Phasing of Millennial Climate Events and Northeast Atlantic Deep-Water Temperature Change Since 50 ka BP

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The observation that Greenland and Antarctic temperatures have followed a specific 'asymmetrical' pattern on millennial time-scales sets rigid constraints on any viable theory of abrupt climate change. The further observation that the very same asymmetry is also reflected in planktonic and benthic $\delta^{18}O$ measurements from the Northeast Atlantic has extended this constraint to include a specific response in the ocean. Here we present records of deep-water temperature, $\delta^{18}O$ and $\delta^{13}C$ variability from the Northeast Atlantic that help to shed light on the links between overturning circulation perturbations, sea-level variability and inter-hemispheric climate change on millennial time-scales. Results indicate that while deep-water temperatures in the Northeast Atlantic have tracked Greenland climate, the $\delta^{18}O$ signature of local deep-water ($\delta^{18}O_{dw}$) has varied in a manner more reminiscent of Antarctic temperature variability. The previously identified correspondence of Antarctic warm events with benthic $\delta^{18}O$ minima in the Northeast Atlantic is thus found to apply specifically to $\delta^{18}O_{dw}$ minima, and to extend beyond Marine Isotope Stage 3 to the entirety of the last 50 ka. It is impossible to reconcile completely the Iberian Margin $\delta^{18}O_{dw}$ record with existing reconstructions of millennial sea-level variability, leading to the conclusion that a significant portion of the $\delta^{18}O_{dw}$ record must represent local hydrographic change. This is supported by benthic $\delta^{13}C$ measurements, which suggest the incursion during Greenland stadials of a colder, low-$\delta^{18}O$ and low-$\delta^{13}C$ water-mass, of presumed Antarctic origin. These observations confirm a one-to-one coupling of inter-hemispheric climate events with changes in the Atlantic overturning circulation, but fail to rule in or out a unique mechanism by which they were triggered.

INTRODUCTION

No theory purporting to explain past 'abrupt' millennial climate change can be said to be complete if it fails to account for the distinctly 'asymmetrical' pattern of Greenland and Antarctic climate change [*Blunier and Brook*, 2001; *EPICA community members*, 2006]. As yet, agreement upon a precisely formulated explanation for why Greenland and Antarctic temperatures have remained coupled in this way

(i.e. how millennial climate changes have been effected) remains lacking. Nevertheless, a growing body of proxy and modelling evidence tends to support the hypothesis that changes in the Atlantic meridional overturning circulation (MOC) have at least been implicated in, if not indeed responsible for orchestrating, the asymmetrical pattern of climate change observed at high latitudes [e.g. *Ganopolski and Rahmstorf*, 2001; *McManus et al.*, 2004; *Piotrowski et al.*, 2004; *Schmittner et al.*, 2003; *Skinner and Shackleton*, 2004]. The observations of *Shackleton et al.* [2000] (reproduced and expanded in Figure 1) are of particular interest in this regard, for they strongly suggest that robust clues as to the roles of the ocean circulation and the cryosphere in mediating (or responding to) inter-hemispheric climate change are encrypted in the

North Atlantic benthic calcite $\delta^{18}O$ record. Extracting these clues requires a 'de-convolution' of the Iberian Margin benthic $\delta^{18}O$ record into its thermodynamic (deep-water temperature) and water-$\delta^{18}O$ components, the latter of which may combine a global glacioeustatic signal with local hydrographic effects due to changes in the T - $\delta^{18}O$ signature of deep-water bathing the core site. There is already some evidence [*Chappell*, 2002; *Siddall et al.*, 2003] that only part of the millennial variability in the benthic $\delta^{18}O$ record shown in Figure 1 may be accounted for by millennial ice-volume fluctuations, and hence that the remainder must be due to local hydrographic variability. Here we make use of deep-water temperature estimates, derived from benthic Mg/Ca ratios, to directly evaluate the contribution of local hydrographic change to the relationships first observed by *Shackleton et al*. [2000], and to consider the possible implications of these observations for the mechanisms of inter-hemispheric climate change.

MATERIALS AND METHODS

Stable oxygen and carbon isotope and Mg/Ca measurements have been performed on the infaunal benthic foraminifer *Globobulimina affinis*, picked from core MD99-2334K (37°48′N, 10°10′W; 3,146m) and core MD01-2444 (37°33′N, 10°08′W; 2,460 m), both retrieved from the Iberian Margin. For these analyses, up to 40 *G. affinis* individuals were picked from the 250-300 μm size fraction, crushed between clean glass plates, and purified according to the protocol of [*Barker et al.*, 2003]. Samples were then split into separate aliquots for stable isotope and minor element analysis. Additional (replicate) analyses were performed by N.J. Shackleton on *G. affinis* samples from MD01-2444 that were destined for stable isotope analysis alone, and thus cleaned accordingly. As many specimens as possible of the epibenthic foraminifer *Planulina wuellerstorfii* (>212 μm) and approximately 60 individuals of the planktonic foraminifer *Globigerina bulloides* were also picked from both cores for stable isotope analysis only. The *G. bulloides* isotope results have been previously reported by *Vautravers and Shackleton* [2006]. All foraminifer samples were weighed using a Mettler Toledo Deltarange microbalance, with an estimated precision of ~5 μg.

Minor element analyses were carried out on an ICP-AES (Vista Inc.) as described by *de Villiers et al.*, [2002]. Samples
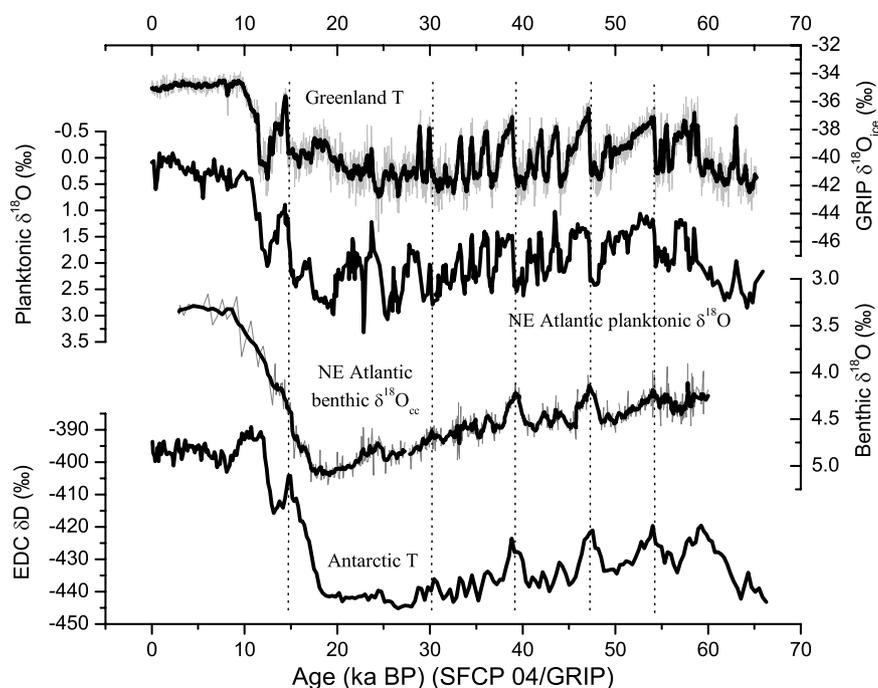

Figure 1. Phasing of millennial changes in the Greenland (top) and Antarctic (bottom) temperature proxy records aligned and compared with the pattern of planktonic (upper middle) and benthic (lower middle) $\delta^{18}O$ variability on the Iberian Margin, following *Shackleton et al*. [2000]. All records have been placed on the SFCP04-GRIP age-scale due to *Shackleton et al*. [2004]. The Antarctic EPICA Dome C (EDC) record has been placed on a consistent age-scale via alignment with the BYRD record, previously synchronised with the Greenland stratigraphy by *Blunier et al*. [2000]. The marine $\delta^{18}O$ records consist of concatenated results from cores MD99-2334K [*Skinner et al.*, 2003] and MD01-2444 [N.J. Shackleton, personal communication; *Vautravers and Shackleton*, 2006]. Vertical dotted lines indicate the timing of selected Greenland stadial–interstadial transitions for reference.

were screened for contamination by reference to Fe/Ca and Mn/Ca ratios (which do not consistently co-vary with benthic Mg/Ca), and dissolution was ruled out as a control on Mg/Ca variability by reference to planktonic shell weight trends (which were negatively correlated with benthic Mg/Ca). Partial dissolution will tend to result in the preferential loss of Mg-rich calcite, thus lowering the Mg/Ca ratio as well as the average weight of the remaining tests [*Barker et al.*, 2005]. Here the opposite is observed, with lighter planktonic shells (more prone to dissolution than benthics) coinciding with higher benthic Mg/Ca.

The calibration of Mg/Ca ratios to deep-water temperatures has been described previously [*Skinner and Elderfield*, 2007; *Skinner et al.*, 2003], and relies on Mg/Ca and temperature constraints from modern, last interglacial, and glacial contexts. The resulting calibration essentially reproduces a previous core-top calibration that made use of living *G. affinis* specimens [*Tachikawa et al.*, 2003], and bears a temperature sensitivity that is very similar to almost all available planktonic foraminiferal Mg/Ca calibrations (~9%). We note that a 'carbonate ion effect' is not expected in this foraminifer, in part since it derives from an anoxic infaunal habitat where pore-water carbonate ion concentration is likely to have remained nearly constant at saturation levels due to the buffering capacity of in situ carbonate dissolution [*Martin and Sayles*, 1996]. In any event, carbonate ion effects [*Elderfield et al.*, 2006] in the *G. affinis* Mg/Ca measurements, if present at all, cannot have been significant on the basis that partial dissolution indicators for planktonic foraminifer shells do not negatively correlate with benthic Mg/Ca in these cores (Mg/Ca would be expected to co-vary with planktonic shell weights on the basis that they will track significant changes in pore- or bottom-water carbonate ion concentration). Benthic Mg/Ca and isotope results from cores MD99-2334K and MD01-2444 were previously presented separately by *Skinner and Shackleton* [2004] and *Skinner et al.* [2003] and by *Skinner and Elderfield* [2007], respectively; here we synthesise the results for the last 50 ka and focus in particular on the de-convolution of the benthic (calcite) $\delta^{18}O$ record.

Stable isotope analyses were carried out on a Micromass Multicarb Sample Preparation System attached to a PRISM mass spectrometer (for small samples) or a SIRA mass spectrometer (for large samples). Measurements of $\delta^{18}O$ and $\delta^{13}C$ were determined relative to the Vienna Peedee Belemnite (VPDB) standard, and the analytical precision was better than 0.08‰ for $\delta^{18}O$ and 0.06‰ for $\delta^{13}C$. The reproducibility of Mg/Ca standards by ICP-AES analysis is ~0.7% on the long term, and of replicate sample measurements is ~2.0%.

In order to provide continuous records of deep-water temperature, $\delta^{18}O$ and $\delta^{13}C$ change since ~50 ka BP, results from

cores MD99-2334K [*Skinner et al.*, 2003] and MD01-2444 [*Skinner and Elderfield*, 2007] have been concatenated. This has been achieved by providing each core with its own age-scale (described below) and then splicing them together. Alternatively, and equivalently, the cores could be first correlated stratigraphically using planktonic $\delta^{18}O$ (or magnetic susceptibility, with less precision), and then assigned a chronology. The fact that the two sediment cores derive from different water depths (offset by ~700m), and therefore will have recorded slightly different hydrographic conditions, has been addressed by adjusting values from the shallower core (MD01-2444) by the current offset between deep-water temperature (and hence temperature-corrected benthic $\delta^{18}O$) and $\delta^{13}C$ at each core site (~0.6°C and 0.1‰, respectively). This approach is best justified by the fact that *time-variant* benthic isotope results from each site are linearly correlated with a slope that is indistinguishable from unity at the 95% confidence level, and with a *y*-intercept value of ~0.14‰ (see Figure 2). Reassuringly, this intercept value is equivalent to a temperature offset between the two core sites of ~0.6°C, as assumed above.

CHRONOSTRATIGRAPHY

The results from both cores have been placed on the modified GRIP (ss09sea) ice-core calendar age-scale of *Shackleton et al*. [2004] (hereafter referred to as GRIP-SFCP04) by correlation of Dansgaard–Oeschger temperature fluctuations that are clearly recorded in the $\delta^{18}O$ of Greenland ice and in planktonic $\delta^{18}O$ from the Iberian Margin [*Shackleton et al.*, 2000] (see Figure 1). Age pins have thus been selected at the mid-point of each stadial–interstadial transition, allowing the transferral of ice-core ages to MD99-2334K and MD01-2444 via the correlation of planktonic foraminiferal $\delta^{18}O$ with GRIP $\delta^{18}O_{ice}$. In core MD99-2334K, the agreement of the assigned ice-core ages with calibrated radiocarbon ages (to within 2σ uncertainty limits) provides additional confidence in the deglacial chronostratigraphy [*Skinner et al.*, 2003].

GRIP-SFCP04 ages represent ice-core ages that have been 'calibrated' on the basis of two absolute age assignments, for Greenland Interstadial (GIS) 4 (29 ka BP) and for GIS 17 (59 ka BP), with ages in between these points being set by glaciological accumulation and thinning rate constraints [*Shackleton et al.*, 2004]. Absolute ages from ~25ka to the present are therefore essentially the same as for the previous GRIP ss09sea ice-core age-scale [*Johnsen et al.*, 2001]. Note that in this context absolute ages are of lesser importance, relative to the stratigraphic correlations that are made between North Atlantic and Greenland temperature transitions [*Shackleton et al.*, 2000], and between Greenland and Antarctic temperature change [*Blunier and Brook*, 2001].

Figure 2. Cross correlation of benthic $\delta^{18}O_{cc}$ values measured in cores MD99-2334K (3,146m) and MD01-2444 (2,637m) since ~34 ka BP. Thick gray line indicates 1:1 relationship. Black solid line indicates linear best-fit to data (regression equation at top left of plot; $R^2 = 0.94$); dotted lines indicate 95% confidence limits for linear best fit. The slope of the linear regression is equivalent to unity at the 95% confidence level.

y = -0.14 + 0.99x

MD01-2444 Mean Benthic $\delta^{18}O_{cc}$ (2,637m)

MD99-2334K Mean Benthic $\delta^{18}O_{cc}$ (3,146m)
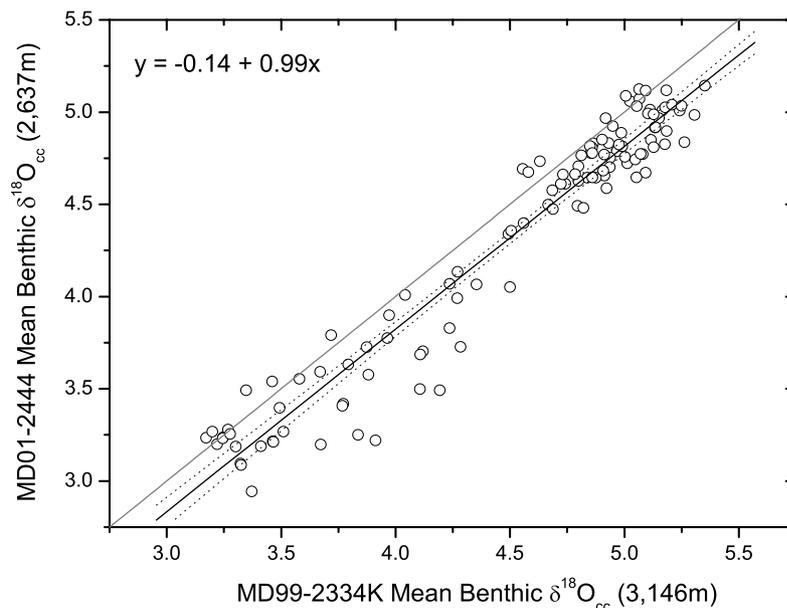
DEEP-WATER TEMPERATURE CHANGE ON THE IBERIAN MARGIN

Figure 3 shows a comparison of Greenland temperature change since ~50 ka BP, with the Mg/Ca-based deep-water temperature record from the Iberian Margin. Two main points emerge from this comparison. The first is that the temperature of the deep Northeast Atlantic appears to have remained loosely coupled with that of Greenland (i.e. the North Atlantic region), with warmer deep-waters generally being exported to the Iberian Margin during the more pronounced interstadial conditions, and with a glacial–interglacial warming trend that began just after the coldest glacial temperatures are recorded over Greenland at ~24 ka BP [*Alley et al.*, 2002]. One plausible interpretation of this loose coupling is that it reflects changes in the contribution of relatively warm North Atlantic sourced deep-water to the Iberian Margin, with a greater representation during interstadial and interglacial times. This 'hydrographic' interpretation has been explored in detail previously [*Skinner and Elderfield*, 2007; *Skinner and Shackleton*, 2004; *Skinner et al.*, 2003], and is supported in principle by a host of independent proxy and modelling studies that also suggest an alternation of northern versus southern deep-water dominance in the deep Atlantic in parallel with millennial and glacial–interglacial climate change [e.g. *Crucifix*, 2005; *Gherardi et al.*, 2005; *Keigwin*, 2004; *Marchitto et al.*, 1998; *McManus et al.*, 2004; *Piotrowski et al.*, 2005; *Piotrowski et al.*, 2004; *Robinson et al.*, 2005; *Sarnthein et al.*, 1994]. An alternative interpretation of course is that the Northeast Atlantic deep-water temperature record

reflects the communication of temperature changes from a single source region in the North Atlantic that was closely coupled with Greenland. Distinguishing between these two interpretations requires a consideration of additional water-mass tracers, such as benthic $\delta^{13}C$ or $\delta^{18}O_{dw}$ (see below).

The second point that should be noted in relation to Figure 3 is that there is by no means a perfect match between local deep-water temperature on the Iberian Margin, and Greenland (and surface-water) temperatures. This stems from two factors, the first being the vagaries of local hydrographic change, which probably cannot be neatly encapsulated by a simple theory of exclusive northern/southern water-mass mixing [*Labeyrie et al.*, 2005; *Skinner and Elderfield*, 2007]. The second factor is that benthic Mg/Ca measurements, even under propitious sedimentary and diagenetic conditions, and performed at relatively high resolution, can remain subject to a significant degree of 'noise'. This noise may be to a large extent 'geological' (i.e. inherent with respect to sedimentary, biological and calcification processes), and underlines the importance of trying to produce similar or consistent deep-water temperature records from proximal core locations in order to learn more about the controls on benthic Mg/Ca.

DEEP-WATER $\delta^{18}O$ CHANGE ON THE IBERIAN MARGIN

One incentive of using benthic Mg/Ca to estimate deep-water temperatures (T_{dw}) is that it allows a 'correction' to be made for the temperature-effects recorded in parallel benthic $\delta^{18}O$ measurements. This is done by using a suitable 'palaeotemperature
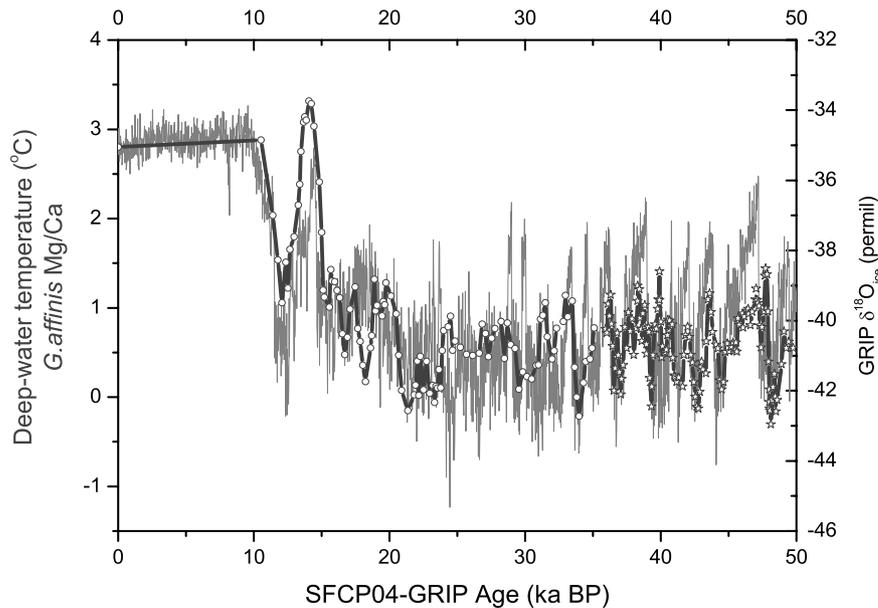
Figure 3. Comparison of deep-water temperature and Greenland temperature variability since ~50 ka BP. Deep-water temperatures are based on benthic Mg/Ca measurements 'spliced' from cores MD99-2334K (solid line with open circles) and MD01-2444 (solid line with stars), on the Iberian Margin. Greenland temperature proxy (thin gray line)is from the GRIP ice-core [*Johnsen et al.*, 2001], placed on the SFCP04-GRIP age-scale (see text).

equation' [e.g. *Bemis et al.*, 1998; *O'Neil et al.*, 1969; *Shackleton*, 1974] in order to isolate the residual $\delta^{18}O$ component that is due to the $\delta^{18}O$ composition of deep-water ($\delta^{18}O_{dw}$). The resulting $\delta^{18}O_{dw}$ value will in turn represent some 'deviation' (which need not be non-zero) from the global mean glacioeustatic $\delta^{18}O_{dw}$ value [*Shackleton*, 1967]. It can be shown that $\delta^{18}O$ values measured in *G. affinis* are linearly correlated with parallel values measured in *P.wuellerstorfii*, with a slope that is statistically indistinguishable from unity, and with an offset of ~0.94‰ [*Shackleton et al.*, 2000]. This observation means that both benthic species have experienced and recorded the same combination of temperature and $\delta^{18}O_{dw}$ conditions, and that both have obeyed the same (equilibrium) 'palaeotemperature equation'. We employ the equation of *O'Neil et al.* [1969], as justified by *Shackleton* [1974]:

$$\delta^{18}O_{dw} = \delta^{18}O_{calcite} + 0.27 - \left(\frac{4.38 - \sqrt{4.38^2 - 0.4(16.9 - T_{dw})}}{0.2} \right)$$

Both T_{dw} and $\delta^{18}O_{dw}$ are conservative tracers, in the sense that their values may only be altered in the ocean interior as a result of mixing processes. Hence coupled T_{dw} and $\delta^{18}O_{dw}$ estimates may provide unique information regarding the evolution of local hydrography, with the caveat that changes in $\delta^{18}O_{dw}$ combine both *local* and *global* components that can be difficult to isolate precisely. This last point is hard to over emphasise, especially in regard to stratigraphic correlations involving benthic $\delta^{18}O$ that are intended to be of

millennial-scale precision or better [*Shackleton*, 2006; *Skinner and Shackleton*, 2005].

Variations in T_{dw} and $\delta^{18}O_{dw}$ calculated in cores MD01-2444 and MD99-2334K are shown in Figure 4 compared with the evolution of Greenland and Antarctic temperature change since ~50 ka BP. What is suggested by this visualisation of the data is that rather similar relationships are observed for T_{dw} and $\delta^{18}O_{dw}$, with respect to Greenland and Antarctic temperature change, as were observed initially by *Shackleton et al*. [2000] for planktonic and benthic $\delta^{18}O$ variations. Note that the relationships observed by *Shackleton et al*. [2000] could initially only be discerned during Marine Isotope Stage (MIS) 3 (see Figure 1). Here it is suggested that the primary cause of the millennial benthic $\delta^{18}O$ variability (i.e. local $\delta^{18}O_{dw}$ change, rather than deep-water temperature per se) in fact persisted across the deglaciation, resulting in $\delta^{18}O$ minima in association with each 'Heinrich stadial' since ~50 ka BP, including Heinrich event 1 and the Younger Dryas. Given what we know about sea-level change since the last glacial maximum, it is clear that the benthic $\delta^{18}O$ record cannot be interpreted in a straightforward manner as a proxy for either deep-water temperature or ice-volume.

As described above, $\delta^{18}O_{dw}$ may vary due to glaciouestatic (ice volume) changes, as well as purely local hydrographic variability. Figure 5 shows a comparison of the reconstructed $\delta^{18}O_{dw}$ record from the Iberian Margin with a record of sea-level change that combines the coral data of *Lambeck et al.*, [2002] with the hydraulic model data of *Siddall et al.*, [2003]. These two sea-level datasets have been reconciled by scaling

them to a maximum glacial $\delta^{18}O_{dw}$ increase of 1.0‰ [*Adkins et al.*, 2002; *Schrag et al.*, 2002]. The coral data (covering the deglacial portion of the record) are shown plotted versus their original radiometric ages, while the data of *Siddall et al.*, [2003] (covering the glacial portion of the record) have been placed on an age-scale that arbitrarily maximises their contribution to Iberian Margin $\delta^{18}O_{dw}$. This approach is conceptually equivalent to that originally adopted by *Siddall et al.* [2003], and posits that: (1) the Red Sea sea-level curve is accurate; and (2) the non-glacioeustatic (hydrographic) residual in the Iberian Margin $\delta^{18}O_{dw}$ signal is minimal. This chronology, as indeed any chronology, represents a set of hypotheses that are entertained precisely for the purpose of considering their implications, and eventually their viability.

While it is clear from the comparison shown in Figure 5 that the bulk of the observed glacial–interglacial $\delta^{18}O_{dw}$ change (~1.2‰) can be attributed to glacioeustacy (equivalent to ~120m sea-level change or ~1.0‰ [*Adkins et al.*, 2002; *Schrag et al.*, 2002]), the same cannot be true of the transient $\delta^{18}O_{dw}$ minima that occurred during the deglaciation, in association with Heinrich event 1 and the Younger Dryas. Instead, these $\delta^{18}O_{dw}$ minima must represent some

deviation from the evolution of 'global average' $\delta^{18}O_{dw}$ due to local changes in deep-water character or mixing, such as the incursion of a deep-water mass analogous to modern Antarctic Bottom Water [*Skinner and Shackleton*, 2004]. It is noteworthy that such hydrographic changes were not restricted to the deep Northeast Atlantic [*Labeyrie et al.*, 2005; *Waelbroeck et al.*, 2006].

During the glacial period (MIS 3) the reasons for millennial $\delta^{18}O_{dw}$ variability are more equivocal, primarily because of the uncertainty of the timing and amplitude of sea-level change during this time interval. Note that, in general, this sort of 'local palaeo-salinity' calculation problem is underdetermined unless one can first claim precise constraints on glacioeustatic effects, and second, justify that local $\delta^{18}O_{dw}$ has responded to glacioeustatic changes approximately as rapidly as they have occurred [e.g. *Duplessy et al.*, 1991]. On the basis of the comparison shown in Figure 5 (i.e. the age-scale hypothesised for the *Siddall et al.* [2003] data, and the hypothesis of a relatively rapid mixing time in the ocean), much of the glacial Iberian Margin $\delta^{18}O_{dw}$ record would indeed be explained by a global glacioeustatic $\delta^{18}O$ signal. This observation essentially represents the assertion that two
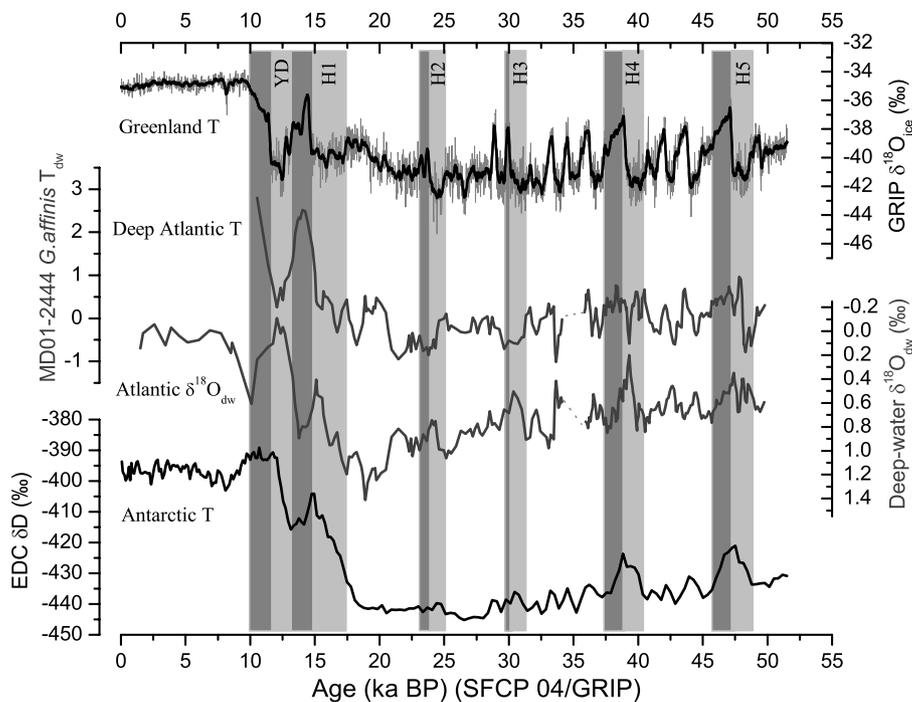

Figure 4. Phasing of millennial changes in the Greenland (top) and Antarctic (bottom) temperature proxy records aligned and compared with the pattern of deep-water temperature (upper middle) and $\delta^{18}O_{dw}$ (lower middle) variability on the Iberian Margin. Deep-water $\delta^{18}O_{dw}$ has been calculated from the benthic $\delta^{18}O$ record shown in Figure 1, using the temperature estimates shown in Figure 3. Vertical light gray bars indicate the timing of 'Heinrich stadials' (H1-5) and the Younger Dryas (YD); vertical dark gray bars indicate their succeeding interstadials.

independent records, each of which is proposed to include or represent a record of glacioeustatic $\delta^{18}O_{dw}$ change, can be correlated with each other reasonably well. A linear regression of these two records (not shown) yields a correlation coefficient R^2 ~0.58, though with a linear slope that is significantly different from 1.

The fact that the cross-correlation of the two records shown in Figure 5 does not yield a linear slope of ~1 indicates that, at most, only a portion of the glacial Iberian Margin $\delta^{18}O_{dw}$ record can be explained by the sea-level curve of *Siddall et al.*, [2003] (in fact sea-level explains only ~40% of the reconstructed $\delta^{18}O_{dw}$ amplitude). Because deep-water temperature does not track benthic (calcite) $\delta^{18}O$ perfectly, the minimum amplitude of $\delta^{18}O_{dw}$ variability that can be obtained (for example, by varying the temperature calibrations that are applied) is approximately equal to the amplitude of original benthic $\delta^{18}O$ signal. A non-glacioeustatic component is therefore required regardless of the amplitude of the reconstructed temperature signal shown in Figure 3. This is most obvious across the deglaciation, which serves as a 'proof of principle', that a residual $\delta^{18}O_{dw}$ component will always be generated when parallel deep-water temperature and benthic $\delta^{18}O$ trends are not exactly equivalent, and furthermore that this $\delta^{18}O_{dw}$ residual need not reflect a 'global average' (glacioeustatic) signal.

In general, a non-glacioeustatic $\delta^{18}O_{dw}$ residual should be attributable to purely local hydrographic change, and should therefore be correlated with independent proxies for deep-water sourcing or mixing, such as $\delta^{13}C$ or deep-water temperature. Figure 6 illustrates a direct comparison of epibenthic $\delta^{13}C$ and $\delta^{18}O_{dw}$ variability on the Iberian Margin, while Figure 7 illustrates a comparison of stadial and interstadial values of deep-water temperature, benthic $\delta^{13}C$ and non-glacioeustatic $\delta^{18}O_{dw}$ (the latter derived by subtracting the two records illustrated in Figure 5). Both of these comparisons reveal a weak yet clearly detectable correlation between indicators of local hydrographic change on the Iberian Margin on millennial time-scales, with maxima/minima in T_{dw}, $\delta^{18}O_{dw}$ and $\delta^{13}C$ tending to coincide as they do today spatially in the deep Atlantic [*Kroopnick*, 1980; *LeGrande and Schmidt*, 2006; *Ostlund et al.*, 1987]. On this basis it might be argued that a portion of the $\delta^{18}O_{dw}$ record is indeed attributable to temporal changes in deep-water character and/or sourcing on the Iberian Margin, analogous to the spatial variability that is observed today in the deep Atlantic.

The above arguments suggest that the Iberian Margin $\delta^{18}O_{dw}$ record may comprise two components, one of which is attributable to ice-volume change, and the other of which (correlated with deep-water temperature and benthic $\delta^{13}C$ variations) is due to local hydrographic variability. This is most clearly demonstrated across the deglaciation, where it is the purely 'hydrographic' component that bears a resemblance to the Antarctic temperature record (see Figure 4). These results support the initial suggestion of *Chappell* [2002] that ice-volume variability can only partially explain the Iberian Margin $\delta^{18}O$ record, though they also contradict the specific assertion that the non-glacioeustatic $\delta^{18}O$ residual can be attributed to temperature effects alone [*Adkins et al.*, 2005; *Chappell*, 2002; *Roche and Paillard*, 2005]. The fact that the deep Northeast Atlantic $\delta^{18}O_{dw}$ record (with its strong 'hydrographic' component) appears to track Antarctic temperature variability suggests a mechanistic link between the two. We briefly explore this possibility below.
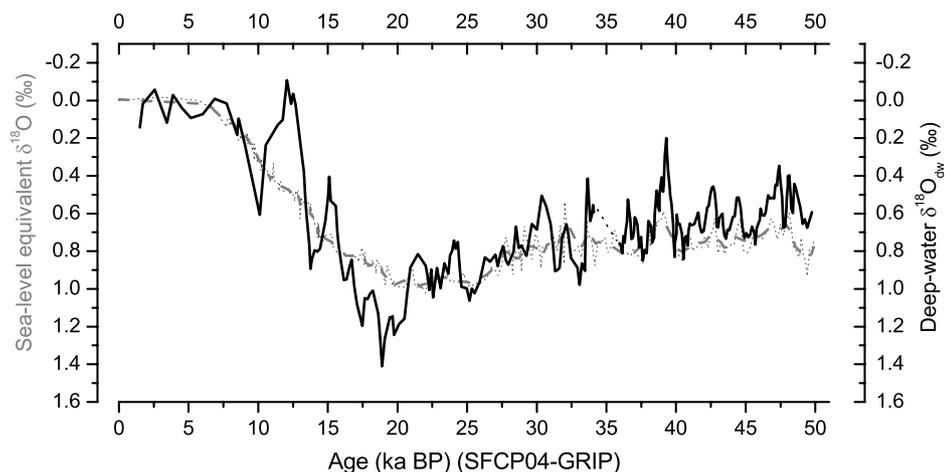

Figure 5. Comparison of reconstructed Iberian Margin deep-water $\delta^{18}O_{dw}$ variability (heavy black line) with a spliced record of sea-level change (gray dashed line) since ~50 ka BP. Records are shown as deviations with respect to modern values. Sea-level change has been converted to equivalent global mean $\delta^{18}O_{dw}$ change, by scaling maximum sea-level drop to 1.0‰ [*Schrag et al.*, 2002].

CONSTRAINING THE MECHANISMS OF INTER-HEMISPHERIC CLIMATE CHANGE?

The leading hypothesis that has been advanced to explain the 'asymmetrical' inter-hemispheric coupling of millennial climate change relies on reversals in the dominant direction of cross-equatorial meridional heat transport in the Atlantic ocean due to perturbations of the overturning circulation [*Knutti et al.*, 2004; *Schmittner et al.*, 2003; *Stocker and Johnsen*, 2003]. According to this hypothesis, southern hemisphere warming would be the direct result of a collapse of the MOC, brought about by 'anomalous' freshwater forcing in the northern North Atlantic. However, this theory is not exclusive and has yet to be explicitly reconciled with additional feedback mechanisms within the climate system. The capacity of relatively small sea-level changes to result in significant ice-sheet feedbacks in the North Atlantic region has recently been underlined by *Fluckiger et al.* [2006]. In addition, numerical modelling experiments performed by *Weaver et al.* [2003] and *Knorr and Lohmann* [2003] have shown in principle that the North Atlantic overturning circulation might also be sensitive to temperature and salinity changes originating in the Southern Ocean. In this context, it seems reasonable to underline at least two distinct and viable mechanisms that may have been involved in inter-hemispheric climate coupling, and which need not be mutually exclusive. The first mechanism would require North Atlantic MOC perturbations that *drive* Antarctic climate change [*Schmittner et al.*, 2003; *Stocker and Johnsen*, 2003], while the second would imply (but not always require) North Atlantic MOC perturbations that *are driven by* Antarctic temperature change. The latter mechanism could provide the basis for both impeding and reinvigorating North Atlantic overturning; via sea-level induced North Atlantic ice-sheet destabilisation in the first instance [cf. *Fluckiger et al.*, 2006], and via buoyancy forcing and/or advective adjustments in the Southern Ocean in the second instance [*Knorr and Lohmann*, 2003; *Weaver et al.*, 2003]. Other hypotheses are also possible [*Wunsch*, 2006], but all must posit a consistent link between Antarctica and Greenland [*Blunier and Brook*, 2001; *EPICA*
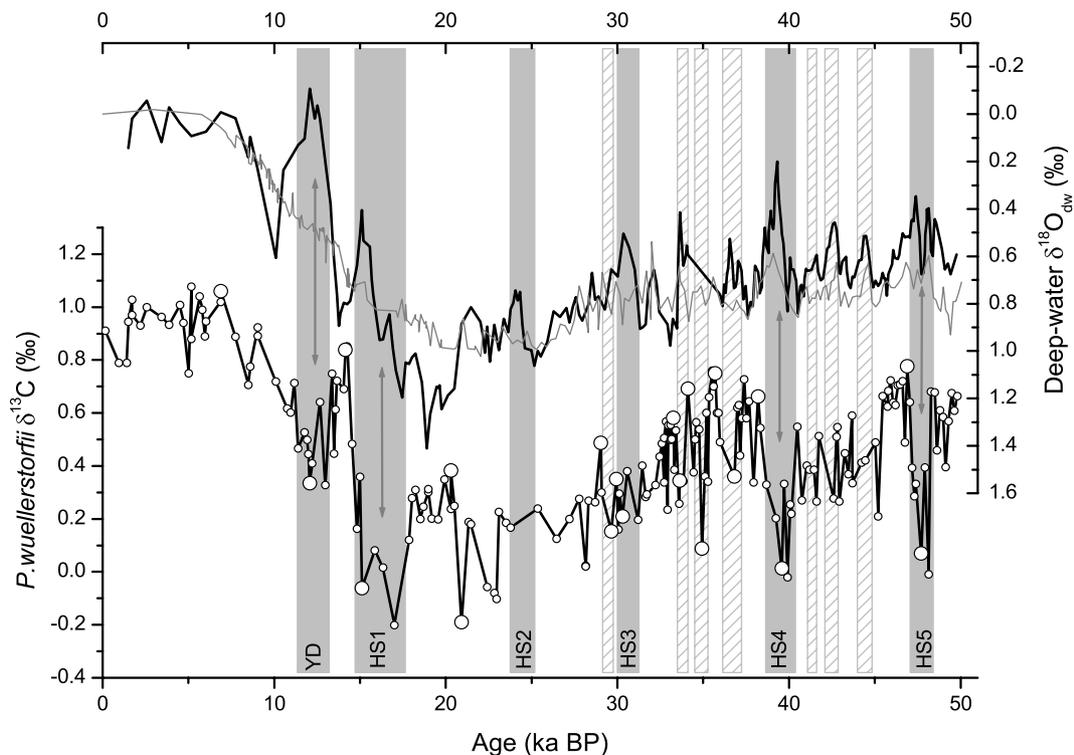

Figure 6. Comparison of deep-water $\delta^{18}O_{dw}$ variability on the Iberian Margin (thick black solid line, upper plot) with 'mean global' $\delta^{18}O_{dw}$ change (thin gray solid line, upper plot) and with local changes in epibenthic $\delta^{13}C$ (black line and open circles, lower plot), used as a proxy for deep-water sourcing/character. Larger open circles in the lower plot (benthic $\delta^{13}C$) indicate stadial and interstadial values for which corresponding deep-water temperature and non-glacioeustatic $\delta^{18}O_{dw}$ values have been interpolated and plotted in Figure 7. Vertical gray bars indicate Greenland stadials (cross-hatched), including 'Heinrich stadials' and the Younger Dryas (filled bars).
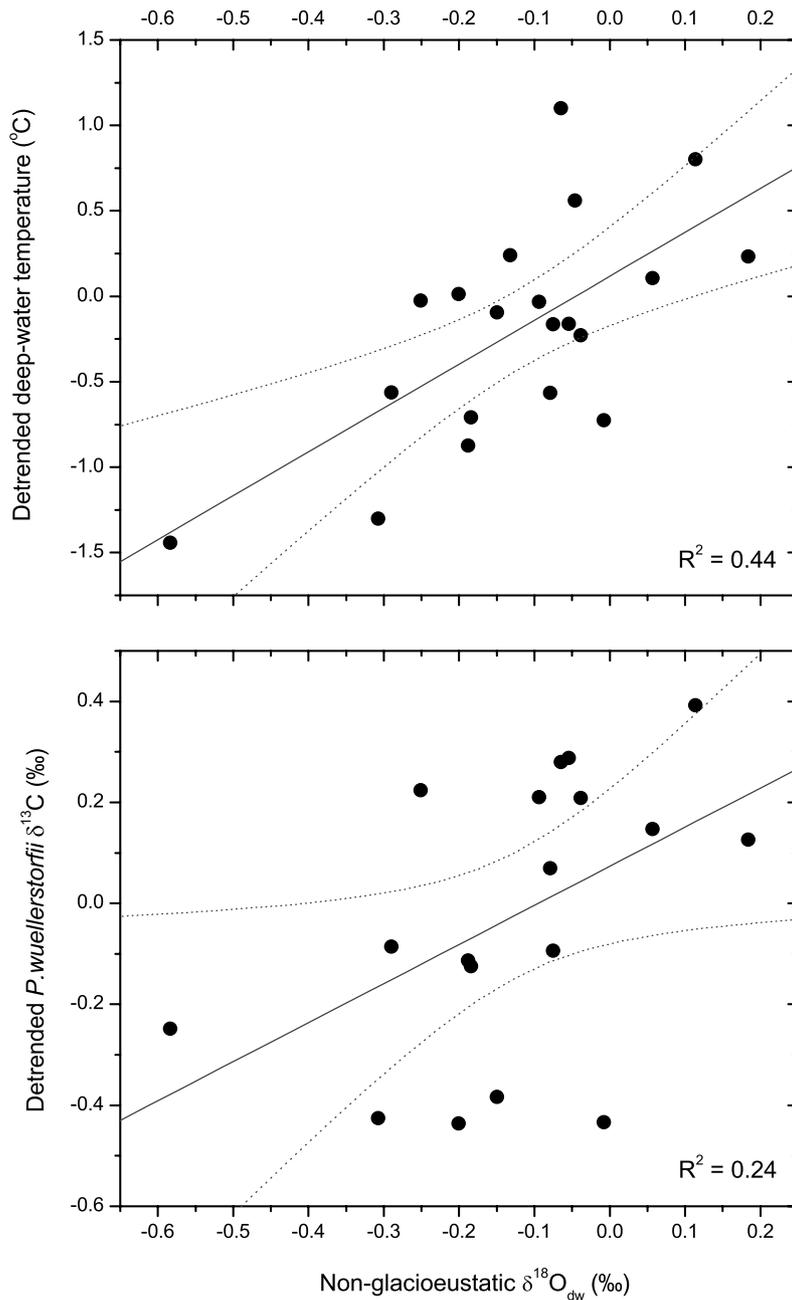
Figure 7. Cross-plots for non-glacioeustatic deep-water $\delta^{18}O_{dw}$ versus parallel deep-water temperature (upper plot) and $\delta^{13}C$ (lower plot) from stadials and interstadials recorded in cores MD99-2334K and MD01-2444, showing a positive correlation between all three indicators of local 'hydrographic change'. Deep-water temperature and $\delta^{18}O_{dw}$ values have been interpolated to coincide with the stadial and interstadial $\delta^{13}C$ values indicated by large open circles in Figure 6. Non-glacioeustatic $\delta^{18}O_{dw}$ values have been derived by calculating the difference between the records plotted in Figure 5. Deep-water temperature and $\delta^{13}C$ values have also been de-trended to remove long-term glacial–interglacial changes, thus permitting a consideration of millennial-scale correlations only. Solid black lines in each plot indicate linear regressions, with confidence limits (dotted lines), and R-squared values.

community members, 2006], as well as implicating some form of hydrographic reorganisation in the deep Atlantic basin across most, if not all, stadial–interstadial events. The results presented here underline the last of these constraints in particular, providing direct palaeoceanographic evidence for a one-to-one link between inter-hemispheric temperature deviations and reversals in the dominant direction of cold deep-water extension in the deep Atlantic, essentially as suggested by *Seidov and Maslin* [2001].

Of crucial importance in determining the relative contributions and timing of eventual 'northern' versus 'southern' drivers in inter-hemispheric climate coupling is the precise phasing of sea-level change [*Clark et al.*, 2002; *Rohling et al.*, 2004], in particular with respect to North Atlantic freshwater forcing, atmospheric CO_2 fluctuations and Antarctic temperature change. On the Iberian margin, the phasing of the 'glacioeustatic' and 'hydrographic' components of the benthic $\delta^{18}O$ record remain to be determined precisely. However, on the basis of the
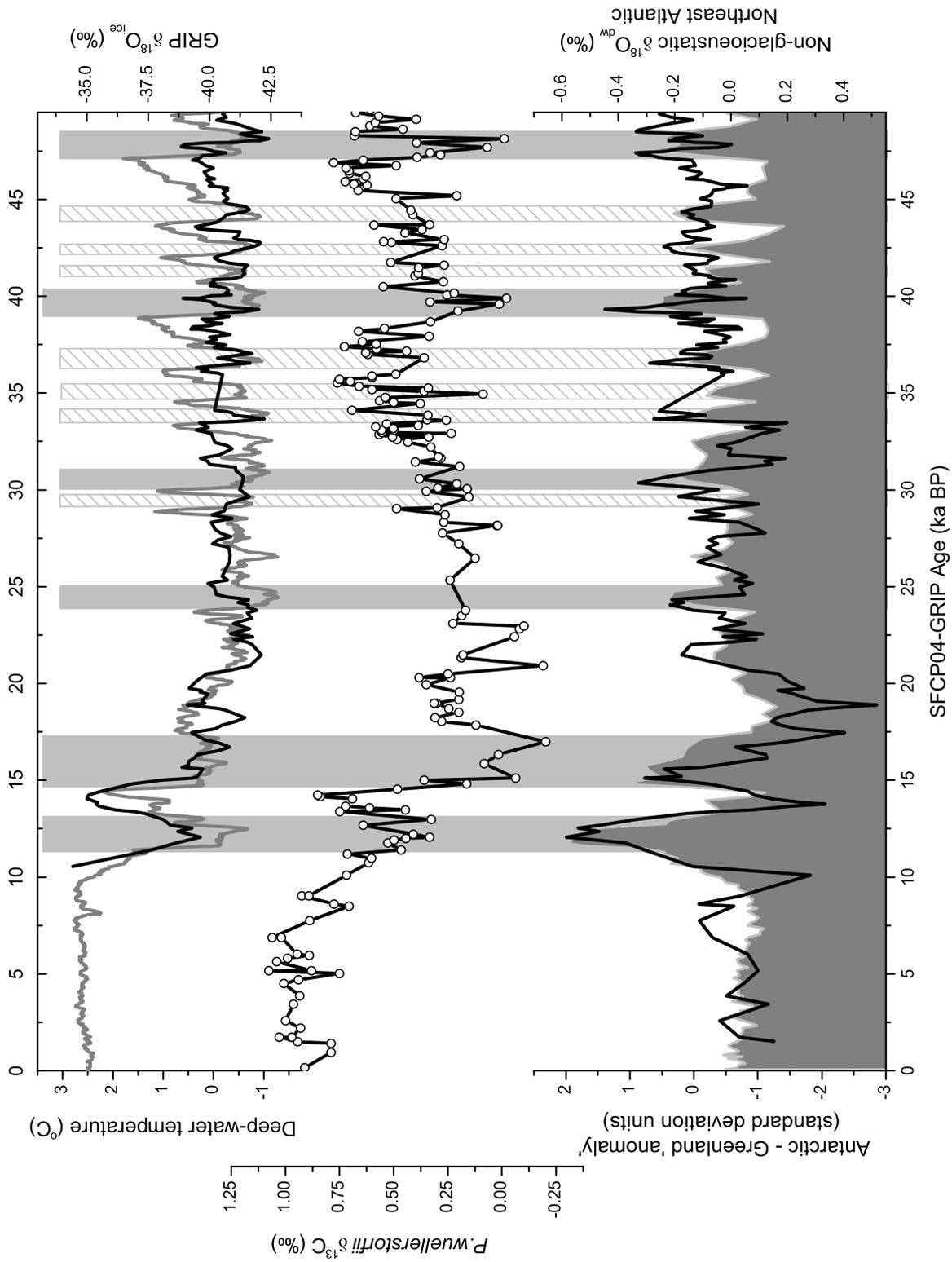
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Figure 8. Comparison of Greenland proxy temperature (top plot; solid black line), Iberian Margin deep-water temperature (top plot; gray solid line), Iberian Margin deep-water temperature (top plot; solid black line), benthic $\delta^{13}C$ (middle plot; solid black line and open circles), as well as the offset between Greenland and Antarctic temperature trends expressed in standard deviation units (bottom plot; solid gray line with shading) versus the proposed non-glacioeustatic component of the Iberian Margin $\delta^{18}O_{dw}$ record, derived by subtracting the two curves plotted in Figure 5 (bottom plot, solid black line). Vertical gray bars indicate Greenland stadials (cross-hatched), including 'Heinrich stadials' and the Younger Dryas (filled bars) as in Figure 6.

hypothesised timing of sea-level change during the last glacia-tion [*Siddall et al.*, 2003] (see Figure 5), it would appear that significant deviations between the trajectories of Greenland and Antarctic temperature change (i.e. Greenland stadials) have been consistently linked to the hydrographic component, and hence deep-water change in the North Atlantic. This is shown in Figure 8, where the divergence between Greenland and Antarctic temperature trajectories (expressed in standard devia-tion units) is compared with the non-glacioeustatic component of the Iberian Margin deep-water $\delta^{18}O_{dw}$ record (as implied by the sea-level curve plotted in Figure 5), as well as local deep-water temperatures and benthic $\delta^{13}C$. Incursions of cold, low-$\delta^{18}O_{dw}$ (less 'evaporated') and low-$\delta^{13}C$ (high nutrient content) deep-water in the Northeast Atlantic are thus shown to coincide with each Greenland stadial, and hence with each significant deviation between Greenland and Antarctic temperature change. Based on the stratigraphy illustrated in Figure 1 and the relationships illustrated in Figure 8 it seems quite clear that the overturning circulation (and more specifically the relative dom-inance of northern versus southern deep-water masses) was indeed implicated in inter-hemispheric climate linkage [*EPICA community members*, 2006; *Seidov and Maslin*, 2001]. Nevertheless it remains impossible to determine to what extent (and at what times) 'north dialled south', and vice versa [*Stocker*, 2003].

CONCLUSIONS

Estimates of deep-water temperature variability in the Northeast Atlantic suggest a loose coupling between the deep Atlantic 'heat budget' and millennial climate change over Greenland during the last ~50 ka. As a result of this associa-tion, it is found that the close phasing of Antarctic warm events and minima in the Northeast Atlantic benthic $\delta^{18}O$ record (first observed by *Shackleton et al.* [2000]) in fact arises due to sim-ilarities in the evolution of Antarctic temperature and local deep-water $\delta^{18}O_{dw}$. The phasing of events noted by *Shackleton et al.* [2000] during MIS 3 is thus found to extend to stadial–interstadial variability of Northeast Atlantic $\delta^{18}O_{dw}$ over the last ~50 ka, including the last deglaciation, and sug-gests a link between Antarctic climate and the extension of southern-sourced deep-water into the North Atlantic.

The Iberian Margin benthic $\delta^{18}O$ record generated by *Shackleton et al.* [2000] has variously been interpreted as reflecting sea-level variability [e.g. *Knutti et al.*, 2004; *Pahnke and Zahn*, 2005; *Siddall et al.*, 2003] or deep-water temperature change [e.g. *Adkins et al.*, 2005; *Chappell*, 2002], with differ-ent implications for the mechanisms that may have been responsible for past abrupt climate change. Based on deep-water temperature estimates we propose that the Iberian Margin benthic $\delta^{18}O$ signal is in fact dominated by local deep-water $\delta^{18}O_{dw}$ variability and that it comprises two main components:

one linked to global sea-level change, and the other linked to Greenland climate (and in particular the deviation of Greenland and Antarctic climate trends) via its association with changes in the Atlantic MOC and incursions of southern-sourced deep-water into the North Atlantic. Although these observations clearly indicate a role for the overturning circulation in inter-hemispheric climate coupling, they fail to unequivocally rule in a unique mechanism, such as the 'bi-polar see-saw' [*Stocker and Johnsen*, 2003]. Completely explaining inter-hemispheric coupling will require better constraints on the timing and ampli-tude of millennial sea-level change, in particular as distin-guished from freshwater delivery to the North Atlantic.

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