Florida Current surface temperature and salinity variability during the last millennium

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[1] The sea-surface and temperature of the Florida Current are key parameters affecting the transport of heat into the North Atlantic, yet little is known about their variability on centennial timescales. Here we report replicated, high-resolution foraminiferal records of Florida Current surface hydrography for the last millennium from two coring sites, Dry Tortugas and the Great Bahama Bank. The oxygen isotopic composition of Florida Current surface water (δ18Ow) near Dry Tortugas increased 0.4‰ during the course of the Little Ice Age (LIA) (~1200–1850 A.D.), equivalent to a salinity increase of 0.8–1.5. On the Great Bahama Bank, where surface waters are influenced by the North Atlantic subtropical gyre, δ18Ow increased by 0.3‰ during the last 200 years. Although a portion (~0.1‰) of this shift may be an artifact of anthropogenically driven changes in surface water ΣCO2, the remaining δ18Ow signal implies a 0.4–1 increase in salinity after 200 years B.P. The simplest explanation of the δ18Ow data is southward migration of the Atlantic Hadley circulation during the LIA. Scaling of the δ18Ow records to salinity using the modern low-latitude δ18Ow–S slope produces an unrealistic reversal in the salinity gradient between the two sites. Only if δ18Ow is scaled to salinity using a high-latitude δ18Ow–S slope can the records be reconciled. Variable atmospheric 18O paralleled Dry Tortugas δ18Ow, suggesting that solar irradiance paced centennial-scale migration of the Inter-Tropical Convergence Zone and changes in Florida Current salinity during the last millennium.


1. Introduction

[2] The Little Ice Age was the most prominent climate event of the last millennium prior to the 20th century. Historical accounts describe severe cold conditions in Europe and North America at intermittent periods during the LIA [Grove, 1988; Bradley and Jones, 1993] and terrestrial paleoclimate data suggest the LIA was a global phenomenon [Bradley and Jones, 1993; Mann et al., 1999; Esper et al., 2002; Dahl-Jensen et al., 1998; Huang et al., 2000; Thompson, 1995; Chinn, 1996; Denton and Karlén, 1973]. Reconstructions of North Atlantic sea surface temperatures (SSTs) show that the eastern subtropical gyre cooled nearly 4°C during the LIA [deMenocal et al., 2000] while the Sargasso Sea cooled by ~1°C [Keigwin, 1996]. The western edge of the gyre apparently cooled by 1°–3°C relative to today during the coldest LIA intervals [Druffel, 1982; Winter et al., 2000]. In both the eastern gyre and Sargasso Sea locations, the LIA was preceded by a relatively warm interval from ~500 to 1200 A.D., the so-called “Medieval Warm Period”.

[3] The Yucatan Peninsula [Hodell et al., 2005] and the Cariaco Basin [Haug et al., 2001] experienced unusually dry Little Ice Age conditions. The records can be interpreted as a southward shift of the Atlantic Inter-Tropical Convergence Zone (ITCZ), either as a result of a greater meridional temperature gradient in the North Atlantic or persistent El Niño–like conditions in the equatorial Pacific [Haug et al., 2001]. Shifts in mean ITCZ position affect evaporation and precipitation in the tropical Atlantic and may act to stabilize perturbations to the Atlantic overturning circulation by influencing Gulf Stream salinity [Vellinga et al., 2002].

[4] Here we present evidence from high-resolution sediment cores that Florida Current salinity increased during the Little Ice Age. The simplest explanation for this pattern is a southward shift of the Atlantic ITCZ and increased evaporation minus precipitation (E – P) in the tropical Atlantic. These data support the suggestion that southward ITCZ migration caused dry LIA conditions in Central and South America [Haug et al., 2001; Hodell et al., 2005] and are consistent with simulations of increased aridity in the tropical North Atlantic in response to Northern Hemisphere cooling [Vellinga et al., 2002; Lohmann, 2003; Vellinga and Wu, 2004]. On the eastern edge of the Florida Current, δ18Ow was anomalously low during the Little Ice Age, perhaps because of southward movement of the Hadley circulation.

2. Oceanographic Setting

[5] We used four sediment cores for this study, two from each side of the Florida Current, the portion of the Gulf
Stream confined to the Straits of Florida. Cores KNR166-2-62MC-A and W167-79GGC were retrieved near Dry Tortugas, on the northwestern edge of the Florida Current (Figure 1). Cores KNR166-2-118MC-A and KNR166-2-125MC-D were retrieved from the Great Bahama Bank, on the eastern edge of Florida Current. Although separated in distance by only 400 km, the surface water at each location represents distinct oceanographic regimes. On the western side of the Florida Current, surface water is fresh because of the advection of low-salinity waters from the tropical Atlantic [Schmitz and Richardson, 1991]. Today, tropical Atlantic and Caribbean surface salinity is controlled largely by the excess rainfall relative to evaporation under the ascending branch of the Hadley circulation, the ITZC. Rainfall associated with the annual migration of the ITZC between 5°S and 15°N [Waliser and Gautier, 1993] is the primary reason surface water salinity is low in this region (Figure 1). This water is advected through the Caribbean and into the Straits of Florida via prevailing wind-driven and meridional overturning transports. Annual average salinity near Dry Tortugas is 36.1 with a seasonal and interannual range of approximately ±0.1 and ±0.2, respectively (J. A. Carton and B. S. Giese, SODA: A reanalysis of ocean climate, unpublished manuscript, 2006, available at http://www.atmos.umd.edu/%7Ecarton/pdfs/carton-giese06.pdf, hereinafter referred to as Carton and Giese, 2006).

3. Methods and Core Top Results

The Great Bahama Bank core sites reflect the influence of both the Florida Current and the Santaren Current, which is relatively salty because of the influence of the subtropical gyre. Higher salinities in the subtropical gyre reflect positive E–P conditions that occur under the dry, descending branch of the Hadley circulation and the convergence of surface water because of Ekman transports. Annual average salinity near Dry Tortugas is 36.2 (Carton and Giese, 2006). The seasonal and interannual range in salinity is approximately ±0.05 and ±0.15, respectively (Carton and Giese, 2006), somewhat smaller than the Dry Tortugas region.

3.1. Age Control

Down-core age control is based on multiple planktonic foraminiferal radiocarbon dates in each core (Table 1). Each date is based on an average of 8 mg G. ruber specimens from the >250 μm size fraction. All three multicore core tops in this study have >1 fraction modern radiocarbon, indicating that deposition occurred primarily after 1950 A. D. The one gravity core (79GGC) has a 14C core top age of 200 years [Lund and Curry, 2004], also indicating the presence of bomb radiocarbon, but to a lesser extent than the multicores. Raw radiocarbon ages were converted to calendar ages using CALIB 4.3 [Stuiver et al., 1998], assuming a surface ocean reservoir age of 400 years. Average sedimentation rates for the cores range from 21 to 42 cm/kyr (Table 1). We sampled 118MC and 125MC every 1 cm, equivalent to time intervals of 24 and 48 years, respectively.
Samples were taken every 0.5 cm in 62MC and 79GGC, or every 16 and 20 years, respectively.

3.2. Sampling Strategy

To address the potential bias of running δ18O and Mg/Ca on separate groups of foraminifera, we decided to use one common population of approximately 100 individual *G. ruber* specimens (212–250 μm size fraction) from each depth interval in 62MC, 118MC, and 125MC. Each sample was sonicated in methanol, siphoned, dried at 70°C, crushed, weighed, and then split for isotopic and Mg/Ca analyses. The crushed sample was evenly distributed on a tray, and split using a fine brush to ensure the subsamples contained approximately equal portions of fine and coarse material. The weight equivalent of 40 *G. ruber* tests was removed and then split into four separate subsamples that are representative of the *G. ruber* population in each sample. A similar comparison is not possible for 79GGC since the variance of δ18O analyses based on multiple measurements of 10 individuals is dominated by sampling rather than analytical uncertainty [Lund and Curry, 2004].

3.3. Stable Isotopes

Stable isotope analyses for 79GGC were run on a Finnigan MAT 252 coupled to a Kiel II carbonate device. Analyses for 62MC, 118MC, and 125MC were run on a Finnigan MAT 253 coupled to a Kiel III carbonate device. Calibration to VPDB scale was made using NBS-19 (δ13C = 1.95‰, 18O = 0±2.0‰). Reproducibility (1σ) of NBS-19 (n = 461) for the MAT 253 system is ±0.08‰ for δ18O and ±0.04‰ for δ13C. The 1σ values for NBS-19 on the MAT 252 were ±0.07‰ for δ18O and ±0.03‰ for δ13C (n > 2200) [Ostermann and Curry, 2000].

The isotope splits for 62MC, 118MC, and 125MC can be used to determine the degree of sample heterogeneity caused by the splitting procedure. If each isotope split represents the δ18O of the sample population, then the variance of the residual δ18O at each depth should be similar to the variance of the external standards. Residual values were determined by subtracting the mean δ18O value at each depth from the individual measurements. The standard deviation of residual values for the cores is 0.09‰, close to the analytical error for NBS-19. The similarity of these statistics indicates that our splitting procedure created subsamples that are representative of the *G. ruber* population. A Mg/Ca analysis was based on 50–60 individual foraminifera (62MC, 118MC, and 125MC). To minimize bias due to matrix effects [Rosenthal et al., 1999], we weighed each Mg/Ca split individually to ensure it fell between 150 and 200 μg. Each sample was cleaned using a sequential three-step process, including clay removal, metal oxide reduction, and organic matter oxidation [Boyle and Keigwin, 1985; Rosenthal et al., 1997]. All samples were run on a Finnigan Element 2 single collector ICP-MS.

3.4. Mg/Ca Analyses

The Mg/Ca analyses were based on either 50–60 individual foraminifera (79GGC) or their weight equivalent from a sample of 100 foraminifera (62MC, 118MC, and 125MC). To minimize bias due to matrix effects [Rosenthal et al., 1999], we weighed each Mg/Ca split individually to ensure it fell between 150 and 200 μg. Each sample was cleaned using a sequential three-step process, including clay removal, metal oxide reduction, and organic matter oxidation [Boyle and Keigwin, 1985; Rosenthal et al., 1997]. All samples were run on a Finnigan Element 2 single collector ICP-MS.

Samples were corrected for drift using a 2% HNO3 blank and a 50 ppm calcium standard. Analytical error was determined using three external standards with Mg/Ca ratios of 1.6, 3.2, and 4.8 mmol/mol. The analytical precision (1σ) for each standard was ±0.03 (n = 12), ±0.04 (n = 11), and ±0.05 mmol/mol (n = 11), respectively. Mg/Ca ratios for the unknowns were typically in the 4 to 5 mmol/mol range (Figures 2 and 3). The average precision (1σ) for Mg/Ca splits, which reflects both analytical and sample processing uncertainty, is ±0.12 mmol/mol (n = 82 pairs). Poorer precision for the unknowns primarily reflects sample heterogeneity due to the cleaning procedure. A Mg/Ca analytical precision of ±0.05 mmol/mol translates to an analytical temperature uncertainty of approximately ±0.1°C (1σ), whereas the combined analytical and sample processing uncertainty is ±0.1°C (1σ).
precision (±0.12 mmol/mol) corresponds to a SST uncertainty of ±0.25°C (1σ).

[14] We converted Mg/Ca values to sea surface temperature using the general calibration equation of Anand et al. [2003], which is based on sediment trap data from 10 planktonic foraminiferal species [Mg/Ca(mmol/mol) = 0.38*exp(0.09*SST(°C))]. We chose this equation because it is not subject to dissolution biases of core top calibration data sets and is therefore more appropriate for the shallow (<700 m) core sites discussed in this paper. We used the generalized sediment trap equation because the errors associated with its preexponential and exponential constants are smaller than for the G. ruber–specific calibration [Anand et al., 2003].

[15] Calculated Dry Tortugas core top SST values are approximately 27.5 (62MC) and 28.5°C (79GGC) (Figures 4b and 4e). The annual average sea surface temperature at these sites is 26.9°C (Carton and Giese, 2006). On the Great Bahama Bank, the core top Mg/Ca-derived SST for 118MC and 125MC is 28.0°–28.5°C (Figures 5b and 5e), compared to an annual average sea surface temperature of 27.2°C (Carton and Giese, 2006). Taking into account the ±1°C (1σ) calibration uncertainty [Anand et al., 2003], it appears that foraminiferal Mg/Ca produces reasonable absolute temperature estimates for these locations. All four core top values are warmer than observed SSTs, however, a discrepancy that is only amplified if G. ruber calcifies closer to 25 m water depth [Fairbanks et al., 1982; Deuser, 1987], where temperatures are ~0.5°C cooler than at the surface (Carton and Giese, 2006). The calibration based on multiple species may create a warm bias in the SST estimates. If we instead use the calibration for G. ruber, [Mg/Ca(mmol/mol) = 0.34*exp(0.102*SST(°C))] [Anand et al., 2003], the time series become 2°C cooler than in the general calibration case, and the core tops fall into better agreement with modern observations. The ~1°C offset between 62MC and 79GGC SSTs is likely due to the strong temperature gradient along the northern edge of the Florida Current which exposes 79GGC to warmer waters and yields higher Mg/Ca–derived SSTs. Spatial SST variability at this scale (~10 km) is not captured by the 0.5° × 0.5° gridding of the Carton and Giese (2006) database.

3.5. The δ18Ow Estimates

[17] We calculated past changes in δ18Ow using the isotopic fractionation versus temperature relationship from Bemis et al. [1998] as presented by Lynch-Stieglitz et al. [1999]. Similar results were obtained using the Kim and O’Neil [1997] paleotemperature equation. To evaluate whether the δ18Ow values are reasonable, we compare the core top δ18Ow to estimates of surface water δ18Ow based on modern δ18Ow–salinity observations. Since the Dry Tortugas sites are heavily influenced by tropical Atlantic surface waters, we used a δ18Ow/salinity regression based on data from the open ocean tropical Atlantic [δ18Ow(‰) = 0.26*S(psu) − 8.44] (G. A. Schmidt et al., Global seawater oxygen 18 database, 1999, available at http://www.giss.nasa.gov/data/o18data/, hereinafter referred to as Schmidt et al., 1999). Assuming annual average salinity values of 36.1 near Dry Tortugas (Carton and Giese, 2006), the predicted core top δ18Ow is 0.95‰, similar to the 62MC and 79GGC core top δ18Ow values of 0.8 ± 0.1‰ and 1.1 ± 0.2‰, respectively (Figures 4c and 4f). For the Great Bahama Bank sites, the annual average salinity (36.2 (Carton and Giese, 2006) and subtropical gyre δ18Ow/salinity regression [δ18Ow(‰) = 0.17*S(psu) − 5.13] (Schmidt et al., 1999) yield a modern δ18Ow of 1.02‰. This value is slightly higher than the 0.7 ± 0.1‰ estimate for the 118MC and 125MC core tops (Figures 5c and 5f).

[18] Although uncertainty in each core top and surface water δ18Ow value may account for some of the discrepancy, three of the four core top values underestimate surface water δ18Ow (62MC, 118MC, and 125MC). One potential source of systematic error is the use of crushed foraminiferal tests for stable isotope analyses. The δ18O values based on crushed foraminifera from 62MC are consistently 0.2‰ lower than δ18Ow values of whole foraminifera from the same samples (Figure 6). We observe no significant offset in δ13C between the whole and crushed samples, so it is unlikely due to fractionation of CO2 in the mass spectrometer. Keigwin and Jones [1990] found a similar pattern: the δ18O of ground foraminifera was 0.2‰ lower than the δ18O of whole foraminifera, with no obvious difference in δ13C values. This offset is likely due to either variable partition-
ing of oxygen isotopes among the products of the calcite-phosphoric acid reaction or diffusion of reaction products away from the particle surface associated with the production of \( \text{CO}_2 \) bubbles [Walters et al., 1972; Keigwin and Jones, 1990]. Given that no fractionation occurs when crushed samples are acidified under elevated pressure to prevent foaming, it appears that the fractionation is most likely tied to the rate at which gas evolves during acidification [Walters et al., 1972]. Swart et al. [1991] suggest that production of gas bubbles may encourage isotopic exchange between \( \text{CO}_2 \) and phosphoric acid, thus affecting the average isotopic value of the evolved carbon dioxide.

The apparent grain-size isotopic effect is analogous to temperature-driven changes in the fractionation factor. As the temperature and the reaction rate increase, the fractionation factor \( \left( \frac{\delta^{18}O_{\text{calcite}}}{\delta^{18}O_{\text{CO}_2}} \right) \) decreases, leading to lower \( \delta^{18}O \) values [e.g., Swart et al., 1991]. For reasons that remain unclear, faster reaction rates associated with crushed samples apparently lead to a greater proportion of \( \text{^16O} \) trapped as \( \text{CO}_2 \) and lower \( \delta^{18}O \). Therefore the low \( \delta^{18}O_w \) in 62MC relative to 79GGC are most likely a function of using crushed foraminiferal tests for 62MC \( \delta^{18}O \) analyses and whole tests for 79GGC (see section 3). The larger error envelope for 79GGC is due to the smaller number of foraminifera used for the stable isotope and Mg/Ca analyses and fewer replicate Mg/Ca measurements.

Figure 4. (left) The (a) \( \delta^{18}O_{\text{c}} \), (b) SST, and (c) \( \delta^{18}O_w \) data from KNR166-2-62MC, based on G. ruber (white) 212–250 \( \mu \text{m} \). (right) The (d) \( \delta^{18}O_{\text{c}} \), (e) SST, and (f) \( \delta^{18}O_w \) data for W167-79GGC. The plotting convention is the same for each of the six subplots, which includes average value at each depth (thin solid line), 100-year running mean (thick solid line), and 95% confidence limits for the 100-year mean (shaded area). SST estimates are based on Mg/Ca analyses and the general temperature calibration of Anand et al. [2003], and \( \delta^{18}O_w \) was calculated using the equation of Bemis et al. [1998]. The \( \delta^{18}O_w \) in each core was highest during the Little Ice Age. Calendar-calibrated radiocarbon dates (triangles) are given in the \( \delta^{18}O_w \) plots. Despite being located only 10 km apart, 62MC \( \delta^{18}O_w \) is approximately 0.2% lower than 79GGC \( \delta^{18}O_w \). This offset is most likely a function of using crushed foraminiferal tests for 62MC \( \delta^{18}O \) analyses and whole tests for 79GGC (see section 3). The larger error envelope for 79GGC is due to the smaller number of foraminifera used for the stable isotope and Mg/Ca analyses and fewer replicate Mg/Ca measurements.

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guishable from the modern seawater estimate (0.95‰). Adding 0.2‰ to the 118MC and 125MC data yields core top \( \delta^{18}O_w \) values of 0.9 ± 0.1‰, also very similar to the seawater estimate for these sites (1.02‰).

4. Down-Core Results and Discussion

4.1. Surface Water \( \delta^{18}O_c \)

[20] Planktonic \( \delta^{18}O_c \) and SST estimates from 62MC and 79GGC yield consistent changes in Florida Current \( \delta^{18}O_w \) during the last 1000 years. In 62MC, \( G. \ ruber \) \( \delta^{18}O_c \) increased by approximately 0.2‰ over the past 900 years, with the largest change occurring from 400 years B.P. to present (Figure 4a). The \( \delta^{18}O_c \) increase is significant at the 95% confidence level and implies that Florida Current surface waters became progressively cooler, saltier, or both. Mg/Ca analyses of \( G. \ ruber \) indicate that SST was stable at 24.5°C from 900 to 500 years B.P., and then increased monotonically to ~25.5°C by 200 years B.P. (Figure 4b). This SST maximum was followed by an abrupt 1°C decrease from 200 to 100 years B.P., and then a steady increase over the past 100 years. Reconciliation of the \( \delta^{18}O_c \) and SST records requires \( \delta^{18}O_w \) variability on the order of 0.4‰ (Figure 4c). Foraminiferal \( \delta^{18}O_c \) in nearby core 79GGC was nearly constant from 1400 to 500 years B.P., while those prior to 700 years B.P. approached modern values. The \( \delta^{18}O_w \) data imply that modern surface water at these sites is anomalously salty in the context of the last 1200 years.

Figure 5. (left) The (a) \( \delta^{18}O_c \), (b) SST, and (c) \( \delta^{18}O_w \) data from KNR166-2-118MC, based on \( G. \ ruber \) (white) 212–250 μm. (right) The (d) \( \delta^{18}O_c \), (e) SST, and (f) \( \delta^{18}O_w \) data for KNR166-2-125MC. The plotting convention is the same as in Figure 4, except that the running average for 125MC was calculated using a 200-year window to compensate for the lower sedimentation rate for this core. The wider error envelope for 125MC reflects the lower data density, even though the smoothing window used was twice as wide as for 118MC. The \( \delta^{18}O_c \) in both cores increased by 0.2‰ over the last 300 years. Mg/Ca–derived SSTs were lower than today from 700 to 300 years B.P., while those prior to 700 years B.P. approached modern values. The \( \delta^{18}O_w \) data imply that modern surface water at these sites is anomalously salty in the context of the last 1200 years.
minifera for the stable isotopic analyses in these two cores (section 3).

[22] On the Great Bahama Bank, significant changes in δ¹⁸Ow and SST occurred during the past millennium, but with different timing than the Dry Tortugas sites. The largest change in 118MC δ¹⁸Ow occurred from 200 years B.P. to present (Figure 5a). A similar shift occurred about 100 years earlier in the 125MC record (Figure 5d). Since 118MC has better sampling resolution and chronological control than 125MC, the δ¹⁸Ow increase most likely occurred after 200 years B.P. Mg/Ca–derived SSTs in 118MC increased gradually by 0.5°C from 500 years B.P. to present (Figure 5b). Core 125MC extends the Great Bahama Bank SST record to 1400 years B.P. and shows a ~0.5°C cooling at 800 years B.P., near the beginning of the LIA, and then a ~0.5°C warming at ~300 years B.P. (Figure 5e). The combined δ¹⁸Oc and SST records require that δ¹⁸Ow along the Great Bahama Bank increased by ~0.3‰ during the past 200 years (Figures 5c and 5f).

4.2. Sea Surface Temperatures

[23] The Great Bahama Bank results are inconsistent with foraminiferal SST estimates from the Sargasso Sea and the magnitudes of LIA cooling inferred from Caribbean corals. Rather than increasing after 400 years B.P., SSTs in the Sargasso Sea apparently decreased [Keigwin, 1996]. This discrepancy may be due to the use of different SST proxies: δ¹⁸Oc for the Sargasso Sea and Mg/Ca for the Great Bahama Bank. The data presented here suggest that high Sargasso Sea δ¹⁸Oc after 200 years B.P. may have been a result of high δ¹⁸Ow rather than low SSTs. Estimates from Caribbean corals indicate SSTs were 1°C–3°C cooler than today during the coldest LIA intervals [Winter et al., 2000], significantly larger than the cooling along the Great Bahama Bank. The disagreement between the coral and Bahamas SST estimates may be due to the brief (~10 years) time windows covered by the corals and signal smoothing in the foraminiferal records caused by bioturbation.

[24] Interestingly, both Great Bahama Bank and Dry Tortugas SSTs warmed by ~0.5°C during the course of the LIA, from about 800 years B.P. to 200 years B.P. (Figure 7). As a result, the ~1.0°C cross-current SST gradient remained nearly constant during this time. Prior to 800 years B.P. and after 200 years B.P., however, the cross-current gradient was 1.5°C–2.0°C, significantly larger than during the LIA interval. We believe this most likely reflects variable formation of Tortugas eddies, which today can shift SSTs at the 62MC and 79GGC core sites by >3°C for months at a time [Fratantoni et al., 1998] (see also APL Ocean Remote Sensing, Gulf of Mexico region, available at http://fermi.jhuapl.edu/avhrr/gm/index.html). Therefore the LIA may have been characterized by reduced Tortugas eddy formation relative to the preceding and subsequent time intervals. This inference is opposite that based only on δ¹⁸Oc from the Dry Tortugas sites [Lund and Curry, 2004]. The apparent SST differences in 62MC and 79GGC over the past 600 years (Figure 7) are not significant at the 95% confidence level (Figures 4b and 4e).

4.3. SST and δ¹⁸Ow Error Analysis

[25] We estimated the 95% confidence limits for SST in two ways (1) based on variance in the raw Mg/Ca data only, and (2) based on both the sample variance and calibration uncertainty. In the latter case we used a Monte Carlo approach that accounts for both measurement error and uncertainty in the preexponential and exponential constants for the general Mg/Ca–SST calibration [Anand et al., 2003]. The resulting 95% confidence limits of approximately ±2°C (not shown) are dominated by the calibration uncertainty. Given the down core SST variability can be replicated to within 0.5°C (Figure 7), we believe incorporating the absolute calibration uncertainty yields artificially conservative errors for estimating relative changes in SST. The calibration error, which largely reflects uncertainty in the calcification temperatures used for calibration [Anand et al., 2003], appears to be most useful for comparing core top SST estimates with modern observations.

[26] The error envelopes for the δ¹⁸Ow time series in Figures 4 and 5 are based on Monte Carlo estimates, where at each depth in each core, δ¹⁸Ow and SST values were randomly sampled from a distribution constrained by the 95% confidence limits for δ¹⁸Ow and SST. For example, the error for 62MC δ¹⁸Ow (Figure 4e) is based on random data from within the error envelopes in Figures 4a and 4b. This approach, which excludes calibration error, yields 95% confidence limits for δ¹⁸Ow of approximately ±0.1‰. If Mg/Ca–SST calibration error is included, then the uncertainty for δ¹⁸Ow expands to approximately ±0.4‰ (95% confidence limit), similar to that calculated by Schmidt et al. [2004a] for data from the Caribbean.

[27] Whether the δ¹⁸Ow variability in 62MC is significant depends on the method used to calculate confidence limits. If we incorporate the calibration error, then the δ¹⁸Ow signal does not exceed the 95% confidence level. At the one-sigma
level (± 0.2‰), however, the $^{18}$O$_{w}$ increase from 700 to 200 years B.P. is significant (Figure 4c). If we choose to exclude the Mg/Ca–SST calibration error, then smaller-scale features, such as high $^{18}$O$_{w}$ at ~200, 400, and 550 years B.P., begin to appear. Similar arguments can be made for 79GGC, 118MC, and 125MC. Overall, the ±0.4‰ uncertainty for $^{18}$O$_{w}$ appears to be too conservative for these core sites. For the remainder of this paper, we assume the down core errors for $^{18}$O$_{w}$ (and SST) are better represented by the confidence limits based on sampling uncertainty (Figures 4 and 5). Although this is not entirely satisfactory, the issue can only be resolved through refinement of the Mg/Ca–SST calibration and verification of the records presented here using additional high-resolution $^{18}$O$_{w}$ time series. We are encouraged, however, by the strong similarity of the replicate $^{18}$O$_{w}$ records at both Dry Tortugas and the Great Bahama Bank (section 5.2).

### 4.4. Carbonate Ion and $^{13}$C Suess Effects

Great Bahama Bank $^{18}$O$_{c}$ (and hence calculated $^{18}$O$_{w}$) may be influenced by decreasing surface water pH associated with the input of anthropogenic CO$_2$ over the past 200 years. As pH decreases, the greater proportion of bicarbonate to carbonate ion leads to isotopically heavier dissolved inorganic carbon, since bicarbonate is the more enriched of the two species [Zeebe, 1999]. In laboratory culturing experiments, *G. ruber* $^{18}$O$_{w}$ varies systematically with [CO$_3^-$] by a slope of $-0.002 \pm 0.001\%$/μmol/kg) [Spero et al., 1999]. During the last 200 years, surface North Atlantic subtropical gyre dissolved inorganic carbon (DIC) has increased ~60 μmol/kg because of anthropogenic CO$_2$ emissions [Sabine et al., 2004]. Assuming constant near-surface alkalinity between the LIA and present (2400 μEq/kg; GEOSECS station 31), we estimate that [CO$_3^-$] decreased by approximately 40 μmol/kg over this time interval (see http://geosci.uchicago.edu/~archer/cgimodels/pco2.html), equivalent to a carbonate ion isotope effect on *G. ruber* $^{18}$O$_c$ of +0.08 ± 0.04‰. Therefore approximately 0.1‰ of the $^{18}$O$_{w}$ increase from the LIA to today may be attributed to decreasing surface water pH. If this is true, then the remainder (0.2‰) reflects increasing salinity of the surface North Atlantic subtropical gyre since the LIA.

Since carbonate ion concentration affects both the oxygen and carbon stable isotopes of foraminiferal calcite, it should be possible to verify the effect of pH on $^{18}$O$_{w}$ using $^{13}$C$_{c}$. Using a [CO$_3^-$] decrease of 40 μmol/kg and a *G. ruber* $^{18}$O$_{c}$/[CO$_3^-$] relationship of $-0.009 \pm 0.001\%$/μmol/kg) [Spero et al., 1999], the expected shift in $^{13}$C$_{c}$ from LIA to present would be +0.35‰, similar to that observed in 118MC and 125MC (Figure 8 top). However, if *G. ruber* $^{13}$C$_{c}$ reliably records $^{13}$C of dissolved inorganic carbon ($^{13}$C$_{DIC}$), it should reflect not only pH variability but also the influence of $^{13}$C-depleted CO$_2$ from fossil fuel and biomass burning (the $^{13}$C Suess effect). Surface Atlantic $^{13}$C$_{DIC}$ decreased by 0.8‰ over the past 200 years [Bohm et al., 1996], consistent with model estimates of mixed layer $^{13}$C$_{DIC}$ driven by decreasing $^{13}$C of atmospheric CO$_2$ [Francey et al., 1999]. Taking into account both the carbonate ion and Suess effects, the change in $^{13}$C$_{c}$ from 200 years B.P. should be approximately $-0.4\%$, as opposed the observed signal of +0.3–0.4‰.

The 0.8‰ offset between predicted and observed $^{13}$C$_{c}$ is either due to real changes in $^{13}$C$_{c}$ or is an artifact of *G. ruber* vital effects. Increasing surface layer $^{13}$C$_{DIC}$ since the LIA would likely enhance biological carbon isotope fractionation and increase $^{13}$C$_{c}$ [Jasper and Hayes, 1990; Hofmann et al., 1999]. An increase in export production of $^{13}$C-depleted organic matter could also enrich mixed layer $^{13}$C of DIC. The situation is further complicated by the tendency of *G. ruber* $^{13}$C$_{c}$ to be lower than ambient seawater $^{13}$C$_{DIC}$. This offset is most likely a function of foraminiferal respiration, metabolic and growth rates, and symbiont photosynthesis, all of which are ecologically sensitive to oceanographic variability [Curry and Matthews, 1981; Spero and Lea, 1993; Russell and Spero, 2000]. Regardless of the exact cause, it is clear *G. ruber* $^{13}$C$_{c}$ on its own cannot be used to validate the effect of pH on $^{13}$C$_{c}$.

The problematic nature of using *G. ruber* $^{13}$C$_{c}$ as a proxy for $^{13}$C$_{DIC}$ is further highlighted by the clear existence of the $^{13}$C Suess effect in benthic foraminiferal $^{13}$C records from the Great Bahama Bank (Figure 8 bottom). At water depths ranging from 250 to 700 m, four separate benthic time series show anomalous $^{13}$C depletion during the past 200 years. The decrease in $^{13}$C$_{c}$ near the core tops is unprecedented during the last millennium and the largest $^{13}$C depletion occurs in the shallowest cores.
of the Mississippi River and altered routing of oceanic currents may have played a minor role, the observed \(\delta^{18}O_w\) signal is too large to be explained by these mechanisms. Instead, southward migration of the Atlantic Hadley circulation during the Little Ice Age, in combination with a mean shift in thermocline \(\delta^{18}O_w\) are the most likely mechanisms responsible for changes in \(\delta^{18}O_w\) near Dry Tortugas and the Great Bahama Bank.

5.1. Local/Regional \(\delta^{18}O_w\) Effects

[33] Input of isotopically light Mississippi River water may have influenced the Dry Tortugas \(\delta^{18}O_w\) records, but only to a small extent. Today, the Mississippi River has an annual average flow rate of 0.015 Sv [Walker et al., 1994] and a \(\delta^{18}O_w\) signature of \(-7\)‰ SMOW [Coplen and Kendall, 2000]. Transport in the upper 100 m of the Florida Current is approximately 9 Sv [Schmitz and Richardson, 1991] with a \(\delta^{18}O_w\) signature of \(-0.9\)‰ (section 3).

[34] If the isotopic signature of the oceanic and fluvial end-members is held constant, the \(\delta^{18}O_w\) increase in 62MC from 700 to 200 years B.P. (0.4‰) would require Mississippi discharge to change by a factor of 33 (e.g., 0.015 Sv at 200 years B.P. and 0.5 Sv at 700 years B.P.). During the 2-month peak of the 1993 flood event, one of the most severe floods of the last century, the Mississippi discharge rate was 0.03 Sv [Walker et al., 1994], a doubling of the average rate, making even the 1993 flood too small to influence Florida Current \(\delta^{18}O_w\) in a significant way. Given the unrealistic river flows required to create the observed changes in \(\delta^{18}O_w\) and that these flow rates would have to persist for centuries, it is highly unlikely that variability in Mississippi River discharge significantly influenced Florida Current \(\delta^{18}O_w\) during the last millennium.

[35] On the Great Bahama Bank, variable current routing may have affected surface water \(\delta^{18}O_w\) because the 118MC and 125MC core sites are located near the confluence of the Santaren and Florida Currents. Today, the Santaren Current supplies on average approximately 2 Sv of warm, salty water to the main flow of the Florida Current [Atkinson et al., 1995; Leaman et al., 1995]. Lower Santaren Current flow during the LIA would likely result in lower salinity at the core sites. Despite the 0.2 psu difference in salinity on either side of the Florida Current, their \(\delta^{18}O_w\) values are nearly equal (section 3). Therefore reduced flow of the Santaren Current relative to the Florida Current during the LIA would not necessarily lead to lower surface water \(\delta^{18}O_w\) along the Great Bahama Bank. If anything, increased Florida Current salinity during the LIA should lead to saltier conditions at the 118MC and 125MC core sites. What is remarkable is that LIA salinities on the GBB are anomalously low relative to core top values, despite the proximity of the cores to the Florida Current (Figure 1).

5.2. Hadley Cell Migration

[36] The salinity changes on the Great Bahama Bank were very different than those at Dry Tortugas during the past 1200 years. From 400 to 200 years B.P., \(\delta^{18}O_w\) was higher than today near Dry Tortugas, it was lower than today on the Great Bahama Bank (Figure 9). Beginning at 200 years B.P., \(\delta^{18}O_w\) increased by \(-0.2\)‰ to \(-0.3\)‰ at Dry

5. Sources of Salinity Variability

[12] Florida Current \(\delta^{18}O_w\) variability during the last millennium was most likely due to large-scale changes in oceanic precipitation and evaporation. While variable flow...
The largest cross-current difference in δ¹⁸O₇ occurred during the Little Ice Age, while the smallest offsets occurred from approximately 0 to 100 years B.P. and 900 to 1200 years B.P. North-south movement of the Atlantic Hadley circulation is the most likely explanation of the observed δ¹⁸O₇ variability.

[37] Given that Florida Current surface salinity is heavily influenced by E – P in the tropical Atlantic, reduced precipitation in this region during the Little Ice Age is the most logical cause of the observed changes. This interpretation is consistent with anomalously dry LIA conditions in the Yucatan Peninsula [Hodell et al., 2005] and northern South America near the Cariaco Basin [Haug et al., 2001]. Comparison of Dry Tortugas δ¹⁸O₇ to the aridity record from the Cariaco Basin suggests a strong climatic link between the two (Figure 10). Cariaco Basin percent Ti, which is proportional to fluvial input, shows a long-term decrease beginning ~600 years B.P., with distinct minima clustered between 400 and 150 years B.P. [Haug et al., 2001]. The long-term drying of the Cariaco Basin region ended abruptly at 200 years B.P. Overall, the changes observed in 62MC and 79GGC δ¹⁸O₇ parallel those in Cariaco Basin percent Ti. One mechanism that could cause this common change is southward ITZ migration, which would dry the watershed for the Cariaco Basin [Haug et al., 2001], and remotely force increased salinity in the Florida Current.

[38] On an interannual basis, southward migration of the Atlantic ITZ can be caused by either an increased Atlantic cross-equatorial SST gradient or El Niño conditions in the equatorial Pacific [Chiang et al., 2002]. Even slight (~0.2°C) cooling of the North Atlantic can force southward migration of the ITZ, which in turn reduces precipitation in the tropical North Atlantic and creates a positive salinity anomaly in this region [Vellinga et al., 2002; Vellinga and Wu, 2004]. The Florida Current δ¹⁸O₇ maxima and Cariaco Basin percent Ti minima from 400 to 150 years B.P. coincide with an interval of unusually cool conditions in several Northern Hemisphere records [Lamb, 1995; deMenocal et al., 2000; Dahl-Jensen et al., 1998; Huang et al., 2000]. This correspondence suggests that LIA cooling in the North Atlantic enhanced the cross-equatorial SST gradient and triggered a southward shift in mean ITZ position.

[39] Beginning at ~900 years B.P., δ¹⁸O₇ increased on the Florida Margin and decreased on the Great Bahama Bank, potentially reflecting southward migration of the Hadley cell and increased E – P in the tropical Atlantic and decreased E – P in the subtropical gyre. Several paleoclimate records indicate cooling in the North Atlantic region at about this time [Dahl-Jensen et al., 1998; Esper et al., 2002; deMenocal et al., 2000]. The sharp decrease in δ¹⁸O₇ after 200 years B.P. near Dry Tortugas and increase on Great Bahama Bank is consistent with warming since the LIA and an abrupt movement of the Hadley circulation to the north, causing relatively higher salinity in the subtropical gyre and lower salinity in the tropical Atlantic.

5.3. Surface Ocean Versus Thermocline δ¹⁸O₇-S

[40] The modern relationship between δ¹⁸O₇ and salinity in the tropical and subtropical Atlantic surface waters is described by a δ¹⁸O₇/S slope of approximately 0.2‰ per unit salinity change (Figure 11a). In the subtropical gyre, δ¹⁸O₇ is high relative to the tropics since evaporation discriminates against ¹⁸O. The resulting ¹⁶O-enriched vapor rains out in the tropics and high-latitude Atlantic. At high latitudes, δ¹⁸O decreases because of the preferential condensation of ¹⁸O at cooler temperatures, making the δ¹⁸O₇/
Slope, this would imply LIA salinities of 36.5 at Dry Tortugas and 35.7 at the Bahamas (Figure 11b). This is unrealistic since it requires the surface North Atlantic subtropical gyre to be fresher than the tropical Atlantic (i.e., a reversal of the Hadley cell distribution of E – P). A more reasonable scenario involves southward migration of the Hadley circulation and its associated E – P patterns (Figure 11c). In this case, the tropical Atlantic becomes saltier along the 0.2‰ slope, but the subtropical gyre becomes fresher through greater influence of 18O-depleted surface water from higher latitudes (equivalent to the thermocline 18O/S line). The situation depicted in Figure 11c allows 18O at Great Bahama Bank to be lower than Dry Tortugas and minimizes the LIA salinity difference between the two sites. Although less extreme than the first case, this scenario still requires a salinity gradient the reverse of today.

Surface water 18O is not only a function of evaporation and precipitation, but also the 18O of water with which it mixes. For the surface North Atlantic, this mixing component is the thermocline. If thermocline 18O and salinity varied during the LIA, it would likely be along its modern slope of 0.5‰. Since the thermocline dominates the surface layer from a volumetric standpoint, the entire low-latitude surface Atlantic 18O/S relationship would also move along a thermocline slope to retain mass balance [Craig and Gordon, 1965]. In this way, a 0.1‰ increase in mean thermocline 18O could account for the 0.1‰ increase at Dry Tortugas during the LIA (Figure 11d). The thermocline scenario would also require a positive shift in Great Bahama Bank 18O, however, opposite that observed.

The scenario most consistent with the sign and magnitude of 18O variability in our records involves a combination of the thermocline and Hadley circulation models (Figure 11c). Higher mean thermocline 18O reduces the implied LIA salinity change at Dry Tortugas and southward migration of the Hadley cell and provides a mechanism for introducing 18O-depleted surface waters into the subtropical gyre. This approach avoids the pitfall of a reversed salinity gradient by moving each site in opposite directions along lines with 0.5‰ slopes. In fact, the only way to accommodate the opposing changes in 18O on either side of the Florida Current is to invoke the influence of high-latitude and thermocline waters. Using today’s low-latitude 18O-S relationship to estimate past changes in salinity yields unreasonable results (Figure 11b). If we use instead the high-latitude slope, then LIA salinities at Dry Tortugas and Great Bahama Bank were nearly equal. This does not mean that salinities in the subtropical gyre and tropical Atlantic were also equal. Today the salinity offset between these regions is on the order of ~1, whereas the offset between the coring sites is only ~0.3. During the LIA therefore the salinity gradient between the tropics and subtropical gyre was likely lower than today, but probably not zero.

5.4. El Niño Southern Oscillation

Of course, the North Atlantic does not act independently of the remainder of the climate system. Anomalous
Figure 11. (a) Modern $\delta^{18}O_{w}$-S regressions for the North Atlantic, including the tropics, subtropical gyre, and thermocline (data from G. A. Schmidt et al., Global seawater oxygen 18 database, 1999, available at http://www.giss.nasa.gov/data/o18data/). Since the thermocline is ventilated at midlatitudes to high latitudes, this line also represents the high-latitude surface North Atlantic. On the basis of modern annual average salinity of 36.1 the Dry Tortugas sites plot along the tropical relationship ($\delta^{18}O_{w} = 0.95$ (shaded circle)) where $E - P$ (evaporation minus precipitation) is negative. The Great Bahama Bank sites, with average salinity of 36.2, plot along the subtropical gyre line ($\delta^{18}O_{w} = 1.02$ (shaded square)), where $E - P$ is positive. Today’s average $\delta^{18}O_{w}$-S for North Atlantic Deep Water (NADW) is noted with the solid square [Craig and Gordon, 1965]. (b) Little Ice Age $\delta^{18}O_{w}$ values at Dry Tortugas (solid circle) and Great Bahama Bank (solid square). Dry Tortugas values were $\sim$0.1‰ heavier than today, while those on the Great Bahama Bank were $\sim$0.2‰ lighter. Using the modern low-latitude slopes, this scenario implies the Dry Tortugas sites were saltier than the Bahamas, an unrealistic possibility. (c) Southward Hadley cell migration case (see text). Dry Tortugas values increase along the low-latitude line as before, but now the Bahamas are influenced by a greater proportion of high-latitude $^{18}O$-depleted water and therefore move along the 0.5‰ slope. The LIA salinity values remain reversed but to a lesser extent than the previous scenario. (d) An increase in the mean thermocline $\delta^{18}O_{w}$ of 0.1‰ moving all the lines upward along the 0.5‰ slope, easily accounting for the LIA Dry Tortugas $\delta^{18}O_{w}$ (solid circle) but also causing the Bahamas to increase by 0.1‰ (solid square), opposite the observed change. (e) A combination of the models in Figures 11c and 11d), where the mean thermocline $\delta^{18}O_{w}$-S increases and the Hadley circulation migrates southward. This is the most realistic model for the data as it eliminates the pitfall of a reversed salinity gradient.
atmospheric convection in the eastern equatorial Pacific during El Niño events forces subsidence and decreased precipitation over the tropical North Atlantic [Chiang et al., 2002]. El Niño events also cause the main locus of rainfall to migrate into the central equatorial Pacific, away from its more typical location in the western equatorial Pacific (WEP) [see http://www.cdc.noaa.gov/ENSO]. Coral δ¹⁸O records show a decreased equatorial Pacific zonal SST gradient during the LIA, possibly caused by more frequent El Niño events during this time [Hendy et al., 2002; Cobb et al., 2003]. If the interannual model can be applied to longer timescales, El Niño–like conditions would lead to decreased precipitation in the northern tropical Atlantic, as is observed, but also in the WEP. Available evidence shows that this did not occur. Low δ¹⁸Ow in the WEP during the LIA [Stott et al., 2004] is inconsistent with the simple interannual model (Figure 10). Either ENSO played little role in Atlantic ITCZ movement or the MD81 δ¹⁸Ow record reflects climatic processes other than ENSO.

5.5. Role of Solar Variability

[45] Reduced sunspot counts [Eddy, 1976], higher atmospheric Δ¹⁴C (Δ¹⁴Catm) [Stuiver et al., 1991] and greater δ¹⁸O Be flux [Bard et al., 2000] all imply solar irradiance was lower than today during the LIA. High Δ¹⁴Catm during the Maunder, Spörer, and Wolf sunspot minima matches intervals of high Florida Current δ¹⁸Ow and low western equatorial Pacific δ¹⁸Ow (Figure 10). The δ¹⁸Ow records appear to lag Δ¹⁴Catm by 50–100 years, assuming surface reservoir ages were similar to modern values over the past 1200 years. A similar lag is observed when comparing δ¹⁸Ow with modeled estimates of radiocarbon production rate [Bond et al., 2001; Marchal, 2005]. If this lag is correct, then it appears that solar variability paced not only migration of the North Atlantic subpolar front [Bond et al., 2001] but also centennial-scale hydrologic variability in both the tropical Atlantic and Pacific.

[46] Numerical simulations of the evaporation-precipitation (E – P) response to reduced solar irradiance lack the positive E – P anomaly in the tropical North Atlantic and negative E – P anomaly in the tropical South Atlantic characteristic of southward ITCZ migration [Rind and Overpeck, 1993]. An imposed 1°C–2°C cooling across the entire North Atlantic is capable of creating the tropical Atlantic E – P dipole pattern, however [Rind and Overpeck, 1993]. Dynamics of the ocean and atmosphere coupling not yet included in solar irradiance models may create very different temperature and E – P anomalies. Reduced northward surface heat transport associated with weaker meridional overturning or wind-driven subtropical gyre circulation may provide the cooling necessary to force ITCZ movement. Available paleoclimate evidence for reduced North Atlantic Deep Water production during the LIA is equivocal [Keigwin and Boyle, 2000].

6. Conclusions

[47] The surface hydrography of the Florida Current was highly variable during the last millennium. Near Dry Tortugas, salinity increased by 0.8–1.5 during the course of the Little Ice Age. Similar although larger changes in tropical Atlantic salinity and Northern Hemisphere temperature are observed on glacial-interglacial and millennial timescales [Schmidt et al., 2004a, 2004b]. Thus this pattern of climate variability is not necessarily a manifestation of large continental ice sheets as it can occur under climatic boundary conditions very similar to today. Variable solar irradiance apparently paced shifts in Florida Current salinity during the last millennium, implying the transport of salt into the North Atlantic is sensitive to modest changes in solar forcing on centennial timescales. The lack of evidence for variable Atlantic overturning during the LIA [Keigwin and Boyle, 2000] suggests that southward ITCZ migration and increased Gulf Stream salinity may stabilize changes in North Atlantic Deep Water production, as is predicted by model studies [Vellinga et al., 2002; Vellinga and Wu, 2004].

[48] On the Great Bahama Bank, where surface waters reflect the influence of the North Atlantic subtropical gyre, SST and δ¹⁸Ow records display different variability than their Dry Tortugas counterparts. Sea surface temperatures decreased by 0.5°C at the beginning of the LIA, consistent with the timing of cooling in several Northern Hemisphere temperature records. SSTs increased by 0.5°C about 200 years prior to the LIA termination, however, suggesting a more heterogeneous temperature anomaly pattern at the end of the LIA. If δ¹⁸Ow on the Great Bahama Bank represents primarily salinity rather than anthropogenically driven pH effects, then surface salinity increased ~0.4 during the last 200 years, making the modern values the saltiest of the last millennium.

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