A Global Atmospheric Diffusion Simulation Model for Atmospheric Carbon Studies

G.I. Pearman and P. Hyson

ABSTRACT

A two-dimensional atmospheric transport simulation model is presented which is designed to be used for the modelling of the transport of tropospherically inert tracers such as CCl₃F and CO₂. The model is under development and in the present version only incorporates transport by diffusion.

Horizontal and vertical transport coefficients are adjusted to obtain model distributions of CCl₃F which are similar to those observed. The model is then used to simulate the annual variation of CO₂ concentration observed at various monitoring stations, by adjusting the monthly net CO₂ exchange at each latitude band.

It was found that earlier estimates of this exchange at high latitudes (> 53°N) in the northern hemisphere need to be increased by up to 150% and those at mid latitudes by 250%, in order to obtain agreement between simulation and observation.

A simulation of the stable carbon isotopic fractionation occurring during this net exchange of carbon produced local annual variations of the $^{13}C/^{12}C$ ratio which are in good agreement with observed values.

Further development of the model includes transport by advection and air-ocean exchange of carbon.

1. INTRODUCTION

Atmospheric diffusion simulation models are used extensively. Such simulations have been important in developing an understanding of the sensitivity of ozone to the addition of natural or anthropogenic trace gases, in assessing the hemispheric or global impact of regional releases of pollutants, and more recently, in examining situations where tropospheric or surface chemistry interacts with dynamics to influence the distribution of particular compounds. The present study involves the use of a diffusion simulation model to aid in the interpretation of primarily tropospheric observations of carbon dioxide (CO₂). The thesis has been that large-scale exchanges of carbon between the Earth's surface (the oceans and biosphere) and the atmosphere will be reflected in the distribution of the gas in the atmosphere. The quasi-steady state distributions of CO₂ in the atmosphere, and perturbations from that state...
on time scales of months or decades can be interpreted in terms of interreservoir exchanges and their variabilities. The model facilitates the interpretation of observations as well as providing a method for testing the likely atmospheric results of biogeochemical perturbations.

This report gives a description of an on-going study with some tentative results.

2. THE MODEL

The model is a zonally averaged, multiple box model in which vertical and horizontal transport is assumed to occur via Fickian diffusion. While the number of boxes in the model can be easily modified, the present description is of a model in which the atmosphere is divided meridionally into 20 boxes subtending equal surface areas. Each of these boxes is divided into five equal mass boxes in the vertical (i.e., at 200 mb intervals), with the upper "stratospheric" box being further subdivided into 4. Figure 1 shows the dimensions of each of the 160 boxes.

Vertical diffusion between boxes is accomplished through the flux-gradient relationship,

\[ F = \rho K_i \frac{\delta c}{\delta z} \]  

where \( \rho \) is density (g cm\(^{-3}\)), \( K_i \) is the eddy-diffusion coefficient relevant to exchange at location \( i \) (cm\(^2\) s\(^{-1}\)) and \( \frac{\delta c}{\delta z} \) is the vertical concentration gradient (mixing ratio cm\(^{-1}\)). The rate of change of concentration in the \( i \)th box due to exchange with the \( i + 1 \) box is given by

\[ \frac{dC_i}{dt} = \frac{F \cdot A_i}{M} \]  

where \( A_i \) is the cross sectional area through which the flux is proceeding and \( M \) the mass of the atmosphere in the \( i \)th box. Thus using finite differences,

\[ \frac{\Delta C_i}{\Delta t} = \frac{\rho \cdot K_i \cdot A_i}{M \cdot \Delta z} \cdot (C_{i+1} - C_i) \]  

where \( \Delta z \) is the distance between the centroids of the exchanging boxes. It can be further shown that the change in \( C_i \) over a time period \( \Delta t \) is given by

\[ \Delta C_i = \frac{C_{i+1} - C_i}{2} \left( 1 - e^{-\zeta \cdot \Delta t} \right) \]  

where,

\[ \zeta = \frac{2 \cdot \rho \cdot K_i \cdot A_i}{M \cdot \Delta z} \]  

The expression \( 1 - e^{-\zeta \cdot \Delta t} \) in equation (4) can be satisfactorily replaced by a qua-
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Figure 1: Schematic representation of the box structure of the present version of the diffusion simulation model.

A quadratic approximation which significantly decreases computation time. Time steps of one day have been used throughout these studies although for most purposes this could be increased by a factor of 2 or more. The expressions for horizontal diffusion are analogous.

Values for the horizontal eddy diffusivities were based on the empirical relationship between wind variance \( \sigma^2 \) and \( a \) reported by Czeplak and Junge (1974),

\[
K = \frac{1}{a} \sigma^2
\]

where \( a = 2.4 \times 10^{-5} \text{ s}^{-1} \). The global distribution of the standard deviation of the meridional component of wind was taken from Newell et al (1969). The resultant distribution of horizontal eddy-diffusivities is shown in Table 1a, while the resultant values for the product \( \zeta \cdot \Delta t \) are given in Table 1b.

The vertical diffusivities were based on those used by Chang (1976) and are tabulated together with the corresponding values for \( \zeta \cdot \Delta t \) in Tables 2a and b.

As the product \( \zeta \cdot \Delta t \) represents the exponent in equation (4), Tables 1b and 2b can be used to derive the time constant for exchange between boxes \( \tau \), where

\[
\tau = \frac{1}{\zeta}
\]

Inspection of the tables will reveal time scales for mixing vertically between boxes to be of order 2–13 days in the troposphere and of order one year in the stratosphere, while horizontal interbox mixing occurs on time scales of 2 to 20 days. In interpreting
these results one should remember that the interbox mixing lengths vary widely in both the vertical and horizontal.

3. VALIDATION OF DIFFUSION PARAMETERIZATION

The ability of the model to simulate the real atmospheric diffusion of gaseous constituents has been tested using trichlorofluoromethane (F-11; CC\(_1\)F\(_3\)) as a tracer. F-11 release data (McCarthy et al., 1977) from 1956 to 1977, together with the global meridional distribution of relative source strengths (Fraser, personal communication), were used to simulate the source of the tracer in the model. The model was run without any photochemical sink in the stratosphere. The results of the simulation were then compared to atmospheric observations of F-11 (e.g. Fraser and Pearman, 1978; Rasmussen et al., 1976; Inn et al. 1977; and unpublished data from NOAA-ARL-GMCC monitoring programmes).

Vertical gradients through the troposphere of a few parts per trillion by volume (pptv) were simulated by the model for recent years and these were consistent with observations. The model did, however, transfer slightly too much F-11 into the stratosphere so that the 1978 model stratosphere contained about 18% of the total atmospheric loading. This is somewhat higher than the estimated content of \(\sim 5\%\) (NAS, 1976) even given the lack of a photochemical sink.

Meridional mixing in the model was slightly less effective than in the real atmosphere, giving a difference in hemispheric mean concentration from north to south of 5.0% of the global mean concentration, compared with an observed difference of 4.8%.

In general the agreement between simulation and observation was encouraging considering the method of representing the transport and the fact that advection was not simulated. Improvement of the model with respect to the method of representing both eddy-diffusion, advection, seasonality and anisotropy is continuing. However, it is believed that further improvements will not greatly influence the general conclusions presented below.

4. BIOSPHERIC EXCHANGE OF CARBON

The biospheric exchange of carbon is the prime cause of variability of CO\(_2\) concentrations in the atmosphere. Machta (1971) used estimates provided by H. Leith of the seasonal variation in biospheric CO\(_2\) exchange (Table 3) in his own simulation model. In the present study, these data were subdivided into geographical zones applicable to the CSIRO model (Table 4).

When the model is run with these biospheric data acting on the surface boxes, we find, as did Machta, that the perturbations generated in the northern hemisphere atmosphere are approximately a factor of two smaller than those observed.

Verification of the simulation output was made against the mean annual cycle in concentration observed at Barrow, Alaska (Komhyr and Harris, 1976), in the upper atmosphere over the north Atlantic (Bolin and Bischof, 1970), at Mauna Loa, Hawaii
A global atmospheric diffusion simulation model (Keeling et al., 1976a; and C.D. Keeling, personal communication), in the mid-troposphere over Bass Strait, Australia (Beardsmore et al., 1978) and at the South Pole (Keeling et al., 1976b).

If we assume that seasonal variations in oceanic CO₂ exchange do not contribute significantly to the annual variation in atmospheric CO₂ concentration, we can proceed to modify the biospheric input data to produce an agreement between the simulation and observations. In particular, if we commence with the assumption that the phase of the annual exchanges as represented in Table 4 are correct, then, the amplitude and phase of variations simulated for a particular location can result only from the modification of the magnitude of exchange rates at each latitude band.

By running the simulation a number of times, scaling the biospheric exchange by different amounts for each latitude band, it was possible to obtain a distribution of exchange rates which give reasonable amplitude and phase agreement between the simulation and observation for all locations with the exception of Mauna Loa. It was found that the simulation consistently gave an annual cycle at Mauna Loa which preceded the observed cycle by ~3 weeks. High latitude northern hemispheric exchanges, the influence of which are lagged by the time they diffuse to Mauna Loa latitudes, would need to be unreasonably large to phase shift the effects of local exchanges. Indeed, it was found necessary to delay the timing of the low latitude exchange (10°-30°) rates as given by Lieth by one month to match the observed observations.

The results of the simulation, after these modifications to the biospheric input, are shown in Figure 2 through 5, while the global distribution of biospheric exchanges required to achieve this level of agreement between simulation and observation, is given in Table 5. The agreement is considered to be satisfactory considering that further refinements will require attention to second-order effects such as the improved parameterization of the diffusion itself, anisotropy, the inclusion of advection, the possible role of oceanic-induced seasonality, the asymmetrical distribution of exchange within a season, and the ability of a single observing station to represent a section of the atmosphere represented by a box in the diffusion model.

It should be mentioned that in order to obtain this level of agreement, southern hemisphere exchange data were not modified from the Lieth data. It was considered reasonable that net exchange should approach zero equatorwards of ~10°N and ~10°S. High latitude (~53-90°N) exchange rates were increased by a factor of 2.5 above those of Lieth, and those for latitudes between ~10 and 37°N were increased by a factor of ~2. The greatest increase was required in the region of ~37-53°N where the exchange was increased by a factor of 2.5 above that suggested by the original Lieth data.

While the amplitude of the annual concentration variations simulated in the southern hemisphere agree reasonably well with the observations without any adjustment of the local exchange rates, the phase agreement is not so good. When the model is run with the southern hemisphere biosphere turned off, one finds that ~2/3 of the amplitude of the southern hemisphere CO₂ variation is due to northern hemispheric influence. Given the large distances for transport between the northern
Figure 2: Annual variation of CO₂ concentration at Barrow, Alaska, both observed (−) and modelled (−−−).

Figure 3: As Figure 2 for the 5–7 km altitude range over the North Atlantic.

Figure 4: As Figure 2 for Mauna Loa, Hawaii.

Figure 5: As Figure 2 for Samoa, Bass Strait and the South Pole.
hemisphere areas contributing to annual variability and the observation sites in Australia and Antarctica, it is reasonable to conclude that phase errors due to the model may be responsible at least in part for these difficulties. However, it is also most likely that seasonality in oceanic exchange might play a significant role in establishing the annual variability in the southern hemisphere.

5. STABLE CARBON ISOTOPES

The simulation model is constructed so that carbon-12 and carbon-13 are accounted for independently thus enabling the ratio of these isotopes to be determined at any location or time. While experiments with this facility are preliminary, the following simulation may be of interest.

As an initial condition, the model atmosphere is given a C\(^{13}/C\(^{12}\) ratio corresponding to \(\delta^{13}= -7\%\). We observe that when atmospheric CO\(_2\) of this isotopic composition is assimilated by plants, the biospheric carbon which results obtains a value of \(\delta^{13}= -25\%\). This means that the atmospheric ratio of \(C^{13}/(C^{12} + C^{13})\) is modified due to biological fractionation by a factor 0.9821.

The model was therefore run with the condition that whenever biospheric uptake occurred, fractionation was affected according to the above value. The second condition was that all carbon returned to the atmosphere during biospheric release was released with a C\(^{13}/C\(^{12}\) ratio of 0.010956. The results of this simulation are given in Figure 6. Clearly the deficiency in such a simulation is that it ignores the gross exchange between the carbon which occur between the atmosphere and ocean and atmosphere and biosphere. Development of techniques to cope with this is progressing. The result of such exchange is that a given atmospheric perturbation in isotope ratio will be to some extent diluted into a larger effective reservoir. However, it is possible, at least in the case of biospheric turn-over, that the mixing into the biospheric reservoir is rather incomplete. During the period of the year when there is a net uptake of carbon by the biosphere, the daytime (or gross) uptake exceeds the net uptake by an amount of carbon which is returned to the atmosphere almost immediately or during the next night. If the source of this respiratory carbon is the carbon which was taken up within the previous few days, then the isotopic composition of the carbon release will be identical to that taken up, and no diluting effect will be observed. To some extent this will be the case with atmospheric-ocean exchange, although in that case, vertical mixing is more likely to influence the net isotopic effect.

With these comments in mind, it is interesting to compare the simulated annual isotopic variations with those few observations available. Keeling (1961) shows a seasonal variation in \(\delta^{13}\) for air “over barren ground and over the Pacific Ocean near the west coast of North America”, to be \(\sim 0.5\%\) and Keeling et al. (1979) show variations at several North American sites and Mauna Loa, which are of similar magnitudes and thus consistent with the model results. In both papers the relationship between CO\(_2\) concentration and isotopic variation is clearly demonstrated. Data of Craig (unpublished data) for Mauna Loa and the North Pacific show seasonal ampli-
tudes for $\delta^{13}$ of ~0.3‰ and 0.6‰ respectively. This is again consistent with the magnitude of variation produced by the model.

![Graph](image)

**Figure 6:** Model generated annual variation of CO$_2$ stable isotopic ratio expressed as $\delta^{13}$ where

$$
\delta^{13} = \left( \frac{(C^{13}/C^{12})_{\text{air}}}{(C^{13}/C^{12})_{\text{standard}}} - 1 \right) \cdot 1000.
$$
A global atmospheric diffusion simulation model

Table 1: a) Horizontal eddy diffusion coefficients used in the simulation model. Box and level numbers refer to locations in atmosphere shown in Figure 1.

b) Meridional eddy diffusion transfer factors, the inverse of which represents the characteristic mixing time between respective box centroids.

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1b. Meridional transfer factors (days^{-1})

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Table 2: As for Table 1 except that coefficient and transfer factors refer to eddy diffusion in the vertical.

### 2a. Vertical diffusion coefficients (m²/s)

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### 2b. Vertical transfer factors (10⁻³ day⁻¹)

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Table 3. Carbon dioxide release (+) to or uptake (−) from the atmosphere, $10^{11}$ kg C month$^{-1}$. From Machta (1974).

Biospheric sources and sinks

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Table 4: Average monthly biospheric exchange per model latitude band (g CO₂) based on the data in Table 3.

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Total g CO₂ per month
.333E+14 .357E+14 .615E+14 .638E+14 .228E+14 -.151E+15

Table 5: Average monthly biospheric exchange of carbon dioxide (kg CO₂ per latitude band) required to simulate the observed annual variations of atmospheric CO₂ concentration. Each latitude band has an area of 2.55X10⁷ km².

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Total g CO₂ per month
A global atmospheric diffusion simulation model

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