Supporting Online Material

Sampling methods and Mg/Ca analysis

Foraminiferal tests were picked from pre-washed sediment samples sieved to the 300-355 µm size fraction. Use of narrow size fractions helps to restrict mass variations due to external shell dimensions but cannot entirely account for variability in the modal size distribution between samples. In order to correct for size variability, shell weights were normalized to the mean size for that sieve-size fraction.

Mg/Ca measurements were made on the same shell population as used for weight and size analyses. Tests were gently crushed between glass slides before being ultrasonicated in water and methanol to remove clay particulates. Organic matter was removed by boiling in an alkali buffered H₂O₂ solution. Analysis was performed on a Varian Vista ICP-AES using the intensity ratio calibration method (SI). Mg/Ca ratio measurements are quoted with an accuracy and precision of <0.3%.

Methods for temperature and salinity estimation

Temperature and salinity records from NEAP 8K

When combining multiple proxies to investigate past oceanic conditions, complications arise due to differences in the source material used. For example, combining faunal or alkenone-derived temperatures with foraminiferal δ¹⁸O data to reconstruct δ¹⁸O seawater involves assumptions regarding the temporal and spatial distribution of several faunal species. Mg/Ca thermometry has the potential advantage that it may be used on the same samples as weight and δ¹⁸O analyses thus reducing much of the associated uncertainty.

The downcore record of Mg/Ca in G. bulloides is shown in Fig. 4A. Mg/Ca ratios start to increase after their minimum value at about 30-35ka, considerably earlier than the δ¹⁸O data from the same core. Maximum values of Mg/Ca occur at about 12ka in the early Holocene. Fig. 4B shows Mg/Ca values from NEAP 8K as compared with faunal and alkenone temperature estimates from a nearby core (BOFS 16K, 59°28’N, 23°24’W, water depth 2370m) (S2). The timing of temperature shifts associated with foraminiferal Mg/Ca reconstructions show greater similarity with alkenone-based estimates than with those based on faunal transfer functions of foraminiferal species assemblages. This apparent dichotomy has been observed previously in a subtropical core (S3) and is probably due to the variability of seasonal and depth habitats associated with the various proxy mediums. It does however highlight the importance of using the same medium (e.g. a single species of foraminifer) when combining multiple proxies.

Mg/Ca temperature calibration

Several published temperature calibrations based on core-top and culturing experiments exist for Mg/Ca ratios in Globigerina bulloides (S3–S5). The general form of such calibration equations is that of an exponential relationship between Mg/Ca and temperature:
Mg/Ca = A exp (B x T)

where T is the calcification temperature and A and B are first order coefficients. The numerical value of B is commonly quoted as 0.1. This is true for several other species of planktonic foraminifera as well as *G. bulloides* (S3). The pre-exponential coefficient, A, for *G. bulloides* is quoted as being between 0.47 and 0.56 (S3, S5).

A potential problem with use of established calibrations is that they are not based on temperatures lower than about 10°C. The average modern temperature for June at 30m water depth at the site of NEAP 8K is approximately 9.7 °C (S6). The range in predicted values of Mg/Ca at this temperature is 1.25 to 1.48 mmol.mol⁻¹ (B = 0.474 - 0.56). The core-top Mg/Ca ratio measured in NEAP 8K is approximately 1.9 mmol.mol⁻¹. This would suggest a predicted temperature of 12.2° - 13.9 °C. A potential cause of differences between predicted and observed Mg/Ca ratios (a particular problem at low temperatures where small changes in Mg/Ca imply large shifts in temperature) is one of current debate in the field of Mg/Ca thermometry and is beyond the scope of the present study. Suffice to say that each application of Mg/Ca thermometry should be considered individually.

In calibrating the Mg/Ca data for NEAP 8K, some straightforward assumptions were made. The general form of the equation used was that above with the value of B taken as 0.1, in line with all other calibrations for *G. bulloides*. The value of A was then adjusted to give a best-fit temperature for the core-top Mg/Ca value (A = 0.72; gives T = 9.74 °C for a Mg/Ca ratio of 1.9 mmol.mol⁻¹). Calculated calcification temperatures for *G. bulloides* in NEAP 8K are shown in Fig. 4C. The 5-6 °C range in temperatures over the last 50 ka is in line with that predicted from faunal estimates but considerably less than for the alkenone reconstruction (range >10 °C) from BOFS 16K (Fig. 5b) (S2). It is conceivable that *G. bulloides* may adapt to changing climatic conditions by altering its season or depth of habitat. Adaptability of this kind would lead to temperature signals that were effectively dampened responses to absolute, depth and/or season specific, temperature variations.

### δw and salinity records

Derivation of paleosalinity records is a nontrivial matter (S7, S8). Errors arise from the differences between temporal and spatial seawater δ¹⁸O:salinity (δw:S) relationships and through the use of nonconcurrent temperature measurements (S8). By using temperature and δ¹⁸O records from a single foraminiferal species, the latter source of uncertainty can be reduced. The modern δw:S relationship observed in surface waters of the North Atlantic is essentially linear (S9). However, difficulties in constraining the isotopic composition of freshwater sources and the effects of sea-ice forming processes (pertinent to studies at 60 °N in the glacial North Atlantic) result in considerable uncertainty in the temporal δw:S relationship at a particular location.

For calculation of paleosalinities in NEAP 8K, the paleo-mixing line equation of (S8) is utilised:

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δw = δw_F + ((δ_1 - δw_F) / (S_m + S_i)) S
\]
where $\delta_F$ is the net isotopic content of high-latitude precipitation at the time of interest (set at -17‰ for the current study), $S_m$ is modern global salinity, $\delta_i$ and $S_i$ are the effects of ice volume on $\delta w$ and $S$ respectively. Both $\delta_i$ and $S_i$ are calculated as a function of sea level fall ($\Delta$): $S_i = 34.7\Delta / (3800-\Delta)$, $\delta_i = 0.011\Delta$ ($S10$). Shackleton’s ($S11$) sea level curve as derived from the Vostok $\delta^{18}O_{atmospheric}$ record (with a correction for the dole effect) was used in the calculation of $\delta_i$ and $S_i$. Values of $\delta w$ were calculated from $\delta^{18}O_{carbonate}$ ($\delta c$) and $T_{calcification}$ using Shackleton’s ($S12$) paleotemperature equation (Fig. 5c):

$$T = 16.9 - 4.38(\delta c - \delta w) + 0.1(\delta c - \delta w)^2$$

These were converted from the PDB scale to VSMOW by addition of 0.27‰. The total error in $\delta w$ is probably in the order of $\pm 1.1-1.5$ ($S5$).

**Sensitivity of calculated $pCO_2(aq)$ values**

**Temperature and Salinity**

The strong temperature dependence of $K_{0,1,2}$ is a significant source of uncertainty in our calculation of $pCO_2(aq)$. Calculated $pCO_2(aq)$ increases with increased temperature. Paleo-temperature estimates from Mg/Ca thermometry are typically quoted with an error of $\pm 1^\circ C$. This results in a corresponding uncertainty in $pCO_2(aq)$ of $\pm 10$µatm.

As compared with temperature, $K_{0,1,2}$ are not as sensitive to salinity variations. Hence even quite large changes in this parameter do not lead to considerable shifts in calculated $pCO_2(aq)$. The most important contribution of salinity variability to the calculations outlined in this paper is the corresponding change in TA (see below). Combined uncertainties in $T$ and $S$ result in an overall uncertainty in of $\pm 10.5$µatm (Fig. S1A).

**Choice of Alkalinity**

The value of TA used to calculate $pCO_2(aq)$ (Fig. 3B) was set at 2320 µeq.kg$^{-1}$ although it is likely that TA has varied over the last 50ka. Total alkalinity in the modern surface ocean is linearly related to salinity although it is not clear whether this relation is constant through time. The record of calculated salinities for NEAP 8K suggests that these may have been up to 7% higher during glacial times. This is significantly higher than that expected from the lower LGM sea level and would correspond to a TA increase of 160 µmol.kg$^{-1}$ if the modern linear relationship were preserved. Fig. S1B shows the calculated record of $pCO_2(aq)$ if TA is treated as a function of salinity, as compared with keeping TA fixed at 2320 µeq.kg$^{-1}$. It is perhaps unlikely that TA was as much as 7% higher during the LGM and thus the true curve probably lies somewhere between the two.

Glacial [CO$_3^{2-}$] in North Atlantic surface waters was about 30% higher than the modern value (Fig. 3A). Therefore, even a 7% increase in TA would result in an increase of glacial [CO$_3^{2-}$]/[TA]. By considering the Bjerrum plot for carbonate, borate and water alkalinitities, it follows that as [CO$_3^{2-}$]/[TA] increases so the total contribution of carbonate
alkalinity to TA ([HCO₃⁻]+2[CO₃²⁻]/[TA]) will decrease. Hence the value of 0.96 used for modern CA/TA should be reduced when calculating $p_{CO_2(aq)}$ during the LGM. This would effectively lower calculated values of $p_{CO_2(aq)}$ during the LGM and shift the record toward that for fixed TA (Fig S1B).

**Fig. S1.** (A) Calculated $p_{CO_2(aq)}$ showing errors associated with uncertainties in temperature and salinity. (B) Calculated $p_{CO_2(aq)}$ attained by allowing TA to vary as a function of salinity compared to that attained by fixing TA at 2320 µeq.kg⁻¹.

**Supporting references**