PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

The past 5000 Years History of Solar Modulation of Cosmic Radiation from ^{10}Be and ^{14}C Studies

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The past 5000 years history of solar modulation of cosmic radiation from ^{10}Be and ^{14}C studies

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^{10}Be is produced in a similar way as ^{14}C by the interaction of cosmic radiation with the nuclei in the atmosphere. Assuming that the ^{10}Be and ^{14}C variation are proportional and considering the different behaviour in the Earth system, the ^{10}Be concentrations in ice cores can be compared with the ^{14}C variations in tree rings. A high correlation is found for the short-term variations (^{14}C-Suess-wiggles). They reflect with a high probability production rate variations. More problematic is the interpretation of the long-term trends of ^{14}C and ^{10}Be. Several explanations are discussed.

The reconstructed CO_2 concentrations in ice cores indicate a rather constant value (280 ± 10 p.p.m. by volume) during the past few millenia. Measurements on the ice core from Byrd Station, Antarctica, during the period 9000 to 6000 years BP indicate a decrease that might be explained by the extraction of CO_2 from the atmosphere–ocean system to build the terrestrial biomass pool during the climatic optimum.

Introduction

After W. F. Libby introduced the radiocarbon dating method around 1950, by studying ^{14}C:C ratios in absolutely dated samples like tree-rings several authors investigated the constancy of the atmospheric ^{14}C:C ratio.

H. E. Suess pointed both at the existence of a long-term trend and short-term wiggles in the ^{14}C-bristlecone tree-ring record (Suess 1971, this Symposium). Whereas the existence of the long-term trend was confirmed in a number of laboratories, it was roughly a decade before general agreement regarding the existence of the short-term wiggles was reached. Today a very precisely measured record of the ^{14}C trend over the past 9000 years exists (Stuiver & Kra 1986). With the development of accelerator mass spectrometry about a decade ago, it became possible to measure additional long-lived radioisotopes (^{10}Be, ^{26}Al, ^{36}Cl) in natural archives.

In this paper we compare the ^{14}C record measured in tree-rings with ^{10}Be records measured on ice cores and deduce the present knowledge of short-term and long-term variations of cosmic radiation.

The ^{14}C-variations

Figure 1 shows the present continuous record on ^{14}C:C variations, expressed in per mille deviations from a standard, determined on dendrochronologically dated tree-rings (Stuiver & Kra 1986).

Additional information on the ^{14}C:C behaviour during the glacial–post-glacial transition exists from peat bog studies (Oeschger *et al.* 1980) and plant macrofossils deposited in lake sediments (Zbinden *et al.* 1990), which indicated variations of the order of 50–100‰ during

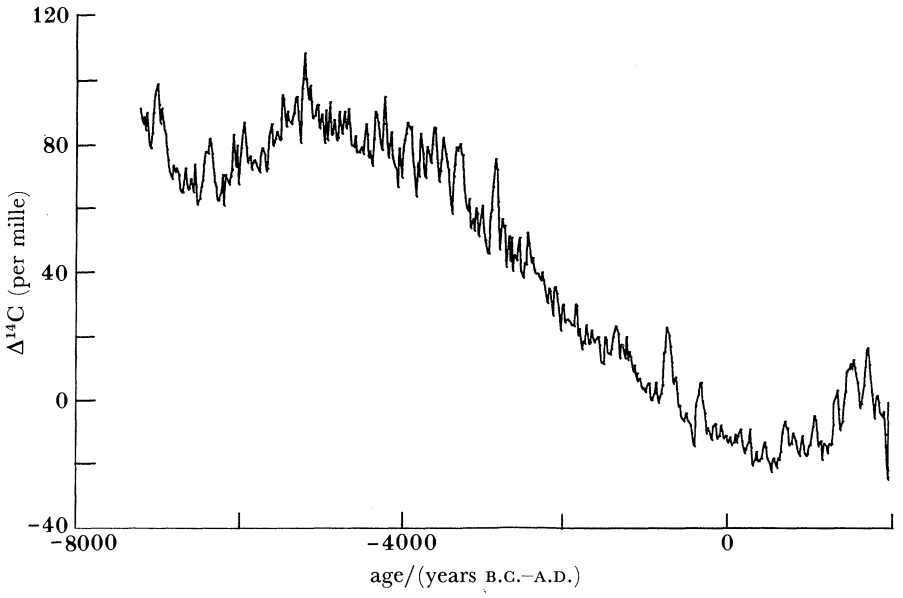

FIGURE 1. Atmospheric $^{14}C:C$ variations expressed in per mille deviation $\Delta^{14}C$ from a standard (Stuiver & Kra 1986).

this period of major climate change. It is likely that these ^{14}C variations from 13000 to 10000 years before present (BP) mainly reflect the effects of carbon system changes as a result of repartitioning of carbon between atmosphere, ocean, biosphere, sediments, etc., as well as possible changes in the system dynamics, like the mixing of ocean surface and deep water. The variations during the Holocene probably reflect mainly changes in the ^{14}C production rate. In the following we draw some independent conclusions on the stability of the carbon system during the past *ca.* 20000 years to investigate the potential influence of carbon system variations on the atmospheric $^{14}C:C$ ratio.

HOW STABLE IS THE ATMOSPHERIC CO_2 CONCENTRATION AND THE CARBON SYSTEM DYNAMICS IN GENERAL?

Information on the stability of the carbon system during the past 30 years can be drawn from the direct CO_2 concentration observations at Mauna Loa, Hawaii, and at the South Pole, which started in 1985. In spite of the very large yearly exchange fluxes between atmosphere and ocean and atmosphere and biomass, which together correspond to 25% of the atmospheric CO_2 per year, the atmospheric CO_2 concentration reacted relatively monotonously to the anthropogenic input (figure 2). The annual increase of CO_2 was *ca.* 1 ± 1 p.p.m. CO_2 per year, varying not more than *ca.* 3% of the atmospheric CO_2 content. Figure 3 shows the CO_2 increase reconstructed from measurements of the air occluded in an ice core from Siple Station, Antarctica, over the past 200 years (Neftel *et al.* 1985; Siegenthaler & Oeschger 1987). Again a very homogeneous increase of atmospheric CO_2 is observed and deconvolution of the increase by using a carbon cycle model resulted in a total CO_2 production from fossil fuel combustion as well as from deforestation and land managing that might be very close to the actual ones (Siegenthaler & Oeschger 1987). Based on this work we conclude that the carbon cycle, in

† p.p.m. is parts per million by volume.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

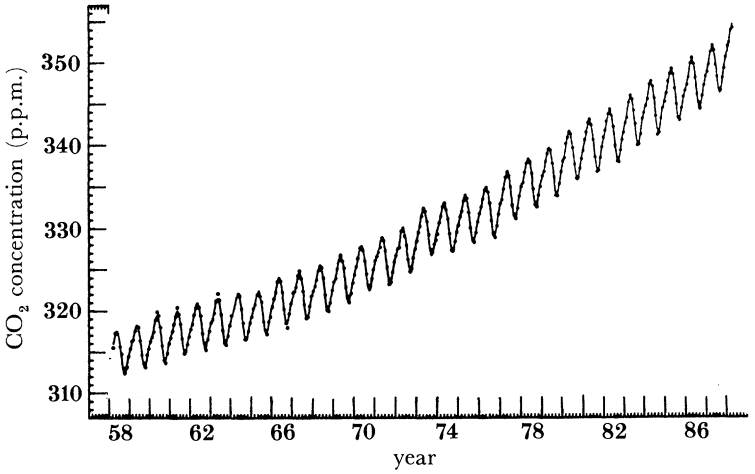

FIGURE 2. Concentrations of atmospheric CO_2 (p.p.m.) at Mauna Loa Observatory, Hawaii, from 1958 to 1988.

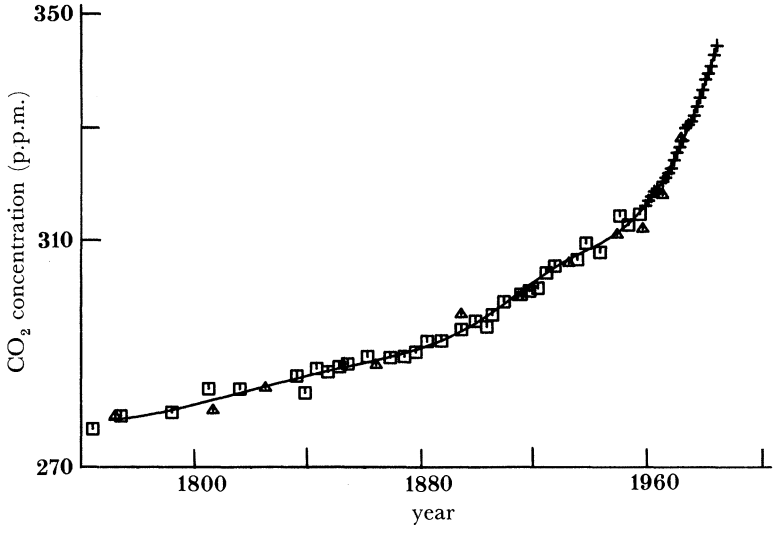

FIGURE 3. Atmospheric CO_2 increase in the past 200 years as indicated by measurements on air trapped in old ice from Siple Station, Antarctica, (Neftel *et al.* 1985) and by the annual mean values from Mauna Loa. The full line is a spline fit through all data (Siegenthaler & Oeschger 1987).

particular the CO_2 exchange atmosphere–ocean, during the past 200 years functioned essentially as it does at present.

Going further back in time there are indications of minor variations in atmospheric CO_2, e.g. an increase by *ca.* 5 p.p.m. between 1200 and 1300 A.D.

Estimates using carbon cycle models indicate that such small variations probably had little influence on the atmospheric ^{14}C:C ratio. A somewhat different situation might have existed at the beginning of the Holocene when the CO_2 system was adjusting to the new conditions. CO_2 measurements on the Byrd Station core, Antarctica (figure 4) showed a decrease from *ca.* 280 p.p.m. to 255 p.p.m. during a period of *ca.* 4000 years. An obvious explanation for this change is the regrowth of soils and plants on the continents in areas that during the glacial had been covered by ice.

The decrease in atmospheric CO_2 was about 10 % and, based on the assumption of an ocean

behaving, regarding its chemistry, like the present one, the 10% CO_2 decrease in the atmosphere and the CO_2 partial pressure in the surface ocean correspond according to a rough estimate to an extraction of CO_2 from the atmosphere–ocean system of *ca.* 400 Gt C.

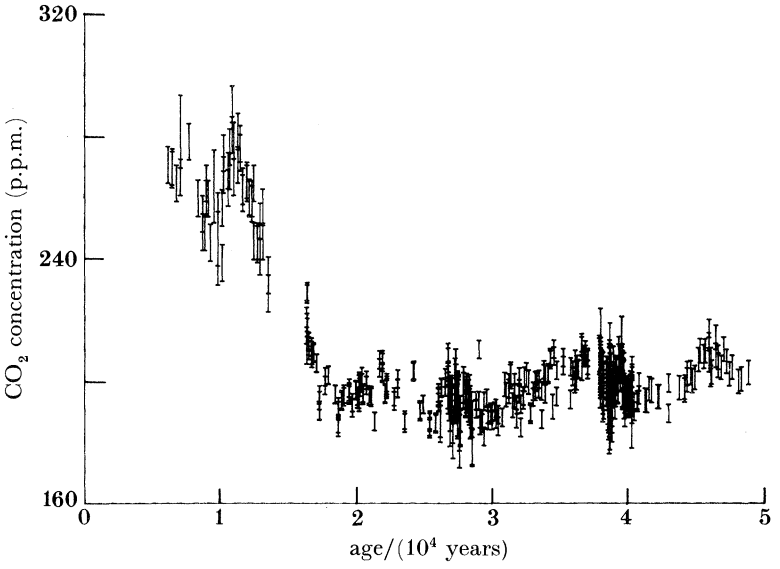

FIGURE 4. CO_2 concentrations of the past 5000–50000 years, measured on air trapped in an ice core drilled at Byrd Station, Antarctica (B. Stauffer & J. Schwander, personal communication).

It is not impossible that this phenomenon had some impact on the long-term ^{14}C:C trend. In fact 7000 years BP the ^{14}C record (figure 1) shows a decline of the ^{14}C:C ratio by *ca.* 30‰. For final conclusions the ^{14}C record should be extended further back in time and model estimates should be made.

Figure 4 shows the CO_2 concentrations during the glacial–post-glacial transition 12000–10000 years BP measured on the ice core from Byrd Station, Antarctica. Evident are the relatively low CO_2 concentrations during glacial period and the parallel increase with the rising $\delta^{18}O$ values that mark the climatic transition. The data are not dense enough to exclude a CO_2 increase in two steps as suggested from studies of ice cores from Greenland (Dansgaard & Oeschger 1989).

There is little doubt that the CO_2 concentration change from 180 to 280 p.p.m. during this period had an influence on the ratio ^{14}C:C in atmospheric CO_2, because on the one side the increase probably was due to CO_2 with a lower ^{14}C:C ratio escaping from the ocean. In addition, with the new partitioning of CO_2 also the CO_2 system dynamics changed. Indicators for the ^{14}C:C variation during this period have been obtained from plant macrofossils in Swiss lakes (Zbinden *et al.* 1990).

THE ^{10}Be VARIATIONS

Both ^{14}C and ^{10}Be are produced by the interaction of cosmic rays and their secondaries with atomic nuclei of air gases. Though the detailed production mechanisms for the two isotopes are different, it is a valid assumption to consider as a first approximation the relative ^{14}C and ^{10}Be production variations to be equal.

Their geochemical pathways are very different: whereas ^{14}C is oxidized to $^{14}CO_2$, ^{10}Be is attached to aerosol particles. $^{14}CO_2$ mixes rapidly with atmospheric CO_2, but is also in exchange with the carbon in the ocean and in the biomass and soils. ^{10}Be is precipitated after a residence time of 1–2 years and can be found in polar ice and in sediments. Thus short-term production variations in principle are recorded in polar ice and can be studied in annual layers.

In the case of ^{14}C, production fluctuations are considerably damped and delayed. Attenuation factors and delay times are given for different models in figure 5 (Siegenthaler *et al.* 1980). Typical examples of attenuation factors for the periods 11 years, 200 years and 2000 years are *ca.* 100, *ca.* 20 and *ca.* 8, respectively.

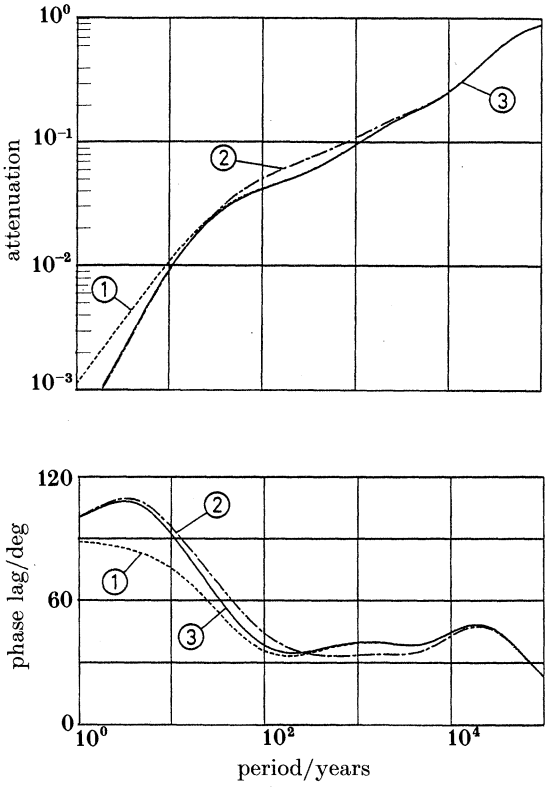

FIGURE 5. Attenuation and phase-shift of tropospheric ^{14}C concentration for a sinusoidal variation of the ^{14}C production rate. (1) Well-mixed atmosphere, model includes biosphere; (2) troposphere separated from stratosphere, no biosphere; (3) troposphere and stratosphere separated, model includes biosphere (Siegenthaler *et al.* 1980).

This large damping of ^{14}C production variations, which renders their detection difficult, is partly compensated by the fact that ^{14}C can be measured relative to a tracer, i.e. ^{14}C:C in CO_2, which is globally well mixed within a few years in the atmosphere. Because of its short atmospheric residence time ^{10}Be is not well mixed in the atmosphere and is therefore influenced by local parameters. Also a valid carrier is missing.

The ^{10}Be concentrations are expressed in atoms per gram of water. Both the atmospheric transport of the aerosols and the dilution with water may vary from year to year.

Because the damping of ^{10}Be-production due to the short residence time is essentially

negligible and the production variations are relatively large (20–30 % for the 11-year cycle) the terrestrial noise does not prevent the observation of the so-called 11-year cycle of the ^{10}Be production due to solar modulation of the cosmic-ray flux.

Comparisons between ^{10}Be and ^{14}C

The first attempts to investigate if ^{10}Be in ice cores shows variations comparable with those of ^{14}C in tree rings were performed on an ice core from station Milcent in central Greenland (Beer *et al.* 1983; Oeschger *et al.* 1987). The ice core covers the past 800 years. Large variations in the ^{10}Be concentrations have been found. For a comparison with the measured ^{14}C variations, a carbon cycle model was used to calculate the ^{14}C variations assuming that the ^{10}Be record directly reflects the ^{14}C production variations. The agreement is satisfactory. The three periods of high ^{14}C:C (corresponding to the Maunder, Spörer and Wolf solar minima) also appear in the calculated ^{14}C:C record. However, whereas the ^{10}Be based ^{14}C maximum agrees during the Wolf in amplitude with the measured one, during the Spörer Minimum it underestimates and during the Maunder Minimum it overestimates it.

^{10}Be in the Camp Century ice core, North Greenland

Thanks to the collaboration with Dr C. C. Langway, SUNY, Buffalo, U.S.A., it was possible to measure the ^{10}Be concentrations in the ice core from Camp Century, North Greenland, which covers the past *ca.* 100000 years. The data on both ^{10}Be and δ^{18}O are given in figure 6.

The most striking feature is that after a flat part with minor fluctuations, both records show a rapid increase at 250 m above bedrock. This height above bedrock corresponds to the transition from Pleistocene to Holocene.

This parallel change in ^{10}Be and δ^{18}O can be explained by the assumption of a roughly constant ^{10}Be flux and a precipitation rate that was smaller during the cold period (δ^{18}O values more negative) than during the warm post-glacial period. During the glacial period, therefore, ^{10}Be was considerably less diluted by water vapour than during the post-glacial one. In fact the variations of the ^{10}Be concentrations have been used to estimate accumulation rate changes during the past 160000 years in the Vostok core (Raisbeck *et al.* 1987).

There are also some interesting differences in the ^{10}Be and δ^{18}O profiles: below 100 m above bedrock ^{10}Be concentrations are low, comparable with the Holocene level, with the exception of one value, whereas the δ^{18}O still indicates glacial conditions. At present we cannot explain this phenomenon.

The small variations in δ^{18}O and ^{10}Be during the Holocene show no correlation, except for the range 1200–1250 m above bedrock, which corresponds to the Maunder Minimum period of sunspots (1645–1715 A.D.) known as a cold period in Europe. Most of the ^{10}Be variations therefore have to be explained by another phenomenon than terrestrial modulation.

For a further analysis the ^{10}Be variations were again converted in apparent ^{14}C variations, based on the assumption that the production variations of ^{10}Be and ^{14}C had been proportional and that the carbon cycle model also holds for the entire post-glacial period. Because both the ^{10}Be and the ^{14}C records are strongly affected by terrestrial phenomena in the glacial and the glacial–post-glacial transition period this comparison was only performed for the period 7000 B.C. till present (figure 7). The choice of this boundary for the assumption of more or less

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

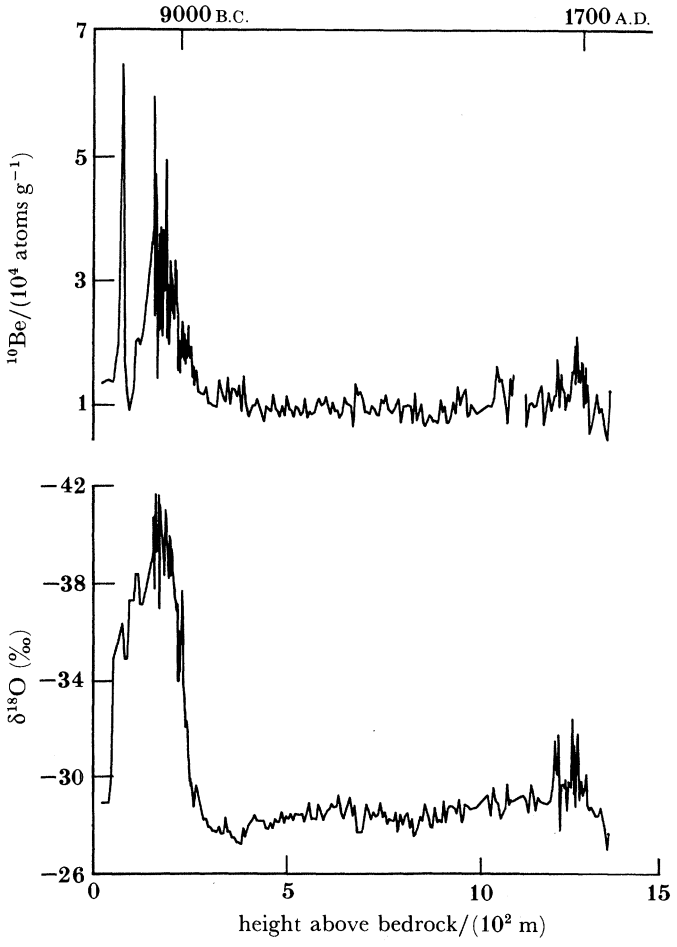

FIGURE 6. ^{10}Be concentrations and δ^{18}O from the Camp Century ice core, Greenland (Beer *et al.* 1988). The strong changes at 250 m correspond to the transition into the last glaciation (*ca.* 9000 B.C.).

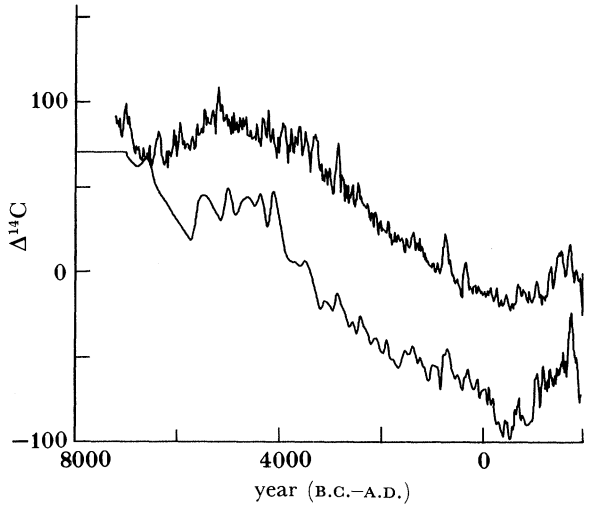

FIGURE 7. Comparison of measured Δ^{14}C (figure 1) with calculated Δ^{14}C based on the ^{10}Be concentrations of figure 6*a*. The calculated curve is shifted by 50 per mille for easier comparison.

stable conditions both for the ^{10}Be and the CO_2 system is somewhat arbitrary and it cannot be excluded that the long-term trends are influenced by after effects following the transition to the Holocene.

Such effects, however, probably would have little influence on the short-term variations that we discuss next. To compare the short-term ^{10}Be and ^{14}C variations from both the model-generated ^{14}C data and the tree ring data of the period 3000 B.C. to 1800 A.D., the long-term trend was removed by applying a binomial high-pass filter. The resulting curves are shown in figure 8 (Beer *et al.* 1988). They are very similar; for example, the prominent maxima of the tree-ring record at *ca.* 2800 B.C., 900 B.C., 300 B.C., 800 A.D., 1100 A.D. and 1700 A.D. (Maunder Minimum) are also found in the ^{10}Be-based model curve. Also the cross-correlation function was calculated. The correlation coefficient has a clear maximum for a phase shift near zero years; its value of 0.58 is highly significant. Although the amplitudes of the ^{10}Be-based ^{14}C variations are slightly higher (4.4% compared with 3.9%) the general agreement is satisfactory.

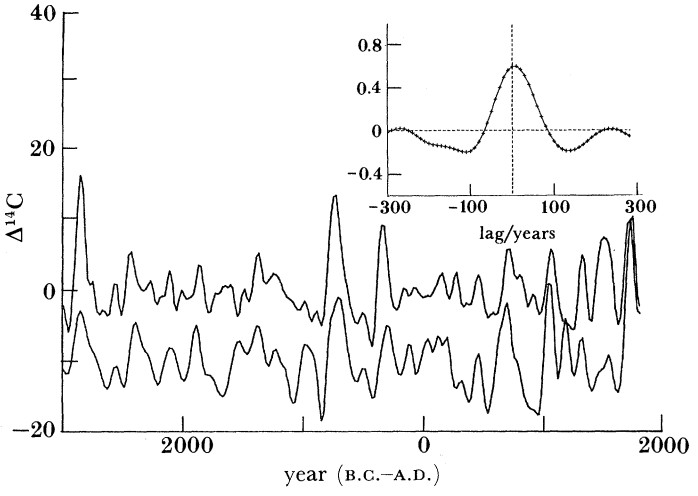

FIGURE 8. (*a*) Expansion of the period 3000 B.C. to A.D. 2000 of figure 7 after removing the long-term trend (Beer *et al.* 1988). The upper curve represents the measured, the lower one the calculated $\Delta^{14}C$ values. The lower curve is shifted by 10 per mille. (*b*) Cross-correlation between the two curves as a function of lag time.

From this comparison we conclude with two points.

1. Both the ^{10}Be and the ^{14}C variations have a common cause, i.e. variations in the production rates.

2. The good agreement between the direct ^{14}C- and the ^{10}Be-deduced short-term variations indicates that for the period investigated the carbon cycle model (Siegenthaler & Oeschger 1987) describes very well the response of the carbon system to an atmospheric disturbance with a characteristic time of *ca.* 200 years.

THE LONG-TERM EFFECTS IN ^{10}Be AND ^{14}C

The long-term effects in the cosmogenic isotope records are of lesser relevance to the topic of this Symposium and we summarize the major results and conclusions briefly. Beer *et al.* (1988) studied how far the ^{10}Be long-term record of the past 9000 years can be explained by

the hitherto determined long-term variation of the geomagnetic dipole moment. They came to the conclusion that the observed ^{10}Be date and those derived from the geomagnetic modulation are not in good agreement. Possible explanations for the poor agreement might be (1) the fact that a high-latitude record reflects to some degree regional production of ^{10}Be and might not therefore be ideal for the study of the geomagnetic effects that are relatively small in the polar atmosphere; (2) the reconstruction of the geomagnetic dipole moment from palaeomagnetic intensity data is hampered by serious uncertainties.

When comparing the measured ^{14}C record with the ^{10}Be-derived record one can draw the following conclusions regarding the long-term behaviour (figure 7).

The ^{14}C decrease from 5000 B.C. to 500 A.D. is not reflected in the ^{10}Be concentration record. It can, however, be simulated by assuming that the production rate of cosmogenic radionuclides during the last *ca.* 10000 years of the Pleistocene was higher by 20% than during the Holocene leading to higher ^{14}C levels in all the carbon reservoirs. The decrease during the Holocene can be understood as an exponential adjustment to the new steady state corresponding to the lower production rate. Such an adjustment would also occur for a constant ^{14}C production if after the last glaciation the CO_2 system had not been in equilibrium regarding ^{14}C production and decay. Astonishingly high are the ^{10}Be concentrations during the past 1000 years, partly as a result of the three periods of a quiet Sun, partly as a result of a decrease of the geomagnetic field intensity. The increase of the ^{14}C:C ratio during this period is also reflected in the direct ^{14}C measurements.

To obtain a perfect picture of the ^{10}Be production one should in principal measure its precipitation history over the entire globe. Samples suited for this purpose are hard to get from mid-latitude glaciers, but correspondingly high resolved information from Antarctica is available (G. M. Raisbeck, this Symposium). The Wolf, Spörer and Maunder Minima are also visible in these records. The Maunder Minimum is less pronounced than in the Greenland cores.

This raises the question whether transport and deposition, or polarity effects caused the different distribution of cosmic-ray produced ^{10}Be in the two polar area during the periods of the quiet Sun.

Conclusions

The ^{10}Be record in ice cores and the ^{14}C record in tree rings give insights into Earth and planetary system processes that hitherto seemed beyond the potential of the natural sciences. In the following some of the major results are summarized.

1. During quiet periods of the Earth system the records essentially give information on the production rate of ^{10}Be and ^{14}C, whereas in periods of major climatic change, like the glacial–post-glacial transition, *ca.* 15000–10000 years BP, the fluctuations in the records are dominated by Earth system changes.

2. A major result is the widely consistent information on the periods of the quiet Sun with high cosmogenic isotope production; during the past 5000 years seven periods of high ^{10}Be and ^{14}C concentrations have been identified. They provide a basis to study the possible effects of variations of solar properties on climate.

3. The good agreement between the short-term ^{14}C variations measured on tree rings and those models calculated from the short-term ^{10}Be variations supports the validity of the carbon cycle model used, at least regarding variations with characteristic times of *ca.* 200 years.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

However, with potential effects of global change this model may lose part of its predictability.

4. More difficult is the explanation of the long-term ^{14}C change. There are two possible reasons.

(i) The decrease of ^{14}C after last glaciation until 500 A.D. may reflect a carbon system not in equilibrium regarding production and decay after the drastic effects during the glacial–post-glacial transition. At maximum such a disequilibrium was of the order of $\Delta^{14}C = 100‰$, which seems at the low side of expectations.

(ii) There is geomagnetic modulation of production as partly suggested from the ^{10}Be record.

5. Astonishing is the increase in ^{14}C and ^{10}Be after 500 A.D. Possible reasons are a decrease in the Earth's magnetic field; an unusual summation of periods of a quiet Sun; meteorological effects in the ^{10}Be deposition process; some unknown carbon cycle effects regarding the ^{14}C increase.

To answer some of the questions raised here, there is an urgent need for additional ^{10}Be measurements on ice cores in both hemispheres, but also attempts to resolve the 11-year cycle in the ^{14}C record, especially during, for example, the solar Maunder Minimum.

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PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES