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Late Holocene variability of upper North Atlantic Deep Water temperature and salinity

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[1] Magnesium/calcium ratios in benthic foraminifera (*Cibicidoides pachyderma*) from a sediment core on the Laurentian Slope (1854 m) exhibit strong millennial-scale fluctuations during the past 4000 years. We convert these data to seawater paleotemperatures using a new monospecific linear equation. Results suggest that the temperature of upper North Atlantic Deep Water (dominated by Labrador Seawater today) has varied by at least 2°C during the late Holocene. Millennial scale coolings coincide with previously identified periods of increased drift ice and regional glacier advances, including the Little Ice Age. Paired oxygen isotope measurements indicate that salinity and perhaps density were reduced during the cold periods. We discuss possible mechanisms for transmitting this cold, fresh signal from surface waters to intermediate depths. Our reconstructed late Holocene ranges in upper North Atlantic Deep Water properties greatly exceed those of the instrumental record and imply that large changes may be yet to come.

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1. Introduction

[2] The warm climate of the present Holocene epoch was punctuated by a series of widespread, millennial scale cooling events. In the subpolar North Atlantic, petrologic tracers of ice rafted debris (IRD) document nine multicentury episodes of southward expansion of drift ice and hence polar waters since the end of the Younger Dryas cold period (∼11,500 cal BP) [*Bond et al.*, 1997, 2001]. Each event was accompanied by sea surface cooling and freshening as inferred from planktonic foraminiferal assemblages and $\delta^{18}O$. The most recent of these events was the Little Ice Age (LIA) [*Jennings and Weiner*, 1996; *Bond et al.*, 2001], which occurred in several stages between roughly AD 1300 and 1870 [*Grove*, 2001]. It was marked by alpine glacier advances throughout

Europe and North America and is well documented in historical records of unusually harsh winters, increased sea ice, and poor crop yields [*Denton and Karlén*, 1973; *Grove*, 1988]. Earlier Holocene IRD peaks appear to correspond to additional alpine glacier expansions [*Denton and Karlén*, 1973], more winter-like atmospheric conditions over Greenland [*O'Brien et al.*, 1995], strengthening of the cool Canary Current off western Africa [*deMenocal et al.*, 2000], and suppression of the southwest Indian monsoon [*Gupta et al.*, 2003].

[3] *Bond et al.* [1997] further demonstrated that the quasiperiodicity of ice rafting events during the Holocene is indistinguishable from that during glacial times ($\sim$1500 ± 500 years), indicating that these events represent a pervasive mode of late Quaternary climate variability. Under glacial conditions, when North Atlantic Deep Water (NADW) formation was apparently close to a stability threshold (due to overall lower sea surface salinities), the freshwater (iceberg) discharge associated with the 1–2 kyr cycle might have been enough to switch NADW between strong and weak modes [*Stocker and Wright*, 1991; *Rahmstorf*, 1995]. Changes in meridional heat transport associated with such switches have been invoked to explain the rapid warmings and coolings (Dansgaard/Oeschger cycles) characteristic of the last glacial period [*Broecker et al.*, 1985; *Keigwin and Boyle*, 1999; *Rahmstorf*, 2002; *Schmittner et al.*, 2002]. This begs the question whether or not smaller changes in NADW formation occurred during the Holocene, when the system was more stable. Recently, *Bond et al.* [2001] proposed that solar forcing drove the 1–2 kyr Holocene cycle, and they additionally suggested that variable thermohaline overturning may have been an important amplifier. Several studies have attempted to constrain millennial scale NADW variability during the Holocene, but results have been equivocal, with possible reductions during some cold events [*Bond et al.*, 1997; *Bianchi and McCave*, 1999; *Keigwin and Boyle*, 2000; *Oppo et al.*, 2003]. All of this work to date has focused on the denser components of NADW, near Arctic overflow regions and in the deep (>4000 m) western North Atlantic basin.

[4] Here we investigate the late Holocene history of upper NADW, which is presently dominated by Labrador Sea Water (LSW). LSW accounts for about half of the net transformation of warm waters to cold waters in the North Atlantic today [*Schmitz and McCartney*, 1993] and can be traced as a low-salinity, well-oxygenated, low-potential vorticity water mass between $\sim$1000 and 2000 m depth [*Talley and McCartney*, 1982] (Figure 1). The formation and properties of LSW are known to be very sensitive to modern surface ocean conditions. For example, fresh water imported into the Labrador Sea by the "Great Salinity Anomaly" of the late 1960s (combined with diminished winter storm activity) reduced the depth of Labrador Sea convection to less than 1000 m, allowing LSW to become warmer and saltier through lateral mixing [*Lazier*, 1995; *Dickson et al.*, 1996]. Intensification of LSW renewal in the early 1970s and again in the late 1980s led to intermediate water cooling and freshening. Such changes in LSW properties impact mid-depth waters throughout the North Atlantic within 4–6 years [*Sy et al.*, 1997; *Curry et al.*, 1998].

[5] To expand this temperature and salinity history to centennial/millennial timescales we use benthic foraminiferal Mg/Ca and δ^{18}O. The Mg content of benthic foraminiferal calcite has been shown to increase with calcification temperature [*Rosenthal et al.*, 1997; *Martin et al.*, 2002; *Lear et al.*, 2002]. By combining Mg-derived bottom water temperatures with benthic foraminiferal δ^{18}O we can reconstruct seawater δ^{18}O and, by extension, salinity.

2. Materials and Methods

[6] Giant gravity core KNR158-4-09GGC and its accompanying multicore 10MC were raised from a depth of 1854 m on the Laurentian Slope south of Newfoundland (44°50′N, 54°54′W) (Figure 1). This location is well situated for monitoring upper NADW properties because it is directly in the path of LSW as it flows southward as part of the Deep Western Boundary Current. Twelve AMS radiocarbon dates from the planktonic foraminifer *Globigerina bulloides* (>150 μm) were converted to calibrated ages [*Stuiver and Reimer*, 1993] using

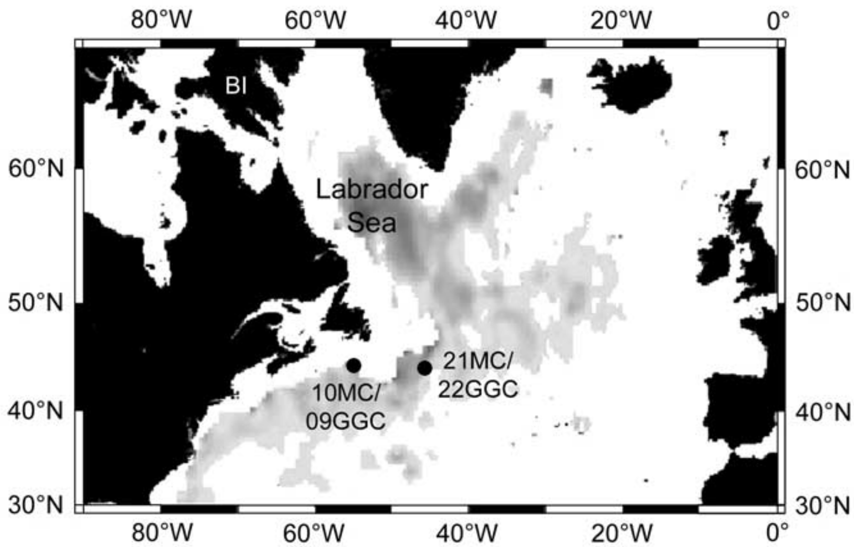

Figure 1. Map of the North Atlantic showing the potential vorticity minimum ($1-11 \times 10^{-12}$ m^{-1} s^{-1}), a useful tracer of LSW, at 1800 m depth (gray). This illustrates the spreading of LSW from the Labrador Sea northeastward into the Irminger Basin, eastward across the Mid-Atlantic Ridge, and southwestward as part of the Deep Western Boundary Current. Core site KNR158-4-10MC/09GGC is located at 1854 m depth, directly in the path of the boundary current. KNR158-4-21MC/22GGC [*Bond et al.*, 2001] is also shown. "BI" indicates Baffin Island. Map made by R. Curry (Woods Hole Oceanographic Institution, personal communication, 2002) using data from the HydroBase 2 package (http://www.whoi.edu/science/PO/hydrobase).

CALIB-98 and a local ΔR of 92 ± 27 years [*Campana*, 1997] (Figure 2). The two cores' stratigraphies were spliced together using percent $CaCO_3$, and the ages fit with a 6th order polynomial. This age model indicates sedimentation rates of ~20 cm/kyr during the late Holocene, more than adequate to resolve millennial-scale climate variability. The multicore was sampled every 0.5 cm and the gravity core every 1 cm, giving temporal resolution of up to 25 years.

[7] Mg/Ca ratios were measured in the benthic foraminifer *Cibicidoides pachyderma*. Samples consisting of ~5–15 individuals each (>150 μm) were reductively and oxidatively cleaned following the methods of *Boyle and Keigwin* [1985] as modified by *Boyle and Rosenthal* [1996]. Measurement was by ICP-AES on a Jobin-Yvon Panorama 2000, with an analytical precision of ±0.018 mmol/mol (1σ). Average reproducibility of sample splits was ±0.045 mmol/mol (pooled standard deviation, dof = 5). Of 103 measurements, 3 were excluded because of apparent contamination (>0.4 mmol/mol higher than replicates or both adjacent samples). We converted Mg/Ca to in situ temperature using both a new linear calibration:

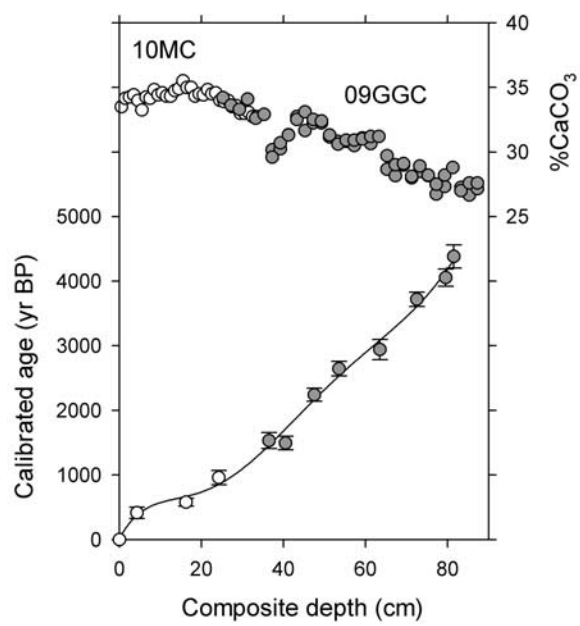

Figure 2. Top: Percent $CaCO_3$ measurements in KNR159-4 10MC (open) and 09GGC (gray) used to splice the two cores' stratigraphies together (GGC top corresponds to 8.25 cm in MC). Bottom: Calibrated radiocarbon ages in 10MC (open) and 09GGC (gray) with ±2σ error bars and 6th order polynomial age model.

$$\mathrm{Mg/Ca} = 0.25\mathrm{T} + 0.35 \tag{1}$$

and a published [*Martin et al.*, 2002] exponential calibration:

$$\mathrm{Mg/Ca} = 0.85e^{0.11\mathrm{T}} \tag{2}$$

(see section 3).

[8] For $\delta^{18}O$, ~2–5 individuals of *C. pachyderma* or *Uvigerina peregrina* (>150 μm) were analyzed on a Micromass Optima mass spectrometer. Analytical precision is ±0.06‰ (1σ). *U. peregrina* $\delta^{18}O$ values were adjusted by the average offset from paired *C. pachyderma* values (−0.52‰ ± 0.09, n = 11), which is within the range of previous observations [e.g., *Mix and Fairbanks*, 1985; *Keigwin*, 1998]. We then used the *Cibicidoides* and *Planulina* temperature equation:

$$\delta^{18}O_{\mathrm{foram(PDB)}} - \delta^{18}O_{\mathrm{seawater(SMOW)}} + 0.27 = -0.21\mathrm{T} + 3.38 \tag{3}$$

[*Lynch-Stieglitz et al.*, 1999] and Mg/Ca-derived temperatures to solve for seawater $\delta^{18}O$. Salinity was estimated using the equation

$$\delta^{18}O_{\mathrm{seawater(SMOW)}} = 0.386\mathrm{S} + \left(6.41 \times 10^{-3}\right)\mathrm{S}^2 - 21 \tag{4}$$

which is based on modern $\delta^{18}O$ and salinity measurements for LSW, Labrador Shelf Water, and local (Arctic) runoff [*Khatiwala et al.*, 1999; *Houghton and Fairbanks*, 2001]. The curvature of this equation approximates the influence of brine rejection. If convection occurred outside the Labrador Sea during some periods of the late Holocene, the resulting equation would be similar because Arctic runoff has a relatively uniform $\delta^{18}O$ value [*Khatiwala et al.*, 1999]. Potential increases in brine formation represent a larger source of uncertainty.

3. Mg/Ca Temperature Calibration

[9] While a number of studies have clearly defined the exponential relationships between planktonic foraminiferal Mg/Ca and temperature [e.g., *Nürnberg et al.*, 1996; *Mashiotta et al.*, 1999; *Elderfield and Ganssen*, 2000], less is known about benthic foraminifera, especially for waters colder than ~5°C [*Rosenthal et al.*, 1997; *Martin et al.*, 2002; *Lear et al.*, 2002]. Recent core top calibrations for *Cibicidoides* [*Lear et al.*, 2002; *Martin et al.*, 2002] use exponential fits that combine two or three species. At the cold temperatures typical of LSW, these calibrations are quite "flat," i.e., a given change in Mg/Ca is equivalent to large range in temperature (much larger than for warmer waters). As discussed below, this flatness results in an improbably large temperature range (up to 7°C) when applied to our downcore data. We therefore suggest that the existing calibrations are too flat at the cold end, and reexamine the calibration of *C. pachyderma*.

[10] The recently published *Cibicidoides* calibrations [*Lear et al.*, 2002; *Martin et al.*, 2002] rely heavily on earlier measurements [*Russell et al.*, 1994; *Rosenthal et al.*, 1997; Y. Rosenthal, previously unpublished data] that have been corrected for a significant analytical bias. Since the nature of this bias is not well understood, and since the *Rosenthal et al.* [1997] samples were not rigorously cleaned, we focus here on the new data generated by *Lear et al.* [2002] and *Martin et al.* [2002]. We also consider *C. pachyderma* separately from *C. wuellerstorfi*. The *C. pachyderma* data, from Little Bahama Bank and the Hawaiian Islands [*Lear et al.*, 2002], are shown in Figure 3. *Lear et al.* [2002] argued that at least four of these samples suffer from contamination by high-Mg calcite overgrowths or were reworked from shallower water depths (open symbols in Figure 3), and we add a fifth point to this suspect group because it is a >5σ outlier (gray circle). A simple straight line (equation (1)) is an adequate fit to the remaining *C. pachyderma* data (r^2 = 0.88). Note that much of the scatter about the line may be attributed to estimated bottom water temperatures being inappropriate for the non-modern ages of the core top foraminifera.

[11] Although we are not confident that the "true" *C. pachyderma* relationship is linear, we suggest that (i) at the cold end it is closer to linear than to the published exponential equations, and (ii) an exponential equation is not justified by the current non-suspect *C. pachyderma* data. The equation of

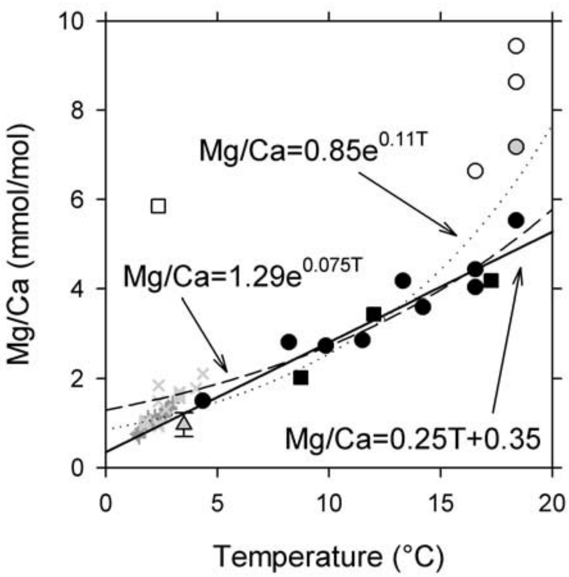

Figure 3. Core top *C. pachyderma* Mg/Ca measurements from Little Bahama Bank (circles) and the Hawaiian Islands (squares) vs. estimated modern in situ bottom water temperature [*Lear et al.*, 2002]. Data suspected by *Lear et al.* [2002] of being contaminated or reworked from shallower water depths are shown as open symbols; we add one point to this group (gray circle). A straight line (equation (1); solid) is a good fit to the remaining *C. pachyderma* data (filled). Also shown are an exponential fit to the filled core top data ($r^2 = 0.88$; dashed line) and the exponential *Cibicidoides* equation (2) from *Martin et al.* [2002] (dotted line). Gray circle is a >5σ outlier relative to the solid and dashed fits (filled data are all <2σ). Linear fits to *C. wuellerstorfi* data from *Martin et al.* [2002] (r^2 = 0.89; dark gray crosses) and *Lear et al.* [2002] (r^2 = 0.72; light gray exes) have slopes that are distinct from *C. pachyderma* (t-test $P < 0.001$ and 0.06, respectively), arguing that these species should be treated separately. Although carbonate undersaturation has been suggested as an explanation for the steeper trend of the *C. wuellerstorfi* data [*Martin et al.*, 2002], data from mostly supersaturated waters [*Lear et al.*, 2002] exhibit an indistinguishable slope (both 0.39). The mean late Holocene (past 4000 years) *C. pachyderma* Mg/Ca value and ±2σ spread in KNR158-4-10MC/09GGC, plotted vs. estimated modern bottom water temperature, is shown as a gray triangle.

Martin et al. [2002] (equation (2); dotted line in Figure 3) is a poor fit to the "good" core top data. An exponential fit to these data (dashed line) is not an improvement ($r^2 = 0.88$) over the linear fit and is incompatible with our downcore data (gray triangle). The incorporation of Mg into foraminiferal calcite is likely related to temperature through both inorganic thermodynamics (expected to be exponential) and physiological processes [*Rosenthal et al.*, 1997]. The latter must be of great importance because foraminiferal Mg/Ca ratios are about an order of magnitude lower than expected from inorganic precipitation [*Mucci*, 1987]. Hence the resulting temperature relationship need not be strongly exponential. Clear differences have been documented between various benthic species, with some exhibiting little or no curvature [*Toyofuku et al.*, 2000; *Lear et al.*, 2002].

[12] In the figures below, we present our *C. pachyderma* results using both our linear equation (1) and *Martin et al.*'s [2002] exponential equation (2). The linear equation gives conservative (minimum) estimates of past temperature and salinity changes, while the exponential gives extreme (and, we suggest, unrealistic) estimates. True paleotemperatures and paleosalinities may fall somewhere in between, but probably much closer to the linear estimates. Therefore we limit our main discussion to the linear results (sections 4.1, section 4.3), though we briefly comment on the exponential results (section 4.2). Note that linear Mg/Ca temperature equations have been used previously for *Cibicidoides* [*Rathburn and De Deckker*, 1997; *Billups and Schrag*, 2002] and for shallow-water benthic foraminifera [*Toyofuku et al.*, 2000]. Clearly, more calibration data are required to better resolve this important issue.

4. Results and Discussion

4.1. Inferred Intermediate Water Temperature and Salinity History

[13] KNR158-4-10MC/09GGC *C. pachyderma* Mg/Ca values range from 0.6 to 1.4 mmol/mol and display several large oscillations during the past 4000 years (*C. pachyderma* is rare or absent prior to this time) (Figures 3a and 4). Inferred temperatures calculated from our linear equation (1) range from 1.2°C to 4.2°C, with nearly all of the data falling between 1.5°C and 3.5°C. Modern bottom water temperatures at the core site are ~3.5°C, suggesting that late Holocene temperatures were frequently lower but rarely higher than today. Although it has been suggested that benthic

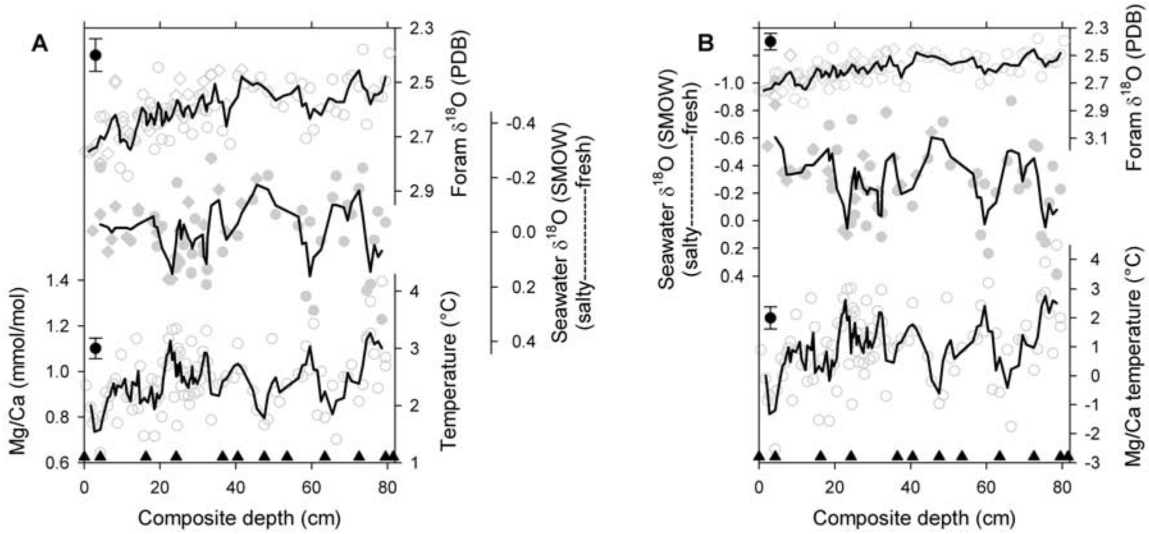

Figure 4. (A) Benthic foraminiferal results from KNR158-4-10MC/09GGC. Bottom: *C. pachyderma* Mg/Ca ratios and inferred in situ temperatures calculated using linear equation (1). Modern temperature at this site is ~3.5°C. Filled circle shows the average reproducibility of sample splits (±0.045 mmol/mol). Top: δ^{18}O of *C. pachyderma* (circles) and *U. peregrina* adjusted by −0.52‰ (diamonds). Modern predicted value is 2.7‰. Filled circle shows analytical precision (±0.06‰). Middle: Seawater δ^{18}O inferred from the foraminiferal δ^{18}O and Mg/Ca. Modern predicted value is 0.3‰. In each record the line is a three-depth moving average. Mg/Ca and foraminiferal δ^{18}O are plotted on the same scale with respect to temperature. Triangles indicate the depths of radiocarbon measurements. (B) Same as in (A), except Mg/Ca temperatures (and resulting seawater δ^{18}O) calculated using exponential equation (2).

foraminiferal Mg/Ca might be further reduced by low bottom water calcite saturation states [*Martin et al.*, 2002], our site is sufficiently supersaturated ($\Delta CO_3^{2-} \approx 45$ μmol kg^{-1}) that significant effects due to changing circulation or biological productivity are unlikely.

[14] The most recent Mg/Ca oscillation shifts from relatively warm intermediate waters (~3°C) during the Medieval Warm Period (~900–750 cal BP) to relatively cold waters (~1.5–2.5°C) during the LIA (since 750 cal BP) (Figure 5a). The cooling since 750 cal BP is significantly distinct from the prior millennium (1800–750 cal BP) ($P \ll 0.0001$). Bottom water temperatures during the Medieval Warm Period were evidently similar to modern instrumental values. Although the top of 10MC is modern (contains excess radiocarbon from the nuclear testing of the mid-late 20th century), our shallowest Mg/Ca measurement is from slightly deeper and likely contains an admixture of foraminifera from the late LIA stratigraphic level; hence we do not record modern temperatures. Two additional prominent cold events, of similar magnitude to the LIA, are centered at ~2100 and 3200 cal BP. Given slight uncertainties in the respective age models, the earlier cooling may correspond to *Bond et al.*'s [2001] drift ice event 2 (~3400–2700 cal BP). Our 2100 cal BP cooling falls between events 1 and 2 and may be related to a more subtle IRD increase at that time (event "1a" in Figure 5a). We also record a slight double cooling that is synchronous with *Bond et al.*'s [2001] event 1 doublet. Overall the correlation between the Mg/Ca three-depth moving average and the *Bond et al.* [2001] record is weak but significant (r = 0.22, $P = 0.04$, n = 93).

[15] Perhaps more striking is the correspondence between the Mg/Ca record and the history of Neoglacial advances on Baffin Island at the head of the Labrador Sea [*Davis*, 1985] (r = 0.43, $P < 0.0001$, n = 93; Figure 5b). Although the precision of the Baffin Island timescale suffers from the limits of lichenometry, each of the four major advances mapped on southern Cumberland Peninsula (Cumberland, Pangnirtung, Kingnait, and Snow Creek) appears to be coincident with inter-

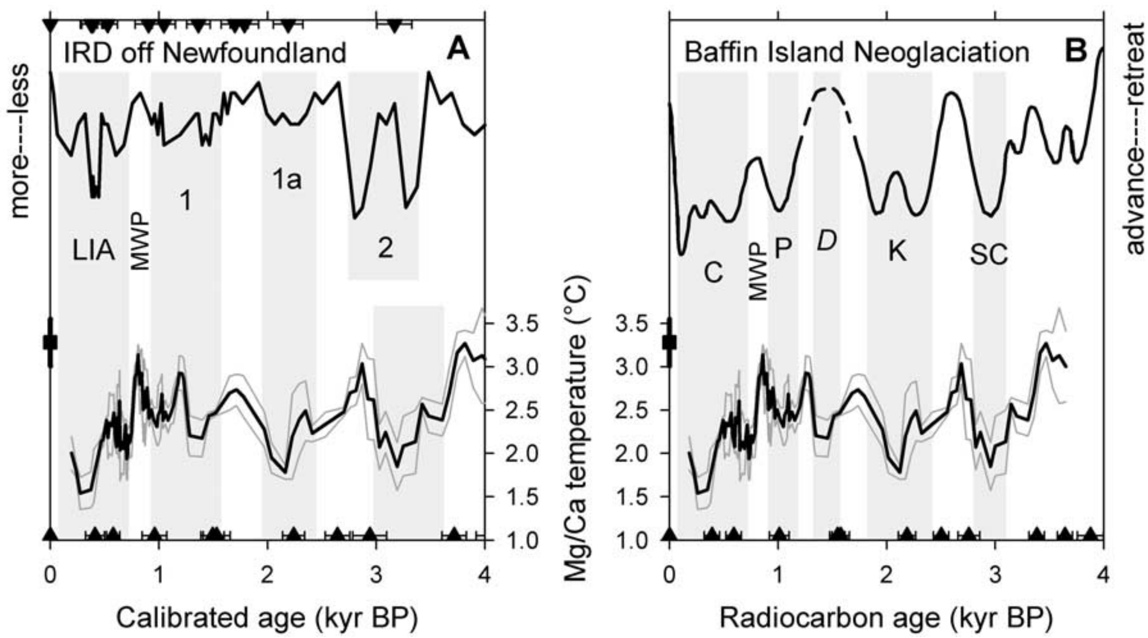

Figure 5. Comparisons of KNR158-4-10MC/09GGC *C. pachyderma* Mg/Ca in situ temperatures (linear equation (1); three-depth moving average with $\pm 1\sigma$ standard error envelope) to regional climate records. Square indicates mean in situ temperature and $\pm 1\sigma$ range for the core of LSW over the past 65 years [*Yashayaev et al.*, 2003]. (A) Percent detrital $CaCO_3$ (inverted) in KNR158-4-21MC/22GGC off Newfoundland (44°18′N, 46°16′W) vs. calibrated age [*Bond et al.*, 2001]. Numbers refer to Holocene IRD events, with probable correlatives in KNR158-4-10MC/09GGC indicated by gray bars. "LIA" and "MWP" refer to the Little Ice Age and Medieval Warm Period as manifested in these two cores. (B) Neoglacial advances on Baffin Island's southern Cumberland Peninsula reconstructed from moraines dated by lichenometry, on a radiocarbon timescale [*Davis*, 1985]. Letters indicate Baffin Island's five major advances: *C*umberland (LIA), *P*angnirtung, *D*orset (seen elsewhere on Baffin Island but not shown in this reconstruction), *K*ingnait, and *S*now *C*reek. In both panels, triangles indicate radiocarbon dates with $\pm 2\sigma$ ranges.

mediate water cooling. A fifth major advance recognized elsewhere on Baffin Island (Dorset) dates to ~1500 ^{14}C yr BP and may also correspond to cooling at our core site. Together with the comparison to *Bond et al.*'s [2001] drift ice record, this suggests that upper NADW cooling is linked to regional surface cooling.

[16] Benthic foraminiferal (*C. pachyderma* and *U. peregrina*) $\delta^{18}O$ is shown at the top of Figure 4a. Although the magnitude of variability in the $\delta^{18}O$ and Mg/Ca records is comparable in terms of temperature, there is little direct correspondence between the two tracers except that both imply a general cooling toward the top of the core. This lack of close similarity suggests that the foraminiferal $\delta^{18}O$ record is significantly affected by variations in seawater $\delta^{18}O$. Inferred seawater $\delta^{18}O$ values (derived by combining foraminiferal $\delta^{18}O$ with Mg/Ca temperatures in equation (3)) are generally highest during warm periods and lower during cold periods (Figure 4a). Since global ice volume has changed little during the late Holocene, the main factor influencing seawater $\delta^{18}O$ is expected to be salinity, with high latitude fresh water being low in $\delta^{18}O$. To get a sense of the magnitude of this influence, we convert our seawater $\delta^{18}O$ into inferred upper NADW salinities using our simple mixing model (equation (4)). Resulting values range from 34.3 to 35.0 PSU, with colder periods being generally fresher (Figure 6a). If sea ice formation was more important during cold periods, this range would be somewhat reduced due to brine rejection (cold waters would be saltier for a given seawater $\delta^{18}O$ value) [*Dokken and Jansen*, 1999].

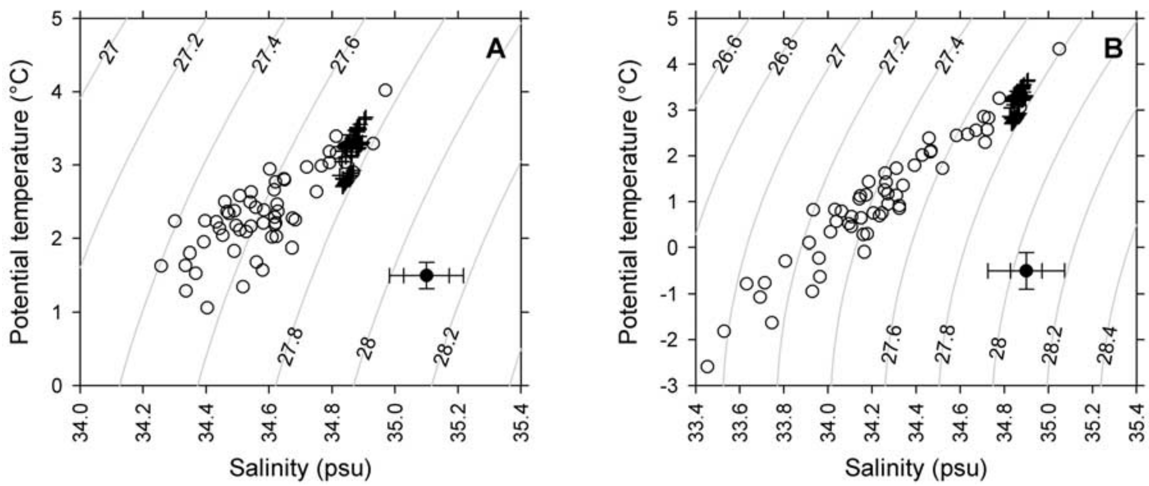

Figure 6. (A) Inferred upper NADW potential temperature (θ) and salinity over the past 4000 years, with isopycnals of σ_θ. KNR158-4-10MC/09GGC downcore in situ temperature is derived from *C. pachyderma* Mg/Ca using linear equation (1), and plotted (as θ) vs. salinity inferred from paired $\delta^{18}O$ measurements (circles). Filled circle shows impact of measurement errors: vertical error bars are based on reproducibility of Mg/Ca samples (±0.045 mmol/mol), and horizontal error bars are based on precision of $\delta^{18}O$ measurements (±0.06‰), assuming no correlation to Mg/Ca error (inner bars) or propagated with Mg/Ca error (outer bars). These ranges indicate the degree to which any individual point may move with respect to the rest. Additional uncertainty associated with the temperature and salinity equations would tend to impact the field as a whole. In particular, increased brine rejection during cold periods would make the field more isopycnal. Also shown are hydrographic measurements in the core of LSW over the period 1938–2001 [*Yashayaev et al.*, 2003] (crosses). (B) Same as in (A), except Mg/Ca temperatures (and resulting salinities) calculated using exponential equation (2).

[17] The positive covariance between temperature and salinity indicated by our foraminifera is also characteristic of the instrumental record. Since 1938 LSW has varied by 0.9°C (2.7–3.6°C θ) and 0.08 PSU (34.83–34.91 PSU) mainly due to variations in the strength of deep convection, but with little change in density [*Yashayaev et al.*, 2003] (Figure 6a). Over the past 4000 years, our site has experienced about three times the temperature range and perhaps nine times the salinity range recorded by LSW hydrographic measurements. The inferred paleosalinities do fall within the range of historical Labrador Sea surface values, which dropped below 34 PSU during the Great Salinity Anomaly [*Lazier*, 1980]. If our paleosalinity estimates are reasonably accurate, then upper NADW densities have varied significantly over millennial timescales, in contrast to the instrumental record of the past 65 years (Figure 6a). Salinity appears to have dominated the paleodensity field, with cold, fresh periods having reduced densities. Alternatively, if brine rejection was substantially increased during cold periods, then corresponding salinities would be higher than in our reconstruction [*Dokken and Jansen*, 1999] and the late Holocene variations would be more isopycnal.

4.2. Comments on the Exponential Results

[18] Our observation of colder, fresher upper NADW during late Holocene cold events is robust whether we use a linear or exponential Mg/Ca temperature calibration. Using the *Martin et al.* [2002] exponential equation (2), however, results in a surprisingly large temperature range: −2.5°C (below the freezing point of seawater) to 4.5°C, though nearly all of the data fall between −1.0°C and 3.2°C. Could this large range possibly be accurate? One argument against its validity is that temperature fluctuations would have to have been opposed by very large, mirror-image changes in seawater $\delta^{18}O$ to result in nearly unchanged foraminiferal $\delta^{18}O$ (Figure 4b). Although temperature and salinity commonly covary in the ocean,

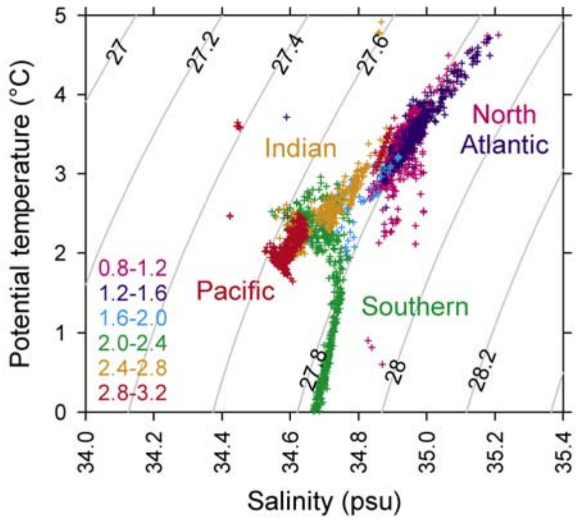

Figure 7. Modern potential temperature and salinity measurements for the world ocean in the depth interval 1800 to 1900 m (eWOCE bottle data, www.ewoce.org), with isopycnals of σ_θ (same scale as Figure 6a). Colors indicate concentrations of dissolved PO_4 (μmol/kg), here used to distinguish between ocean basins. Pacific waters (red) are roughly 0.15 σ_θ less dense than equivalent depth waters in the North Atlantic (pink and blue).

such an exact cancellation of the foraminiferal $\delta^{18}O$ signal seems too fortuitous (note in Figure 4a that much more of the seawater $\delta^{18}O$ variability appears to be "preserved" by the foraminifera when the linear temperature equation is used).

[19] Perhaps more importantly, inferred salinities and resulting densities reach implausibly low values (far lower than any equivalent depth waters in the modern ocean; see Figure 7) when the exponential temperature equation is used (Figure 6b). The only mechanism that might reconcile the exponential temperatures with realistic densities would be if brine formation was extremely important during the cold phases. If intermediate waters actually remained nearly isopycnal through time, then surface waters may have been subducted at much colder temperatures and only slightly lower salinities, with much lower seawater $\delta^{18}O$ negating the expected foraminiferal $\delta^{18}O$ increases. Again, however, we have the problem of the coincidental cancellation. We therefore view the exponential paleotemperatures as improbable.

4.3. Implications for Deep Convection

[20] The coincidence of cold, fresh conditions in North Atlantic surface waters [*Bond et al.*, 1997, 2001] and at our intermediate depth site suggests downward transmission of the signal through the water column. It is not yet clear, however, whether this occurred by deep convection (like today), by slow diapycnal mixing (like in the modern North Pacific), or by some combination of the two.

[21] Taken at face value, our reconstructed salinities suggest that intermediate (and thus shallower) depths had significantly lower densities during cold periods. In this way, North Atlantic hydrology may have been more analogous to that of the modern North Pacific. At the water depth of our core, the North Pacific is roughly 0.15 σ_θ less dense than the North Atlantic because of diapycnal mixing with very fresh, cold waters above (Figure 7). This difference is comparable to our late Holocene range: the mean density decrease between warm (>3°C θ) and cold (<2°C θ) periods is 0.19 ± 0.03 σ_θ (standard error). A significant freshening of the North Atlantic sea surface due to increased drift ice [*Bond et al.*, 1997, 2001] and reduced evaporation during cold periods could have therefore suppressed deep convection in the Labrador Sea and impacted intermediate depths through slow mixing.

[22] As surface salinities increased toward the ends of cold periods, deep convection may have resumed at lower densities than today because the threshold for convective instability would have been lower. It is not presently possible for us to discern between slow mixing and low-density convection because they would have similar impacts on temperature and salinity at our single core site. Multiple cores from a range of water depths are required to determine whether the water column was stratified (indicative of diapycnal mixing) or vertically homogenous (indicative of deep convection). We note that estimated late Holocene ranges of winter Labrador Sea surface temperature (~2.6–4.2°C) and salinity (~34.4–34.9 PSU) [*Hillaire-Marcel et al.*, 2001] are comparable to our intermediate water reconstructions, suggesting an apparent lack of strong stratification.

[23] Alternatively, if our reconstructed cold period salinities are negatively biased by unaccounted-for brine rejection [*Dokken and Jansen*, 1999] (i.e., if cold period salinities were higher than we have estimated), then corresponding densities would not be so low and Labrador Sea deep convection would have been easier to maintain. Using modern North Atlantic circulation as a guide to the past, cooling and freshening of intermediate waters may actually signal enhanced LSW formation during cold periods like the LIA. The crosses in Figure 6a illustrate the historical ranges of intermediate water temperature and salinity attributable to the presence (colder, fresher) or absence (warmer, saltier) of modern LSW [*Lazier*, 1995; *Dickson et al.*, 1996; *Yashayaev et al.*, 2003]. Since our reconstructed properties far exceed these ranges (even if forced to be isopycnal), they cannot be explained by a simple waxing and waning of modern-like LSW; rather, LSW would had to have formed at considerably lower temperatures and salinities during the cold periods.

[24] Of course, cold, fresh intermediate waters may have also formed by deep convection outside of the Labrador Sea during cold periods. If fresh water lowered sea surface densities throughout the North Atlantic, then overall convection depths may have been reduced, with the relatively shallow LSW being the first NADW component to disappear. Intermediate waters could have instead formed through brine rejection in a colder basin such as the Greenland Sea, at densities similar to those of modern LSW and at the expense of lower NADW. A reconstruction of Labrador Sea stratification based on dinocyst assemblages and foraminiferal δ^{18}O suggests that LSW formation was established after ~8,000 cal BP [*Hillaire-Marcel et al.*, 2001], but with millennial scale oscillations in sea surface density that may have impacted convection (resolution is too low for a detailed comparison to our record).

[25] Additional high resolution temperature and salinity reconstructions from other locations and water depths are clearly required to better identify sources and extents of the various NADW types through time. Planktonic Mg/Ca and δ^{18}O paired with independent sea surface salinity proxies [*de Vernal et al.*, 1997] may also help determine if brine formation was an important factor during late Holocene cold periods.

5. Conclusions

[26] Recent hydrographic studies have documented significant trends in upper and lower NADW properties over the past several decades [*Lazier*, 1995; *Sy et al.*, 1997; *Hansen et al.*, 2001; *Dickson et al.*, 2002]. Our results demonstrate that changes in upper NADW properties over the past 4000 years have greatly exceeded the range of variability evident in the instrumental record. Intermediate waters became colder and fresher during periods of increased drift ice and glacier expansions. Since future climatic responses to anthropogenic forcing are predicted to exceed the range of natural late Holocene variability [*Crowley*, 2000; *Intergovernmental Panel on Climate Change*, 2001], it is reasonable to hypothesize that even larger changes in NADW properties lie ahead. Such changes may fundamentally alter the distribution of heat throughout the Atlantic and thereby impact our future climate.

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