. 1 was first converted to $^{14}C$ age, which was then objectively mapped (i.e., converted then mapped, not mapped then converted). Therefore the mapped variable is different in the two figures. Contours are 100 years apart.

Fig. 3. Objectively mapped natural $^{14}C$ abundance within 300 m from the seafloor and deeper than 3500 m. Radiocarbon abundance is expressed in $\Delta^{14}C$ (‰).
Fig. 4. These horizontal maps together with meridional sections of the Pacific (Fig. 5) clearly illustrate the extent of the model-data mismatch in the Pacific. The Low-Low model, which has low Kv, is generally too stratified and as a consequence has too negative $\Delta^{14}C$ everywhere in the deep ocean (Fig. 4(a)) when

Fig. 4. Simulated natural $^{14}C$ abundance on the 3500 m level. Contours are 10‰ apart. Note the different color scales used in this figure and Fig. 1. Radiocarbon abundance is expressed in $\Delta^{14}C$ (‰).

Fig. 5. Observed and modeled meridional sections of natural $^{14}C$ abundance in the Pacific. These sections represent basin-wide zonal mean. Contours are 30‰ apart. Radiocarbon abundance is expressed in $\Delta^{14}C$ (‰).

Fig. 4. These horizontal maps together with meridional sections of the Pacific (Fig. 5) clearly illustrate the extent of the model-data mismatch in the Pacific.

The Low-Low model, which has low Kv, is generally too stratified and as a consequence has too negative $\Delta^{14}C$ everywhere in the deep ocean (Fig. 4(a)) when
compared to the observed (Fig. 1). Poor ventilation in this model’s deep ocean is most prominent in the North Pacific, where Δ\(^{14}\)C is almost −370‰e, which is more than 100‰e too negative or about 1000 years too old. This is largely a result of the lack of deep vertical mixing in the Southern Ocean in this model in contrast to the real ocean (Fig. 5). Also, the location of the oldest water in this model is found in the northwestern margin of the Pacific (Fig. 4(a)), on the opposite side of the basin from the observed. Furthermore, whereas the core of the North Pacific Deep Water (NPDW) is located in the depth range of 2000–3500 m according to where we find the most negative natural Δ\(^{14}\)C (Fig. 5(a)), the depth range of NPDW is less well defined in the Low-Low model (Fig. 5(b)).

The High-High and Med-Low models, which have higher vertical mixing coefficients than the Low-Low model, are able to ventilate the deep Pacific to a greater extent everywhere (Figs. 4(b) and (c)). In fact the High-High model has too positive Δ\(^{14}\)C in the deep North Pacific compared to the observed, while the Med-Low model has about the right NPDW Δ\(^{14}\)C. As in observation, these two models have deep vertical mixing in the Southern Ocean and a tongue of relatively well ventilated (more positive Δ\(^{14}\)C) bottom water moving from the deep Southern Ocean northward in the Pacific sector (Figs. 3 and 5). Also, these models have relatively well defined NPDW core that is consistent with observation.

Perhaps the most puzzling feature that is common in all of the horizontal maps shown in Fig. 4 is the apparently missing NADW. In the real ocean on the 3500 m level, there is a north-south natural Δ\(^{14}\)C gradient in the Atlantic basin of almost 100‰e that indicates the presence of NADW (Fig. 1). However, none of the models show such a large gradient; it is perhaps only 20‰e. This does not mean of course that NADW does not exist at all in these models. Such a gross model deficiency would make any ocean general circulation models completely inadequate. In these as well as in most other coarse resolution models, NADW exists at a much shallower depth range (Doney and others, in prep.). Meridional sections of natural \(^{14}\)C and salinity in the Atlantic basin from the Low-Low model show that the core of NADW is located in depth of about 2000 m and is largely absent at 3500 m (Fig. 6). This feature is essentially the same in the other two models. In the real ocean, NADW is found between 1500 and 4000 m. The excessively shallow NADW found in these and other ocean general circulation models has to do with their inability to form deep waters with the appropriate temperature and salinity properties. Our analysis of the thirteen ocean circulation models participating in the second phase of OCMIP (Dutay et al., 2002), which is not shown here, indicates that some models do not have waters as dense as NADW anywhere in their model domain.

The presence of NADW in these models at a shallower depth is readily confirmed by examining the natural \(^{14}\)C abundance at 2000 m (Fig. 7). In all models, we find as large a north-south gradient in natural Δ\(^{14}\)C as observed. Also, the southward advective pathway of NADW along the western boundary is clearly evident. Unlike the deep Pacific, where intermodel difference in natural Δ\(^{14}\)C was large, the Atlantic at 2000 m does not show large Δ\(^{14}\)C difference. The exception is in the far south, east of the Drake Passage, where the influence of
Fig. 6. Simulated meridional sections of natural $\Delta^{14}$C and salinity in the Atlantic. These sections represent basin-wide zonal mean from the Low-Low model. The other two models likewise show shallow NADW (see text). Radiocarbon abundance and salinity are expressed respectively in $\Delta^{14}$C and PSU, both in unit of ‰.

Fig. 7. Simulated natural $^{14}$C abundance on the 2000 m level in the Atlantic basin. Radiocarbon abundance is expressed in $\Delta^{14}$C (‰).
Circumpolar Deep Water is visible in the case Low-Low model (Fig. 7(a)). That there is a smaller intermodel natural $\Delta^{14}C$ difference in the deep Atlantic than in the deep Pacific is not surprising, because NADW was in contact with the atmosphere more recently than NPDW. Since the preindustrial atmosphere by definition has a fixed $\Delta^{14}C$ of 0‰, any water mass that was at the surface recently would have had its $\Delta^{14}C$ reset by the atmosphere, subject to the extent of air-sea gas equilibration. In contrast, NPDW is farthest removed from the atmosphere. Therefore, there is more “time” for model errors to accumulate.

4. DISCUSSION AND CONCLUSIONS

The three ocean carbon cycle models presented here exhibited a wide range in how well they could simulate the modern deep ocean natural $\Delta^{14}C$. One model ventilated the deep ocean excessively, another insufficiently, and the other roughly right. Because of this difference, we expect dissimilar responses by these models to anthropogenic carbon perturbation over centennial time scale that involves the deep ocean.

Figure 8 illustrates the three models’ cumulative oceanic uptake of anthropogenic carbon when forced only by changing atmosphere pCO$_2$. That is,
the models remain unchanged in their circulation and boundary conditions. The atmospheric forcing function is a combination of past pCO$_2$ changes as measured on samples of ice core air bubbles and air and a future stabilization scenario. The simulated anthropogenic carbon inventories in all three models increase with the rising atmospheric pCO$_2$. Oceanic uptake continues to increase even after the atmospheric pCO$_2$ is stabilized at 650 ppm in the year 2200. There is a continued net transfer of carbon from the atmosphere to the ocean, because the atmospheric pCO$_2$ is apparently higher than the areally weighted mean surface ocean pCO$_2$ during the entire simulation period.

The cumulative uptake is highest at all times in the High-High model, which has the largest Kv of all the models. The Med-Low and Low-Low models have successively less anthropogenic carbon inventory with decreasing strength of vertical mixing. A higher Kv ventilates the deep ocean more efficiently and thereby brings to the surface waters uncontaminated with anthropogenic signal. These “fresh” waters are able to absorb more anthropogenic carbon. The most stratified Low-Low model has the least inventory, because it is unable to rapidly mix down the surface waters that have absorbed anthropogenic carbon. Therefore, the surface concentration of anthropogenic carbon in this model is the highest.

The predicted anthropogenic carbon inventories diverge more with time, because the oceanic uptake increasingly involves the deep ocean. In the earlier part of the simulation period, uptake only occurs in the surface mixed layer and shallower isopycnal layers, which by their relatively small volumes inherently has a limit to how much uptake can occur. Also, the atmospheric pCO$_2$ and hence air-sea pCO$_2$ gradient that drives the carbon flux are still relatively low. For these reasons, intermodel difference is relatively small. In the year 2000, the three models’ inventories are roughly within 20 Pg-C of about 135 Pg-C. By the end of the simulation period, the High-High model has almost 1000 Pg-C compared to 750 Pg-C in the Low-Low model.

It should be emphasized that these results were obtained from ocean-only models and not from coupled models. Therefore, these results do not account for possible changes in surface ocean temperatures, alkalinity, and stratifications in the future that all affect anthropogenic carbon uptake. For example, surface water warming would reduce the solubility of CO$_2$. Surface warming would also increase the buoyancy of the surface waters, which would strengthen stratification and reduce vertical mixing. Additional changes in the surface buoyancy can be expected from changes in the hydrological cycle. An increasingly acidic ocean (Caldeira and Wickett, 2003) would begin to dissolve CaCO$_3$ accumulating on the sea floor and thereby increase the oceanic alkalinity budget and hence CO$_2$ buffer capacity. All these effects are neglected in our illustration of how different models respond to a scenario of future atmospheric pCO$_2$ forcing (Fig. 8).

While we noted above that the Med-Low model has the best deep natural $\Delta^{14}$C distribution of the three models, we should not necessarily assume that the future uptake results from this model would be the most accurate. These coarse resolution models have a number of simplifications and parameterizations of the real ocean physics. While the particular ways in which they are implemented may
be appropriate under the modern conditions, this may not be so under different boundary conditions in the future. Nevertheless, the utility of the observed natural radiocarbon distribution (Figs. 1, 2, 3 and 5) in validating ocean carbon cycle is not diminished, because there is even less confidence in the long-term future predictions of models that fail to simulate the observed. In fact, the new natural radiocarbon data from WOCE and other surveys offer perhaps the sole means to quantitatively evaluate on centennial time scale ocean carbon cycle models and identify the better models from the worst (Matsumoto et al., in press).

Acknowledgements—Financial support was provided by NSF grant OCE-9819144 to R. Key.

REFERENCES


K. Matsumoto (e-mail: katsumi@ni.aist.go.jp) and R. M. Key (e-mail: key@princeton.edu)