

Tropical Atlantic coral oxygen isotopes: glacial–interglacial sea surface temperatures and climate change

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Abstract

We have generated a detailed oxygen isotope time-series from the fossil coral reefs from offshore Barbados. The Barbados coral-based record is a unique paleoceanographic data set with an equivalent sedimentation rate in excess of 600 cm/kyr where not only is the annual signal uniquely preserved but seasonality as well. Oxygen isotope values during the late glacial and LGM (16–20 ¹⁴C kyrs; 19–24 calendar kyrs) are $\sim 2.3\text{‰}$ heavier than corresponding living specimens, and indicate a regional cooling on the order of 4.5°C. There is also an isotopic expression of the Bølling-Pre-Boreal climate oscillation, with values reflecting a cooling during the Younger Dryas. Seasonality, a key diagnostic of the state of the climate system, remained the same or slightly less than present and indicates that the observed coolings were a change in the mean state of the western tropical Atlantic. Pan-tropic cooling during the last glaciation is best explained by a change in the radiative balance of the tropics. Variable tropical sea surface temperatures during climate oscillations such as the Younger Dryas challenge the paradigm that climate change only cascades from variations in North Atlantic deep water production. © 2001 Elsevier Science Ltd All rights reserved.

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1. Introduction

Recent evidence from fossil corals indicates that the deep mixed layer region of the western equatorial Atlantic was $\sim 5^{\circ}\text{C}$ cooler than at present at the last glacial maximum (Guilderson et al., 1994). Colder sea surface temperatures (SSTs) at Barbados (13°N,

59.5°W) may reflect either local or regional cooling of the western tropical Atlantic. These results are in stark contrast to the $\leq 2^{\circ}\text{C}$ temperature change indicated in the CLIMAP transfer function paleotemperature estimates (e.g. CLIMAP, 1976, 1981), but consistent with the available terrestrial evidence for a cooler tropical Atlantic from vegetation reconstructions (e.g. Bush and Colinvaux, 1990; Maley, 1989, 1991; Leyden et al., 1994; Colinvaux et al., 1996), tropical snowlines (summarized in Broecker and Denton, 1989), and from noble gas paleotemperature estimates in a low altitude Brazilian aquifer (Stute et al., 1995). Open marine planktonic foraminiferal stable isotope ($\delta^{18}\text{O}$) records tend to have a smaller glacial–interglacial amplitude than the Barbados

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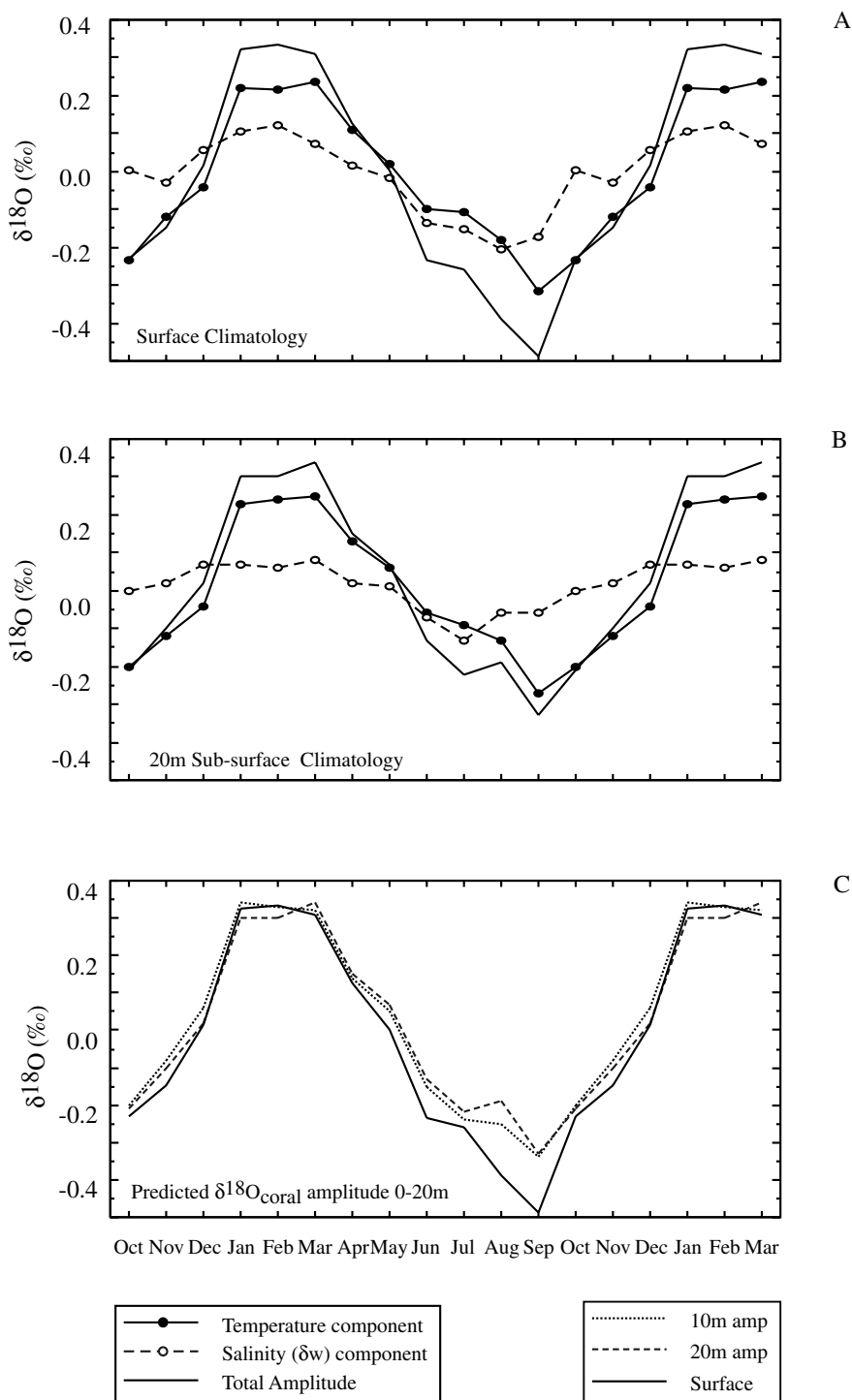


Fig. 1. Barbados (59°W, 13.5°N) climatological sea surface temperature and salinity (Levitus and Boyer, 1994) scaled into $\delta^{18}\text{O}$ space using 0.22‰/°C (Weber and Woodhead, 1972). Salinity is scaled using the local water line (Fairbanks et al., 1992). The combined seasonal $\delta^{18}\text{O}$ amplitude of a coral living in the surface mixed layer is $\geq 80\%$ temperature.

coral sequence with a corresponding smaller paleo-temperature estimate (e.g. Broecker, 1986).

The original Barbados *A. palmata* $\delta^{18}\text{O}$ sequence reached only to 15.4 ^{14}C kyrs without any samples within the classically defined Younger Dryas chronozone. We have sampled additional coral specimens from the mixed coral or buttress zone of the Barbados paleo-reef sequence and determined their oxygen isotope composition to resolve two key questions. First, is the difference between the open marine records and the Barbados record a consequence of their respective lengths and resolution? Second, with a more detailed record is there evidence of the Younger Dryas event? The additional oxygen isotope analyses extend the Barbados time series to ~ 20 ^{14}C kyrs BP and increase the resolution during the deglaciation.

1.1. Barbados regional and local hydrography

The western tropical Atlantic is characterized by a deep thermally mixed surface-layer (DML, ≥ 60 m) and minimal sea surface temperature variations (Picaud et al., 1984; Levitus and Boyer, 1994). Hydrocasts taken directly offshore Barbados indicate that the local hydrography is indistinguishable from the open ocean without any discernible “island effect”. Casts taken between November 1961 and April 1964 at Station 1B (Beers et al., 1965), and more recent stations off the south coast (Stansfield et al., 1995) show a surface thermally mixed layer in excess of 60 m. There is a steep seasonal halocline beginning at 25–30 m in conjunction with the seasonal salinity minimum and a salinity maximum at the top of the permanent thermocline. Barbados’ coasts are bathed by waters brought in by the seasonally varying Guiana and North Equatorial Currents (NEC). Present-day SSTs range from 26–29°C, reaching maximum temperatures in October. Salinity follows a general seasonal cycle (33.5–35.5‰) with freshening from January through late July–August as a consequence of Amazon run-off that is advected into the North Brazilian Coastal Current (NBCC) and into the Guiana Current (Borstad, 1982; Müller-Karger, 1988). In late August, salinity begins to rise due to both the decreasing Amazon discharge and the northward migration of the Inter-Tropical Convergence Zone (ITCZ), which retroflects the NBCC into the

North Equatorial Counter Current (NECC) (Richardson and Walsh, 1986; Müller-Karger, 1988).

1.2. Influences on (Barbados) $\delta^{18}\text{O}$ coral

Embedded in the oxygen isotopic composition of corals is information on temperature and the oxygen isotopic composition of the water (δ_w) in which the coral resided. Variations in δ_w include a eustatic or global component over glacial–interglacial time-scales due to ice volume (e.g. Fairbanks, 1989) and local or regional variations. The Barbados coral sequence is a direct record of sea level history (i.e. ice volume, Fairbanks 1989, 1990), and so it is possible to subtract an estimate of the mean global change in δ_w due to ice volume (e.g. Guilderson et al., 1994). The residual $\delta^{18}\text{O}$ signal is a combination of temperature and local δ_w (salinity) variations. Present day salinity variations at Barbados are dominated by the influence of variable Amazon discharge.

Comparison of the modern seasonal temperature cycle (converted into $\delta^{18}\text{O}\text{‰}$) and δ_w variations shows that the seasonal amplitude recorded in coral $\delta^{18}\text{O}$ residing in the uppermost surface waters at Barbados is $\sim 80\%$ temperature (Fig. 1). The oxygen isotopic composition of modern Barbados surface waters closely follows the salinity cycle and has a zero salinity intercept of -5.2‰ (Fairbanks et al., 1992). This zero salinity end-member is consistent with the measured δ_w of Amazon River water and its major tributaries which average $\sim -5.5\text{‰}$ (Mortatti et al., 1987; Karr and Showers, 1992). Oceanic precipitation has a minimal impact on δ_w (and salinity) and thus we are left with a simple two component mixing model between an Amazon end-member (0‰ salt, -5.2‰ δ_w) and a sub-tropical one ($\sim 36.1\text{‰}$ salt, 1.0 – 1.2‰ δ_w) (Fairbanks et al., 1992). Solving this mixing model for either salt or δ_w constrains the mean annual fraction of Amazon derived water at Barbados to between 3 and 8%. Changing the Amazon δ_w end-member 1.2‰ in either direction results in only a $\sim 0.05\text{‰}$ shift in the local δ_w value, and thus has minimal impact on paleo-temperature estimates. Doubling (halving) the Amazon input would decrease (increase) δ_w by $\sim 0.22\text{‰}$ and is a possible source of error in paleo-SST estimates.

2. Methods

Circum-Caribbean reefs are characterized by a distinct coral depth zonation (Goreau, 1959; Lewis, 1960). The reef crest zone, constrained to the upper ~5 m of the water column, is composed almost exclusively of *A. palmata*. Below this or just off the reef crest is the mixed coral or buttress zone, which begins at ~6–10 m and extends to 15–30 m depending on the underlying geomorphology. A very diverse coral assemblage including high abundances of the hemispherical or boulder variety of *Montastrea annularis* (Ellis and Solander) and *Porites asteroides* (Lamarck) with living coral cover often reaching 90% characterizes this zone.

Cores from the submerged coral reefs off the south coast of Barbados (Fairbanks, 1989) and a new suite of shallow cores taken in 1995 from the same locality have been sampled for this study. Cores of *A. palmata*, *M. annularis*, and *P. asteroides* recovered from the drowned reefs were cut, parallel to the vertical growth axis, into ~8 mm thick slabs using a diamond blade saw continuously flushed with water. Slabs were rinsed in distilled water, sonicated in deionized water (which was replaced through a continuous flow system), and dried at <45°C. Slabs were X-radiographed to provide sampling templates. Samples were rejected if they exhibited borings and encrusting algae with stringers into the corals as deduced via X-radiographs or visual microscopic examination. Skeletal architecture of the paleo *M. annularis* specimens indicates that they are of the hemispherical or boulder variety (D. Dodge, pers. comm.). This in conjunction with their association with *Diploria*, *P. asteroides*, and to some extent *A. palmata* indicates that the buttress zone was recovered. This association of specimens probably grew in water less than 20 m deep. Additional supporting evidence is inferred from the constraints on the break in species morphology in *M. annularis* where platy forms are only found deeper than 20 m (e.g. Baker and Weber, 1975). The offshore deglacial sequence is remarkably pristine relative to Holocene or recently dead samples, most probably due to the rapid vertical growth rates of the drowned reefs as they kept up with rising sea level.

Linear growth rates of individual corals were determined by using the low density to high-density band distance as an approximate annual chronometer (e.g.

Knutson, 1972; Dodge and Thompson, 1974). To minimize growth-rate kinetic induced effects, we only analyzed specimens with similar linear growth rates and from the same relative position within the corals. Typical linear extension rates were 5 mm/year for all specimens analyzed and ranged from 3 to 7 mm/year. Two to four parallel transects encompassing 4–8 years per transect were extracted using a micro dental drill along the main vertical growth axis. To minimize the potential for skeletal inhomogeneities, we attempted to only sample theca (walls), except for the *P. asteroides* specimens whose skeletal architecture is too fine for theca sampling. *A. palmata* specimens were only sampled along the upper growing surface as deduced from the skeletal architecture observed in X-radiographs. We used X-ray diffraction analysis of drilled and ground specimens to confirm that the aragonite was pristine as indicated by the absence of any calcite peaks (detection limit 0.5 wt%). The sampling protocol did not convert aragonite to calcite. Living specimens collected over the last ~20 years were treated in a similar fashion to the offshore cores.

Samples (100–200 µg) were reacted at 90°C in a Carousel-48 device and analyzed on a Finnigan MAT 251 isotope ratio mass spectrometer. Stable isotope analyses were also performed on powders (~60 µg) recovered by discretely sampling with a 300 µm diameter drill bit along the axis of a glacial, Younger Dryas, and a modern *M. annularis* recovered from 17.5 m water depth as well as glacial and deglacial *A. palmata*. Calibration is through secondary standards calibrated via NBS (now NIST SRM) standards 16, 17, 19, and 20. Analytical precision of concurrently analyzed standards is better than ±0.05‰ (1-sigma) for both oxygen and carbon. Data is reported in δ-notation relative to Pee Dee Belemnite using the defined values of NBS-20 (δ¹⁸O – 4.14‰ PDB, δ¹³C – 1.06‰ PDB). We have not corrected the absolute values of our results for the “acid-alpha” difference between aragonite (samples) and calcite (standards).

Radiocarbon and ²³⁰Th/²³⁴U chronological control is provided by previously reported radiocarbon and Th/U TIMS measurements (Fairbanks 1989; 1990; Bard et al., 1993), supplemented by additional measurements (Fairbanks et al., in prep.). The majority of samples through the deglaciation have been

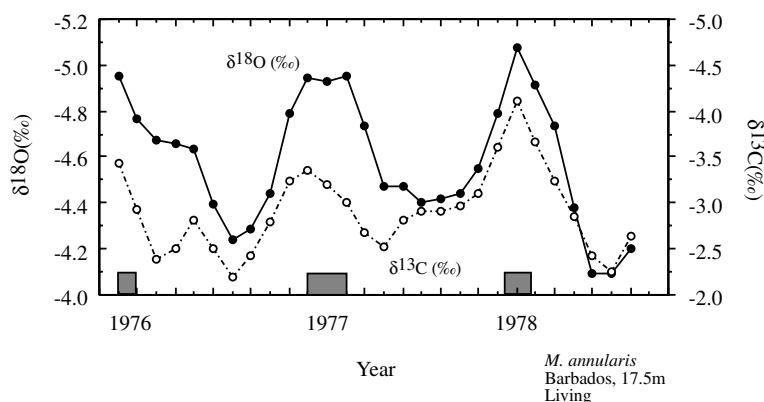


Fig. 2. Sequential stable isotope analysis of a living *M. annularis* recovered (year) from 17.5 m water depth off Barbados. Hatched bars mark high density bands. Note that the seasonal amplitude in $\delta^{18}\text{O}$ is the same as that predicted in Fig. 1.

dated with both techniques. *M. annularis* specimens older than 15.7 ^{14}C kyrs have yet to be Th/U dated. Absolute ages for these specimens are interpolated using a least squares regression through the appropriate coral data (Fairbanks et al., in prep.). Ice volume residual $\Delta\delta^{18}\text{O}$ is calculated in a similar fashion to Guilderson et al. (1994). In specimens older than the last glacial maximum as determined by the *A. palmata* Barbados sea level record, the residual is calculated using an ice volume component of 1.1‰.

3. Results

Stable isotope analysis of a modern (living) *M. annularis* from 17.5 m water depth matches our estimate of the influence of δ_w on $\delta^{18}\text{O}$ in corals residing in the buttress zone (Fig. 2). We assume fossil corals from this zone will also record surface water characteristics. We note that the absolute value of the analyses on this specimen are $\sim 0.6\text{‰}$ higher than those measured by Fairbanks and Dodge (1979) on a *M. annularis* from a nearby location. A similar offset on the exact same Bermuda specimen used by Fairbanks and Dodge (1979) was also determined. The offset is consistent with conversion of aragonite to calcite and indicates that pre-treating samples with low temperature oxygen plasma cleaning (used by Fairbanks and Dodge) is inappropriate for aragonitic coral samples.

Oxygen isotope data (Table 1; Fig. 3(A)) are presented relative to that of the respective living

specimens to remove the genera-specific disequilibrium (Weber and Woodhead, 1972). In general, offset tracts from the same coral are within 0.1‰ (1-sigma) of each other. The updated multiple coral species oxygen isotope record exhibits the classic Bølling through Pre Boreal oscillation previously recognized only in records from higher northern latitudes (Ruddiman and McIntyre, 1981). Our estimated correction of the global ice volume component further enhances the oxygen isotope oscillation during the Younger Dryas chronozone (Fig. 3(B)). Late glacial (16–20 ^{14}C kyrs; 17–24 cal kyrs) $\delta^{18}\text{O}$ values are $\sim 2.3\text{‰}$ heavier than corresponding living specimens.

4. Discussion

4.1. Glacial–interglacial $\delta^{18}\text{O}$

In order to convert $\delta^{18}\text{O}$ into temperature the mean ocean change due to ice volume must be removed. The Barbados sea level record constrains the change in relative sea level to 121 ± 5 m (Fairbanks, 1989). The maximum relationship between sea level and ice volume $\delta^{18}\text{O}$ determined by Fairbanks and Matthews (1978) is 0.11‰ per 10 m of sea level, or $\sim 1.3\text{‰}$ with a eustatic lowering of 121 m. Measuring the oxygen isotopic composition of deep-sea pore waters Schrag et al. (1996) estimate a mean ocean change due to ice volume of just over 1‰. This discrepancy needs to be resolved to accurately estimate the true glacial–interglacial SST amplitude recorded in oxygen isotope

Table 1
Oxygen isotope data

Sample ID	Species	Age (kys BP)	^a	$\delta^{18}\text{O}$ (‰)	Relative to present day (‰)	$\Delta\delta^{18}\text{O}$ (‰) ^b
Living Aps	<i>A. palmata</i>	0.00		−3.75	0.00	0.00
Living Pasts	<i>P. asteroides</i>	0.00		−3.86	0.00	0.00
Living Manns	<i>M. annularis</i>	0.00		−4.59	0.00	0.00
RGF 95ic 1-4	<i>A. palmata</i>	2.59	*	−3.31	0.44	0.42
RGF 95ic 1-7	<i>A. palmata</i>	4.00	I	−3.37	0.38	0.36
RGF 7-4-2	<i>P. asteroides</i>	7.42	*	−3.79	0.07	−0.08
RGF 7-5-5	<i>A. palmata</i>	8.45	*	−3.22	0.53	0.27
RGF 7-9-6	<i>A. palmata</i>	9.10	*	−3.15	0.60	0.31
RGF 7-10-2	<i>A. palmata</i>	9.15	*	−3.21	0.54	0.24
RGF 7-12-2	<i>A. palmata</i>	9.26	*	−2.93	0.82	0.48
RGF 7-13-1	<i>A. palmata</i>	9.55	I	−3.04	0.71	0.36
RGF 7-16-2	<i>A. palmata</i>	9.73	*	−3.05	0.70	0.31
RGF 7-19-3	<i>P. asteroides</i>	10.06	*	−2.99	0.87	0.47
RGF 7-25-2	<i>A. palmata</i>	10.84	I	−3.22	0.53	0.04
RGF 7-26-2	<i>A. palmata</i>	10.95	*	−3.14	0.61	0.12
RGF 7-27-4	<i>A. palmata</i>	11.09	*	−2.64	1.11	0.60
RGF 16-12-6	<i>A. palmata</i>	11.40	I	−2.57	1.18	0.52
RGF 16-12-7	<i>A. palmata</i>	11.41	*	−2.43	1.32	0.66
RGF 9-2-5	<i>M. annularis</i>	11.89	*	−2.95	1.64	1.03
RGF 15-1-2	<i>M. annularis</i>	12.00	*	−2.62	1.96	1.36
RGF 12-9-3	<i>P. asteroides</i>	12.19	I	−2.21	1.65	0.97
RGF 12-9-5	<i>A. palmata</i>	12.26	*	−2.38	1.37	0.67
RGF 15-2-4	<i>M. annularis</i>	12.75	I	−2.47	2.11	1.48
RGF 15-2-5 top	<i>M. annularis</i>	12.76	C	−2.42	2.17	1.53
RGF 15-2-5 lower	<i>M. annularis</i>	12.78	I	−2.47	2.12	1.48
RGF 12-21-2	<i>A. palmata</i>	13.35	*	−2.48	1.27	0.44
RGF 12-21-7	<i>A. palmata</i>	13.64	*	−2.22	1.53	0.69
RGF 12-21-10	<i>A. palmata</i>	13.73	*	−2.11	1.64	0.80
RGF 9-12-7	<i>A. palmata</i>	14.46	*	−1.87	1.88	0.79
RGF 9-13-3	<i>A. palmata</i>	14.69	*	−2.20	1.55	0.44
RGF 13-2-4	<i>M. annularis</i>	15.18	I	−2.59	2.00	1.02
RGF 15-10-6	<i>A. palmata</i>	17.15	I	−1.50	2.25	1.05
RGF 13-8-6	<i>A. palmata</i>	17.16	I	−1.33	2.42	1.17
RGF 13-8-7	<i>A. palmata</i>	17.17	I	−1.58	2.17	0.92
RGF 9-20-2	<i>A. palmata</i>	18.02	*	−1.63	2.12	0.88
RGF 9-21-9 ^c	<i>A. palmata</i>	18.44	*	−1.40	2.35	1.10
RGF 9-21-10 ^c	<i>A. palmata</i>	18.30	I	−1.39	2.36	1.10
RGF 9-21-11 ^c	<i>A. palmata</i>	18.24	*	−1.56	2.19	0.94
RGF 9-24-2	<i>A. palmata</i>	18.84	*	−1.38	2.37	1.08
RGF 9-24-4	<i>A. palmata</i>	18.90	*	−1.55	2.20	0.91
RGF 14-5-7	<i>M. annularis</i>	18.90	I	−1.89	2.70	1.51
RGF 14-8-2	<i>M. annularis</i>	19.65	I	−2.39	2.20	0.91
RGF 15-14-4	<i>M. annularis</i>	19.95	I	−2.32	2.27	1.19
RGF 15-15-8	<i>M. annularis</i>	20.28	I	−2.41	2.17	1.09
RGF 15-18-4	<i>M. annularis</i>	21.01	I	−2.31	2.28	1.19
RGF 15-18-5	<i>M. annularis</i>	21.01	I	−2.41	2.17	1.09
RGF 15-20-6	<i>M. annularis</i>	21.45	I	−2.14	2.45	1.36
RGF 9-34-7	<i>P. asteroides</i>	21.96	I	−1.97	1.89	0.80
RGF 15-22-7	<i>M. annularis</i>	21.97	I	−2.49	2.10	1.01
RGF 9-34-8	<i>P. asteroides</i>	21.98	*	−1.61	2.25	1.15
RGF 9-34-10	<i>P. asteroides</i>	22.00	I	−1.76	2.10	1.01

Table 1 (continued)

Sample ID	Species	Age (kyrs BP)	^a	$\delta^{18}\text{O}$ (‰)	Relative to present day (‰)	$\Delta\delta^{18}\text{O}$ (‰) ^b
RGF 9-35-2	<i>P. asteroides</i>	22.01	I	−1.58	2.28	1.18
RGF 9-35-6	<i>M. annularis</i>	22.36	I	−2.30	2.29	1.20
RGF 9-37-2	<i>M. annularis</i>	22.74	I	−2.24	2.35	1.26
RGF 9-37-3	<i>M. annularis</i>	22.84	I	−2.25	2.34	1.25
RGF 9-38-1	<i>M. annularis</i>	23.65	I	−2.16	2.42	1.34
RGF 15-25-3	<i>M. annularis</i>	23.95	C	−1.89	2.70	1.61
RGF 15-25-4	<i>M. annularis</i>	24.07	I	−2.32	2.26	1.18
RGF 15-25-5	<i>M. annularis</i>	24.09	C	−2.29	2.30	1.21

^a Absolute chronologic control: * explicitly Th/U dated; I = interpolated using the Barbados ¹⁴C-Th/U data set (Fairbanks et al., in prep.); C = band counted from dated portion of continuous coral head.

^b Ice volume residual $\Delta\delta^{18}\text{O}$ as explained in the text (cf. Guilderson et al., 1994).

^c Th/U ages are indistinguishable from each other.

records. In terms of temperature, the $\sim 0.2\text{‰}$ difference equates to $\sim 1^\circ\text{C}$ at the LGM. We use the ocean ice volume estimate of Fairbanks (1989) and for samples prior to the LGM as deduced from the Barbados *A. palmata* sea level record we use 1.1‰ .

The updated multiple coral species record confirms the earlier *A. palmata* $\Delta\delta^{18}\text{O}$ glacial-modern amplitude of $\sim 2.3\text{‰}$. Correcting for global ice-volume yields a $1.0\text{--}1.2\text{‰}$ residual, which if strictly temperature indicates a $5 \pm 0.5^\circ\text{C}$ cooling. This amplitude is greater than that derived from planktonic foraminifera in tropical Atlantic cores which average $\sim 1.8\text{‰}$ (Broecker, 1986). The glacial–interglacial oxygen isotope amplitude in only the mixed layer dwelling *Gs. ruber* and *Gs. sacculifer* vary spatially from 1.0 to 2.3‰ (Fig. 4(A)) throughout the tropical Atlantic and seem to confirm Broecker’s mean Atlantic G-I $\Delta\delta^{18}\text{O}$ of 1.8‰ . However, this variability cannot reflect real oceanographic differences in either temperature or the δ_w — salinity field. In many cases $\Delta\delta^{18}\text{O}$ values from adjacent sites differ in excess of 0.4‰ , representing temperature differences in excess of 2°C or salinity differences of 2‰ using the modern δ_w — salinity relations (Fairbanks et al., 1992). This variability is likely a function of sedimentation rate, bioturbation, sample resolution, and to some extent dissolution; cores with low sedimentation rates, as expected, tend to have lower amplitudes indicative of an effective bioturbation filter. Higher sedimentation rate cores ($\geq 5\text{ cm/kyr}$) have amplitudes of 2.0‰ or more (e.g. Slowey and Curry, 1995; Curry and Oppo, 1997). Analysis of the tropical Atlantic

$\Delta\delta^{18}\text{O}$ as a function of estimated sedimentation rate (Fig. 4(B)) shows the expected response for a strong filtering mechanism (e.g. Peng et al., 1979). The Barbados coral record is a unique paleoceanographic data set because of its discrete nature which is irrecoverable from open tropical marine records. It has an equivalent sedimentation rate in excess of 600 cm/kyr or more than $200\times$ that of the average tropical Atlantic record of $\leq 3\text{ cm/kyr}$ (e.g. Mix, 1987) where not only the annual signal is uniquely recovered but seasonality as well. It is thus possible that the amplitude difference between the coral results and those from more conventional piston cores is a consequence of resolution (sedimentation rate) and bioturbation.

We have attempted to model bioturbation by interpolating the Barbados record at 200 year intervals and “resedimented” it at 5 cm/kyr with a 9 cm homogeneous mixed zone. Although our treatment of bioturbation is simplistic, the synthetic record bears some resemblance to the “high-accumulation” record of Curry and Oppo not only in terms of the glacial–interglacial amplitude but also the deglacial structure (Fig. 5). Obviously mismatches can occur due to the differences between the recording/sampling mechanisms and chronologic control. Open ocean isotopic records are generated at a more regular resolution (with respect to depth downcore) whereas the coral data are not as regular but have better chronologic control. The effect of bioturbation has implications for biomarker paleo-SST estimates as well. It has been shown that organic compounds have higher apparent (bioturbation) diffusion coefficients than

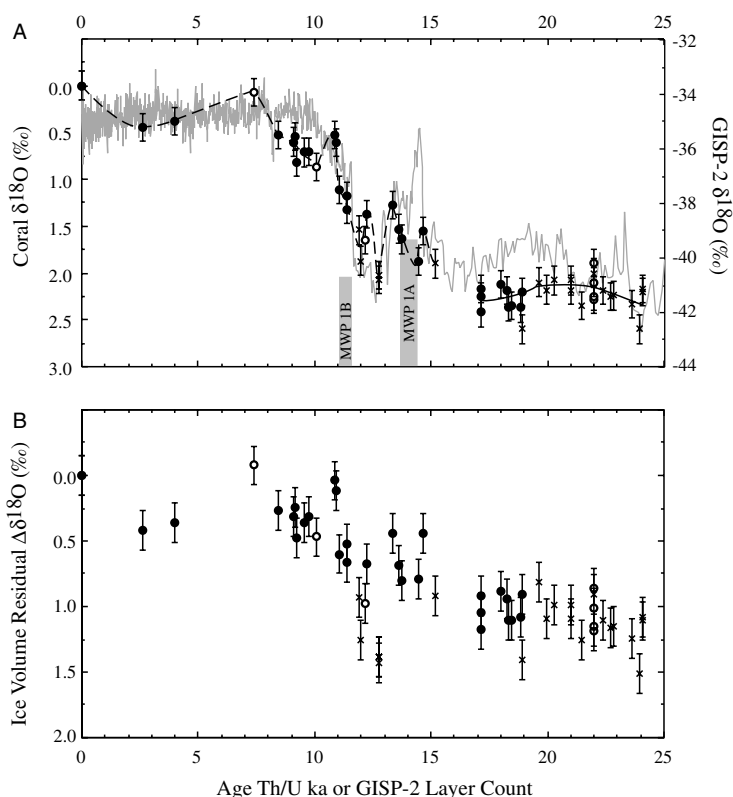


Fig. 3. (A) Barbados multi-coral oxygen isotope record to absolute (Th/U) time and the GISP-2 $\delta^{18}\text{O}$ ice record to layer count years (Stuiver and Grootes, 2000). Coral values are relative to results on living late-20th century Barbados specimens: *A. palmata* (solid circle) -3.75‰ , *P. asterioides* (open circle) -3.86‰ , *M. annularis* (crosses) -4.49‰ . Offset, parallel transects within each coral specimen are in general within $\pm 0.1\text{‰}$ of each other and we use this value as our error bars. MWPI-A and B as defined in Fairbanks (1989 and 1990) are marked. (B) Ice-volume (estimate) corrected residual to time. Coral $\delta^{18}\text{O}$ has been adjusted to account for ice volume using the Barbados sea level curve over the last 20 ka and an estimated ice-volume component of 1.1‰ between 20 and 24 ka. Glacial-modern amplitudes are $\sim 2.3\text{‰}$ or correcting for ice volume $1.0\text{--}1.2\text{‰}$ which if strictly temperature indicates $\sim 5^\circ\text{C}$ cooling during the late glacial and LGM.

particles; organic compounds are bioturbated deeper and more efficiently than the equivalent physical mixing of particles such as foraminifera (Wheatcroft et al., 1990; Smith et al., 1993; Pope et al., 1996). As a consequence, it is likely that there is signal amplitude attenuation over glacial–interglacial timescales in biomarker derived estimates from low sedimentation rate cores. We therefore argue that the Barbados isotope record is not anomalous compared to open marine records when bioturbation is taken into account, and interpret it as a regional (western tropical Atlantic) record.

Although we do not have any direct estimate of the glacial Amazon $\delta^{18}\text{O}$ composition, the analysis of the influence of the modern Amazon suggests that changing

its oxygen isotopic composition by as much as 1.2‰ would have a minimal ($\sim 0.05\text{‰}$) impact on the mean composition of Barbados surface waters. Glacial Amazon streamflow is also unknown. Glacial aridity in the Amazonas, a contentious subject (e.g. Colinvaux et al., 1999), if true would likely have resulted in a reduced Amazon streamflow. If we take a halving of the streamflow as an upper-limit with a corresponding $\sim 0.2\text{‰}$ increase in mean δ_w , then our temperature estimate would be reduced to 4°C . Thus although we can not explicitly convert our late-glacial and LGM oxygen isotope data to temperature, we can place a reasonable constraint on the SSTs being $\sim 4.5^\circ\text{C}$ colder reflecting a regional not local cooling over the interval $\sim 17\text{--}24$ kyrs BP ($16\text{--}20$ ^{14}C kyrs BP).

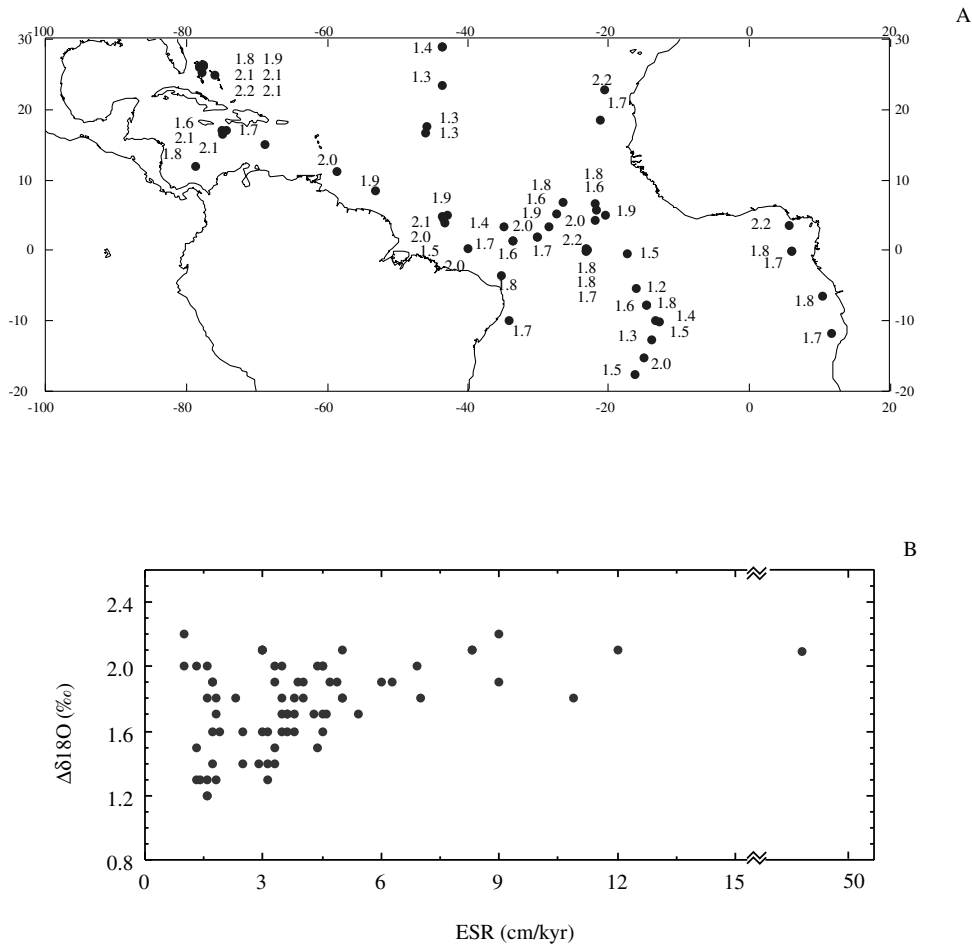


Fig. 4. Atlantic *Gs. ruber* and *Gs. sacculifer* glacial-interglacial $\Delta\delta^{18}\text{O}$ (A) map view and (B) as a function of sedimentation rate (cm/kyr). Stable isotope results presented were tabulated from the literature and can be found in Billups and Schrag (2000).

4.2. Deglacial $\delta^{18}\text{O}$

The Barbados multi-coral time-series has an expression of the Bølling–Younger Dryas–Pre Boreal where oxygen isotope value return towards glacial values. The ice volume correction further enhances the oxygen isotope oscillation during this interval. However, during deglaciation the oxygen isotope composition of any surface Atlantic $\delta^{18}\text{O}$ record cannot be precisely corrected for ice volume, particularly when the waning ice sheets are discharging their melt-water into the Atlantic basin. Deviations from the global mean ice volume correction over any given time interval result from the time it takes to

isotopically homogenize the signal throughout the hydrosphere and secondly, differences in the oxygen isotopic composition of the individual waning ice masses. We employ a mean ice volume correction (Fairbanks, 1989) that does not take into account the transient nature of the glacial meltwater as it mixes out of Atlantic surface waters. Thus, implicitly included in the residual ($\Delta\delta^{18}\text{O}$) signal is some fraction of the “ice volume” signal not adequately subtracted due to the mostly North Atlantic input of glacial melt water and subsequent mixing into the world’s oceans. Irregardless, there is a $\delta^{18}\text{O}$ (climatic) expression of the Bølling–Pre Boreal climate oscillation(s) in the tropical Atlantic.

The Barbados coral record is not the only high-resolution tropical Atlantic record to record climate change in this interval. The grey-scale record of the Cariaco basin on the north coast of Venezuela clearly shows trade wind intensity changes over this interval (Hughen et al., 1996). Our Younger Dryas results are primarily based on *M. annularis* specimens from the mixed coral zone and no deeper than ~20 m. Using the modern salinity- δ_w relation and hydrographic data, there is a $\sim 0.1\text{‰}$ difference in the mean δ_w at 20 m with that at the surface. If we use the same arguments on the effect of variable Amazon discharge as presented earlier and use a halving of the Amazon input as an upper limit, then we have a cumulative “salinity” effect on the order of 0.3‰ . After correcting for ice-volume the *M. annularis* derived residual is $\sim 1.35\text{‰}$ ($n=5$), the *P. asteroides* residual is 1‰ ($n=1$) as is that from a single *A. palmata* within the chronozone. Thus it seems that the open western tropical Atlantic as monitored at Barbados has an expression of the Younger Dryas with a cooling perhaps as much as that observed during the glaciation (0.7 to $\sim 1.0\text{‰}$). It would be incorrect to state that the chronozone was all the same amplitude. There is an interesting correspondance with a sequence of colder–warmer–colder values at Barbados and a similar potential interpretation of the GISP-2 ice core record (Fig. 3). However, the age-model errors in the respective records are such that the records could also be out of phase with respect to this additional small scale oscillation within the classic chronozone.

4.3. Implications on mechanisms of cooling

Seasonality is a key diagnostic of the state of the climate system and also climate change mechanisms. The cooling at Barbados which we interpret to be at the very least a regional western Atlantic phenomenon could be the result of increased seasonality or a change in the mean state. Sequential sampling of glacial and deglacial coral samples record $\delta^{18}\text{O}$ seasonality similar to or slightly less than present (Fig. 6). This observation indicates that for those intervals where we have sampled (for brevity, not all are shown) the cooling is a change in the mean state and not the result of increased seasonality. There are a number of mechanisms by which the

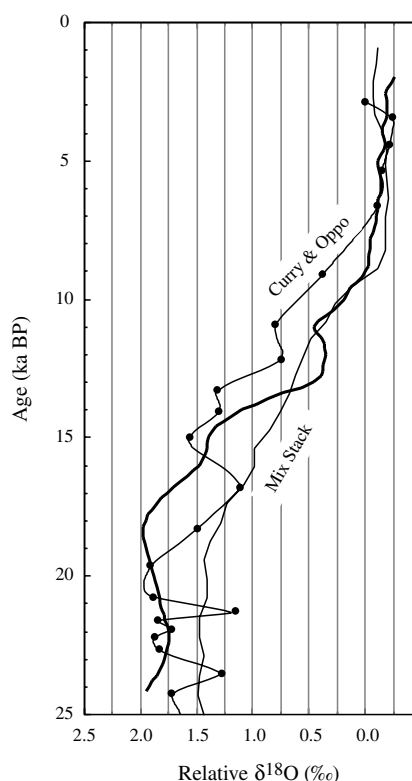


Fig. 5. Comparison of tropical Atlantic $\delta^{18}\text{O}$ time series: Barbados (thick solid line), the stacked tropical Atlantic record of Mix (1987, thin solid line), and the record of Curry and Oppo (1997, thin line with closed circles). The Barbados record has an equivalent sedimentation rate in excess of 600 cm/kyr , whereas most sediment cores have sedimentation rates $\leq 3\text{ cm/kyr}$. The stacked record, originally on a ^{14}C timescale, has been recast to calendar time using a third order polynomial relation. We have synthetically bioturbated the Barbados record with a 5 cm/kyr sedimentation rate and a 9 cm homogenous mixed layer. Our synthetic record recreates, to a decent first approximation, the open marine record of Curry and Oppo (1997) from a core with a 5 cm/kyr sedimentation rate. By necessity, this treatment of bioturbation is overly simplistic not taking into account temporal variations in accumulation and bioturbation, and completely disregards dissolution. It is evident that there is not a discrepancy in the glacial-interglacial $\delta^{18}\text{O}$ amplitude between the open marine tropical Atlantic and Barbados coral records, when bioturbation is taken into account.

western tropical Atlantic could have cooled during the glacial: a surface advective process whereby colder waters derived from the Canary Current or colder South Equatorial Current chill the western tropical Atlantic; an overall cooling of the surface and thermocline of the tropics as a consequence of

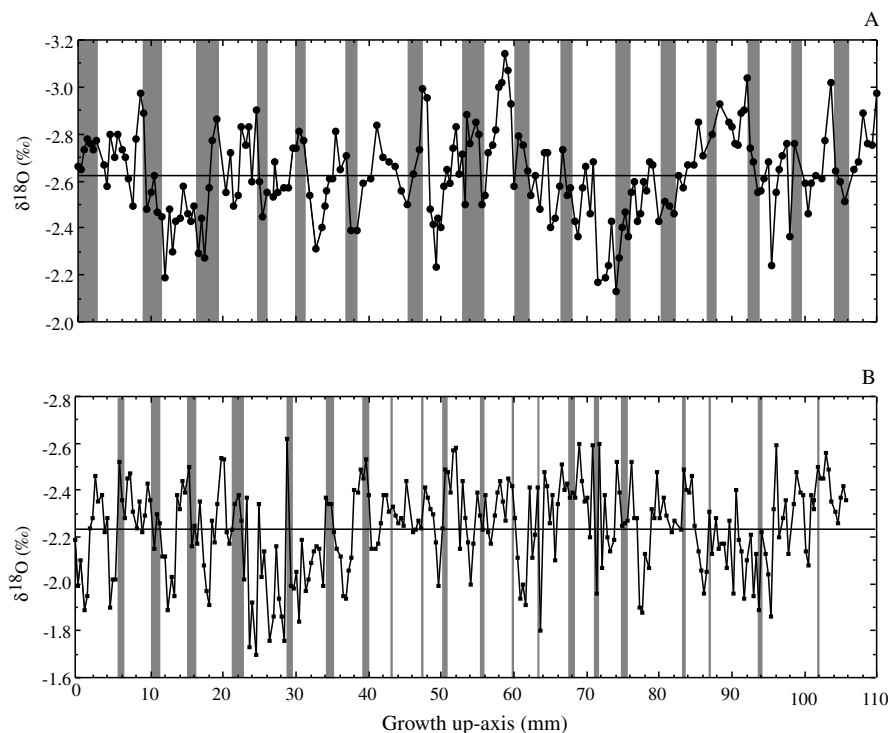


Fig. 6. (A) $\delta^{18}\text{O}$ seasonality during the Younger Dryas and (B) the glacial as recorded in Barbados *M. annularis* corals. Stippled boxes denote high density bands which are approximate annual markers. Mean value of the glacial specimen is -2.24‰ and -2.63‰ for the Y-D specimen. Axial extension is measured in mm with increasing growth upward. Although it is possible that there is some attenuation of the seasonal amplitude due to sampling, the lack of increased seasonality implies that the inferred cooling is the result of a change in the mean state.

increased thermocline ventilation; or a radiative balance change as a consequence of cloudiness.

Ravelo et al. (1990) clearly documented that during the LGM the mixed layer planktonic foraminifera assemblage remained in place in the western tropical Atlantic indicating that the hydrography of the western Atlantic was similar to present (i.e. a deep thermal mixed layer). Late Pleistocene solar radiation at low latitudes is nearly the same as present and if we assume that surface radiation ($\sim 200 \text{ W/m}^2$) and surface current velocities for the NEC and SEC ($14 \pm 4 \text{ cm/s}$ with a fast axis speed of $\sim 30 \text{ cm/s}$, respectively; Richardson and Walsh, 1986) were similar to or within 30% of today's it is hard for cold "point sources" to be the cause of the cooling. The deep mixed layer results in a large thermal capacitance which in conjunction with reasonable transit times from point sources allows for significant warming of a given parcel of water. In an elegant study, Slowey

and Curry (1995) documented that there is enhanced thermocline ventilation with a $\sim 4^\circ\text{C}$ cooling and a shallowing of the permanent thermocline at 26°N . They infer a southward migration of the northern edge of the subtropic gyre (where the thermocline waters outcrop) and increased Ekman pumping due to more vigorous winds. The obvious question is whether or not this is directly connected with the cooling observed further south at Barbados and our interpretation that the whole of the western tropical Atlantic cooled.

Studies of the ventilation of the modern tropical Atlantic thermocline have documented several interesting features germane to this discussion. The first is that although there is significant ventilation of the thermocline especially in the western portion of the basin near Brazil, the waters feeding the ventilation are relatively shallow otherwise there would be a physical manifestation in the form of cool surface

temperatures (Philander and Pacanowski, 1986). This upwelling is thought to be a non-local response in conjunction with downwelling between 3° and 8°N. In a separate study using transient tracers, Reverdin et al. (1993) determined that the tropical thermocline is dominated by waters from the southern hemisphere not waters from the northern hemisphere. This indicates that the shallow circulation of the tropical Atlantic is not a closed meridional cell and includes both hemispheric exchange and (some) diapycnal mixing. Variations in the depth of the thermocline are not necessarily correlated with the absolute vertical velocity; a shallower thermocline does not necessarily mean increased vertical velocity and there are additional non-local effects which can decouple simple east–west and north–south thermocline accommodation. Undoubtedly the increased ventilation of the thermocline at 26°N (which reflects increased Ekman pumping in the subtropics) and the cooling of the tropical Atlantic are related in a gross climate sense. However, it is unclear if there is a direct dynamic pathway indicating cause and effect.

Results on reconstructing mixed layer depth have also been obtained in the western Pacific warm pool (Andreasen and Ravelo, 1997) and in conjunction with paleo-SST estimates imply similar hydrography and temperatures that were 4–5°C colder (Patrick and Thunell, 1997). A pan-tropic SST change is best (more easily) explained by a change in the radiative balance of the tropics. This is perhaps the strongest evidence leading towards water vapor (e.g. low level clouds and tropical albedo) as the likely amplifier of climate change during the glacial (e.g. Ramanathan et al., 1989; Webb et al., 1997; Seager et al., 1999).

Late Pleistocene climate including the LGM and the Younger Dryas, have been used as a natural experiment to test the sensitivity of climate models in the hope of understanding the potential of future climate change in an evolving trace gas world (e.g. Manabe and Stouffer, 1997 and references therein). Of pivotal importance to these model simulations is the ultimate cause of the Younger Dryas and the means by which the climate oscillation is propagated through the climate system (atmosphere versus thermohaline circulation), and in like fashion, the cause of its abrupt termination. The sea level record of Fairbanks (1989, 1990) does not indicate a large melt-water trigger (i.e. MWPI-A) as the cause of the Younger Dryas:

MWPI-A precedes the beginning of the Younger Dryas by ~1200 years (Fairbanks, 1990) and its isotopic signature is well documented in the Gulf of Mexico (e.g. Leventer et al., 1982; Fairbanks et al., 1992). A key component of the conventional view of the Younger Dryas is that glacial melt-water entering the high north Atlantic would lower salinities reducing both convection and deep thermohaline circulation (Broecker et al., 1988; Manabe and Stouffer, 1997 among others). Model experiments which are used to study the melt-water induced climate change exhibit regional North Atlantic cooling with the effects observed “downstream” in Europe. Reduced thermohaline circulation would tend to raise tropical SSTs through a reduction in meridional ocean transport, in effect keeping heat in the tropics (Manabe and Stouffer, 1997). Therefore these models can not explain a tropical cooling during the Younger Dryas in the tropical Atlantic through a simple change in the thermohaline circulation.

If a synchronous cold Younger Dryas is found tropic wide, then it will be necessary to look for an alternative mechanism to generate and propagate this “anomaly”. Accepting the current generation of coupled ocean-atmosphere models as reflecting the response of the climate system to a given forcing such as freshwater discharge into the North Atlantic, then the synchronicity and similar amplitude of such an event indicates that variations in deep thermohaline circulation is not the direct forcing nor the propagating mechanism of the Younger Dryas climate anomaly. This is in part because a reduction in deep thermohaline circulation has the effect of keeping heat in the tropics through reduced poleward heat transport in addition to an overall reduction in the replacement of tropical waters from below (e.g. Manabe and Stouffer, 1997). At present, absolute chronological control is insufficient to determine the relative phasing of many of the various climate records over this interval, and one can only say that the records as currently envisioned are synchronous within errors (Lowell et al., 1995; Peteet 1995; Thompson et al., 1995). If it is born out that the physical expression of the Younger Dryas in the tropics and mid to high-latitudes of both hemispheres is indeed synchronous and the result of a single forcing mechanism, then this indicates a more “even handed” distributor of the climate oscillation than deep

thermohaline circulation and challenges the paradigm that the climate system only cascades from variations in North Atlantic Deep Water variations. In a simple model, it has been shown that the coupled tropical atmosphere–ocean system responds to Milankovitch forcing capable of altering the mean state of the tropical climate (Clement et al., 1999). This begs the question as to whether the origins of the Younger Dryas do not lie in the high north Atlantic, but the tropics (Clement and Cane, 1999).

5. Conclusions

We have determined the oxygen isotopic composition of coral specimens from the mixed coral or buttress zone of the Barbados paleo-reef sequence. Today, the oxygen isotope composition of corals from the mixed coral zone is dominated by temperature (relative to salinity and local precipitation), an assumption more than likely valid in past times. Late glacial (16–20 ^{14}C kyrs; 17–24 cal kyrs) $\delta^{18}\text{O}$ values are $\sim 2.3\text{‰}$ heavier than corresponding living specimens and reflect a cooling on the order of 4.5°C . The oxygen isotope record exhibits an expression of the Bølling–Younger Dryas–Pre Boreal where values return toward glacial values. A simple bioturbation model, when applied to the Barbados time-series, recreates quite well the planktonic foram records from the open tropical Atlantic. The coolings observed in the Barbados coral sequence reflect a regional change in the mean state of the western tropical Atlantic, and not an increase in seasonality. If truly synchronous with higher latitudes, a physical expression of the Younger Dryas in the tropics indicates a more “even handed” distributor of climate change than deep thermohaline circulation and challenges the paradigm that the climate system cascades from variations in North Atlantic Deep Water variations.

The Laurentide and Fennoscandian Ice Sheets supplied the bulk of ice during the last ice age (Denton and Hughes, 1981) and thus the relative rate of sea level change recorded in Barbados corals reflects the destruction of these masses, and is therefore a proxy for high latitude climate. Internally consistent chronologies from the Barbados coral series (sea-level and SST) indicate that rising tropical SSTs lead the major recessions of the continental ice sheets during the last deglaciation. A

similar internally consistent sequence is observed in the GISP-2 ice core record. Methane which is interpreted to reflect tropical climate (temperature and precipitation) is low over the interval 28–17 cal kyrs when it begins to rise, prior to the oxygen isotope expression of the Bølling (Brook et al., 1996). Moreover, Guilderson et al. (1994) noted that the warming observed at Barbados during the last deglaciation preceded re-initiation of modern deep thermohaline circulation, and the increase in atmospheric pCO_2 . It is not clear if some external forcing is driving the variations in tropical temperatures, or if the tropical temperatures are responding to an internal mechanism associated with the hydrologic cycle intimately linked with meridional energy transport. The indirect evidence leads towards a synergy between low latitude and high latitude climate which is not necessarily surprising; the tropics is the Earth’s primary redistribution point for incoming radiation. Tropical SSTs are a direct link in the export of energy to higher latitudes through fast feedback processes via the focusing effect for defining regions of convection and “atmospheric moisture pumps” which export latent and sensible heat to both hemispheres. Variable tropical SSTs are diagnostic of the sensitivity of the climate system at least on glacial–interglacial timescales.

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