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A deep-sea coral record of North Atlantic radiocarbon through the Younger Dryas: Evidence for intermediate water/deepwater reorganization

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[1] Our record of Younger Dryas intermediate-depth seawater $\Delta^{14}C$ from North Atlantic deep-sea corals supports a link between abrupt climate change and intermediate ocean variability. Our data show that northern source intermediate water (~1700 m) was partially replaced by ^{14}C-depleted southern source water at the onset of the event, consistent with a reduction in the rate of North Atlantic Deep Water formation. This transition requires the existence of large, mobile gradients of $\Delta^{14}C$ in the ocean during the Younger Dryas. The $\Delta^{14}C$ water column profile from Keigwin (2004) provides direct evidence for the presence of one such gradient at the beginning of the Younger Dryas (~12.9 ka), with a 100‰ offset between shallow (<~2400 m) and deep water. Our early Younger Dryas data are consistent with this profile and also show a $\Delta^{14}C$ inversion, with 35‰ more enriched water at ~2400 m than at ~1700 m. This feature is probably the result of mixing between relatively well ^{14}C ventilated northern source water and more poorly ^{14}C ventilated southern source intermediate water, which is slightly shallower. Over the rest of the Younger Dryas our intermediate water/deepwater coral $\Delta^{14}C$ data gradually increase, while the atmosphere $\Delta^{14}C$ drops. For a very brief interval at ~12.0 ka and at the end of the Younger Dryas (11.5 ka), intermediate water $\Delta^{14}C$ (~1200 m) approached atmospheric $\Delta^{14}C$. These enriched $\Delta^{14}C$ results suggest an enhanced initial $\Delta^{14}C$ content of the water and demonstrate the presence of large lateral $\Delta^{14}C$ gradients in the intermediate/deep ocean in addition to the sharp vertical shift at ~2500 m. The transient $\Delta^{14}C$ enrichment at ~12.0 ka occurred in the middle of the Younger Dryas and demonstrates that there is at least one time when the intermediate/deep ocean underwent dramatic change but with much smaller effects in other paleoclimatic records.

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1. Introduction

[2] European lake records of the last deglaciation are punctuated by two reappearances of the arctic Dryas flower [*Mangerud et al.*, 1974] that each signal an abrupt cooling in the otherwise generally warming trend of the termination. The Younger Dryas cold event has since been recognized in terrestrial [*Mathewes et al.*, 1993; *Siegenthaler et al.*, 1984], ice [*Alley et al.*, 1993; *Dansgaard et al.*, 1982] and marine records [*Broecker et al.*, 1989; *Lehmann and Keigwin*, 1992] across the Northern Hemisphere in a wide variety of tracers. As recorded in the GISP2 ice core, the Younger Dryas is a ~1300 yearlong abrupt return to cold temperatures and low accumulation rate conditions from 12.9 to 11.6 ka that is the last of a series of glacial era rapid climate shifts (Figure 1a) [*Alley et al.*, 1993; *Dansgaard et al.*, 1993; *Grootes et al.*, 1993]. This event is unique among the many millennial-scale Dansgaard-Oeshger (DO) oscillations and Heinrich events that punctuate the glacial period because it occurred during the glacial termination. However, the Younger Dryas's age means that radiocarbon can be used to understand the causes of abrupt shifts in the climate system in ways that are unavailable to most of the previous glacial period.

[3] In particular, radiocarbon measurements from the deep ocean can provide an important test of one leading theory for the cause of the Younger Dryas. Variations in the North Atlantic ventilation rate will both alter the poleward heat transport associated with North Atlantic Deep Water (NADW) formation and significantly change the ^{14}C content at depth. According to the "salt oscillator" theory, Atlantic salinity is modulated by both ice sheet formation/melting and the export of salt out of the basin by NADW [*Broecker et al.*, 1990a]. When Atlantic salinity is reduced, the surface density in the high-latitude north becomes insufficient for surface water to sink, thus turning "off" North Atlantic Deep Water (NADW) formation. As an

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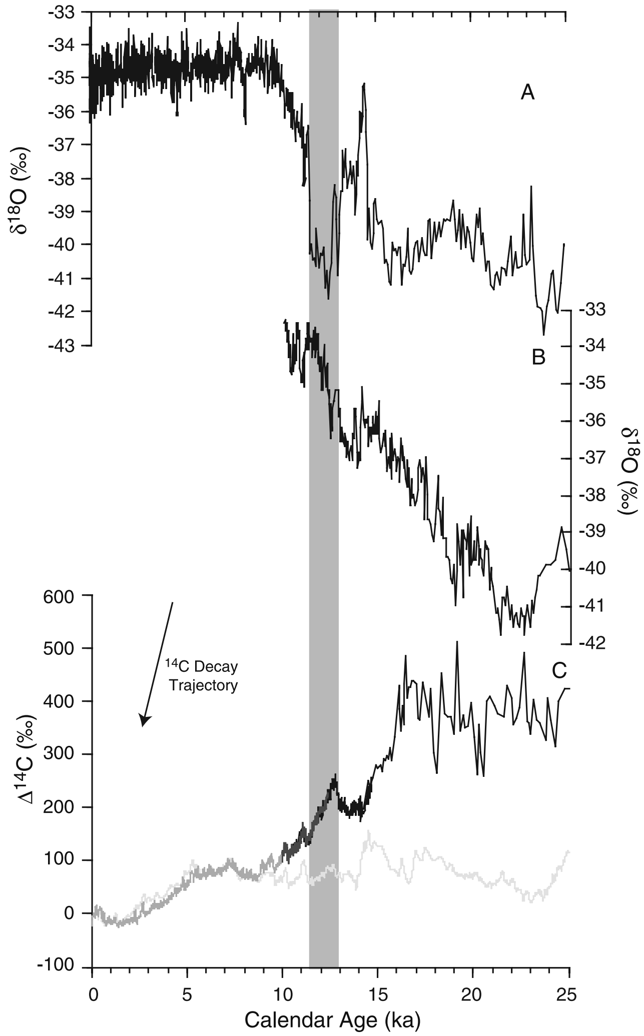

Figure 1. A 25 kyr record of climate from (a) GISP2 δ^{18}O [*Grootes et al.*, 1993] shows dramatic cooling in Greenland during the Younger Dryas that is not seen in the Antarctic record of (b) Byrd δ^{18}O [*Blunier and Brook*, 2001]. A comparison between (c) the atmospheric Δ^{14}C record compiled from the tree ring record (dark shaded curve) [*Stuiver et al.*, 1998; *Friedrich et al.*, 2004] and the varved sediments of the Cariaco Basin (solid curve) [*Hughen et al.*, 2000, 2004a, 2004b], and the ^{10}Be-based Δ^{14}C reconstruction (lower estimate, light shaded curve) [*Muscheler et al.*, 2004] demonstrates that the atmospheric Δ^{14}C changes during the Younger Dryas are larger than Δ^{14}C changes caused by changes in the ^{14}C production rate alone. The Younger Dryas is highlighted in the records as a vertical shaded bar.

extension to the salt-oscillator hypothesis, the "bipolar seesaw", accounts for the asynchronous connection between the Arctic and Antarctic ice core records of temperature [*Blunier and Brook*, 2001; *Blunier et al.*, 1998; *Broecker*, 1998; *Sowers and Bender*, 1995]. In this case, the density gradient between sinking regions in the south and in the north swings back and forth with NADW "on" conditions cooling the Southern Hemisphere by drawing heat from the south to the north and NADW "off" conditions leading to the rapid coolings seen in the Greenland ice cores.

[4] These theories are crucially dependent on the flux of deep water formed in the North Atlantic, yet most of our deep ocean tracers do not contain an intrinsic measure of rate. Nutrient tracers such as δ^{13}C and Cd/Ca allow for an estimate of the relative proportions of deep source waters. A record of deep (4450 m) Atlantic Cd/Ca measured in benthic foraminifera from the Bermuda Rise indicates that deepwater nutrients increased during the Younger Dryas reflecting an increased southern source influence [*Boyle and Keigwin*, 1987]. At the same time, intermediate water (965 m) nutrients from the Bahama Banks declined reflecting an increased contribution from northern source water [*Came et al.*, 2003; *Marchitto et al.*, 1998; *Rickaby and Elderfield*, 2005]. The evidence suggests that at the start of the Younger Dryas, NADW shoaled and was replaced by deep water from a southern source. However, the volumetric reduction of northern source water at the beginning of the Younger Dryas does not necessarily mean that its flux was reduced. A more direct estimate of overturning rate through the Younger Dryas comes from (^{231}Pa/^{230}Th) ratios in deep-sea sediments [*McManus et al.*, 2004]. This record implies that while the overturning rate of the North Atlantic was lower during the Younger Dryas as compared to today, it was not nearly as reduced as during Heinrich 1.

[5] In the modern ocean we estimate the overturning rate of the deep ocean by measuring the ^{14}C content of dissolved inorganic carbon [*Broecker and Peng*, 1982; *Stuiver et al.*, 1983]. Four factors largely determine this number. The Δ^{14}C of the atmosphere when the water was last at the surface and the surface/atmosphere offset (reservoir age) set the initial ^{14}C concentration of newly formed deep water. After leaving the surface, mixing with other water masses and in situ aging will then cause Δ^{14}C to evolve with time. To calculate deep ocean ventilation rates from Δ^{14}C measurements, we need to isolate this in situ aging component. With our modern understanding of end-member Δ^{14}C values and measurements of any other conservative mixing tracer (like temperature and salinity), the radiocarbon ventilation age is just the Δ^{14}C deficit relative to conservative mixing. Ideally, we would use this approach for the past ocean as well [*Adkins and Boyle*, 1999]. However, three problems complicate paleoradiocarbon interpretation; ^{14}C is normally our chronometer and therefore cannot also be a water mass tracer, mixing calculations are complicated by nonconservative behavior of the tracers, and water mass end-member variability is sometimes poorly constrained.

[6] Several methods have been used to overcome the chronometer problem. High-resolution, independent stratigraphy itself can be used to calculate the past Δ^{14}C of the atmosphere [*Hughen et al.*, 2000, 2004a] and surface ocean [*Shackleton et al.*, 2004; *Siani et al.*, 2001; *Waelbroeck et al.*, 2001] as long as it is tied back to a calendar age scale. In addition, comparison of benthic and planktonic foraminiferal radiocarbon ages from the same time horizon in a sediment core, provides an estimate of the ventilation age of the deep sea by measuring the vertical age

gradient in the past without having to know the exact calendar age of the samples [*Broecker et al.*, 1990b; *Duplessy et al.*, 1989; *Shackleton et al.*, 1988]. *Keigwin* [2004] examined benthic/planktonic foraminifera pairs from a suite of North Atlantic sediment cores and demonstrated that the ^{14}C profile during the early part of the Younger Dryas consisted of ^{14}C-depleted water beneath ^{14}C enriched water with a transition between the two at ~2400 m. This implies that well ventilated water from the north did not penetrate below this front. *Skinner and Shackleton* [2004] generated a Δ^{14}C time series at 3000 m in the northeast Atlantic using a correlation between their measured planktonic foraminiferal δ^{18}O record and that of Greenland ice to estimate an independent calendar age for each time horizon. The data point that falls within the Younger Dryas interval indicates that deep water was radiocarbon-depleted compared to the data point ~200 years before.

[7] Another solution to the chronometer problem is to use a second radioactive clock to account for the radiocarbon decay since the organism grew. Modern deep-sea corals accurately record the Δ^{14}C of dissolved inorganic carbon [*Adkins et al.*, 2002] and fossil samples can be precisely dated using U-Th techniques [*Cheng et al.*, 2000]. Two timescales of Δ^{14}C history are available in the deep-sea coral archive. A time series with resolution similar to a sediment core can be constructed by comparing results from different coral specimens. In this case, the time span of interest is bounded only by the calendar age distribution of the samples collected. In addition, finely spaced measurements within individual corals span very brief (~100 years) time intervals with ~10 year resolution [see *Adkins et al.*, 1998, 2004; *Robinson et al.*, 2005]. This resolution is similar to that of ice cores and is ultimately constrained by the growth pattern of the coral.

[8] Five previous studies have used coupled U-Th and ^{14}C ages in deep-sea corals to determine the Δ^{14}C of past seawater [*Adkins et al.*, 1998; *Frank et al.*, 2004; *Goldstein et al.*, 2001; *Mangini et al.*, 1998; *Schroder-Ritzrau et al.*, 2003]. *Adkins et al.* [1998] demonstrated that western North Atlantic intermediate/deep water Δ^{14}C decreased significantly (by ~70‰) between 13.7 and 12.9 ka. *Schroder-Ritzrau et al.* [2003] found a similar decrease in Δ^{14}C between 13.9 and 13.0 ka in the eastern North Atlantic, though the shallow depth (240 m) and proximity to the coast suggests that these samples are not representative of the deep sea. Their other corals from Younger Dryas intermediate water show atmosphere/ocean Δ^{14}C offsets similar to the modern, with the exception of one sample from 11.4 ka that has a larger depletion relative to the atmosphere. *Frank et al.* [2004] show that by 10.2 ka the Δ^{14}C offset between the atmosphere and intermediate ocean (~730 m) was similar to that observed in a modern coral. Here we add to the growing body of deep-sea coral data and measure Δ^{14}C in North Atlantic samples to investigate changes in deep-water ventilation and organization over the Younger Dryas cold period.

[9] One other aspect of radiocarbon during the Younger Dryas is important to our study. Using planktonic foraminifera from the Cariaco basin and an age model that is tied to the GISP2 isotope record, *Hughen et al.* [2000] have documented the Δ^{14}C of the surface waters through the Younger Dryas. By assuming there is a constant ~400 year offset between the local surface waters and the atmosphere, we can use the Cariaco record as a proxy for $\Delta^{14}C_{atm}$. In an indirect, but sensitive, way this record reflects the mean overturning rate of the deep ocean. The inventory of atmospheric ^{14}C is set by the balance of inputs from cosmic ray production and outputs due to both the in situ radioactive decay of ^{14}C and the carbon exchange with other reservoirs (equation (1)).

$$\frac{d^{14}C_{atm}}{dt} = Production - \lambda^{14}C_{atm} - Ocean\ Exchange \quad (1)$$

Over centennial and millennial timescales, this balance is dominated by two terms, the production rate and the rate of ^{14}C uptake by the oceans. Therefore trends in the record of $\Delta^{14}C_{atm}$ can be compared with those of production and $\Delta^{14}C_{deep\ ocean}$ with one important caveat: production rate changes will be felt for a longer time in the whole ^{14}C system than variations in the ocean exchange term because production rate variations alter the inventory of ^{14}C atoms, while the ocean term only reorganizes the existing ^{14}C atoms between reservoirs [*Muscheler et al.*, 2004].

[10] During much of the Younger Dryas the ^{14}C production rate was balanced by the atmospheric loss terms giving rise to an "age plateau" in many sedimentary records. However, at the initiation of this period the Cariaco basin record of $\Delta^{14}C_{atm}$ shows a 70‰ rise over ~200 years starting at 13.0 ka [*Hughen et al.*, 2000] (Figure 1c). With a roughly constant radiocarbon production rate [*Muscheler et al.*, 2004], the observed peak in Younger Dryas $\Delta^{14}C_{atm}$ is well above that expected from production alone. Since decay in the deep ocean is the largest sink for radiocarbon, and North Atlantic Deep Water (NADW) formation is the primary mode of ^{14}C transport to the deep reservoir in the modern ocean [*Broecker and Peng*, 1982], the initial sharp peak in Younger Dryas $\Delta^{14}C_{atm}$ implies a decrease in the ocean uptake, specifically a reduction in the rate of NADW formation, that persisted for ~200 years. The subsequent decline in $\Delta^{14}C_{atm}$ is consistent with a reinvigoration of NADW formation or the activation of another ^{14}C sink that brings the ^{14}C system back toward steady state with atmospheric production. In this paper we present new measurements of the deep ocean Δ^{14}C in the North Atlantic and discuss them as a complement to the detailed record of $\Delta^{14}C_{atm}$ from the Cariaco Basin.

2. Samples and Methods

[11] We routinely screen new fossil deep-sea coral samples for their calendar age. Previously we have used a relatively imprecise, but high throughput, quadrupole ICP-MS technique [*Adkins and Boyle*, 1999]. With the advent of multicollector magnetic sector ICP-MS we have switched to precisely dating every sample [*Robinson et al.*, 2005]. We selected 7 North Atlantic *Desmophyllum dianthus* (Esper, 1794) corals with U-Th calendar ages that fall within the Younger Dryas (13.0 to 11.5 ka) from our larger sample pool. Our samples are from the Smithsonian invertebrate collection (1 sample) and from a DSV *Alvin*

Table 1. *D. cristagalli* Sample Locations

Sample	Coral Identification	Collection Site	Latitude, N	Longitude, W	Depth, m
YD-1	ALV-3891-1459-003-002	Gregg Seamount	38°56.9′	61°1.6′	1176
YD-2	ALV-3891-1758-006-003	Gregg Seamount	38°56.9′	61°1.7′	1222
YD-3	Smithsonian 48735.1	Azores	37°57.5′	25°33.0′	1069–1235
YD-4	ALV-3890-1407-003-001	Manning Seamount	38°13.6′	60°27.6′	1778
YD-5	ALV-3887-1549-004-012	Muir Seamount	33°45.15′	62°35.3′	2372
YD-6	ALV-3887-1549-004-007	Muir Seamount	33°45.15′	62°35.3′	2372
YD-7	ALV-3887-1549-004-009	Muir Seamount	33°45.15′	62°35.3′	2372

cruise to the New England seamounts in May–June 2003 (6 samples) (Table 1).

2.1. Reconstructing Δ^{14}C

[12] To reconstruct Δ^{14}C in the past ocean we measure the conventional ^{14}C age of the coral and use the measured U-Th calendar age to account for closed system radioactive decay since the time of aragonite precipitation according to the expression:

$$\Delta^{14}C = \left(\frac{e^{-\frac{^{14}C\ Age}{Libby\ Mean\ Life}}}{e^{-\frac{U/Th\ Cal\ Age}{True\ Mean\ Life}}} - 1 \right) \times 1000‰$$

where the Libby Mean Life is 8033 years and the True ^{14}C Mean Life is 8267 years [*Stuiver and Polach*, 1977]. Conventional ^{14}C ages are δ^{13}C normalized to account for isotopic fractionation and Δ^{14}C is a measure of the relative difference between this normalized ^{14}C/^{12}C ratio and a standard [*Stuiver and Polach*, 1977].

2.2. U-Th Calendar Ages

[13] U-Th calendar ages were determined for a top portion (~1 g) from each coral. Because the calendar age error is comparable to the lifetime of each coral, only one calendar age measurement was necessary for each coral. Calendar ages for samples closer to the base of the coral were estimated by assuming a 1 mm/yr vertical extension rate [*Adkins et al.*, 2004; *Cheng et al.*, 2000]. Smithsonian sample 48735.1 was U-Th dated by TIMS [*Cheng et al.*, 2000], and the New England Seamount samples were U-Th dated by MC-ICPMS [*Robinson et al.*, 2005].

2.3. Conventional Radiocarbon Ages

[14] To measure a ^{14}C age, a thecal section composed of portions of a S1 septum and the adjacent smaller septa (2–3 mm thick) was cut out of each coral using a small diamond tipped saw attached to a Dremel rotary tool (Figure 2). Visible contamination on the coral surface was mechanically abraded away with the saw, and any holes formed by endolithic deep-sea organisms were milled out with a drill bit. Each thecal section was cut transversely into pieces (14–50 mg each) that were cleaned and leached (>40% mass removal in final leach just prior to graphitization) by the procedure of *Adkins et al.* [2002]. The resulting 10 mg pieces were hydrolyzed in phosphoric acid, and the evolved CO_2 was graphitized under H_2 on an iron catalyst before ^{14}C analysis [*Vogel et al.*, 1984]. Radiocarbon ages were measured at the Lawrence Livermore National Laboratory Center for Accelerator Mass Spectrometry (sample YD-3) and at the University of California, Irvine Keck Carbon Cycle Accelerator Mass Spectrometry (UCI-KCCAMS) Laboratory (all other samples).

3. Results

[15] Our ~2000 yearlong Δ^{14}C time series consists of measurements from 7 individual coral skeletons with a sequence of 3 to 7 ^{14}C measurements along each coral transect. U-series and ^{14}C results are summarized in Tables 2 and 3, respectively. The corals fall into two categories: those that contain large within-coral variation in their Δ^{14}C values (YD-3,4) and those with essentially constant Δ^{14}C over the entire skeletal transect (YD-1,2,5,6,7) (Figure 3). Interpreted as a Δ^{14}C record of the seawater that bathed these corals, our data show that intermediate water (<2000 m) Δ^{14}C increased by ~10–20‰ through the Younger Dryas and exhibited a transient enrichment, of magnitude ~40–50‰, in the middle of the Younger Dryas (~12 ka). Because of their uniformity in Δ^{14}C, the data within each of the low-variability Δ^{14}C corals have been averaged together in the plots that follow.

[16] Contamination with modern carbon, an issue for all corals [*Chiu et al.*, 2005], was especially problematic for coral YD-4 from the New England Seamounts (12.2 ka). A slight stain persisted on sample YD-4b after acid leaching and the Δ^{14}C result for this sample was elevated with respect to samples YD-4a and c (Figure 3b, shaded squares). If this contamination were composed of modern $CaCO_3$ or contained adsorbed CO_2, the contamination, and not a change in the environmental conditions, could conceivably cause the Δ^{14}C enrichment. Sample YD-4b would have to contain 1% modern $CaCO_3$ to cause the ~30‰ Δ^{14}C enrichment. The leaching experiment of *Adkins et al.* [2002] showed that an acid leach resulting in 5–10% sample loss was sufficient to remove any significant contaminating carbon. In the case of sample 4b, 43% of the sample mass was leached away, so it is unlikely that an exterior coating of $CaCO_3$ or adsorbed CO_2 significantly above background levels persisted. In this case, the stain composed far less than 1% of the sample, and since the stain most likely contained organic carbon, which is not oxidized in acidic solution, it is again unlikely to be the cause of the measured ^{14}C enrichment. For macroporous surface corals it is possible to overleach samples that have secondary calcite overgrowths [*Chiu et al.*, 2005]. As our corals have a nonporous morphology and the YD is not old enough for this process to greatly alter our ages, we do not consider any of our signals to be analytical artifacts. Furthermore, given

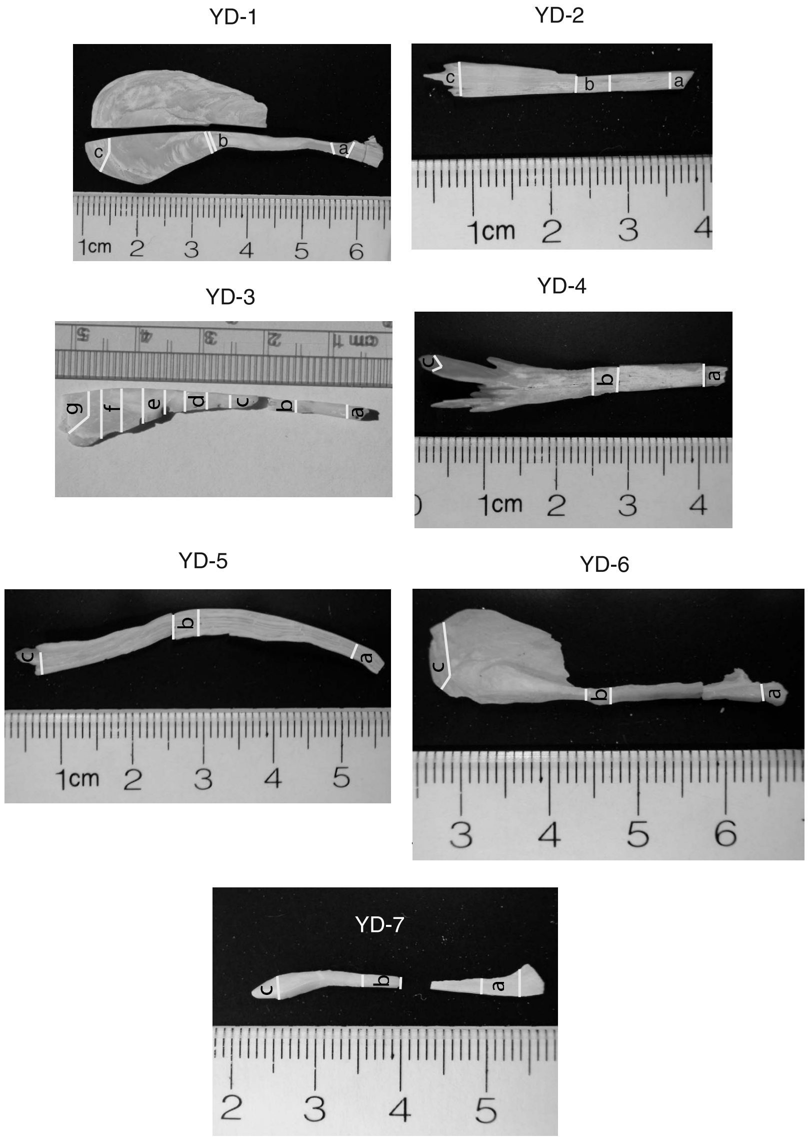

Figure 2. *D. dianthus* deep-sea coral sections sampled for ^{14}C ages. Samples are marked with their corresponding sample numbers.

that one other coral also shows elevated $\Delta^{14}C$ concurrently, we believe that the environmental signal in YD-4b is robust.

[17] Calcite blanks contain less ^{14}C than samples from a radiocarbon dead (>50 ka) deep-sea coral samples (Figure 4). The long-term fraction modern averages (measured at UCI-KCCAMS) for our calcite blanks and a 240 ka deep-sea coral are 0.0012 ± 0.0005 and 0.0039 ± 0.0018 (2σ), respectively. For all of the data reported here, we have

Table 2. *D. dianthus* U/Th Calendar Ages[a]

Sample	Coral Identification	^{238}U (2σ), ppm	^{232}Th (2σ), ppb	$d^{234}U_{Measured}$ (2σ), ‰	$^{230}Th/^{238}U$ (2σ)	U/Th Calendar Age, years B.P.	U/Th 2σ, years B.P.	$d^{234}U_{Initial}$, (2σ), ‰
YD-1	ALV-3891-1459-003-002	4.496 (0.003)	0.879 (0.006)	139.6 (1.1)	0.1141 (0.0005)	11,330	120	144.2 (1.1)
YD-2	ALV-3891-1758-006-003	3.665 (0.003)	0.797 (0.008)	140.7 (1.1)	0.1154 (0.0006)	11,450	130	145.4 (1.1)
YD-3	Smithsonian 48735.1	3.554 (0.003)	0.330 (0.006)	145.6 (1.3)	0.1200 (0.0010)	11,960	120	150.6 (1.3)
YD-4	ALV-3890-1407-003-001	3.361 (0.002)	1.889 (0.008)	144.8 (1.1)	0.1244 (0.0006)	12,220	300	149.9 (1.1)
YD-5	ALV-3887-1549-004-012	3.286 (0.003)	0.561 (0.012)	142.5 (1.1)	0.1259 (0.0006)	12,590	110	147.7 (1.1)
YD-6	ALV-3887-1549-004-007	4.039 (0.003)	0.584 (0.008)	139.6 (1.1)	0.1266 (0.0006)	12,700	100	144.7 (1.1)
YD-7	ALV-3887-1549-004-009	3.360 (0.002)	0.159 (0.007)	143.3 (1.1)	0.1266 (0.0006)	12,700	70	148.5 (1.2)

[a]Calendar ages are in years before the date of U-series measurement.

Table 3. *D. dianthus* Radiocarbon Ages and $\Delta^{14}C_{water}$[a]

Sample	Laboratory Identification	Sample Span From Coral Base, mm	^{14}C Age, ^{14}C years	Error (2σ), ^{14}C years	$D^{14}C_{Water}$, ‰	Propagated Error (2σ) From ^{14}C Age, ‰	Propagated Error (2σ) From Cal Age, ‰	Average $D^{14}C_{Water}$, ‰	Error ($2\sigma_{mean}$), ‰
					ALV-3891-1459-003-002				
YD-1									
a	4722	5.8–8.3	10070	70	131	9	16	137	9
b	4726	29.7–30.6	9940	60	145	7			
c	4709	49.7–54.2	10000	50	134	6			
					ALV-3891-1758-006-003				
YD-2									
a	4718	0.0–3.3	10060	70	148	9	18	151	8
b	4719	11.1–15.8	9970	60	158	7			
c	4711	30.5–35.1	10040	50	146	6			
					Smithsonian 48735.1				
YD-3									
a	45610	0.0–3.5	10780	80	116	11	17	137	20
b	45539	11.1–15.6	10420	60	166	9			
c	45535	17.3–21.6	10500	80	153	11			
d	45540	24.9–28.2	10370	60	172	9			
e	45541	31.3–33.7	10660	70	129	10			
f	45538	36.3–38.3	10750	70	116	9			
g	45536	41.8–45.6	10800	70	108	10			
					ALV-3890-1407-003-001				
YD-4									
a	4715	0.0–2.9	11010	60	119	7	42	126	23
b	4710	14.9–18.9	10780	60	149	7			
c	4720	39.7–42.6	11030	60	111	7			
					ALV-3887-1549-004-012				
YD-5									
a	4713	0.0–4.8	11380	60	119	7	15	116	5
b	4723	27.8–31.8	11410	70	111	8			
c	4716	50.4–54.4	11340	70	118	8			
					ALV-3887-1549-004-007				
YD-6									
a	4712	0.0–3.0	11500	60	116	7	13	121	10
b	4727	20.2–22.7	11370	70	131	8			
c	4717	39.7–41.9	11460	60	116	7			
					ALV-3887-1549-004-009				
YD-7									
a	4725	3.4–8.2	11470	70	119	8	9	126	10
b	4708	16.6–20.9	11340	60	136	7			
c	4721	31.3–34.2	11420	80	122	10			

[a]Calendar ages have been converted to years before 1950.

adjusted the measured fraction modern using the larger blank associated with the 240 ka coral and its corresponding larger uncertainty. Replacing the deep-sea coral blank with the calcite blank would give a $\Delta^{14}C$ that is ~10‰ more enriched than we report in this paper. The uncertainty in the deep-sea coral ^{14}C background defines the detection limit for our deep-sea coral ^{14}C ages (~45 ka). This uncertainty also limits the precision of our measured past $\Delta^{14}C$ values. In Figure 5 we propagate the two blank uncertainties (calcite and coral) through the $\Delta^{14}C$ calculation over a range of calendar age errors and find that for the 10–12 ka samples in this study, our $\Delta^{14}C$ errors are primarily governed by the calendar age uncertainty (1%). For older samples, however, more precise background measurements will be needed to produce a meaningful $\Delta^{14}C$ reconstruction.

4. Discussion

[18] From our deep-sea coral data set we have compiled a time series of $\Delta^{14}C$, at essentially one location (~39°N in the western Atlantic) and several different depths that spans the Younger Dryas interval. While conservative or passive tracer data would be very helpful, we do not have any new constraints on the mixing ratios of separate end-member waters masses for our new radiocarbon data. However, several existing radiocarbon data sets from the deep ocean and the atmosphere let us place constraints on

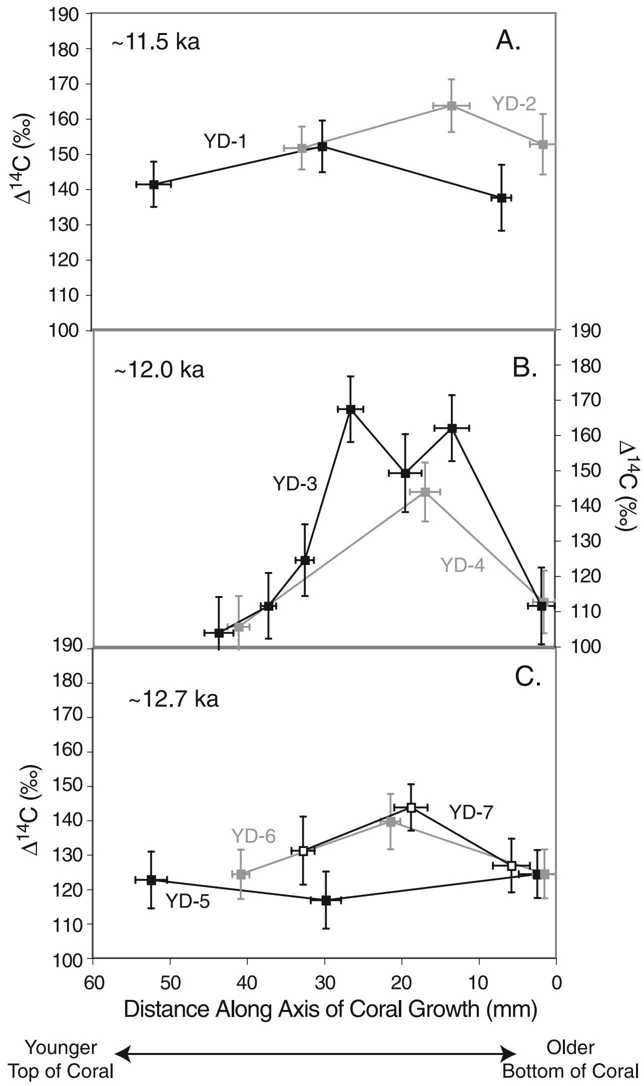

Figure 3. *D. dianthus* Younger Dryas Δ^{14}C results for individual coral transects. These ~100 year long time series show significant variability only at ~12.0 ka in the middle of the Younger Dryas. The other corals at 11.5 and 12.7 ka show no significant variation over their lifetimes.

the role of deep circulation changes in the climate shifts of this time. As described in section 1, the $\Delta^{14}C_{atm}$ record of *Hughen et al.* [2000, 2004a, 2004b] is the complement to our intermediate water/deepwater record. In addition, the Δ^{14}C profile for the North Atlantic from *Keigwin* [2004] is an important reference point for the Δ^{14}C of the waters deeper than our coral samples. The deep Δ^{14}C record of *Robinson et al.* [2005] presents some new benthic-planktonic radiocarbon pairs for the waters below 3000 m and before the start of the Younger Dryas that indicate the deep portion of Keigwin's record was established before 13.0 ka. So, we assume that the "older" Southern Source waters below ~2500 m were in place at or before the start of the Younger Dryas and that our new data set constrains the behavior of the waters above this depth.

[19] The new deep-sea coral data are shown in Figure 6 along with 2 data points from *Adkins et al.* [1998], the record of atmospheric Δ^{14}C from the Cariaco Basin [*Hughen et al.*, 2000, 2004b], and the GISP2 ^{10}Be-based Δ^{14}C reconstruction [*Muscheler et al.*, 2004]. Over the beginning of the Younger Dryas, the ocean Δ^{14}C record at ~1700 m in the North Atlantic is consistent with the inverse of the atmospheric Δ^{14}C record. From 13.0 to 12.8 ka, atmospheric Δ^{14}C rose steeply, while intermediate water/deepwater Δ^{14}C dropped by ~70‰ over less than ~800 years (though we believe the drop was probably much shorter than this, see below). If ocean circulation and air-sea exchange processes were unchanged over this time period (14–11 kyr B.P.), the Δ^{14}C of the deep water would follow the $\Delta^{14}C_{atm}$. Instead, the observed drop in ocean Δ^{14}C (and the rise in $\Delta^{14}C_{atm}$) is evidence that deep-ocean ^{14}C exchange was reduced, probably because of a decrease in the rate of NADW formation and subsequent invasion of ^{14}C-depleted southern source water. After this initial drop, the intermediate water data show a slow ramp up of Δ^{14}C mirroring the gradual decrease of the atmospheric record.

[20] If the Younger Dryas was initiated by a cessation of deepwater formation at 13.0 ka, as implied by the $\Delta^{14}C_{atm}$ and the GISP2 δ^{18}O records, two possible end-member states exist for the intermediate/deep water at our site. The water may stagnate, or it may be replaced by water from another source. To distinguish between the two, we

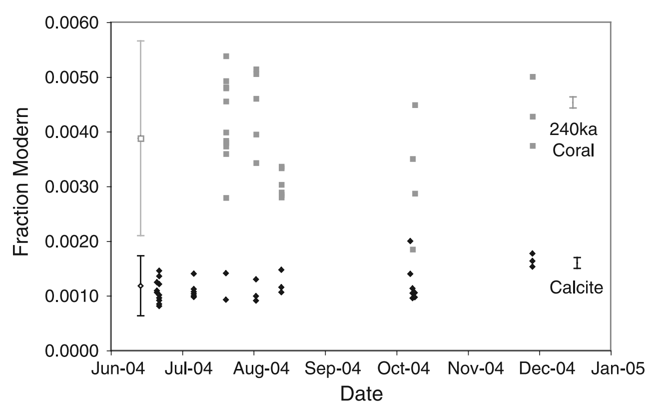

Figure 4. Deep-sea coral blanks (shaded squares) consistently contain more ^{14}C than calcite blanks (solid diamonds). These blank measurements demonstrate that our oldest coral contains some amount of refractory ^{14}C that cannot be cleaned away. The average fraction moderns (measured at University of California, Irvine Keck Carbon Cycle Accelerator Mass Spectrometry) are 0.0012 ± 0.0005 and 0.0039 ± 0.0018 (2σ) for our calcite blanks and a 240 ka deep-sea coral, respectively (open symbols on the left). Analytical uncertainty for each measurement is given by the error bars on the right. To account for this refractory blank, the result from the 240 ka coral is used to blank correct our sample results.

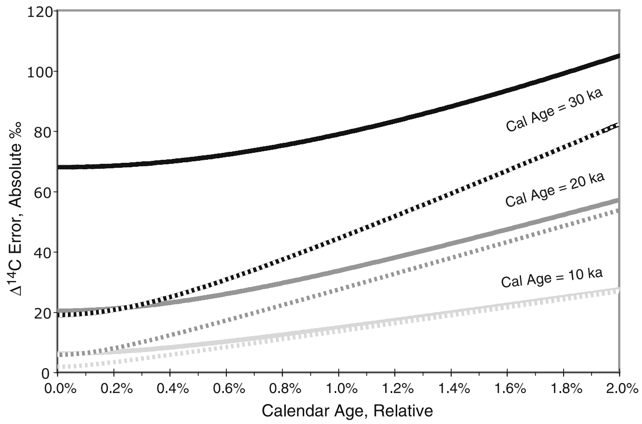

Figure 5. Propagated error curves over a range of calendar age uncertainty for a $\Delta^{14}C_{water}$ arbitrarily picked to be 100‰ plotted for three calendar ages (10, 20, and 30 ka). Solid lines have a radiocarbon blank uncertainty based on the measured 240 ka coral data. Dotted lines have a blank uncertainty based on the measured calcite data shown in Figure 4. For the 10–12 ka samples in this study our $\Delta^{14}C$ errors are primarily controlled by calendar age uncertainty. For older samples, however, more precise background measurements will be needed to reduce the propagated error.

assume that the $\Delta^{14}C$ of JFA2 (160 ± 30‰ at 13.7 ka) is a value representative of pre-Younger Dryas conditions right up to the cutoff at 13.0 ka. If the intermediate water simply stagnated, it would evolve from its initial value at 13.0 ka along the $\Delta^{14}C$ decay trajectory noted in Figure 6. Decay from 160‰ at 13.0 ka to our next data point at 12.9 ka (JFA17) would produce only a ~13‰ depletion (to ~147‰) compared to the observed 70‰ depletion (to ~90‰). Younger Dryas deepwater formation would have had to cease completely at 13.2–13.4 ka, before any observable change in GISP2 $\delta^{18}O$ or $\Delta^{14}C_{atm}$, in order to account for the measured depletion by stagnation alone. Since stagnation can account for only a minor fraction (~20%) of the observed depletion, a rearrangement of water masses, must account for at least ~80% of the depletion. The movement of $\Delta^{14}C$-depleted southern source water into this region at the expense of northern source water would result in a rapid shift of the $\Delta^{14}C$ signature to a more depleted value. The magnitude and speed of the transition depend on the size of the gradient that exists in the water and the rate of the water mass reorganization. The size of this transition is large, especially when considering the entire range for the deep Atlantic today is ~85‰, and implies that large, mobile $\Delta^{14}C$ gradients existed in the Younger Dryas intermediate/deep Atlantic. More samples in the period between 14.0 and 12.7 ka would constrain the phasing of intermediate water $\Delta^{14}C$ change and the start of the Younger Dryas that we assume here.

[21] Two samples at the onset of the Younger Dryas, separated by 210 calendar years and ~600 m depth, illustrate these gradients in a vertical water column profile that we compare to a modern profile from the Atlantic expedition of GEOSECS [*Stuiver and Ostlund*, 1980] (Figure 7). Together with the benthic/planktonic foraminiferal $\Delta^{14}C$ profile from *Keigwin* [2004], we see that the early Younger Dryas profile is higher in absolute value and spans a much larger $\Delta^{14}C$ range (~300‰ range from shallow to deep) than the modern profile (~15‰ range from shallow to deep). The deep-sea coral $\Delta^{14}C$ profile also highlights the water column structure

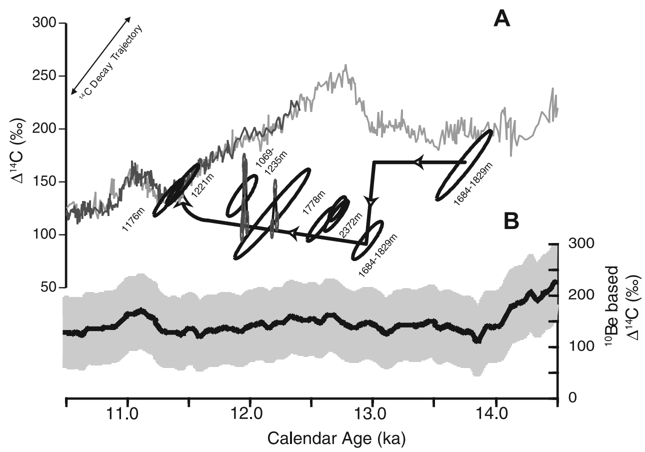

Figure 6. (a) *D. dianthus* radiocarbon results plotted with atmospheric $\Delta^{14}C$ [*Friedrich et al.*, 2004; *Hughen et al.*, 2004b]. (b) The ^{10}Be-based $\Delta^{14}C$ reconstruction from the Greenland ice cores and a simple ^{14}C model of the modern ocean circulation [*Muscheler et al.*, 2004]. Coral collection depths and the trajectory of projected closed system ^{14}C decay are plotted with the data. Error estimates (2σ) for the coral data are ellipses because the calculated $\Delta^{14}C$ is itself dependent on the calendar age (the x axis). When $\Delta^{14}C$ errors are dominated by calendar age uncertainty, the major axis of the ellipse is elongated and tends toward an angle equal to the rate of ^{14}C decay. When the ^{14}C age uncertainty dominates the overall $\Delta^{14}C$ errors, the major axis tends toward the vertical. For all points, except the corals with variable $\Delta^{14}C$ at 12.0 ka, the error ellipses are based on the weighted 2σ standard errors of the triplicate ^{14}C age measurements. For the two corals at 12.0 ka the ellipses are based on the average analytical uncertainty of ±70 years because the separate age measurements do not come from the same parent population. Two relevant points from *Adkins et al.*'s [1998] study of North Atlantic deep-sea corals are also plotted (samples JFA2 (13.7 ka) and JFA17 (12.9 ka)). While production can account for the observed $\Delta^{14}C$ hump at 11.0 ka, production is not sufficient to account for the observed Younger Dryas peak at 12.8 ka. (b) Curve through the deep-sea coral data demonstrating how the data are consistent with behavior that is inverse to the atmosphere, which is evidence for a slowdown of North Atlantic intermediate water/deep water formation.

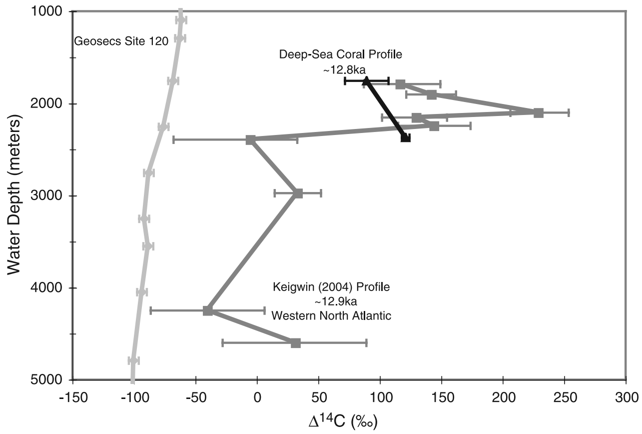

Figure 7. Profile of $\Delta^{14}C$ at the beginning of the Younger Dryas. Our deep-sea coral $\Delta^{14}C$ profile is consistent with *Keigwin*'s [2004] profile. To convert *Keigwin*'s [2004] benthic/planktonic foraminifera age differences to $\Delta^{14}C$, we converted the planktonic ^{14}C ages to calendar ages using Calib5.0 and then calculated $\Delta^{14}C$ for the deep water using the ^{14}C age of the benthic foraminifera. Comparing the modern profile from Geochemical Ocean Sections Study (GEOSECS) station 120 with the Younger Dryas profile reveals the presence of relatively enriched $\Delta^{14}C$ in the Younger Dryas ocean (because of higher ^{14}C production rates in the past) and the existence of a steep gradient at ~2400 m.

above 2400 m. A $\Delta^{14}C$ inversion is present with the intermediate depth water (1684–1829 m) ~35‰ depleted relative to deeper water (2372 m). This "slanted" shape to the $\Delta^{14}C$ profile is also seen in Keigwin's data and probably reflects the presence of a southern source intermediate water, analogous to modern AAIW, at this northerly latitude that is less dense than the recently ventilated northern source water at ~2000 m depth. This sort of lateral water mass movement, as opposed to the deepening and shoaling of GNAI/DW, has been observed for other times during the deglaciation [*Robinson et al.*, 2005]. Based on the time series of $\Delta^{14}C$ at ~1700 m in Figure 6, we imagine that the profile in Figure 7 evolves to higher $\Delta^{14}C$ values above 2000 m over the course of the Younger Dryas.

[22] Capitalizing on the decadal resolution possible in a single coral, we note that the variability of the within coral transect results vary depending on the timing within the Younger Dryas. As noted earlier, the three corals at the beginning and two at the end of the Younger Dryas show no significant variability in $\Delta^{14}C$ over their lifetimes (with an uncertainty ~10‰). This consistency is in sharp contrast to the two coral records at ~12.0 ka, from opposite sides of the North Atlantic basin and separated by more than 500 m depth, that both show a transient ~40‰ $\Delta^{14}C$ enrichment over their lifetimes (Figures 3 and 6). With their overlapping calendar age errors and similar $\Delta^{14}C$ enrichments, we interpret the $\Delta^{14}C$ record in these corals to reflect the same event on opposite sides of the North Atlantic. This pulse occurred rapidly, and the speed of the transition requires a shift in the water composition. It is likely that "young" northern source

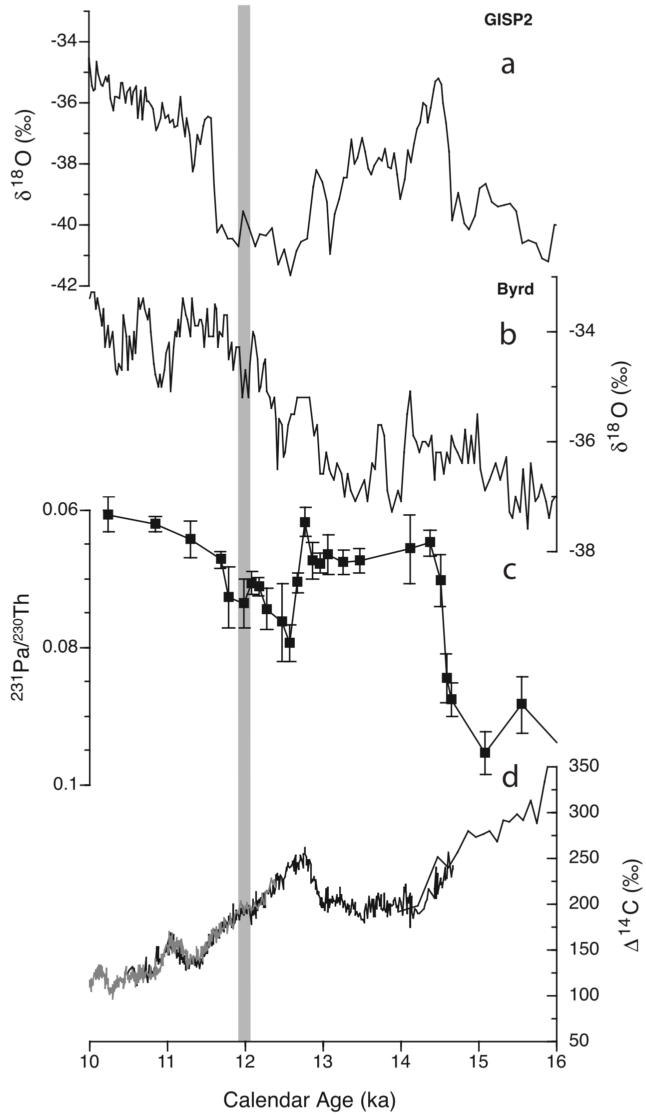

Figure 8. A comparison of our observed transient intermediate/deep ocean event at 12.0 ka to (a) GISP2 $\delta^{18}O$ [*Grootes et al.*, 1993], (b) Byrd $\delta^{18}O$ [*Blunier and Brook*, 2001], (c) (Pa/Th) [*McManus et al.*, 2004], and (d) records of atmospheric $\Delta^{14}C$ [*Kromer and Becker*, 1993; *Spurk et al.*, 1998; *Burr et al.*, 1998; *Hughen et al.*, 2000, 2004a, 2004b]. The sizable intermediate water/deepwater event that we observe in both basins of the North Atlantic is not clearly observed in the ice core records of northern or southern $\delta^{18}O$. The (Pa/Th) record also does not show a large shift in the strength of the meridional overturning circulation. The atmospheric $\Delta^{14}C$ record, however, does record a slight leveling out shift in slope that could indicate a perturbation to the carbon cycle.

water briefly dominated the water masses at this site, pushing out the more depleted water originating from a similar depth but spreading from the south.

[23] The $\Delta^{14}C$ enriched part of the transient event approaches the $\Delta^{14}C_{atm}$, a situation that is not observed in the modern ocean, even in surface water. The trend in atmospheric $\Delta^{14}C$ (that sets the initial $\Delta^{14}C$ of the water) and the trajectory of ^{14}C decay are very similar from the $\Delta^{14}C_{atm}$ peak through the end of the Younger Dryas. Therefore a ^{14}C enriched water mass could have formed anywhere in this time interval and evolved parallel to the atmospheric trend once isolated from the surface, or this enriched $\Delta^{14}C$ water mass could have formed because the initial $\Delta^{14}C$ of the intermediate/deep water that came to bathe the corals was simply more enriched relative to the atmosphere than in the modern ocean (i.e., a younger "reservoir age"). *Stocker and Wright* [1996, 1998] used a "2.5-D" model to investigate the ocean response to a slowdown in North Atlantic overturning caused by the input of fresh water to the high-latitude north and found that surface reservoir ages were reduced to ~200 years (25‰) at 39°N. This result, if applicable to the initial $\Delta^{14}C$ of intermediate/deep water, could account for our observed $\Delta^{14}C$ within error. Given the close match of the tree ring [*Friedrich et al.*, 2004] and Cariaco Basin [*Hughen et al.*, 2004b] records back to 12.4 ka, it is unlikely that the atmospheric record of $\Delta^{14}C$ is underestimated through part of the Younger Dryas, but an increased reservoir age correction to the Cariaco Basin record would result in a higher peak and a steeper atmospheric decline from 12.8 ka to 12.4 ka, which would be sufficient to explain the enriched $\Delta^{14}C$ that we observe.

[24] The 12.0 ka transient event is without an equal magnitude counterpart in any other record of climate during the Younger Dryas (Figure 8). GISP2 $\delta^{18}O$ [*Grootes et al.*, 1993] records a very small warming, and the (Pa/Th) record [*McManus et al.*, 2004] is consistent with a small decrease in the deep North Atlantic circulation rate. The slight flattening of the atmospheric $\Delta^{14}C$ record [*Friedrich et al.*, 2004; *Hughen et al.*, 2004b] at 12.0 ka could be interpreted as a slowdown in NADW formation (in agreement with (Pa/Th). However, the surface coral reconstruction of $\Delta^{14}C_{atm}$ is much more variable, obscuring any "kink" in the record [*Burr et al.*, 1998]. Furthermore, we observe that Antarctic $\delta^{18}O$ from the Byrd ice core increases steeply just prior to 12.0 ka (~2‰ over ~300 years), which suggests that this transient event may have originated in the south (Figure 8b). However it was caused, our data from this transient event show that the intermediate water of the North Atlantic can be quite variable with little associated atmospheric effect. Our within coral transects span about 100 years of climate history. This is certainly long enough to see changes in the ice core records and tree ring data shown in Figure 8, though it is close to the limit of resolution for the ice. On the other hand, 100 years is too short a time span for the Pa/Th record from the Bermuda Rise to record a robust signal. With this in mind, the fact that there is any hint of a change in these other records at 12.0 ka implies that our deep-sea coral data are not an analytical artifact. However, we do believe that our data set is climatically more sensitive to this 12.0 ka event because of the presence of large $\Delta^{14}C$ gradients in the Younger Dryas ocean. So, our signal is hard to see in other records because of both inherent temporal resolution of the other archives and the muted nature of the climate signal outside of deep $\Delta^{14}C$.

[25] After the atmospheric $\Delta^{14}C$ peak, $\Delta^{14}C_{atm}$ declines for the remainder of the Younger Dryas (12.8–11.5 ka) while the $\Delta^{14}C$ of intermediate water/deepwater approaches the atmosphere. This is consistent with the reinvigoration of NADW formation bringing more ^{14}C into the deep North Atlantic from the atmosphere. At the close of the Younger Dryas, two deep-sea corals show that intermediate water/deep water $\Delta^{14}C$ (~1200 m) becomes indistinguishable from atmospheric $\Delta^{14}C$. This observation is harder to explain than its "young water" counterpart at 12.0 ka. While this is a surprising result, the end of the Younger Dryas is an exceptional period and we can think of one explanation for our surprising data. If the open ocean mode of convection were interrupted during the Younger Dryas, surface ocean water would more fully exchange with the atmosphere. A restart of the convection would then simultaneously transport this enriched surface water $\Delta^{14}C$ to intermediate depths and cause a steep drop in $\Delta^{14}C_{atm}$. This scenario is consistent with our observed enriched corals at ~1200 m and the steep drop in $\Delta^{14}C_{atm}$ at ~11.5 ka. However, very young "reservoir ages" for high-latitude surface waters, or their precursors in the tropics, have not been observed. After this transient, the system must return to a steady state where intermediate depths are older than the atmosphere [*Frank et al.*, 2004], but the end of the Younger Dryas is clearly a time where transients dominate the system.

5. Conclusions

[26] Because changes in the $\Delta^{14}C$ of the intermediate/deep ocean occur too fast to be accounted for by radioactive decay alone, we conclude that our deep-sea coral measurements of North Atlantic intermediate water/deep water $\Delta^{14}C$ primarily reflect the rapid reorganization of water masses during the Younger Dryas. Our data indicate that, along with the rise in atmospheric $\Delta^{14}C$ and the drop in Greenland temperatures, ^{14}C-depleted southern source water came to bathe our North Atlantic coral growth sites consistent with a shoaling of or a reduction in NADW formation. The magnitude of the $\Delta^{14}C$ changes we observe implies that large $\Delta^{14}C$ gradients existed in the intermediate/deep ocean. One such gradient is illustrated by *Keigwin*'s [2004] vertical profile of the water column that shows a transition to depleted $\Delta^{14}C$ at ~2400 m. In addition, the age "inversion" above the main ^{14}C gradient in this profile is probably due to a ^{14}C-depleted southern source water, analogous to modern Antarctic Intermediate Water, reaching our site. A transient ~40‰ enrichment in $\Delta^{14}C$ over ~100 yr at 12.0 ka on both sides of the North Atlantic basin shows that deep water is capable of rapid, transient reorganization events

with a muted effect in the atmosphere. The identification of additional Younger Dryas deep-sea corals that fill in gaps between the existing data points and the development of a deep-sea coral proxy to gauge the effect of conservative mixing will further refine our understanding of this abrupt climate event.

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