CARBON EXCHANGE BETWEEN ATMOSPHERE AND OCEANS IN A LATITUDE-DEPENDENT ADVECTION-DIFFUSION MODEL

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ABSTRACT. Marine transport of inorganic and organic carbon is simulated by means of a computer model in which the oceans are divided into a high and low latitude region. Water transport (and with it carbon transport) is reproduced 1) as downwelling of surface waters at low latitudes, and 2) in general, as different depth-dependent turbulent diffusion in both deep-sea regions. The model is calibrated with pre-bomb ^{14}C and validated against perturbations of total carbon, $^{13}C/C-$ and $^{14}C/C-$ ratios; it is compatible with carbon release from fossil fuels and from biogenic sources.

INTRODUCTION

In view of possibly serious climatic consequences of increased atmospheric CO_2 (greenhouse-effect), efforts have been made to model the cycling of carbon through the global environment. The reliability of a simulation model is increased according to the number of independent observations it can verify. A major problem is whether the terrestrial biota are to be regarded as a source or a sink for additional atmospheric CO_2. The model presented here has been developed to account for at least part of an additional CO_2 input into the atmosphere resulting from forest destruction and desertification.

ATMOSPHERE AND BIOSPHERE

The model atmosphere is subdivided into tropospheric and stratospheric parts. The total atmospheric CO_2 content in the initial stationary state was taken to be 608Gt C ($\hat{=}$ 285ppm). Exchange between the two atmospheric compartments is linearly donor-controlled with a stratosphere-to-troposphere exchange time of 2 years (Machta, 1973). Data for fossil fuel input into the troposphere is taken from Rotty (1981) for the time period 1860-1978 (Fig 1). Natural production of ^{14}C in the stratosphere ($P_{nat}^{14}(t)$) is assumed time dependent according to the variation of Aa indices (Stuiver and Quay, 1980):

$$P_{nat}^{14}(t) = P_o^{14} + (0.43 - 0.024 \cdot Aa) \frac{O_e \cdot \theta}{N_A} \qquad (1a)$$
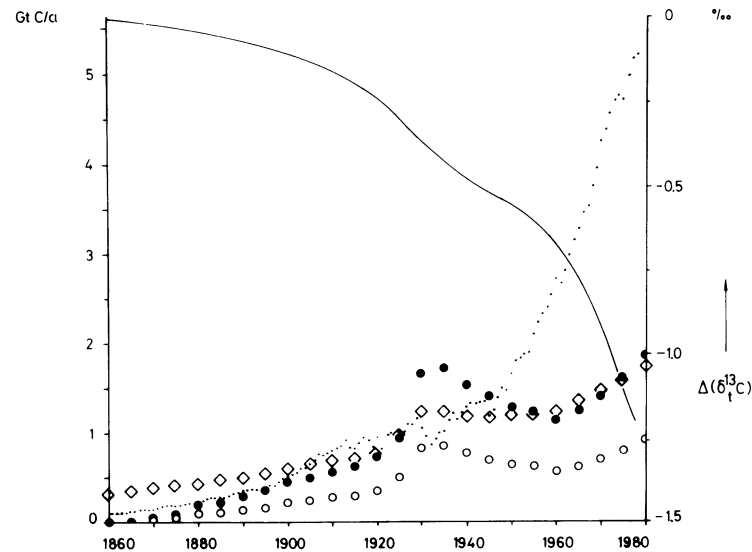
Fig 1. $\Delta(\delta_t^{13}C)$ in response to total carbon inputs into the troposphere (solid line)
.... = fossil fuel release
oooo = net biogenic input as introduced into the model calculations
        $(F_{b,net})$
●●●● = total biogenic release $(F_{b,tot})$

◇◇◇ = $2 \cdot F_{b,net}$ as estimated in Kohlmaier et al (1981)

O_e : earth's surface $5.1 \cdot 10^{18} cm^2$, θ: $31.6 \cdot 10^6 sec/a$, N_A: Avogadro's number.
P_o^{14}: stationary ^{14}C production. The bomb ^{14}C production $(P_b^{14}(t))$ is taken from Killough (1980). The total ^{14}C production is then

$$P^{14}(t) = P_{nat}^{14}(t) + P_b^{14}(t) \qquad (1b)$$

        Since the description of the oceanic part of the global carbon cycle is our primary concern here, only a summary of the terrestrial biosphere part will be given (see Kohlmaier et al, 1981, for details). The terrestrial biota are subdivided into five major biomes (tropical forests, tropical grasslands, extra-tropical forests, extra-tropical grasslands and deserts) each consisting of living biomass and humus. In order to model regrowth of cleared forest areas, the living biomass of these biomes was further subdivided into 20 compartments that extend over varying areas each representing a certain growth state. Under such assumptions we can model a steady state for a constant forest clearing rate. The total amount of CO_2 released through the perturbation of the biota is $F_{b,tot}$, the net flux

to the atmosphere resulting from deforestation ($F_{b,net}$) is then given by the difference between $F_{b,tot}$ and the net ecosystem production ($F_{b,NEP}$) (fig 1).

   Non-tropical forests are assumed to be affected by wild-fires and clearings, whereas tropical forest areas are diminished by commercial clearings and increased permanent agriculture. Tropical forest areas cleared for shifting cultivation are assumed to be regrown, and soil erosion and charcoal formation are accounted for (Buringh, 1979; Seiler and Crutzen, 1980). Former model calculations (Kohlmaier et al, 1981) yielded a total net biospheric input of ca 2Gt C/a during 1980. Best agreement with the measured atmospheric ^{13}C data is obtained, if 50 % of this input is introduced into the model calculations.

OCEANS

The model ocean consists of two regions of high (c) and low (l) latitudes, respectively (fig 2).
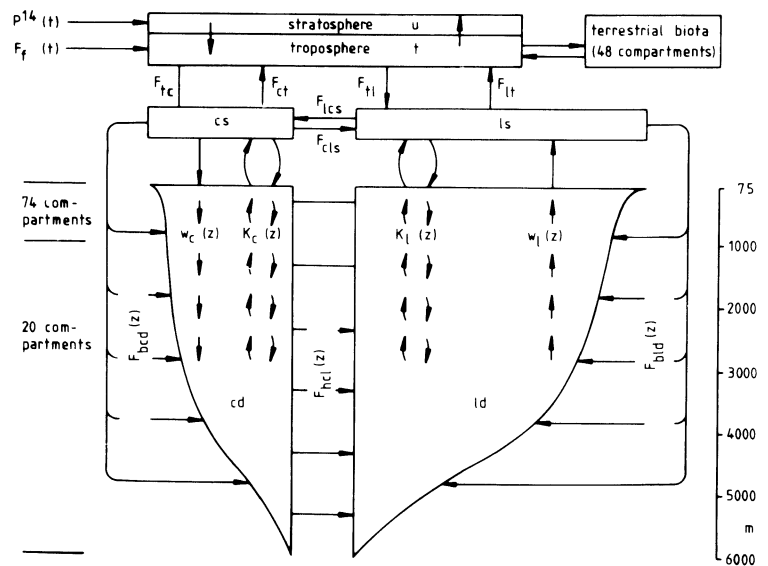

Fig 2.   Compartments and fluxes in the atmosphere-ocean-sub-system of the carbon cycle model
l = ocean region of warm surface water or low latitudes
c = ocean region of cold surface water or high latitudes
s = surface ocean compartments
d = deep sea
F_{bcd}, F_{bld} = sinking of biogenic matter and dissolution in the high
          and low latitudes deep sea, respectively
F_{hcl} = horizontal flux due to water mass movement
w_c, w_l = sinking and upwelling velocities
K_c, K_l = eddy diffusion constants in the deep sea of high and low
          latitudes, respectively

The subdivision represents 1) different CO_2 uptake behaviors
of cold and warm surface waters resulting in a net flux of CO_2
from equatorial oceans via the atmosphere to high latitudes
surface waters, 2) different turbulent diffusivities in each
deep sea region, 3) advection, i e, the sinking of cold surface
water of high latitudes which spreads into low latitudes in the
deep sea where it upwells. The surface water compartments are
connected by fluxes in either direction, the difference of
which corresponds to the net deep circulation of water masses.
In the computer model it is possible to set the boundary
latitude between the two ocean regions alternatively at either
40, 50, or 60°N and S, respectively, but only one model with an
ocean subdivision at 40° is presented here. This model structure
does not take into account that regions of deep water formation
are rather restricted, nor does it show that at least part of
the surface water that sinks is not derived from horizontally
admixed surface water but from deeper layers and thus can only
partially reach isotopic equilibrium with atmospheric $^{13}C/C$
and $^{14}C/C$.

     The depth of both surface compartments is chosen as 75m.
The depth interval from 75 to 1000m has been subdivided into
37 compartments in each region, each extending over a depth of
25m. The deep sea from 1000 to 6000 m is represented by 10 com-
partments 500m deep in each region (Oeschger et al, 1975). The
volume of each compartment in either region is determined from
the hypsometric curve according to Menard and Smith (1966).

     Exchange between atmosphere and oceans is proportional to
the respective CO_2 partial pressures:

$$F_{tc} = G_c/V_m \cdot p_t \qquad\qquad F_{ct} = G_c/V_m \cdot p_{cs} \qquad\qquad (2a,b)$$

$$F_{t1} = G_1/V_m \cdot p_t \qquad\qquad F_{1t} = G_1/V_m \cdot p_{1s} \qquad\qquad (2c,d)$$

V_m: molar volume, $0.0224m^3$/mole, G_c, G_1: total gas exchange (in
m^3/a) in either ocean region, where $G_c + G_1 = G$.

     The CO_2 partial pressures in the surface ocean compartments
are calculated from the corresponding ΣC considering the
chemical equilibria which determine the pH in surface waters
(calculation procedure, see Bacastow, 1981). Temperatures of 11
and 25°C, respectively, in either surface compartment directly
influence solubility and other equilibrium constants.

     Fluxes of isotopes ^{13}C and ^{14}C from compartment i to j are
generally formulated as

$$F_{ij}^{13} = \overline{\alpha_{ij}^{13}} \, Q_i^{13} \, F_{ij} \qquad\qquad F_{ij}^{14} = \overline{Q_i^{14}} \, F_{ij} \qquad\qquad (3a,b)$$

$\overline{\alpha_{ij}^{13}}$ = corrected fractionation factor for ^{13}C; Q^{13} = isotopic ratio of mole numbers n^{13}/n or concentrations c^{13}/c (n and c refer to total carbon)

Q^{14} = isotopic ratio for $^{14}C/C$ from fractionation corrected $\Delta^{14}C$ values: $Q^{14} = (\Delta^{14}C/1000 + 1) \cdot Q_{std}^{14}$

$Q_{std}^{14} = 1.176 \cdot 10^{-12}$ (Karlén et al, 1964); F_{ij} = total carbon flux

$\overline{\alpha_{ij}^{13}}$ is derived from α_{ij}^{13} according to

$$\overline{\alpha_{ij}^{13}} = \alpha_{ij}^{13}/[1-(1-\alpha_{ij}^{13})\cdot Q_{i0}^{13}] \qquad (4)$$

0 = stationary state. α_{tc}^{13} and α_{tl}^{13} are calculated according to Vogel, Grootes and Mook (1970), yielding corrected values of 0.99837 and 0.99842 for high and low latitudes, respectively. Considering isotopic equilibration among CO_2, HCO_3^- and CO_3^{2-}, as given by Emrich, Ehhalt and Vogel (1970), leads to values for $\overline{\alpha_{ct}^{13}}$ (0.98872) and $\overline{\alpha_{lt}^{13}}$ (0.99028).

Variation of total carbon content (in mole numbers n) in deep-sea compartment, i, of either region is described as:

$$\dot{n}_{cd}(i) = F_{ad,c}(i) - F_{ad,c}(i+1) - F_{hcl}(i)$$
$$- F_{diff,c}(i) + F_{diff,c}(i+1) + F_{bcd} \qquad (5a)$$
$$\dot{n}_{ld}(i) = F_{ad,l}(i+1) - F_{ad,l}(i) + F_{hcl}(i)$$
$$+ F_{diff,l}(i+1) - F_{diff,l}(i) + F_{bld}(i) \qquad (5b)$$

Vertical advection fluxes are denoted by F_{ad}, horizontal fluxes being F_{hcl}; F_{diff} are turbulent diffusion fluxes and F_{bd} represent the transport of biogenic matter to the deep sea and its subsequent dissolution. Indices, i, refer to the upper boundary of compartment, i. The individual fluxes are

$$F_{ad,c}(i) = W_d(i)\, c_{cd}(i-1) \qquad F_{ad,l}(i) = W_d(i)\, c_{ld}(i) \qquad (6a,b)$$

$$F_{hcl}(i) = [W_d(i) - W_d(i+1)]\, c_{cd}(i) \qquad (6c)$$

$$F_{diff,c}(i) = \kappa\, O_s\, \overline{h}(i)\, K_c(i)[c_{cd}(i) - c_{cd}(i-1)]/\Delta z(i) \quad (6d)$$

$$F_{diff,l}(i) = (1-\kappa)O_s\,\overline{h}(i)\, K_l(i)[c_{ld}(i) - c_{ld}(i-1)]/\Delta z(i) \quad (6e)$$

$W_d(i)$ is the water mass passing through the upper boundary area $O_s\overline{h}(i)$, such that $W_d(i) = \kappa\, O_s\, \overline{h}(i)\, w_c(i) = (1-\kappa)\, O_s\, \overline{h}(i)\, w_l(i)$ where κ is the ratio of high latitudes region to total ocean

surface area (0.332); $w_c(i)$ and $w_1(i)$ are downwelling and up-
welling velocity, respectively. $K_c(i)$ and $K_1(i)$ are turbulent
diffusion constants. The boundary conditions for the two bottom
compartments are:

$$W_d(i+1) = 0, \quad K_c(i+1) = 0, \quad K_1(i+1) = 0 \tag{7a-c}$$

The related differential equations for ^{13}C and ^{14}C differ sig-
nificantly in two ways: 1) for ^{14}C, the decay term $\lambda n_{cd}^{14}(i)$
and $\lambda n_{1d}^{14}(i)$ must be added, 2) the isotopic composition of the
biogenic flux is always assumed proportional to the respective
biotic isotopic ratios.

The time behavior of the surface compartments can now be
described by:

$$\dot{n}_{cs} = F_{tc} - F_{ct} + F_{1cs} - F_{cls} - F_{ad,c}(1) + F_{diff,c}(1) - F_{csb}$$

$$\dot{n}_{ls} = F_{t1} - F_{1t} + F_{cls} - F_{1cs} + F_{ad,1}(1) + F_{diff,1}(1) - F_{lsb}$$

$$\tag{8a,b}$$

where F_{cls} and F_{1cs} denote the horizontal fluxes connecting the
surface compartments.

$$F_{cls} = W_{cls} \, c_{cs} \qquad\qquad F_{1cs} = W_{1cs} \, c_{ls} \tag{9a,b}$$

$$W_{1cs} - W_{cls} = W_d(1) \tag{9c}$$

$$F_{csb} = \sum_i F_{bcd}(i) \qquad\qquad F_{lsb} = \sum_i F_{bld}(i) \tag{9d,e}$$

According to equations (5a,b) and (6a-e), the concentrations
for the stationary state ($\dot{n}_{cd0} = \dot{n}_{1d0} = 0$) in compartments
(i-1) can be derived from the stationary concentrations in
compartments i and (i+1). Under consideration of the additional
decay term for ^{14}C analogue equations yield the stationary
concentrations for the rare carbon isotopes. From boundary
conditions (7) follows that, once a set of bottom concentrations
for each of the state variables is given, all stationary deep-
sea and surface-ocean compartment concentrations can be com-
puted subsequently.

The calibration of the model deep-sea proceeds as follows:
1) Depth-dependent functions for $W_d(i)$ and $K_c(i)$ are assumed.
2) $\Delta_{cd0}^{14}C(47)$ is varied so that a certain value for $\Delta_{1s0}^{14}C$
results.
3) $F_{bcd}(i)$ is varied to reproduce data for $c_{cd0}(z)$.
4) $\delta_{cd0}^{13}C(47)$ is varied in such a way that the resulting
$\delta_{cs0}^{13}C$ falls within the range of 1.5-2.5‰.

5) $K_1(i)$ is determined for each compartment separately by varying its numerical value to agree with a prescribed depth profile of $\Delta_{1d0}^{14}C(i)$ (fig 4).

6) $F_{bld}(i)$ is varied to approximately reproduce $c_{1d0}(i)$ which may lead to major changes in $\Delta_{1d0}^{14}C(i)$ so that 5) has to be repeated.

7) $\delta_{cd0}^{13}C(47)$ is varied to reproduce a prescribed $\delta_{1d0}^{13}C(i)$ depth profile.

1) $W_d(z)$ is assumed a function with two inflection points, one at 1000 and one at 4000m, reflecting both intermediate and deep water spreading into the low latitudes deep sea. Equal volumes of intermediate and deep water are formed per unit time. A value of 15Sv for the deep-water formation is adopted here which is on the low end of the respective estimation range (Dietrich et al, 1975; Arons and Stommel, 1967; Weiss, Östlund and Craig, 1979; Broecker, Peng and Engh, 1980). On the whole, 30Sv ($\cong 10^{15} m^3/a$) are assumed to sink down in high latitudes resulting in an upwelling velocity of ca 5m/a in low latitudes just below the surface compartment.

$K_c(z)$ is described by

$$K_c(z) = 10^5 m^2/a \cdot \exp(-z/500m) \qquad (10)$$

yielding a depth-averaged $\overline{K}_c = 8330 \ m^2/a$,

2) As discussed below, the model requires the assumption of a value of $\Delta_{cs0}^{14}C = -35\%o$. This can be explained by the fact that the model does not allow for relatively "old" high-latitude deep water to upwell and become part of the surface water in this region.

3) $c_{cd0}(z)$ is taken from Takahashi, Broecker and Brainbridge (1981, table S-14, GEOSECS Antarctic data). A mean water density of $1.026g/cm^3$ was used to convert mmoles/kg to moles/m^3. $F_{bcd}(i)$ and $F_{bld}(i)$ are normalized exponential functions (table 1).

4) A stationary value of $\delta_{bsc0}^{13}C = -20 \ \%o$ is used.

5) $\Delta_{1d0}^{14}C(i)$ for the depth range from 1000 to 6000m is calculated as an area weighted mean of the GEOSECS data (Stuiver, Östlund and McConnaughey, 1981) for the Atlantic and Pacific ocean (50°N to 50°S). These layers are probably left essentially unaffected by perturbations of the atmospheric $^{14}C/C$ ratio. From the pre-industrial values for $\Delta_{1s0}^{14}C$ estimated by Stuiver (1980) (-40%o) and determined from corals (Druffel and Linnick, 1978) (-50%o) the average of -45%o is adopted here. To obtain $\Delta_{1d0}^{14}C(i)$ for 75 to 1000m depth, a smooth curve connecting the

TABLE 1. Transfer coefficients of deep-sea model for 75 and 1000m

z (m)	K_c (m^2/a)	K_1 (m^2/a)	w_1 (m/a)
75	100,000	8000	5
1000	16,500	800	3.1

F_{csb} = 24Gt C/a

F_{lsb} = 8.4Gt C/a   75% NPP_{oc}

$\int\limits_{z=1000m}^{6000m} F_{bcd}(z)dz = 1.25$Gt C/a

$\int\limits_{z=1000m}^{6000m} F_{bld}(z)dz = 0.18$Gt C/a   3.3% NPP_{oc}

NPP_{oc}: marine net primary production (De Vooys, 1979: 43.5GtC/a)
25% NPP_{oc} are dissolved in the surface ocean.

$\Delta^{14}_{ld0}C$ values of the lower layers with $\Delta^{14}_{ls0}C$ is constructed. This curve can be varied to obtain a $K_1(i)$ distribution that best reproduces the bomb ^{14}C input into the thermocline.

6) $c_{ld0}(z)$ is taken as an area weighted mean of the data of Takahashi, Broecker and Bainbridge (1981). As in 3), a mean water density of 1.026g/cm^3 is used. 7) According to Craig (1970), a mean $\delta^{13}_{bls0}C = -15\%o$ is inserted, which means that ca 30% of the dissolved material is $CaCO_3$.

The stationary state assumption for the entire ocean with respect to ΣC and ^{14}C is then used in calibrating atmosphere ocean exchange:

$$0 = G/V_m \cdot p_{a0} - G_c/V_m \cdot p_{cs0} - G_1/V_m \cdot p_{ls0} \tag{11a}$$

$$0 = G/V_m \cdot \overline{Q^{14}_{t0}} \cdot p_{a0} - G_c/V_m \cdot \overline{Q^{14}_{cs0}} \cdot p_{cs0} - G_1/V_m \cdot \overline{Q^{14}_{ls0}} \cdot p_{ls0} - \lambda n^{14}_{oc,0} \tag{11b}$$

($n^{14}_{oc,0}$: total stationary ^{14}C inventory in the oceans)

Because of $G = G_1 + G_c$, equations (11a,b) yield:

$$G = V_m \cdot \lambda \cdot n^{14}_{oc,0} \cdot \{\overline{Q^{14}_{t0}} \cdot p_{a0} - [\overline{Q^{14}_{cs0}} \cdot (p_{ls0} - p_{a0})p_{cs0} + \overline{Q^{14}_{ls0}} \cdot (p_{a0} - p_{cs0})p_{ls0}]/(p_{ls0} - p_{cs0})\}^{-1} \tag{12}$$

The water mass transport from low to high latitudes is calculated using a stationary state for n^{14}_{cs0}:

$$W_{lcs} = \{F_{ct0}^{14} - F_{tc0}^{14} - \kappa O_s \overline{h}(1)/\Delta z(1) \, K_c(1) \, [c_{cd0}^{14}(1) - c_{cs0}^{14}]$$

$$+ \lambda \, n_{cs0}^{14} + F_{csb}^{14} \}/(c_{ls0}^{14} - c_{cs0}^{14}) \tag{13}$$

Since p_{cs0} and p_{ls0} are essentially unknown, an iterative procedure beginning with arbitrary values for p_{cs0} and p_{ls0} is employed to obtain stationarity for atmosphere and surface ocean compartments with respect to ΣC and ^{14}C. The entire atmosphere-ocean model is then run without perturbation for 360 years to yield stationary $^{13}C/C$ ratios which may also slightly shift p_{cs0}, p_{ls0}, and $\Delta_0^{14}C$ values.

The calculated atmosphere ocean exchange time τ_{as} = 6.7a corresponds to a stationary exchange flux of 90.9Gt C/a or a mean CO_2 exchange of 20.9 moles/m^2/a. The value for τ_{as} given here is to be compared with 5 to (15+5)a (Broecker and Peng, 1974; Oeschger et al, 1975; Peng et al, 1979), the center of estimates being between 6-8a. For the stationary state $F_{1t0} - F_{t10}$ = 3.9Gt C/a is calculated.

TABLE 2. Quantities related to atmosphere-ocean-exchange

p_{cs0} = 242ppm	p_{lso} = 302ppm
F_{tc0} = 26Gt C/a	F_{t10} = 64.9Gt C/a
F_{ct0} = 22.1Gt C/a	F_{1t0} = 68.8Gt C/a
τ_{ts} = 5.7a	τ_{as} = 6.7a
W_{cls} = 0.179·10^{15}m^3/a	W_{lcs} = 1.309·10^{15}m^3/a

RESULTS AND DISCUSSION

As stated above, the atmospheric initial partial pressure and the biospheric input were chosen to reproduce the tropospheric partial pressure and $\delta_t^{13}C$. The average CO_2 partial pressure as measured at Mauna Loa and at the South Pole (Bacastow and Keeling, 1981) was 316ppm in 1959 and 334.3ppm in 1978. The model calculations yield values of 314.2ppm for 1959 and 337.1ppm for 1978. Whereas the measured average airborne fraction with respect to fossil input was 49.3%, the airborne fraction as obtained from the model is 42.2%. For the entire calculation (1860-1980), an airborne fraction of 49.6% results.

$\delta_t^{13}C$ shows a difference of -0.57‰ between 1956 and 1978 which is within the uncertainty range of (-0.65 + 0.13)‰ as obtained from atmospheric ^{13}C measurements (Keeling, Bacastow, and Tans, 1980). Clearly, the biogenic model input can be increased for $\Delta(\delta_t^{13}C)$ to closer approach -0.65‰ for the period in question, but then agreement with the atmospheric CO_2 par-
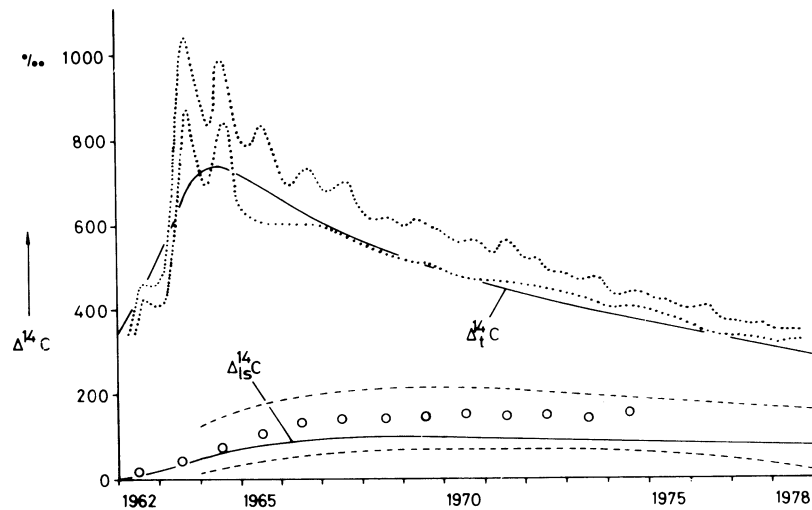
Fig 3. $\Delta^{14}C$ in troposphere ($\Delta_t^{14}D$) and low latitudes surface ocean ($\Delta_{ls}^{14}C$) in response to bomb ^{14}C input (solid lines); envelopes of significant data for the troposphere (....) and for surface ocean (- - - -) (Nydal, Lövseth, and Skogseth, 1980); ooo = coral measurements (Druffel and Linick, 1978).

tial pressure data would be decreased. The $\Delta(\delta_t^{13}C)$ for the period 1860 - 1940 is -0.48%o which neither agrees with Freyer's (1979) tree-ring measurements (-1.6%o) nor with Francey's (1981) (0%o).

The model response to the ^{14}C bomb input is shown in figure 3, which shows $\Delta^{14}C$ in the troposphere and the low latitudes surface ocean as calculated. The agreement with data observed for the troposphere seems good although the calculated values lie at the lower end of the data range. $\Delta_{ls}^{14}C$ is significantly lower than the coral measurements. In view of the broad band of data, however, higher reliability cannot be expected. In general, the model response to ^{14}C perturbations is characterized by a more pronounced relaxation than is observed. This may be explained by, at least, one of the following: 1) K_c is assumed too large; 2) in contrast to the real ocean, cold surface water is described as attaining nearly perfect isotopic equilibrium with the atmosphere, thereby taking up more ^{14}C and ^{13}C than observed; 3) the ^{14}C bomb input, as estimated by Killough (1980), is too small. 1) and 2) lead to an overemphasized uptake capacity of the high latitudes deep sea.
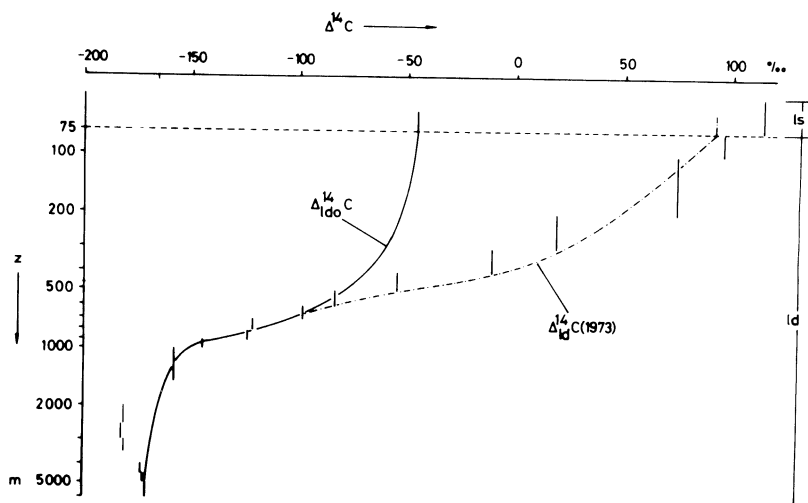
Fig 4. Depth distribution of $\Delta^{14}C$ in low latitudes for a stationary stage (solid line) and for the year 1973 (_·_·_·) as compared to averages of the GEOSECS data (Stuiver and Quay, 1980) ($|\,|\,|\,|\,|$).

The behavior of bomb ^{14}C in the low latitudes deep sea as calculated from the model is given in figure 4; the agreement with the GEOSECS data for 1973 seems good. The Suess effect is calculated as $\Delta(\Delta_t^{14}C) = -18.6‰$ in 1950 ($\Delta_{t0}^{14}C = 0.6‰$) which should be compared to the results of tree-ring measurements of $-21‰$ (Lerman, Mook, and Vogel, 1970) and ca $-23‰$ (De Jong and Mook, 1982).

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