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High concentration of atmospheric ^{14}C during the Younger Dryas cold episode

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THE various reservoirs of the global carbon cycle, with their very different residence times, are linked by a complex and evolving system of exchanges for which natural radiocarbon is the most robust tracer[1]. Any change in the sizes of these reservoirs, or the exchange rates between them, could perturb the $^{14}C/^{12}C$ ratio of each other reservoir, and the smallest of them—the atmosphere—would be the most sensitive. In particular, high-resolution reconstructions of past atmospheric $^{14}C/^{12}C$ ratios may provide important clues to the mechanisms of abrupt climate change. Annually laminated lake sediments potentially provide an optimal record in this respect, as they preserve information about both past atmospheric ^{14}C levels and climate changes, providing absolutely dated material beyond the range of tree-ring chronologies and, unlike corals, directly monitor ^{14}C concentrations in atmospheric CO_2. Here we report the relationship between atmospheric ^{14}C concentration and climate changes during the Younger Dryas and early Holocene periods, derived from analyses of the annually laminated sediments of Lake Gościąż, in central Poland. We find that atmospheric ^{14}C concentrations during the Younger Dryas were abnormally high, which we interpret as a reduced ventilation rate of the deep ocean, most probably as a result of a decrease in intensity of the North Atlantic Deep Water formation.

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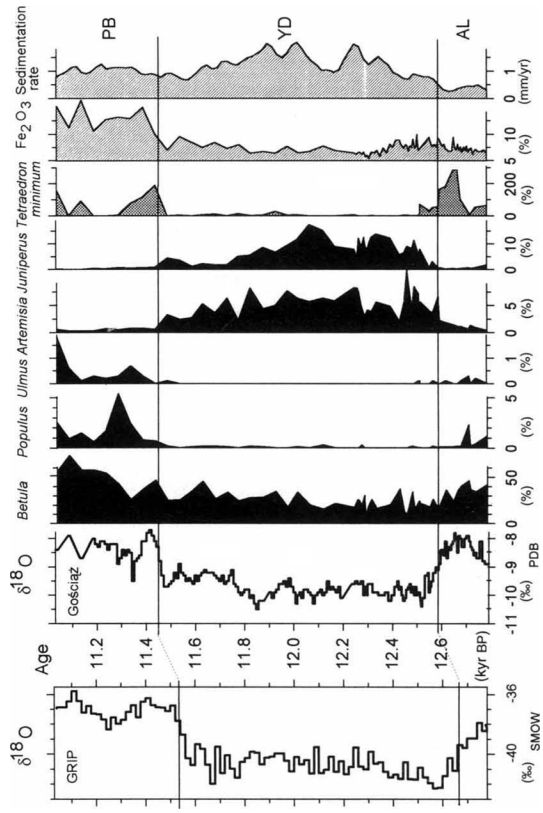

FIG. 1 Combined diagram of proxy climate data recorded in the Late Glacial and Early Holocene part of the Lake Gościąż sediment, compared with the $\delta^{18}O$ record in the GRIP site, central Greenland. PB, Preboreal; YD, Younger Dryas; AL, Allerød; SMOW, standard mean ocean water; PDB, Pee Dee Belemnite. $\delta^{18}O = [(^{18}O/^{16}O)_{sample}/(^{18}O/^{16}O)_{standard}] - 1$. The transition from the Younger Dryas to the Preboreal is illustrated by a definite reduction in *Juniperus* and *Artemisia* pollen frequencies, reflecting the gradual retreat of open shrub–herb communities. A distinct rise in *Populus tremula* type coincident with the appearance of *Ulmus* pollen at the transition, followed by a large maximum in *Betula*, correspond to an expansion of forest communities responding to a change of climate. A more comprehensive discussion of available pollen data is given elsewhere[7,39]. The rapid expansion of forests and formation of low-redox-potential soils is also reflected in the abrupt increase in content of iron, which could more easily be leached from soils under more reductive conditions. The decrease in sedimentation rate (with no significant change in annual accumulation of $CaCO_3$) and the decrease in the content of aluminium (not shown) suggests a limitation of erosion at that time. The development of the alga *Tetraedron minimum* in the lake at the Younger Dryas/Preboreal transition indicates rising summer temperatures[39]. The main changes in vegetation are synchronous with those of $\delta^{18}O$ of authigenic carbonates. The scale of the calendar age (±120 years) is based on a wiggle-match of AMS radiocarbon dates to the German oak chronology. The record of sedimentation rate has been smoothed.

Because of natural variations in atmospheric ^{14}C concentration, the calibration curve showing the relationship between radiocarbon and calendar age is not linear. One of its most prominent features is the plateau at the radiocarbon age of ~10,000 ^{14}C yr BP (ref. 2), being also the generally accepted date of the Pleistocene/Holocene or Younger Dryas/Preboreal boundary[3]. It complicates an absolute dating and synchronization of that boundary between different sites when the timescale is reconstructed by radiocarbon dates. But even for the best absolute chronologies currently available (central Greenland ice cores and German tree-ring series), a significant discrepancy exists between estimates of the age of the boundary[4-6]. This discrepancy can be interpreted either as a real delay in climate warming in Europe with respect to the North Atlantic or as a result of an as yet undetected error in counting annual ice layers and/or an incorrect match between German pine and oak chronologies[2]. Another possibility of synchronization of that boundary is given by direct comparison of its position with respect to the worldwide synchronous ^{14}C plateau. This cannot be done in ice cores, but is possible in varved lacustine sediments.

The archive analysed was the sediment of Lake Gościąż in central Poland[7]. The lake (52° 30′ N, 19° 20′ E) contains two basins (24.6 m and 12.5 m), separated by a shallowing to ~6 m. The sediment is annually laminated, consisting of layers dominated by authigenic calcite and deposited during summer months, and layers rich in organic matter, deposited in the autumn/winter seasons. The varve-to-varve correlation of cores from both basins enabled the construction of continuous chronology containing 9,662 ± 90 varves for the lower part of the sediment. Because of the poor quality of the lamination in the upper part, this chronology is considered floating.

The abrupt climate changes at the Late Glacial/Holocene transition left a distinct signature in the sediments of Gościąż Lake (Fig. 1). The fluctuations in ^{18}O isotope composition of authigenic carbonates deposited in the lake coincide with major changes in the vegetation cover in the area[7]. The Gościąż $\delta^{18}O$ record ($\delta^{18}O$ is defined in Fig. 1 legend) closely resembles those observed in GRIP and GISP2 ice cores from central Greenland[5,6] (compare Figs. 1 and 3). The Younger Dryas/Preboreal transition in the Gościąż sediments is marked by ~2‰ increase in $\delta^{18}O$, completed in ~70 years. Whereas a significant part of $\delta^{18}O$ signal in ice cores is probably related to changes of sources of moisture over central Greenland[8], the prevailing influence of the westerly circulation maintained over the Europe during the glacial period[9] makes the interpretation of our $\delta^{18}O$ record meaningful in terms of temperature change.

The accelerator mass spectrometry radiocarbon dating was done using plant macrofossils of terrestrial origin. The suite of dates (Fig. 2) enabled synchronization of the Lake Gościąż chronology with the German oak (younger) and German pine (older) chronologies[2], using the 'wiggle-matching' procedure[10]. Matching the set of youngest Gościąż samples to German oak chronology, we date the Younger Dryas/Preboreal boundary to 11,440 ± 120 BP. Thus, within the indicated error, this transition in central Europe was synchronous with that recorded in Greenland ice cores (11,550 ± 90 BP, ref. 5; 11,640 ± 250 BP, ref. 11).

The early Holocene dates of analysed macrofossils (Fig. 2) clearly show a shift of Lake Gościąż varve chronology with respect to that of German pines. The wiggle-match of Gościąż to pine ^{14}C dates is excellent, enabling precise synchronization of both chronologies. The ^{14}C ages of two plateaux shown by our data agree very well with that reconstructed in high-precision dendrocalibration, indicating that systematic errors in dating and/or problems related to rebedding and contamination of Gościąż macrofossils are negligible. But the synchronization requires either a 'compression' of Lake Gościąż varve chronology or the revision of the tentative dendro-match between German oak and pine chronologies. The rejection of 200 varves from the 800-year sequence between 9,700 and 10,500 BP seems unlikely because the chronology is replicated in several cores from two separate lake basins. The revision of the dendro-match, on the other hand, would expand the plateau of 8,900 ^{14}C BP to ~600 years. In fact, such a long plateau is suggested by data from Soppensee[12], where five dates of 8,900–9,000 ^{14}C BP occur in a time span of 700 years.

Independently of the absolute age, we can determine the position of the Younger Dryas/Preboreal boundary at the 10,000 ^{14}C BP plateau. The warming in Poland, as abrupt as in Greenland (Fig. 3), began about 250 years before the end of the plateau. On the other hand, the boundary in German pines, defined as the beginning of slow rises of $\delta^{13}C$ and δD in wood cellulose[4], coincides with the end of the plateau. There is no doubt that the changes of isotopic composition in German wood started ~200 years after the main $\delta^{18}O$ increase in Lake Gościąż, during a period of distinct development of elm trees in Poland, that is, after the Younger Dryas period. Because both regions are situated only 1,000 km apart, exposed to the prevailing westerly circulation, it is very unlikely that the major warming began 200 years earlier in the region further east. Moreover, the increase in $\delta^{13}C$ and δD in trees continued for ~500 years, contrary to a very fast change observed in both Gościąż and ice core $\delta^{18}O$ data. We therefore conclude that the increase of $\delta^{13}C$ and δD in German pines is delayed somewhat and transformed with respect to the abrupt climate warming at the beginning of the Holocene.

The profiles of $\delta^{13}C$ and δD were derived from pine trunks excavated from Danube valley deposits[4]. Major warming at the

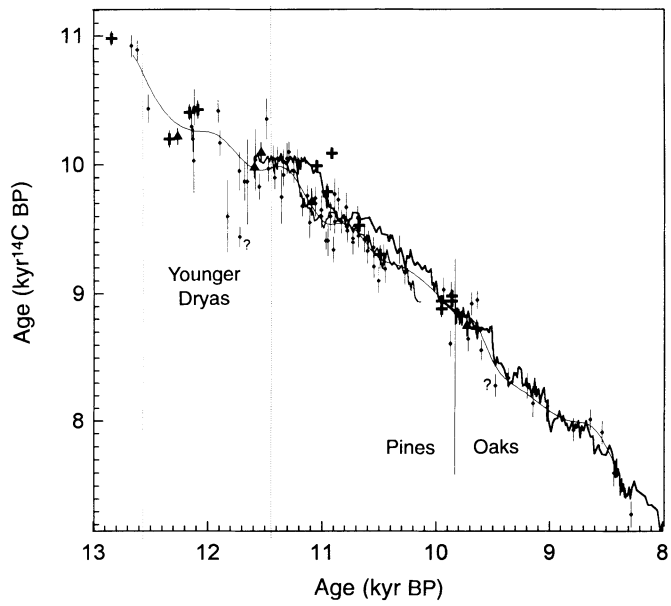

FIG. 2 AMS radiocarbon dates of terrestrial macrofossils extracted from the Lake Gościąż sediments, superimposed on the radiocarbon calibration data. The ages of terrestrial macrofossils are free of hard water or reservoir effect. To obtain enough material for dating (~1 mg), the macrofossils were collected from five precisely correlated cores. Apart from typical short-living macrofossils widely used in dating lake sediment (birch nutlets, birch fruit scales, bud scales and willow leaves), fragments of pine peridermis were collected. The suitability of peridermis fragments for dating sediments was checked by separate dating of pine peridermis and bud scales found in the same layers. The agreement of dates of paired scales and peridermis is satisfactory. Circles, Lake Gościąż macrofossils; triangles, Barbados corals[14], crosses, Huon Peninsula corals[15]; heavy and light broken lines, German pines in original position[2], and shifted by 200 years, respectively. The calendar timescale of Gościąż points (±120 years) is based on a ^{14}C wiggle-match to German oaks. The smoothed curve is a cubic spline fitted to the Gościąż dates. The two samples indicated by question marks were probably contaminated.

Younger Dryas/Preboreal boundary must have lead to a marked influx of isotopically depleted water from the melting of Alpine glaciers to the Danube river and its flood plains. Even today, the Danube river is depleted in deuterium by about 10–15‰ during summer, owing to meltwater from the Alps (D. Rank, personal communication). As the δD of wood cellulose is mainly controlled by isotopic composition of environmental water[13], the apparent 200-year delay of the analysed pines in response to an improving climate could result from enhanced contribution of isotopically depleted meltwater. The subsequent gradual increase of δD may then reflect a diminishing contribution of this source, when compared with local precipitation. The delay of δ^{13}C of wood cellulose to major warming is not well understood, but, it might result from a combination of the effects of increased water supply and the biological response to increased temperature.

In the late Allerød and Younger Dryas, the radiocarbon dates of Gościąż macrofossils (Fig. 2) generally confirm the U/Th calibration[14–16], showing a plateau at 10,400 ^{14}C BP and a rapid decline of ^{14}C age at the beginning of the Younger Dryas. The wiggles of the radiocarbon calibration curve reflect the changes of ^{14}C concentration in the atmosphere. Our data (Fig. 4*a*) confirm the large increase of Δ^{14}C at the onset of the Younger Dryas, by 40–70‰ within 300 years. (Δ^{14}C is defined in Fig. 4 legend.) This increase, demonstrated by four data points determined in three independent laboratories on two completely different types of material, would then be the largest known in the Late Glacial and Holocene records of atmospheric ^{14}C concentration. On the other hand, the decrease in Δ^{14}C by 50–70‰ at the Younger Dryas/Preboreal boundary, is the largest and most rapid of declines occurring regularly between 12.3 and 9.8 kyr BP. The striking coincidence in distinct climate and radiocarbon variations suggests the involvement of some common causes.

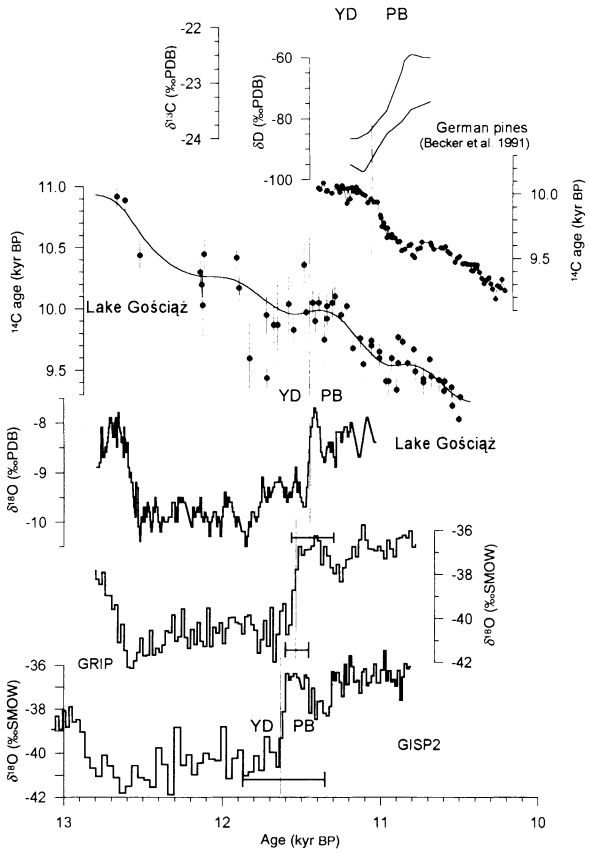

FIG. 3 Synchronization of the ^{14}C calibration curve and stable-isotope records representing the Late Glacial and Early Holocene. YD, Younger Dryas; PB, Preboreal. The match of Lake Gościąż and German pine chronologies (from ref. 4) requires a shift of one of chronologies by ~200 years. The changes in δ^{13}C and δD of German pines at the YD/PB boundary[4] are represented by trend curves. Horizontal bars at the YD/PB transitions in Gościąż and Greenland ice represent age uncertainties ($\pm 1\sigma$).

Among possible causes of atmospheric Δ^{14}C variations are the changes of production rate of ^{14}C and reorganizations in the global carbon cycle. It is commonly accepted that the medium-term variations are connected with changing solar activity, whereas the secular Δ^{14}C trend in the Holocene reflects changes in the geomagnetic field[17]. The past geomagnetic variations have been reconstructed with a relatively good time resolution[18]. We used these data to simulate variations of atmospheric ^{14}C concentration. We ran simple box models[19,20] with all parameters kept constant except for ^{14}C production rate, dependent on magnetic field according to Lal[21]. The results given by different models were very similar. In Fig. 4*b* we present the residuals between the smoothed curve of the observed Δ^{14}C and averages of those calculated from geomagnetically induced production rates.

The results of the calculations are sensitive to the choice of initial conditions and the model adjustment of the absolute value of ^{14}C production. In the original adjustment, the ^{14}C production rate corresponding to the current geomagnetic field maintains the steady-state Δ^{14}C = 0‰ in the preindustrial atmosphere. The simulated curve that best fits the early Holocene data was obtained with an initial steady state of Δ^{14}C = 230‰ and a production rate systematically lowered by 5%. Consequently the variability in geomagnetic field is then a good candidate for explaining the long-term Δ^{14}C decrease. In contrast, because of the buffering effect of the ocean, the concentration of atmospheric radiocarbon is not sensitive to hypothetical short-term variations of dipole moment in the range allowed by uncertainties in palaeomagnetic reconstructions. Moreover, the presence of an undetected magnetic anomaly concomitant with a major climate change seems very unlikely. We thus conclude that the geomagnetic Late Glacial/Early Holocene rapid shifts of atmospheric Δ^{14}C cannot be explained only by changes of the Earth's magnetic field.

The other mechanism might be changes in solar activity, although there is no direct evidence of this. But the regular occurrence of 180–220-year-wide Δ^{14}C maxima over the Holocene indicates that they could be induced by solar events similar to those responsible for the Spörer and Maunder minima[17]. Therefore, unless contradictory evidence is shown, we can only assume that the Late Glacial solar-wind-induced ^{14}C variations were similar to those observed in the Holocene. Although the amplitude of observed Δ^{14}C variations during the Late Glacial is similar, the duration (1000–2500 years) is distinctly larger than those attributed to the Sun in the Holocene[22].

The general property of the global carbon cycle is that nearly all radiocarbon, which is produced in the atmosphere, decays in the deep ocean. For that reason, the concentration of radiocarbon in atmosphere depends on the size of the atmospheric reservoir and the rate of carbon exchange with the ocean. The large increase in atmospheric CO_2 concentration by 80 p.p.m. between 17 and 10 kyr BP is well documented[23], a change that might decrease the Δ^{14}C by 25–75‰ (refs 24, 25). The rise in CO_2 concentration could then explain the 20‰ decrease between the mean pre-Younger Dryas and Holocene Δ^{14}C residuals (Fig. 4*b*), but it is too slow to be responsible for the Δ^{14}C maximum within the Younger Dryas.

It is also accepted[22] that the winds were stronger in glacial time and the Younger Dryas[26], and the enhanced rate of carbon exchange at the atmosphere–ocean interface[22] could decrease Δ^{14}C by 20‰. But that mechanism works in the opposite direction and cannot explain the observed Δ^{14}C variations.

The last possible mechanism is the exchange with the deep oceans. This is especially important because a significant part of total gas exchange is related to the formation of North Atlantic

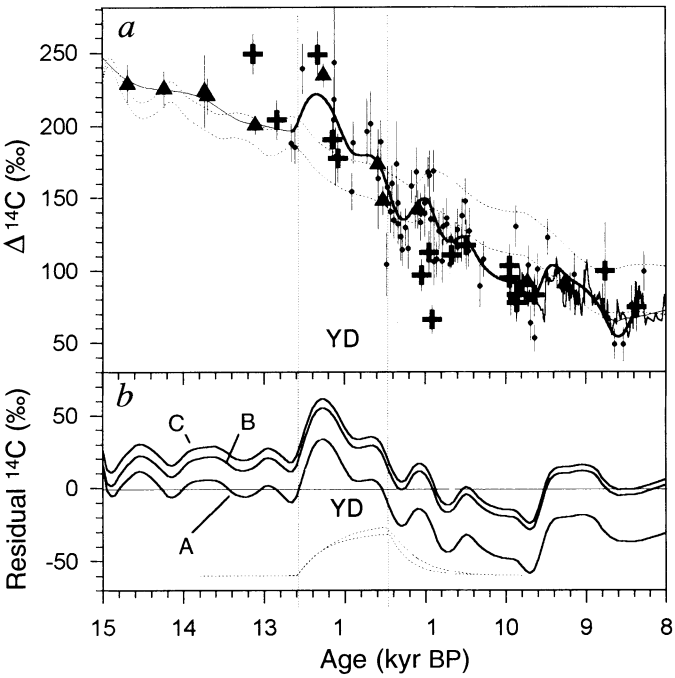

FIG. 4 *a*, Reconstruction of atmospheric ^{14}C concentration in the Late Glacial and Early Holocene. $\Delta^{14}C$ is defined as the relative deviation of radiocarbon concentration from that in the standard of the modern biosphere, and is expressed in per mil. The symbols are the same as in the previous figures. For clarity, the calibration data based on German pines are not included. The uncertainties in $\Delta^{14}C$ values resulting from error in the calendar age of the Gościąż varve chronology are neglected. The cubic splines fitted to Gościąż data (heavy smooth line) and to the coral data (thin smooth line), put together, serve as input data for model calculations (see text). The two broken lines represent $\Delta^{14}C$ curves, calculated on the basis of geomagnetically induced production rates: upper curve, $\Delta^{14}C_0 = 230‰$, ^{14}C production rate from Tric *et al*[18]; lower curve, production rate systematically lowered by 5%. *b*, Differences between the reconstructed ^{14}C concentrations (data from Gościąż macrofossils and corals) and that calculated on the basis of geomagnetically induced production rates. A, $\Delta^{14}C_0 = 230‰$, ^{14}C production rate from Tric *et al*[18]; B, production rate systematically lowered by 5%; C, $\Delta^{14}C_0$ lowered by 10‰. The lower broken line shows the amplitude of atmospheric $\Delta^{14}C$ change, simulated by the PANDORA box model[20], by changing the NADW flux from 30 Sv to 10 Sv and the Pacific loop from 15 Sv to 20 Sv and back, at the onset and termination of the Younger Dryas (YD), respectively. The upper broken line shows the amplitude of atmospheric $\Delta^{14}C$ change, simulated by the 13-box model[36], by changing the NADW flux from 20 Sv to 10 Sv and back, at the onset and termination of the YD, respectively.

Deep Water (NADW), and switching NADW formation 'on' and 'off' would be a plausible mechanism of abrupt climate changes[27]. Many palaeoceanographic reconstructions[28–31] give firm evidence for decreased NADW formation during both the Younger Dryas and the Late Glacial Maximum (LGM). That point has been challenged by two recent studies of benthic $\delta^{13}C$ (refs 32, 33). One hypothesis to explain the discrepancies relies on a shutdown of NADW transport to the deep ocean, but with transport to intermediate depths persisting during the LGM and the Younger Dryas[34].

The earlier ^{14}C dating of contemporaneous benthic and planktonic foraminifera suggested that the overall ventilation of the deep oceans at the LGM was much weaker than today[35]. More recent data[20] suggest that the age of NADW during the LGM was twice as great as it is today, but the age of Pacific deep water was greater only by 70 ± 105 years. To simulate the corresponding change in atmospheric $\Delta^{14}C$ we used the standard PANDORA[20] and a 13-box model[36]. By reducing the NADW flux by a factor of three and slightly increasing the circulation in the Indo-Pacific, the deep Atlantic beomes twice as old, the age of the deep Pacific increases by 150 years, and the atmospheric $\Delta^{14}C$ increases by about 30‰, in reasonable agreement with the observed amplitude of atmospheric $\Delta^{14}C$ changes at the boundaries of the Younger Dryas (Fig. 4*b*). A very similar $\Delta^{14}C$ wiggle (35‰) is obtained when the NADW flux is simply halved.

The $\Delta^{14}C$ record within the Younger Dryas, suggests that the lowest oceanic ventilation is in the earlier part of the Younger Dryas. This seems to be consistent with the amelioration of climate in the second half of the Younger Dryas, reconstructed in Lake Gościąż sediment[7] and other regions in Europe[37]. The rapid increase in atmospheric $\Delta^{14}C$ requires a decrease in deep-ocean ventilation in the early stage of the Younger Dryas that is significantly larger than the two-fold decrease in NADW flux alone. Unfortunately, available benthic/planktonic data are too scarce to reconstruct directly the magnitude of ventilation change in the Younger Dryas and, for the moment, cannot give a clear view of what actually occurred in the ocean during that period.

Recently it has been pointed out[38] that the possibility that the thermohaline circulation was shut down during the Younger Dryas, though tempting, could be eliminated by studying atmospheric $^{14}C/^{12}C$ ratios. The abnormally high ^{14}C levels during the Younger Dryas shown by our results do not contradict the hypothesis that the thermohaline circulation decreased during the Younger Dryas. □

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