Steady-State and Transient Modeling of Tracer and Nutrient Distributions in the Global Ocean

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ABSTRACT

We have completed the studies using the inorganic carbon cycle in the zonally averaged ocean circulation model to calculate anthropogenic uptake of $C0_2$ and bomb radiocarbon. While our results are in broad agreement with previous studies, we have learned that horizontal mixing processes due to gyre circulation are important for transient tracer experiments over a few decades. These findings are in press (Stocker et al., 1994). Using the inorganic carbon cycle model we have started to look at the distributions of $\delta^{13}C$ in the ocean. The model is able to reproduce faithfully the air-sea fractionation of $\delta^{13}C$.

The effect of changing sea surface temperature in the middle and low latitudes of the world ocean on $\rho C0_2^{atm}$ is studied in this model and compared to the organic carbon cycle model of the Hamburg group. We find significant differences in sensitivity and are in the process of investigating possible reasons.

Incorporation of the organic component is still ongoing. In the present version the surface concentrations of phosphate are still too high indicating strong upwelling.

RESEARCH RESULTS

In this final report we describe the research results that have been obtained under U.S. Department of Energy Agreement No. DE-FG02-91ER61202 from June 1, 1993 to July 31, 1997. Stocker has taken up a faculty position with the Physics Institute, University of Berne (Switzerland) as of October 1, 1993. Part of the work described here was done by Jean Lynch-Stieglitz as part of her Ph.D. thesis at LDEO and her post doctoral period at WHOI. She is now a professor at Lamont-Doherty Earth Observatory.

INORGANIC CARBON CYCLE

The apparent gap between the 1-dimensional box models and the few 3-dimensional models of the global carbon cycle has been closed by developing an inorganic component to the zonally averaged ocean circulation model of Wright and Stocker (1992). We have performed $2xC0_2$-experiments and simulated the uptake of anthropogenic CO_2 over the last 200 years. Results are in broad agreement with the previous models. The obvious advantages of the 2-d model are its efficiency and the ability to yield detailed information about the geographical distribution of tracer uptake.

We demonstrated that the model is able to simulate tracer uptake on timescales of decades using bomb radiocarbon. In order to obtain realistic surface distributions is was necessary to introduce additional meridional mixing in the top 500 meters. This mixing was related to the wind-driven gyre circulation, and we have proposed a parameterization of this effect. The results are in press (Stocker et al., 1994).

AIR-SEA COMPONENT OF $\delta^{13}C$

The stable isotope ^{13}C was added as a tracer to the inorganic carbon cycle model. Surface fluxes of ^{13}C follow those of ^{12}C but with a temperature-dependent equilibrium fractionation factor (Mook et al., 1974) as well as a small kinetic fractionation (Siegenthaler and Münnich, 1981). Although the distribution of $\delta^{13}C$ in today's ocean is primarily driven by biological cycling (Kroopnick, 1985), air-sea exchange does effect the $\delta^{13}C$ signature to a lesser degree (Broecker and Maier-Reimer, 1992). The air-sea exchange signature $\delta^{13}C_{as}$ can be calculated by using the nutrient content to subtract the portion of the oceanic $\delta^{13}C$ that can be attributed to nutrient cycling.

The $\delta^{13}C$ distribution in the inorganic carbon cycle model agrees well with the observed b 13C values in both the surface and the deep ocean (Figures 1 and 2). Although both stable isotopes of carbon, ^{12}C and ^{13}C reach steady state in the model within about 2000 years, the distribution of the ratio $\delta^{13}C$ requires about 10,000 years of integration. The relatively fast 2-D ocean model proves ideal for making the long integrations necessary to study the distribution and evolution of $\delta^{13}C$.

In addition, ^{13}C has been incorporated also into the organic carbon cycle models. For this, it is necessary to include not only the fractionation due to air-sea exchange, but also the fractionation due to biological processes. Primary production in the surface ocean discriminates against ^{13}C making organic material approximately 20‰ lighter than surface ocean water. There is no fractionation of carbon isotopes upon production and dissolution of calcium carbon shell material.

The organic carbon cycle model is still being improved by exploring the range of parameters and evaluating different formulations of vertical particle fluxes and new production.

CARBON CYCLE MODEL COMPARISON

The solubility of CO_2 in sea water is strongly temperature-dependent. In order to reassess the sensitivity of typical glacial-to-interglacial SST changes in various models we have started a comparison with the Hamburg 3-dimensional carbon cycle model (Heinze et al., 1991). Instead of changing SST and hence the circulation, solubility was decreased in the latitude band of 40°S to 40°N. Since ρCO_2^{atm} is predicted in both models, the outgassing due to decreased solubility results in an increase of ρCO_2^{atm} which is partially compensated by an increased uptake of $C0_2$ in the high latitudes.

Sensitivity experiments for the different models indicate that the 2-D model is less sensitive than the 1-D or the 3-D models (Figure 3). At present, only a tentative comparison is possible because the 3-D carbon cycle model is organic and seasonal whereas the other two are inorganic and annual mean. This question will be addressed in the continuation of this project.

ORGANIC CARBON CYCLE

The organic carbon cycle model consists of 6 tracers: the 3 isotopes of carbon ($\sum^{12}C$, ^{13}C, and ^{14}C), alkalinity, phosphate and oxygen. In the ocean interior these tracers are mixed by advection, diffusion and convection, and the carbon isotopes are exchanged with the atmosphere through gas exchange at the sea surface. The temperature-dependent fractionations from ^{13}C and ^{14}C are now included. The formulation of the biological pump follows that of Maier-Reimer (1993).

We have now relaxed our previous boundary condition of restoring to observed surface values in that we relate new production to the surface phosphate concentration. Further, $\rho C0_2^{atm}$ is also predicted assuming that the global carbon inventory is constant. All carbon isotope air-sea exchange fluxes are now calculated as functions of the $\rho C0_2$-contrast between the ocean and the atmosphere.

PUBLICATIONS

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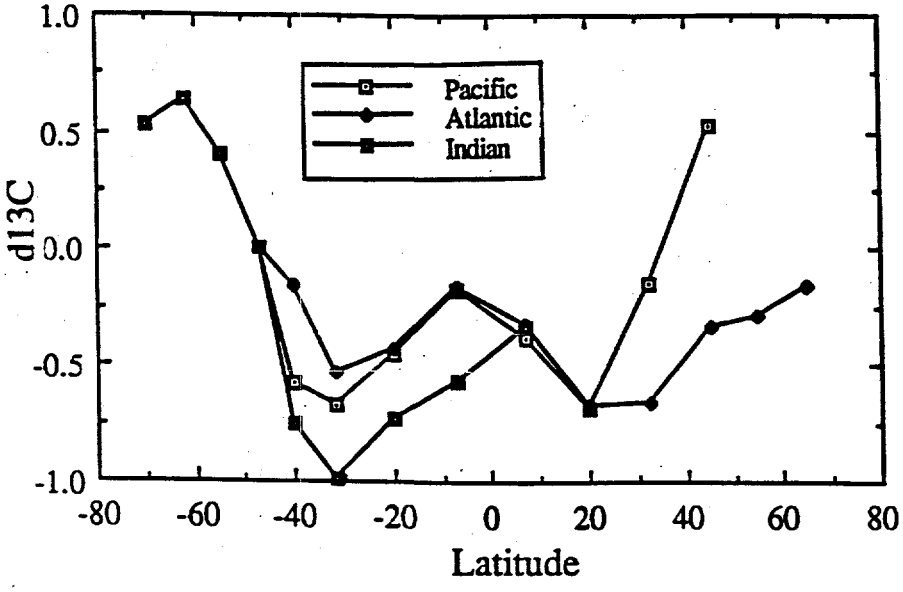
FIGURES

Figure 1: Comparison of the surface ^{13}C due to air-sea exchange in the 2-d inorganic carbon cycle model (top) and the observations (bottom) for the three ocean basins. Note that the model does not include the anthropogenic decrease of atmospheric ^{13}C.

Figure 2: Distribution of the ^{13}C due to air-sea exchange in Atlantic (top) and Pacific (bottom).

Figure 3: Sensitivity of various carbon cycle models to changing the solubility of $C0_2$ in surface water. Dots represent the 1-d, squares the 2-d and triangles the 3-d model results. The equilibration factor defined as the relative change of ρCO_2^{atm} increases with a decrease in solubility.


Pacific
Atlantic
Indian
d13C
1.0
0.5
0.0
-0.5
-1.0
-80
-60
-40
-20
0
20
40
60
80
Latitude

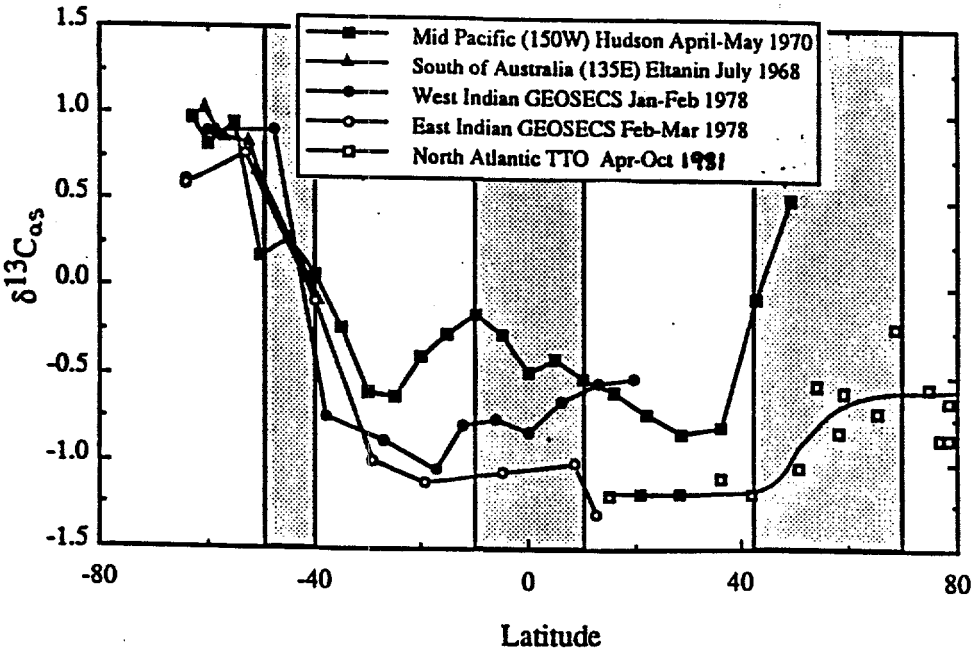
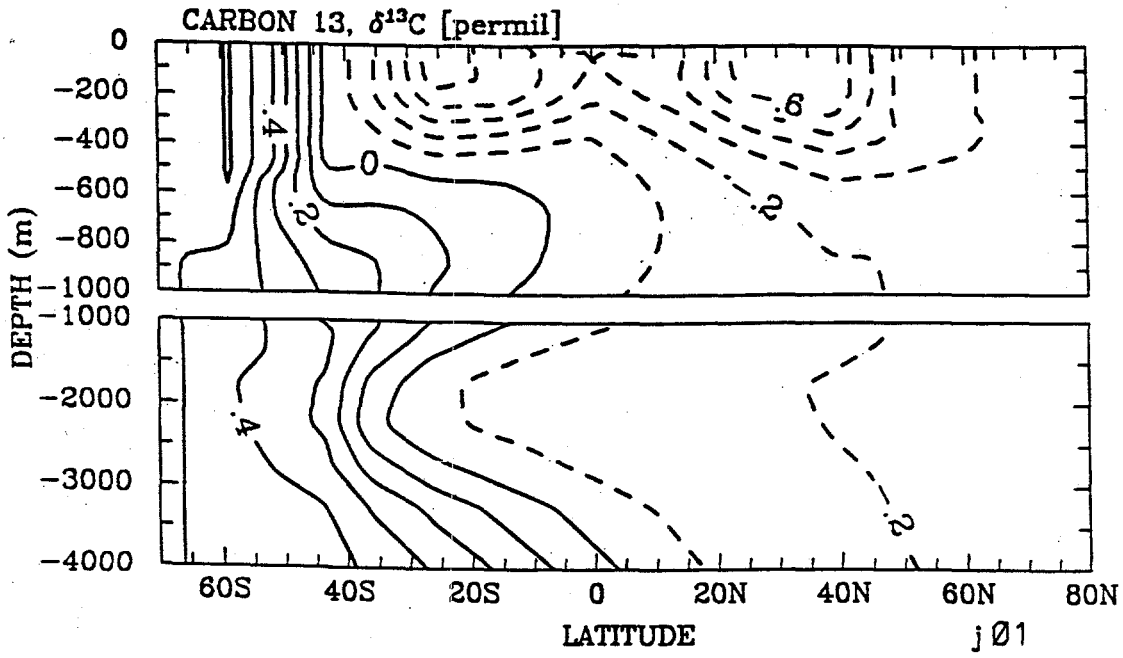
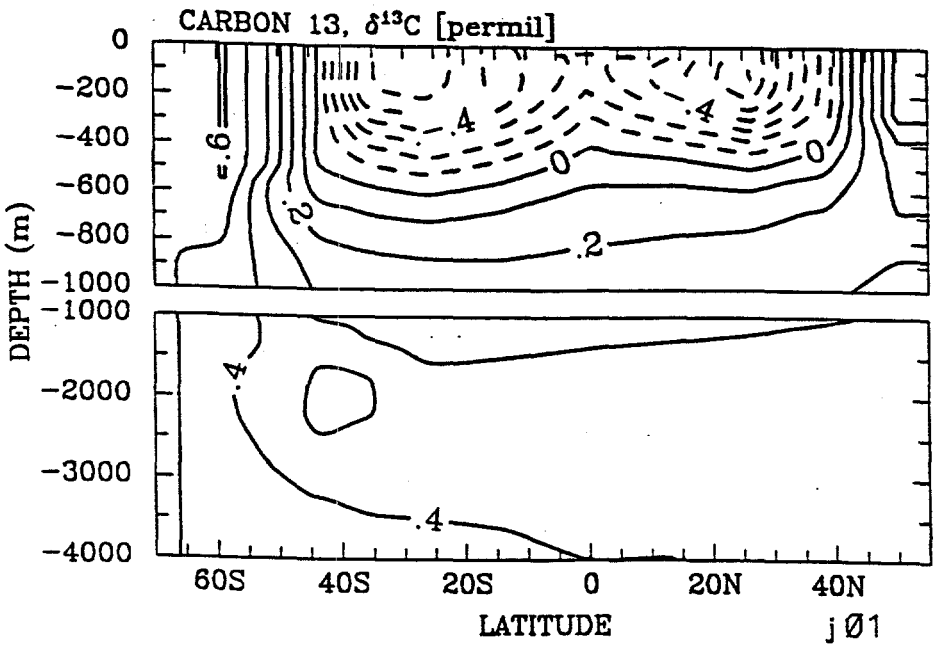

Mid Pacific (150W) Hudson April-May 1970
South of Australia (135E) Eltanin July 1968
West Indian GEOSECS Jan-Feb 1978
East Indian GEOSECS Feb-Mar 1978
North Atlantic TTO Apr-Oct 1981
δ13Cα,s
1.5
1.0
0.5
0.0
-0.5
-1.0
-1.5
-80
-40
0
40
80
Latitude

Figure 1


CARBON 13, δ13C [permil]
0
-200
-400
-600
-800
-1000
-1000
-2000
-3000
-4000
DEPTH (m)
.4
.2
0
.6
.2
.4
.2
60S
40S
20S
0
20N
40N
60N
80N
LATITUDE
j Ø1


CARBON 13, δ13C [permil]
0
-200
-400
-600
-800
-1000
-1000
-2000
-3000
-4000
DEPTH (m)
.6
.2
.4
.4
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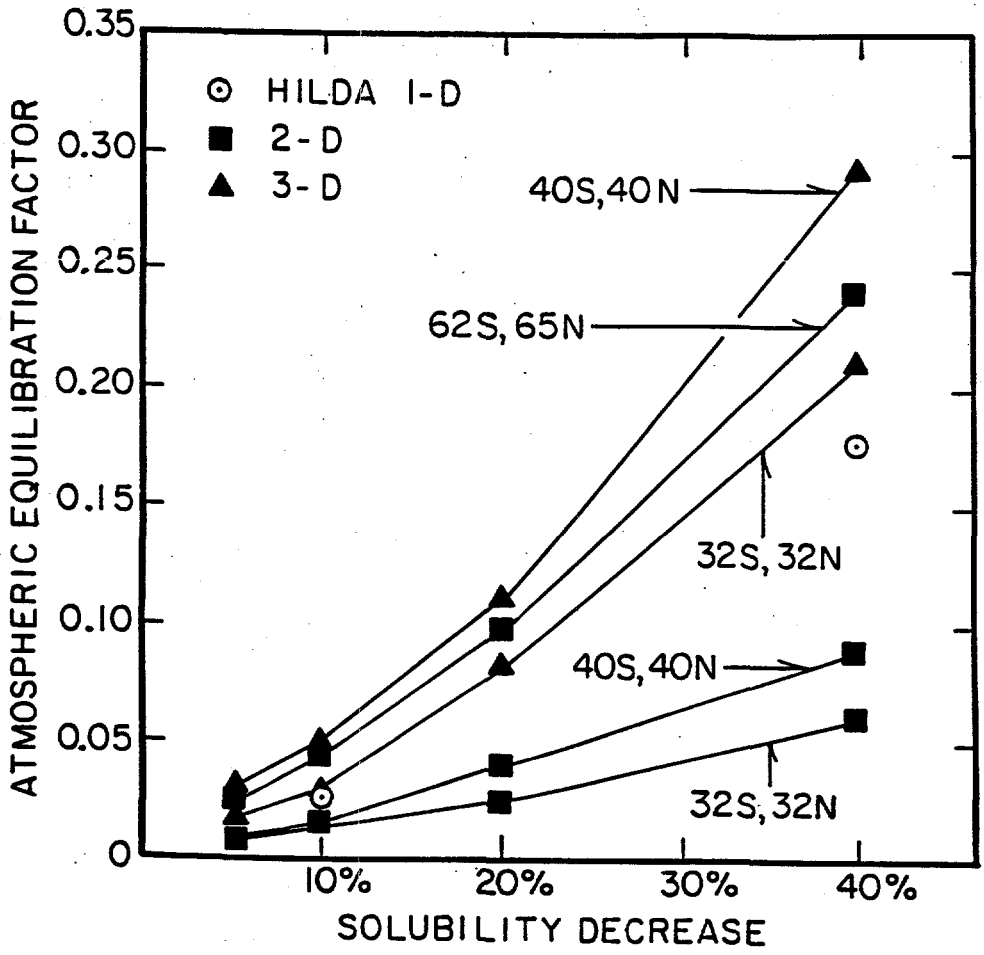
.4
.4
60S
40S
20S
0
20N
40N
LATITUDE
j Ø1

Figure 2


0.35
0.30
0.25
0.20
0.15
0.10
0.05
0
ATMOSPHERIC EQUILIBRATION FACTOR
HILDA 1-D
2-D
3-D
40S,40N
62S,65N
32S,32N
40S,40N
32S,32N
10%
20%
30%
40%
SOLUBILITY DECREASE