PALEOCEANOGRAPHY, VOL. 23, PA1102, doi:10.1029/2007PA001519, 2008

Click Here for Full Article

Evidence for a widespread climatic anomaly at around 9.2 ka before present

Dominik Fleitmann,[1,2] Manfred Mudelsee,[3] Stephen J. Burns,[4] Raymond S. Bradley,[4] Jan Kramers,[1] and Albert Matter[1]

Received 5 July 2007; revised 14 November 2007; accepted 8 January 2008; published 13 March 2008.

[1] The 8.2 ka event was triggered by a meltwater pulse (MWP) into the North Atlantic and resultant reduction of the thermohaline circulation (THC). This event was preceded by a series of at least 14 MWPs; their impact on early Holocene climate has remained almost unknown. A set of high-quality paleoclimate records from across the Northern Hemisphere shows evidence for a widespread and significant climatic anomaly at ~9.2 ka B.P. This event has climatic anomaly patterns very similar to the 8.2 ka B.P. event, cooling occurred at high latitudes and midlatitudes and drying took place in the northern tropics, and is concurrent with an MWP of considerable volume (~8100 km^3). As the 9.2 ka MWP occurs at a time of enhanced baseline freshwater flow into the North Atlantic, this MWP may have been, despite its relatively small volume, sufficient to weaken THC and to induce the observed climate anomaly pattern.

Citation: Fleitmann, D., M. Mudelsee, S. J. Burns, R. S. Bradley, J. Kramers, and A. Matter (2008), Evidence for a widespread climatic anomaly at around 9.2 ka before present, *Paleoceanography*, *23*, PA1102, doi:10.1029/2007PA001519.

1. Introduction

[2] Approximately 8.47 ± 0.3 ka ago, ~163,000 km^3 of freshwater was released from glacial lakes Agassiz and Ojibway into the North Atlantic [*Barber et al.*, 1999; *Teller and Leverington*, 2004], triggering sudden and widespread cooling in the North Atlantic region [*Alley et al.*, 1997; *Alley and Ágústsdóttir*, 2005]. Temperatures decreased by 1.5° to 3°C in Europe and North America [*von Grafenstein et al.*, 1999; *Hu et al.*, 1999] and, farther afield, the hydrological cycle in the Northern Hemisphere tropics weakened considerably [e.g., *Fleitmann et al.*, 2003; *Dykoski et al.*, 2005] (Figure 1). Marine sediments and climate model simulations suggest that this climatic anomaly termed the "8.2 ka event" was triggered by a slowdown of the thermohaline circulation (THC) by ~40% [*LeGrande et al.*, 2006] in response to a meltwater-induced freshening of the North Atlantic [e.g., *Alley and Ágústsdóttir*, 2005; *Wiersma and Renssen*, 2006; *Ellison et al.*, 2006]. The meltwater pulse (MWP) responsible for the 8.2 ka event is the final one in a series of at least 14 similar events documented for the early Holocene [*Teller and Leverington*, 2004], but the possible climatic impacts of these smaller outbursts are not well documented. On the basis of an ensemble of recently published and revised paleoclimate records we provide evidence for a notable widespread climatic anomaly at around 9.2 ka B.P. (Figure 1). We suggest that this event also resulted from an MWP, but one of much smaller magnitude, only ~5% of that which resulted in the 8.2 ka event (~8100 km^3 or 0.26 sverdrup if released within 1 year; 1 sverdrup = 1 Sv = 1 × 10^6 m^3 s^{-1}). Because the magnitude and climatic anomaly pattern associated with the 9.2 ka event is nearly identical to that associated with the 8.2 ka event, our results suggest that early Holocene climate was much more sensitive to freshwater forcing than previously thought.

[1] Institute of Geological Sciences, University of Bern, Bern, Switzerland.
[2] Formerly at Department of Geosciences, University of Massachusetts, Amherst, Massachusetts, USA.
[3] Climate Risk Analysis, Hannover, Germany.
[4] Department of Geosciences, University of Massachusetts, Amherst, Massachusetts, USA.

0883-8305/08/2007PA001519$12.00

2. Statistical Methods

[3] Detecting an anomaly in measured climate time series is a serious statistical task for two reasons. First, the anomaly ("signal") is a manifestation of an anomalous process (e.g., MWP) that occurred against a background climate process that itself has potential time dependences in the trend and also the variability. Second, the anomalies, which appear as extreme peaks in a record, should not interfere with the estimation of trend and variability; that is, the estimation method has to be robust. Methods to be avoided are, for example, the running mean for trend and the running standard deviation for variability estimation, because these are nonrobust methods and lead to highly inflated values in the presence of extremes [*Lanzante*, 1996]. The statistical method should, furthermore, not only detect anomalies but also quantify their size and the duration over which they occurred.

[4] We used the running median ($2\,k + 1$ window points) as estimator of the time-dependent trend and the running median of absolute distances to the median (MAD) as estimator of the time-dependent variability. Both median and MAD are standard tools in robust statistics [*Tukey*, 1977; *Hampel*, 1985]. The 95% confidence band, which is employed to define the extremes detection threshold, is given by median ±2.96 MAD. (A normal distribution with

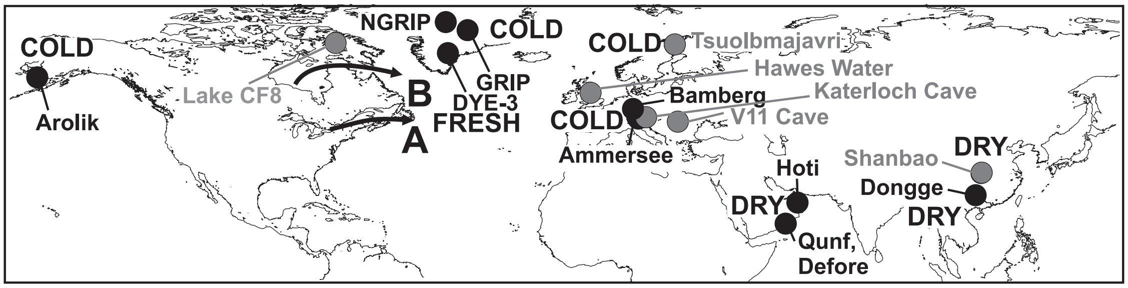

Figure 1. Map showing the location of climate proxy records (black circles) presented in Figure 2. Also shown are proxy records showing evidence for a climatic anomaly at around 9.2 ka B.P. (grey circles). Climatic anomalies associated with the 9.2 ka event are also shown. Black arrows show the routing of Lake Agassiz [*Teller and Leverington*, 2004; *Barber et al.*, 1999] at 9.2 (labeled A) (via St. Lawrence Bay) and 8.2 ka B.P. (labeled B) (via Hudson Bay).

standard deviation σ has MAD = 1.48 σ and a 95% confidence interval of $\pm 2\sigma$.) The duration of an anomaly is then given by the time points the detection threshold is crossed. The size of an anomaly is the maximum of the peak value minus median, divided by the MAD. We selected following values of k for the unevenly spaced time series: NGRIP, k = 25; GRIP, k = 25; DYE-3, k = 25; Arolik Lake, k = 11; Bamberg tree ring, k = 500; Ammersee, k = 33; Hoti Cave (H5), k = 132; Qunf Cave (Q5), k = 69; Defore Cave (S4), k = 218; Dongge Cave, k = 43. This choice leads to average window widths on the order of 600 to 1000 years; that is, it permits us to explore millennial-scale background and variability; see also *Rohling and Pälike* [2005, Table S2], who adopted a similar smoothing value (750 years). Estimations were made using the Fortran 90 program CLIM-X-DETECT [*Mudelsee*, 2006], specifically designed for the purpose of anomaly detection. CLIM-X-DETECT has implemented the efficient calculation of the running median with the updating scheme after *Härdle and Steiger* [1995].

3. Results and Discussion

[5] Detecting short-lived (<10^2 years) climatic events, even if they are of considerable magnitude, is difficult because many paleoclimate records do not have sufficient temporal resolution, chronological precision, or sensitivity to detect decadal-scale climatic anomalies [*Alley and Ágústsdóttir*, 2005; *Rohling and Pälike*, 2005]. We have identified ten paleoclimate records that provide clear evidence for a notable climatic anomaly at ~9.2 ka B.P. (Figures 2a–2j and 3). Perhaps the most compelling evidence comes from three Greenland ice cores, DYE-3 (65°18′N, 37°64′E), GRIP (72°58′N, 37°64′E) and NGRIP (75°10′N, 43.83′E), which reveal a distinct minimum in $\delta^{18}O_{ice}$ at 9.2 ± 0.06 ka B.P. ("present" is defined as 1950 A.D.) on the recently revised GICC05 timescale [*Vinther et al.*, 2006]. With $\delta^{18}O_{ice}$ being a function of air temperature [*Johnsen et al.*, 2001], the observed negative isotopic excursions indicate a short-lived cooling episode. A distinct cold/wet climatic anomaly at ~9.17 ka B.P. is also evident in the biogenic silica record from Arolik Lake (65°18′N, 37°64′E) in the Alaskan Subarctic [*Hu et al.*, 2003], where climate is strongly influenced by the North Atlantic (Figure 2d). In central Europe, an ostracod $\delta^{18}O$ time series from Lake Ammersee (47°59′N, 11°07′E) also shows clear evidence for a distinct cold episode at ~9.18 ka B.P (Figure 2f) [*von Grafenstein et al.*, 1999]. Using an inferred $\delta^{18}O_p$ gradient of 0.58‰/°C [*von Grafenstein et al.*, 1999] for Lake Ammersee, the estimated drop in mean annual air temperature at 9.2 ka B.P. is ~1.6°C in central Europe. Cooling is also evident in an annually precise tree ring width record from Bamberg (49°53′N, 10°53′E), Germany (Figure 2e) [*Spurk et al.*, 2002]. Here, low tree ring widths, indicative of poor growing conditions in summer, are observed at around 9.25 ka B.P. (Figures 2e and 3).

[6] In the Asian monsoon domain a total of four thorium-uranium dated stalagmite $\delta^{18}O_{calcite}$ records show clear evidence for a weak and short-lived (<10^2 years) monsoon anomaly centered at ~9.2 ka B.P. In Oman a positive anomaly in $\delta^{18}O_{calcite}$ centered at 9.22 ± 0.10 ka B.P. is evident in three stalagmites: H5 from Hoti Cave (23°05′N, 57°21′E) [*Neff et al.*, 2001; *Fleitmann et al.*, 2007], Q5 from Qunf Cave (17°10′N, 54°18′E) and S4 from Defore Cave (17°07′N, 54°05′E) (Figures 2h–2j and 3). In China, the well-dated Dongge Cave (25°17′N; 108°50′E) [*Dykoski et al.*, 2005] also shows a positive anomaly in $\delta^{18}O_{calcite}$ at ~9.17 ± 0.08 ka B.P. (Figures 2g and 3). As $\delta^{18}O_{calcite}$ in all these stalagmite records is primarily a function of the amount of monsoon precipitation, with more negative $\delta^{18}O$ values reflecting higher monsoon precipitation and vice versa [*Neff et al.*, 2001; *Fleitmann et al.*, 2003; *Dykoski et al.*, 2005], the 9.2 ka event in the Asian monsoon domain is associated with a notable drop in monsoon precipitation. Overall, there seems to be strong evidence for a hemispheric climatic anomaly at around 9.2 ka B.P. Estimating the duration of the 9.2 ka event is difficult as its end seems to be either gradual or stepwise, but its duration is less than between 200 and 150 years in all records presented here (Figures 2a–2j and 3). The brevity of the 9.2 ka B.P. event precludes its detection in many lower-resolution records; a problem that is also specific to the short-lived 8.2 ka event which has been, even after several years of intensified "anomaly hunting" [*Alley and Ágústsdóttir*, 2005], unam-

biguously identified in only a few paleoclimate records [e.g., *Rohling and Pälike*, 2005; *Wiersma and Renssen*, 2006]. Therefore, it is not surprising that the 9.2 ka event has not yet been detected in more paleoclimate records.

[7] Despite the relatively small number of climate records showing a distinct climatic anomaly at ~9.2 ka B.P., several lines of evidence suggest that the event is indeed a widespread and synchronous climatic perturbation. First, the event is evident in a set of high-quality climate records spread widely across climatic zones. Second, the 9.2 ka event is reproduced within a climatic zone, such as in three ice core records from Greenland or four speleothem records from the Asian monsoon domain. Therefore, we can exclude any local climatic effects. Third, the 9.2 ka event is a significant climatic anomaly which either reaches or exceeds the 95% confidence band in all records presented (Figures 2a–2j). Fourth, within the age uncertainties of each time series the 9.2 ka event seems to be synchronous across the latitudinal transect (Figure 1). In the most precisely dated Greenland ice core and Bamberg tree ring records, the 9.2 ka event is centered at around 9.25 ka B.P., a timing that is in good agreement with thorium-uranium-dated stalagmites from Oman and China which place the event at 9.21 ± 0.08 ka B.P. (mean age of all four stalagmite records presented in Figures 2g–2k). Fifth, climatic anomalies at ~9.25 ka B.P. are identical to those associated with the 8.2 ka event, namely, strong cooling in the North Atlantic, moderate cooling in Europe and a reduction in precipitation in the Indian and Asian monsoon domain (Figures 1 and 2a–2j). Furthermore, the 9.2 ka climatic anomalies are almost identical in magnitude to those associated with the 8.2 ka event (Table 1). Sixth, there is further evidence for the 9.2 ka event in other paleoclimate records shown in Figure 1. Subfossil midge (Chironomidae) assemblages from the eastern Canadian arctic (Lake CF8, Baffin Island) (Figure 1) reveal a distinct cold period, summer temperatures dropped more than 3°C, at ~9.2 ka B.P. [*Axford et*

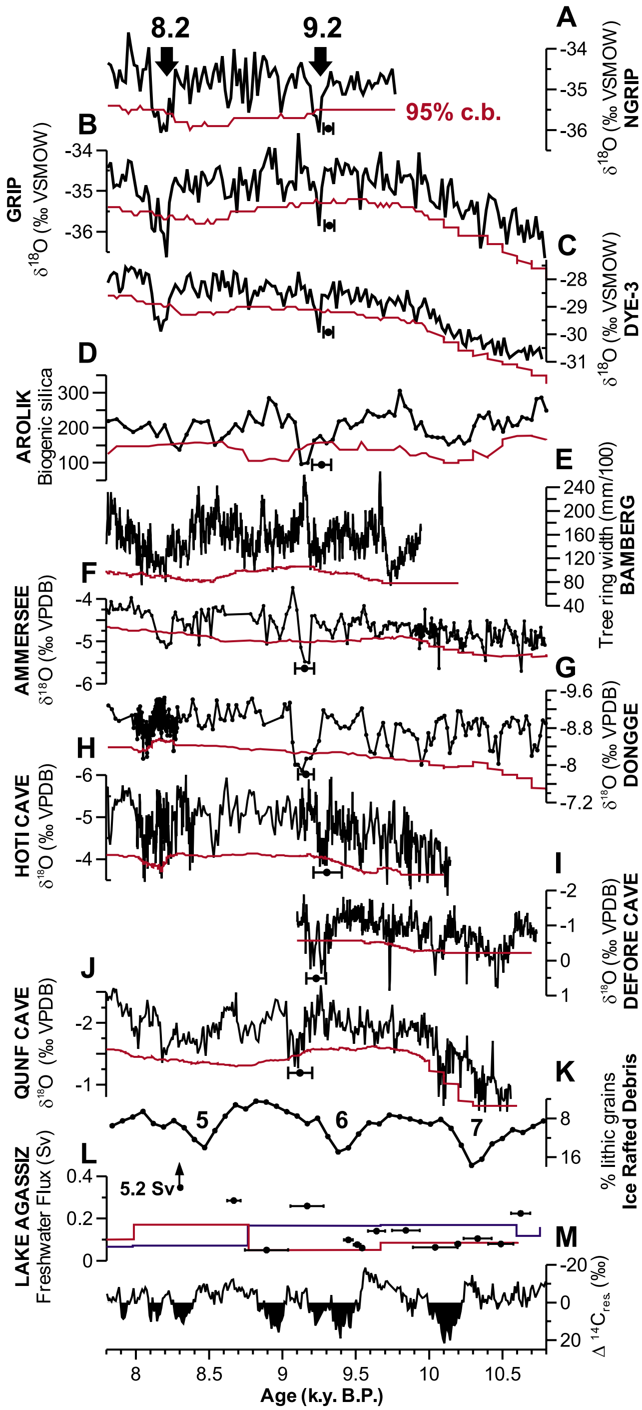

Figure 2. Comparison of early Holocene climate proxy records showing evidence for the 9.2 ka event. (a) The $\delta^{18}O_{ice}$ profiles of NGRIP, (b) GRIP, and (c) DYE-3. Note chronologies of all three ice cores are based on the GICC05 timescale [*Vinther et al.*, 2006]. (d) Arolik lake record from the Alaskan Subarctic [*Hu et al.*, 2003]. (e) Smoothed tree ring width time series (three-point moving average) from Bamberg (Germany) [*Spurk et al.*, 2002]. Thinner tree rings suggest less favorable growth conditions during summer. (f) Ostracod $\delta^{18}O$ record from Lake Ammersee [*von Grafenstein et al.*, 1999]. Lower $\delta^{18}O$ values suggest colder air temperatures. Stalagmite $\delta^{18}O$ profiles from (g) Dongge [*Dykoski et al.*, 2005], (h) Hoti [*Neff et al.*, 2001; *Fleitmann et al.*, 2007], (i) Defore [*Fleitmann et al.*, 2007], and (j) Qunf caves [*Fleitmann et al.*, 2003, 2007]. In all stalagmite-based time series, lower $\delta^{18}O$ values coincide with higher summer monsoon precipitation and vice versa. (k) Stacked North Atlantic marine record of ice-rafted debris (numbers denote so-called "Bond events") [*Bond et al.*, 2001]. (l) Meltwater outbursts in sverdrups (1 Sv = 10^6 m^3 s^{-1}) from Lake Agassiz and Ojibway into the North Atlantic [*Teller and Leverington*, 2004]. Note each outburst has been interpreted as occurring within ~1 year. Solid lines mark baseline flow of freshwater via the St. Lawrence (blue line) and Hudson (red line) (see Figure 1) [*Clark et al.*, 2001]. (m) Detrended atmospheric $\Delta^{14}C_{res}$ [*Stuiver et al.*, 1998]. Positive values indicate higher solar irradiance and vice versa. Thick red lines mark 95% confidence bands (c.b.) which were calculated as described in statistical methods. Dots with error bars show chronological uncertainties of individual records.

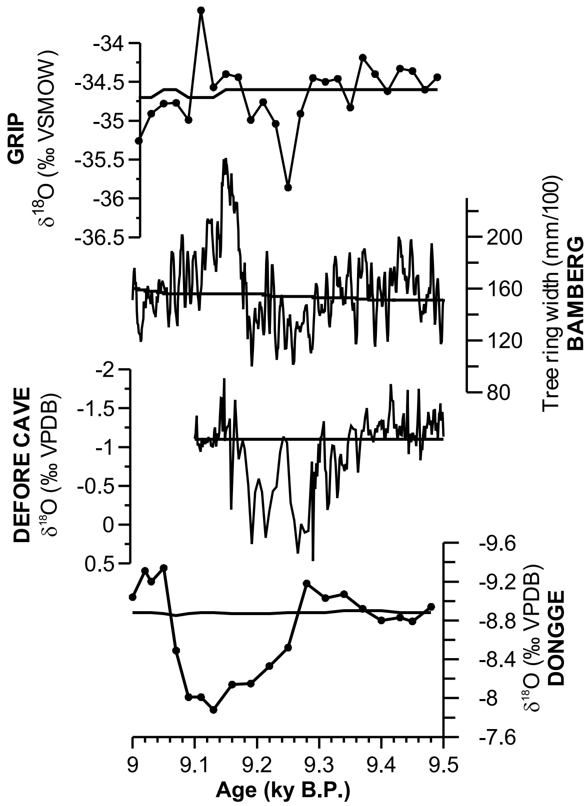

Figure 3. Detailed comparison between GRIP [*Vinther et al.*, 2006], Bamberg tree ring [*Spurk et al.*, 2002], Defore Cave [*Fleitmann et al.*, 2007] and Dongge Cave [*Dykoski et al.*, 2005] stalagmite records. Thick black line without circles marks the median (MAD) as determined with CLIM-X-DETECT [*Mudelsee*, 2006].

al., 2006]. In Finland, chironomid (midges) assemblages in lake sediments document a drop of 0.8–1.5°C in summer temperature at ~9.2 ka B.P. [*Korhola et al.*, 2002]. In NW England decreasing δ^{18}O values of authigenic calcite in a lake sediment core from Hawes Water (Figure 1) reveal a drop in summer temperature at around 9.35 ka B.P. [*Marshall et al.*, 2007], although chronological uncertainties are around ±300. In Austria (Katerloch Cave) and Romania (V11 Cave), stalagmite δ^{18}O profiles show evidence for a sharp cooling episode [*Tamas et al.*, 2005; *Boch et al.*, 2007], and from China, a short-lived dry episode is evident at 9.2 ka B.P. in a δ^{18}O stalagmite monsoon record from Shanbao Cave (31°4′N; 110°26′E) [*Shao et al.*, 2006], in excellent agreement with the Dongge Cave δ^{18}O record farther south (Figure 2g).

[8] Accepting that a widespread climatic event took place at ~9.2 ka B.P., what might have been its origin? The 9.2 ka event does not coincide with a period of strongly reduced solar irradiance in the detrended tree ring ^{14}C time series, which is a commonly accepted proxy for solar output [e.g., *Stuiver et al.*, 1998; *Beer et al.*, 2000]. A distinct minimum in solar irradiance is centered at ~9.4 ka B.P. (Figure 2m), but the well-constrained chronologies of the ice core and the annually precise tree ring width records do not permit a shift in the 9.2 ka event by several decades to match this notable solar minimum in the tree ring ^{14}C record (Figure 2k). Likewise, in all records presented the 9.2 ka event clearly postdates Bond event 6 [*Bond et al.*, 2001] by at least 150 years (Figure 2k), suggesting that they are not associated or perhaps are due to chronological uncertainties of the stacked IRD record (e.g., variable ^{14}C reservoir ages and/or low sedimentation rates). On the basis of these observations, solar forcing of the 9.2 ka event seems to be rather unlikely. Volcanic forcing is also unlikely as none of the ice core sulphate profiles show signs of strong volcanic activity at around 9.2 ka B.P. [e.g., *Zielinski et al.*, 1996]. However, if compared to the record of meltwater outbursts from Lake Agassiz, the 9.2 ka event matches one of the largest early Holocene MWPs at 9.17 ± 0.11 ka B.P. (lake stage "Stonewall") [*Teller and Leverington*, 2004], when estimated ~8100 km^3 or 0.26 Sv (if released within 1 year) were discharged through the St. Lawrence Strait into the North Atlantic (Figure 2l). Although the precise timing and volume of this MWP is still not well constrained, this association suggests that the 9.2 ka event may have been triggered, like the 8.2 ka event [e.g., *Alley and Ágústsdóttir*, 2005; *Ellison et al.*, 2006], by a freshwater-induced reduction in the formation of North Atlantic Deep Water (NADW) and weakening of the THC. One strong argument for this hypothesis is the fact that the spatial climatic anomaly pattern at 9.2 ka B.P. is consistent with that expected following a weakening of THC, namely, cooling in the high latitudes and midlatitudes and drying in parts of the northern tropics [*Alley et al.*, 1997; *Vellinga and Wood*, 2002; *Alley and Ágústsdóttir*, 2005; *Rohling and Pälike*, 2005; *Stouffer et al.*, 2006]. However, the estimated volume of the MWP at 9.17 ± 0.11 ka B.P. is only 5% of that released at 8.47 ± 0.3 ka B.P. (Figure 2h), but it is nevertheless ~90% of the volume injected at the onset of the Younger Dryas [*Teller and Leverington*, 2004]. Consequently, two key questions arise: (1) Is there direct evidence in marine sediment records from the Atlantic for a weakening in THC? (2) Is such a small volume of freshwater injected into the North Atlantic sufficient to perturb THC and to trigger such a widespread climatic event?

[9] Marine sediments from the North Atlantic, the ideal source of information on the mode of the THC, do not provide conclusive evidence for a reduction of the THC at ~9.2 ka B.P. This is in part due to low sampling resolution

Table 1. Comparison of the Climatic Anomalies Associated With the 9.2 and 8.2 ka Events[a]

Proxy Record	Proxy	Anomaly 8.2 ka	Anomaly 9.2 ka	Timing, ka B.P.
NGRIP	δ^{18}O (VSMOW)	−1.40	−1.30	9.25
GRIP	δ^{18}O (VSMOW)	−1.70	−1.30	9.25
DYE-3	δ^{18}O (VSMOW)	−1.80	−1.30	9.25
Arolik	biogenic silica	−65	−113	9.13
Ammersee	δ^{18}O (PDB)	−0.77	−0.95	~9.2
Hoti Cave	δ^{18}O (PDB)	−1.47	−1.00	9.29
Qunf Cave	δ^{18}O (PDB)	−0.75	−0.70	9.11
Dongge Cave	δ^{18}O (PDB)	−0.64	−1.00	9.17
Defore Cave	δ^{18}O (PDB)		−0.73	9.26

[a]Anomalies for all records are calculated maximum deviation from the median as defined by CLIM-X-DETECT [*Mudelsee*, 2006].

(typically >50–100 years or higher), chronological uncertainties due to variable ^{14}C marine reservoir ages, and bioturbation, factors that hinder the detection of such a short-lived anomaly. We note that these shortcomings have also proven an obstacle for detecting the 8.2 ka event in marine sediments from the Atlantic [*Alley and Ágústsdóttir*, 2005]. Nevertheless, there is some evidence for a reduction in NADW formation and weakening of the THC respectively at ~9.2 ka B.P. in at least two marine cores from the Atlantic. A carbon isotope record of the epifaunal benthic foraminifera (*Cibicidoides wuellerstorfi*) in the North Atlantic shows an interval of reduced NADW formation at around 9.3 ka B.P. [*Oppo et al.*, 2003]. Further evidence for a weakening of the THC comes from an aragonite dissolution record (based on the pteropod *Limacina inflata*) from Northern Brazil, where heavier corroded shells at around 9.2 ka B.P. indicate a reduced influence of less corrosive NADW because of a weakening in THC [*Arz et al.*, 2001]. However, chronological uncertainties of both marine sediment records preclude an unambiguous correlation. Thus, current evidence based on marine sediments from the Atlantic neither fully support nor contradict our hypothesis that a weakening of the Atlantic THC triggered the 9.2 ka climatic anomaly.

[10] Regarding the second question, climate models which uniformly suggest that both a higher baseline flow of freshwater (e.g., enhanced river discharge to the Arctic Ocean) or a large MWP can lead to a reduction of the Atlantic THC; particularly if the freshwater is injected close to the relatively small areas of NADW formation (e.g., Labrador Sea) and/or if the mean state of the THC is already close to an instability [e.g., *Wood et al.*, 2003; *LeGrande et al.*, 2006; *Stouffer et al.*, 2006; *Rennermalm et al.*, 2006]. Intercomparison between climate models (ranging from models of intermediate complexity to fully coupled atmosphere-ocean general circulation) show a weakening in THC by ~30% (mean of 14 models) in response to a freshwater input of only 0.1 Sv over a period of 100 years [*Stouffer et al.*, 2006]. Although these experiments were performed under modern climatic conditions, they nevertheless reveal the sensitivity and stochastic response of the THC to small freshwater perturbations. In part because the mean climate during the early Holocene was somewhat different than today and may have made THC more sensitive to freshwater forcing, we suggest that the 9.2 ka B.P. MWP may have been sufficient to impact THC for the following reasons. (1) The injection of freshwater at 9.17 ± 0.11 occurred through the Gulf of St. Lawrence (Figure 1) into the North Atlantic, a routing that injects freshwater close to key areas of NADW formation [*Teller and Leverington*, 2004]. (2) The 9.2 ka event was preceded by a series of MWPs of variable volume (between 0.06 and 0.28 Sv, Figure 2l) which may have preconditioned the THC for the MWP at 9.2 ka B.P. (3) The 9.2 ka MWP is superimposed on enhanced baseline freshwater flow of approximately 0.2 Sv into the North Atlantic because of ongoing melting of the remnant ice sheets (Figure 2l) [*Clark et al.*, 2001]. (4) A weakening of the THC and cooling over the North Atlantic would result in an increase of sea ice [*Stouffer et al.*, 2006]. A greater extent and increased thickness of sea ice would reflect more solar radiation and reduce the ocean-atmosphere heat exchange, and thereby further reduce surface air temperatures over the North Atlantic. Therefore, the relatively small volume 9.2 ka MWP may have been sufficient to invoke a reduction in THC and to lead to a short-lived climatic perturbation in the Northern Hemisphere, despite the fact that it was an order of magnitude smaller than the later MWP at 8.47 ± 0.3 ka B.P. and the resultant 8.2 ka event.

[11] Because the 8.2 and 9.2 ka events have so much in common, a weakening in the strength of Atlantic THC due to an MWP seems to be, based on paleoclimate and model data, the most plausible mechanism. If so, the 9.2 ka event may provide crucial additional insights into the threshold behavior of the THC, which is important in the context of future climate scenarios predicting a freshening of the North Atlantic [e.g., *Wood et al.*, 2003]. However, we must emphasize that more records from other parts of the globe are needed to confirm the occurrence of the 9.2 ka event., and to better constrain its timing and duration. In this spirit we declare, according to *Alley and Ágústsdóttir* [2005], the ‘‘anomaly hunting’’ season for the 9.2 ka event opened.

[12] **Acknowledgments.** This research was supported by the U.S. National Oceanic and Atmospheric Administration (NOAA) and the Swiss National Science Foundation (grants 2021-52472.97 and 2000-059174.99). We are grateful for critical comments by four anonymous reviewers and E. J. Rohling on the manuscript.

References

Alley, R. B., and A. M. Ágústsdóttir (2005), The 8k event: Cause and consequences of a major Holocene abrupt climate change, *Quat. Sci. Rev.*, *24*, 1123–1149.

Alley, R. B., P. A. Mayewski, T. Sowers, M. Stuiver, K. C. Taylor, and P. U. Clark (1997), Holocene climatic instability: A prominent, widespread event 8200 yr ago, *Geology*, *25*, 483–486.

Arz, H. W., S. Gerhardt, J. Pätzold, and U. Rohl (2001), Millennial-scale changes of surface- and deep-water flow in the western tropical Atlantic linked to Northern Hemisphere high-latitude climate during the Holocene, *Geology*, *29*, 239–242.

Axford, Y., J. P. Briher, G. H. Miller, and D. R. Francis (2006), Just the 8.2 event: Dynamic early Holocene climate in arctic Canada, *Eos Trans. AGU*, *87*(52), Fall Meet. Suppl., Abstract PP41C-03.

Barber, D. C., et al. (1999), Forcing of the cold event of 8,200 years ago by catastrophic drainage of Laurentide lakes, *Nature*, *400*, 344–348.

Beer, J., W. Mende, and R. Stellmacher (2000), The role of the Sun in climate forcing, *Quat. Sci. Rev.*, *19*, 403–415.

Boch, R., C. Spötl, and J. Kramers (2007), Early Holocene climate events recorded in fast growing stalagmites from the SE-fringe of the Alps (Austria), *Geophys. Res. Abstr.*, *9*, Abstract 09777, SRef-ID:1607-7962/gra/EGU2007-A-09777.

Bond, G., B. Kromer, J. Beer, R. Muscheler, M. N. Evans, W. Showers, S. Hoffmann, R. Lotti-Bond, I. Hajdas, and G. Bonani (2001), Persistent solar influence on north Atlantic climate during the Holocene, *Science*, *294, 2130–2136.*

Clark, P. U., S. J. Marshall, G. K. C. Clarke, S. W. Hostetler, J. M. Licciardi, and J. T. Teller (2001), Freshwater forcing of abrupt climate change during the last glaciation, *Science*, *293*, 283–287.

Dykoski, C. A., R. L. Edwards, H. Cheng, D. X. Yuan, Y. J. Cai, M. L. Zhang, Y. S. Lin, J. M.

Qing, Z. S. An, and J. Revenaugh (2005), A high-resolution, absolute-dated Holocene and deglacial Asian monsoon record from Dongge Cave, China, *Earth Planet. Sci. Lett.*, *233*, 71–86.

Ellison, C. R. W., M. R. Chapman, and I. R. Hall (2006), Surface and deep ocean interactions during the cold climate event 8200 years ago, *Science*, *312*, 1929–1932.

Fleitmann, D., S. J. Burns, M. Mudelsee, U. Neff, J. Kramers, A. Mangini, and A. Matter (2003), Holocene forcing of the Indian monsoon recorded in a stalagmite from southern Oman, *Science*, *300*, 1737–1739.

Fleitmann, D., et al. (2007), Holocene ITCZ and Indian monsoon dynamics recorded in stalagmites from Oman and Yemen (Socotra), *Quat. Sci. Rev.*, *26*, 170–188.

Hampel, F. R. (1985), The breakdown points of the mean combined with some rejection rules, *Technometrics*, *27*, 95–107.

Härdle, W., and W. Steiger (1995), Optimal median smoothing, *Appl. Stat.*, *44*, 258–264.

Hu, F. S., D. Slawinski, H. E. Wright, E. Ito, R. G. Johnson, K. R. Kelts, R. F. McEwan, and A. Boedigheimer (1999), Abrupt changes in North American climate during early Holocene times, *Nature*, *400*, 437–440.

Hu, F. S., D. Kaufman, S. Yoneji, D. Nelson, A. Shemesh, Y. Huang, J. Tian, G. Bond, B. Clegg, and T. Brown (2003), Cyclic variation and solar forcing of Holocene climate in the Alaskan Subarctic, *Science*, *301*, 1890–1893.

Johnsen, S. J., D. Dahl-Jensen, N. Gundestrup, J. P. Steffensen, H. B. Clausen, H. Miller, V. Masson-Delmotte, A. E. Sveinbjornsdottir, and J. White (2001), Oxygen isotope and palaeotemperature records from six Greenland ice-core stations: Camp Century, Dye-3, GRIP, GISP2, Renland and NorthGRIP, *J. Quat. Sci.*, *16*, 299–307.

Korhola, A., K. Vasko, H. T. T. Toivonen, and H. Olander (2002), Holocene temperature changes in northern Fennoscandia reconstructed from chironomids using Bayesian modelling, *Quat. Sci. Rev.*, *21*, 1841–1860.

Lanzante, J. R. (1996), Resistant, robust and non-parametric techniques for the analysis of climate data: Theory and examples, including applications to historical radiosonde station data, *Int. J. Climatol.*, *16*, 1197–1226.

LeGrande, A. N., G. A. Schmidt, D. T. Shindell, C. V. Field, R. L. Miller, D. M. Koch, G. Faluvegi, and G. Hoffmann (2006), Consistent simulations of multiple proxy responses to an abrupt climate change event, *Proc. Natl. Acad. Sci. U. S. A.*, *103*, 837–842.

Marshall, J. D., et al. (2007), Terrestrial impact of abrupt changes in the North Atlantic thermohaline circulation: Early Holocene, UK, *Geology*, *35*, 639–642.

Mudelsee, M. (2006), CLIM-X-DETECT: A Fortran 90 program for robust detection of extremes against a time-dependent climate records, *Comput. Geosci.*, *32*, 141–144.

Neff, U., S. J. Burns, A. Mangini, M. Mudelsee, D. Fleitmann, and A. Matter (2001), Strong coherence between solar variability and the monsoon in Oman between 9 and 6 kyr ago, *Nature*, *411*, 290–293.

Oppo, D. W., J. E. McManus, and J. L. Cullen (2003), Palaeo-oceanography: Deepwater variability in the Holocene epoch, *Nature*, *422*, 277–278.

Rennermalm, A. K., E. F. Wood, S. J. Déry, and A. J. Weaver (2006), Sensitivity of the thermohaline circulation to Arctic Ocean runoff, *Geophys. Res. Lett.*, *33*, L12703, doi:10.1029/2006GL026124.

Rohling, E. J., and H. Pälike (2005), Centennial-scale climate cooling with a sudden cold event around 8,200 years ago, *Nature*, *434*, 975–979.

Shao, X. H., Y. J. Wang, H. Cheng, X. G. Kong, J. Y. Wu, and R. L. Edwards (2006), Long-term trend and abrupt events of the Holocene Asian monsoon inferred from a stalagmite $\delta^{18}O$ record from Shennongjia in central China, *Chin. Sci. Bull.*, *51*, 221–228.

Spurk, M., H. H. Leuschner, M. G. L. Baillie, K. R. Briffa, and M. Friedrich (2002), Depositional frequency of German subfossil oaks: Climatically and non-climatically induced fluctuations in the Holocene, *Holocene*, *12*, 707–715.

Stouffer, R. J., et al. (2006), Investigating the causes of the response of the thermohaline circulation to past and future climate changes, *J. Clim.*, *19*, 1365–1387.

Stuiver, M., P. J. Reimer, and T. F. Braziunas (1998), High-precision radiocarbon age calibration for terrestrial and marine samples, *Radiocarbon*, *40*, 1127–1151.

Tamas, T., B. P. Onac, and A. V. Bojar (2005), Late glacial-middle Holocene stable isotope records in two coeval stalagmites from the Bihor Mountains, NW Romania, *Geol. Q.*, *49*, 185–194.

Teller, J. T., and D. W. Leverington (2004), Glacial Lake Agassiz: A 5000 yr history of change and its relationship to the $\delta^{18}O$ record of Greenland, *Geol. Soc. Am. Bull.*, *116*, 729–742.

Tukey, J. W. (1977), *Exploratory Data Analysis*, 688 pp., Addison-Wesley, Reading, Mass.

Vellinga, M., and R. A. Wood (2002), Global climatic impacts of a collapse of the Atlantic thermohaline circulation, *Clim. Change*, *54*, 251–267.

Vinther, B. M., et al. (2006), A synchronized dating of three Greenland ice cores throughout the Holocene, *J. Geophys. Res.*, *111*, D06102, doi:10.1029/2005JD006079.

von Grafenstein, U., H. Erlenkeuser, A. Brauer, J. Jouzel, and S. J. Johnsen (1999), A mid-European decadal isotope-climate record from 15.500 to 5000 years B.P., *Science*, *284*, 1654–1657.

Wiersma, A. P., and H. Renssen (2006), Model-data comparison for the 8.2 ka BP event: Confirmation of a forcing mechanism by catastrophic drainage of Laurentide lakes, *Quat. Sci. Rev.*, *25*, 63–88.

Wood, R. A., M. Vellinga, and R. Thorpe (2003), Global warming and thermohaline circulation stability, *Philos. Trans. R. Soc. London, Ser. A*, *361*(1810), 1961–1974.

Zielinski, G. A., P. A. Mayewski, L. D. Meeker, S. Whitlow, and M. S. Twickler (1996), A 110,000-yr record of explosive volcanism from the GISP2 (Greenland) ice core, *Quat. Res.*, *45*(2), 109–118.

R. S. Bradley and S. J. Burns, Department of Geosciences, Morrill Science Center, University of Massachusetts, Amherst, MA 01003-9297, USA.

D. Fleitmann, J. Kramers, and A. Matter, Institute of Geological Sciences, University of Bern, Bern CH-3012, Switzerland.

M. Mudelsee, Climate Risk Analysis, D-30167 Hannover, Germany.