

Temporal changes in North Atlantic circulation constrained by planktonic foraminiferal shell weights

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[1] Records of planktonic foraminiferal shell weights for *Globigerina bulloides*, covering Termination I from four proximal sites at waters depths from 1150 to 4045 m in the northeast Atlantic, demonstrate the influence of dissolution superimposed upon initial shell weight variability. Records of shell weight, unaffected by dissolution, may be used as a reference for interpreting deeper records in terms of preservation history. Combining records of planktonic shell weight (and benthic $\delta^{13}\text{C}$) from shallow and deep sites suggests that maximum oceanic stratification and incursion of southern sourced deep waters in the North Atlantic occurred at about 18–20 ka, defining the glacial mode of Atlantic circulation. Reduced stratification and enhanced preservation in deeper waters reflect conditions during Heinrich event 1. A state similar to the modern mode of Atlantic circulation was attained by about 10 ka. **INDEX TERMS:** 1050 Geochemistry: Marine geochemistry (4835, 4850); 4267 Oceanography: General: Paleooceanography; 4825 Oceanography: Biological and Chemical: Geochemistry; 4283 Oceanography: General: Water masses; 4532 Oceanography: Physical: General circulation; **KEYWORDS:** foraminiferal shell weights, dissolution, North Atlantic circulation

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1. Introduction

[2] The use of shell weights of planktonic foraminifera as an indicator of dissolution in the deep ocean was proposed by Lohmann [1995] and has been developed by Broecker and Clark [2001a, 2001b] as a proxy for bottom water carbonate ion concentration, $[\text{CO}_3^-]$. Weight loss, caused by dissolution of the carbonate tests of foraminifera, can be calibrated with bottom water $[\text{CO}_3^-]$ and used to reconstruct paleo- $[\text{CO}_3^-]$ gradients using cores from different water depths [Broecker and Clark, 2001a, 2001b].

[3] The shell weight method is based on a number of assumptions. One is the implicit assumption that foraminiferal dissolution is related to bottom water $[\text{CO}_3^-]$, ignoring supralysocline dissolution [Archer *et al.*, 1989] unless a reasonably constant offset between bottom water and pore water $[\text{CO}_3^-]$ is assumed. Of more fundamental importance is the assumption that the “initial weights” of foraminiferal shells (i.e., before dissolution) are reasonably constant through time such that weight loss may be determined by comparison with core-top samples. However, it has been observed that there is significant variability in size-normalized weights of foraminiferal shells from core tops recovered from shallow water depths [Barker and Elderfield, 2002]. This is thought to be due to the influence of carbonate saturation during shell growth, i.e., surface

water $[\text{CO}_3^-]$ [Spero *et al.*, 1997; Bijma *et al.*, 1999]. There may be additional causes of variability in “initial shell weight,” such as growth habitat and food supply [Broecker and Clark, 2001a], but the influence of surface water $[\text{CO}_3^-]$ is strongly supported by observations that glacial age shells from the North Atlantic were significantly heavier than those of Holocene age and that this increase could be attributed to the glacial increase in surface ocean $[\text{CO}_3^-]$ in line with lowered glacial atmospheric CO_2 [Barker and Elderfield, 2002].

[4] Given the variability in the shell weights of undissolved foraminifera, it follows that, before records of shell weight may be interpreted in terms of dissolution and bottom water $[\text{CO}_3^-]$ variability, they must be adjusted to allow for changes in initial weight. The most straightforward means of achieving this is to use the record of a shallow site, located well above the lysocline, as a reference site. Here, we apply this method using records of shell weight from four sites on a depth transect from 1150 to 4045 m in the North Atlantic. Our approach is to use the record of shell weight at the shallowest site to constrain initial weight variability and apply this to interpret deeper records in terms of weight loss and dissolution. Using records rather than time slices means we are able to track the evolution of deep water $[\text{CO}_3^-]$ through time.

2. Background and Methods

2.1. Cores and Analytical Methods

[5] The cores used in this study were chosen to span a significant range in water depth within a limited geographic

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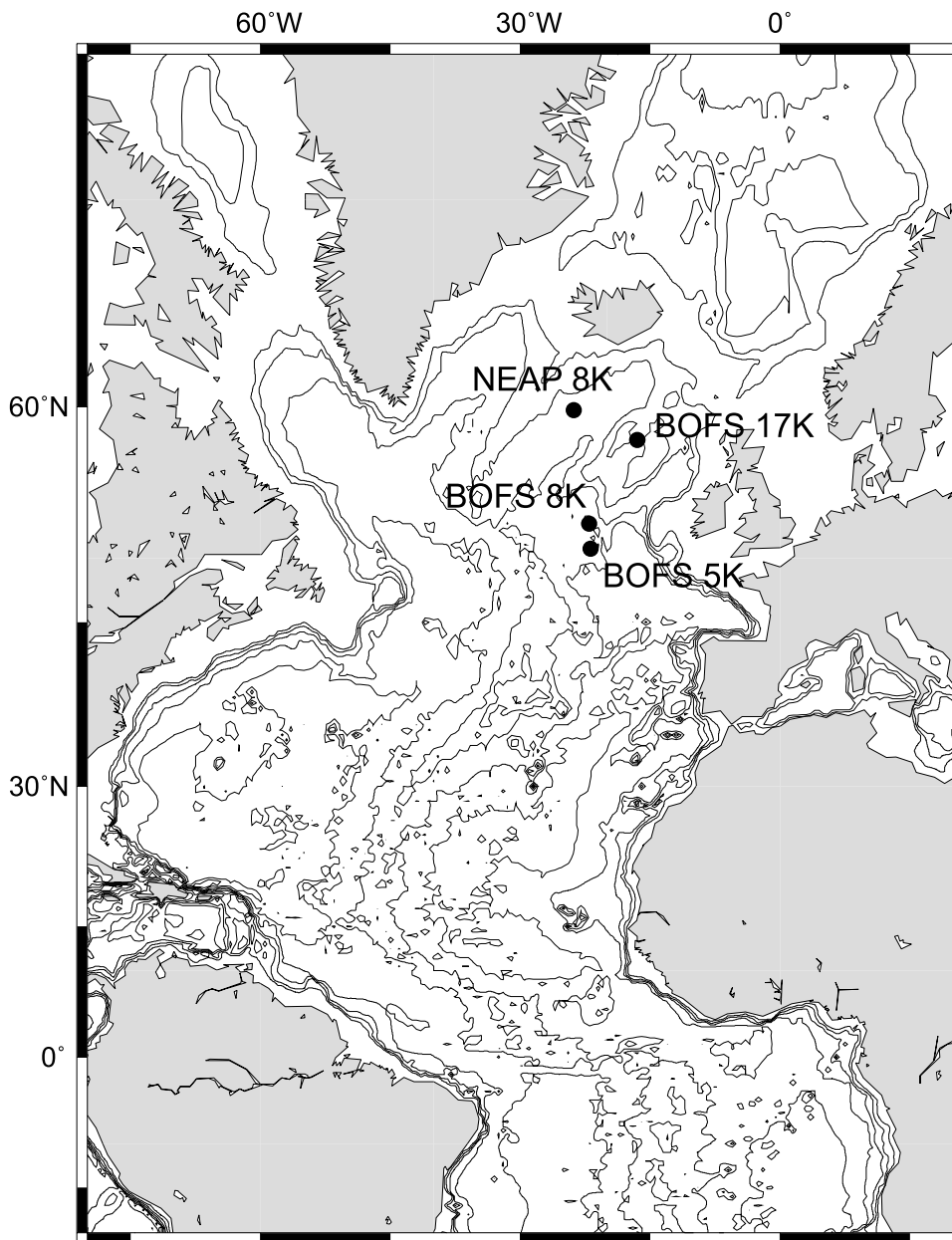


Figure 1. Locations of the four cores used in this study. Details are given in Table 1.

ical area in the northeast Atlantic Ocean (Figure 1, Table 1) under the assumption that the shell weight records from each core would have been similar before post depositional alteration by dissolution. The choice of cores was partially determined by the availability of preexisting proxy data to aid in cross correlation between cores and development of age models [Manighetti et al., 1995; Maslin et al., 1995].

[6] Tests of *G. bulloides* were picked from prewashed core sediment sieved to 300–355 μm . For weight analysis an ideal sample contained about 50 individual tests. However, the abundance of *G. bulloides* during glacial intervals in these cores is very low [Maslin et al., 1995]. Therefore fewer tests were used where numbers were limited although no sample contained less than 10 individual tests. Weight

measurements were made using a microbalance (0.1 μg). We estimate the error associated with natural variation to be about $\pm 0.5 \mu\text{g}$. Size analysis was performed on digital photomicrographs of individual sample populations. Each image contained the same set of reference shells allowing postanalysis correction for variations in light intensity and

Table 1. Core Locations for the Present Study

Core ID	Latitude, deg N	Longitude, deg W	Water Depth, m
BOFS 17K	58.0	16.5	1150
NEAP 8K	59.8	23.8	2360
BOFS 5K	50.7	21.9	3547
BOFS 8K	52.5	22.1	4045

focusing precision. The day to day precision of reference shell measurements was generally better than 0.5% for length determinations. This compares with downcore variations of >10%. Stable isotope measurements of O and C were made by M. Hall at the Godwin Laboratory, Cambridge. Scanning electron microscope (SEM) images of foraminiferal shells were made using a Jeol JSM 820.

2.2. Age Model Development

[7] The existing radiocarbon-based age model for BOFS 5K [Manighetti et al., 1995; Maslin et al., 1995] was revised in order to improve the chronological precision across the last deglacial. This chronology was then adopted for the other cores by cross-correlation. A general stratigraphic framework for BOFS 5K was attained from the percentage maximum of ice rafted debris (IRD) defining Heinrich layer (HL) 1 (62–78 cm) and HL 2 (106–108 cm), and from a maximum of Vedde ash particles (46–52 cm; semiquantitatively determined on the >150 μm size fraction) marking the Younger Dryas (Figure 2). To obtain a more precise chronology, we recalibrated the planktonic ^{14}C ages of Manighetti et al. [1995] with the CALIB program (version 4.3) [Stuiver and Reimer, 1993; Stuiver et al., 1998] and with the data set of Voelker et al. [2000] in the case of the ^{14}C age >20,000 ^{14}C years at 114 cm (Table 2). Instead of correcting the measured ^{14}C ages with a constant reservoir age of 400 years [Bard, 1988; Manighetti et al., 1995] we applied a local paleoreservoir age of 1900 ^{14}C years at the base of HL1 [Waelbroeck et al., 2001] and 800 ^{14}C years at the top of the YD [Haflidason et al., 2000; Waelbroeck et al., 2001].

[8] Dates were omitted from the age model at three levels: (1) The *G. inflata* date at 44 cm was ignored because it is about 5 ka too old for unknown reasons [Manighetti et al., 1995]. (2) The two *G. bulloides* dates at 73 and 74 cm were derived from a minimum in *G. bulloides* abundance below a fivefold abundance increase at 70 cm. As a result of downward mixing of Bølling-aged *G. bulloides* tests into HL1 [Manighetti et al., 1995] their ages are about 1 ka reservoir-corrected ^{14}C years younger than the corresponding *N. pachyderma* (s) also dated at 74 cm (Table 2). Although *N. pachyderma* (s) was dated from within an abundance decrease and might therefore be slightly biased toward being too old, we used this date in our age model because it provided a reasonable age estimate for HL1. (3) The date of 23.36 cal ka at 100 cm appeared several hundred years too old for its position 5 cm below HL2. Given the relatively poor calibration of ^{14}C dates below 13.0 ^{14}C ka [Stuiver et al., 1998] and to avoid improbable fluctuations in sedimentation rates we instead correlated the top of HL2 at 105 cm to the onset of the Dansgaard-Oeschger interstadial 2 at 23.53 cal ka [Stuiver and Grootes, 2000].

[9] No ^{14}C date was available between the base of HL1 and the top of the YD (74–46 cm). Here, an additional age constraint was provided by a maximum in right-coiling over total *N. pachyderma* at 60 cm, which was assumed to represent the Bølling warm spell and, accordingly, was assigned an age of 14.5 cal ka [Stuiver and Grootes, 2000].

[10] The revised age model results in the Vedde ash and a concomitant IRD maximum being in line with dates for the YD [Stuiver and Grootes, 2000] and the IRD maximum and

meltwater-induced planktonic $\delta^{18}\text{O}$ minimum of HL1 in line with dates for Heinrich event 1 [Bard et al., 2000]. Moreover, the *N. pachyderma* coiling ratios as an approximation of paleo-SST closely match the $\delta^{18}\text{O}$ based temperature record from Greenland [Stuiver and Grootes, 2000] (Table 2).

[11] As discussed above, the ages of *G. bulloides* at 73/74 cm are biased by bioturbation and are therefore too young with respect to other sediment components (e.g., IRD and *N. pachyderma* (s)). However, they may represent a good average age for *G. bulloides* at that core depth. Therefore an alternate *G. bulloides*-specific age model was constructed for comparison. This age model uses an average of the *G. bulloides* ages at 73 and 74 cm instead of the *N. pachyderma* (s) age at 74 cm and does not include the nearby *N. pachyderma* coiling ratio maximum as an age marker.

[12] Owing to their close proximity, the chronology for BOFS 8K could be derived from BOFS 5K by correlation of their IRD concentrations (Figure 2a). Use of this correlation is supported by the good agreement of coiling ratios in *N. pachyderma* (Figure 2b). BOFS 17K and NEAP 8K are located significantly to the north of the main IRD belt [Bond et al., 1992] and so no useful correlations could be made using percent IRD. Instead the benthic oxygen isotope records of BOFS 17K and BOFS 8K were used to cross-correlate between these two cores (Figure 2c). The oxygen isotope records for *G. bulloides* were used to correlate between NEAP 8K and BOFS 17K since these two cores are fairly proximal to one another (Figure 2d). Age models for BOFS 8K, 17K and NEAP 8K are given in Table 3.

2.3. Correction of Weight Data for Size Variability

[13] The shell weight method makes use of narrow size fractions to limit weight variability due to varying shell size (hence the notion of size normalized weight). Considerable size variability can occur even within a fairly narrow sieve size range [Elderfield and Barker, 2001] and can lead to significant weight variability within a single sample if strict control on the specific morphotype selected is not employed. This can be addressed by measuring the length and cross-sectional area of individual shells within a sample population and using these measurements to apply a size normalization to samples [Barker and Elderfield, 2002].

[14] Measured shell lengths from BOFS 17K show no particular trend through time (Figure 3). Those from BOFS 8K reveal that glacial age tests are significantly (up to 10%) smaller than those from the Holocene section of the core. As discussed later, the glacial section of BOFS 8K has suffered significant dissolution. Therefore it seems likely that the reduction in size is due to the loss of carbonate from the outer surface of the test during dissolution (see section 3.2). Since this phenomenon is a primary consequence of dissolution, it seems sensible not to correct for it. It is therefore suggested that the size normalization procedure employed by Barker and Elderfield [2002] should not be used if dissolution is suspected. We used an optimum sample size of 50 individuals for this study. This is the same as used by Broecker and Clark [2001a, 2001b] and should be sufficient to overcome the greater proportion of natural size variability without further normalization. The results for NEAP 8K are from Barker and Elderfield [2002] where only 24 individuals were used. Since we do not expect significant dissolu-

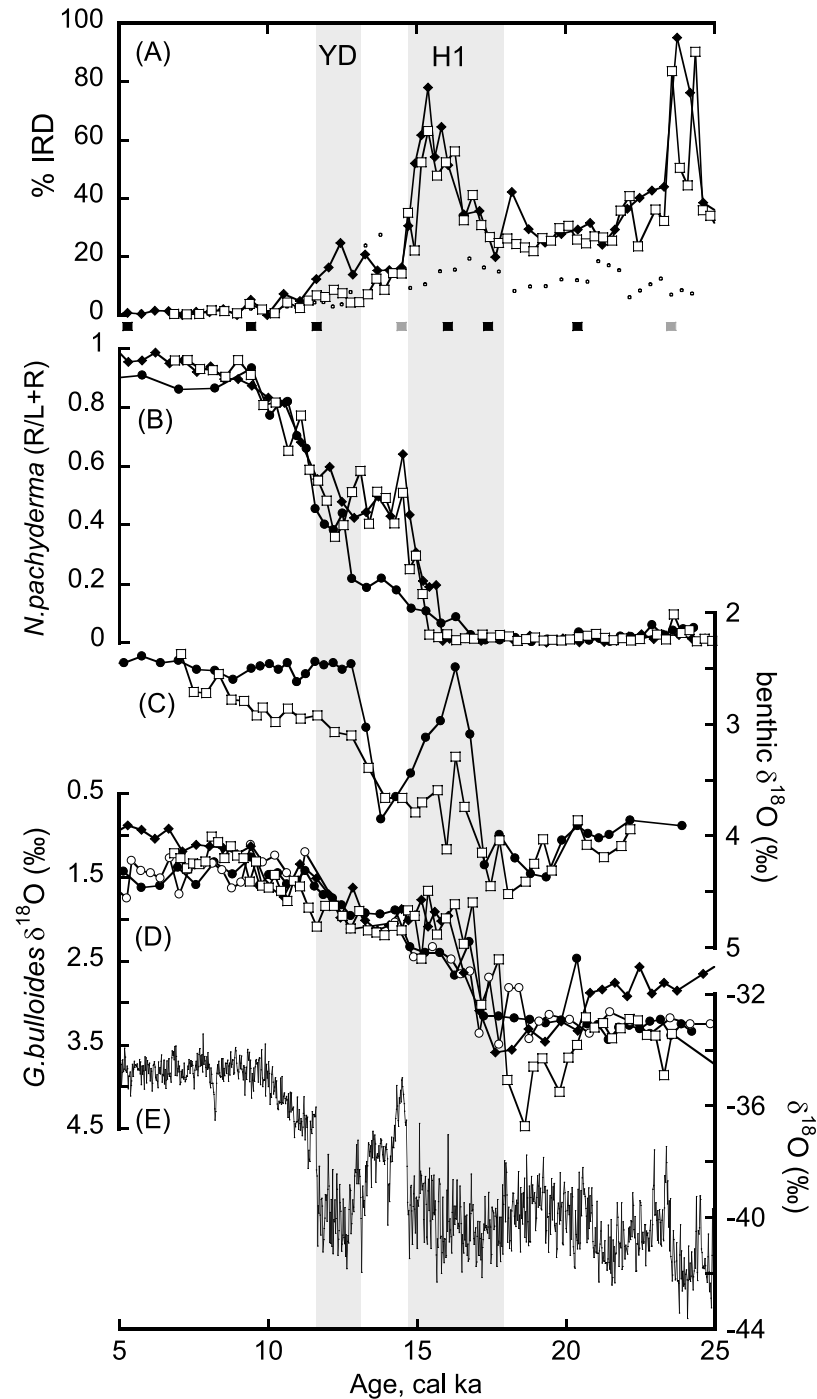


Figure 2. For all parts; solid circles represent BOFS 17K, hollow circles are NEAP 8K, solid diamonds are BOFS 5K, and hollow squares are BOFS 8K. (a) Percent IRD from BOFS 5K and 8K used for correlation. Also shown are data from BOFS 17K (small circles in this case) (data from *Maslin et al.* [1995]). ^{14}C age measurements for BOFS 5K are indicated by black squares. Gray squares indicate other dating constraints described in the text. (b) Coiling ratios for *N. pachyderma* for the BOFS cores (data from *Maslin* [1992]). BOFS 5K and 8K show a warming at the onset of the Bølling interstadial. (c) Benthic $\delta^{18}\text{O}$ data from BOFS 17K and BOFS 8K. (d) The $\delta^{18}\text{O}$ from *G. bulloides* for the four cores (data for the BOFS cores are from *Manighetti et al.* [1995]; data for NEAP 8K are from the NEAPACC data CD). See text for full description of age model development. (e) $\delta^{18}\text{O}$ from the GISP2 ice core [Stuiver and Grootes, 2000].

Table 2. Age Model for BOFS 5K

¹⁴ C Ages		Age Model							
Depth, cm	Foram Species	¹⁴ C Age	Reservoir Age	Corrected ¹⁴ C Age	Calibrated Ages	Calibration Used	Depth, cm	Cal Age, kyr	Comment ^a
0	<i>G. bulloides</i>	2020 ± 140	400	1620	1567	Calib 4.3	0	1.57	
20	<i>G. inflata</i>	4970 ± 130	400	4570	5300	Calib 4.3	20	5.3	
38	<i>G. bulloides</i>	8860 ± 130	400	8460	9436	Calib 4.3	38	9.44	
44	<i>G. inflata</i>	14970 ± 180	400	14570					A
46	<i>G. bulloides</i>	11180 ± 150	800	10380	12097, 12012, 11971	Calib 4.3	46	11.64	B
46	<i>N. pachyderma</i> (s)	10770 ± 130	800	9370	11322, 11243, 11197	Calib 4.3	46	11.64	B
							60	14.5	C
73	<i>G. bulloides</i>	12860 ± 150	400	12460	14322	Calib 4.3			D
74	<i>G. bulloides</i>	12990 ± 160	400	12590	15068, 14747, 14378	Calib 4.3			D
74	<i>N. pachyderma</i> (s)	15370 ± 180	1900	13470	16064	Calib 4.3	74	16.06	
78	<i>N. pachyderma</i> (s)	16520 ± 180	1900	15720	17388	Calib 4.3	79	17.38	E
80	<i>N. pachyderma</i> (s)	16510 ± 170	1900	15710	17376	Calib 4.3	79	17.38	E
90	<i>N. pachyderma</i> (s)	17650 ± 220	400	17250	20414	Calib 4.3	90	20.41	
100	<i>N. pachyderma</i> (s)	20210 ± 280	400	19810	23360	Calib 4.3			F
							105	23.53	G
114	<i>N. pachyderma</i> (s)	22960 ± 330	400	22560	25500	Voelker et al. [2000]	114	25.5	

^aComments are as follows: A, rejected as too old [Manighetti et al., 1995]; B, average of calibrated ages of Nps and Gb dates; C, *N. pachyderma* coiling ratio max. correlated to Bølling in GISP2; D, culled because of downcore mixing [Manighetti et al., 1995], used in bulloides-specific age model; E, average of 78 and 80 cm; F, culled because slightly too old with respect to top H2; G, top H2 correlated to DO2 warming in GISP2.

tion at this site we use the size normalized data of *Barker and Elderfield* [2002].

2.4. Correction for Latitudinal Offsets

[15] The basis for correcting for initial weight variability is that the weight records of the deeper cores would have been the same as the shallowest core had no weight loss by dissolution occurred. For this reason, our chosen core locations occupy a fairly limited geographical region. However, the bathymetry of the North Atlantic and the availability of suitable core material results in core sites spanning several degrees of latitude. Because modern shell weights vary with surface water [CO_3^{2-}] which itself varies with latitude (Figure 4), initial shell weights from the more southerly cores are expected to be slightly heavier than those from the more northerly sites. In order to incorporate this offset into our study, we have made a correction to three of the four cores (the most northerly site, that of NEAP 8K, being taken as the reference location) using a straightforward relation between shell weight and latitude:

$$\text{shell weight} = 42.2 \times \exp(-0.0228 \times \text{N.Lat}). \quad (1)$$

For the most southerly site (BOFS 5K) the correction decreases initial shell weights by 2.5 μg . We cannot predict

Table 3. Age Models for BOFS 8K, 17K, and NEAP 8K

BOFS 8K		BOFS 17K		NEAP 8K	
Depth, cm	Age, cal ka	Depth, cm	Age, cal ka	Depth, cm	Age, cal ka
20	8.98	0	0.89	0	0
40	11.09	28	9.44	118	9.44
64	14.5	50	12.80	136	11.27
72	15.39	64	16.29	138	12.80
80	16.59	70	17.76	144	14.29
126	23.32	80	20.39	162	17.27
136	24.62	90	22.15	196	22.85
174	29.35				

how the relation between shell weight and latitude might vary with time; for example, changes in surface temperature or nutrient gradients between sites might cause offsets to vary through time. However, it is difficult to assess how such changes might affect the results from this study and therefore we make the assumption here that the offsets

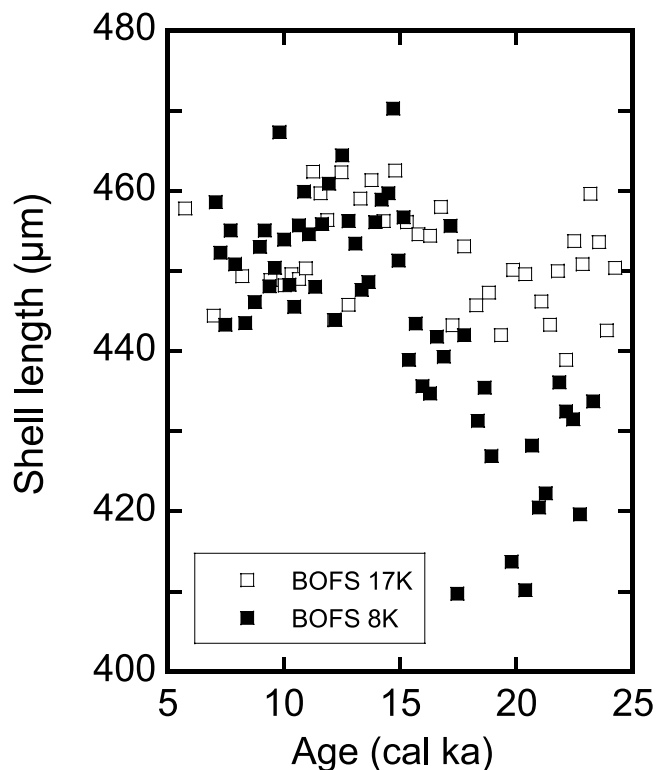


Figure 3. Measured shell lengths from BOFS 17K and 8K. Shells from the glacial section of BOFS 8K are significantly smaller than during the Holocene as a result of partial dissolution.

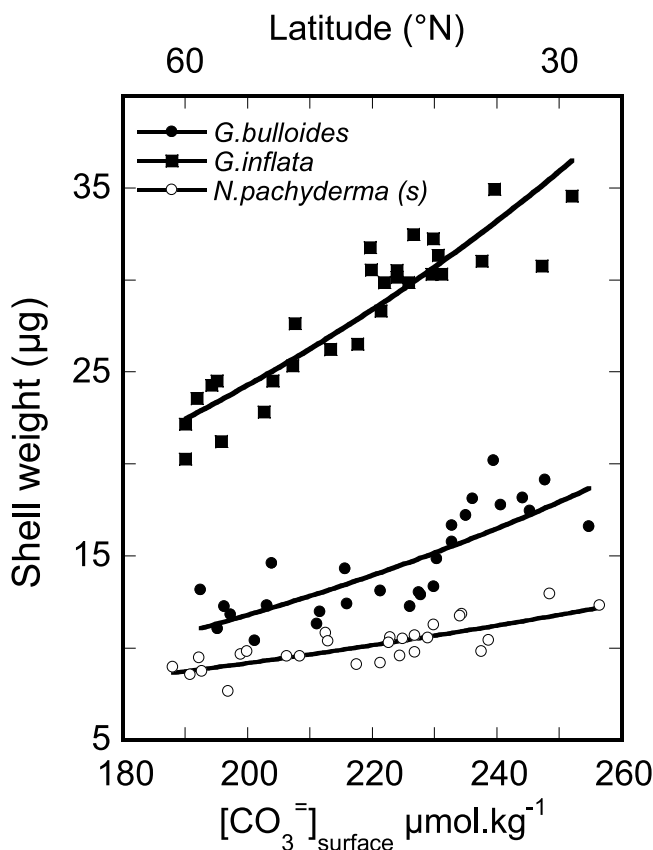


Figure 4. Shell weights for three species of planktonic foraminifera plotted against surface ocean $[\text{CO}_3^-]$ and latitude. The southward increase in surface ocean $[\text{CO}_3^-]$ in the modern North Atlantic is due mainly to the increases in temperature and the ratio of alkalinity to dissolved inorganic carbon between 60 and 30°N. The relation between weight and latitude for *G. bulloides* was used to correct for offsets between the core sites [after Barker and Elderfield, 2002].

between core sites remain constant. All subsequent calculations are performed on the adjusted data.

2.5. Calibration of Shell Weight Loss With Bottom Water $[\text{CO}_3^-]$

[16] At present there is no calibration of weight loss versus bottom water $[\text{CO}_3^-]$ available for *G. bulloides*. Broecker and Clark [2003] present a revised calibration for *G. sacculifer* after the comments of Bijma et al. [2002] concerning the depth dependency of calcite saturation. Since *G. sacculifer* and *G. bulloides* are closely ranked in Berger's [1970] solubility index (5th and 8th out of 22 respectively), we have used Broecker and Clark's [2003] calibration for *G. sacculifer* in this work.

2.6. Delta Records

[17] Shell weight data (and also benthic $\delta^{13}\text{C}$ measurements) from NEAP 8K, BOFS 5K and BOFS 8K (the deeper cores) were subtracted from data from BOFS 17K (the shallowest core) to provide records of difference (delta

records). Absolute records were first converted to a common timescale by sampling at 100 year intervals using a linear interpolation algorithm so that differences could be calculated on the common timescale. Delta records were not calculated on records later than 9.5 ka since the relative age control within the Holocene sections of these cores is not sufficiently robust. By converting absolute records to records of difference, it is possible to visually appreciate the evolving gradients between the shallow and deep sites through time. An advantage of using records of Δ shell weight is that the difference between weight records describes variations in preservation between the two sites irrespective of how the record from the shallow site is interpreted. The same is true for $\Delta\delta^{13}\text{C}$ in that global changes in $\delta^{13}\text{C}$, superimposed upon the absolute records will be cancelled out.

3. Results and Discussion

3.1. Modes of Atlantic Deep Ocean Circulation

[18] Circulation in the Atlantic ocean is thought to have three distinct modes of operation; modern, glacial and Heinrich [Sarnthein et al., 1994; Alley and Clark, 1999; Rahmstorf, 2002]. In the modern mode, North Atlantic Deep Water (NADW) forms in the Nordic Seas and flows south to fill the North Atlantic with low nutrient water down to abyssal depths. During the last glacial period (glacial mode) deep water formation shifted southward and was constrained to shallower depths (producing Upper North Atlantic Deep Water, UNADW) by the incursion of high nutrient deep water sourced from the south, so-called Southern Ocean Water (SOW). This gave rise to a strong vertical nutrient gradient in the glacial North Atlantic as compared with the rather homogenous water mass observed today [e.g., Boyle and Keigwin, 1982; Oppo and Lehman, 1993; Sarnthein et al., 1994]. During the Heinrich mode of circulation, deep water formation essentially ceases and the North Atlantic may be filled by waters of Antarctic origin [Vidal et al., 1997; Zahn et al., 1997; Elliot et al., 2002]. There is some debate over which mode of the MOC dominated during the Younger Dryas (YD) period [Jansen and Veum, 1990; Veum et al., 1992; Lehman and Keigwin, 1992; Boyle, 1995]. Evidence from the western North Atlantic basin suggests a dramatic reduction in NADW production during the YD [e.g., Boyle and Keigwin, 1987; Keigwin et al., 1991] whereas cores from the eastern North Atlantic tend to lack a distinct deep water YD signal [Jansen and Veum, 1990; Veum et al., 1992; Sarnthein et al., 1994]. This may indicate that a change in circulation at this time was restricted to the western basin or that low sedimentation rates and bioturbation have combined to remove the signal in other records [Boyle, 1995; Alley and Clark, 1999].

[19] Benthic $\delta^{13}\text{C}$ measurements were made on BOFS 8K (this study) and BOFS 17K [Bertram et al., 1995] (Figure 5). Holocene benthic $\delta^{13}\text{C}$ values are similar for the two cores (around 1‰) and agree with other studies suggesting a fairly homogeneous water column with respect to $\delta^{13}\text{C}$ and nutrient concentrations [Kroopnick, 1980; Sarnthein et al., 1994; Beveridge et al., 1995]. During the glacial and deglacial periods the two records display very different trends. The

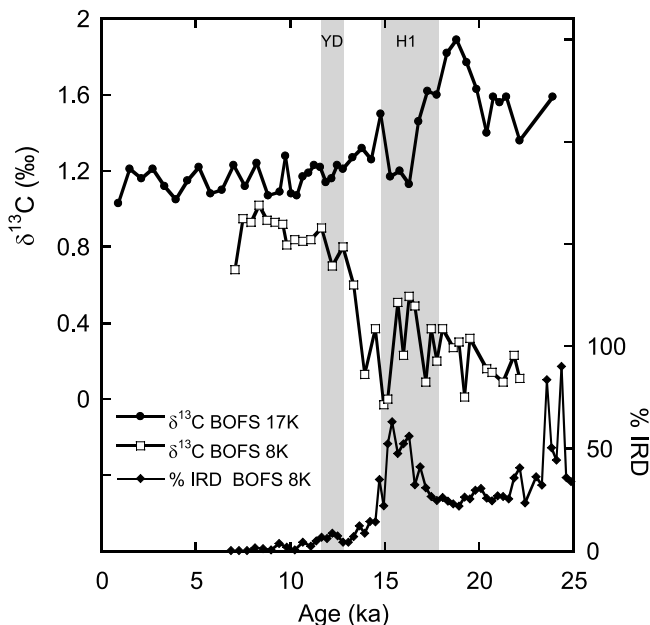


Figure 5. Benthic $\delta^{13}\text{C}$ data from BOFS 17K (*Cibicidoides ssp* [Bertram *et al.*, 1995]) and BOFS 8K (*C. wuellerstorfi*, this study). Holocene values are similar for the two cores. The glacial section is characterized by a strong gradient between the shallow and deep sites.

deeper core site reveals glacial values of around 0‰ whereas values from the shallow site are around 1.6‰. The enhanced gradient in benthic $\delta^{13}\text{C}$ between shallow and deep sites reflects the stratification of the North Atlantic with respect to $\delta^{13}\text{C}$ and nutrient distributions during of the glacial mode of circulation. A minimum in benthic $\delta^{13}\text{C}$ in BOFS 17K occurs during the period 15–18 ka (approximately equivalent to Heinrich event 1 [Sarnthein *et al.*, 2000]). This phenomenon has been observed in several other cores in the eastern North Atlantic, most predominantly from sites above about 3 km water depth, and has been inferred to represent a major reduction in NADW production in the North Atlantic [Keigwin and Lehman, 1994; Vidal *et al.*, 1997; Zahn *et al.*, 1997]. A slight increase in $\delta^{13}\text{C}$ is observed in BOFS 8K at this time. Such an increase has not been noted previously. However, this core is somewhat deeper than most published deglacial benthic $\delta^{13}\text{C}$ records that display a clear signal pertaining to H1.

[20] A recent study by Raymo *et al.* [2004], concerning vertical gradients in $\delta^{13}\text{C}$ in the N. E. Atlantic throughout the Pleistocene, makes the suggestion that northern sourced deep waters can have highly variable $\delta^{13}\text{C}$ values depending on the region and mode of deep water formation. It follows that such variability should be considered when interpreting records of benthic $\delta^{13}\text{C}$ in terms of circulation changes. While this is obviously an important point, the main focus of that study was on intermediate depths (1145–2301 m). The deepest site in the study (DSDP 607; 41°N, 33°W, 3427 m) showed fluctuations that were arguably influenced by deep waters of a southern origin [Raymo *et al.*, 2004]. As the study here focuses mainly on the difference between

intermediate and deeper waters (1150 m versus 4045 m) we suggest that we observe the varying influence of northern versus southern sourced deep waters related to changes in North Atlantic circulation.

3.2. Records of Shell Weight and Deep Ocean $[\text{CO}_3^-]$

[21] Records of shell weight from the North Atlantic reveal significant temporal variability (Figure 6). The two shallower cores (BOFS 17K and NEAP 8K) show a decrease in shell weight through the deglaciation which has been interpreted in terms of decreasing surface water $[\text{CO}_3^-]$ in parallel with increasing atmospheric CO_2 [Barker and Elderfield, 2002]. The two deeper cores show a very different trend. The Holocene records show a decrease in shell weight from about 12 ka, in line with the shallower cores, but the deglaciation records show an increase in shell weights, opposite to the shallower sites. Shell weights of glacial age in BOFS 8K, the deepest core, are similar to those from the early Holocene in the same core. If these results were interpreted solely in terms of dissolution it would be concluded that no appreciable change in bottom water $[\text{CO}_3^-]$ occurred between glacial and Holocene times in the deep North Atlantic. From comparison with the shallowest site, BOFS 17K, it is clear, however, that glacial age shells from BOFS 8K must have suffered significant weight loss through dissolution. SEM imagery provides qualitative evidence of this. A glacial age sample of *G. bulloides* from BOFS 17K shows a heavily calcified outer structure with clear evidence of spine bases, typical of well-preserved shells (Figures 7a and 7b) [Dittert and Henrich, 2000], whereas a glacial sample from BOFS 8K shows significant smoothing of the outer surface, loss of spine bases and etching of interpore areas (Figures 7c and 7d), indicative of partial dissolution [Dittert and Henrich, 2000].

[22] The Δ shell weight records display several significant features (Figures 8 and 9): a maximum at around 19 ka, followed by an overall decrease during the termination to a minimum in the early Holocene at around 10 ka. A sharp decrease occurs just prior to 17 ka, during the first half of the Heinrich 1 (H1) ice-rafting event. Using the alternate age model for *G. bulloides*, as described in section 2.2, results in a shift in the decrease of Δ shell weight toward the latter half of H1 (dashed line in Figure 9). These features are also apparent in the benthic $\Delta\delta^{13}\text{C}$ record (Figure 9).

[23] Broecker and Clark [2001b] applied the shell weight method to a glacial time slice to reconstruct deep water $[\text{CO}_3^-]$ in the equatorial Atlantic. They observed a gradient in $[\text{CO}_3^-]$ in the glacial Atlantic similar to that expected from nutrient proxy reconstructions. However, shell weights from intermediate depths, being heavier than their modern counterparts, were interpreted as representing higher glacial $[\text{CO}_3^-]$ at these depths. Following Barker and Elderfield [2002], Broecker and Clark [2003] recalculated their shell weight data to allow for changes in initial weight and found that their weight measurements no longer demanded enhanced preservation at intermediate depths during glacial times. According to our study, $\Delta[\text{CO}_3^-]$ ($= [\text{CO}_3^-]_{\text{in situ}} - [\text{CO}_3^-]_{\text{saturation}}$) at the deepest site, BOFS 8K, reached an extreme value of $-24 \mu\text{mol.kg}^{-1}$ at about 19 ka and had a mean value of about $-16 \mu\text{mol.kg}^{-1}$ during the Last Glacial

Maximum (LGM) period (taken as 19–23 ka after *Mix et al.* [2001]) (Figure 9). This may be compared with an estimate of $-29 \mu\text{mol.kg}^{-1}$ at 4 km in the equatorial Atlantic during the LGM [*Broecker and Clark, 2003*]. Modern values

of $[\text{CO}_3^-]$ in the deep equatorial Atlantic are about $30 \mu\text{mol.kg}^{-1}$ lower than at 52°N .

3.3. Glacial-Interglacial Changes in Shell Weight

[24] The records of shell weight shown in Figure 6 demonstrate clearly the interaction between initial weight variability and postdepositional weight loss due to dissolution. The similarity between all weight records during the Holocene suggests that their preservation histories were similar throughout this period although quite different before about 10 ka. It may be suggested that the glacial weight maximum from 20–17 ka, observed in the two shallowest cores, is a result of better preservation rather than changes in surface ocean $[\text{CO}_3^-]$ and initial weight variability. However, the site of NEAP 8K (2360 m) lies at or just below the glacial transition between UNADW and SOW [e.g., *Oppo and Lehman, 1993*]. This suggests that glacial age shells in this core would, if anything, be lighter were dissolution the only control on shell weight. Additionally, if core-top shells in BOFS 17K (the shallowest core) had lost 45% by weight through dissolution (the glacial to Holocene difference) we may expect to observe physical evidence of dissolution, such as that illustrated in Figures 7c and 7d. Moreover, we should expect deeper sites to have dissolved to an even greater extent (with large offsets between all depths), due to the pressure dependence of CaCO_3 solubility (given a modern vertical profile of in situ $[\text{CO}_3^-]$). We do not observe these features.

[25] A potentially complicating factor is that of varying pore water/bottom water offsets in $[\text{CO}_3^-]$. CO_2 released during respiration of organic matter (so called respiratory CO_2) within the upper few centimeters of the sediment column will tend to lower $[\text{CO}_3^-]$ within sediment pore waters and may lead to dissolution of carbonate [*Emerson and Bender, 1981*]. Assessing the influence of pore water dissolution in sediment cores is nontrivial. The extent of dissolution will depend variously on the relative rain rates of inorganic and organic carbon, the rates of oxic versus anoxic carbon degradation in the sediments and presumably the saturation state of overlying bottom waters [*Archer, 1991*]. Attempts to reconstruct glacial productivity in the North Atlantic have given ambiguous results [e.g., *Coppedge and Balsam, 1992; Manighetti and McCave, 1995*]. Organic carbon contents as a paleoproductivity indicator is complicated by the presence of terrigenous organic matter [*Villanueva et al., 1997*] but nevertheless should give some indication of the extent of respiratory CO_2 available for pore water dissolution (although arguably it is the amount of organic matter that has been oxidized which is the controlling factor here, rather than that remaining). The organic and inorganic carbon deposition rates for the

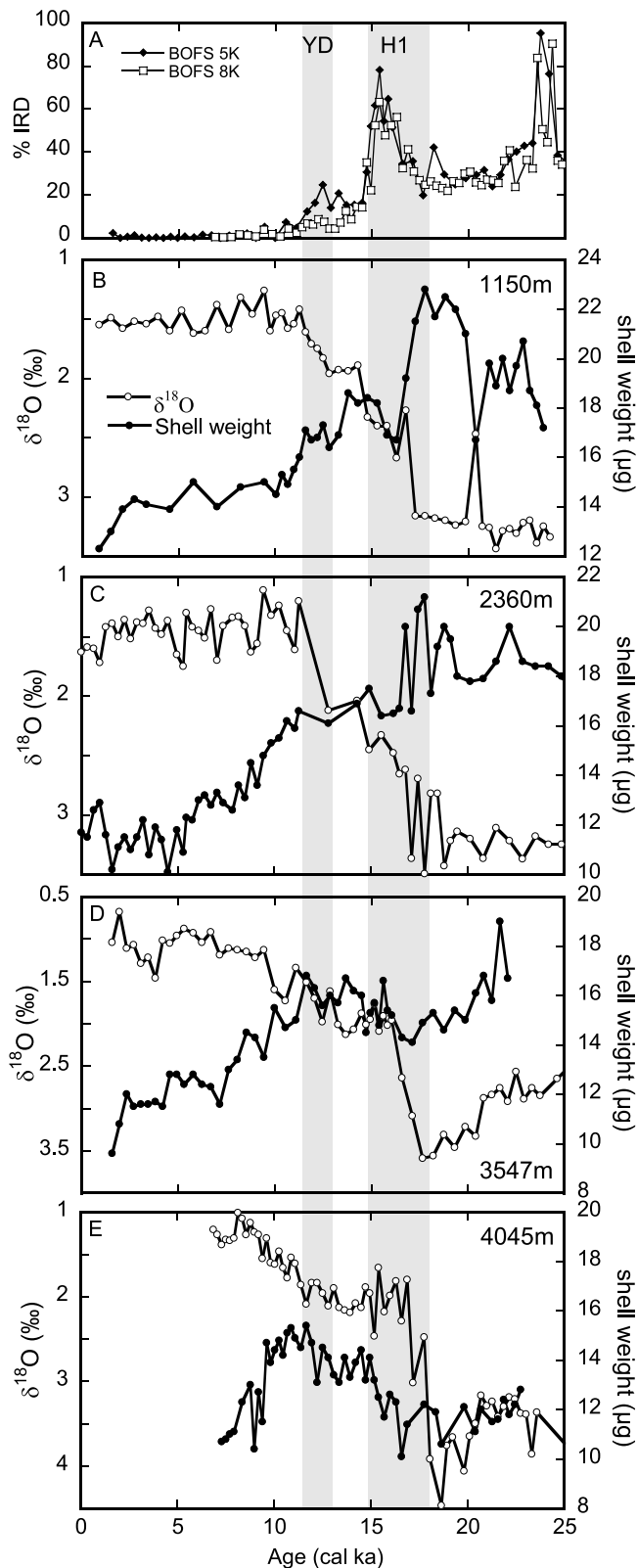


Figure 6. Downcore records of shell weight and $\delta^{18}\text{O}$ for *G. bulloides* (corrected for latitudinal offsets; see section 2.4). Differences between the records are due to differences in preservation between the sites. Deeper sites show a strong decrease in weight relative to the shallower sites during the glacial period. This reflects the incursion of corrosive southern sourced deep water into the North Atlantic at this time.

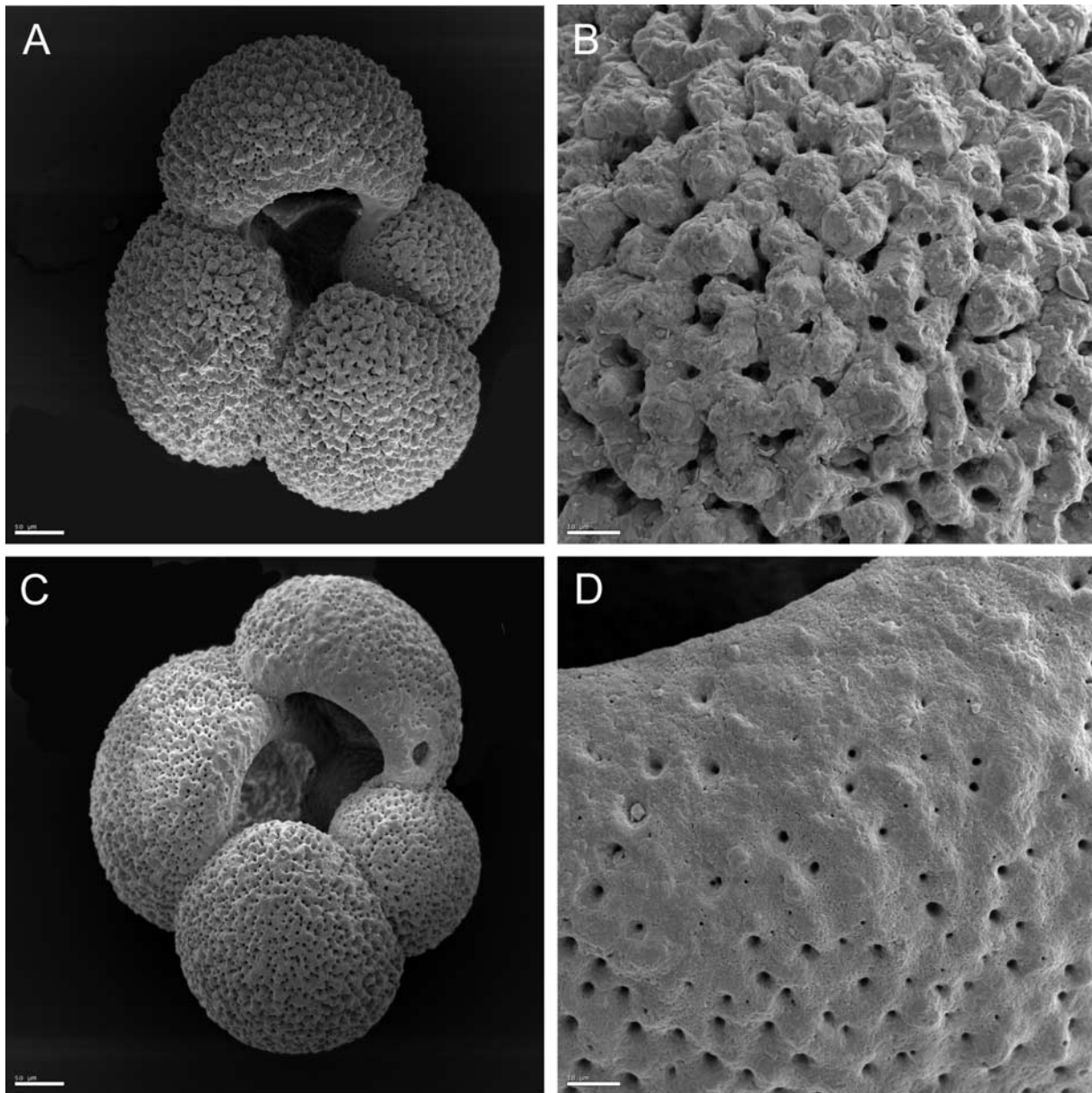


Figure 7. SEM images of *G. bulloides* from the glacial sections of (a and b) BOFS 17K and (c and d) BOFS 8K. The specimen from the shallow site (BOFS 17K) shows a well-preserved heavily calcified external structure as compared to the deeper specimen which shows clear evidence of dissolution. Scale bars are 50 μm in Figures 7a and 7c and 10 μm in Figures 7b and 7d.

BOFS cores were studied by *Manighetti and McCave* [1995]. Their results suggest that the flux of organic matter in BOFS 17K was less (of the order 30–40%) during the glacial period than during the Late Holocene (that in BOFS 8K was increased by approximately the same proportion although BOFS 5K also experienced a decrease). This may lead to the suggestion that increased glacial age shell weights in BOFS 17K are a result of decreased dissolution in pore waters. However, the flux of inorganic carbonate to all locations in the study was reduced by an even greater extent (around 70%) during the glacial period. This would mean that the rain ratio of organic to inorganic carbon most likely increased, leading to a possible increase in pore water dissolution in all

sites. This discussion does not take into account more recent work by *Armstrong et al.* [2002] which suggests that the relative rain rates of organic and inorganic carbon arriving at the seafloor may actually be rather constant, due to the “ballasting” of organic matter by CaCO_3 .

[26] Recent genetic studies on planktonic foraminifera have revealed the existence of several subspecies (cryptospecies, or genotypes) of *G. bulloides* [*Darling et al.*, 2003]. If different subspecies had fixed but distinct shell weights, it may be argued that the mean weight changes observed in this study are a result of changing surface water conditions and the relative contributions from two or more subspecies rather than changes in surface ocean carbonate chemistry

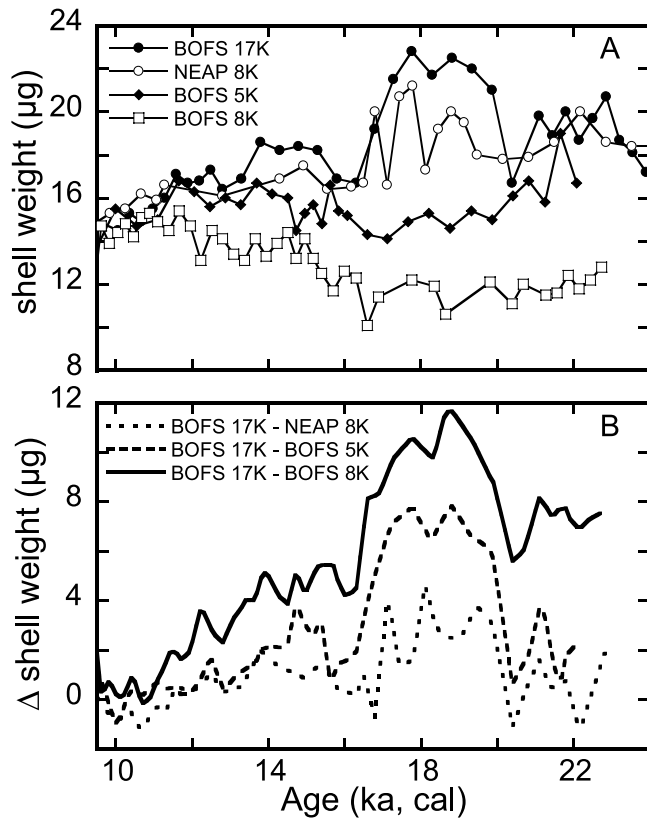


Figure 8. (a) Shell weights (corrected for latitudinal offsets) and (b) Δ shell weight records for the four cores. Calculation of Δ records is explained in section 2.6.

and or dissolution. In light of this possibility, several populations of *G. bulloides* from down core samples in ODP core 982 (57.5°N, 15.9°W, 1134 m), situated very near to BOFS 17K, were analyzed for individual shell weights (Figure 10). The results from three intervals with differing mean population weight suggest a normal distribution within each sample. If the changes in weight were due to “end-member” mixing, a bimodal distribution may be expected.

[27] The presence of nutrient enriched, low $\delta^{13}\text{C}$, waters at depth in the glacial North Atlantic (Figure 5) suggests a lowering of deep water $[\text{CO}_3^{2-}]$. This prediction is borne out by measurements of Cd/Ca and Zn/Ca in benthic foraminifera [Marchitto *et al.*, 2002] and time slice reconstructions of bottom water $[\text{CO}_3^{2-}]$ [Broecker and Clark, 2001b, 2003]. A lowering of $[\text{CO}_3^{2-}]$ in the glacial deep North Atlantic would explain the dissolution of foraminiferal tests at the sites of BOFS 5K and BOFS 8K (Figure 8). The apparent weight loss in NEAP 8K during the period 17–20 ka may be a result of dissolution. Alternatively it may be an artefact of in-mixing of lighter shells since *G. bulloides* is relatively sparse in this section of the core. The same cannot be true of the two deeper cores since this process would tend to increase the average weight of glacial aged shells in these cores. In these cores the glacial weight loss is most likely to be a result of dissolution.

[28] Shell weights in the two shallowest cores show a decrease during H1. This could be interpreted as dissolution

in the shallower cores, in response to a strongly reduced MOC during Heinrich events [e.g., *McManus et al.*, 2004], rather than a decrease in initial shell weight. This would mean that deeper waters might not have experienced an increase in $[\text{CO}_3^{2-}]$ to the extent suggested by the Δ shell weight records. While this possibility is not unrealistic (benthic $\delta^{13}\text{C}$ at the site of BOFS 17K does suggest the presence of less ventilated waters during H1) we would still expect a decrease in initial shell weight during H1 (shells in the shallow cores are lighter at the end of H1 than at the beginning). Since the two deeper records show a general increase in shell weight during the same interval, this still suggests that bottom waters at these sites became less corrosive during H1.

[29] Holocene shell weights in BOFS 17K appear heavier than in the other cores after about 9 ka (Figure 6). This could mean that dissolution has occurred in the three deeper sites for most of the Holocene but this would be contrary to studies which demonstrate that dissolution occurs only below about 4 km in the modern North Atlantic [Broecker and Clark, 2003]. As mentioned earlier, changes in surface temperature or nutrient gradients between sites may also play a role in determining offsets in shell weight between the cores. Since we have little way of determining and quantifying what such effects might be we can only highlight them as an uncertainty within our interpretation. It is clear that a future attempt to repeat this approach should exploit a depth transect of cores that underlies a very restricted surface area so as to further minimize possible offsets in initial shell weights. On the other hand, the major relative changes in (planktonic) shell weights between the shallow and deep sites during deglaciation are matched by relative changes in benthic $\delta^{13}\text{C}$ between the same sites (Figure 9). This lends support to the suggestion that in general, weight differences are caused by the superposition of a “benthic” signal (i.e., dissolution) on the shell weight records from deeper sites rather than differences in initial weight.

3.4. Records of $\Delta\delta^{13}\text{C}$ and Δ Shell Weight

[30] From the record of $\Delta\delta^{13}\text{C}$ it can be inferred that the North Atlantic was most stratified with respect to nutrients between 18 and 20 ka (Figure 9). The decrease in benthic $\delta^{13}\text{C}$ in BOFS 17K during H1 corresponds to a slight increase in $\delta^{13}\text{C}$ in BOFS 8K, resulting in a decrease in the gradient between the two sites. A decrease in stratification might reflect increased vertical mixing in a stagnating water column or the increased dominance of southern sourced deep water throughout the North Atlantic basin during H1. In the latter case, the increase in $\delta^{13}\text{C}$ at the deeper site would suggest that SOW was better ventilated with respect to its nutrient content during H1 than during the LGM. It is important to keep in mind that proxies for chemical stratification do not describe dynamic properties of oceanic circulation, i.e., increased stratification does not demand a sluggish circulation. In this case, maximum stratification occurred at the end of the LGM (18–20 ka) (while NADW production was active in the form of UNADW) and decreased throughout H1, presumably as a result of the reduction in UNADW production. The decrease in stratification into the Early Holocene then reflects the resumption of modern NADW formation.

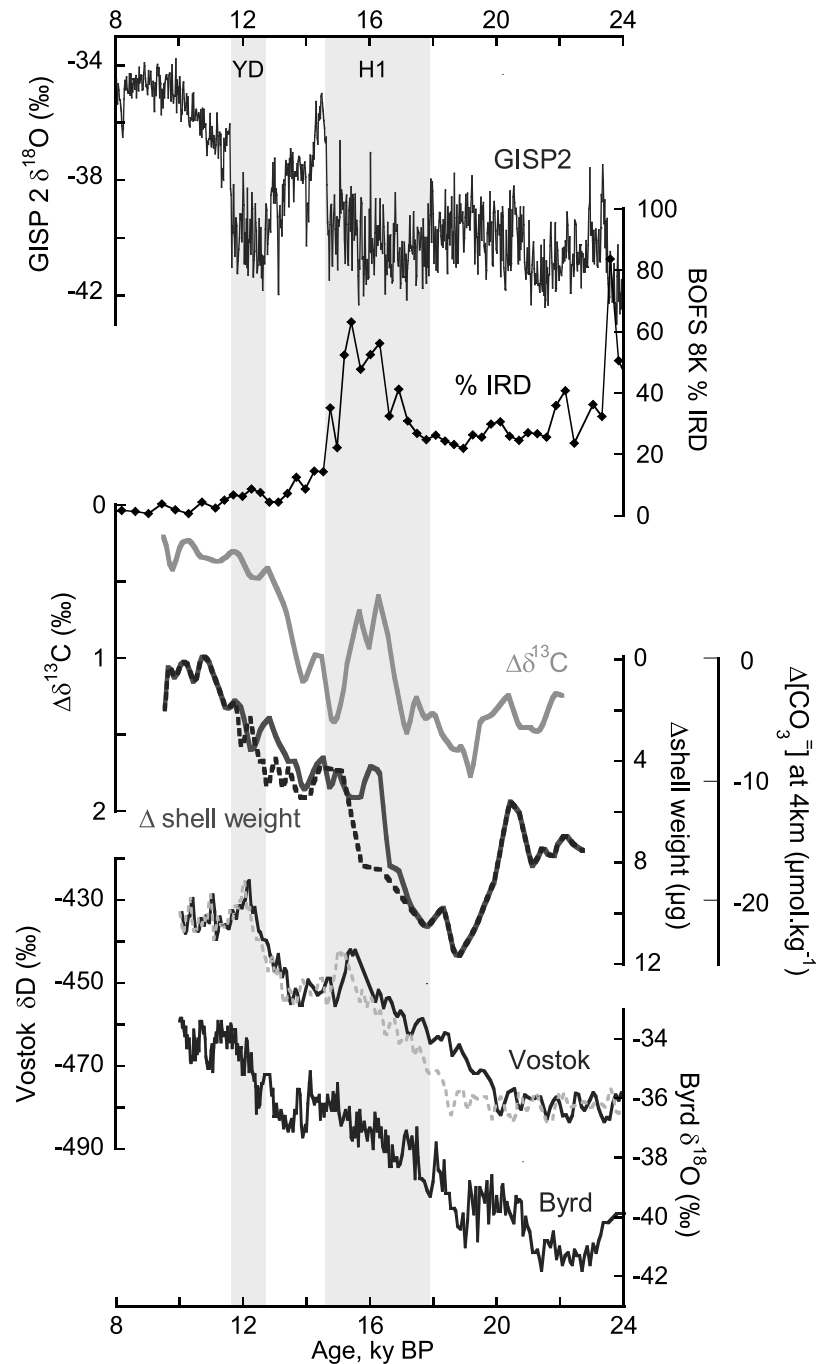


Figure 9. Plots of $\Delta\delta^{13}\text{C}$ and Δ shell weight for BOFS 17K and 8K. See text for explanation and note reversed scale. Dashed line in the Δ shell weight record represents the alternative age model constructed specifically for *G. bulloides* (see section 2.2). Also shown are $\delta^{18}\text{O}$ values from the GISP 2 [Stuiver and Grootes, 2000] and Byrd [Johnsen *et al.*, 1972] ice cores on the GISP 2 timescale [Meese *et al.*, 1994; Blunier and Brook, 2001] and the δD record from Vostok [Jouzel *et al.*, 1987]. The Vostok record is plotted on the GRIP timescale [Blunier *et al.*, 1998] (dashed line) and on the GISP 2 timescale via CH_4 correlation [Blunier and Brook, 2001] (solid line). See color version of this figure at back of this issue.

[31] A maximum in $\Delta\delta^{13}\text{C}$ occurs during the period 13.5–15.5 ka, spanning the latter half of H1 and the Bølling interstadial. According to Sarnthein *et al.* [2000] the Holocene mode of the MOC started with full intensity at the onset of the Bølling interstadial at 14.7 ka (prior to a

partial collapse near the end of the YD). The maximum in our record of $\Delta\delta^{13}\text{C}$ at this time appears to be mainly a function of deep water variability (the record of BOFS 8K displays a temporary return to glacial values) although values from the shallow site are slightly higher than

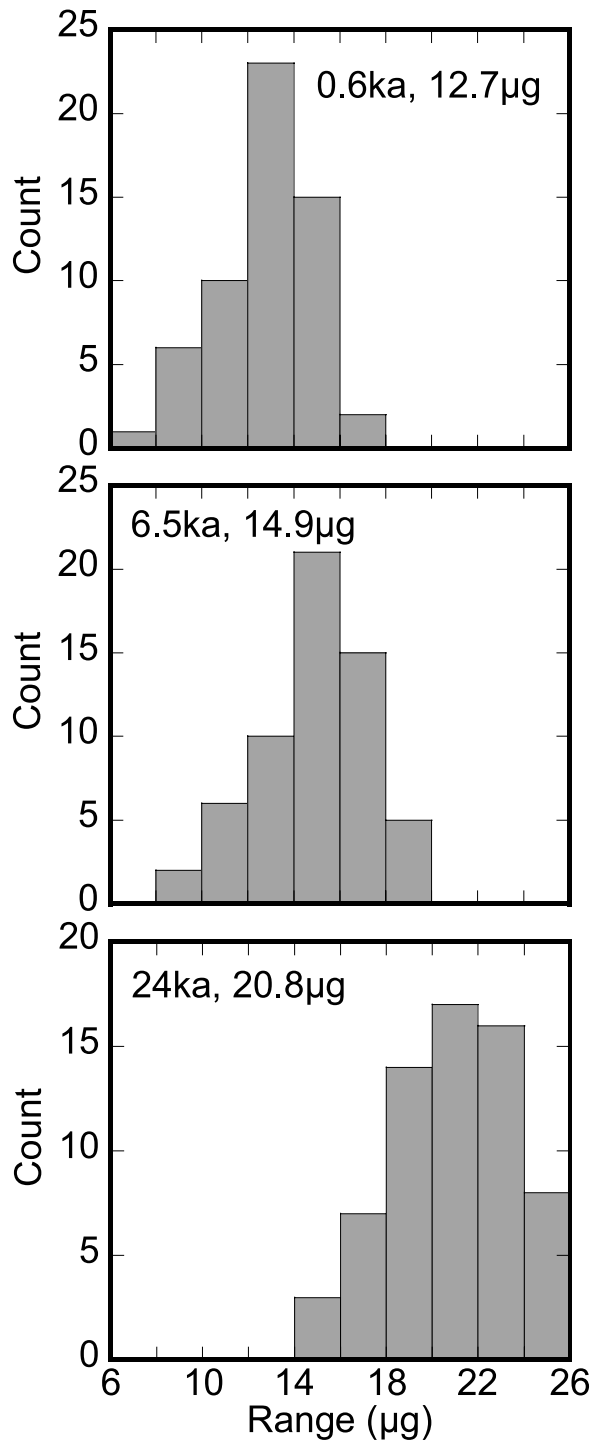


Figure 10. Histograms of individual shell weights of *G. bulloides* picked from ODP core 982 showing normal distribution of shell weights as the mean population weight increases from 0.6 ka (12.7 µg) to 24 ka (20.8 µg).

modern. This suggests that although deep water formation may have been active at this time, it had not reached the full extent of its modern mode with deeper waters still poorly ventilated with respect to intermediate and shallow depths. The record of $\Delta\delta^{13}\text{C}$ reaches its minimum value (similar to

that expected today) at around 10 ka, suggesting that the modern mode of MOC was probably active by this time.

[32] While the record of Δ shell weight describes the difference in preservation between the deep and shallow sites, it cannot be interpreted directly as the gradient in deep water $[\text{CO}_3^-]$. This is due to the pressure dependence of CaCO_3 solubility. The concentration of CO_3^- in the modern North Atlantic is fairly constant down to abyssal depths and at present no part of the water column is undersaturated with respect to CaCO_3 (calcite). If the profile of $[\text{CO}_3^-]$ was shifted toward lower values without any change in gradient, deeper waters would become undersaturated significantly earlier than those at shallower depths. Thus dissolution at deeper depths does not necessarily imply a change in the gradient of $[\text{CO}_3^-]$. The site of BOFS 17K (1150 m) is presently bathed by NADW and was presumably under the influence of UNADW during the last glacial period. Since surface water $[\text{CO}_3^-]$ was probably higher during the glacial period [Sanyal *et al.*, 1995; Lea *et al.*, 1999], we should expect glacial UNADW to have had a similar or even higher $[\text{CO}_3^-]$ than modern NADW. This would suggest that the site of BOFS 17K was safe from dissolution throughout the period of interest (but see section 3.3). If no weight loss has occurred in the shallow core, it may be argued that the calculated record of Δ shell weight is a record of weight loss in BOFS 8K and is therefore a record of varying $[\text{CO}_3^-]$ at the deeper site (Figure 9).

[33] If the North Atlantic above about 2 km depth experienced an increase in $[\text{CO}_3^-]$ during the glacial then dissolution in the deeper sites (BOFS 5K and 8K) does in fact reflect a change in the gradient of $[\text{CO}_3^-]$. This was observed by Broecker and Clark [2001b, 2003] and again reflects the incursion of SOW at depth. Our record of Δ shell weight may then be viewed as a function of SOW incursion into the deep northeast Atlantic. Since there is little known about absolute CO_3^- concentrations of deep waters forming during the last glacial, the record of Δ shell weight should be considered as a qualitative record of SOW invasion. More strictly, the record is a combined function of the chemistry and northward flux of southern sourced waters.

[34] The record of Δ shell weight (Figure 9) suggests that maximum incursion of low $[\text{CO}_3^-]$ SOW occurred at around 19 ka, at the same time as the maximum in stratification recorded by $\Delta\delta^{13}\text{C}$. Since this was a period of maximum dissolution in the deeper sites, due to maximum incursion of southern sourced deep water, we may expect to see a decrease in benthic $\delta^{13}\text{C}$ in BOFS 8K; in fact a modest increase is observed after about 20 ka. A possible explanation for this discordancy may be the decoupling of $\delta^{13}\text{C}$ and $[\text{CO}_3^-]$. $\delta^{13}\text{C}$ of dissolved inorganic carbon (DIC) in seawater is influenced by air-sea exchange, independent of biological interactions. In the modern ocean, high southern latitude surface waters have the highest (positive) values of $\delta^{13}\text{C}_{\text{as}}$ (the air-sea exchange signature), mainly due to their low temperatures [Lynch-Stieglitz *et al.*, 1995]. During glacial times the extent of sea-ice around Antarctica was significantly increased relative to today [Crosta *et al.*, 1998; Gersonde *et al.*, 2003]. This would have led to decreased air-sea exchange and therefore a decrease in $\delta^{13}\text{C}_{\text{as}}$ [Stephens and Keeling, 2000]. The expected decrease of

benthic $\delta^{13}\text{C}$ in BOFS 8K would have coincided with the onset of sea-ice reduction (and corresponding increase in $\delta^{13}\text{C}_{\text{as}}$) in the Southern Ocean at around 19 ka as reconstructed by *Shemesh et al.* [2002]. Thus the two signals may act to cancel one another out. However, the same mechanism may well counter the argument for more corrosive bottom waters at this time since an increase in air-sea exchange (and concomitant decrease in surface ocean $p\text{CO}_2$) would most likely lead to an increase in $[\text{CO}_3^-]$.

[35] Δ shell weight decreases during the first half of H1 in line with the decrease in $\Delta\delta^{13}\text{C}$. Using the alternate age model for *G. bulloides*, as described in section 2.2, results in a shift in the decrease of Δ shell weight toward the latter half of H1 (dashed line in Figure 9). The decrease in weight loss could reflect a decline in southern sourced waters during H1. Alternatively the decrease in Δ shell weight may be the result of an increase in the $[\text{CO}_3^-]$ of SOW perhaps due to an increase in the flux (and corresponding reduction in aging) of this water mass during H1. If this is the case, it suggests that the shutdown of NADW production in the North Atlantic was accompanied by enhanced production of deep waters sourced from the south; a scenario reminiscent of the bipolar seesaw (or polar asynchronicity) hypothesis [*Blunier et al.*, 1998; *Broecker*, 1998]. In light of this we have plotted $\delta^{18}\text{O}$ values (indicative of temperature) from the GISP 2 [*Stuiver and Grootes*, 2000] and Byrd [*Johnsen et al.*, 1972] ice cores and δD from Vostok [*Jouzel et al.*, 1987] together with our delta weight and $\Delta\delta^{13}\text{C}$ records from the N. Atlantic (Figure 9). Comparison of the Δ shell weight curve with the isotope record from Greenland (GISP2) reveals little similarity. However, there is some similarity between the Δ shell weight and Antarctic temperature records. Perhaps it is not too surprising that the record of Δ shell weight, which is strongly influenced by the flux and chemistry of deep waters originating from the Antarctic region, should resemble Antarctic climatic variability.

[36] A recent study by *McManus et al.* [2004], based on $^{231}\text{Pa}/^{230}\text{Th}$ ratios, provides compelling evidence that meridional overturning in the western North Atlantic at least, was shut down completely during H1 (~ 17.5 – 14.7 ka). This scenario would lend support to the notion of a stagnating water column at this time rather than an increase in the presence of southern sourced deep water. Another study by *Skinner and Shackleton* [2004] suggests that ^{14}C ventilation ages of Atlantic deep waters may have significantly increased during the latter stages of H1, although this would not necessarily demand a more sluggish circulation (see *Skinner and Shackleton* [2004] for an explanation of the uncertainties involved in interpreting past ^{14}C ventilation ages). On the other hand, a study of paired benthic Mg/Ca and $\delta^{18}\text{O}$ measurements from a core off the Iberian margin (MD99-2334; 37.8°N , 10.2°W , 3146 m) suggests that the deep N. E. Atlantic was dominated by cold, fresh, southern sourced deep waters during Northern Hemisphere stadial events over the last 34 ka [*Skinner et al.*, 2003]. This argues toward a more active circulation during H1. Our shell weight data adds some independent proxy evidence for a gradual reinforcement of ventilation of the deep Atlantic during H1, possibly by southern source water.

[37] The record of Δ shell weight shows a plateau during the maximum in $\Delta\delta^{13}\text{C}$ in the latter part of H1 and the Bølling interstadial. Use of the alternative age model specifically for *G. bulloides* shifts this plateau to within the Bølling and earliest YD. The fact that dissolution occurred at the deeper sites during this period (weight loss is also apparent in BOFS 5K) confirms that the MOC had probably not fully recovered to its Holocene mode by this time.

[38] Records of benthic foraminiferal $\delta^{13}\text{C}$ and Cd/Ca from the west and central Atlantic [e.g., *Boyle and Keigwin*, 1987; *Keigwin and Lehman*, 1994; *Marchitto et al.*, 1998] have been used to demonstrate a weakening of the MOC during the Younger Dryas (YD) period (11.6–12.9 ka [*Hughen et al.*, 1996; *Stuiver and Grootes*, 2000]). While our records of $\Delta\delta^{13}\text{C}$ and Δ shell weight suggest that circulation had not yet recovered to its Holocene mode during the YD they do not show any particular features that might be interpreted as a marked change in circulation at this time. A possible explanation for this is that the sedimentation rate in our cores is too low to detect a YD signal or that any signal has been lost due to bioturbation. Alternatively, the records suggest that any change in the MOC during the YD did not strongly affect the $\delta^{13}\text{C}$ gradient and deep water $[\text{CO}_3^-]$ of the northeast Atlantic.

[39] Core-top weight data in the equatorial Atlantic imply that weight loss through dissolution commences below about 4 km in the modern ocean [*Broecker and Clark*, 2003]. The record of Δ shell weight attains its minimum value (approximately zero) at around 10 ka (Figure 9) and suggests that a modern mode of MOC was probably active by about 10 ka.

4. Conclusions

[40] Records of shell weight in the North Atlantic demonstrate the interplay between initial weight variability and postdepositional weight loss. In order to interpret a record of shell weight in terms of preservation variability, it is necessary to account for any changes in initial shell weight. It has been demonstrated that these two signals can be differentiated by use of multiple records from proximal sites and different water depths.

[41] Results suggest that maximum incursion of SOW and stratification of the North Atlantic occurred between about 18–20 ka defining the full glacial mode of the Atlantic MOC. During H1, chemical stratification was reduced and the deep North Atlantic became less corrosive. This could be interpreted as a reduction in the presence of SOW at this site and the homogenization of the North Atlantic while NADW production was switched off. Alternatively, it may reflect a change in the flux and chemistry of SOW during the period, suggesting enhanced production of southern sourced deep waters during the shutdown of NADW. The modern mode of the MOC was attained by about 10 ka at the start of the Holocene period.

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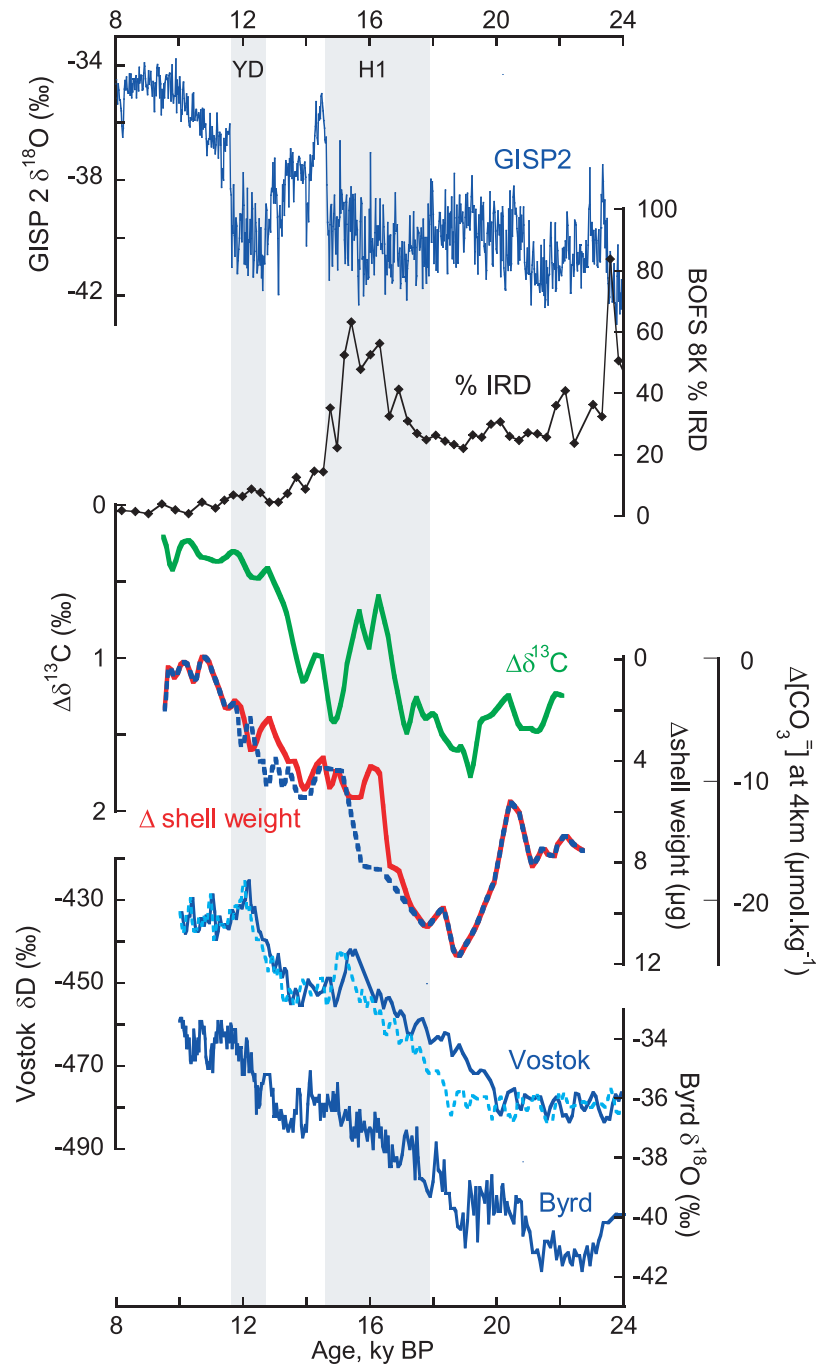


Figure 9. Plots of $\Delta\delta^{13}\text{C}$ and $\Delta\text{shell weight}$ for BOFS 17K and 8K. See text for explanation and note reversed scale. Dashed line in the $\Delta\text{shell weight}$ record represents the alternative age model constructed specifically for *G. bulloides* (see section 2.2). Also shown are $\delta^{18}\text{O}$ values from the GISP 2 [Stuiver and Grootes, 2000] and Byrd [Johnsen et al., 1972] ice cores on the GISP 2 timescale [Meese et al., 1994; Blunier and Brook, 2001] and the δD record from Vostok [Jouzel et al., 1987]. The Vostok record is plotted on the GRIP timescale [Blunier et al., 1998] (dashed line) and on the GISP 2 timescale via CH_4 correlation [Blunier and Brook, 2001] (solid line).