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Reconstructions of the ^{14}C Cosmogenic Isotope Content from Natural Archives after the Last Glacial Termination

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Abstract—Data on the content of the ^{14}C cosmogenic isotope in tree rings, which were obtained as a result of laboratory measurements, are often used when solar activity (SA) is reconstructed for previous epochs, in which direct observations are absent. However, these data contain information not only about SA variations but also about changes in the Earth climatic parameters, such as the global temperature and the CO_2 content in the Earth's atmosphere. The effect of these variations on the ^{14}C isotope content in different natural reservoirs after the last glacial termination to the middle of the Holocene is considered. The global temperature and the CO_2 content increased on this time interval. In this case the ^{14}C absolute content in the atmosphere increased on this time interval, even though the ^{14}C to ^{12}C isotope concentration ratio (as described by the $\Delta^{14}C$ parameter) decreased. These variations in the radiocarbon absolute content can be caused by its redistribution between natural reservoirs. It has been indicated that such a redistribution is possible only when the rate of carbon exchange between the ocean and atmosphere depends on temperature. The values of the corresponding temperature coefficient for the 17–10 ka BC time interval, which make it possible to describe the carbon redistribution between the ocean and atmosphere, have been obtained.

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1. INTRODUCTION

As is known, the ^{14}C cosmogenic isotope is produced in the Earth's atmosphere under the action of galactic cosmic rays (GCRs) that penetrate into the Earth's atmosphere from outer space. In this case the GCR intensity near the Earth's orbit does not remain the same but rather changes under different SA. This fact makes it possible to use the measured ^{14}C isotope concentrations in tree rings in order to study past SA. The data on the ^{14}C isotope content in tree rings and in the Earth's atmosphere (e.g., (Reimer et al., 2009)) cover a time interval including several ten thousand years, during which the climate changed. Radiocarbon data also contain information about these changes. To reconstruct the radiocarbon production rate in the Earth's atmosphere in the past, it is necessary to separate the effect of the climatic and solar influences on the ^{14}C isotope content in the Earth's atmosphere.

In this work we consider the time interval from 17 ka BC to the middle of the Holocene (~5000 years BC). From ≈18–17 ka BC, the temperature and carbon dioxide amount in the Earth's atmosphere simultaneously increased. The interglacial period (the Holocene), with a relatively stable warm climate, started at ≈9.5 ka BC. The global temperature change on this time interval is shown in Fig. 1a based on the indirect data (Marcott and Shakun, 2015). Figure 1b also presents data on a change in the carbon dioxide amount in the Earth's atmosphere (Monnin et al., 2004) and variations in the relative (the ratio to the stable ^{12}C isotope concentration) content of the ^{14}C isotope in the atmosphere ($\Delta^{14}C$) (Reimer et al., 2009) (Fig. 1c).

Based on these data, we can calculate (see, e.g., (Kuleshova et al., 2015)) the change in the ^{14}C isotope content in the atmosphere using the formula:

$$\frac{N_a(t)}{N_a(t_0)} = \frac{CO_2(t)}{CO_2(t_0)} \times (1 + \Delta^{14}C(t)/100)/(1 + \Delta^{14}C(t_0)/100), \quad (1)$$

where $N_a(t)$ is the absolute content of the ^{14}C isotope in the atmosphere at instant t, $CO_2(t)$ is the carbon dioxide concentration in the atmosphere, $\Delta^{14}C$ is given in percent, and t_0 is the initial instant. The calculation results are presented in Fig. 1d. Figure 1d indicates that the ^{14}C isotope absolute content increases before the Holocene, which evidently results from the redistribution of CO_2 (including $^{14}CO_2$) between the ocean and atmosphere upon climate warming (see also (Roth and Joos, 2013)).

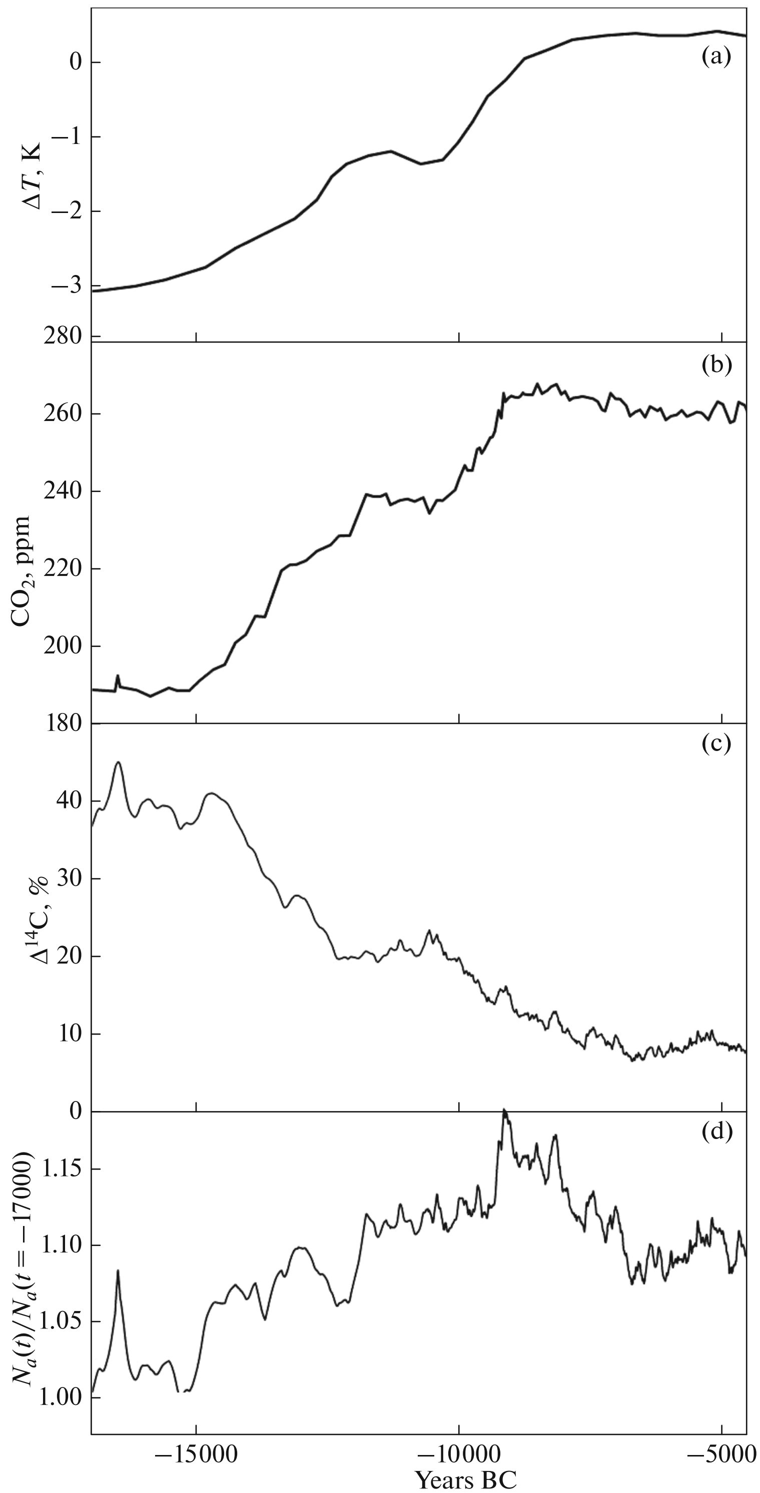

Fig. 1. (a) Change in the global temperature according to (Marcott and Shakun, 2015); (b) the carbon dioxide concentration in the Earth's atmosphere (Monnin et al., 2004); (c) a change in the relative ^{14}C isotope abundance in the atmosphere (Reimer et al., 2009); (d) a change in the ^{14}C absolute content in the atmosphere as calculated based on the above data.

2. CHANGE IN THE ^{14}C ISOTOPE CONTENT IN NATURAL ARCHIVES

The ^{14}C isotope exchange between natural reservoirs can be described by the five-reservoir model (e.g., (Dorman, 1978)), which will be used by us in further calculations. We will calculate $N_b(t)$, $N_h(t)$, $N_{mO}(t)$, and $N_O(t)$, which are the ^{14}C contents in the biosphere, and humus, respectively, as well as in the

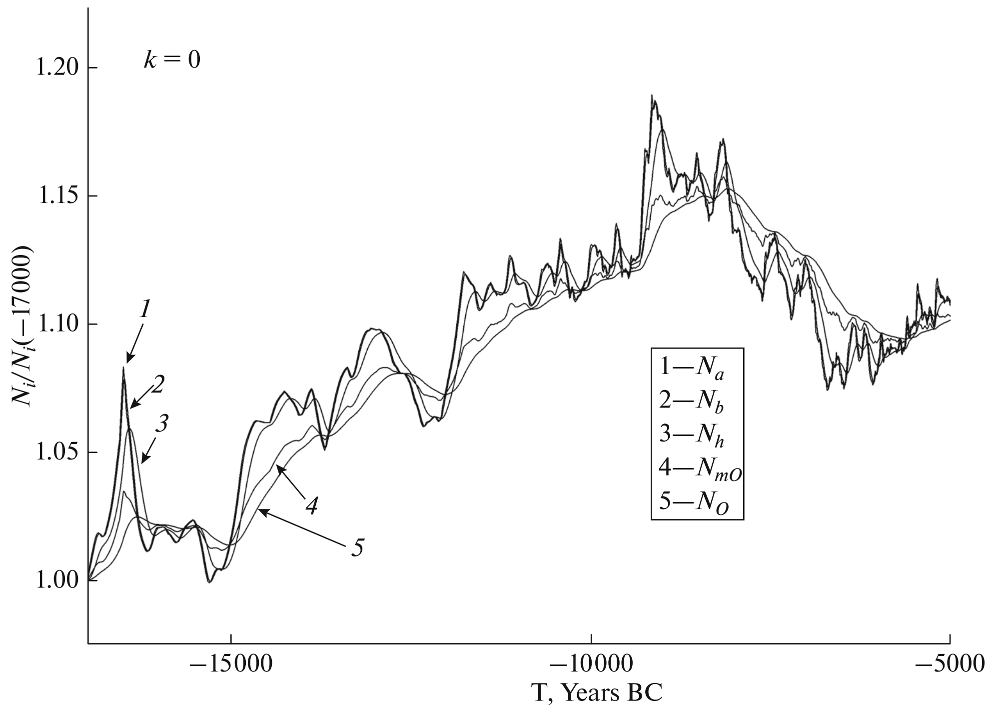

Fig. 2. Change in the ^{14}C isotope content in the atmosphere (N_a), biosphere (N_b), humus (N_h), and in the upper (N_{mO}) and deep (N_O) ocean layers at $k = 0$ and the following initial conditions: $N_a(t_0) = 6.36\text{E} + 09$ cm^{-2}, $N_b(t_0) = 3.24\text{E} + 09$ cm^{-2}, $N_h(t_0) = 10.6\text{E} + 09$ cm^{-2}, $N_{mO}(t_0) = 7.42\text{E} + 09$ cm^{-2}, $N_O(t_0) = 4.66\text{E} + 11$ cm^{-2}, and $t_0 = -17$ ka BC.

upper (mixed) and deep ocean layers at different instants.

In this case it is important to note that the ^{14}C isotope exchange rates between these reservoirs are actually variable, and this model should be adapted in order to take into account the change in the climatic parameters, especially during the periods when they change strongly (see, e.g., (Dergachev and Ostryakov, 1978)). The end of the glacial period and the transition to the Holocene, as well as the Little Ice Age, undoubtedly belong to such periods. The adaptation of the five-reservoir model for the Little Ice Age was described in (Kudryavtsev et al., 2013; Koudriavtsev et al., 2014; Kuleshova at al., 2015]). In these works the exchange rate between the upper ocean and atmosphere λ_{mOa} was modeled by the relationship $\lambda_{mOa} = (1 + k\Delta T)\lambda^0_{mOa}$, where k is the temperature coefficient, and ΔT is global temperature variations.

In the present work, we will also use this relationship for the rate temperature dependence λ_{mOa} and will try to find the possible values of temperature coefficient k, which will allow us to describe the ^{14}C isotope redistribution between the ocean and atmosphere during heating but for the end of the last glacial period and the transition to the Holocene. For the exchange rate between the upper ocean and atmosphere λ^0_{mOa} and other exchange rates, we used the values presented in (Koudriavtsev et al., 2014). Based on Figs. 1a and 1b, we assume in this case that the carbon system was almost equilibrium 17 ka BP. Therefore, during the first calculation stage, we find the equilibrium (stationary) values for the radiocarbon content in different reservoirs for each selected value of coefficient k. We will use these values as initial conditions in further calculations (see Figs. 2–4). We should note here that the initial values $N_a(t_0)$, $N_b(t_0)$, and $N_h(t_0)$ are identical for different k values and $N_{mO}(t_0)$ and $N_O(t_0)$ are different, since we vary only the exchange rate between the ocean and atmosphere in this work.

Figures 2–4 illustrate the calculations for different values of temperature coefficient k. Figure 2 indicates that the ^{14}C content almost simultaneously increases in all reservoirs and that radiocarbon is not redistributed between reservoirs at $k = 0$, i.e., for the case when the temperature dependence of the exchange rate between the upper ocean and atmosphere is neglected.

When coefficient k increases to 0.03 K^{-1}, the ^{14}C isotope content in the ocean increases more slowly than in the atmosphere: the maximal increase in the content of this isotope is about 18 and 6% in the atmosphere and ocean, respectively.

When parameter k increases to 0.04 K^{-1}, the ^{14}C isotope content in the ocean and its total value in all reservoirs (Fig. 3, curve *6*) tend to remain unchanged on the 17–9 ka BC time interval: fluctuations about a certain average level are observed. These fluctuations

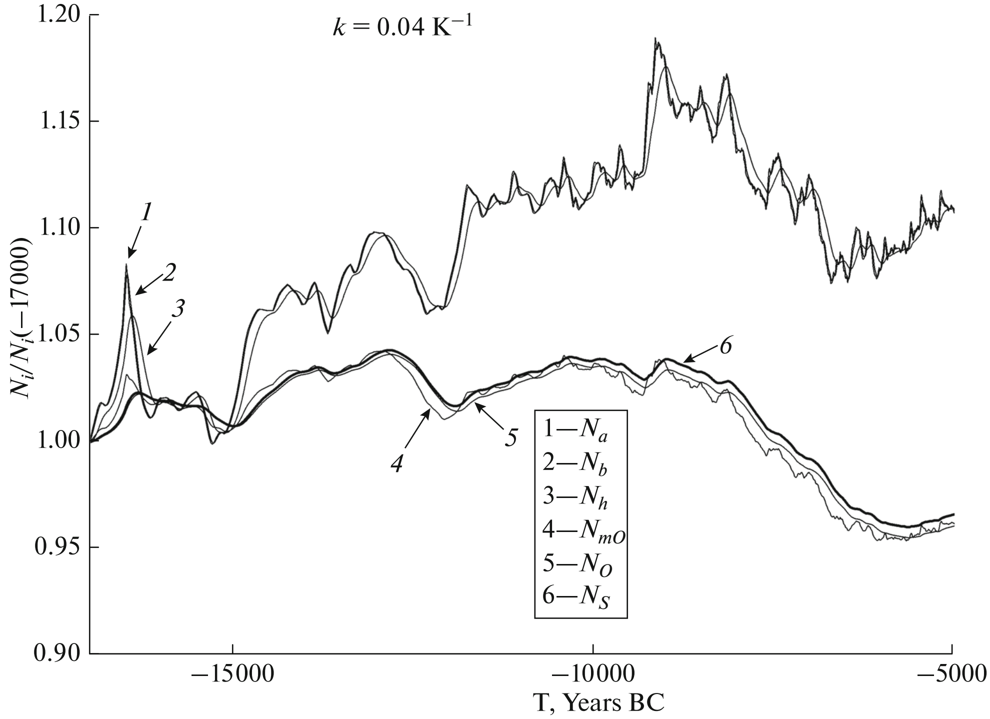

Fig. 3. Change in the ^{14}C isotope content in the atmosphere (N_a), biosphere (N_b), humus (N_h), and in the upper (N_{mO}) and deep (N_O) ocean layers at $k = 0.04$ K^{-1} and the following initial conditions: $N_a(t_0) = 6.36$E + 09 cm^{-2}; $N_b(t_0) = 3.24$E + 09 cm^{-2}; $N_h(t_0) = 10.6$E + 09 cm^{-2}; $N_{mO}(t_0) = 8.39$E + 09 cm^{-2}; $N_O(t_0) = 5.27$E + 11 cm^{-2}; $t_0 = -17$ ka BC; N_S is the total ^{14}C in all reservoirs.

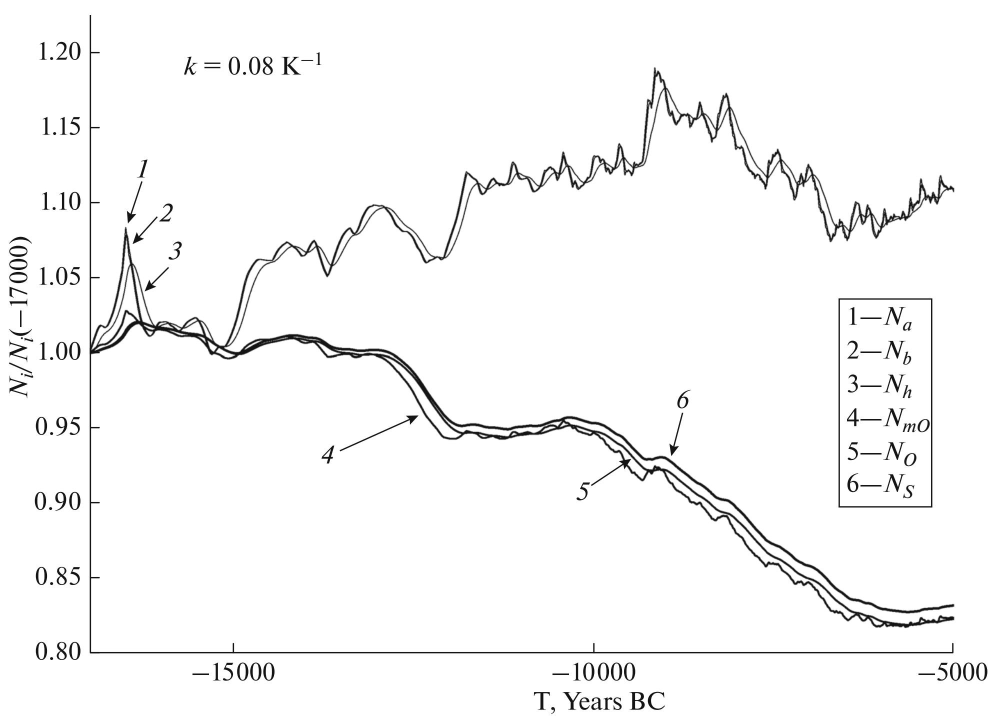

Fig. 4. Change in the ^{14}C isotope content in the atmosphere (N_a), biosphere (N_b), humus (N_h), and in the upper (N_{mO}) and deep (N_O) ocean layers at $k = 0.08$K^{-1} and the following initial conditions: $N_a(t_0) = 6.36$E + 09 cm^{-2}; $N_b(t_0) = 3.24$E + 09 cm^{-2}; $N_h(t_0) = 10.6$E + 09 cm^{-2}; $N_{mO}(t_0) = 9.65$E + 09 cm^{-2}; $N_O(t_0) = 6.07$E + 11 cm^{-2}; $t_0 = -17$ ka BC; N_S is the total ^{14}C in all reservoirs.

can evidently be related to the SA and geomagnetic field variations. In this case we should note that the ^{14}C isotope content in the ocean is much higher than in the other reservoirs; therefore, when this isotope is redistributed between reservoirs, its relative change in the ocean will be minimal.

When the temperature coefficient continues increasing, the pattern begins to change into the opposite one: already at $k = 0.05$ K^{-1}, not only the ^{14}C isotope content in the ocean but also the total radiocarbon start decreasing from 11–10 ka BC. At $k = 0.08$ K^{-1}, the ^{14}C content in the ocean and the total radiocarbon in all reservoirs tend to decrease for the entire time interval considered (Fig. 4, curve *6*).

3. CONCLUSIONS

On the assumption of the presented calculation results, we can conclude that the ^{14}C isotope (and carbon dioxide) redistribution between the ocean and atmosphere can be described by the five-reservoir model with a temperature coefficient of $k \approx 0.04$ K^{-1} for the time interval from 17 ka BC to approximately the beginning of the Holocene. Such a value of coefficient k can be caused by the fact that an increase in the surface water layer temperature by 1 K results in an increase in the partial pressure of CO_2 dissolved in water by ≈ 4%, and that the total CO_2 flow through the ocean surface is proportional to a difference in carbon dioxide partial pressures in the surface water layer and the atmosphere (e.g., Byutner, 1986; Takahashi et al., 1993, 2009; Malinin and Obraztsova, 2011).

In addition, Fig. 3 demonstrate that the calculated total amount of this isotope decreases on the ~9–6 ka BC time interval. This specific feature can result from the fact that data concerning possible changes in the exchange rates between the atmosphere and biosphere, as well as between the biosphere and humus, were absent during this time interval. That is, the simplified model considered by us, which takes into account only the change in the exchange rate between the ocean and atmosphere, can be applied only to the time interval from 17 ka BC to the beginning of the Holocene, when the glacier retreats and the global temperature rises by ~2 K. At the same time, to adapt the five-reservoir model to the Holocene, we should also take into account changes in other reservoirs.

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