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Proxy benchmarks for intercomparison of 8.2 ka simulations

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Abstract. The Paleoclimate Modelling Intercomparison Project (PMIP3) now includes the 8.2 ka event as a test of model sensitivity to North Atlantic freshwater forcing. To provide benchmarks for intercomparison, we compiled and analyzed high-resolution records spanning this event. Two previously-described anomaly patterns that emerge are cooling around the North Atlantic and drier conditions in the Northern Hemisphere tropics. Newer to this compilation are more robustly-defined wetter conditions in the Southern Hemisphere tropics and regionally-limited warming in the Southern Hemisphere. Most anomalies around the globe lasted on the order of 100 to 150 yr. More quantitative reconstructions are now available and indicate cooling of $\sim 1\,^{\circ}$C and a $\sim 20\,\%$ decrease in precipitation in parts of Europe as well as spatial gradients in δ^{18}O from the high to low latitudes. Unresolved questions remain about the seasonality of the climate response to freshwater forcing and the extent to which the bipolar seesaw operated in the early Holocene.

1 Introduction

The 8.2 ka event is likely one of the best examples from the past of the climate system's response to North Atlantic freshwater forcing. Several lines of evidence support the hypothesis that the drainage of proglacial Lake Agassiz into the Hudson Bay at about 8.2 calendar kiloyears before present (calendar ka BP) slowed the Atlantic Meridional Overturning Circulation (AMOC) and caused the climate anomalies observed in a wide variety of proxy records. This evidence includes the stratigraphic record of lake drainage (Barber et al., 1999), reconstructions of sea level rise (Li et al., 2012; Tornqvist and Hijma, 2012), geochemical reconstructions from the Hudson Strait and northwest Labrador Sea of freshwater discharge (Carlson et al., 2009; Hoffman et al., 2012), proxy indicators of AMOC weakening (Ellison et al., 2006; Kleiven et al., 2008), and climate model experiments testing the linkage between freshwater forcing and climate change (LeGrande et al., 2006; Wiersma and Renssen, 2006). There are some remaining uncertainties about the forcing of the 8.2 ka event, including the the possibility of multiple freshwater releases (Gregoire et al., 2012; Teller et al., 2002; Tornqvist and Hijma, 2012), the pathway of freshwater once it reached the North Atlantic (Condron and Winsor, 2011), and the contribution of other climate forcings around that time (Rohling and Pälike, 2005). Yet, the 8.2 ka event is unique among past meltwater events in that the hypothesized forcing has been quantified and the duration is short enough to make model simulations of the event very feasible. Furthermore, the early Holocene background climate state was not too dissimilar from the present, with two main differences: the increased (decreased) seasonality of insolation in the Northern (Southern) Hemisphere due to orbital forcing and the presence of a remnant Laurentide Ice Sheet both before and after the 8.2 ka event (Carlson et al., 2008; Renssen et al., 2009). For these reasons, the 8.2 ka event was selected for a model intercomparison for the third phase of the Paleoclimate Modelling Intercomparison Project (PMIP3; Morrill et al., 2012).

Paleoclimate proxy data are essential as a benchmark for the model intercomparison. The last global syntheses of proxy data around 8.2 ka were published in 2005–2006 and came to several common conclusions (Alley and Ágústsdóttir, 2005; Wiersma and Renssen, 2006; Morrill and Jacobsen, 2005; Rohling and Pälike, 2005). The most robust

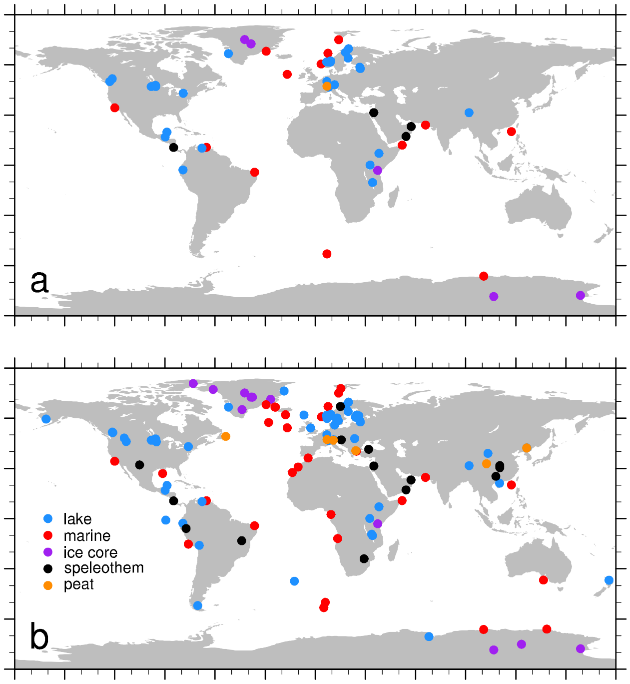

Fig. 1. Location of high-resolution proxy records spanning 8.2 ka that were available in **(a)** 2005 (Morrill and Jacobsen, 2005) and **(b)** 2012.

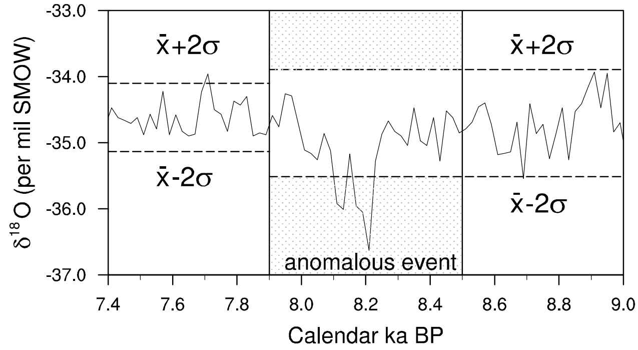

Fig. 2. Diagram of method used to detect climate anomalies at 8.2 ka, as described in text.

finding was cold anomalies in Greenland of up to 7 °C and in Europe of about 1 °C. All also agreed on the lack of signal in the Southern Hemisphere, though few records were available at the time. Differing conclusions were reached about precipitation changes in the Northern Hemisphere tropics, with some studies arguing for drying in specific regions and another claiming that these anomalies were too long-lived to be the actual 8.2 ka event (Rohling and Pälike, 2005).

Since these previous syntheses were published, the number of high-resolution records spanning the 8.2 ka event has doubled. In this paper, we compile and analyze these proxy records. Our main goals are to update previous conclusions reached about climate anomalies at 8.2 ka, particularly those regarding the tropics and Southern Hemisphere. We also place special attention on presenting measures of the duration and magnitude of climate anomalies that can be used to evaluate model output quantitatively.

2 Dataset description and analysis methods

We selected previously published proxy records for our analysis based on several criteria. First, the records have a sampling resolution of 50 yr or better over the interval 7.9 to 8.5 calendar ka. This cutoff was chosen so that detection of a short event (∼ 150 yr) would be feasible. Second, the records have age models with an estimated precision of better than several hundred years taking into account the precision of radiocarbon or U-Th dating and uncertainties that arise from age model interpolation between age control points. This is long relative to the estimated duration of the 8.2 ka event, but better precision is not currently available for the majority of paleoclimate records spanning this time. Third, the proxies measured have well-supported climatic interpretations based on knowledge of modern processes. A total of 262 time series from 114 sites met the above criteria (Fig. 1, Table S1).

The number of sites has doubled since the last global syntheses of the 8.2 ka event were published in 2005–2006 (Fig. 1), both globally and for each continent. A large proportion of the sites meeting our selection criteria are from Europe. North America is also fairly well represented, and other regions more sparsely sampled. The majority of sites included in this study are either lacustrine or marine. This, too, is relatively unchanged from previous syntheses. Data from about half of the sites have been publically archived and are now available as a consolidated dataset from the World Data Center for Paleoclimatology (ftp://ftp.ncdc.noaa.gov/pub/data/paleo/8.2ka/8.2ka-data.csv) and as Supplement to this article. For the other half, we digitized records for the statistical analysis.

Climate anomalies were identified in these records using a statistical test following the approach of Morrill and Jacobsen (2005). First, we detrended those records with significant long-term linear trends using linear regression; this is necessary because our statistical approach loses sensitivity when background trends are present. Then, for each individual record, we measured the mean and variability of the background climate state surrounding each event by calculating the mean ($\bar{x}$) and standard deviation (σ) of proxy values for two windows between 8.5–9.0 and 7.4–7.9 calendar ka BP (Fig. 2). These windows were chosen to bracket the event, while accommodating errors in the age models of several hundred years. A small number of time series contained too few data points in one of these windows for a robust calculation of $\bar{x}$ and σ; for these records, we shifted the windows by 100–200 yr after making certain that this would not impinge upon any possible anomalous event. Given that many of these proxy records contain substantial noise and that just one outlier data point can have a large impact on the calculated standard deviation, we also calculated a series of standard deviations for each window that successively left out one data value at a time. Then, we used the lowest standard deviation along with its corresponding mean to define the upper and lower bounds of background climate variability as $\bar{x}+ 2\sigma$ and $\bar{x}- 2\sigma$. The two windows commonly had different values for $\bar{x}$ and σ, so we used the maximum and minimum values for $\bar{x}+ 2\sigma$ and $\bar{x}- 2\sigma$, respectively, in order to make the stricter test (Fig. 2). Next, we identified all values in the proxy time series between 7.9–8.5 calendar ka that were beyond these respective bounds. Since, on average, about 5 % of data points will fall outside the 2σ bound, other criteria were set for limiting false positives. Only excursions with at least two (three for records with sub-decadal resolution) adjacent anomalous values with the same signed anomaly were identified. This condition makes it statistically unlikely ($p < 0.05$) that the excursions are due to random variations in the time series (Feller, 1966).

For records with a detected climate anomaly and a resolution of 15 yr or better, we also report on event duration using the moving two-tailed z-test method of Wiersma et al. (2011). Thirteen precipitation records and nine temperature records met this criterion. We limited this analysis to the highest-resolution records because only these were sampled densely enough in time to be meaningfully compared to climate model output. Data between 7.9–8.5 calendar ka BP were sampled in overlapping 30-yr increments and their means compared to the mean and variance of the background climate, defined as the periods between 7.4–7.9 and 8.5–9.0 calendar ka BP. Like Wiersma et al. (2011), we defined the duration of the 8.2 ka event as the longest stretch of consecutive overlapping windows whose z-values were all significant at the 99 % level.

The number of proxies that quantitatively estimate temperature and precipitation has grown greatly since 2005. We used these to calculate anomalies near 8.2 ka by again comparing values between 7.9–8.5 calendar ka BP to the average of all data between 7.4–7.9 and 8.5–9.0 calendar ka BP. We report quantitative estimates in two ways: as the single maximum anomaly value and as a mean value calculated over a subjectively determined time interval covering the 8.2 ka event. The subjective approach is necessary because the resolution of many of these records is not high enough to permit a more objective measure of event duration, such as the z-test. For the few records that met our resolution criterion for the moving z-test, calculations of maximum and mean anomalies using the more objective definition of event duration were not substantially different (within a few tenths of a degree or a few tenths of a per mil) from those obtained using our subjective approach, and we present the latter. We note that the mean anomaly over a defined time interval is a measure that has been useful for discussing the magnitude of the

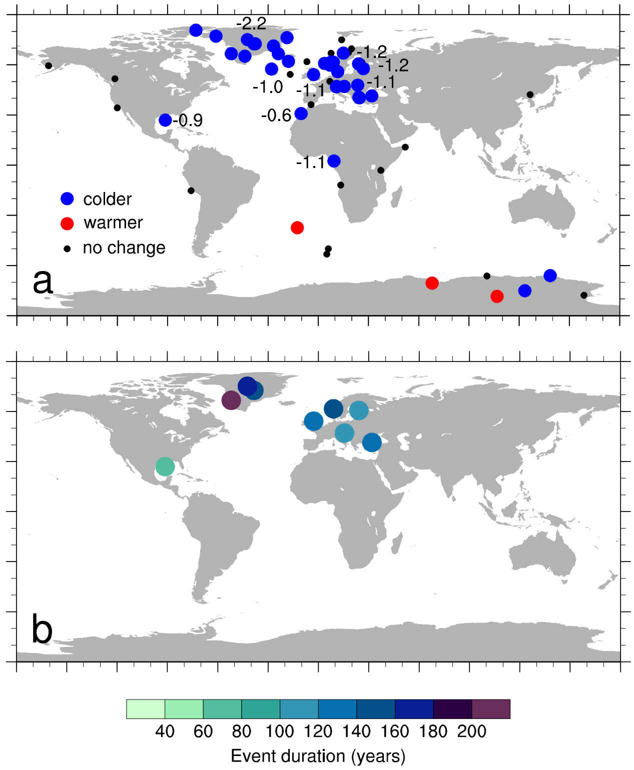

Fig. 3. (a) Temperature anomalies relative to early Holocene background climate (defined as the average between 7.4–7.9 and 8.5–9.0 calendar ka) detected near 8.2 ka by the method described in text. Black dots indicate sites with temperature proxies that did not have an identifiable anomaly. Values plotted are quantitative mean annual temperature estimates in degrees Celsius and are also provided in Table 1. **(b)** Duration of temperature anomalies in high-resolution (better than 15 yr/sample) proxies, as determined using the method of Wiersma et al. (2011).

8.2 ka event (e.g., Thomas et al., 2007; Kobashi et al., 2007) and is a quantity that is easily compared to model output.

3 Climate anomaly patterns at 8.2 ka

3.1 Temperature

Temperature-sensitive proxies indicate cold anomalies around the North Atlantic at 8.2 ka (Fig. 3a), a result common to previous syntheses. New to this study is some evidence for warm anomalies in the Southern Hemisphere (Fig. 3a). These occur in lake records from Nightingale Island in the South Atlantic (Ljung et al., 2008) and Amery Oasis in Antarctica (Cremer et al., 2007) as well as the deuterium record from Vostok (Petit et al., 1999). At the same time, however, several additional records from the Southern Hemisphere indicate cooler conditions at 8.2 ka. Thus, temperature change in the Southern Hemisphere appears to have been regionally heterogeneous.

Isotopic records from the annual-resolved Greenland ice cores estimate the duration of temperature anomalies at 8.2 ka very precisely at 150–160 yr (Thomas et al., 2007; Kobashi et al., 2007). Our analysis of event duration using the moving z-test yields similar values for the GISP2 and NGRIP ice cores in Greenland (160–180 yr, Fig. 3b). According to the moving z-test, event durations in Europe appear to be somewhat shorter than those in Greenland (100–160 yr; Fig. 3b).

Reconstructed mean annual temperature anomalies (MAT) around the circum-North Atlantic are between −0.6 and −1.2 °C with the exception of Greenland, which seems to have experienced larger cooling (Table 1, Fig. 3a). A few estimates are available for summer and winter temperatures. Three pollen records of winter temperature from the Aegean Sea, Greece and Romania have 8.2 ka anomalies that are greater than those for MAT in the same region (Table 1; Pross et al., 2009; Feurdean et al., 2008; Dormoy et al., 2009). A third site in Northern Europe, Vanndalsvatnet (Nesje et al., 2006), shows a winter warming, which may not be coeval with the 8.2 ka event since it immediately precedes a significant cooling. At another site, Gardar Drift (Ellison et al., 2006), the magnitude of winter cooling is quite similar to the amount of summer cooling. Lower-resolution records with quantitative reconstructions paint an equally complex picture, including inferences of greatest cooling in summer (Magny et al., 2001), greatest cooling in the winter (Bordon et al., 2009) and equal summer and winter cooling (Rousseau et al., 1998). Thus, from these data, it is still ambiguous whether winter temperatures cooled more than summer temperatures, as suggested for the 8.2 ka event (Rohling and Pälike, 2005) and for other past freshwater events (Denton et al., 2005).

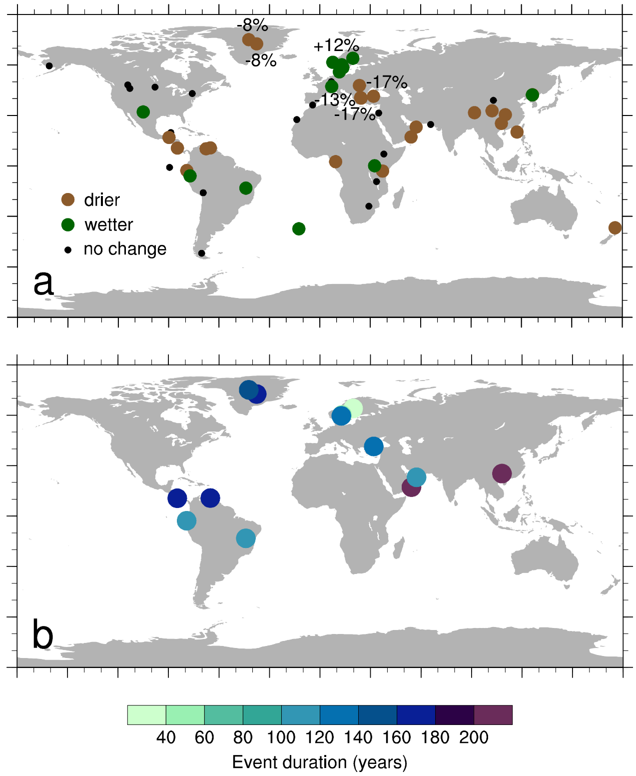

Fig. 4. (a) Precipitation anomalies relative to early Holocene background climate (defined as the average between 7.4–7.9 and 8.5–9.0 calendar ka) detected near 8.2 ka by the method described in text. Black dots indicate sites with precipitation proxies that did not have an identifiable anomaly. Values plotted are quantitative mean annual precipitation estimates, expressed as a percent difference from values averaged for 7.4–7.9 and 8.5–9.0 calendar ka BP, and are also presented in Table 2. **(b)** Duration of precipitation anomalies in high-resolution (better than 15 yr/sample) proxies, as determined using the method of Wiersma et al. (2011).

3.2 Precipitation

The pattern of precipitation anomalies at 8.2 ka includes drier conditions over Greenland, the Mediterranean, and Northern Hemisphere tropics and wetter conditions over Northern Europe and parts of the Southern Hemisphere tropics (Fig. 4a). While reduced rainfall in the Northern Hemisphere tropics at 8.2 ka was noted in previous syntheses, new records from South America showing wetter conditions strengthen support for the idea that the mean position of the Intertropical Convergence Zone shifted southward (Cheng et al., 2009; van Breukelen et al., 2008). While previous syntheses documented decreased precipitation in the Mediterranean (Magny et al., 2003), the pattern of wetter conditions in Northern Europe is a newer result. Many of the records from Northern Europe are indicators of increased runoff associated with the spring snowmelt (Hammarlund et al., 2005; Hede et al., 2010; Zillén and Snowball, 2009; Snowball et al., 1999, 2010) while the inference of dry conditions in southern Europe comes from pollen-based reconstructions for mean annual precipitation (Pross et al., 2009; Feurdean et al., 2008; Dormoy et al., 2009).

According to the moving z-test, most of the high-resolution precipitation anomalies last on the order of 100 to 150 yr (Fig. 4b). The exceptions to this general conclusion are two shorter anomalies of 30 to 50 yr in Sweden (Snowball et al., 1999, 2010) and two longer anomalies of 230 to 280 yr in the Asian monsoon region (Dykoski et al., 2005; Wang et al., 2005; Fleitmann et al., 2003). The Swedish lake records likely record changes in erosion related to spring snowmelt runoff and their shorter event duration might reflect differences in sampling for extreme events as opposed to a change in the mean state. Longer anomalies in Asia were originally discussed by Rohling and Pälike (2005) and attributed to a multi-century cooling upon which the 8.2 ka event might be superimposed. Since 2005, however, there are new precipitation records from the Northern Hemisphere tropics with event durations of < 150 yr (Fig. 4b), lending support to the

Table 1. Quantitative temperature anomalies from early Holocene background values.

Site	Proxy type	Maximum (°)	Mean (°C)	Error[a] (°C)	Duration (yr)
		Mean Annual Temperature			
GISP2	δ^{15}N	−3.3	−2.2[b]	1.1	120
Ammersee, Germany	δ^{18}O	−1.3	−1.1	N/A	90
Lake Rõuge, Estonia	pollen	−2.6	−1.2	0.9	280
Lake Arapisto, Finland	pollen	−2.2	−1.2	0.9	200
South Iceland (thermocline)	Mg/Ca	−1.2	−1.0	1.0	80
Steregoiu, Romania	pollen	−1.6	−1.1	N/A	190
Gulf of Mexico (surface)	Mg/Ca	−1.3	−0.9	1.1	120
Cape Ghir (surface)	alkenone	−0.7	−0.6	˜1	250
Cape Ghir (thermocline)	Mg/Ca	−1.0	−0.6	0.7	80
Gulf of Guinea (surface)	Mg/Ca	−1.9	−1.1	1.2	140
		Winter Temperature			
Tenaghi Philippon, Greece	pollen	−4.0	−2.8	2.5	140
Aegean Sea	pollen	−9.1	−5.9	6.4	120
Gardar Drift (surface)	forams	−1.6	−1.3	∼ 1	80
Vanndalsvatnet, Norway	pollen	2.5	1.7	2.6	240
Steregoiu, Romania	pollen	−5.6	−4.2	2.6	110
		Summer Temperature			
Aegean Sea	pollen	−4.2	−2.6	3.7	120
Hawes Water, UK	chironomid	−1.5	−0.8	1.0	90
Gardar Drift (surface)	forams	−2.1	−1.7	∼ 1	60

[a] Root mean square errors for the calibration as reported by original investigators, N/A = not available. [b] Value based on oxygen isotope temperature sensitivity, inferred from δ^{15}N measurements and applied to GISP2 δ^{18}O time series.

conclusion that precipitation did decrease in these areas coincident with the 8.2 ka event.

There are just six sites with quantitative precipitation reconstructions, all of which are mean annual quantities in either Greenland or Europe. Five of these sites show precipitation decreases, including 8 % in central Greenland and 13–17 % in southeastern Europe (Table 2; Hammer et al., 1997; Rasmussen et al., 2007; Pross et al., 2009; Feurdean et al., 2008). The sixth record, Vanndalsvatnet, shows a precipitation increase, but again there is some ambiguity in the record as to which of several fluctuations might actually be the 8.2 ka event (Nesje et al., 2006).

3.3 Other changes

Some of the proxy records we analyzed reflect climate variables other than temperature and precipitation, or show the combined influences of temperature and precipitation (e.g., glacier advances). These records and their detected 8.2 ka anomalies are shown in Fig. 5. Of particular interest are indications of reduced AMOC (Arz et al., 2001; Ellison et al., 2006), glacier advances in Europe and North America (Menounos et al., 2004; Matthews et al., 2000; Nesje et al., 2001) and strengthening of the Asian winter monsoon (Yancheva et al., 2007). We also included sea ice in this discussion, even

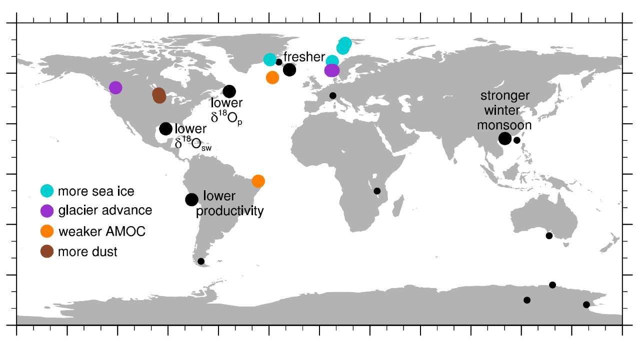

Fig. 5. Climate anomalies relative to early Holocene background climate (defined as the average between 7.4–7.9 and 8.5–9.0 calendar ka) detected near 8.2 ka that were not easily categorized in terms of temperature or precipitation. Anomalies observed at more than one site are plotted in color and described by the legend in the lower left, while anomalies observed at just one site are plotted as large black dots and described by accompanying text. Small black dots indicate sites without an identifiable climate anomaly.

though it has a strong connection to temperature, because it is a variable predicted by climate models and because it participates in important ocean feedbacks. Significantly, several records near convection areas in the North Atlantic indicate sea ice expansion at 8.2 ka (Jennings et al., 2002; Moros et

Table 2. Quantitative mean annual precipitation anomalies from early Holocene background values.

Site	Proxy type	Maximum (%)	Mean (%)	Error* (%)	Duration (yr)
GRIP	ice accumulation	−28	−8	∼ 5	120
NGRIP	ice accumulation	−18	−8	∼ 5	150
Tenaghi Philippon, Greece	pollen	−27	−17	14	110
Aegean Sea	pollen	−24	−13	35	120
Vanndalsvatnet, Norway	pollen	20	12	24	240
Steregoiu, Romania	pollen	−25	−17	17	110

* Root mean square errors for the calibration as reported by original investigators and scaled as a percentage of reconstructed early Holocene background precipitation.

al., 2004; Hald and Korsun, 2008; Sarnthein et al., 2003). Lastly, two varved lake records in central North America show an increase in dust flux at 8.2 ka, possibly related to exposure of Lake Agassiz sediments (Hu et al., 1999; Dean et al., 2002).

With the advent of oxygen isotope-enabled climate models, one of the more comprehensive tests of 8.2 ka simulations uses δ^{18}O anomalies. In Table 3, we present δ^{18}O anomalies separated into three categories: precipitation, surface water and carbonate. Each of these categories reflects different climatic signals. The δ^{18}O of carbonate, which is precipitated from groundwater or surface water, combines the greatest number of signals. These include the δ^{18}O signature of the host water, which records the combined influence of precipitation δ^{18}O and any evaporative enrichment, as well as the temperature-dependent fractionation of oxygen that occurs during carbonate formation. The δ^{18}O of surface water is derived from carbonate δ^{18}O using an independent temperature time series to subtract this temperature-related fractionation. It is most direct to compare modeled δ^{18}O to reconstructed seawater or precipitation values, but for some proxies, such as cave δ^{18}O, measured values can reflect precipitation δ^{18}O changes if changes in ambient temperature and evaporative enrichment are negligible. This may be less true for δ^{18}O of lake carbonates which, depending on the residence time of water in the lake, can be significantly changed through evaporative enrichment.

We also note that some of these δ^{18}O values were measured relative to the Standard Mean Ocean Water (SMOW) standard while others were relative to the Pee Dee Belemnite (PDB) standard. The SMOW and PDB scales are offset by ∼ 30 ‰, but are otherwise linearly related on a nearly 1 : 1 line (Coplen et al., 1983; Clark and Fritz, 1997). Thus, Table 3 combines anomaly values from the SMOW and PDB scales with no conversion between the two.

In Greenland, ice cores record a decrease of −0.8 to −1.2 ‰ (Fig. 6, Table 3). In the North Atlantic and Europe, the decrease is generally less, on the order of −0.4 to −0.8 ‰. These isotopic anomalies are generally thought to reflect temperature effects on the δ^{18}O of precipitation, although there could be some source effect from the meltwater added to the North Atlantic as well (LeGrande et al., 2006).

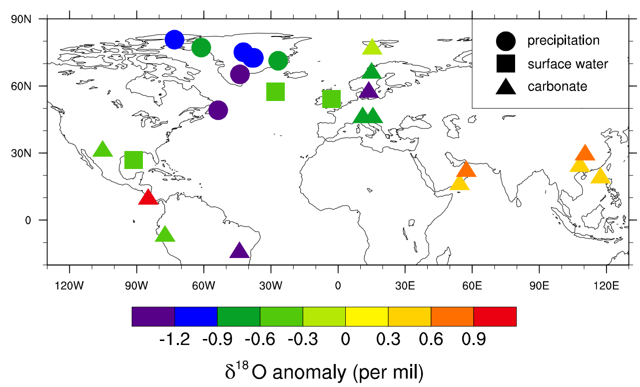

Fig. 6. Anomalies in δ^{18}O detected using method described in text. Sites plotted here are also provided in Table 3.

The smaller changes outside of Greenland are in line with the smaller temperature changes reconstructed quantitatively from Europe (Sect. 3.1). The Northern Hemisphere tropics record an increase of 0.4 to 0.8 ‰, indicating decreased precipitation amount. Conversely, the Southern Hemisphere tropics experienced a decrease of −0.5 to −1.3 ‰, as precipitation likely increased.

4 Discussion and conclusions

The most robust features of the 8.2 ka event from proxy records include: mean annual cooling in the North Atlantic and Europe on the order of ∼ 1 °C; event duration generally of 100 to 150 yr for both temperature and precipitation; decreased precipitation in the Asian monsoon region, Central America and northern South America; and decreases in δ^{18}O of −0.8 to −1.2 ‰ in Greenland and −0.4 to −0.8 ‰ in Europe, and increases in δ^{18}O of 0.4 to 0.8 ‰ in the Northern Hemisphere tropics. These anomalies are all supported by consistent evidence from multiple sites and are unambiguous enough that simulations of the 8.2 ka event should reproduce them.

There are a number of proxy observations that seem likely to hold true, but are somewhat less certain because they have been found at only a few sites. These include: strengthened

Table 3. Oxygen isotope anomalies from early Holocene background values.

Site	Material	Maximum (‰)	Mean (‰)	Duration (yr)
δ^{18}O of precipitation				
GISP2	ice	−1.9	−1.1	140
GRIP	ice	−2.0	−1.1	140
NGRIP	ice	−1.9	−1.0	140
Agassiz	ice	−2.0	−1.0	140
Camp Century	ice	−1.3	−0.8	160
Renland	ice	−1.8	−0.9	120
Dye 3	ice	−1.9	−1.2	140
Nordan's Pond Bog, Canada	peat cellulose	−3.0	−2.6	80
δ^{18}O of surface water				
Gulf of Mexico	seawater	−0.6	−0.4	150
Hawes Water, UK	lake water	−0.9	−0.6	110
Gardar Drift	seawater	−0.7	−0.4	180
δ^{18}O of carbonate				
Ammersee, Germany	ostracod	−0.8	−0.6	90
Katerloch Cave, Austria	cave	−1.3	−0.7	130
Igelsjon Lake, Sweden	bulk lake	−2.7	−2.0	250
Okshola Cave, Norway	cave	−1.0	−0.8	20
Svalbard	benthic forams	−0.4	−0.2	70
Pink Panther Cave, USA	cave	−0.8	−0.4	270
Venado Cave, Costa Rica	cave	2.0	1.0	80
Tigre Perdido Cave, Peru	cave	−1.0	−0.5	170
Padre Cave, Brazil	cave	−1.8	−1.3	60
Qunf Cave, Oman	cave	0.7	0.4	250
Hoti Cave, Oman	cave	1.1	0.8	30
Dongge Cave, China	cave	0.9	0.4	170
Heshang Cave, China	cave	1.1	0.8	130
South China Sea	planktic forams	0.4	0.4	40

Asian winter monsoon; increased precipitation in the Southern Hemisphere tropics; and reductions in precipitation on the order of 10 % and 20 % for Greenland and southern Europe, respectively. We have enough confidence in these observations that they could be used for model-proxy comparison, but we would not necessarily make strong statements about model skill based on whether a model can reproduce these anomalies.

Both of these sets of proxy anomalies are changes that are expected given our current understanding of how freshwater forcing of the North Atlantic impacts climate. When the AMOC slows, reduction in northward oceanic heat transport cools the Northern Hemisphere (e.g., Manabe and Stouffer, 1997). Decreased precipitation in the Northern Hemisphere is, in general, expected due to cooler sea surface temperatures and more sea ice, both leading to less evaporation from the North Atlantic, as well as decreased specific humidity in a colder atmosphere according to the Clausius–Clayperon relationship (Vellinga and Wood, 2002). Strengthening of the Asian winter monsoon is another expected consequence of a colder Northern Hemisphere (Sun et al., 2012).

It is important to emphasize that the uncertainty in the quantitative calibrations of climate is similar to the magnitude of the reconstructed climate anomalies at 8.2 ka (Tables 1, 2). For example, standard errors for most of the mean annual temperature calibrations are $\sim 1\,^{\circ}$C regardless of proxy type (Table 1). Also for the pollen calibrations, the magnitude of reconstructed climate anomalies depends strongly on the particular reconstruction technique used, particularly for seasonal temperature (Table 1; Dormoy et al., 2009; Peyron et al., 2011). This level of uncertainty reduces the confidence that can be placed in the quantitative reconstructions and limits to some extent their usefulness for model comparison. The fact that the reconstructed anomalies in mean annual temperature are consistent across vastly different proxy types does suggest, however, that they still have some utility.

Another consideration in reducing discrepancies between nearby proxy sites, as well as between models and data, is the seasonality of proxy records. Some proxies necessarily indicate seasonal patterns (e.g., organisms that grow during the summer will only record warm season conditions) while others reflect annual means (e.g., lake water balance integrates over the annual cycle); however, our temperature and precipitation compilations presented in Figs. 3 and 4 do not discriminate between annual mean and seasonal signatures. Table S1 indicates the seasonality of each proxy, if this information is known. One goal for future compilations is to improve the separation of seasonal signals.

Some other patterns are suggested by proxy records, but so far are too uncertain to be used as benchmarks. These include: winter temperature decreases in Europe of up to 4 to 5.5 °C that are larger than summer temperature decreases; and regional variability in cold and warm anomalies in the Southern Hemisphere high latitudes. Each relates to unresolved questions about the impacts of North Atlantic freshwater forcing. Denton et al. (2005) suggest that wintertime changes were more extreme than those in summer during abrupt events of the last glacial because the Northern Hemisphere was closer to a sea ice related temperature threshold in the winter. While some proxy records support this seasonal pattern, others indicate substantial summer changes (e.g., Hoffman et al., 2012; Winsor et al., 2012; Young et al., 2012). It is unclear whether a similar sea ice threshold was in play during the early Holocene. While reduction of northward heat transport in the Atlantic might be expected to warm the Southern Hemisphere, as happened at times of North Atlantic freshwater forcing during the last glacial (EPICA community members, 2006), this pattern is ambiguous in proxy records of the 8.2 ka event. It remains to be explained whether oceanic heat transport changes were not large enough at 8.2 ka to cause widespread Southern Hemisphere warming, or if fundamental differences between Holocene and last glacial climate determine the likelihood of a bipolar see-saw response.

Supplementary material related to this article is available online at: http://www.clim-past.net/9/423/2013/cp-9-423-2013-supplement.zip.

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