Deep-Sea Coral Evidence for Rapid Change in Ventilation of the Deep North Atlantic 15,400 Years Ago

Jess F. Adkins,† Hai Cheng, Edward A. Boyle, Ellen R. M. Druffel, R. Lawrence Edwards

Coupled radiocarbon and thorium-230 dates from benthic coral species reveal that the ventilation rate of the North Atlantic upper deep water varied greatly during the last deglaciation. Radiocarbon ages in several corals of the same age, 15.41 ± 0.17 thousand years, and nearly the same depth, 1800 meters, in the western North Atlantic Ocean increased by as much as 670 years during the 30- to 160-year life spans of the samples. Cadmium/calcium ratios in one coral imply that the nutrient content of these deep waters also increased. Our data show that the deep ocean changed on decadal-centennial time scales during rapid changes in the surface ocean and the atmosphere.

Records from Greenland ice cores have revealed that the glacial polar climate shifted extremely rapidly several times. From 18,000 to 40,000 years ago (ka), glacial climates were periodically punctuated by rapid returns to milder conditions, called interstadials, that lasted for hundreds of years (1). An abrupt change, fewer than several decades, marked the end of the Younger Dryas (YD) cooling event (11.5 ka) (2, 3). Analogous changes in sea surface properties have been found in sediment cores from the North Atlantic, Caribbean Basin, and Santa Barbara Basin (4). The global extent of these surface ocean and atmosphere correlations shows that these reservoirs change coherently and abruptly during major climate transitions.

The role of the deep ocean in these events has been uncertain. As the deep ocean is the largest reservoir of heat and mass in the climate system, it plays an important role in the mechanism of rapid climate change. Shifts in deep circulation patterns, recorded by the stable carbon isotope composition of benthic foraminifera in deep-sea sediments, have been shown to correspond to interstadials of longer duration (5). One problem has been that the time resolution of sediments is limited by bioturbation of the upper few centimeters. We present coupled 230Th and 14C dates and Cd/Ca ratios from deep-sea corals and show that this new archive records a very rapid deep circulation change at the beginning of the last deglaciation.

Today the deep North Atlantic Ocean is partially ventilated by North Atlantic Deep Water (NADW), a low-nutrient water mass that is formed in the Nordic and the Labrador Seas. NADW reaches to bottom depths in the modern western Atlantic Ocean as far south as about 40°N and then spreads southward with a core at ~3000 m (6). However, the large-scale circulation at the last glacial maximum (LGM) was different (7, 8). Nutrient-rich bottom waters of a southern origin spread to 60°N (9-11) and NADW shoaled to form its glacial analog, glacial North Atlantic intermediate/deep water (GNAI/DW). The vertical boundary between low-nutrient GNAI/DW and southern source waters lay between 1500 and 2000 m (measured between 50° and 60°N) (11). Our deep-sea coral samples are from 40°N in the western basin of the North Atlantic and a water depth around 1800 m. They lie near the boundary between glacial water masses and are a sensitive indicator of the timing of deglacial deep circulation shifts.

Similar to surface reef-building corals, some deep, aragonitic, solitary corals have paired light and dark density bands. This...
banding structure and growth rates of about 0.2 to 1 mm per year provide the potential for annual to decadal records of deep-ocean change (12, 13). Deep-sea corals can live at depths of 60 to 6000 m, but most live between 500 and 2000 m (14). Corals have been dredged from the ocean floor since the days of the Challenger expedition (1872–1876) and thousands of samples exist in collections. Because most of these samples had not been dated, we developed an age screening method to sort samples (15). The deep-sea scleractinia Desmophyllium cristagalli and Scleroneis sp. are two of several cosmopolitan species that are abundant in dredge collections.

We measured paired thermal ionization mass spectrometry (TIMS) uranium series dates (13, 16) and accelerator mass spectrometry (AMS) radiocarbon dates (17) on six corals (Table 1). From these paired dates, we calculated the initial coral 14C/12C ratio, denoted Δ14C (18), which is also the Δ14C value of the water in which the coral grew. In the modern ocean, we know the initial Δ14C value for deep-water masses and therefore can calculate a radiocarbon deficiency from a deep Δ14C value measured downstream. However, the initial Δ14C value of surface waters at deep-water formation zones has changed with time. This initial water Δ14C value is determined largely by the atmospheric Δ14C history, which depends on the production rate of 14C in the upper atmosphere and the exchange of 14C between active carbon reservoirs. To avoid this problem of variable initial Δ14C, we calculated a 14C projection age to remove the effects of changing atmospheric radiocarbon contents on the measured deep Δ14C for each of our samples (19). This procedure extrapolates back from the deep value along a closed system 14C decay path to the intersection with the known atmospheric Δ14C record. The difference between the age of the intersection and the calendar age of the coral is the 14C projection age. The more conventional ventilation age, measured relative to the sea surface, is calculated by subtracting the surface ocean reservoir age from the 14C projection age of the sample (Fig. 1A). We subtracted the modern North Atlantic reservoir age of 400 years from the 14C projection ages to obtain the ventilation age value relative to the tropical surface ocean for all our data. During the YD the high-latitude North Atlantic reservoir age was 600 to 800 years versus the modern value of 400 years (20). Subtracting much more than 400 years from the projection age of our data at 13.7 ka would imply a negative ventilation age. Therefore, during the Bolling-Allerød warm period, the reservoir age must have been close to the modern value, with an upper limit of 700 years. Although our data are not from the YD, they provide an indication of how much the reservoir age can change.

During deglaciation sea level increased abruptly (16, 21) (Fig. 1A). Two major melt-water pulses from the Northern Hemisphere ice sheets influenced thermohaline circulation during this deglacial time period.

Table 1. AMS radiocarbon and TIMS uranium series dates (13, 16) and accelerator mass spectrometry (AMS) radiocarbon dates (17) on six corals (Table 1). From these paired dates, we calculated the initial coral 14C/12C ratio, denoted Δ14C (18), which is also the Δ14C value of the water in which the coral grew. In the modern ocean, we know the initial Δ14C value for deep-water masses and therefore can calculate a radiocarbon deficiency from a deep Δ14C value measured downstream. However, the initial Δ14C value of surface waters at deep-water formation zones has changed with time. This initial water Δ14C value is determined largely by the atmospheric Δ14C history, which depends on the production rate of 14C in the upper atmosphere and the exchange of 14C between active carbon reservoirs. To avoid this problem of variable initial Δ14C, we calculated a 14C projection age to remove the effects of changing atmospheric radiocarbon contents on the measured deep Δ14C for each of our samples (19). This procedure extrapolates back from the deep value along a closed system 14C decay path to the intersection with the known atmospheric Δ14C record. The difference between the age of the intersection and the calendar age of the coral is the 14C projection age. The more conventional ventilation age, measured relative to the sea surface, is calculated by subtracting the surface ocean reservoir age from the 14C projection age of the sample (Fig. 1A). We subtracted the modern North Atlantic reservoir age of 400 years from the 14C projection ages to obtain the ventilation age value relative to the tropical surface ocean for all our data. During the YD the high-latitude North Atlantic reservoir age was 600 to 800 years versus the modern value of 400 years (20). Subtracting much more than 400 years from the projection age of our data at 13.7 ka would imply a negative ventilation age. Therefore, during the Bolling-Allerød warm period, the reservoir age must have been close to the modern value, with an upper limit of 700 years. Although our data are not from the YD, they provide an indication of how much the reservoir age can change.

During deglaciation sea level increased abruptly (16, 21) (Fig. 1A). Two major melt-water pulses from the Northern Hemisphere ice sheets influenced thermohaline circulation during this deglacial time period.

![Fig. 2. Cd/Ca data and radiocarbon ages from D. cristagalli sample JFA 24.8. This sample from a depth of 1794 m at 40°N in the western basin of the Atlantic Ocean was 230Th-dated to be 15,410 ± 170 years old. Radiocarbon ages from three samples of a single septum show a 670 ± 60 years continuous increase in 14C age, over and above the growth trend, toward younger ages at the top of a sample. Cd/Ca data were sampled at about 2.3 mm average resolution along the line connecting the top and bottom 14C samples. These data also show a continuous trend to older, more poorly ventilated waters during the lifetime of the coral. Our growth rate study using 230Th dating in five modern D. cristagalli constrains the extension rate to be no less than about 0.2 mm/year and therefore the life span of this sample is no more than about 160 years (13). This evidence for a rapid change in deep ocean circulation corresponds to the first major deglacial event observed in many climate records (see Fig. 1).](image-url)
Large inputs of fresh water to the North Atlantic may have reduced or halted NADW formation and caused abyssal circulation to stagnate (22). Our deep-sea coral data show that ventilation at 1680 to 1830 m was relatively rapid during meltwater pulse 1a (at about 14.0 ka) and somewhat more sluggish during the period of reduced melting, about 1000 years afterward (Fig. 1A). One possible explanation of this data is that during rapid ice-sheet melting into the North Atlantic, the well-ventilated NADW shoaled. This signature is evident in the radiocarbon age of the 13.7-ka coral. By 12.9 ka, when melting was reduced, the NADW sank to deeper depths. As a result, the coral site was bathed by an older water mass. This scenario also agrees with the glacial nutrient data discussed earlier.

All four of the corals at 15.4 ka have the same 230Th age but they have different radiocarbon ages. The calculated ventilation ages vary by a factor of 2 (legend to Fig. 1). Separate AMS 14C dates for the tops and bottoms of three of these corals show an age reversal in each sample (Table 2). In terms of radiocarbon ages, the biologically younger part of each coral (the top), is older than the biologically older portion (the bottom). We interpret this as a change in the ventilation age of the water during the lifetimes of the corals. From 230Th dates on modern D. cristagalli, we expect that these corals live for no longer than about 160 years (14). The largest 14C age difference, 670 ± 60 years for JFA 24.8, implies a Δ14C difference of about 80 per mil. This is near the range of prebomb Δ14C values in the entire water column of the modern Atlantic Ocean (23).

To check for the large water mass switch implied by these data, we also measured the Cd/Ca ratios in JFA 24.8. The Cd/Ca ratios change by nearly a factor of 2 through the coral (Fig. 2). These data are consistent with the 14C evidence for a large change in circulation in less than 160 years at this site (24). At 15.4 ka North Atlantic sea surface temperatures (SSTs) increased dramatically (Fig. 1C). Bolling-Allerød warming began over Greenland at about 14.6 ka (Fig. 1B). Benthic Cd/Ca values at the Bermuda Rise also increased around 15.4 ka (Fig. 1D). In addition, benthic δ13C values (25) from many North Atlantic sediment cores decreased at the same time, implying that bottom waters became more nutrient rich (10). Our data thus imply that the deep ocean can change during rapid climate events at a rate comparable to that of the atmosphere and the surface ocean. Whether the deep ocean is merely responding to or is actively modulating these rapid climate changes is not clear.

On the basis of a modern calibration of the sensitivity of Cd/Ca in D. cristagalli to the water [Cd], the Cd partition coefficient (D) is about 1.6 (26). Data from benthic foraminifera, which have a D value of 2.9, show a Holocene Cd/Ca value in NADW of 0.07 μmol/mol (Fig. 1). This corresponds to a D. cristagalli ratio of 0.04 μmol/mol (1.6/ 2.9 = 0.27). If we take the value of JFA 24.8 at the top as representative of pure southern source water mass (0.19 μmol/mol), then the bottom of JFA 24.8 (0.11 μmol/mol) grew in a 50:50 mixture of northern and southern water masses. We are justified in this two end-member approach because there is known to be a sharp gradient between these water masses at 2000 m and 60°N during the LGM (11). Apparently, the bottom of this coral grew in the transitional zone of this glacial North Atlantic gradient in nutrients. During the lifetime of the coral, this nutrientline shoaled and as a result the coral was bathed in pure southern source (old) waters. In less than 160 years at 15.4 ka the glacial nutrient gradient between northern and southern source water masses in the North Atlantic shoaled by at least 200 m. At the same time, surface ocean and atmospheric climate show a rapid transition to the warming of the Bolling-Allerød period. The reservoir of heat and mass in the deep ocean is in communication with surface reservoirs on short time scales during rapid climate changes. Cd/Ca ratios in sediment core EN120-GGGC (Fig. 1D) show that the abyssal Atlantic Ocean was in the process of becoming better ventilated when the 1800-m corals show more poorly ventilated waters. Presumably the nutricline at 2000 m shoaled because GNAI/DW became denser than the underlying southern source water (11). However, this is not a firm conclusion because the uncertainty between the two time scales cannot constrain the phasing of events.

The coupled 14C and Cd/Ca data allow us to deconvolve the radiocarbon data and calculate a radiocarbon transit time of this southern source water mass to 40°N. The top to bottom 14C age difference in JFA 24.8 is equal to the radiocarbon age of the southern source water mass (SO) minus a 50:50 mixture of this water mass and the radiocarbon age of the shallow, northern source water mass (NA):

\[
\Delta t = (0.5(\text{SO}) + 0.5(\text{NA})) - 670
\]

The 14C age of the deeper water mass (SO) is expanded into an initial value at its southern source (initial south) and an age-average transit time for the water mass to reach the coral location (t). This location is close enough to the northern source zone that NA is essentially the northern initial value (initial north):

\[
(\text{initial south} - \text{initial north}) + t = 1340
\]

In the modern western Atlantic Ocean the initial ages of the northern and southern waters are about 560 and 1400 years, respectively (27). If we assume the modern initial Atlantic age difference (840 years) between northern and southern waters for the period around 15.4 ka, then the coral data imply an average transit time of 500 years for the southern source water mass. This is significantly longer than today's transit time. However, 15.4 ka may be a time of rapid variation in atmospheric Δ14C (28) and a time of older Pacific deep waters than today (19). Either of these differences from the modern climate could have led to an increase in the age contrast between northern and southern source waters at their formation zones. Consequently, we cannot calculate precisely how much of the 500-year residual is due to more sluggish deep circulation and how much is due to variability in the initial radiocarbon ages at deep-water formation zones.

| Sample | Position | UCID | CAMS | Radiocarbon age (14°C years B.P.) | Age | Age diff. | Length (cm) | Lifetime (years) |
|--------|----------|------|------|---------------------------------|------|----------|------------|-----------------|
| JFA 20.10 | Top | 1529 | 30,183-30,184 | 14,120 ±1σ | 50 | 40 | 140 | 60 | 3.0 | 30-150 |
| | Bottom | 1530 | 30,185-30,186 | 13,980 ±1σ | 50 | 40 | 140 | 60 | 2.8 | 30-140 |
| JFA 24.19 | Top | 1531 | 30,137-30,188 | 14,820 ±1σ | 50 | 40 | 140 | 60 | 3.2 | 30-160 |
| Middle | 2064 | 30189-30,190 | 14,520 ±1σ | 40 | 670 | 60 | 3.2 | 30-160 |
| Bottom | 1532 | 30,189-30,190 | 14,300 ±1σ | 40 | 670 | 60 | 3.2 | 30-160 |
| JFA 24.8 | Top | 1533 | 30,139-30,192 | 14,300 ±1σ | 40 | 670 | 60 | 3.2 | 30-160 |
| Middle | 2065 | 30,189-30,190 | 14,470 ±1σ | 50 | 670 | 60 | 3.2 | 30-160 |
| Bottom | 1534 | 30,193-30,194 | 13,850 ±1σ | 40 | 670 | 60 | 3.2 | 30-160 |

www.sciencemag.org • SCIENCE • VOL. 280 • 1 MAY 1998
Local Orbital Forcing of Antarctic Climate Change During the Last Interglacial
Seong-Joong Kim,* Thomas J. Crowley, Achim Stössel

During the last interglacial, Antarctic climate changed before that of the Northern Hemisphere. Large local changes in precession forcing could have produced this pattern if there were a rectified response in sea ice cover. Results from a coupled sea ice–ocean general circulation model supported this hypothesis when it was tested for three intervals around the last interglacial. Such a mechanism may play an important role in contributing to phase offsets between Northern and Southern Hemisphere climate change for other time intervals.

One of the perplexing problems in Pleistocene climatology involves the factors responsible for Antarctic climate change. Although orbital insolation variations play a major role in driving Pleistocene climate change (1), precessional component of orbital forcing is almost out of phase between the Northern Hemisphere (NH) and Southern Hemisphere (SH), so any conditions favorable for glaciation and deglaciation in the NH should result in the opposite response in the SH. For more than 20 years it has been known that although SH cooling in the Pleistocene accompanied NH glaciation, SH climate led NH climate into and out of the last (and other) interglacials. That is, the SH warmed and cooled before the NH (1). Carbon dioxide also increased before NH glacier retreat (2). Yet standard explanations for SH climate change rarely focus on local forcing changes around Antarctica. Most explanations involve more remote processes such as changes in atmospheric carbon dioxide concentration (3), variations in North Atlantic Deep Water (NADW) heat transport to the Antarctic (4), or lowering of sea level causing expansion of the Antarctic ice sheet. There is, however, a modest (~1°C) contribution from mean annual changes in the local radiation budget at the highest latitudes as a result of synchronous NH-SH obliquity changes at the 41,000-year period (5).

Here, we show that local forcing at the precessional period (19,000 and 23,000 years), which is out of phase between the NH and SH, may also be important in SH climate change. We based our study on the hypothesis that seasonal changes in the Antarctic summer may be proportionately more important than in the Antarctic winter because sea ice is much closer to the freezing point in summer. This relation may allow a