Can deep ocean carbonate preservation history inferred from atmospheric pCO₂ account for ¹⁴C and %CaCO₃ profiles on the Ontong–Java Plateau?

Katsumi Matsumoto a,b,*, Wallace S. Broecker a,b, Elizabeth Clark a, Daniel C. McCorkle c, William R. Martin c, Irka Hajdas d

a Lamont-Doherty Earth Observatory of Columbia University, Palisades, NY 10964, USA
b Department of Earth and Environmental Sciences, Columbia University, New York, NY, USA
c Woods Hole Oceanographic Institution, Woods Hole, MA 02543, USA
d ITP, ETH, Hoenggerberg, CH-8093 Zürich, Switzerland

Received 28 August 2000; accepted 13 June 2001

Abstract

Bulk ¹⁴C and ²¹⁰Pb excess measurements were made on box cores from water depths of ~2300 and ~4200 m from the equator on the Ontong–Java Plateau. A sediment mixed layer thickness of approximately 6 cm and a mixing rate of 20–45 cm²/kyr are inferred from these radionuclide measurements and model simulations of their depth profiles. Using these sediment mixing characteristics, the model is then forced with a deep sea carbonate preservation history derived from an ice core atmospheric pCO₂ record, in order to examine whether the derived history can account for depth profiles of ¹⁴C age and CaCO₃ content from the deep western equatorial Pacific. Results show that ¹⁴C and %CaCO₃ data from the deep site can be simulated simultaneously with some success using the assumed carbonate preservation history and a time-varying non-carbonate flux function. However, the model is unable to use the same forcing functions to simulate the ¹⁴C and %CaCO₃ data from the shallower core. These results suggest that local carbonate accumulation is not driven simply by generalized deep water carbonate chemistry inferred from atmospheric pCO₂. Potentially important local processes and aspects of the model that may have contributed to the results are also discussed. © 2001 Elsevier Science B.V. All rights reserved.

Keywords: C-14; Pb-210; bioturbation; deglaciation; Ontong–Java Plateau; deep-sea environment; sediments; Pacific Ocean

1. Introduction

The burning of fossil fuels since the beginning of the Industrial Revolution has increased the atmospheric CO₂ content from about 280 to over 360 ppm. This excess atmospheric CO₂ will be repartitioned amongst the accessible surface carbon reservoirs of biosphere, atmosphere, and hydrosphere. Of these, the hydrosphere (essentially...
the ocean) will eventually take up most of the excess CO$_2$ because of its sheer size. Over a time scale of thousands of years, the increasingly acidic ocean must then be neutralized by the vast amount of CaCO$_3$ in surficial sediments of the sea floor. This neutralization of fossil fuel CO$_2$ will in part depend on the rate of sediment mixing, which stirs the insoluble residue left after dissolution down deeper and the underlying CaCO$_3$ up into the dissolution zone.

While biologists have studied the mixing characteristics of marine sediments through behavior observations of single species of benthic animals (e.g., [1]), geochemists have measured downcore distributions of various tracers, which provide information on community-wide sediment mixing. Lithologic tracers such as tektites and volcanic ashes as well as radioactive tracers of wide-ranging half-lives including $^7$Be, $^{14}$C, $^{32}$Si, $^{210}$Pb, $^{228}$Th, and $^{234}$Th have been measured, which together with models provide estimates of mixing rate and mixed layer depth (see [2] for review).

In this study, we present downcore profiles of high-resolution $^{14}$C and excess $^{210}$Pb ($^{210}$Pb$_{ex}$) measurements from four box cores raised from the Ontong-Java Plateau (OJP) in the western equatorial Pacific. As with previous studies that have presented $^{14}$C data from OJP [3–6], we use the new $^{14}$C and $^{210}$Pb$_{ex}$ data together with a sediment mixing model to characterize the local sediment mixing.

In addition, a new feature in this study is to examine whether the depth profiles of bulk carbonate content (%CaCO$_3$) and $^{14}$C age from the deep Pacific can be explained by a history of carbonate accumulation as inferred from a record of atmospheric pCO$_2$. This was partly motivated by a recent demonstration that a natural 20–25 ppm rise in atmospheric CO$_2$ concentration over the course of the Holocene [7] is consistent with reconstructed carbonate ion content changes in the deep Atlantic and Pacific for the same period [8]. Here we force a simple sediment mixing model with an externally determined carbonate accumulation function that we believe is geologically reasonable on a global scale and ask whether the simulated results are consistent with local data.

### 2. Materials and methods

We sampled two ‘Soutar’ box cores from OJP in the equatorial Pacific for AMS radiocarbon analyses: MW91-9-B56 (0°, 162°E, 4040 m) and MW91-9-B36 (0°, 158°E, 2310 m). All $^{14}$C measurements were made on bulk carbonate at ETH in Zürich and were reservoir-age corrected by 400 years [9]. Bulk carbonate content for the same two cores was determined by the standard coulometric method at Lamont-Doherty Earth Observatory.

From the same region, we sampled two other ‘Soutar’ box cores, MW91-9-B28 (0°, 158°E, 2323 m) and MW91-9-B66 (0°, 163°E, 4440 m), for $^{210}$Pb$_{ex}$ analyses. Measurements were made at Woods Hole Oceanographic Institution following the method of Wheatcroft and Martin [10].

We constructed a sediment mixing model based on the equation governing the distribution of tracers subject to diffusional mixing, advection by sediment accumulation, and in situ reaction in a constant porosity sediment [11]:

$$\frac{\partial C}{\partial t} = \frac{\partial}{\partial z} \left( K(z) \frac{\partial C}{\partial z} \right) - S \frac{\partial C}{\partial z} + R(C, z, t)$$

where $C$ is the desired concentration of the tracer, $z$ is depth in the sediment, $t$ is time, $K(z)$ is diffusion coefficient or mixing rate, $S$ is sediment accumulation rate, and $R$ is the rate of local production or consumption of $C$ at $z$ and $t$. We follow the common practice and make the simplifying assumption that $K$ is constant within the finite depth of mixing $0 \leq z \leq L$. Once advected below the mixed layer, the sediment becomes part of the historical record that is free of diffusive mixing. In the case of $^{210}$Pb and $^{14}$C, $R$ is set to zero, because the model does not explicitly treat carbonate dissolution. We assume that dissolution takes place at the sediment–water interface [12,13] and before bioturbation has the chance to mix the carbonate rain into the sediment mixed layer. The model implicitly accounts for dissolution by varying the flux of fresh carbonate that is reservoir-age corrected by 400 years.
Here $^{14}$C refers to the abundance of radiocarbon atoms relative to the stable $^{12}$C atoms associated only with CaCO$_3$ and not with any other carrier of carbon atoms in the deep sea sediments. Since the model essentially moves around the mass of sediments, tracers in ratios must be converted to some kinds of concentration for them to be simulated properly. The $^{14}$C/$^{12}$C ratio is thus converted to a 'pseudo-concentration' by multiplying it by %CaCO$_3$. The model treats the pseudo-concentration as the tracer concentration $C$ in Eq. 1. The conversion ensures that the $^{14}$C/$^{12}$C ratio in each of the mixed layer grid boxes reflects not just a simple average of the original $^{14}$C/$^{12}$C ratios before being mixed but an average weighted by the abundance of its carrier. An application of pseudo-concentration is described for the sediment model in [3] but is commonly employed whenever $^{14}$C age is simulated as a passive tracer, such as in ocean biogeochemistry models of various sophistication (e.g., [14,15]). There is no need to do similar conversion for $^{210}$Pb$_{ex}$, since its unit in dpm/gdw (decays per minute per gram of dry weight) is already in concentration.

The vertical grid spacing is 1 cm and the time step is 5 years for $^{14}$C modeling. A 1 cm model resolution is higher than the measurement resolution. The 5 year time step is very small compared to the $^{14}$C half-life of 5730 years and represents a balance between our desire to have as large a time step as possible in order to minimize the simulation time and the short time step required by the 1 cm spatial resolution for numerical stability. Model time step is shortened to approximately 2 weeks and spatial resolution is increased to 0.1 cm for modeling $^{210}$Pb$_{ex}$, whose half-life is only 22.3 years. A 100 year simulation time for $^{210}$Pb$_{ex}$ is sufficient to reach an equilibrium regardless of its initial profile, and we are only seeking steady-state profiles for $^{210}$Pb$_{ex}$. For radiocarbon, we seek time-dependent profiles. Since we have no knowledge of $^{14}$C ages below the deepest

Fig. 1. Model sensitivity of $^{14}$C profile to changes in: (a) sediment accumulation rate $S$, (b) mixed layer depth $L$, and (c) mixing rate $K$. Here $K$ is constant within $L$. Note that the $^{14}$C gradient below the mixed layer is entirely a function of $S$, while the $^{14}$C gradient within the mixed layer is mostly sensitive to $K$ and weakly dependent on $L$. As reference, $^{14}$C measurements from BC36 (2310 m water depth) are shown with star marks and BC56 (4040 m) with circles; no effort was made to match the data.
measurements in BC36 and BC56, we assume as initial conditions equilibrium $^{14}$C profiles that are achieved with constant carbonate and non-carbonate fluxes that approximately yield the average sedimentation rates and the deepest $\%$CaCO$_3$ of both cores.

The model is stepped forward in time with adjustable parameters of $L$ (cm), $S$ (cm/kyr), and $K$ (cm$^2$/kyr). We present in Fig. 1 the sensitivity of the model to changes in these parameters. The mixed layer $^{14}$C profiles become more vertical with higher $K$, approaching the infinite mixing regime. In contrast, the mixed layer $^{14}$C gradient is only weakly dependent on $L$ and not sensitive at all to $S$ (unless $S$ is very large, in excess of \(\sim 30\) cm/kyr, according to the model). Below the mixed layer, however, the $^{14}$C profile is determined by $S$ only. The core top age is older for slowly accumulating sediments, deeper mixed layers, and higher $K$, all of which allow old sediments to be stirred upward.

We use the model for two main purposes. The first is to characterize sediment mixing on OJP at the core sites. For this, the entire depth profile of $^{210}\text{Pb}_{\text{ex}}$ concentration and mixed layer $^{14}$C gradient are simulated with the model as a function of $K$. The second is to examine whether sedimentary depth profiles of $\%$CaCO$_3$ and $^{14}$C age from the deep Pacific can be explained largely by the history of deep sea carbonate preservation as reflected in the atmospheric pCO$_2$ record.

The carbonate chemistry of seawater suggests that the deep sea carbonate ion content will be inversely related to atmospheric pCO$_2$. We assume that the global carbonate preservation rate will follow the time derivative of atmospheric pCO$_2$ history (Fig. 2), based on observation of deglacial preservation maxima in the deep sea [16,17]. There is as yet no broadly accepted explanation for the past variations in atmospheric pCO$_2$, which was 80–100 ppm lower during peak glacial periods compared to interglacial periods [18,19]. However, leading hypotheses all call on changes in the ocean carbon cycle that transfer CO$_2$ into the deep ocean by mechanisms such as enhanced biological pump and increased stratification (see recent reviews [20,21]). These mechanisms, working backwards during deglaciation, would allow the deep sea CO$_2$ to leak out to the atmosphere, raise the deep sea carbonate ion concentration, and deepen the lysocline [22]. If the riverine input of alkalinity is unchanged, a deepened lysocline would upset the ocean alkalinity balance in favor of its removal by CaCO$_3$ burial. This temporary imbalance is restored on a time scale of thousands of years as excess alkalinity removal over input brings down the deep sea carbonate ion content and shoals the lysocline. It is this process of carbonate compensation that is expected to yield a peak in carbonate preservation during termination that is captured by the time derivative of atmospheric pCO$_2$ history. We use the ice core pCO$_2$ measurements from Taylor Dome, Antarctica [7,23] to derive a carbonate preservation history since the last glaciation (Fig. 2).

3. Results

3.1. Characterizing sediment mixing

As expected for a short-lived radionuclide, $^{210}\text{Pb}_{\text{ex}}$ concentration in both BC28 and BC66 is highest near the sediment–water interface and quickly decays with increasing depth (Figs. 3 and 4). The actual core top concentration of BC66 is higher by a factor of four compared to
that of BC28; however, estimates of total inventory of sedimentary $^{210}\text{Pb}_{\text{ex}}$ at the two sites are not significantly different (Table 1).

The sedimentary $^{210}\text{Pb}_{\text{ex}}$ inventory was estimated here by integrating with depth the $^{210}\text{Pb}_{\text{ex}}$ activity weighted by downcore porosity measurements (Table 1). The total $^{210}\text{Pb}_{\text{ex}}$ inventory for BC28 is 23.1 dpm/cm$^2$ and that for BC66 is 25.7 dpm/cm$^2$; the difference is only about 10%. The total $^{210}\text{Pb}_{\text{ex}}$ inventory is independent of sediment mixing characteristics but reflects the overlying water column $^{210}\text{Pb}$ production (atmospheric deposition and in situ production from radioactive decay of the parent $^{226}\text{Ra}$) and the net removal of $^{210}\text{Pb}$ from the water column by surface scavenging and deep regeneration. These factors are difficult to quantify. Also, large uncertainties associated with lateral transport of $^{210}\text{Pb}$ by ocean circulation and by postdepositional sediment transport prevent quantitative interpretation of the 10% $^{210}\text{Pb}_{\text{ex}}$ inventory difference at the two

Table 1

<table>
<thead>
<tr>
<th>$^{210}\text{Pb}$ measurements</th>
</tr>
</thead>
<tbody>
<tr>
<td>Box core 28 (2323 m water depth)</td>
</tr>
<tr>
<td>Core depth (cm)</td>
</tr>
<tr>
<td>0.0–0.2</td>
</tr>
<tr>
<td>0.2–0.4</td>
</tr>
<tr>
<td>0.4–0.6</td>
</tr>
<tr>
<td>0.6–0.85</td>
</tr>
<tr>
<td>0.85–1.0</td>
</tr>
<tr>
<td>1.0–1.2</td>
</tr>
<tr>
<td>1.2–1.4</td>
</tr>
<tr>
<td>1.4–1.6</td>
</tr>
<tr>
<td>1.6–1.9</td>
</tr>
<tr>
<td>1.9–2.1</td>
</tr>
<tr>
<td>2.1–2.5</td>
</tr>
<tr>
<td>2.5–3.0</td>
</tr>
<tr>
<td>3.0–3.5</td>
</tr>
<tr>
<td>3.5–4.0</td>
</tr>
<tr>
<td>4.0–4.5</td>
</tr>
<tr>
<td>4.5–5.0</td>
</tr>
<tr>
<td>5.0–6.0</td>
</tr>
<tr>
<td>6.0–7.0</td>
</tr>
<tr>
<td>7.0–8.0</td>
</tr>
<tr>
<td>8.0–9.0</td>
</tr>
<tr>
<td>9.0–10.0</td>
</tr>
</tbody>
</table>

From all BC28 measurements, 1.5 dpm/gdw was subtracted. Likewise, 6 dpm/gdw was subtracted from all BC66 measurements. Using 2.7 g/cm$^3$ as the mineral density, the $^{210}\text{Pb}_{\text{ex}}$ inventory for BC28 is 23.12 dpm and for BC66 it is 25.72 dpm. Assuming steady state, the flux of $^{210}\text{Pb}_{\text{ex}}$ to the sea floor is 0.72 and 0.80 dpm/cm$^2$/yr for BC28 and BC66 respectively. See text.
sites, except to note simply that the integrated effects of these factors that determine the sedimentary $^{210}$Pb$_{ex}$ inventory must be similar at the two sites.

Assuming steady state for sedimentary $^{210}$Pb$_{ex}$ inventory, we can estimate the supply rates of $^{210}$Pb$_{ex}$ at the two sites. A $^{210}$Pb$_{ex}$ inventory of 23.1 dpm/cm$^2$ for BC28 requires a flux of 0.72 dpm/cm$^2$/yr to maintain it at steady state. For BC66, the required $^{210}$Pb$_{ex}$ flux is 0.80 dpm/cm$^2$/yr. For comparison, we have computed equivalent fluxes from $^{210}$Pb measurements in the water column, sediments, and sediment traps in the Pacific Ocean (Fig. 5). The different types of measurements require dissimilar sets of assumptions to calculate the $^{210}$Pb flux [24], and these measurements come from different regions of the Pacific. Nevertheless, it is clear from Fig. 5 that most of the estimated $^{210}$Pb fluxes lie within 0.2–1.4 dpm/cm$^2$/yr, especially near the equatorial region, and that the $^{210}$Pb$_{ex}$ fluxes estimated in this study are consistent with the compiled fluxes.

Model simulations of $^{210}$Pb$_{ex}$ data using the calculated $^{210}$Pb$_{ex}$ fluxes as inputs indicate that the appropriate mixing rate $K$ is about 45 cm$^2$/kyr for the shallower core BC28 (Fig. 3) and 20 cm$^2$/kyr for the deeper core BC66 (Fig. 4).

These rates yield the smallest mismatch to the observation in the least squared sense and are largely independent of both $S$ and $L$, because for a (radioactive) mean life of $^{210}$Pb (32 years), particle mixing dominates over advection and decay. In 32 years, sedimentation by a typical $S$ at OJP of 2–3 cm/kyr (Table 2) advects signal down by 0.06–0.10 cm. By comparison, the random walk distance traveled by diffusion during the same period of time given by $\sqrt{K}$-time for a reasonable $K$ of 30 cm$^2$/kyr is 0.98 cm. The effect of $L$ on $^{210}$Pb$_{ex}$ model output is also minimal given that a typical $L$ of 9.8 cm [2] is much too large compared to the distance that a $^{210}$Pb$_{ex}$ atom can travel during its typical life time either by particle mixing or by advection or combined.

We have also estimated $K$ from simulations of the mixed layer $^{14}$C gradient (Fig. 1), which like the modeled $^{210}$Pb$_{ex}$ profiles is largely a function of $K$. The results show that simulation mismatch (error) from observation is minimal when $K$ is approximately 25 cm$^2$/kyr for BC36 and when $K$ is about 40 cm$^2$/kyr for BC56 (Fig. 6). These estimates of $K$ agree reasonably well with the earlier estimates from $^{210}$Pb$_{ex}$ data, especially in light of the fact that the range of $K$ values reported in the literature ($10^{-1}–10^5$ cm$^2$/kyr) is six orders of magnitude [2].
3.2. Model simulations forced by carbonate preservation function

For our simulations of the depth profiles of %CaCO₃ and ¹⁴C age from BC36 and BC56, we set the mixed layer depth $L$ equal to 6 cm for both cores, based on the observed ¹⁴C profile in each core (Figs. 7 and 8). The choice of $L$ does not affect the results of our mixed layer ¹⁴C simulations, but we note that our 6 cm estimate is smaller than the 9 cm and 10 cm values calculated previously [25] using:

$$h = \tau A \left[ e^{(t/\tau)} - 1 \right]$$  \hspace{1cm} (2)

where $h$ is the desired depth (equivalent to $L$), $\tau$ is mean life of the ¹⁴C atom, $t$ is core top age, and $A$ is sediment accumulation rate. This equation does not account for $K$ explicitly and implicitly assumes very high mixing rates in the range of 100 cm²/kyr. For lower $K$, as suggested here, Eq. 2 overestimates $h$.

For the mixing rates $K$, we used the best estimates from simulations of the mixed layer ¹⁴C gradient for each core (Fig. 6) with the added constraint that the estimated $K$ not exceed the $K$ obtained from simulations of the ²¹⁰Pbex profiles (Figs. 3 and 4). This constraint is based on the observation of a negative relationship between $K$ and the timescale of the radioactive tracer used for the mixing rate estimate [26,27], because selective feeding by animals on recently rained, organic-rich sediments tend to preferentially mix shorter-lived radiotracers. We thus chose $K$ values of 25 cm²/kyr for BC36 (the ¹⁴C-derived rate, which was lower than the ²¹⁰Pbex-derived rate), and 20 cm²/kyr for BC56 (where the ²¹⁰Pbex rate was lower than the ¹⁴C-derived rate of 40 cm²/kyr). We kept $K$ constant through each model run.

Using the carbonate flux history derived from atmospheric pCO₂ record (Fig. 2), we modeled the depth profiles of %CaCO₃ and ¹⁴C age at the water depth of this core (4040 m) is supported by a recent deep sea carbonate ion content reconstruction [8], which used the coarse fraction of carbonate content as paleocarbonate ion proxy. Calibration study of this proxy indicates that it is sensitive to the ambient seawater carbonate ion content in modern seawater at approximately 4000 m and deeper but rather insensitive at shallower water depths [28]. Therefore, since the correlation between atmospheric pCO₂ and carbonate ion content at the water depth of this core (4040 m) is supported by a recent deep sea carbonate ion content reconstruction [8], which used the coarse fraction of carbonate content as paleocarbonate ion proxy. Calibration study of this proxy indicates that it is sensitive to the ambient seawater carbonate ion content in modern seawater at approximately 4000 m and deeper but rather insensitive at shallower water depths [28]. Therefore, since the correlation between atmospheric pCO₂ and carbonate ion content at the water depth of this core (4040 m) is supported by a recent deep sea carbonate ion content reconstruction [8], which used the coarse fraction of carbonate content as paleocarbonate ion proxy. Calibration study of this proxy indicates that it is sensitive to the ambient seawater carbonate ion content in modern seawater at approximately 4000 m and deeper but rather insensitive at shallower water depths [28]. Therefore, since the correlation between atmospheric pCO₂ and carbonate ion content is unclear at the water depth of the shallower core BC36 (2310 m), we did not use the carbonate flux history derived from atmospheric pCO₂ to simulate the %CaCO₃ and ¹⁴C data from BC36.

Table 2
AMS radiocarbon measurements

<table>
<thead>
<tr>
<th>Box core 36 (2310 m water depth)</th>
<th>Box core 56 4040 m water depth</th>
</tr>
</thead>
<tbody>
<tr>
<td>Core depth (cm)</td>
<td>¹⁴C age (years)</td>
</tr>
<tr>
<td>0–1</td>
<td>2785±55</td>
</tr>
<tr>
<td>2–3</td>
<td>2965±60</td>
</tr>
<tr>
<td>4–5</td>
<td>3250±55</td>
</tr>
<tr>
<td>6–7</td>
<td>3560±60</td>
</tr>
<tr>
<td>8–9</td>
<td>4280±60</td>
</tr>
<tr>
<td>10–11</td>
<td>4530±60</td>
</tr>
<tr>
<td>12–13</td>
<td>4320±60</td>
</tr>
<tr>
<td>14–15</td>
<td>5050±60</td>
</tr>
<tr>
<td>19–20</td>
<td>7785±65</td>
</tr>
<tr>
<td>29–30</td>
<td>9620±90</td>
</tr>
</tbody>
</table>

All ages are corrected by 400 years to account for the reservoir effect. Overall sediment accumulation rates can be estimated using the deepest datum in each core: 3.1 cm/kyr for BC36 and 1.8 cm/kyr for BC56.
Shown in Fig. 7 are results from two simulations of BC56 using the derived carbonate preservation flux function. Profiles of the control simulation (solid lines) show as good a fit as one can obtain to the 14C data using a constant non-carbonate flux function. The carbonate preservation flux function here is scaled arbitrarily, but there is only a very limited range in scaling that is appropriate to be consistent with the overall sediment accumulation rate. With a constant non-carbonate flux function, the model underestimates the mixed layer age by several hundred years and the deepest 14C age by almost 2500 years. Most significantly, the control run is entirely unsatisfactory with regard to the %CaCO3 data. We therefore allow the non-carbonate flux to change with time to better match the %CaCO3 data in the second simulation. With a slight increase in non-carbonate flux prior to the Holocene and a slight decrease in the early Holocene (Fig. 7c), the model is able to simulate well the %CaCO3 data and to improve somewhat the match to the 14C profile (dashed lines), although the deepest 14C datum is still underestimated by ~2000 years.

We now use this time-varying non-carbonate flux function in modeling the shallow core BC36, because the detrital flux at each core site should be similar, barring any postdepositional sediment transport, since all the cores were raised from a limited geographical region. Here we show two simulations using two distinct histories of carbonate preservation, each trying to match the 14C and %CaCO3 data (Fig. 8). The results show that...
while the model is able to simulate each set of measurements with some success, though not entirely satisfactory especially with regard to $^{14}$C data, it is unable to simulate both sets simultaneously. These two simulations in some sense represent the endmember simulations. Other simulations fall on a spectrum defined by these two simulations and show that a better match to either the $^{14}$C or the $\%$CaCO$_3$ data compromises the match to the other data set.

Although we do not show results, we note for the sake of completeness that the model is unable to simulate simultaneously both $^{14}$C and $\%$CaCO$_3$ data from BC36 using the carbonate flux function scaled to the time derivative of atmospheric pCO$_2$ record, regardless of the history of non-carbonate flux. The model likewise cannot simulate simultaneously the two sets of data from both cores using carbonate flux functions scaled directly to an interpolated atmospheric pCO$_2$ record rather than to its time derivative. Although the physical meaning of the latter experiment is unclear, as we expect a deglacial carbonate preservation that is approximated by the time derivative \cite{16,17}, we mention these experiments to reinforce our point that the model cannot match both $^{14}$C and $\%$CaCO$_3$ data at the same time.

4. Discussion and conclusions

Simulations presented here examined whether depth profiles of sedimentary $\%$CaCO$_3$ and $^{14}$C age from OJP can be accounted for by the carbonate preservation history derived from atmospheric pCO$_2$ record. The results are not encouraging.

The model can simulate simultaneously both the $\%$CaCO$_3$ and $^{14}$C data from the deep core BC56 with some success using the derived carbonate flux function, if the non-carbonate rain rate is allowed to vary with time (Fig. 7, dashed lines). However, the model is unable to use the same variations in non-carbonate flux and simulate both the $\%$CaCO$_3$ and $^{14}$C profiles from the shallower core BC36, whatever the carbonate flux function may be (Fig. 8). This indicates that non-carbonate, mostly clay materials, did not accumulate in a similar manner at these two sites despite being located in essentially the same geographical area. Some kind of postdepositional sediment transport, such as winnowing or sediment focusing, is required to account for the dissimilar detrital accumulation.

Also, the model can simulate the $\%$CaCO$_3$ and $^{14}$C profiles from BC36 with some success only individually, using two completely dissimilar carbonate flux functions (Fig. 8). While it is reassuring that these carbonate fluxes at BC36 are in general higher than at the deeper site of BC56, which we expect from the increase in carbonate undersaturation with depth, these two carbonate accumulation functions are unlike the carbonate preservation function derived from atmospheric pCO$_2$ record. The dissimilarity indicates that the local carbonate accumulation is not driven simply by the seawater carbonate chemistry inferred from atmospheric pCO$_2$ record, which is a global signal. There must be some local processes that mask the global imprint.

Potentially important local processes that we have not accounted for in the model include in situ carbonate dissolution driven by respiration CO$_2$ (e.g., \cite{29,30}), variation in surface productivity, and postdepositional sediment transport. Also, the assumption that mixing rate $K$ is constant with time and in space is not strictly valid, as there may exist a feedback between food abundance, its decomposition, and the intensity of bioturbation \cite{31}. The effect of variable $K$ on the depth profiles of $\%$CaCO$_3$ and $^{14}$C below the mixed layer is limited though (Fig. 1). Finally, the governing equation (Eq. 1) which formulates bioturbation as diffusional mixing is not mechanistic. Vertical mixing by animals in reality is not a diffusive process but advective, although the use of Eq. 1 is justified in the sense that the equation has been successful at fitting geochemical data over the years and that rapid horizontal mixing coupled with vertical advection can produce vertical profiles characteristic of diffusion \cite{32}. Nevertheless, the separate treatment of $K$ and $L$ is artificial \cite{31}, as both characteristics result from the same activities of benthic deposit feeders \cite{33}.

We conclude by noting that, despite the fact that our model experiments could not account for the $\%$CaCO$_3$ and $^{14}$C data from the shallower
core BC36, it is still possible that some combination of clay and carbonate accumulation functions may yield reasonable fits to both data. This can only be determined in an inverse model. While this work indicates that such functions have little to do with deep sea carbonate preservation linked to atmospheric pCO₂, they may offer interesting insights into which local processes are important in the formation of deep sea sediments.

Acknowledgements

We thank M. Fleisher, T.-H. Peng, and J. Shepherd for useful discussions at early stages of this study. Robert F. Anderson kindly provided unpublished sediment trap ²¹⁰Pb data from JGOFS Equatorial Pacific study and was very helpful in explaining subtle aspects of ²¹⁰Pb to the senior author. We thank Robin Keir and an anonymous reviewer for critical reading of the manuscript. Financial support was provided by NSF Grants ATM9730546 to W.S.B., OCE9730932 to D.C.M., and OCE9810962 to W.R.M. This is LDEO contribution 6170 and WHOI contribution 10434.

References


