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DISSOLVED ORGANIC MATTER AND THE GLACIAL-INTERGLACIAL pCO₂ PROBLEM

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Abstract. Two box models, one analytical and one numerical, are used to investigate systematically a broad range of oceanic circulation changes on the atmospheric carbon dioxide (CO₂) concentration. A number of oceanic carbon cycle models have failed to reproduce the 30% increase in CO₂ partial pressure (pCO₂) during the last deglaciation as reconstructed from polar ice cores. We apply therefore this approach of exploring the model's parameter space to examine the effect of long-lived dissolved organic matter on the system. The results from the two models complement each other, in terms of insight versus detail. Carbon is usually assumed to be transported from the surface into the deep ocean through the sedimentation of particulate matter. If there exists in the ocean a pool of dissolved organic matter (DOM) with a regeneration time comparable with the advection time, then the associated carbon can also be advected, resulting in a different distribution. Such a DOM reservoir acts as a smoother on the spatial distribution of nutrients in the sea, particularly in the equatorial intermediate waters. We establish the role of intermediate waters as one of the key components of the oceanic carbon cycle and show that DOM reduces the sensitivity of the carbon cycle to oceanic circulation pattern changes, mainly because of its smoothing effect. Consequently, the possible existence of DOM species with a time constant of the order of a century tends to reduce, rather than enhance, the glacial-interglacial difference in pCO₂ levels due to changes in the thermohaline circulation.

1. INTRODUCTION

The analysis of air bubbles trapped in polar ice has revealed that the atmospheric concentration of carbon dioxide (CO₂) was approximately 30% lower during the last glacial maximum than during preindustrial time [Barnola et al., 1987; Delmas et al., 1980]. Such a decrease in greenhouse gas concentration may be responsible for as much as half of the temperature difference between a glacial age and today. It is therefore important to understand the origin of this variability in the CO₂ partial pressure pCO₂.

The ocean being the largest reservoir of carbon able to exchange CO₂ with the atmosphere on timescales shorter than a few thousands of years, it has been proposed that these pCO₂ variations are the result of changes in the oceanic circulation or in the ocean's biological productivity. This view is further supported by 13C/12C, Cd/Ca and calcium carbonate dissolution records, which indicate considerable reorganization of the ocean's geochemistry and circulation during this time interval [Boyle, 1992; Broecker et al., 1990; Curry and Lohmann, 1983; Duplessy et al., 1988]. Simple box models of the ocean [Ghil et al., 1987; Stommel, 1961; Welander, 1988] as well as more elaborate, but still idealized ones [Quon and Ghil, 1992; Thual and McWilliams, 1992] also show that the thermohaline circulation can easily flip from one steady state circulation to a completely different one. The sensitivity of the carbon cycle to the ocean's circulation changes is therefore of considerable interest.

Numerous box models have been formulated in order to understand the basic mechanisms involved, from models with only three or four boxes [Knox and McElroy, 1984; Lyle and Pisias, 1990; Marino et al., 1992; Sarmiento and Toggweiler, 1984; Siegenthaler and Wenk, 1984] to models including approximately 15 boxes representing the major ocean water masses [Keir, 1988; Michel, 1992]. Three-dimensional ocean general circulation models (GCMs) have also been employed to
understand the problem of this glacial-interglacial difference in atmospheric carbon dioxide [Heinze et al., 1991; Maier-Reimer and Hasselmann, 1987]. However, depending on the model, only a 20-30 ppm difference, at best, can be explained, and therefore no model or mechanism has given a satisfactory explanation for the 80-90 ppm documented difference.

In all these models it was assumed that the transit of organic matter in the water column was the result of falling particulate matter. However Sugimura and Suzuki [1988], using a new measurement method, have reported levels of dissolved organic matter (DOM) at intermediate levels in the water column several times higher than what was commonly expected — "standard" techniques yield surface water concentrations of dissolved organic carbon which fall in the 60-90 μmol L⁻¹ range, while the new technique gives concentrations around 300 μmol L⁻¹ near the surface and still above 100 μmol L⁻¹ up to 500 m deep. The idea that organic matter may not only be falling down to the bottom of the ocean but also be advected horizontally seems to produce modeling results in better agreement with nutrient concentrations measured in the sea [Bacastow and Maier-Reimer, 1991; Toggweiler, 1988]. It is therefore tempting to look at the effect of organic matter on the glacial age pCO₂ problem.

In section 2 of this paper a three-box model is formulated in order to investigate the role of the different oceanic-circulation parameters. Using simple analytical formulae, we show that the vertical distribution of nutrients is a key to understand the relationship between the ocean's deep circulation and atmospheric pCO₂. The existence of a nutrient maximum in intermediate waters, as seen in today's ocean, implies a carbon reservoir just a few hundred meters below the sea surface and is associated with high atmospheric pCO₂. During glacial time, the slow-down of the general deep circulation reduces this nutrient maximum. The carbon's presence at greater depth in the water column lowers the atmospheric pCO₂. This mechanism depends on the regeneration of particulate organic matter in intermediate waters; hence introducing DOM into the model tends to reduce the effect oceanic circulation changes can have on atmospheric pCO₂, as seen in section 3.

The results in these two sections are mostly qualitative, so we introduce a four-box model in section 4 to obtain numerical estimates of the effect described above. A glacial-interglacial pCO₂ difference of 20-30 ppm can be obtained in the DOM-free version of the model; this is fairly close to most other model results in the litterature. The addition of DOM, which acts as a smoother on the vertical nutrient distribution, gives results in the 10-20 ppm range, and makes the glacial-interglacial pCO₂ problem more difficult.

2. THREE-BOX ANALYTICAL MODEL

2.1. Model Description

We describe here a simple model with only one surface box, an intermediate box and a deep box, in order to understand the relative role of the global thermohaline deep circulation (the parameter d), the upwellings (u), the vertical diffusion in the water column (the mixing m) and the biological production (see Figure 1). A separate polar-water box, the importance of which has been demonstrated by Sarmiento and Toggweiler [1984], is introduced in section 4. The schematic diagram of Figure 1 represents the simplest version of a two-stage extension of Broecker's [1982] bio pump, designed to explore the role of intermediate waters as a key link between the thermohaline circulation and the atmospheric CO₂ concentration. Two versions of this three-box model are considered: with and without DOM.

In order to estimate the equilibrium atmospheric pCO₂ we need to know the dissolved inorganic carbon concentration (DIC) and the alkalinity of the surface box. We also need to assume a simple relationship between the new production (P) and the concentration of limiting nutrient in the surface layer. We have therefore nine variables in the DOM-free version of

Fig. 1. Schematic diagram of the three boxes for the analytical model. The oceanic circulation is represented though the three circuits with fluxes d, u, and m, respectively. The new production P is a function of the nutrients in the surface box, and ϵ is the fraction of this production reaching the deep box by sedimentation.
the model: the nutrient content \( N_i, N_2, N_3 \), the alkalinity \( A_i, A_2, A_3 \), and the DIC content \( C_i, C_2, C_3 \) of each box: 1, surface; 2, intermediate; 3, deep. These three global conservation equations:

\[
\begin{align*}
\nu_1 N_1 + \nu_2 N_2 + \nu_3 N_3 &= N, \\
\nu_1 A_1 + \nu_2 A_2 + \nu_3 A_3 &= A, \\
\nu_1 C_1 + \nu_2 C_2 + \nu_3 C_3 &= C,
\end{align*}
\]

where \( \nu_i \) is the ratio of the volume of box \( i \) to the total volume of the ocean and \( N, A, \) and \( C \) are the mean value of the nutrient, alkalinity and DIC concentrations in the sea.

In this model the river input compensate exactly for the sedimentary output, and consequently the mean values \( N, A, \) and \( C \) are assumed to be constant. From the nine equations describing, for each concentration, the balance between sources and sinks in any box, any subset of six equations will be independent. The equilibrium values of all nine variables can then be found by solving, in addition to \((1a)-(1c)\), for example the six equations corresponding to the surface and the deep box:

\[
\begin{align*}
(u + d) (N_1 - N_3) + m (N_2 - N_3) + \epsilon P &= 0, \\
(u + d) (A_1 - A_3) - (2 r_{ca} - r_n) P &= 0, \\
(d (C_1 - C_3) + m (C_2 - C_3) + (r_c + r_{ca} \epsilon') P &= 0,
\end{align*}
\]

where the nutrients and the biological production \( P \) are in phosphate units, \( r_n, r_{ca}, \) and \( r_c \) are the Redfield ratios for nitrate, calcite and organic carbon respectively (i.e., the "stoechiometric coefficients" of organic matter with respect to phosphate), \( \epsilon \) is the fraction of the new production \( P \) reaching the deep box before being oxidized (of the order of 10% - 20%), and \( \epsilon' \) is the fraction of the calcite produced \( (r_{ca} P) \) reaching the deep box before being dissolved (of the order of 90%). The DIC balance takes into account both "soft" and "hard" components of the new production \( (r_{ca} P) \) and \( r_c P \), and the alkalinity balance both the contributions of \( CO_3^2- \) ions \( (2 r_{ca} P) \) and of nitrate \( (r_n P) \).

To incorporate DOM, we need the three additional variables \( D_1, D_2, D_3 \) representing its concentration in each box (in phosphate units), along with the corresponding conservation equations:

\[
\begin{align*}
(u + d) (D_2 - D_1) - \frac{\nu_1}{\tau} D_1 - Q &= 0, \\
(m + d) (D_3 - D_2) + u (D_3 - D_2) - \frac{\nu_2}{\tau} D_2 &= 0, \\
(d (D_1 - D_3) + m (D_2 - D_3) - \frac{\nu_3}{\tau} D_3 &= 0,
\end{align*}
\]

where \( \tau \) is the time constant for DOM regeneration, and \( Q \) is the part of the biological production going into dissolved form. Equations \((2a)-(2f)\) must be changed to

\[
\begin{align*}
(u + d) (N_2 - N_1) - \frac{\nu_1}{\tau} D_1 - Q &= 0, \\
(d (N_1 - N_3) + m (N_2 - N_3) + \epsilon P + \frac{\nu_2}{\tau} D_3 &= 0, \\
(u + d) (A_2 - A_1) - (2 r_{ca} + r_n) P - r_n \frac{\nu_1}{\tau} D_1 - Q &= 0,
\end{align*}
\]

where we assume that Redfield ratios are the same for organic and particulate matter.

2.2. The Nutrient Profile

In order to establish the importance of intermediate waters for the atmospheric pCO2, we discuss first the structure of the nutrient profile. Rearranging \((2a)\) and \((2b)\) gives

\[
N_2 - N_3 = \frac{d}{d+u-\epsilon} \frac{P}{P + \frac{\nu_1}{\tau} D_1 - Q}.
\]

The existence of a nutrient maximum in intermediate waters (i.e. \( N_2 > N_3 \)), as seen in today’s ocean, is the result of \( \epsilon \) being smaller than \( d/(d+u) \). Conversely, in order to get a deep ocean rich in nutrients, as the geological records suggest for glacial time [Boyle and Keigwin, 1982; Duplessy et al., 1988; Kallel et al., 1988], we can either increase the upwellings \( u \), decrease the general circulation \( d \), or increase the proportion \( \epsilon \) of particulate organic matter reaching the deep ocean (see Figure 2). All three possibilities are compatible with current reconstructions of past ocean circulation and paleoproductivity. They may therefore very well act together to change the distribution of nutrients in the sea. Still, to eliminate this nutrient maximum, the values needed for \( \epsilon, d, \) and \( u \) are very different from their present magnitudes. On the scale of the Atlantic Ocean a reasonable estimate for \( d \), corresponding mainly to North Atlantic Deep Water (NADW), would be around 15 or 20 Sv (1 Sv = 10^6 m^3/s), and for \( u \), near 5 or 10 Sv.

To represent the whole world ocean, we should add to \( d \) the Antarctic Bottom Water (AABW), about 20 Sv, and the upwelling in the Indian and in the Pacific Oceans. In either case, \( d \) is larger than \( u \). On the other hand, \( \epsilon \) varies greatly from one place to another (from approximately 1% in the open ocean to more than 20% along continental margins [Berger et al., 1989]) and from one season to another (in mid-latitudes, \( \epsilon \) is much larger during spring blooms, when the phytoplankton population increases dramatically, while the grazer population is still quite low). However, even with enhanced upwelling stimulating the productivity near the equator and along the continents, or with more violent spring blooms resulting from more brutal changes in the seasonal thermocline depth, it seems hard to believe that the value of \( \epsilon \) can become as high as 30% or 40%. Even so, \( d \) would have to be no larger than roughly one third of \( u \) (for example, 5 Sv of NADW formation against 15 Sv of upwelling). As we will see later on, the addition of a polar box softens this conclusion, and less dramatic changes are needed to achieve a plausible geochemical simulation of the glacial-age ocean. Nevertheless, this simple relationship between \( \epsilon, d, \) and \( u \) shows us that only major changes in the ocean circulation or in the productivity can account for the paleodata from deep-sea cores.

Including DOM, \((5)\) is replaced by

\[
d (A_1 - A_3) + m (A_2 - A_1) + (2 r_{ca} \epsilon' - r_n \epsilon) P \\
- r_n \frac{\nu_1}{\tau} D_1 - Q = 0,
\]

\[
(u + d) (C_2 - C_1) - (r_c + r_{ca}) P + r_c \frac{\nu_1}{\tau} D_1 - Q = 0,
\]

\[
d (C_1 - C_3) + m (C_2 - C_3) + (r_c \epsilon + r_{ca} \epsilon') P \\
+ r_c \frac{\nu_1}{\tau} D_3 = 0,
\]

where we assume that Redfield ratios are the same for organic and particulate matter.
Fig. 2. Plot of the term \( \frac{d}{d+u} \) in (5), compared with the fraction \( e \) of organic matter reaching the deep ocean. The nutrient maximum of intermediate waters can disappear when the ratio \( u/d \) is large enough compared with \( e \).

\[
N_2 - N_3 = \left( \frac{d}{d+u} - e \right) \frac{P}{d+m} - (D_2 - D_3), \quad (6a)
\]
\[
D_2 - D_3 = G(d,u,m) Q, \quad (6b)
\]

where \( G(d,u,m) \) is a rational function of \( d, m, u \), whose exact expression is given in Appendix A; its first-order behavior is given by

\[
G(d,u,m) \sim \left( \frac{u}{d+u} \right) \frac{d + m + \frac{\gamma_2}{\lambda}}{d+u} \left( \frac{d+u}{d+m} + \frac{\gamma_3}{\lambda} \right), \quad (7)
\]

where \( \lambda \) is the ratio \( \psi_3/(\psi_2+\psi_3) \) (approximately 0.75 for an intermediate box of depth 1000 m and a total ocean depth of 4000 m). The sign of \( G \) is the sign of \( \left[ \mu - (\psi_2/\psi_3) d \right] \) in (7) as well as in the exact expression. For a 1000-m-deep intermediate box and a 4000-m-deep ocean, \( (\psi_2/\psi_3) = 0.33 \) and \( G \) is positive for reasonable present-day values of \( u \) (5-10 Sv) and \( d \) (15-20 Sv). Consequently, the nutrient maximum \( (N_2 - N_3) \) decreases when the DOM production \( Q \) increases. Hence DOM acts as a smoother on the vertical nutrient distribution. In an ice-age scenario, with \( u \) higher than at present and \( d \) lower, the value of \( G \) will be larger, and the "smoothing effect" of DOM more significant (see Figure 3). This will affect the sensitivity of atmospheric pCO\(_2\) to oceanic-circulation changes during a glacial-interglacial transition.

2.3. Atmospheric CO\(_2\)

This simple three-box model does not include explicitly the atmosphere. However, since it has only one surface box, the carbon dioxide in the atmosphere will equilibrate with the DIC in this box; the resulting equilibrium partial pressure of CO\(_2\) will be the same in the atmosphere and in the surface waters. Consequently, pCO\(_2\) can be calculated from the temperature, salinity, alkalinity and carbon content of the surface box. This can be done exactly, but the purpose of this simple model is to show how the pCO\(_2\) varies when parameters change, rather than give a precise numerical value for given parameter settings.

For a given temperature and a given salinity, Figure 4a shows that the equilibrium pCO\(_2\) of the water increases with the DIC concentration minus the alkalinity. On this diagram, the slope of the isolines is close to 1.2, so that pCO\(_2\) is, to first order, a decreasing function of \( (A_1 - 1.2 C_1) \), or equivalently an increasing function of \( (C_1 - 0.8 A_1) \). This function behaves exponentially. For \( t = 15^\circ \text{C} \) and \( S = 35 \% \), the approximate formula is

\[
pCO_2 = 144 e^{4.2 (C_1 - 0.8 A_1)} \quad (8)
\]

The relative errors in pCO\(_2\) when using (8) versus the exact values are shown in Figure 4b. Oceanic values for carbon and alkalinity lie close to the main diagonal of Figures. 4a and 4b. For these \( C \) and \( A \) values the error in using (8) is less than 10% or even 5%.

With this simplification, and knowing the values of \( A_1 \) and \( C_1 \) from (1) and (2), we get

\[
pCO_2 \quad \text{"increases with"} \quad (C_1 - \mu A_1),
\]

where \( \mu = 0.8 \). This yields, according to (1a)-(1c) and (2c)-(2f), that

\[
pCO_2 \quad \text{"increases with"}
\]

\[
- \left[ r_c - (2\mu - 1) r_{ca} + \mu r_n \right] \frac{P}{d+u} \left[ 1 - \frac{d}{d+m} \right] \quad (9a)
\]

\[
+ \lambda \left[ (2\mu - 1) r_{ca} \epsilon' - (r_c + \mu r_n) \epsilon \right] \frac{P}{d+m}.
\]

From Redfield ratios following Broecker and Peng [1982], i.e., \( r_c = 106, r_{ca} = 21 \) and \( r_n = 15 \), along with \( \epsilon = 0.1 \) and \( \epsilon' = 0.9 \), one has
Fig. 3. Plot of the nutrient content in intermediate waters, normalized to the no-DOM situation, as a function of DOM production, compared to particulate production. When the deep circulation $d$ is small (5 Sv), the nutrient maximum disappears as soon as approximately 30% of the new production is exported as DOM.

Fig. 4. (a) Plot of $pCO_2$ in surface waters as a function of total dissolved inorganic carbon $C$ and alkalinity $A$, calculated from the chemical equilibrium at a given salinity of 35% and a temperature of 15°C. Note that, to first order, $pCO_2$ behaves like the carbon concentration $C$ minus the alkalinity $A$. (b) Relative error in $pCO_2$ values when using the approximate equation (8), instead of the exact dependence of $pCO_2$ on $C$ and $A$ used in the four-box model.

The final result is that

$$pCO_2 \text{ "increases with" } -105 \frac{P}{d+u} + 105 \frac{\rho}{d+u} - \frac{d}{d+m} (9b)$$

The term $P/(d+u)$ represents the effect of the biological pump compared to the upwelling of carbon-rich intermediate waters. The term $P/(d+m)$ represents the effect of the biological detritus reaching the deep ocean compared to the deep mixing and the general circulation. Finally, the term $[P/(d+u)] [P/(d+m)]$ in (9b) represents the effect of the ocean circulation on the nutrient and carbon distribution in intermediate versus deep waters, as seen previously while discussing the nutrient maximum (cf. (5)).
With DOM, (9) is replaced by
\[ p\text{CO}_2 \text{ "increases with"} \]
\[-(r_c - (2 \mu - 1) r_{ca} + \mu r_n) \frac{P}{d + u} \frac{1 - \lambda}{d + m} + \frac{\lambda}{(2 \mu - 1) r_{ca} - (r_c + \mu r_n)} \frac{P}{d + m} - (r_c + \mu r_n) H(d,u,m) Q, \]
where \( H(d,u,m) \) is a rational function of \( d, u, m \).

We get:
\[ p\text{CO}_2 \text{ "increases with"} \]
\[-105 \frac{P}{d + u} + 105 \lambda \frac{P}{d + u} \frac{d}{d + m} \]
\[-0.5 \lambda \frac{P}{d + m} - 118 \frac{Q}{d + u} + 118 \lambda (1 - \lambda) \frac{d}{d + m + \lambda \frac{\Sigma 
u_2}{\tau} d + u} \]
where we used the first order approximation
\[ H(d,u,m) = \lambda \cdot G(d,u,m) + \frac{1}{d + u}. \]

3. ATMOSPHERIC CO2 AND THE DEEP CIRCULATION

3.1. Effect of the General Circulation on \( p\text{CO}_2 \)

The role of the thermohaline turnover \( d \) in changing \( p\text{CO}_2 \) when no DOM is incorporated in the model can be deduced from (9a). Care is required in interpreting the role of the fraction \( P/(d+u) \), since \( P \) is an increasing function of \( d \) (and \( u \)).

In this simple three-box model, it can be shown (cf. Appendix B) that, independently of the parameterization used for \( P \), the ratio \( P/(d+u) \) will still decrease when \( u \) (or \( d \)) increases.

This follows from the intuitive fact that, considering biological productivity as a carbon pump fed by the incoming nutrients (i.e. proportional to \( d+u \)) then \( P/(d+u) \) increases with increasing thermohaline overturning. But the rate of increase will be much smaller. Looking only at the first factors, for \( \lambda = 0.75 \), we get 105 \( \lambda = 79 \), and 118 \( \lambda(1-\lambda) = 22 \).

From Figure 5 it is clear that the introduction of DOM, as anticipated, reduces the sensitivity of \( p\text{CO}_2 \) to changes in the deep circulation.

3.2. Polar Regions

While the three-box model is awkward in trying to describe specifically what happens in high latitudes, we can still use (9) and (10) in order to get some insight into the basic mechanisms. In this model, by taking a very large mixing \( m \), the intermediate and the deep box will behave like one box, the surface layer has to be chosen thicker, \( u \) interpreted for the present purpose as the deep convection, and \( d \) defined with an opposite sign. Equation (10b), with \( m \) taken very large, tells us that DOM is irrelevant to the discussion here; the only remaining terms are \( Q/(d+u) \) which is proportional to \( P/(d+u) \), and \( P/(d+u) \) itself. Accordingly, \( p\text{CO}_2 \) will behave like \(-P/(d+u) \); it will decrease when the deep convection \( u \) decreases, or when \( d \) increases. The effect of \( d \) in high latitudes is, not suprisingly (its sign being reversed), the opposite of its effect in low latitudes. Without intermediate waters, however, and the term \( 1 - \lambda \) \( (d/(d+m)) \) of (9a), changes of \( d \) in the polar regions will not dominate atmospheric \( p\text{CO}_2 \) changes.

3.3. Summary of Analytical Results

During a glacial-interglacial transition it is reasonable to assume that \( u \) decreases as a result of weakened mid-latitude winds and \( d \) increases as a result of enhanced deep-water formation. The resulting effect on atmospheric \( p\text{CO}_2 \) will be due mainly to the term \( 1 - \lambda \) \( (d/(d+m)) \) in (9a), which represents the stratification of the nutrient distribution between intermediate and deep waters. A weakening of this stratification, due for example to the advection of DOM, will result in a carbon cycle less sensitive to deep-ocean circulation changes. This corresponds to the intuitive idea that the presence of carbon-rich waters in the nutrient maximum, resulting from the rapid regeneration of falling particulate matter, can affect \( p\text{CO}_2 \) through the action of \( d \) or \( u \). If part of this organic matter is advected as a dissolved compound instead of being regenerated in the first thousand meters, less DIC will be available near the surface and the atmospheric \( p\text{CO}_2 \) will be less sensitive to the circulation pattern.
Fig. 5. Plot of the "glacial-interglacial" pCO$_2$ difference as a function of DOM production, where the glacial stage is defined here by $d = 5$ Sv. The efficiency of ocean circulation changes to modify pCO$_2$ is lowered by the presence of DOM.

Fig. 6. Schematic diagram of the four boxes for the numerical model. Seven loops represent the oceanic circulation; their fluxes are given by $d$, $t$, $u$, $m$, $c$, $e$, and $f$. The new production ($P_L$ in low latitudes, $P_H$ in high latitudes) is parameterized in the same way as before. In high latitudes, all the production $P_H$ goes to the deep box, as well as a fraction $e$ of $P_L$.

This simple three-box model is able, somewhat surprisingly, to reproduce the general behavior of the oceanic carbon cycle, but it cannot give acceptable numerical estimates of oceanic concentrations and consequently of atmospheric pCO$_2$. In order to obtain an estimate of the atmospheric CO$_2$ variation due to changes in the oceanic circulation, a more realistic model, (with an explicit calculation of pCO$_2$ and with a polar box), is developed in the next section.

4. A SIMPLE FOUR-BOX MODEL

4.1. Model Description

A more satisfactory representation of the oceanic carbon cycle is obtained by using four boxes: a high-latitude cold surface box and a low-latitude warm surface box represent respectively a sink and a source of atmospheric CO$_2$; an intermediate box, necessary to model the stratification of nutrients in the ocean, the importance of which was demonstrated in the previous chapter; and the deep box as the main reservoir. A schematic diagram of the boxes and of the fluxes between them is shown on Figure 6. There are seven independent loops describing the circulation and the mixing between the boxes: $d$ and $t$ represent the general thermohaline circulation, $m$ the deep mixing, $u$ the wind-driven upwelling, $c$ the high-latitude deep convection, $e$ the mixing between warm surface waters and cold surface waters, and $f$ the mixing along isopycnals.

The advection scheme is the same as in the analytical model described in section 2: the water mass flux $f_{ij}$ between a box $i$ with a concentration $C_i$ and a box $j$ with a concentration $C_j$...
transports a dissolved-matter flux $f_j C_i$. The parameterization of the biological new production is given by the Michaelis-Menton formula, which we write here as a function of the nutrients in box number $i$ ($i = 1$ or 2):

$$P_i = P_{\text{max}} \frac{N_i}{k_i + N_i} .$$

(12)

The values of $P_{\text{max}} = 3$ molC m$^{-2}$ yr$^{-1}$ and of $k_i = 0.5$ mmolPO$_4$ L$^{-1}$ are taken the same in the two surface boxes: the high nutrient content in the polar box ($i = 1$) is therefore the result of the deep convection $c$ and of the mixing $f$ with intermediate waters "along isopycnals." As in section 2, the Redfield ratios are $r_c = 106$, $r_{ca} = 21$, and $r_n = 15$ and the remineralization and dissolution coefficients are $e = 0.1$ and $e' = 0.9$ (see Table 1).

### 4.2. Glacial-Interglacial Transition

First, the model is investigated in its simplest possible form, as described above. The effect that the seven circulation parameters $u$, $c$, $d$, $t$, $f$, $e$ and $m$ have on atmospheric pCO$_2$ and on the nutrient in intermediate waters is depicted in Figure 7. In the real ocean, high-latitude surface waters are fed essentially by thermocline water ($t$) and not by surface water ($d$). The high-latitude deepwater formation will therefore be carried mainly by the parameter $t$. In this diagram, variations of each parameter are shown and compared to a standard (glacial) state, defined by $e = 5$ Sv; $f = 10$ Sv; $c = 20$ Sv; $d = 5$ Sv; $m = 10$ Sv; $t = 20$ Sv; $u = 25$ Sv. The resulting equilibria are shown in the plane of atmospheric pCO$_2$ and nutrient content of the intermediate box. The concentrations in each box are given in Table 2 for the standard values of the circulation parameters.

For an interglacial-age scenario, stronger deepwater formation in high latitudes ($t$) than during a glacial age is a very reasonable assumption [Broecker et al., 1990; Shackleton et al., 1988]. This corresponds, during glacial stages, to the slow-down of Broecker's conveyor belt. Therefore a glacial-interglacial transition can be represented by an increase in $t$. We would also like to have higher upwelling rates $u$ during glacial stages than today: in many coastal areas and in equatorial regions the productivity was probably higher than [Arrhenius, 1952; Mix, 1989; Sarthein et al., 1987, 1998]; this is consistent with stronger winds associated with stronger temperature gradients.

The changes in the high-latitude convection $c$ and in the "mixing" $f$ are far from being so easily understood. Paleoproduction in high latitudes is poorly known, since the possible paleoproduction indicators do not always agree. But not indicator suggests dramatic changes during glacial-interglacial transitions, for example Bareille's [1991] silica fluxes, while some suggest even a possible decrease [Labeyrie and Duplessy, 1985; Mortlock et al., 1991]. This could suggest lower values for $c$ and $f$ during a glacial stage, but $^{14}$C isotopes do not agree with C/D/Ca results [Boyle, 1992] with respect to deep-water nutrient content, implying that we do not fully understand the geochemistry of the southern ocean in this time period.

Changes in $e$ do not affect the results very much, and it is difficult to find physical phenomena able to produce considerable changes in $m$, which represents the diffusive mixing in the deep ocean. We then define a standard glacial-interglacial transition as an increase in $t$, a decrease in $u$ and a small increase in $c$. Other possible transitions are also shown in Figure 7. As discussed in sections 2 and 3, all of them associate an increase in pCO$_2$ with an increase in the nutrient content of intermediate waters. This is mainly due to the circulation parameter $t$ (cf. the discussion of $d$ and pCO$_2$ in section 3.1), and illustrate well the relationship between atmospheric pCO$_2$ and the nutrient stratification. Oxygen and carbon isotopes have also been calculated for each set of parameters, and compare reasonably well with oceanic values.

The standard transition described here explains a $26.8$ ppm atmospheric pCO$_2$ difference, which is far from the documented $80$ ppm difference. A more realistic glacial-interglacial transition would probably involve variations in several other parameters. Since the system is only weakly nonlinear, a model with many more boxes would not give too different a result. It seems therefore that, within the context of this type of box model, changes in the oceanic circulation alone cannot

### TABLE 1. Symbols and Their Numerical Values

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Volumes of the Four Boxes</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$v_1$</td>
<td>17</td>
<td>Mkm$^2$ (10$^6$ km$^2$)</td>
</tr>
<tr>
<td>$v_2$</td>
<td>26</td>
<td>Mkm$^2$</td>
</tr>
<tr>
<td>$v_3$</td>
<td>234</td>
<td>Mkm$^2$</td>
</tr>
<tr>
<td>$v_4$</td>
<td>1107</td>
<td>Mkm$^2$</td>
</tr>
<tr>
<td><strong>Areas of the Surface Boxes</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\Sigma_1$</td>
<td>80.3</td>
<td>Mkm$^2$ (10$^6$ km$^2$)</td>
</tr>
<tr>
<td>$\Sigma_2$</td>
<td>265.9</td>
<td>Mkm$^2$</td>
</tr>
<tr>
<td><strong>Salinities and Temperatures</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$s_1$</td>
<td>33.75</td>
<td>$%_o$</td>
</tr>
<tr>
<td>$s_2$</td>
<td>35.45</td>
<td>$%_o$</td>
</tr>
<tr>
<td>$t_1$</td>
<td>4.5</td>
<td>$\degree_C$</td>
</tr>
<tr>
<td>$t_2$</td>
<td>23.8</td>
<td>$\degree_C$</td>
</tr>
<tr>
<td><strong>Exchange Velocity for CO$_2$</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$E_1$</td>
<td>0.11</td>
<td>molC m$^2$ yr$^{-1}$</td>
</tr>
<tr>
<td>$E_2$</td>
<td>0.07</td>
<td>molC m$^2$ yr$^{-1}$</td>
</tr>
<tr>
<td><strong>Mean Concentrations in the Sea</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$A$</td>
<td>2.4</td>
<td>meqAlk L$^{-1}$</td>
</tr>
<tr>
<td>$N$</td>
<td>2.145</td>
<td>mmolPO$_4$ L$^{-1}$</td>
</tr>
<tr>
<td>$C$</td>
<td>2.3</td>
<td>mmolDIC L$^{-1}$</td>
</tr>
<tr>
<td><strong>Redfield Ratios and Remineralization Coefficients</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$r_c$</td>
<td>106</td>
<td></td>
</tr>
<tr>
<td>$r_{ca}$</td>
<td>21</td>
<td></td>
</tr>
<tr>
<td>$r_n$</td>
<td>15</td>
<td></td>
</tr>
<tr>
<td>$e$</td>
<td>0.1</td>
<td></td>
</tr>
<tr>
<td>$e'$</td>
<td>0.9</td>
<td></td>
</tr>
<tr>
<td><strong>Michaelis-Menton Formula Parameter Values</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$P_{\text{max}}$</td>
<td>3</td>
<td>molC m$^2$ yr$^{-1}$</td>
</tr>
<tr>
<td>$k_b$</td>
<td>0.5</td>
<td>mmolPO$_4$ L$^{-1}$</td>
</tr>
<tr>
<td><strong>Oceanic Circulation</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$d$</td>
<td>5 - 10</td>
<td>Sverdrups (10$^6$ m$^3$ s$^{-1}$)</td>
</tr>
<tr>
<td>$t$</td>
<td>20 - 40</td>
<td>Sv</td>
</tr>
<tr>
<td>$m$</td>
<td>30 - 5</td>
<td>Sv</td>
</tr>
<tr>
<td>$u$</td>
<td>20 - 10</td>
<td>Sv</td>
</tr>
<tr>
<td>$c$</td>
<td>20 - 30</td>
<td>Sv</td>
</tr>
<tr>
<td>$e^*$</td>
<td>5 - 10</td>
<td>Sv</td>
</tr>
<tr>
<td>$f$</td>
<td>10 - 15</td>
<td>Sv</td>
</tr>
</tbody>
</table>
explain the low carbon dioxide content of the atmosphere during the last glacial maximum. Other changes should be taken into account, mainly variations in the mean concentrations of nutrients, carbon and alkalinity. These are certainly not constant in the real world, because of sea level changes, as well as variations in river input and in sediment preservation.

4.3. Comparison With Other Box Models

We have chosen here to study the effect of oceanic-circulation changes alone, everything else being held constant. This accounts for some of the differences with some other simple box models.

In the work by Knox and McElroy [1984], the new production in high latitude is increased arbitrarily by a factor of 100. In the work by Sarmiento and Toggweiler [1984] and in the work by Siegenthaler and Wenk [1984], the high-latitude production $P_h$ is an independent parameter, and large pCO$_2$ variations are obtained when increasing $P_h$ (in the former) or keeping it constant (in the latter), while decreasing the high-latitude convection $c$. Here $P_h$ is calculated as a function of the nutrient content, itself a function of the circulation, and it increases with increasing $c$. In addition, the circulation parameters are changed at most by a factor of two. The large pCO$_2$ variations found in these earlier papers are therefore ruled out here.

The model of Marino et al. [1992] (published while this paper was under review) is an extension of Knox and McElroy model. Though the circulation parameters are allowed to differ from present-day values much more than here (deepwater formation is totally suppressed and intermediate-water formation is multiplied by six for the glacial stage) the resulting pCO$_2$ variations (31 ppm) are similar to the results of our model.

**TABLE 2. Concentrations for the Standard Glacial and Interglacial Circulation in the Absence of DOM**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Glacial</th>
<th>Interglacial</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>pCO$_2$ atmosphere</td>
<td>254.6 ppm</td>
<td>281.4 ppm</td>
<td>ppm</td>
</tr>
<tr>
<td>Phosphate</td>
<td>1  1.033 μmol L$^{-1}$</td>
<td>2  0.149 μmol L$^{-1}$</td>
<td></td>
</tr>
<tr>
<td>Production</td>
<td>1  2.021 mol C m$^{-2}$ yr$^{-1}$</td>
<td>2  0.689 mol C m$^{-2}$ yr$^{-1}$</td>
<td></td>
</tr>
<tr>
<td>Alkalinity</td>
<td>1  2.356 meq L$^{-1}$</td>
<td>2  2.323 meq L$^{-1}$</td>
<td></td>
</tr>
<tr>
<td>Oxygen</td>
<td>1  319.5 μmol L$^{-1}$</td>
<td>2  215.5 μmol L$^{-1}$</td>
<td></td>
</tr>
<tr>
<td>Carbon</td>
<td>1  2.133 mmol L$^{-1}$</td>
<td>2  1.965 mmol L$^{-1}$</td>
<td></td>
</tr>
<tr>
<td>pCO$_2$</td>
<td>1  248.3 ppm</td>
<td>2  257.6 ppm</td>
<td>ppm</td>
</tr>
<tr>
<td>Carbon 13</td>
<td>1  1.44 per mil</td>
<td>2  1.76 per mil</td>
<td></td>
</tr>
<tr>
<td>$^{13}$C atmosphere &amp; Δ$^{13}$C</td>
<td>1  -5.97 per mil</td>
<td>2  -37.0 per mil</td>
<td>per mil</td>
</tr>
</tbody>
</table>

Fig. 7. Summary of four-box model results without DOM. Each line represents a change in the ocean circulation. A "standard" glacial-interglacial transition is defined by simultaneous changes in $t$, $u$, and $c$, and the interglacial atmospheric pCO$_2$ is 26.5 ppm higher.
Much larger variations of pCO₂ are achieved, however, by these authors when changing the global budget of nutrients, carbon and alkalinity, and changing the coefficients of the biological parameterization.

4.4. Effect of Dissolved Organic Matter

As mentioned earlier, a new method for measuring dissolved organic matter (DOM) has given several times the amounts found by traditional methods [Sugimura and Suzuki, 1988]. Yet no one else has reported similar results and these measurements remain controversial. According to these authors, this DOM pool shows marked variations with depth, suggesting that it is actively involved in the transport of carbon, while the DOM fraction already known from other measurement methods is very long-lived (several thousands of years) and consequently more or less uniformly distributed in the ocean [Williams and Druffel, 1987]. In order to exhibit a strong gradient between 300 and 500 m as found by Sugimura and Suzuki, this new DOM must be more labile and be regenerated with a characteristic time of the order of a century.

The addition of the new, larger and more rapidly recycled DOM pool into the model is straightforward: the production of DOM is considered to represent 50% of the new production, which is then advected by the water and slowly transformed back into inorganic matter through biological consumption, with an exponential decay time constant of 200 years. Simulations with the same parameter values as in the DOM-free case give the results summarized in Figure 8. The concentrations in each box are given, for the standard states, in Table 3. We notice that the nutrient content is much lower than in the no-DOM situation: the effect of DOM is to smooth the action of the biology. This was also noticed in GCM simulations of the present ocean [Bacastow and Maier-Reimer, 1991], which show a smaller nutrient maximum in the equatorial intermediate waters, in better agreement with in situ measurements. The addition of DOM in these GCM simulations also leads to disappearance of an unwanted anoxic region in the intermediate equatorial waters. In our model, the low-latitude region is much too wide to represent the equatorial zone only and no anoxia is observed, but we note that the oxygen level increases in intermediate waters when DOM is incorporated, which is a corollary of the reduction in nutrient content.

In the presence of DOM the geochemical stratification of the ocean is weakened, and as a result the pCO₂ variations between the different circulation regimes are smaller: the standard transition allows a 15.3 ppm pCO₂ difference and, from the figure, it is clear that alternate scenarios would give similar results. This smoothing effect depends only weakly on the exact parameterization of DOM: if a bigger fraction of the new production is transformed into DOM, or if we take a longer time constant for the decay into inorganic matter, the smoothing action of DOM increases, while opposite assumptions lead to a DOM decrease. This means that the smoothing effect depends mainly on the size of the DOM pool in the ocean. If this pool is not negligible, then it makes the size of pCO₂ glacial-interglacial differences more difficult to explain.

Our model results appear to be quite robust: the diagrams are very similar in both cases, with changes in the oceanic circulation having more or less the same effects. Many different possible parameterizations of the biological component of the system have been tried, nonlinear formulas for the new production, for the regeneration of the falling particulate matter, etc., without changing the general behavior described in the "standard case." Models with more boxes have also been explored, but the main results given here do not seem to depend on the geographic details that would be provided by a more realistic model.

![Graph](image)

Fig. 8. Same as Figure 7, but when DOM is incorporated in the model. Nutrient levels are much lower in the intermediate box than previously and the resulting atmospheric pCO₂ difference smaller.
The introduction of dissolved organic matter (DOM) into the system only makes the explanation even more difficult. If the DOM pool is negligible, we can at best explain a 30 ppm difference, with a more probable value between 10 and 20 ppm, and the presence of DOM reduces this effect by as much as 50%.

The carbonate response of the sediment could perhaps provide us with another 10 or 20 ppm [Boyle, 1988], but this would still be insufficient. This is entirely consistent with ocean GCM experiments [Heinze et al., 1991] which take into account some simple interactions with the sediment and give a pCO2 difference of 28 ppm between the present and a glacial-age ocean circulation. Numerous hypotheses have been formulated in order to account for such a discrepancy between the models and the paleoclimatic data; most of them involve increased biological production at high latitudes, the last one being the fertilization of polar oceans by iron [Martin and Fitzwater, 1988; Martin et al., 1990]. Reconstructions of paleoproductivity, however, are in disagreement with a large-scale increase in productivity in polar regions [Labeyrie and Duplessy, 1985; Mortlock et al., 1991], and no indication of anoxia, as would be implied by the high productivity levels required by such mechanisms, is to be found in those regions during the last glaciation.

In addition, recent measurements of 13C in the carbon dioxide trapped in air bubbles from Antarctic ice [Leuenberger et al., 1992], as well as plant analysis [Marino et al., 1992], indicate that atmospheric 13C was probably lower during the last glacial maximum. Keir [1992] argues convincingly this makes it more difficult to account for the lower atmospheric pCO2 at that time, because increasing ocean productivity and, to a smaller extent, reducing the nutrient maximum through changes in ocean circulation as done in this paper, also increases surface water and atmospheric 13C. Furthermore, Keir [1991] used a three-box model and a 13-box model to show that an increase in the biological carbon pump efficiency requires a considerable reduction of the amount of land biota to account for lower 13C in the deep ocean. Hence the decrease in intermediate-depth nutrients during glacial time may be better explained by changes in the ocean circulation or in the biological recycling.

It is conceivable that some crucial elements of the oceanic carbon budget have been overlooked, like CaCO3 sedimentation or river inputs; it may also be that our representation of the biological processes coupled with the mixed-layer dynamics is oversimplified [Simonot et al., 1988]. In any case the changes in ocean circulation studied in this paper are not sufficient to explain the lower pCO2 level of the last glacial maximum, and dissolved organic matter makes matters only more difficult. If the carbonate response of the sediment could perhaps provide us with another 10 or 20 ppm [Boyle, 1988], but this would still be insufficient. This is entirely consistent with ocean GCM experiments [Heinze et al., 1991] which take into account some simple interactions with the sediment and give a pCO2 difference of 28 ppm between the present and a glacial-age ocean circulation. Numerous hypotheses have been formulated in order to account for such a discrepancy between the models and the paleoclimatic data; most of them involve increased biological production at high latitudes, the last one being the fertilization of polar oceans by iron [Martin and Fitzwater, 1988; Martin et al., 1990]. Reconstructions of paleoproductivity, however, are in disagreement with a large-scale increase in productivity in polar regions [Labeyrie and Duplessy, 1985; Mortlock et al., 1991], and no indication of anoxia, as would be implied by the high productivity levels required by such mechanisms, is to be found in those regions during the last glaciation.

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neglected in a first approximation of the oceanic carbon cycle model. Although the glacial-interglacial pCO2 problem remains unsolved, the additional difficulties involved by dissolved organic matter probably do not exacerbate the situation.

APPENDIX A: SOLUTION OF THE THREE-BOX MODEL WITH DOM.

The best way to solve (1), (3), and (4) is use the variables \( (N_i - D_i), (A_i - r_n D_i) \) and \( (C_i + r_c D_i) \) that are solutions of (2). We then obtain directly:

\[
N_2 - N_3 = (\text{terms without DOM}) - (D_2 - D_3)
\]

\[
pCO_2 = \frac{[(C_1 - A_1) - (C_i - A_i)](v_2 + v_3)}{(v_2 + v_3)}
\]

with \( D = v_1 D_1 + v_2 D_2 + v_3 D_3 \) being the mean DOM concentration.

To solve (3), we can use the intermediate variables

\[
D_{13} = D_1 - D_2, \quad D_{23} = D_2 - D_3.
\]

We then obtain the system of two equations with two unknowns:

\[
\begin{align*}
\left( \frac{v_3}{T} m - \frac{v_2}{T} d \right) D_{13} & = \left( \frac{v_2}{T} (m + v_3) + \frac{v_3}{T} (m + d + u) \right) D_{23} \\
\left[ \frac{v_3}{T} (u + d + v_1) + \frac{v_1}{T} (d + m) \right] D_{23} & = \left[ \frac{v_3}{T} (d + u) - \frac{v_1}{T} m \right] D_{23} + \frac{v_3}{T} Q
\end{align*}
\]

Its solution is

\[
D_{23} = \frac{v_3}{T} \left( \frac{v_3}{T} u - \frac{v_2}{T} d \right) Q
\]

\[
D_{13} = v_3 \left( \frac{v_3}{T} (m + v_3) + \frac{v_3}{T} (d + m + u) \right) Q
\]

with

\[
\Delta = \left( \frac{v_3}{T} \right)^2 \left[ d^2 + m d + m u + d u + v_1^2 (m + d + u) \right]
\]

\[
+ \frac{v_3 v_2}{T} \left[ d^2 + m d + m u + d u + \frac{v_1}{T} v_2 + \frac{v_2}{T} (d + u) + \frac{v_1}{T} m \right]
\]

\[
+ \frac{v_3 v_1}{T} \left[ v_1 d^2 + m d + m u + d u \right] + \frac{v_1}{T} \frac{v_3}{T} \frac{v_1}{T} m .
\]

To simplify this expression, we notice that \( v_1 \) is negligible compared to \( v_3 \). We can therefore use the approximate expression:

\[
\Delta = \left( \frac{v_3}{T} \right)^2 \left[ d^2 + m d + m u + d u + \frac{v_2}{T} \right]
\]

The functions \( G(d,u,m) \) and \( H(d,u,m) \) introduced in Section 2 are given as functions of \( D_{13} \) and \( D_{23} \) by

\[
G = D_{23} / Q,
\]

\[
H = [(D_{13} - D_{23}) + \lambda D_{23}] / Q = \frac{1}{d + u} + \lambda G .
\]

This yields (7) and (11) used in sections 2 and 3.

APPENDIX B: EFFECT OF PRODUCTIVITY AND UPWELLING IN THE THREE-BOX MODEL.

Using Broecker and Peng's [1982] Redfield ratios, and assuming that \( m \) is of the same order of magnitude as \( d \) (14C calibration of numerous box models suggests that \( m \) might actually be 2 or 3 times larger than \( d \), it follows from (9) that \( d(d+m) \) is of the order of (or smaller than) one half. This means that the dominant term of the right-hand side of (9b) is clearly the first one, and to zeroth order, pCO2 will behave like \( P(d+u) \). An increase in productivity without any change in the ocean circulation does decrease the pCO2. An increase in the upwelling rate \( u \) must be looked at more carefully, because \( P \) itself depends on the circulation. However, in the realm of this simple three-box model, it can be shown that, independently of the parameterization used for \( P \), the ratio \( P(d+u) \) will still decrease when \( u \) (or \( d \)) increases. This is actually the result of the quite intuitive idea that biology, considered as a carbon pump fed by the incoming nutrients carried by (\( u + d \)), cannot be more than 100% efficient.

This implication can be easily verified using either a crude parameterization for \( P \), like \( P = \mu (d+u) N \), i.e., the production is proportional to the incoming nutrients (for \( \mu = 1 \), there are no nutrients left in the surface box), or a more realistic one like the Michaelis-Menton formula:

\[
P = \frac{P_{max} N_i}{k_s + N_i}
\]

[Dugdale, 1967] and then solving the resulting equations (1a), (2a), and (2b), of the first or the second degree accordingly. More generally, if we assume that \( P \) is a function of \( N_i \) only, increasing with \( N_i \), then, from the solution of (1a), (2a), and (2b), we have \( \partial N_i / \partial u \) and \( \partial N_i / \partial d \) as functions of \( \partial P / \partial u = (dP/dN_i) \partial N_i / \partial u \). We can therefore obtain \( \partial (N_i - N_i) / \partial u \) as a function of \( dP/dN_i \) only, and from (5a) deduce that \( P(d+u) \) will decrease when \( u \) increases:

\[
\frac{dP}{d+u} = \frac{1 + v_3 (u + d + m) + v_2}{d + u + m} \frac{dP}{dN_i}
\]

On the other hand, if the deep mixing \( m \) is small enough compared to \( d \), then the factor \( 1 - \lambda (d/d+m) \) can be as low as \( 1 - \lambda = 0.25 \) and the two terms on the right-hand side of (5a) are of the same order of magnitude. Even in this case, some dramatic assumptions have to be made on \( \varepsilon \) and \( \varepsilon' \) in order to reduce pCO2 by increasing \( u \). Boyle [1986] succeeds in doing so in a four-box model, by taking \( \lambda = 0.87 \), \( m = 0 \), and \( \varepsilon = \varepsilon' = 0.2 \). However, in a box model having diffusive mixing between the deep and intermediate box, it seems hopeless to obtain a reduction in atmospheric CO2 by an increase in upwelling, as is supposed to have occurred during glacial maxima due to an intensification of the pole-to-equator temperature difference and hence of the zonal winds.

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