^{14}C VARIATIONS CAUSED BY CHANGES IN THE GLOBAL CARBON CYCLE

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ABSTRACT. A box-diffusion model for the carbon cycle is used to estimate the magnitude of ^{14}C variations caused by changes of reservoir sizes and exchange fluxes in the global carbon system. The influence of changes in atmospheric CO_2 concentration, biomass, CO_2 exchange rate between atmosphere and ocean, and ocean mixing is considered. Steady-state ^{14}C concentrations as well as the transients are calculated. For changing biomass, atmospheric CO_2 levels and ^{13}C/^{12}C ratios are also calculated.

Carbon-cycle-induced ^{14}C variations may have been significant in the transition period from Glacial to Postglacial when drastic changes in environmental conditions took place within short time periods, while they were probably less important during the climatically more stable Postglacial.

Changes of the oceanic circulation, as supposedly occurred, are considered the most important factor, besides variations of the production rate, affecting the global distribution of ^{14}C. ^{14}C variations due to changes of the atmospheric CO_2 level or the air-sea-exchange probably did not exceed one to a few percent. Fluctuations of the forest biomass, which may have occurred between Glacial and Postglacial, hardly affected the ^{14}C concentration over a long term.

Responses of the atmospheric ^{14}C concentration are also calculated for variations of the ^{14}C production rate by cosmic radiation. The following cases are considered: a step change, square-wave changes producing "wiggles", and sinusoidal variations.

INTRODUCTION

The possible causes of variations of the atmospheric ^{14}C concentration have been discussed by many authors (see, eg, several contributions in Olsson, 1970). There are two distinct groups of such causes: variations of the rate of ^{14}C production in the atmosphere, and reservoir changes within the global carbon cycle system. In this paper, we mainly consider the second group, which has been given less attention than variations of the production rate. In particular, we will discuss the influence on the atmospheric ^{14}C/C ratio (and briefly on CO_2 concentration and ^{13}C/^{12}C ratio) of the following processes: CO_2 exchange flux between atmosphere and ocean, pattern and rate of the ocean circulation, build-up of biosphere from Glacial to Postglacial. A few results for production variations are also presented.

Significant variations of reservoir parameters can be expected during periods of major climatic change. We shall, therefore, mainly discuss changes associated with the transition from Glacial to Postglacial. A quantitative assessment of past changes in the carbon system is hardly possible at present; the aim of our discussion rather is to establish reasonable ranges for the different parameter.

A change of a reservoir parameter can be regarded as a perturbation of the coupled CO_2–^{13}C–^{14}C system. Correspondingly, we distinguish in each case between concentration changes from the initial to the final (steady) state and the time dependent excursion during the transition period.

Quantitative results were obtained by means of the box-diffusion model described by Oeschger and others (1975) and Siegenthaler and

Oeschger (1978). Figure 1 shows the exchanging reservoirs considered and the values of some of the key parameters that were used in the calculations. Sedimentation is neglected, as well as other minor fluxes, because it removes only about 0.5×10^{-5} parts of the total oceanic amount of carbon (or ^{14}C) per year. The set of time-dependent equations for the transients was solved numerically. At every time step first the new CO_2 levels and $^{13}C/^{12}C$ ratios were determined. Then the true (uncorrected) $^{14}C/C$ ratios were obtained and from these through the standard correction formula the ^{13}C-corrected $\Delta^{14}C$ values. The fractionation factors used for air-sea exchange are those given by Siegenthaler and Münnich (in press).

Steady-state distribution of ^{14}C

In this section, we discuss the factors that determine the steady-state ^{14}C distribution. For the sake of clarity we will make a few minor approximations in the calculations, so that some numerical results are slightly different from those obtained by means of the computer version of the model, used for calculating time-dependent concentrations, which does not include significant approximations.

In a steady-state, the rates of ^{14}C production, Q, and decay must be equal:

$$Q/\lambda = R_a\,N_a + R_b\,N_b + R_{oc}\,N_{oc} \qquad (1)$$

where N_i = mass of carbon in reservoir i,
      R_i = mean $^{14}C/C$ ratio in reservoir i, corrected for isotopic fractionation to $\delta^{13}C$ (PDB) $= -25\%_{\circ}$,
      λ = decay constant of ^{14}C, 1/8270 yr,
      a, b, oc refer to atmosphere, terrestrial biosphere, and ocean.

Since most of the exchanging carbon (~95 percent) is found in the ocean, the global mean $^{14}C/C$ ratio is approximately equal to the average ratio for the ocean. Using the numbers given in figure 1, the value for the ocean is 84.0 percent and the global mean 84.8 percent. Shifts in the carbon system that change the distribution of ^{14}C will, therefore, have little influence on the average $^{14}C/C$ ratio of the whole ocean, as long as the total amount of exchanging carbon remains constant. An interesting asymmetry for larger changes of the system is illustrated by considering extreme assumptions. The minimum possible atmospheric ^{14}C concentration is 84.8 percent; it would result from infinitely fast mixing of the whole system. The theoretical maximum value, obtained by assuming that the atmosphere does not exchange at all with other reservoirs, so that all ^{14}C would decay in the atmosphere, is 5870 percent.

The concentration difference between atmosphere (100 percent) and ocean (84 percent), corresponding to an apparent ^{14}C age of 1440 years for oceanic carbon, is the result of a balance of ^{14}C exchange between atmosphere and ocean, ocean mixing, and radioactive decay. The CO_2 exchange flux between air and sea essentially determines the ^{14}C difference between atmosphere and mixed layer (= surface-ocean). This follows from

a stationary ^{14}C balance for the ocean, stating that ^{14}C decay must be balanced by the net influx from atmosphere to mixed layer (index m):

$$\lambda \, R_{oc} \, N_{oc} = k_{am} \, N_a \, R_a - k_{ma} \, N_m \, R_m = k_{am} \, N_a \, (R_a - R_m) \qquad (2)$$

$k_{am} \, N_a$ is the CO_2 flux from atmosphere to mixed layer; at steady-state it equals $k_{ma} \, N_m$, the flux from mixed layer to atmosphere. The exchange coefficient k_{am} is the reciprocal of the mean atmospheric residence time for CO_2 with respect to exchange with the ocean. With the numbers given in figure 1 we obtain $k_{am} = 1/7.5$ years.

In the box-diffusion model, ocean mixing and, therefore, the ^{14}C distribution within the ocean is determined by the eddy diffusivity K. With the approximation $R_{oc} = R_d$ (R_d = mean concentration in deep-sea below mixed layer) and the mean ocean depth $h_{oc} = 3800m$, we have (eq (19) of Oeschger and others, 1975):

$$R_{oc} = R_m \frac{\tanh (\sqrt{\lambda/K} \, h_{oc})}{\sqrt{\lambda/K} \, h_{oc}} \qquad (3)$$

Eq (3) determines the eddy diffusivity K for given R_{oc} and R_m.

We express the atmospheric $^{14}C/C$ ratio, R_a, as a function only of the production rate and the carbon reservoir parameters. R_{oc} can be eliminated from eq (1) by means of (2) and (3). Furthermore, we assume that the mean biospheric ^{14}C concentration is approximately equal to the atmospheric value

$$R_b \cong R_a \qquad (4)$$

We thus obtain

$$R_a = \frac{Q/\lambda}{N_a + N_b + N_{oc} \dfrac{1}{m + \dfrac{\lambda N_{oc}}{k_{am} \, N_a}}} \qquad (5)$$

with

$$m = \frac{\sqrt{\lambda/K} \, h_{oc}}{\tanh (\sqrt{\lambda/K} \, h_{oc})}$$

The rightmost term in the denominator of eq (5) corresponds to the amount of ^{14}C in the ocean divided by the atmospheric $^{14}C/C$ ratio.

In addition to (5) we assume conservation of carbon

$$N_{tot} = N_a + N_b + N_{oc} = \text{constant} \qquad (6)$$

An increase of N_a or N_b is therefore compensated by a corresponding decrease of N_{oc}. The influence of small changes in the parameters on R_a can be indicated by replacing N_{oc} by $(N_{tot} - N_a - N_b)$ in (5) and forming the total derivative:

$$\frac{dR_a}{R_a} = -0.11 \, \frac{dK}{K} - 0.047 \, \frac{d \, k_{am}}{k_{am}} - 0.050 \, \frac{d \, N_a}{N_a}$$
$$- 0.009 \, \frac{d \, N_b}{N_b} + 0.22 \, \frac{d \, h_{oc}}{h_{oc}} + \frac{dQ}{Q} \qquad (7)$$

Except for changes of the production rate Q, all the sensitivities are small which shows that changes in reservoir sizes and exchange parameters probably produced moderate ^{14}C variations only.

Using eqs (5) to (7) we will now discuss the influence of individual factors on the stationary ^{14}C concentration. It is important to note that changes in different quantities are often coupled. If, for instance, the total biomass N_b decreases, not all the CO_2 produced is absorbed by the ocean, but part of it remains in the atmosphere. Time-dependent variations, as calculated by means of the computer version of the model, will also be presented.

CO$_2$ exchange between atmosphere and ocean

It is not easy to see directly by which physical mechanisms the exchange coefficient k_{am} (respectively, the CO_2 residence time in the atmosphere, $1/k_{am}$) is determined. Therefore, the CO_2 exchange flux between atmosphere and ocean is better written as (see, eg, Broecker and Peng, 1974, for a general discussion of gas exchange):

$$k_{am} N_a = w_{am} A_{oc} C_a = w_{am} A_{oc} N_a/V_a;$$

$$k_{am} = \frac{w_{am} A_{oc}}{V_a} \tag{8}$$

where w_{am} = mean transfer velocity (piston velocity) for gas exchange
　　　　A_{oc} = free ocean surface
　　　　C_a = atmospheric CO_2 concentration
　　　　V_a = volume of the atmosphere normalized to temperature and pressure at sea level
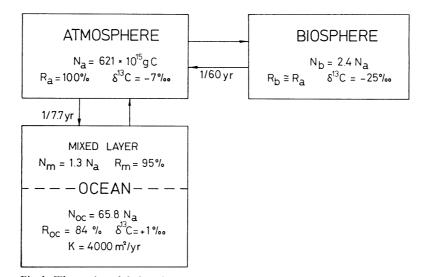

Fig 1. The major global carbon reservoirs and fluxes. N_i = amount of carbon in reservoir i, R_i = $^{14}C/C$ ratio.

The transfer velocity w_{am} is independent of the CO_2 concentration, so that we see from eq (8) that the exchange coefficient k_{am}, and therefore, the atmospheric residence time of CO_2, remains constant if the atmospheric CO_2 level varies. The exchange flux, $N_a k_{am}$, is directly proportional to the CO_2 concentration.

The exchange flux enters as a whole into the oceanic ^{14}C balance (2), so that the three factors w_{am}, A_{oc}, and N_a affect the ^{14}C distribution in the same way and we need not calculate the effects of changes in these quantities separately. (This is not strictly true for N_a, since a change in N_a involves also a shift in the reservoir sizes, not only in the exchange flux). Qualitatively, a reduction of the exchange flux corresponds to a higher resistance for ^{14}C transfer, with the result of a larger ^{14}C difference, $(R_a - R_m)$, between atmosphere and mixed layer. As stated above, the amount of ^{14}C in the ocean, and therefore the left-hand side of eq (2), is insensitive to changes in the carbon system. Therefore, the product $(k_{am} N_a) (R_a - R_m)$ is approximately constant also. If the CO_2 exchange flux *increases* by a factor a, the ^{14}C difference $(R_a - R_m)$ across the air-sea interface *decreases* by approximately the same factor a; and vice versa. The ^{14}C value for the mixed layer is essentially not affected, since the inner-oceanic differences are determined only by oceanic mixing. Evaluation of eq (5) shows that, for an increase or a decrease of the exchange flux $k_{am} N_a$ by a factor of 2, the steady-state ^{14}C concentration R_a changes by -2.3 percent and $+4.7$ percent, respectively (fig 2a). The responses are obviously asymmetrical. The various factors that influence the CO_2 exchange flux will now be discussed.

Ocean surface area

At the maximum of the last glaciation, 18,000 BP, the sea level was about 120m lower and the volume of the ocean three percent smaller than now (Shackleton and Opdyke, 1973). Furthermore, the area of sea ice was larger. A rough estimate yields that the free ocean surface, A_{oc}, was reduced by about 5 percent, compared to now, due to the sea level drop, and by perhaps 7 percent due to the enhanced ice cover (CLIMAP, 1976), in total by about 12 percent. This corresponds to a 12 percent decrease of k_{am} which leads, according to eq (7), to an increase of the atmospheric ^{14}C concentration by 0.6 percent, a rather small change if we consider that the rise of the sea level from Glacial to Postglacial extended over several thousand years.

Transfer velocity for gas exchange

Gas transfer is controlled by diffusion through a thin surface layer in the liquid, the thickness of which depends on near-surface turbulence and, therefore, on the wind stress (eg, Liss and Slater, 1974). The transfer velocity increases with increasing wind velocity probably more than proportionally (Kanwisher, 1963; Jähne, Münnich, and Siegenthaler, 1979). Changes in the global wind regime can, therefore, influence the CO_2 exchange and, consequently, the ^{14}C concentration difference between atmosphere and surface ocean. During glacial times the meridional temperature gradient was steeper than now which may have caused stronger

winds. Therefore, the mean transfer velocity may have been larger during the Glacial than during the Postglacial, involving a reduction of the ^{14}C difference across the air-sea interface. This effect would be opposite to that of a reduced sea surface area.

A quantitative estimate is difficult. A 20 percent increase of the world-wide mean wind velocity $\bar{u}$ over the ocean during the Glacial appears rather large. For proportionality to $\bar{u}^2$, the corresponding increase of the transfer velocity would be about 50 percent, which would imply a ^{14}C decrease in the atmosphere of about 2 percent.

Atmospheric CO₂ concentration

The partial pressure of CO_2 in the atmosphere depends on chemical equilibria between dissolved CO_2, $HCO_3{}^-$, and $CO_3{}^{--}$ in surface sea
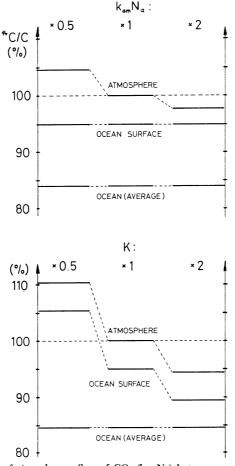
Fig 2. Influence of a) exchange flux of CO_2 ($k_{am}N_a$) between atmosphere and ocean, b) ocean mixing rate, represented by eddy diffusivity K. Oceanic ^{14}C values are nearly unaffected by the CO_2 exchange rate; on the contrary, the ^{14}C difference atmosphere-mixed layer is essentially uninfluenced by the oceanic mixing rate.

water. At 18,000 BP, the average surface temperature of the ocean was about 2.3°C lower (CLIMAP, 1976) and the ocean volume three percent smaller than now. For a temperature decrease of 1°C (from 20°C to 19°C), the atmospheric CO_2 concentration decreases by 2.8 percent at steady-state (Heimann, 1978). Due to the temperature effect alone, the atmospheric CO_2 pressure, therefore, would have been 6.4 percent lower than today. An analysis of the carbonate equilibria shows that, if the volume reduction of three percent was accompanied by a proportional concentration increase of total CO_2 and alkalinity, then the CO_2 pressure increased by only one percent. Thus, the combined effect of temperature and volume change is about a 5 percent lower concentration at 18,000 BP, which implies, according to eq (7), a shift of 0.3 percent in the atmospheric ^{14}C concentration. Therefore, the pure physico-chemical effects on CO_2 and ^{14}C are not significant.

Biological processes may have been more important. Because of marine biological activity, total CO_2, alkalinity, and CO_2 pressure are lower in surface water than in the deep sea. The average CO_2 pressure in the deep sea is about 450ppm, referred to p = 1 atm (Li, Takahashi, and Broecker, 1969); if the oceans were well-mixed and the vertical differences due to the biology were absent, the equilibrium CO_2 concentration would be of the order of 700ppm instead of 290ppm, taking into account that the temperature at the surface is higher than in deep waters. Biological activity in the sea obviously exerts a strong control over the atmospheric CO_2 level; thus, changes in the productivity or composition of the marine biosphere could lead to atmospheric CO_2 variations.

Sediment studies have yielded evidence that during the Quaternary the oceanic carbonate chemistry fluctuated in general parallel to the glacial-interglacial cycles (eg, Broecker, 1971; Berger, 1977). It seems probable that these fluctuations also had their effect on the atmospheric CO_2 concentration, but the processes that occurred are complex and not yet understood in enough detail to permit more than speculative conclusions on past atmospheric CO_2 concentrations.

The results of CO_2 measurements in old polar ice suggest that the atmospheric CO_2 concentration was lower during the Glacial than in the Postglacial (Berner, Oeschger, and Stauffer, 1980). If it was only two-thirds or half of the recent (pre-industrial) level, the steady-state $^{14}C/C$ ratio in the atmosphere at that time was 102.5 percent or 104.9 percent, respectively, of its standard value, due to the different CO_2 level alone.

A special case of CO_2 variation is implied by the build-up of a large forest biomass. This is considered next.

Biomass

Based on $\delta^{13}C$ data in deep sea sediments as well as observations on the continents, Shackleton (1977) suggested that during the last glaciation, the forest biomass was lower, by as much as a factor of 2 to 3, than now. If this is so, a corresponding amount of carbon dioxide was added to the ocean and the atmosphere. The partitioning of this additional carbon dioxide between the two reservoirs is determined by the buffer factor ξ:

the ocean's capacity for excess CO_2 is proportional to N_{oc}/ξ, so that in steady-state, the excess CO_2 is partitioned between atmosphere and ocean according to the ratio $\xi N_a/N_{oc}$ (*cf* Siegenthaler and Oeschger, 1978). As long as interaction with the sediments is not important, ξ has a value of ~ 10, and the partitioning is 1:6.58. (Alkalinity of ocean water may have changed simultaneously, as well as other factors, but only the effects of varying land biomass are considered here.) Reduction of the biomass to 50 percent of its present value would then lead to an increase of the atmospheric CO_2 level by 16 percent. According to eq (7), this yields a ^{14}C decrease of 0.4 percent in the *steady state*. The change is small because the biosphere has approximately the same $^{14}C/C$ ratio as the atmosphere.

The *transient* ^{14}C variation is, however, larger. Figure 3 shows the results of model calculations for the (arbitrary) assumption that the biosphere grew linearly within 500 years from 50 to 100 percent of its present-day size. From 0 to 500 years, the atmospheric CO_2 level would decrease from 338 to 234ppm, $\delta^{13}C$ would increase from -7.4 to $-6.0\%_o$, $\Delta^{14}C$ from -0.4 to $+1.0\%$. After the biospheric growth stops, the trends reverse and the concentrations approach their new (Holocene) equilibrium values of
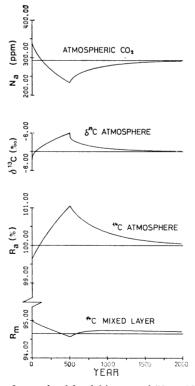

Fig 3. Consequences of growth of land biomass of 50 to 100 percent of its present value within 500 years (constant growth rate). From top to bottom: atmospheric CO_2 concentration (ppm); $\delta^{13}C$ and $^{14}C/C$ ratio in atmospheric CO_2; $^{14}C/C$ ratio in the mixed ocean-surface layer.

293ppm, $-7.0\%_o$ ($\delta^{13}C$) and 0% ($\Delta^{14}C$). The reason for the relatively large transient ^{14}C increase is that as long as the atmospheric CO_2 concentration, and therefore the air-sea exchange flux, is considerably reduced, a larger ^{14}C difference between atmosphere and mixed layer is established, because the rate at which the ^{14}C produced in the atmosphere escapes into the ocean is slowed down.

If the same total biospheric growth would occur within 100 years, the transient changes would be larger: the atmospheric CO_2, $\delta^{13}C$, and $\Delta^{14}C$ at the end of the growth period would be 173ppm, $-3.7\%_o$ and $+2.0\%$, respectively. Shackleton (1977) observed that $\delta^{13}C$ in benthonic and planktonic foraminifera was about $0.5\%_o$ lower in glacial than in interglacial sediments. If interpreted as being due to a change in land biomass, this difference would be expected if the biomass during the glacial was about 50 percent of its present-day value. Shackleton's $\delta^{13}C$ data do not show if there were transient "overshoot" signals, which would indicate rapid biomass changes.

Ocean circulation

Deep waters of the world ocean are formed at the cold ocean surface in high latitudes. North Atlantic Deep Water (NADW) has its origin mainly in the Norwegian Sea; the formation of NADW provides a relatively rapid ventilation of the deep Atlantic Ocean (Broecker, 1979). There is some evidence that at the climax of the last glaciation, the formation of NADW may have been eliminated because the near-surface ocean in the present source of NADW region was stably stratified and the permanent sea-ice cover extended to lower latitudes; the production of Antarctic Bottom Water was possibly also reduced (Schnitker, 1974; CLIMAP, 1976; Streeter and Shackleton, 1979). Deep circulation of the North Atlantic during the Glacial may have been similar to the circulation pattern of the North Pacific.

The present rate of formation of NADW has been estimated between 10×10^6 m³/s (Gordon and Taylor, 1975) and 30×10^6 m³/s (Broecker, 1979). The second region where deep water is formed is the Antarctic Ocean. The estimates for the production of Antarctic Bottom Water vary; about 20×10^6 m³/s appears as a reasonable value (Broecker, 1979). If indeed no NADW was formed during the climax of the last glacial, the global rate of formation of deep water would have been reduced to about half its present-day value.

It is not possible to simulate such changes in a realistic way by means of a simple one-dimensional model. Our model calculations give a qualitative idea of the magnitude and time scales of the corresponding ^{14}C variations. In the box-diffusion model, the vertical mixing rate of the ocean is given by the eddy diffusivity K. Although eddy diffusion does not offer a realistic description of deep water formation, K can be interpreted directly in terms of water age. The mean diffusion time down to depth z is proportional to z^2/K, and it can be shown rigorously that the average age at a given depth (with respect to surface) is proportional to $1/K$. According to eq (5), a decrease or increase of K by a factor 2 leads to an

atmospheric ¹⁴C change by +10.4 or −5.5 percent, respectively. Again, the responses are asymmetric (fig 2b). Therefore, if in 18,000 BP the oceanic circulation was slowed down to half its present rate, the atmospheric ¹⁴C level would have been at the order of ten percent higher than now, providing the ¹⁴C production rate was the same as at present.

Figure 4 shows atmospheric ¹⁴C concentration during a sudden increase of K from 2000m²/yr (ice age) to 4000m²/yr (postglacial). A rapid ¹⁴C decrease would occur, 3.6 percent in the first 100 years. Interesting is the relaxation time, $\tau = (R_a - R_{a\infty})/(dR_a/dt)$. τ is not constant (*ie*, R_a does not decrease to its final value exponentially), but increases from $\tau = 160$ years at t = 0 to $\tau = 420$ years at t = 1000 years.

An interesting question with respect to ¹⁴C is, how fast could changes of the ocean circulation have taken place? Ruddiman and McIntyre (1973) showed through sediment studies that the polar waters in the North Atlantic retreated northward from 17,000 BP to 6000 BP. The retreat was not uniform, but was concentrated in several periods of about 700 to 2000 years' duration, and a marked readvance occurred around 10,000 BP. Such retreats and readvances must have involved fluctuations of the rate of formation of NADW.

In summary, climatic changes may well have been accompanied by changes of the oceanic circulation producing ¹⁴C variations in the atmosphere. It also is possible that in some periods of drastic climatic events, such as during the Late Postglacial, the rate of these ¹⁴C variations was as much as one percent per century.

Responses to changing production rate

At steady state, the relative change of ¹⁴C concentration in any reservoir is obviously equal to the relative change of the production rate. Therefore, only the transient responses need discussion.
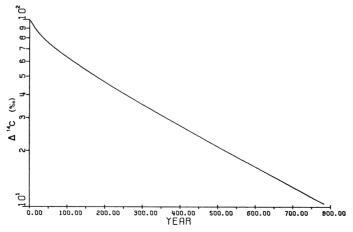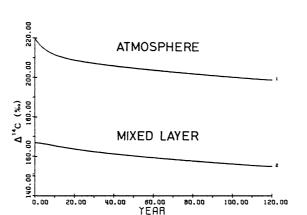
STEP CHANGE OF K

Fig 4. Δ¹⁴C in atmospheric CO_2 after a step change at time 0 from K = 2000m²/yr ("glacial") to K = 4000m²/yr ("postglacial").

According to Barbetti and Flude (1979), the geomagnetic dipole moment during the Late Pleistocene may have amounted to only about two-thirds of its present value. According to Elsasser, Ney, and Winckler (1956), the ^{14}C production varies with the dipole moment approximately to the power of -0.52, so that in the Late Pleistocene, ^{14}C production rate may have been higher by 22 percent than now. Figure 5 shows the calculated atmospheric ^{14}C concentration for a step change of the production rate from 122 to 100 percent of its present value. As indicated by Suess (1968), the mean global ^{14}C concentration would exponentially decrease to its new steady-state value with the ^{14}C half-life, as well as the ^{14}C concentration in the individual reservoirs after some time (fig 5B). First, however, the differences between the various reservoirs would have to adjust to their new values. The time scales for this adjustment are the relaxation times of the carbon exchange system. For instance, the atmospheric ^{14}C level would decrease by one percent within only about 15 years, reflecting
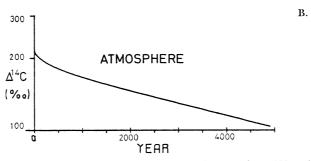

Fig 5. Response to a step change of the ^{14}C production rate from 122 to 100 percent of its present value. A) Initially, the differences within the system adjust to values corresponding to a new dynamic equilibrium; B) asymptotically, the ^{14}C concentration decreases to zero exponentially with the ^{14}C half-life (note logarithmic scale for $\Delta^{14}C$). A well-mixed atmospheric reservoir is assumed for this calculation.

the time necessary for equilibration between atmosphere and mixed layer (fig 5A).

¹⁴C measurements on tree rings from the Postglacial have demonstrated ¹⁴C variations (Suess wiggles) with a characteristic time scale of about 200 years and an amplitude of 1 to 2 percent. In view of the above, it seems unlikely that these fluctuations could have been caused by changes in the carbon cycle; rather, they appear to be due to solar modulation of atmospheric ¹⁴C production (Stuiver and Quay, 1980). Figure 6 shows wiggles produced by a square-wave-like variation of the production rate with a period of 120 years and an amplitude -10 to $+20$ percent; these numbers were chosen according to tree-ring data measured by de Jong, Mook, and Becker (1979). The calculated amplitude of atmospheric ¹⁴C variation is 1.7 percent (peak to peak), of which more than half is reached within only ten years after a jump in production rate. The shape of our synthetic wiggles is different from the curve obtained by de Jong, Mook,
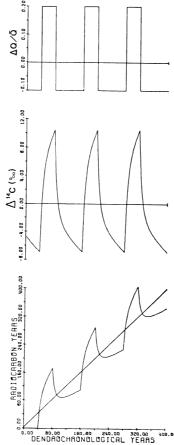

Fig 6. Atmospheric ¹⁴C concentration (center and bottom) for an assumed periodic, square-wave variation of the ¹⁴C production rate Q (top). The atmosphere is assumed well-mixed for this calculation.

and Becker (1979) which exhibits alternating intervals of nearly linearly increasing or decreasing $\Delta^{14}C$. Thus, in reality, ^{14}C production varied less abruptly than assumed by a square-wave function.

Finally, figure 7 shows the response of tropospheric ^{14}C concentration to sinusoidal variations of the production rate, in terms of the attenuation (defined as the ratio between the relative amplitudes of the tropospheric concentration and the production rate) and the phase lag between production and tropospheric response. Similar diagrams based on different models were presented by other authors (Houtermans, 1966; Ekdahl and Keeling, 1973). Three model versions based on different assumptions about the atmosphere and the biosphere are considered:

1) atmosphere is well-mixed, model includes a well-mixed biosphere with a carbon residence time of 60 years;
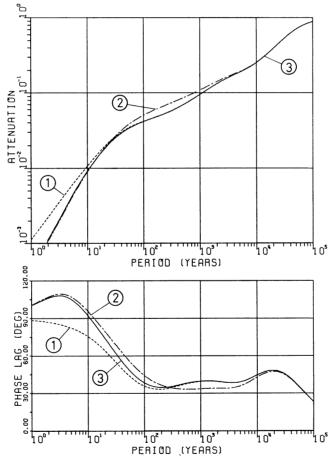
Fig 7. Attenuation and phase-shift of tropospheric ^{14}C concentration for a sinusoidal variation of the ^{14}C production rate. 1) Well-mixed atmosphere, model includes biosphere; 2) troposphere separated from stratosphere, no biosphere; 3) troposphere and stratosphere separated, model includes biosphere.

2) no biosphere, but atmosphere is subdivided into stratosphere and troposphere;

3) model includes biosphere, stratosphere, and troposphere.

The model stratosphere contains 20 percent of the total atmospheric CO_2 and has a residence time of 1.5 years with respect to exchange with the troposphere (Reiter, 1975). Two-thirds of the total ¹⁴C is produced in the stratosphere, one-third in the troposphere.

The subdivision of the atmosphere noticeably affects the results for periods <30 years. The effect of the biosphere, a stronger attentuation of the atmospheric signal, is noticed mainly for periods between about 30 and 2000 years. The response characteristics of the box-diffusion model to sinusoidal production variations are, in general, not very different from those of the multi-box models as discussed by Ekdahl and Keeling (1973).

CONCLUSION

Exchange processes of the global carbon system determine, for a given ¹⁴C production rate, the atmospheric ¹⁴C/C ratio. These processes probably changed during periods of climatic change, mainly in the transition time from the Glacial to the Postglacial. Estimates of the produced ¹⁴C variations are subject to relatively large uncertainties because it is difficult to quantitatively assess the changes within the carbon system.

The most important factor is oceanic circulation. There is evidence that during the Ice Age, the deep waters of the world ocean were renewed at a considerably slower rate than now. For a constant ¹⁴C production rate, consequently, the atmospheric ¹⁴C concentration may have been higher during the last glaciation by perhaps ten percent. During phases of drastic climatic variation, such as during the Late Glacial, rates of change of the order of one percent per century may have been caused by the rapid establishment of a new ocean circulation pattern.

During climatically more stable periods, ¹⁴C variations due to non-stationary carbon cycling were presumably smaller and slower. It appears improbable that the observed ¹⁴C wiggles could have been caused by fluctuating exchange rates.

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