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On the timing and mechanism of millennial-scale climate variability during the last glacial cycle

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Abstract The demonstration that natural climate variability during the last glacial cycle shifted rapidly between remarkable extremes has dramatically revised the understanding of climate change. To further advance our understanding, research continues into the timings, geographic distribution, and nature of the millennial-scale climate extremes, and into the mechanisms for intra- and inter-hemispheric transmission of variability through the climate/ocean system. Complementing the traditional definition of the timings of millennial-scale climate variability from ice-core δ^{18}O records, we here further narrow down the temporal constraints by determining statistically significant anomalies in the major ion series of the GISP2 ice core. This exercise offers an objective definition of the timing of climatic anomalies in Northern Hemisphere palaeoclimate proxy records of the last 110,000 years that significantly improves the potential for inter-calibration of 'ice-core tuned' chronostratigraphies. We then present a process-oriented synthesis of proxy records from the Northern Hemisphere. This leads to a conclusion that the Dansgaard-Oeschger (D-O) style fluctuations in these records are (virtually) in phase, since all fall within a clear (atmospheric) pattern of concerted relative dominance shifts between polar/westerly dominated winter-type conditions and tropical/monsoon dominated summer-type conditions. Finally, we speculate on a monsoon-related mechanism that could help explain the anomalously long duration of D-O interstadials 12, 8, and 1, which coincided with cooling trends in Antarctic records.

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1 Introduction

It has become widely established that the climate/ocean system has undergone strong fluctuations on a millennial time scale during the last glacial cycle (Marine Isotope Stages, MIS, 4, 3 and 2, see overviews in Broecker 2000; Sarnthein et al. 2000). Recent studies have convincingly argued that these fluctuations were not synchronous, or of the same nature, in the Northern and Southern Hemispheres (Blunier et al. 1998; Shackleton et al. 2000). However, phase relationships between records of millennial-scale climate variability within the Northern Hemisphere are less easy to assess, since there are no current techniques for developing exact comparable chronologies between the various records (Sarnthein et al. 2000). Using a novel approach to synchronise records with geomagnetic palaeointensity signals, Kiefer et al. (2001) suggested anti-phase behaviour between the N Atlantic and the North Pacific. This view, if corroborated by future research, would upset the amazing signal similarity that seems apparent between the temperature histories of the Greenland summit (ice core δ^{18}O record) and the NE Pacific Santa Barbara Basin (foraminiferal abundance ratio) (Hendy and Kennett 1999, 2000; Sarnthein et al. 2000). The work of Kiefer et al. (2001) offers hope that multidisciplinary analytical studies may eventually elucidate the phase-relationships, but as yet the matter remains elusive. The present work follows a complementary theoretical approach. We first optimise the constraints to both timing and nature of the millennial-scale variability in the North Atlantic sector, and subsequently present a process-oriented assessment of intra- and inter-hemispheric phase relationships.

Correlation between highly resolved records that show extreme similarity in their structures and general timings is commonly done in an 'event stratigraphic' fashion, which compares records based on well-understood proxies (e.g. for temperature), and is used to hone/tune an existing detailed (e.g. radiocarbon-based) chronology. The Greenland ice-core $\delta^{18}O$ series from GRIP and/or GISP2 (Dansgaard et al. 1993; Grootes et al. 1993) are the commonly used 'master records' for such event-stratigraphic correlations in the Atlantic sector and further afield, including Asia and the Indian and Pacific oceans (among many others Behl and Kennett 1996; Rasmussen et al. 1996; Sirocko et al. 1996, 1999; Bond et al. 1997, 1999; Boyle 1997; Heusser and Sirocko 1997; Schultz et al. 1998; Voelker et al. 1998; Allen et al. 1999; Cacho et al. 1999; Fang et al. 1999; Tada et al. 1999; Hendy and Kennett 1999, 2000; Broecker 2000; Sarnthein et al. 2000). However, the ice-core correlation targets may be more tightly constrained than they have been to date, since: (1) the $\delta^{18}O$ series does not represent the highest-resolved signal for the ice cores; and (2) the $\delta^{18}O$ series reflects only one aspect of the climate system (mainly high-latitude temperature), whereas broader information about the intensity and extent of the polar circulation, which may be more directly reflected in other types of records, is available from non-sea-salt, nss, ion series in the same ice cores.

We here exploit the outlined potential to further constrain the ice-core correlation targets for records of millennial-scale climate variability, by complementing the targets in the GISP2 ice-core $\delta^{18}O$ series with the ages of statistically constrained peak-anomalies in the GISP2 nss ion series (Mayewski et al. 1994, 1997; O'Brien et al. 1995; Yang et al. 1997). Rough 'eye-ball fits' of parts of the ion series to other proxy records have been applied in the past (Mayewski et al. 1997; Bond et al. 1997; Grousset et al. 2000), but to date there has been no statistically rigorous comparison of the entire GISP2 ion series to justify such correlations. Our study starts with a statistical assessment of the main anomalies in the GISP2 nss ion series, noting their ages rounded off to the nearest half century. Note that this rounding off of the ages only concerns the *internal* GISP2 chronological precision, and that the confidence limits to the GISP2 time scale (Meese et al. 1997, in press) are applicable where the *external* age control is concerned. By deriving the GISP2 ages for the events, we develop a relative 'template' chronostratigraphic framework for other palaeoclimate proxy records; we realise that these ages in an absolute sense will be subject to any adjustments that future research may deem necessary to the GISP2 time scale. Next, we compare our GISP2 ages for the ion-series anomalies with the ages found for Heinrich events and Dansgaard-Oeschger (D-O) type oscillations in representative North Atlantic palaeoclimate/ocean records. Although a good general agreement is found, we also identify potential pitfalls that are associated with the sheer variety of proxies discussed in the literature, and the fact that sometimes the initial (radiocarbon) age control is too relaxed. Thereafter, the nature of D-O type variability is considered within a wider northern hemispheric context, where we argue in favour of (virtual) in-phase relationships between many records on the basis of a common atmospheric forcing mechanism that can be identified for the bulk of the observed variabilities. Finally, we briefly speculate on the possible importance of the observed out-of-phase variability between the Northern and Southern Hemispheres (Blunier et al. 1998; Shackleton et al. 2000).

2 Heinrich events and the Dansgaard-Oeschger cycle

N Atlantic marine sediment records since the previous interglacial contain several 'Heinrich layers' of ice rafted debris (IRD), spaced at irregular 5–14 kyr intervals (among many others Heinrich 1988; Bond et al. 1992, 1993, 1997, 1999; Grousset et al. 1993, 2000; Andrews et al. 1994; Manighetti et al. 1995; Stoner et al. 1996; Chapman and Shackleton 1998; Elliot et al. 1998; Voelker et al. 1998; Dokken and Janssen 1999; Paterne et al. 1999; Snoeckx et al. 1999), with compositions that suggest a dominantly Laurentide origin (H1, H5), a Fennoscandian and/or eastern Greenland origin (H3, H6), or an internal provenance change between these source regions (H2, H4) (Grousset et al. 1993, 2000; Snoeckx et al. 1999). Heinrich event-equivalent anomalies are known from palaeoclimate proxy records throughout the Northern Hemisphere (among many others Grimm et al. 1993; Porter and An 1995; Rasmussen et al. 1996; Sirocko et al. 1996; Rohling et al. 1998; Schultz et al. 1998; Hodell et al. 1999; Cacho et al. 1999; Paterne et al. 1999; Broecker 2000; Leuschner and Sirocko 2000).

Heinrich layers represent extreme expressions within a fundamental 1–3 kyr cycle of IRD fluxes, which appears independent of longer-term fluctuations in the global glaciation state (Bond et al. 1997, 1999; Elliot et al. 1998; Voelker et al. 1998; Van Kreveld et al. 2000). It closely resembles the Dansgaard-Oeschger (D-O) cycle in the $\delta^{18}O$ records of the GISP2 and GRIP ice cores from the Greenland summit (Dansgaard et al. 1993; Grootes et al. 1993), which shifts between lighter values indicative of climatic cooling (D-O stadials) and heavier values indicative of warming (D-O interstadials). A typical ~1.5 kyr D-O periodicity was deduced from the GISP2 ion series (Fig. 1), which show that concentrations, and transport fluxes to high latitudes, of atmospheric dust and sea salt were much enhanced during D-O stadials, suggesting dry, dusty and windy climatic conditions (Mayewski et al. 1997). The same periodicity appears in a variety of proxies within the highly resolved and intensively AMS^{14}C dated record of Van Kreveld et al. (2000). Strong surface-water cooling throughout the N Atlantic during HEs underlies their accepted correlation with strong D-O stadials in the ice-core $\delta^{18}O$ record, corroborated by comparison of calibrated AMS^{14}C ages of more recent Heinrich layers with the ages of these $\delta^{18}O$ minima (Bond et al. 1992, 1993, 1999; Rasmussen et al. 1996; Chapman and Shackleton 1998; Voelker et al. 1998; Paterne et al. 1999; Cacho et al. 1999; Broecker 2000; Van Kreveld et al. 2000).

3 Analysis of the GISP2 nss ion series

Using statistical significance tests, we identify the combined strongest peaks in the GISP2 continental-source K^+, Ca^{2+}, and Mg^{2+} ion series presented by Mayewski et al. (1994, 1997), O'Brien et al. (1995) and Yang et al. (1997) (Fig. 1; Table 1). The GISP2 ion series represent ion chromatographic data concerning soluble species. Previous contributions (Grootes et al. 1993; Mayewski et al. 1994, 1997; O'Brien et al. 1995; Biscaye et al. 1997; Yang et al. 1997; Meeker and Mayewski 2002), and references therein, have elaborated the time frame and analytical procedures, evaluated the representation of the total dust flux by soluble fraction data, and statistically compared the ion series in relation to atmospheric circulation patterns (see also discussion of methodologies by Kreutz et al. 1997, in that case for Antarctica). Here, each of the three main ion series is analysed individually, but our conclusions on significant peaks concern consistent signals in all three. Findings are compared with, but independent of, the ice-core $\delta^{18}O$ record (Fig. 1B). The raw data (Fig. 1D–F) show clear 'steps' in the series. Before any analysis can be performed the effect of these trends needs to be

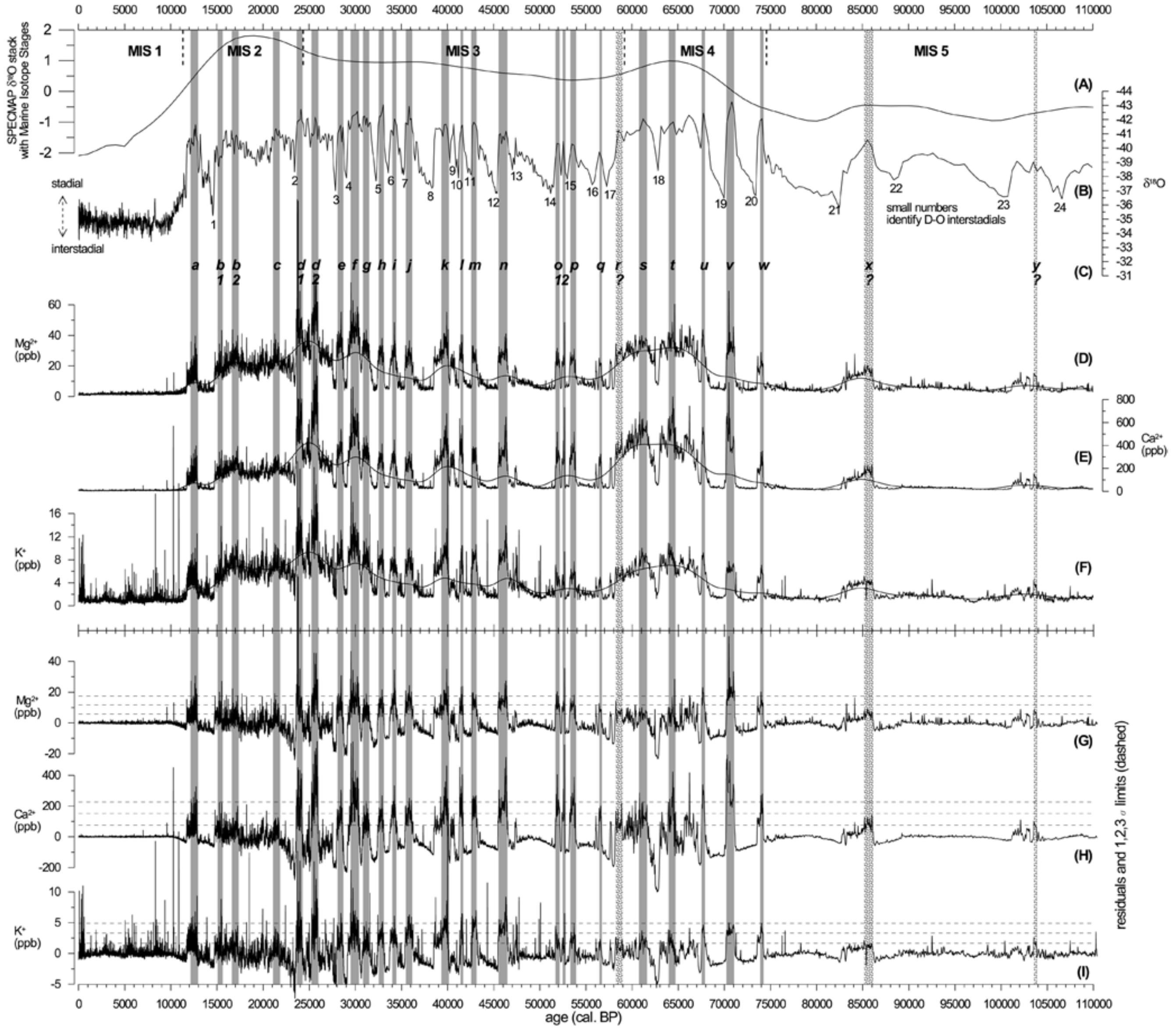

Fig. 1 Identification of intervals where significant ion peaks are observed consistently in the continental-source abundances of the three major GISP2 ion species: **A** SPECMAP stacked marine $\delta^{18}O$ series with Marine Isotope Stages (see Martinson et al. 1987); **B** GISP2 $\delta^{18}O$ series, with *small numbers* indicating the Dansgaard-Oeschger interstadials (Dansgaard et al. 1993; Grootes et al. 1993); **C** labels for the ion-series peaks identified in the present work; **D–F** GISP2 K^+, Ca^{2+}, and Mg^{2+} series. Original data is shown, with spline fit smoothing; **G–I** residuals of the ion series after spline fit detrending. The residuals follow an approximately normal distribution (see also Fig. 2B), allowing simple determination of the 1, 2, and 3σ limits (*dashed lines*). The ion series peaks are consecutively labelled with *bold italics* in **C**, and their intervals, rounded to the nearest half century value, are identified in Table 1. The GISP2 $\delta^{18}O$ record in (**B**) is shown for comparison only, and is *not* used to define the intervals marked as *grey bands*. It is *inversely plotted* for easier comparison with the ion series

removed, and ideally we also want the residuals after trend removal to have a normal distribution. The distribution of the raw data is skew and multimodal (Fig. 2A). Trend removal should remove the multimodality, and to remove skewness we take logarithms of the data (we use natural logarithms but the base is irrelevant). A smoothing spline (Green and Silverman 1994) is applied to remove the changing mean. The smoothing parameter in the spline fit was set to remove the low-frequency signal, and to allow the millennial scale features discussed here to remain (Fig. 1D–F). The histogram of the log data after the trend removal (Fig. 2B) looks normal, and a Q–Q plot shows that up to $\pm 2\,\sigma$ for each ion series the data follow a normal distribution very well. Beyond this both the high and low tails are 'heavy', which implies that the extremes do not come from the underlying distribution. One further method is applied to assess whether the excursions over the 2σ thresholds are significant or due to a random process. Statistics of extremes show that up-crossings of high thresholds even from correlated series form a Poisson process (they occur at random) (Leadbetter et al. 1983), in which case the distances between adjacent up-crossings follow a negative exponential distribution. Hence, we applied a χ^2 test to compare the histogram of distances between up-crossings in the data with that for a negative exponential distribution with the same mean, finding the distribution of up-crossings significantly different from random in all cases. This result did not change when we applied a log transformation to the series to make the distribution more Gaussian, or used the 3σ rather than 2σ limit.

Table 1 Intervals for the significant peaks in the GISP2 ion series presented in Fig. 1. Ages reported to the nearest half century (concerning the internal precision of the GISP2 age model)

Event	GISP2 age-range (ka BP)	Event	GISP2 age-range (ka BP)	Event	GISP2 age-range (ka BP)
a	12.20–12.85	*i*	33.95–34.35	*r*	58.25–58.85
b1	15.25–15.50	*j*	35.40–36.10	*s*	60.80–61.60
b2	16.70–17.25	*k*	39.40–40.05	*t*	63.90–64.40
c	21.15–21.70	*l*	41.30–41.65	*u*	67.55–67.80
d1	23.60–24.20	*m*	42.65–43.10	*v*	70.20–71.10
d2	25.20–25.90	*n*	45.55–46.45	*w*	73.95–74.15
e	28.10–28.55	*o1*	51.75–52.15	*x*	85.20–86.00
f	29.45–30.35	*o2*	52.40–52.70	*y*	103.55–103.80
g	30.90–31.40	*p*	53.40–53.75		
h	32.60–32.95	*q*	56.40–56.60		

Although the three ion series considered display generally coherent behaviour, detailed examination reveals differences in concentration and event sequencing that reflect differences in source area and source strength related to specific atmospheric circulation patterns (Mayewski et al. 1994, 1997; O'Brien et al. 1995; Biscaye et al. 1997; Yang et al. 1997; Meeker and Mayewski 2002). Significant peaks represent periods of high atmospheric dust loading, and are expected to correspond to notable anomalies in other palaeoclimate proxy records. To validate this expectation, the peaks' GISP2 ages are compared with reported ages for IRD events and other anomalies in N Atlantic marine records (Fig. 3).

4 Results and discussion

4.1 Evaluation of the GISP2 ion series-based ages for millennial-scale variability in the North Atlantic

The age-ranges for prominent ion peaks *a*, *d1*, and *d2* agree well with accepted ranges for the Younger Dryas (YD or H0) and H2 (Figs. 1, 3, Table 1), where the predicted two-step nature for H2 (peaks *d1* and *d2*, Fig. 1) matches a distinct bimodality in H2 datings (Fig. 3). Several observations endorse this inferred two-step nature: H2 shows a composite IRD structure in the Bay of Biscay (Grousset et al. 2000) and an internal subdivision in foraminiferal δ^{18}O values in the Nordic Seas (Dokken and Janssen 1999); an IRD peak occurs below the main H2 detrital carbonate layer in the open N Atlantic (Bond et al. 1997); and foraminiferal and alkenone records from the Alboran Sea show two distinct events at H2 time (Cacho et al. 1999). The age range for H1 falls within a broad high in the ion series that is bound by two spikes over the 1–2σ level (Fig. 1G–I), marked *b1* and *b2*. Of these, the latter in particular agrees with datings for H1 (Fig. 3). Comparing our prominent peak *f* with datings for H3, some marked outliers are observed. These likely represent faulty ^{14}C results, since the more comprehensively dated records have been able to overcome/minimise such problems (Fig. 3). Interestingly, Fig. 3 suggests that DC3 in the Labrador Sea may not be equivalent to H3 (*f*), correlating instead with the younger event *e*, which may corroborate geochemical arguments that H3 has a non-Laurentide origin (Grousset et al. 1993; Snoeckx et al. 1999).

More serious ^{14}C dating problems are apparent prior to H3, due to low ^{14}C contents in samples, unknown reservoir ages, and poorly quantified fluctuations in past ^{14}C production. Nevertheless, our predictions for H4, H5, and two intercalated coolings (peaks *k*, *n*, *l*, and *m*, respectively) are strongly supported by the Alboran Sea record (Cacho et al. 1999). That record is exceptional because of excellent age-control, a regular accumulation rate, and arguably limited fluctuations in reservoir age (small-volume well-ventilated basin). This outstanding match, and the support to our H5 prediction from the intensively dated record of Voelker et al. (1998), positively validates our tuning targets for peak D-O stadial conditions.

H6 is beyond radiocarbon dating and is commonly dated from its association with the MIS 4/3 boundary at 59 ka BP (Martinson et al. 1987). Our nearest peak (*r*) is at 58.25–58.85 ka BP, but it is rather weakly developed. Peak *s* at 60.80–61.60 ka BP, or *t* at 63.90–64.40 ka BP, offer more convincing and likely candidates for the timing of H6 (Figs. 1, 3, Table 1). Age discrepancies of this size could result from: (1) mis-interpretation of the MIS4/3 boundary in marine δ^{18}O records due to the H6 δ^{18}O anomaly; (2) the age of the MIS4/3 boundary being ~62 ka BP or ~65 ka BP instead of 59 ka BP, adjustments at or within the ±5 kyr error margin of

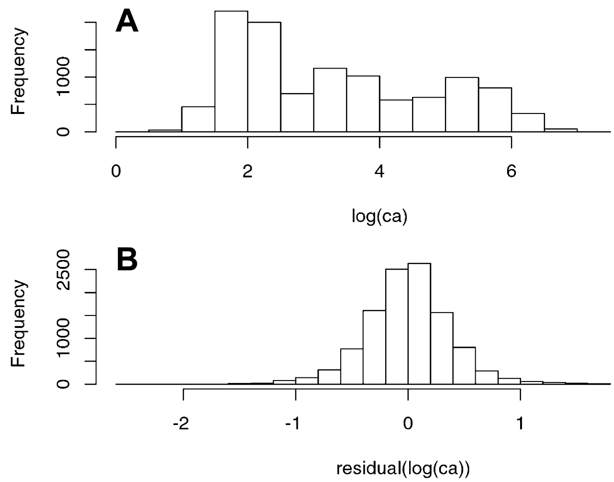

Fig. 2 Histograms of log(Ca^{2+}): **A** data before removing the low frequency trend; and **B** the residuals after trend removal. The residuals follow an approximately normal distribution. Histograms for Mg^{2+} and K^{+} are similar

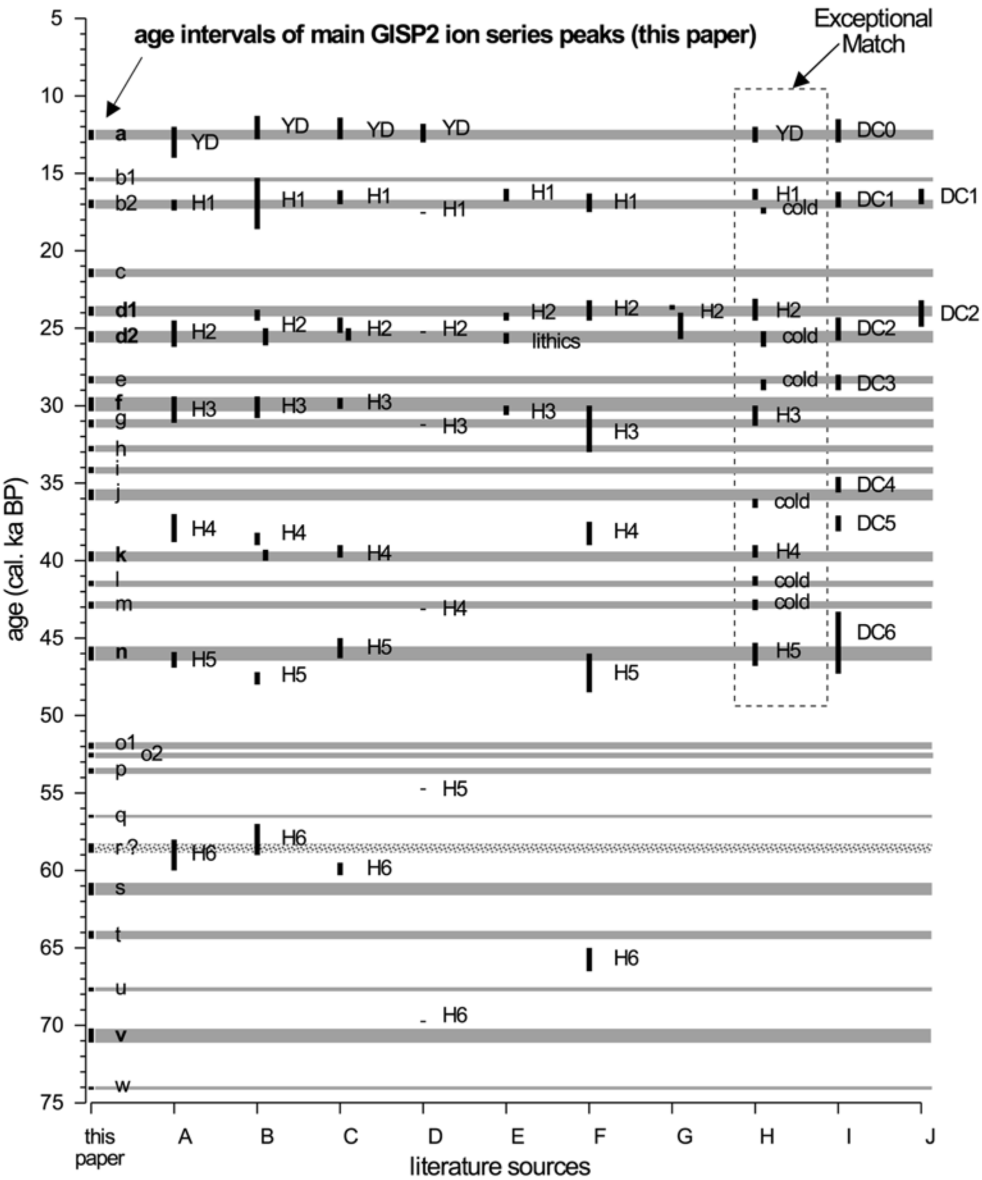

Fig. 3 Age intervals of main ion series peaks determined in this study (Fig. 1; Table 1), compared with literature-based age intervals of Heinrich Events and significant intervening anomalies in N Atlantic cores. Ages that were not reported in calendar years were calibrated using the geomagnetic model prediction (Voelker et al. 1998), which can be applied further back into time than the analysed calibration series. Original age models were followed, i.e., before tuning. Sources are: *A* midway Iceland and Greenland 67.52°N–21.46°W (Voelker et al. 1998); *B* Nordic Seas, 66.41°N–04.34°W (Dokken and Jansen 1999); *C* open N Atlantic, 54.15°N–16.50°W, 49.53°N–24.14°W, and 44.58°N–46.25°W (Bond et al. 1999); *D* open N Atlantic suite of cores between 50.41 to 58.39°N and 16.30 to 22.32°W (Manighetti et al. 1995); *E* open N Atlantic 54.16°N–16.47°W, and 54.15°N–16.50°W (Bond et al. 1997); *F* open N Atlantic 43.03°N–30.03°W (Paterne et al. 1999); *G* Bay of Biscaye ~47°N ~8°W (Grousset et al. 2000); *H* Alboran Sea 36.09°N–02.37°W (Cacho et al. 1999); *I* entrance Labrador Sea 50.12°N–45.41°W (Stoner et al. 1996); *J* Labrador Sea 61.30°N–58.39°W (Andrews et al. 1994)

SPECMAP ages (Martinson et al. 1987); and/or (3) incorrect characterisation of H6. The latter raises the problem of the great variety of criteria used to define HEs: IRD, detrital carbonate, Mg/Ca, magnetic suceptibility, a δ^{18}O anomaly, foraminiferal- or alkenone-derived coolings, etc. Not all yield exactly the same interval. For example, Voelker et al. (1998) use a δ^{18}O anomaly to characterise H6, but show an IRD peak ~20 cm (~2 kyr) below that anomaly. The age of this IRD peak, close to that of peak *s*, is corroborated by an IRD peak at 60–62 ka BP (within MIS4) in the central N Atlantic (Chapman and Shackleton 1998). Evidently, the choice of criteria to define HEs affects the development of an accurate chronostratigraphy. We tentatively suggest that peak *s* seems the most likely target for H6. Note, however, that we predict several potential HE horizons within MIS4 (Fig. 1), implying that reported 'H6' levels are not necessarily time-equivalent to one-another. We especially anticipate that new investigations will demonstrate H-type layers within early MIS4 in the age-range of peak *v* (70.20–71.10 ka BP). The ages of our (weakly defined) events *x* and *y* (MIS 5) agree with rough estimates for HEs 7 and 9 (Chapman and Shackleton 1998) and so may offer potential tuning targets within late MIS 5.

4.2 Phase relationships across the Northern Hemisphere

The close agreements between well-dated millennial-scale events and the ion peaks validate the use of these peaks in conjunction with the ice-core δ^{18}O record to constrain in detail the chronostratigraphies of highly resolved palaeoclimate proxy records, either by 'tuning' following initial radiocarbon age control, or by event-stratigraphic correlation when no/problematic radiocarbon results are available. Of course, this approach assumes knowledge of any phase offset between D-O style cycles in the proxy records and the GISP2 record, which cannot be obtained from current dating techniques. Since in-phase relationships are initially expected only within the high-mid N Atlantic domain, we used only such records to evaluate our predictions (Fig. 3). As discussed, however, many records from the wider Northern Hemisphere (Fig. 4) also show extreme visual similarity to the D-O cycles, with comparable event sequences and structures (Behl and Kennett 1996; Boyle 1997; Schultz et al. 1998; Allen et al. 1999; Tada et al. 1999; Broecker 2000; Hendy and Kennett 2000; Leuschner and Sirocko 2000; Sarnthein et al. 2000). Kiefer et al. (2001) intriguingly suggest that North Pacific signals may be out of phase with the North Atlantic signals, although their correlation of the magnetic palaeointensity records for that scenario is not significantly better than that for the in-phase alternative scenario (the antiphase scenario was preferred because it requires the least variability in sediment accumulation rate for the Pacific core). Until the Kiefer et al. (2001) case for antiphasing can be firmly corroborated, however, we believe that an in-phase relationship between variability in the Pacific and Greenland remains more plausible in view of the extreme signal similarity between the temperature histories of Greenland (GISP2 and GRIP δ^{18}O) and the NE Pacific Santa Barbara Basin (Hendy and Kennett 1999, 2000; Sarnthein et al. 2000). In particular, we emphasise that: (1) the relative durations of the cold and warm phases in both records are distinctly different, and a straight 180° phase-shift would therefore seriously upset the comparability of the overall structures of the two records; (2) longer-term ('sawtooth') trends are highly comparable as well as the millennial-scale trends, and again this agreement would be upset if a 180° phase-shift were assumed; and (3) the record through the last 20,000 years, where radiocarbon age control is less debatable, strongly suggests (virtual) in-phase variability.

In-phase relationships around the Northern Hemisphere would require a rapid signal transmission, which is more likely to be associated with atmospheric forcing mechanisms than with oceanic/thermohaline processes. Reviewing the climatic interpretations offered in the literature for individual highly resolved key records of D-O style variability across the Northern Hemisphere, we note that the records indeed appear to represent a spectrum of local responses in concert with a common, atmospheric, forcing mechanism. The warm events in Santa Barbara coincide with sedimentary expressions of poorly ventilated deep-water conditions (lamination) in the same core. These are reminiscent of the D-O style repetitions of intervals of reduced ventilation in the Sea of Japan described by Tada et al. (1999), who suggested

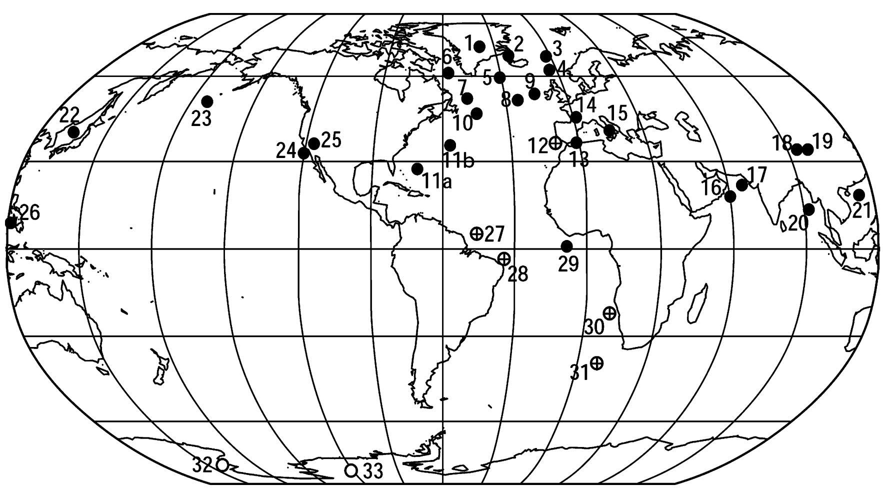

Fig. 4 Locations of key sites displaying the Arctic GISP2/GRIP ice core D-O type cycles (*solid circles*), Antarctic-type millennial-scale cycles (*open circles*), and combinations of the two (*crossed circles*). The Arctic-type signals are defined in the GISP2/GRIP δ^{18}O series of Grootes et al. (1993) Dansgaard et al. (1993). The Antarctic-type signals are defined in D/H and δ^{18}O series of Vostok and Byrd, as summarised in Blunier et al. (1998). Differences in structure, timing, and phase between these two types of millennial-scale variability were discussed in Blunier et al. (1998) and Shackleton et al. (2000) and references therein. Where records show both types of variability, different proxies are involved, with Antarctic-type variability often defined in benthic (bottom-water) δ^{18}O series and Arctic-type variability in surface-water temperature or salinity proxies and/or deep-water ventilation proxies (e.g., δ^{13}C or grey-scale). Numbers refer to: (*1*) Dansgaard et al. (1993), Grootes et al. (1993), also Mayewski et al., (1997); (*2*) Voelker et al. (1998), Van Kreveld et al. (2000); (*3*) Dokken and Jansen (1999); (*4*) Rasmussen et al. (1996); (*5*) Van Kreveld et al. (2000); (*6*) Andrews et al. (1994); (*7*) Stoner et al. (1996); (*8*, *9*, *10*) Bond et al. (1999); (*11a*, *b*) Keigwin and Jones (1994); (*12*) Shackleton et al. (2000); (*13*) Cacho et al. (1999); (*14*) Thouveny et al. (1994); (*15*) Allen et al. (1999); (*16*) Leuschner and Sirocko (2000); (*17*) Schultz et al. (1998); (*18*) Chen et al. (1997); (*19*) Fang et al. (1999); (*20*) Colin et al. (1998); (*21*) Wang et al. (1999); (*22*) Tada et al. (1999); (*23*) Kiefer et al. (2001); (*24*) Behl and Kennett (1996), Hendy et al. (1999, 2000); Heusser (1998); (*25*) Benson et al. (1996), Benson (1999); (*26*) Linsley (1996), De Garidel-Thoron et al. (2001); (*27*) Curry and Oppo (1997); (*28*) Arz et al. (1998); (*29*) McIntyre and Molfino (1996); (*30*) Little et al. (1997); (*31*) Charles et al. (1996); (*32*, *33*) Blunier et al. (1998). For similar compilations, see Leuschner and Sirocko (2000) and Voelker et al. (2002)

a relationship with enhanced summer monsoon conditions. Extending arguments developed in Hendy and Kennett (1999, 2000) and Tada et al. (1999), it is possible to argue for a single causative mechanism that links the two ventilation records. Pacific intermediate water, which ventilates Santa Barbara Basin, mainly originates from the NW Pacific (Sea of Okhotsk), and periods of reduced convective activity in the nearby Sea of Japan are likely to reflect similar conditions in the Sea of Okhotsk. Reduced buoyancy loss during warm/wet events in the NW Pacific (Tada et al. 1999) would reduce the Pacific intermediate water ventilation, and hence would be conducive to the development of anoxic conditions in Santa Barbara Basin. This would be conducive to a (virtual) in-phase relationship between the warm/wet and poorly ventilated intervals in the Sea of Japan, and the warm and poorly ventilated intervals in the Santa Barbara Basin.

Then follows the next question: how do these warm/wet episodes relate to D-O style cycles in loess sections on the Asian mainland and in marine records of the Arabian Sea? Significant insight is offered by another proxy, aeolian dust, measured in the same Sea of Japan cores (Tada et al. 1999). The dust flux was high in cold/dry intervals, and low in warm/wet intervals. Tada et al. (1999) explained this variability in terms of changes in the intensity of the moisture-bringing summer monsoon. They also pointed out, however, that other work related severe loess-flux maxima in China to episodes of intensified winter monsoon circulation that had been considered time-equivalent to the severe D-O stadials when HEs developed in the N Atlantic region (Porter and An 1995). Whether related to weakened summer monsoons, strengthened winter monsoons, or both (Porter 2001), dust-flux maxima in China and over the Sea of Japan are governed by the same system and so will be in phase with one another. Recent Chinese loess studies provided further detail to the relationship between loess fluxes and atmospheric circulation patterns, suggesting that the westerlies jet remained south of the Tibetan Plateau during periods considered equivalent to D-O stadials, causing summer monsoon failure and enhanced winter-type west-northwesterly polar air flow over the Loess Plateau (Fang et al. 2000; Lu and Sun 2000). The inferred weakening of summer monsoon circulation at times of D-O stadials is supported by the climatic interpretation of well-dated records from the Arabian Sea (Schultz et al. 1998; Leuschner and Sirocko 2000). Our process-oriented synthesis therefore suggests with a high level of probability that the records from Santa Barbara Basin, the Sea of Japan, the Loess Plateau, and the Arabian Sea, being intimately related to the same variability that essentially consists of shifts in the relative dominance of winter versus summer monsoon conditions, are (virtually) in phase. The apparent in-phase relationship between D-O interstadials and episodes of intensified boreal summer monsoon circulation agrees with the observation that D-O interstadials in the Greenland ice cores are associated with peaks in the (wetland derived) global methane concentration record (Blunier et al. 1998; Brook et al. 1999).

The methane argument would imply that there is (virtually) no phase lag between the monsoon-regulated records and those from the non-monsoon dominated higher latitudes including the N Atlantic sector. This can be corroborated by comparison of aeolian dust records. Given the dominant Asian provenance in the dust-related GISP2 nss K^+ series (Biscaye et al. 1997; Meeker and Mayewski 2002), and the established correspondence between ice-core ion concentration records and actual atmospheric dust fluxes, it seems plausible that a direct relationship exists between increased loess fluxes and the GISP2 ion peaks during D-O stadials (see also Porter 2001). The GISP2 ion data were previously interpreted in terms of an atmospheric polar circulation index, where peaks and lows correspond to intensified and weakened polar atmospheric circulation (Mayewski et al. 1997). Polar vortex expansion was corroborated by increased polar air outbreaks over the Mediterranean during HEs (severe D-O stadials) (Rohling et al. 1998) and D-O stadials in general (Cacho et al. 1999, 2000, 2001), and would explain coincident cooling and aridity over the N Atlantic and southeastern USA (e.g. Grimm et al. 1993; Chapman and Shackleton 1998; Voelker et al. 1998; Paterne et al. 1999). This mechanism also offers a plausible explanation for the D-O style changes observed in Mediterranean vegetation (Allen et al. 1999), and has even been found to underlie the distinctly in-phase relationship during the Holocene between cold/dusty events in GISP2 and cooling by polar/continental air outbreaks over the NE Mediterranean (Rohling et al. 2002). The inferred alternation between relatively enhanced winter-type conditions (polar vortex expansion/intensification) during D-O stadials, and enhanced summer-type conditions (polar vortex contraction/weakening) for the higher latitudes is in basic agreement with relatively enhanced winter and summer monsoon conditions, respectively (as discussed already).

A compelling case therefore emerges to consider the various records across the Northern Hemisphere to be (virtually) in-phase, based on: (1) the high degree of similarity in event sequences and structures over a very wide spatial domain; and (2) the fact that our process-oriented synthesis highlights a consistent common theme of relative dominance shifts between winter-type and summer-type conditions, ranging all the way across the Northern Hemisphere from polar into monsoonal latitudes. Our arguments corroborate the in-phase relationship between climate variabilities in the high northern latitudes and the tropics suggested in Blunier et al. (1998) and Brook et al. (1999). We cannot assert whether north polar vortex expansion constrained the northward monsoon penetration, or vice versa. However, we note that the apparent pattern of concerted relative dominance shifts between polar/westerly dominated winter-type conditions and tropical/monsoon dominated summer-type conditions represents a primary mode of atmospheric variability, which is well-known to dominate also on shorter (winter versus summer) and longer (glacial versus interglacial) time scales.

4.3 On the Northern–Southern Hemisphere phase relationship

In view of the apparent involvement of shifts in the relative intensities of the boreal monsoons in much of the Northern Hemisphere variability, some speculation may be warranted concerning the observed phase difference between millennial-scale climate variability of the Northern and Southern Hemispheres (Blunier et al. 1998; Shackleton et al. 2000). These studies demonstrated distinct antiphasing for the more prominent (long-duration) Greenland D-O interstadials 12, 8, and 1, with Antarctic temperature changes leading those in Greenland by 1–2.5 kyrs, whereas the shorter-duration events have no significant counterparts in Antarctica (see also Marchal et al. 1999). The crucial methane-based synchronisation of the Antarctic and Greenland records by Blunier et al. (1998) was confirmed by a study of isotopic changes in planktonic (Northern Hemisphere type signal) and benthic (Southern Hemisphere type signal) foraminifera in a single set of samples from a core from the Iberian margin (Shackleton et al. 2000). The latter study also emphasises that there is no strict hemispheric separation; signals evidently can cross the equator (see also Fig. 4), for example via property changes in equator-crossing deep/bottom water masses.

Blunier et al. (1998) argued against an atmospheric linkage between the Southern and Northern Hemisphere variabilities, on the basis of their observation that the methane record synchronously follows the Greenland D-O cycles (interstadials have high methane concentrations; see also Brook et al. 1999), and therefore is out of phase with the Antarctic variability. Following the argument of Blunier et al. (1998), the opposite should be expected if Antarctic temperatures determined tropical temperatures. However, we propose that there may have been an atmospheric link between (some of the) variability in the Antarctic and the tropics that is not directly evident in temperature, but instead operated via a modulation of the intensity and northward penetration of the boreal summer monsoon.

It has been suggested that stronger thermal contrasts on the Southern Hemisphere (colder S Pole) cause intensified trade wind surges from the Southern Hemisphere, enhancing the boreal summer monsoon's intensity and northward penetration (overviews in Pedelaborde 1963; Rossignol-Strick 1985). Analysing the period 1979–1983, Shresta and Murakami (1988) observed that such southerly surges, following a 30–60 day oscillation, indeed form a prominent feature that intensifies convection over the boreal summer monsoon region in the Indian and Pacific sectors. On multi-annual time scales, there appears to be a positive correlation between cold/high phases of the Southern Oscillation Index and enhanced intensities of the Indian summer monsoon (Hastenrath 1991). Concerning multi-decadal time scales, Lamb (1972, p 300) summarised that 40-year averages of the latitudes of the main atmospheric circulation features in the southern Indian Ocean and Australasian sector show displacements of 2–4° latitude in phase with variations in the strength of the east Asian monsoon currents. We here propose that also on longer (centennial to millennial) time scales, cooling in the southern hemisphere may have intensified boreal summer monsoons, causing a 'warm/wet' signature in monsoonal Eurasia with consequent expansion of wetlands (raising natural methane production). If true, then could the monsoon-related increase in interhemispheric latent heat transfer perhaps (partially) explain the anomalously long durations of the D-O interstadials that coincided with marked Antarctic cooling (especially interstadials 12, 8, and 1: Blunier et al. 1998; Marchal et al. 1999; Shackleton et al. 2000)? Assessment of this potentially important atmospheric feedback mechanism calls for the application to 'palaeo-problems' of global coupled Ocean–Atmosphere GCMs that have sufficient spatial and temporal resolution to resolve transient and relatively small-sized features such as the 'trade surges'. This argument complements the more specific requirements to the next generation of palaeoclimate models listed by Pierrehumbert (1999).

5 Concluding remarks

We present further temporal constraints to the cold climatic extremes associated with the Dansgaard-Oeschger oscillation during the last glacial cycle, to complement the traditional temporal constraints from the ice-core $\delta^{18}O$ record. Subsequently, we evaluate phase relationships between the North Atlantic climate cycle and similar-style variability across the globe (Fig. 4), by comparing and contrasting the individual (local) climatic interpretations of highly-resolved climate proxy records to develop a wider hemispheric to global perspective. In the strict sense, none of our interpretations of northern hemispheric/global change in atmospheric circulation is 'new', as the basic concepts (e.g. polar vortex expansion, summer monsoon weakening) were inferred already from the individual records in our literature sources. However, it is the compilation of all these remarkably mutually consistent interpretations that is most convincing in elucidating the overall atmospheric response to the D-O oscillation. We infer that there is a compelling case that atmospheric 'connections' invoked (virtual) in-phase variability across the Northern Hemisphere. As such, our analysis contrasts with the assertion by Kiefer et al. (2001) that there is a near 180° phase-shift between climatic variations in Greenland and the North Pacific.

We conclude that the atmospheric response to the D-O oscillation consisted of a reorganisation of the meridional extent of the main atmospheric circulation features on at least a Northern Hemisphere-wide scale, from polar to tropical/monsoonal latitudes. An equatorward shift appears for D-O stadials, with intensification and expansion of the north polar vortex, enhanced winter monsoon intensities, and with a general expansion/intensification of winter-type circulation features (e.g. northwesterly circulation and polar outbreaks). The enhanced winter monsoon conditions were paralleled by indications of reduced intensities and northward penetration of the summer monsoon system. The opposite tendencies are noted for D-O interstadials. Although individual D-O cycles appear to have different intensities and durations, a mean periodicity appears around ~1500 years (Mayewski et al. 1997; Van Kreveld et al. 2000; Alley et al. 2001). We consider that the anomalously long duration of D-O interstadials that were associated with Antarctic cooling may be (partly) due to enhanced southern trade surges driving stronger boreal summer monsoons and hence an enhanced atmospheric interhemispheric latent heat transfer, although alternative/complementary hypotheses exist as well, focussing more on the role of ocean circulation (e.g. Keeling and Stephens 2001a, b and references therein). To resolve the possible importance of the proposed trade-surge mechanism, a new generation of global coupled ocean–atmosphere GCMs is needed with sufficient spatial and temporal resolution to resolve such transient and relatively small-sized features.

We have not focussed on the nature of the fundamental 'driver' of the ~1500 periodicity. Key insights are that this cycle seems independent from the global glaciation state (Mayewski et al. 1997; Bond et al. 1999), that ^{10}Be and Δ^{14}C records may imply a link with solar variability (Mayewski et al. 1997; Bond et al. 2001) and that global deep water circulation was involved (among many others Bond et al. 1997; Chapman and Shackleton 1998; Dokken and Janssen 1999; Boyle 1997; Bianchi and McCave 1999; Keigwin and Boyle 1999; Sarnthein et al. 2000; Van Kreveld et al. 2000; Alley et al. 2001; Kiefer et al. 2001; Keeling and Stephens 2001a, b). A comprehensive assessment of potential mechanisms driving the ~1500 year variability was presented by Alley et al. (2001), who suggest that stochastic resonance may have enabled any weak (e.g. solar) 'metronome' to pace an internal ocean–atmosphere oscillation.

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