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

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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, Cariaco 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 ^{230}Th and ^{14}C dates and Cd/Ca ratios from deep-sea corals and show that this new archive records a very
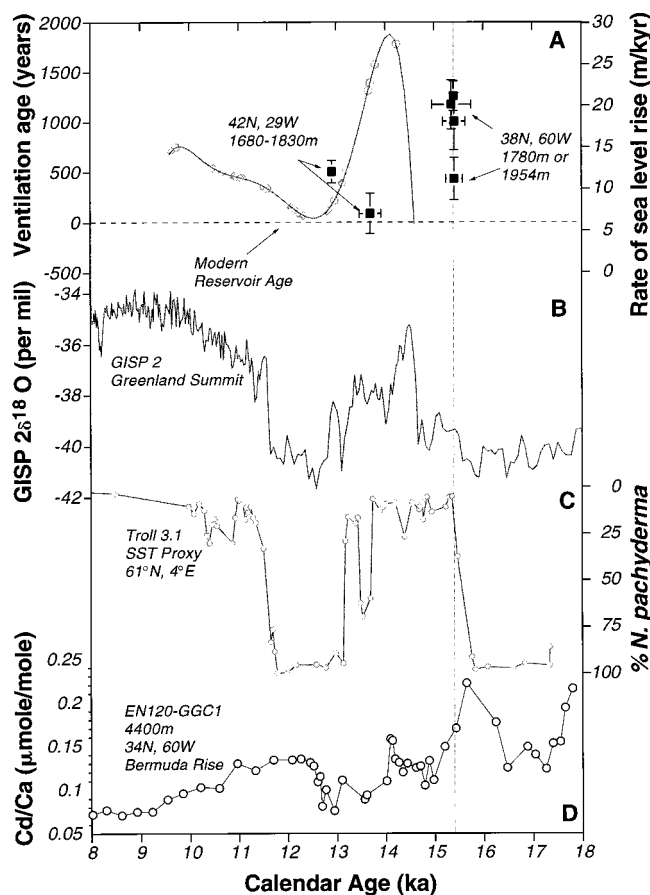
rapid deep circulation change at the beginning of the last deglaciation.

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

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Fig. 1. Deep-sea coral ventilation ages and related climate data. (A) Coral ventilation ages relative to the tropical surface ocean (black squares). Vertical error bars are a combination of analytical errors in the TIMS and AMS measurements as well as uncertainties in the estimates of past atmospheric Δ^{14}C (19). Gray line is the first derivative of a seventh-order polynomial fit to the surface coral sea-level data (16, 21). No attempt was made to account for different depth habitats of particular coral species. Ages of data used to construct the line are shown as gray triangles (16) and circles (21). (B) Oxygen isotopic data from the GISP 2 ice core at 2 m resolution is a proxy for atmospheric temperature at the Greenland Summit (1). (C) Relative abundance of the cold-dwelling planktonic foraminifera N. pachyderma in core Troll 3.1 (29), a proxy for SST. Because the end of the YD cooling occurs during a ^{14}C age plateau, the age model from the original reference has been altered (30) to match the ice core age for the termination of the YD, 11.64 ka (2), and the other control points are left the same (29). (D) Cd/Ca ratios from benthic foraminifera in core EN120-GGC1 at the Bermuda Rise, a proxy for deep ventilation (8, 31). Higher values represent a larger content of nutrient-rich waters, of a southern origin in the Atlantic Ocean, and lower values correspond to a higher content of nutrient-poor waters, of a northern origin in the Atlantic Ocean. The coral data from 15.4 ka correspond to rapid events in all the other tracers from the atmosphere to the deep ocean.

Sample	Depth (m)	Lat.	Long.	Species	^{238}U (ppb)	^{232}Th (ppt)	δ^{234}U (measured)	^{230}Th/^{238}U (activity)	Cal. age (years)	Initial δ^{234}U	^{14}C age (^{14}C years)	Δ^{14}C (‰)	Ventilation (age, years)
JFA 24.8*	1784	38 N	60 W	D. cristagalli	3460 ± 3	1813 ± 51	150.5 ± 1.4	0.15467 ± 0.00081	15,410 ± 170	157.2 ± 1.5	13,570 ± 100	192 ± 26	440 ± 210
JFA 24.19*	1784	38 N	60 W	D. cristagalli	3623 ± 3	1175 ± 44	149.4 ± 1.9	0.15356 ± 0.00056	15,400 ± 110	156.1 ± 2.0	14,270 ± 100	90 ± 19	1260 ± 150
JFA 20.10*	1954	38 N	62 W	D. cristagalli	3943 ± 3	2754 ± 64	148.8 ± 1.2	0.15518 ± 0.00105	15,410 ± 230	155.4 ± 1.3	14,020 ± 110	126 ± 31	1010 ± 280
JFA 24C*	1784	38 N	60 W	D. cristagalli	3858 ± 1	3255 ± 36	144.8 ± 1.3	0.15497 ± 0.00059	15,360 ± 400	151.2 ± 1.4	14,090 ± 60	123 ± 24	1180 ± 250
JFA 2†	1684–1829	42 N	29 W	?	3752 ± 2	1723 ± 26	144.8 ± 1.5	0.13787 ± 0.00068	13,690 ± 220	150.5 ± 1.6	12,120 ± 60	157 ± 20	90 ± 200
JFA 17‡	1684–1829	42 N	29 W	Solenosmilia sp.	3851 ± 1	877 ± 22	147.1 ± 1.3	0.12976 ± 0.00057	12,910 ± 120	152.6 ± 1.3	11,860 ± 60	90 ± 12	510 ± 110

*Desmophyllum cristagalli. †Species not known. ‡Solenosmilia sp.

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 *Desmophyllum cristagalli* and *Solenosmilia* 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 ^{14}C/^{12}C ratio, denoted Δ^{14}C (18), which is also the Δ^{14}C value of the water in which the coral grew. In the modern ocean, we know the initial Δ^{14}C value for deep-water masses and therefore can calculate a radiocarbon deficiency from a deep Δ^{14}C value measured downstream. However, the initial Δ^{14}C of surface waters at deep-water formation zones has changed with time. This initial water Δ^{14}C value is determined largely by the atmospheric Δ^{14}C history, which depends on the production rate of ^{14}C in the upper atmosphere and the exchange of ^{14}C between active carbon reservoirs. To avoid this problem of variable

initial Δ^{14}C, we calculated a ^{14}C projection age to remove the effects of changing atmospheric radiocarbon contents on the measured deep Δ^{14}C for each of our samples (19). This procedure extrapolates back from the deep value along a closed system ^{14}C decay path to the intersection with the known atmospheric Δ^{14}C record. The difference between the age of the intersection and the calendar age of the coral is the ^{14}C projection age. The more conventional ventilation age, measured relative to the sea surface, is calculated by subtracting the surface ocean reservoir age from the ^{14}C projection age of the sample (Fig. 1A). We subtracted the modern North Atlantic reservoir age of 400 years from the ^{14}C 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 Bølling-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 1784 m at 40°N in the western basin of the Atlantic Ocean was ^{238}U/^{230}Th dated to be 15,410 ± 170 years old. Radiocarbon ages from three samples of a single septum show a 670 ± 60 year continuous increase in ^{14}C 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 ^{14}C 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 ^{230}Th 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).
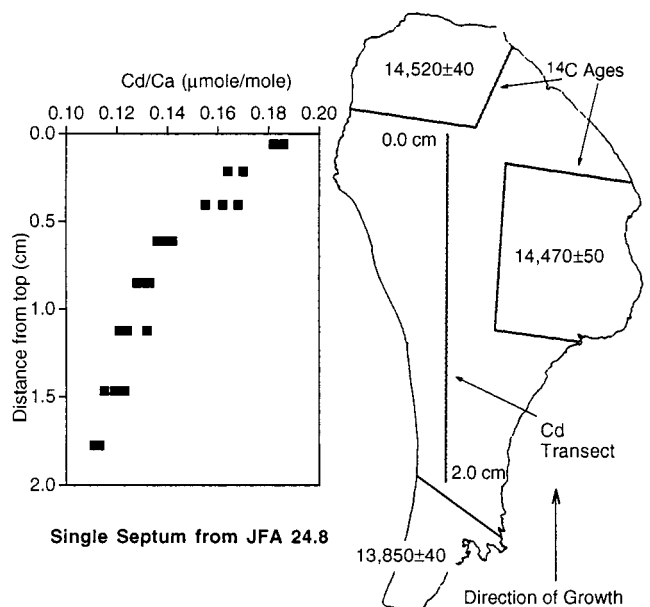
Large inputs of fresh water to the North Atlantic may have reduced or halted NADW formation and caused abyssal circulation to stagnate (*22*). Our deep-sea coral data show that ventilation at 1680 to 1830 m was relatively rapid during meltwater pulse 1a (at about 14.0 ka) and somewhat more sluggish during the period of reduced melting, about 1000 years afterward (Fig. 1A). One possible explanation of these 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 ^{230}Th age but they have different radiocarbon ages. The calculated ventilation ages vary by a factor of 2 (legend to Fig. 1). Separate AMS ^{14}C 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 ^{230}Th dates on modern *D. cristagalli,* we expect that these corals live for no longer than about 160 years (*14*). The largest ^{14}C age difference, 670 ± 60 years for JFA 24.8, implies a Δ^{14}C difference of about 80 per mil. This is nearly the range of prebomb Δ^{14}C 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 ^{14}C 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). Bølling-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 δ^{13}C 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 × 0.07). If we take the value of JFA 24.8 at the top as representative of pure southern source water (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 endmember 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 nutricline 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 Bølling-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-GGC1 (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 ^{14}C 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 ^{14}C 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*):

$$SO - (0.5(SO) + 0.5(NA)) = 670$$

The ^{14}C age of the deeper water mass (*SO*) is expanded into an initial value at its southern source (*initial south*) and an 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*):

$$(initial\ south - initial\ north) + t = 1340$$

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 Δ^{14}C (*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.

Table 2. Radiocarbon dates from the top and bottom of three 15.4-ka *D. cristagalli*. All three corals show an age reversal corresponding to the water mass shift discussed in the text. Samples have different top-to-bottom ages because they grew at slightly different times during the deep circulation event. Lengths are measured from the top and bottom portions of the sample and not the midpoint of the chunk used for AMS dating. Lifetimes are calculated from the growth rate estimates of 0.2 to 1.0 mm per year (*13*). B.P., before present.

Sample	Position	UCID	CAMS	Radiocarbon age (^{14}C years B.P.)				Length (cm)	Lifetime (years)
				Age	±1σ	Age diff.	±1σ		
JFA 20.10	Top	1529	30,183–30,184	14,120	40	140	60	3.0	30–150
	Bottom	1530	30,185–30,186	13,980	40				
JFA 24.19	Top	1531	30,187–30,188	14,500	50	440	60	2.8	30–140
	Middle	2054	36,257	14,500	50				
	Bottom	1532	30,189–30,190	14,060	40				
JFA 24.8	Top	1533	30,191–30,192	14,520	40	670	60	3.2	30–160
	Middle	2055	36,258	14,470	50				
	Bottom	1534	30,193–30,194	13,850	40				

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17. Samples were cleaned of exterior organic carbon and coatings of manganese/iron oxides with successive leaches in a 50:50 mixture of 30% peroxide and 1 M NaOH. A final short leach in a 50:50 mixture of peroxide and 1% perchloric acid was used to remove all remaining organic stains. These precleaning steps removed 15 to 25% of the original coral weight. Immediately before final dissolution and graphitization, corals were soaked in 10% HCl to remove an additional 7 to 30% of aragonite. This procedure overcomes contamination problems from modern carbon that result from storage and handling and organic carbon from polyp and crust material. Samples were carefully chosen to avoid secondary calcification from endolithic activity.
18. The ^{14}C/^{12}C ratio is expressed with the Δ notation, where Δ^{14}C is the per mil deviation from the ^{14}C/^{12}C ratio in 19th-century wood [M. Stuiver and H. A. Polach, Radiocarbon 19, 355 (1977)].
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24. We measured the spatial variation in Cd/Ca ratios from within a single modern D. cristagalli to check for vital effects. These biologically induced nonthermodynamic deviations from equilibrium for a particular tracer have been found to dominate the stable carbon and oxygen isotope signatures of deep-

sea corals (12). A coral from 550 m in the South Pacific Ocean had an average Cd/Ca value of 0.172 μmol/mol with 1 SD of 0.016 μmol/mol. There was no within-coral spatial pattern to this variation. Although D. cristagalli may contain vital effects for Cd/Ca, this 10% SD is smaller than the signal in JFA 24.8 and does not have the top-to-bottom coherence we find in the 15.4-ka sample. For these reasons we believe sample JFA 24.8 has faithfully captured a change in the [Cd] of the past water masses.
25. The carbon isotopic ratio is defined as follows: $\delta^{13}C = [(^{13}C/^{12}C)_{sample}/(^{13}C/^{12}C)_{standard} - 1] \times 1000$, relative to the Pee Dee Belemnite standard [see H. Craig, Geochim. Cosmochim. Acta 3, 53 (1953)].
26. The coral Cd distribution coefficient is the ratio of Cd/Ca in the coral to Cd/Ca in the water. Similar to a core top calibration in foraminifera, we measured Cd/Ca ratios in a suite of modern D. cristagalli that span a range of 0.5 to 3.2 μM [PO$_4$]. Six of these coral fall on a line with a slope of 1.6 coral per water Cd/Ca ratio. This value implies a [PO$_4$] of 3.6 μM for the top of sample JFA 24.8. However, there are three other samples from this calibration that have significantly higher Cd/Ca ratios in the coral than was predicted by the water [Cd]. On the basis of this study, we believe the distribution coefficient for D. cristagalli is 1.6 or higher and that the 3.6 μM [PO$_4$] estimate for JFA 24.8 is a maximum value.
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32. AMS radiocarbon dates were measured by standard procedures at the Lawrence Livermore National Laboratory Center for Accelerator Mass Spectrometry (CAMS). Ventilation ages were calculated with the ^{14}C projection age method described in (19). In three of the corals there is a slight elevation in initial δ^{234}U, indicating a small degree of exchange with a high δ^{234}U reservoir. From the observation that there is no significant correlation between ^{230}Th age and δ^{234}U in different fragments of the same coral (13), we infer that the effect of this exchange on ^{230}Th age is small.
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Local Orbital Forcing of Antarctic Climate Change During the Last Interglacial

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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), the 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 glacial 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

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