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The Younger Dryas cold interval as viewed from central Greenland

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Abstract

Greenland ice-core records provide an exceptionally clear picture of many aspects of abrupt climate changes, and particularly of those associated with the Younger Dryas event, as reviewed here. Well-preserved annual layers can be counted confidently, with only $\approx 1\%$ errors for the age of the end of the Younger Dryas $\approx$ 11,500 years before present. Ice-flow corrections allow reconstruction of snow accumulation rates over tens of thousands of years with little additional uncertainty. Glaciochemical and particulate data record atmospheric-loading changes with little uncertainty introduced by changes in snow accumulation. Confident paleothermometry is provided by site-specific calibrations using ice-isotopic ratios, borehole temperatures, and gas-isotopic ratios. Near-simultaneous changes in ice-core paleoclimatic indicators of local, regional, and more-widespread climate conditions demonstrate that much of the Earth experienced abrupt climate changes synchronous with Greenland within thirty years or less. Post-Younger Dryas changes have not duplicated the size, extent and rapidity of these paleoclimatic changes.

1. Introduction

Abrupt climate changes are of societal interest. Under usual economic assumptions, discounting of the future makes extensive reduction of greenhouse-gas emissions economically unfavorable provided that the changes produced by the greenhouse gases require more than a few decades to occur (e.g., Nordstrom, 1992). One can argue about the appropriateness of such discounting (e.g., Schultz, 1996). However, it is possible that greenhouse-gas emissions could trigger a climate jump that is fast compared to the time scale for significant discounting (e.g., Broecker, 1997; Stocker and Schmittner, 1997).

We thus wish to understand abrupt changes, why they occurred, whether they could recur, and how big they might be if they recurred. In paleoclimatic studies of abrupt changes, global spatial coverage is required to understand patterns. Numerous terrestrial, marine, glaciological and biological proxies are needed, and low-latitude sites will be increasingly important because of proximity to places where people are concentrated today and thus where economic impacts of changes are potentially largest. However, polar ice cores occupy a special place in the study of abrupt climate changes, because the polar ice cores combine high time resolution, reliable transfer functions relating sediment properties to the climatic conditions that produced them, and archival on a common time scale of indicators for local, regional, and hemispheric to global climate.

Dating with subdecadal accuracy into the late glacial or beyond is exceedingly difficult. Because of the subdecadal time scale for abrupt climate changes, we cannot compare independent records from different regions and learn whether an abrupt change was synchronous or diachronous. However, the ice cores collect indications of local climate (snow accumulation and temperature), regional climate (wind-blown materials from well beyond the ice sheet), and broader climates (trapped-bubble records of concentrations of trace gases involved in global biogeochemical cycles), which can be related to each other with little or no uncertainty. The relative timing of changes in ice-sheet temperature, wind-blown sea-salt and Asian dust, and atmospheric methane in a Greenland ice core allows one to assess the spatial distribution of climate changes with better time resolution than is provided by the absolute dating of the ice cores.

Here, I focus on the Younger Dryas cold interval, which ended with one of the largest of the abrupt climate changes that have occurred frequently over most of the last 100,000 years and beyond (e.g., Dansgaard et al., 1993). I concentrate on data from the Greenland Icecore

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Project (GRIP) and Greenland Ice Sheet Project II (GISP2) cores from central Greenland. I review some of the major paleoclimatic indicators in the ice cores, the basis for their transfer functions, and what they reveal about the Younger Dryas.

2. Dating

Thus far, no zero-error dates are available for the end of the Younger Dryas or for other, older abrupt changes. Tree-ring dating (e.g., Becker and Kromer, 1993) offers the best hope, although some difficulty remains in identifying the signal of the Younger Dryas termination in the longest tree-ring record (see Gulliksen, Birks, Possnert & Mangerud, 1998). Some ice cores with multiple annual indicators probably allow zero-error interpolation between independently dated horizons at shallow depths, but the best ice-core ages for the end of the Younger Dryas have errors of $\approx 1\%$, based on internal consistency and comparison to independent estimates (Johnsen et al., 1992; Alley et al., 1993; Meese et al., 1997).

Recent estimates of the age of the Younger Dryas termination, including those of Hughen et al. (1998) and Gulliksen et al. (1998), and those summarized by Alley et al. (1997b), indicate an age close to 11,500 yr before 1950 (11.5 kabp), or perhaps older by up to a century or so. Gulliksen et al. (1998) note that the response time of the paleoclimate recorder may matter at such high time resolutions – the Gulliksen et al. (1998) record, for example, may lag climate by a few years to a decade or so. However, most of the high-resolution paleoclimatic recorders likely have lags that are similar to or short compared to the absolute dating errors.

Dating of events in the Younger Dryas, and the estimation of its length, are more difficult than is dating back to the end of the Younger Dryas. One difficulty is that the onset of the Younger Dryas occurred in a series of steps, but lacks the single, nearly unambiguous step that marks the termination. Based on our GISP2 ice-core dating, many indicators persisted at full Younger Dryas levels for between 1200 and 1300 yr (Alley et al., 1993; Meese et al., 1997; see Fig. 1).

3. Snow accumulation

Relative thicknesses of adjacent annual layers in a seasonally resolved ice core allow determination of relative snow accumulation rates. Estimation of absolute rates requires correction for flow thinning. Because ice on the east side of Greenland flows east, that on the west side flows west, and no crevasse opens in the middle, layers must be progressively stretched and thinned as they are buried.

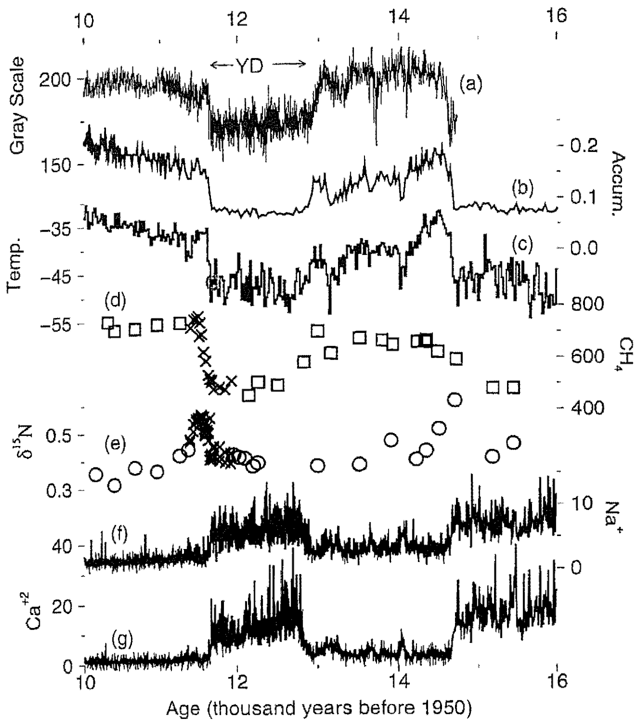

Fig. 1. High-resolution data from the GISP2 ice core and from the Cariaco Basin, covering the abrupt onset of the Bolling/Allerod warm period, the cool Younger Dryas interval (YD), and the warming into and through the Preboreal that followed. (a) Gray-scale data from Hughen et al. (1996, 1998) for a core from the Cariaco Basin, offshore Venezuela. The gray scale is affected by productivity in response to wind-driven upwelling, with lower numbers corresponding to enhanced windiness. Note that the time scale for this record was derived independently of those for the ice core. (b)–(e) Data mostly from Severinghaus et al. (1998), with the accumulation and temperature originally from Cuffey and Clow (1997). (b) Accumulation in to m ice/yr. (c) Temperature converted from ice-isotopic ratios (Grootes & Stuiver, 1997) using the glacial–interglacial calibration of Cuffey et al. (1995), shown in degrees C. The Severinghaus et al. (1998) estimate of temperature in the Younger Dryas is plotted as a circle; uncertainties of about $\pm$ 3°C are not much larger than the symbol. (d) Methane data (ppbv) from Brook, Sowers and Orchardo (1996) (squares) and Severinghaus et al. (1998) (x). (e) $\delta^{15}N$ data in per mil are from Severinghaus et al. (1998), with (o) from their Fig. 2 and (x) from their Fig. 4. (f)–(g) Sodium (f) and calcium (g) data from Mayewski et al. (1997), normalized to their mean concentration in the millennium before the Little Ice Age following Alley et al. (1997) using the accumulation-rate estimates of Alley et al. (1993). Most of the ice-core data, and many related data sets, are available on The Greenland Summit Ice Cores CD-ROM, 1997, National Snow and Ice Data Center, University of Colorado at Boulder, and the World Data Center-A for Paleoclimatology, National Geophysical Data Center, Boulder, Colorado, www.ngdc.noaa.gov/paleo/icecore/greenland/summit/index.html.

In the simplest flow model (Nye, 1963), thinning is proportional to burial. A layer buried halfway from surface to bed will have been thinned by half, a layer buried three-fourths of the way from surface to bed will have lost three-fourths of its original thickness, and so on. More-accurate models (e.g., Dansgaard and Johnsen, 1969) have somewhat faster layer thinning at shallow depths, producing thinner layers and older ages than the Nye time scale. These models must become inaccurate in very old ice because they do not allow for the initial growth of the ice sheet. Numerical models can handle this complication (e.g., Schott et al., 1992). However, this is only likely to be significant in the very oldest ice. More importantly, flow over bedrock bumps or inhomogeneous flow tend to disrupt layers in the lowermost few percent of the ice sheet (e.g., Bender et al., 1994; Alley, et al., 1995b; Chappellaz et al., 1997b).

Complications in correcting layer thickness for flow thinning are related to uncertain knowledge of the history of changes in mode of flow of the ice (and especially in basal velocity), and of changes in the ice-sheet thickness and shape. For central Greenland, these are very minor factors. The ice sheet is modeled to have remained frozen to its bed throughout the last climate cycle (e.g., Firestone et al., 1990), and thus to have had essentially zero basal velocity. Ice flux by internal deformation is proportional approximately to the fifth power of the thickness and the third power of the surface slope (e.g., Paterson, 1994 Chapter 11). Because the Greenland ice sheet sits on an island surrounded by deep water and so cannot change its lateral extent greatly, changes in surface slope over time are tied to changes in thickness in the center of the ice sheet, yielding an eighth-power dependence of ice flux on thickness in central regions. With such a strong dependence, there is little thickness change even for large climate changes (e.g., Raynaud et al., 1997).

Flow-model corrections of annual-layer thicknesses to obtain accumulation rates have included those of Alley et al. (1993), Dahl-Jensen et al. (1993), Cutler et al. (1995), Bolzan et al. (1995), and Cuffey and Clow (1997). Uncertainties in flow correction are basically zero in young ice, increasing to about 15% at 40 ka and then more in older ice (Cuffey and Clow, 1997).

The relation between snow accumulation and precipitation can range from direct to complex. Steep, small ice caps can have wind-scour zones (e.g., Fisher et al., 1983). Sublimation can be important in some areas, and especially those with small snowfall and strong katabatic winds. Surface topography resulting from ice flow over bedrock can create "standing waves" in the snow-accumulation pattern, which the ice flows through (Gow and Rowland, 1965). However, coring in the central region of a large, broad ice cap with sufficiently rapid snow accumulation, as in central Greenland, avoids significant contributions from any of these. "Noise" from snow drifts or other spatial heterogeneity in accumulation affects year-to-year accumulation, but snowdrift anomalies cannot be sustained for multiple years. Hence, accurate reconstructions of accumulation rate are straightforward over tens of thousands of years in central Greenland.

For the GISP2 site in central Greenland, the accumulation rate doubled at the end of the Younger Dryas, from approximately 0.07 ± 0.01 to 0.14 ± 0.02 m ice/yr, with the change in one to a few years. An additional increase of 50–100% after the Younger Dryas brought accumulation to the modern 0.24 m ice/yr (Cuffey and Clow, 1997).

4. Aerosols

Ice cores collect whatever is transported through the air, including materials of marine, continental, biologic, volcanic, atmospheric, and extraterrestrial origins. Changes in ice-core concentrations of these materials can reflect numerous processes at the source, along the path to the ice sheet, during transfer to the ice sheet, or after deposition on the ice sheet.

We can hope that models eventually will simulate all of these processes accurately, and that model validation will require only knowledge of concentrations of materials in the ice cores. For now, there is considerable interest in estimating atmospheric loadings from ice-core data.

A large body of recent work (e.g., Wolff and Bales, 1996) illustrates the complexities of the air-snow transfer process. Some species are reversibly deposited to the snow, some undergo chemical changes in the snow, and some are mobile in the snow. Changes in ventilation/wind pumping may affect the rate of transfer to the snow. Because of such complexity, a small change in concentration of some species may have numerous causes. However, the huge changes observed across abrupt climate events in the Greenland ice cores are inexplicable except by large changes in atmospheric loading of wind-blown materials (e.g., Mayewski et al., 1997).

Colder, drier times have produced dustier ice (De Angelis et al., 1997; Mayewski et al., 1997; Zielinski and Mershon, 1997). Reduced dilution of aerosols during dry times because of lack of water contributed to the dustiness of the ice. However, because the changes in aerosol concentration have been much larger than the changes in snow accumulation, other processes must have been more important.

A simple, data-driven model (Legrand, 1987; Davidson, 1989; Whung et al., 1994; Alley et al., 1995a; De Angelis et al., 1997) allows improved estimates of atmospheric aerosol loading by correcting ice-core concentrations for changes in the dilution effects. Some aerosols are deposited directly (dry deposition) and some are deposited attached to snowflakes (wet deposition). During a dry year, wet deposition will be low. A plot of chemical flux versus water flux (Fig. 2) produces a noisy linear array. The intercept at zero water flux can be used as an estimate of the dry-deposition rate, which increases with the atmospheric loading, and the slope of the line (increase in chemical flux per unit increase in water flux) also increases with the atmospheric loading. Different climate states produce different linear arrays, with increased slope and intercept during colder times reflecting

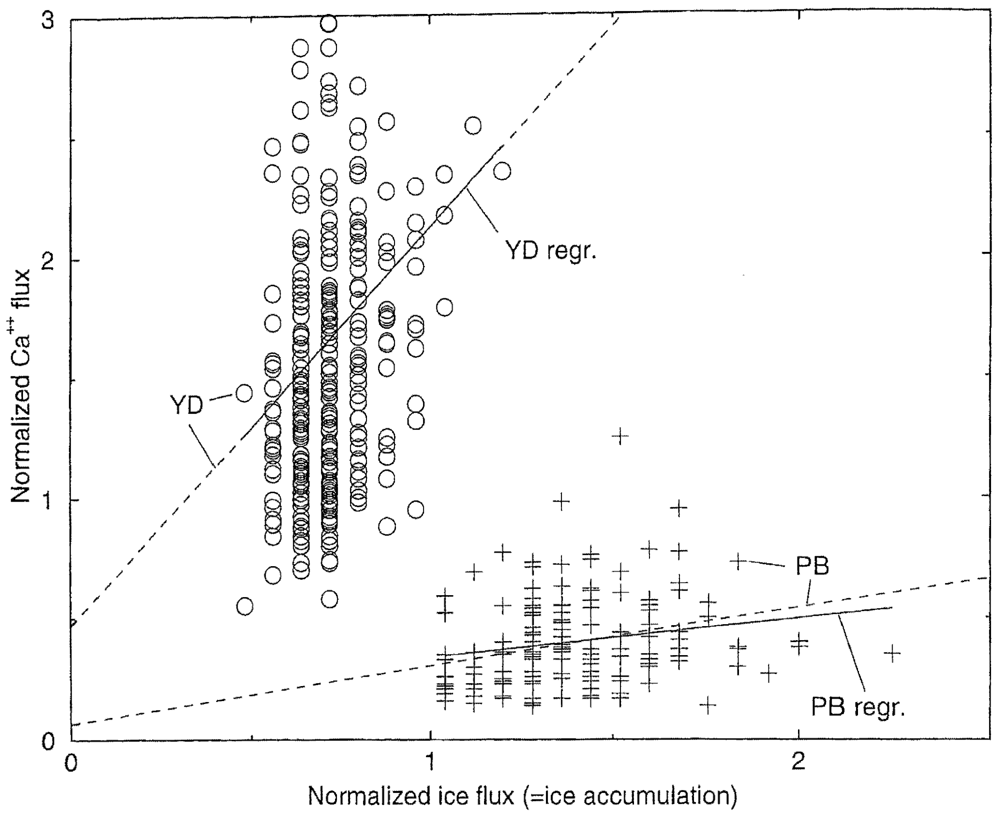

Fig. 2. Comparison of calcium flux to ice flux (snow accumulation rate) for the GISP2 ice core, modified from Alley et al. (1995), with the calcium data from Mayewski et al. (1994, 1997). The data have been normalized by their average values between 11,322 and 14,035 abp. The data sets shown represent 11,322–11,637 abp (during the Preboreal, labeled PB) and 11,759-12,854 abp (Younger Dryas, labeled YD). Solid line segments are individual linear regressions for the YD and for the PB data; dashed lines represent a joint inversion that requires the ratio of the slopes of the lines through the data from the two times to equal the ratio of the intercepts, as hypothesized in the atmospheric-removal model. That ratio is 6.95 in this case, indicating a sevenfold change of calcium loading in the atmosphere sampled by snow accumulation between the Younger Dryas and the few centuries following it (Alley et al., 1995).

in part increased atmospheric loading during colder times (Fig. 2).

Tests of this model show that it does explain a significant fraction of the variance in data sets (Alley et al., 1995a; De Angelis et al., 1997). Application of the model to central Greenland shows that the end of the Younger Dryas produced: a three-fold drop in sea-salt; a four-fold drop in submicron dust, increasing with increasing particle size to a seven-fold drop in 3–10 micron dust; and a seven-fold drop in calcium. An additional two-fold drop in sea-salt, and a slightly bigger drop in calcium, occurred through the early Holocene (Alley et al., 1995a; Alley, Mayewski, Sowers, Stuiver, Taylor and Clark, 1997a).

Several effects may have contributed to these large changes (e.g., Wolff et al., 1993; Mayewski et al., 1994,1997; Zielinski & Mershon, 1997). Higher Younger Dryas loadings of sea-salt can be linked to stronger Younger Dryas winds, which would have increased production of sea-salt aerosol and reduced transport times and so fallout en route. Drier Younger Dryas conditions also would have reduced precipitation scavenging en route. Coarser continental dust during colder times argues for stronger winds then. Changes in circulation patterns or in sites or sizes of source areas cannot be ruled out, although the results of Biscaye et al. (1997) at least suggest that such changes were not large.

Biscaye et al. (1997) used mineralogic and isotopic characteristics of ice-core dust and of dusts from potential source areas to track atmospheric circulation. Biscaye et al. (1997) found a dominant Asian signature in the ice-core dust. They did not sample the Younger Dryas, but found little change in the source area for other climate transitions; by analogy, source-area changes are unlikely to have been dominant for the Younger Dryas. Wolff et al. (1993) argued that independent evidence limits the potential changes in the sizes of Asian source areas. A combination of several causes is likely to have been important in producing observed changes in dustiness, but a role for stronger winds during colder times is almost certainly required (Biscaye et al., 1997).

An upper limit on the wind-speed increase over source regions can be obtained by assuming that higher wind speed was the only cause of increased atmospheric dust loading, and calculating following Petit et al., (1981) the required wind-speed change to explain the dust change. Such a calculation indicates that Younger Dryas winds were $\leqslant 70\%$ stronger than during the centuries or so that followed, and Younger Dryas winds were no more than slightly above double late-Holocene values.

5. Paleothermometry

Paleotemperature receives much attention in scientific and especially in popular treatments. Temperature measurements are invariably local, and so large networks of reliable reconstructions are required to provide useful information. Ice sheets offer the possibility of several independent temperature estimates. Central Greenland has produced exceptionally robust, independent thermometers. Notice that these calibrations are site-specific; similar work is required at other sites, and the local calibrations for GISP2 and GRIP cannot be applied confidently to other sites without such work (cf. Jouzel et al., 1997).

The classical paleothermometer is the stable-isotopic composition of accumulated ice. Because the vapor pressure of "heavy" water (that containing ^{18}O or ^{2}H) is lower than for "ordinary" water (containing only ^{16}O or ^{1}H), condensation and precipitation during air-mass cooling preferentially remove heavy water, depleting the air mass of heavy water so that further precipitation becomes progressively lighter. The colder an air mass is, the more depleted of heavy water it has become and the lighter the precipitation it produces (e.g., Dansgaard, 1964; Robin, 1983).

Numerous factors in addition to temperature can affect the isotopic composition of accumulated snow (reviewed by Paterson, 1994, chapter 15). Some are related to source conditions; the isotopic composition or temperature of ocean surface waters supplying water to precipitating air masses for a site can change over time, changing the isotopic composition of the precipitation. Isotopic composition of precipitation also is sensitive to the path followed, the seasonality of precipitation, adiabatic or isobaric cooling, degree of convective lifting, offset between cloud and ground temperature, sub-cloud sublimation, etc. Site conditions also can affect isotopic composition, which is sensitive to post-deposition sublimation or drifting, diffusion, ice-flow transport from the site of deposition, etc.

Much research (reviewed by Jouzel et al., 1997) demonstrates that: (i) site temperature is the most important factor controlling stable isotopic composition of accumulated snow in most places; (ii) the isotopic paleothermometer can be used reliably; but (iii) the calibration of the paleothermometer can change significantly over time and space, so that local calibrations are desirable. Indeed, the documented large differences between previously expected and observed calibrations in central Greenland suggest extreme caution in use of water isotopes in paleothermometry in the absence of reliable local calibrations. The use of atmospheric models (Fawcett et al., 1997; Krinner et al., 1997; Jouzel et al., 1997) to calibrate the water-isotopic paleothermometer has great promise. Beyond that, borehole paleothermometry and gas-isotopic thermometry have proved especially valuable in Greenland.

The temperature of the ice itself provides a paleothermometer. Ice from the ice age was deposited at a lower temperature than today, and in central Greenland has not had time to warm to modern temperatures (Cuffey et al., 1995; Johnsen et al., 1995; Cuffey and Clow, 1997; Clow et al., 1997; Dahl-Jensen et al., 1997; Dahl-Jensen et al., 1998). Information on old, short-lived events has been lost from the borehole-temperature profile because of diffusion. However, the remaining information is almost entirely a record of surface temperature if ice flow is understood well and significant melting has not occurred, as in central Greenland (e.g., Alley and Koci, 1990; MacAyeal et al., 1991; Cuffey et al., 1992; Cuffey et al., 1994).

Several techniques have been used to "read" the history of surface temperature by inverting the modern borehole-temperature profile (e.g., MacAyeal et al., 1991; Clow et al., 1997; Dahl-Jensen et al., 1997, 1998). An especially valuable technique has been to use the stable-isotopic record of ice-core ice as a shortcut in the inversion (Cuffey et al., 1992, 1994, 1995; Johnsen et al., 1995; Cuffey and Clow, 1997; cf. Paterson and Clarke, 1978). A guess at the relation between ice isotopes and temperature turns the ice isotopes into a provisional surface-temperature history. This surface-temperature history is used to force a model of heat and ice flow, producing an estimated modern temperature-depth profile that can be compared to the observed profile in the ice sheet. The initial guess at the isotope-temperature relation is then adjusted using an inverse procedure to minimize the misfit between calculated and observed borehole temperatures. A poor fit could be due to any number of causes. However, a good fit demonstrates the utility of the ice-isotopic thermometer and calibrates it, because the model used does not have enough free parameters to link two large data sets (ice-isotope ratios and borehole temperatures) unless they are indeed related.

The minimum duration of a paleoclimatic event that is directly resolvable using borehole paleothermometry increases with increasing age, and typically falls between 1/10 of the age and the age. A 1200-yr-long event 12,000 years ago, such as the Younger Dryas, falls near or below this threshold (Firestone, 1995). The borehole-calibration technique then allows one to say that there is temperature information in the higher-frequency variations of the ice stable isotopes such as the Younger Dryas signal, but does not rule out the possibility that high- and low-frequency calibrations differ.

New gas-isotope techniques pioneered by Severinghaus, Sowers, Brook et al., (1998) offer this high-frequency calibration at important transitions. On cold ice sheets with little or no melting, the transformation of snow to ice takes decades to millennia, and there are tens of meters of old snow called firn above the depth at which bubbles are formed. Wind-mixing is limited to the top few meters, so tens of meters of firn exist in which gases exchange with the free atmosphere by diffusion alone.

Gravitational fractionation causes the gases trapped by bubbles forming at the bottom of the firn to be very slightly heavier than the free atmosphere, in predictable and regular ways (Sowers et al., 1989). However, following abrupt climate changes, the trapped-gas composition is also perturbed slightly by thermal diffusion.

Consider the abrupt warming at the end of the Younger Dryas. The time scale for the heat from that warming to penetrate to the depth of bubble formation in central Greenland is around 100 yr. The time scale for diffusive mixing of the air in the pore spaces down to the depth of bubble formation is only a few years. And, in a diffusive medium, gases subjected to a temperature gradient are fractionated, with heavy ones typically moving to where temperature is lower (Severinghaus et al., 1998).

Hence, a very few years after an abrupt warming, the air being trapped in bubbles should shift to heavier values, with this anomaly lasting a few decades before relaxing as the deeper ice warms. This allows two independent paleothermometers as described next, one for the absolute temperature just preceding the abrupt change, and one for the size of the temperature step (Severinghaus et al., 1998).

The transformation of firn to ice is well-understood, at least empirically. Transformation is more rapid at higher temperature or under greater load. The number of years from deposition of snow to formation of bubbles (the gas-age/ice-age difference) is thus a known function of temperature and snowfall if they are constant long enough to reach steady conditions (Herron and Langway, 1980). Nonsteady-state models also are available (e.g., Schwander et al., 1997), but are not needed in this case because the Younger Dryas maintained a sufficiently stable climate long enough to achieve a nearly steady profile down to the close-off depth. Also, the time scale for significant change in the firn depth-density profile is longer than the few years for gas diffusion through the pore spaces to the bubble-formation depth.

The end of the Younger Dryas is well marked in ice-isotopic, ice-chemistry, ice-dust, ice-layer-thickness, and other records (using the main, middle step in the records of Taylor et al., 1997 as discussed below), and is well marked in the gas isotopes. The number of annual layers can be counted between the records of the Younger Dryas termination in the gas and in the ice. The snow accumulation can be estimated from the layer thickness and our understanding of ice flow. Because the gas-age/ice-age difference is a known function of temperature and accumulation, and we know the gas-age/ice-age difference and the accumulation, temperature can be determined (Severinghaus et al., 1998).

Severinghaus et al. (1998) are developing yet another paleothermometer. Following an abrupt warming such as the termination of the Younger Dryas, the depth-density profile does change over a century or longer in response to the temperature and accumulation changes. This affects the thickness of the diffusive gas column in the firn and thus the degree of gravitational fractionation and the isotopic composition of trapped bubbles. Changes in wind speed also could affect the thickness of the diffusive layer in the firn, and so the gravitational fractionation. At the same time, the isotopic composition of trapped gases would be evolving because of changing temperature gradients between the surface and the bubble-formation depth.

Fortunately, gravitational fractionation of gas isotopes is mass-dependent, but thermal gradients fractionate gases based on other factors as well, such that argon is not fractionated thermally as strongly as is nitrogen. By comparing the amplitude of changes in nitrogen and argon isotopes in the gas, Severinghaus et al. (1998) were able to separate thermal from gravitational effects, and learn how large the abrupt temperature change was. (Results are somewhat preliminary pending further measurement of interdiffusion coefficients at appropriate temperatures (Severinghaus et al., 1998).)

All of this paleothermometry yields highly consistent results. The borehole-temperature calibrations of Cuffey et al. (1995) and Johnsen et al. (1995) for the glacial-interglacial transition yield exceptionally large changes of more than 20°C since the last glacial maximum (cf. Clow et al., 1997; Cuffey and Clow, 1997; Dahl-Jensen et al., 1997, 1998). Our understanding of the climate changes associated with the Younger Dryas and the glacial maximum would lead us to expect that the ice-isotopic calibration would be similar for the end of the Younger Dryas and for the glacial–interglacial transition (Alley and Clark, 1999). This is borne out by the Severinghaus et al. (1998) estimates of the Younger Dryas temperature and of the temperature change at the end of the Younger Dryas, which are in good agreement with estimates obtained by converting the ice-isotopic record to temperature using the Cuffey et al. (1995) glacial–interglacial calibration.

The temperature estimated for the Younger Dryas using these techniques is about 15°C colder than modern, with about half of the temperature change to today having occurred in an abrupt step and half in a ramp during the millennia following the Younger Dryas. The Severinghaus et al. (1998) preliminary nitrogen and argon results indicate a 5–10°C step warming at the end of the Younger Dryas, in full agreement. Note that this 5–10°C warming was abrupt, with much of the change in a few decades or less (Severinghaus et al., 1998). Also note from Fig. 1 that much "noise" in the ice-isotopic record, especially near the abrupt jump, introduces some uncertainty in assigning magnitudes to step versus ramp changes in temperature estimated from ice-isotopic ratios.

The reader should remember that paleothermometry here has dealt with surface-temperature changes. Two complications exist. Elevation changes of order 100 m through the atmospheric lapse rate may have produced temperature changes of order 1°C over millennia (Cuffey and Clow, 1997). More importantly, the development of atmospheric temperature inversions over ice sheets is likely to cause surface-temperature changes to be larger than temperature changes averaged through the troposphere. A crude scaling suggests that the free-tropospheric changes may have been about 2/3 of the surface-temperature changes (Cuffey et al., 1995).

6. Atmospheric composition

Trapped gases in ice cores typically provide highly accurate records of past atmospheric composition (e.g., Raynaud et al., 1993). The small isotopic effects from gravitational fractionation and thermal diffusion are well understood and do not significantly change the concentrations of important trace gases. Certain Greenland ice has somewhat anomalous CO_2 concentrations owing to chemical reactions that produce or consume carbon dioxide (Anklin et al., 1997; Smith et al., 1997), but most Greenland ice and the cleaner Antarctic ice produce reliable results.

The Younger Dryas interval did not produce significant anomalies in carbon-dioxide concentration of the atmosphere (Blunier et al., 1997). Any effect of the Younger Dryas on global CO_2 is masked by the deglacial trend of rising CO_2 at orbital frequency (Blunier et al., 1997).

The Younger Dryas interval is marked by strongly reduced methane concentration. Concentrations were almost 700 ppbv before the event and almost 750 ppbv after the event, but fell to 475 ppbv during the Younger Dryas (Chappellaz et al., 1997a). Changes in source strengths in both boreal and tropical/subtropical regions are implicated, based on the size of the change, on reconstructed source strengths, and on changes in the inter-hemispheric gradient of methane between Greenland and Antarctic cores (Chappellaz et al., 1993; Brook et al., 1999).

Ordinarily, dating of gas records in ice cores includes uncertainty in ice-core dates and additional uncertainty from difficulty in estimating the gas-age/ice-age difference. Severinghaus et al. (1998) were able to remove this latter uncertainty almost entirely for the end of the Younger Dryas. The thermal-diffusion gas-isotopic spike in the GISP2 trapped gases shows which bubbles were being closed when the surface of Greenland warmed by many degrees in decades or less (Fig. 1). Severinghaus et al. (1998) have approximately 30-yr resolution in their sampling for gas isotopes and methane. Consecutive samples show the first clear indication of post-Younger Dryas warming and then post-Younger Dryas rise in methane. Hence, methane increase lagged warming in central Greenland by 0–30 yr. The methane rise continued for more than a century. The long rise may be related to the time for wetlands to fill or thaw after conditions became appropriate for those changes, or for methanogenic communities to become fully established (Brook et al., 1999).

The lag of methane behind Greenland warming eliminates all models that appeal to methane changes to explain the termination of the Younger Dryas (Severinghaus et al., 1998). Methane may have provided a small feedback, but even here, the slower rise of methane than of temperature argues against any major role. The high probability that tropical and subtropical regions were involved in the methane change (Chappellaz et al., 1997a; Brook et al., 1999), and the 0–30-yr lag between warming in Greenland and onset of methane rise, argue that near-synchronous climate change affected large regions of the Earth's surface, although we cannot tell exactly which regions were involved or whether the signal was truly global or "only" hemispheric to sub-hemispheric. The observed timing also eliminates models in which the climate change originated in the tropics or elsewhere and was transmitted by slow oceanic processes to the North Atlantic (Severinghaus et al., 1998). The most direct interpretation is that the climate change was transmitted through the atmosphere to broad regions, and a North-Atlantic origin is fully consistent with the data.

7. Derived characteristics

Accurate paleoclimatic reconstructions allow comparisons between different paleoclimatic variables. The ice-isotope/temperature relation and the temperature/snow-accumulation relation have proven instructive in studying the Younger Dryas, and indicate changes in storm tracks and seasonality of processes at the end of the Younger Dryas.

Temperature and snow accumulation are related in many places on ice sheets, with warming increasing water content of the air and so snowfall. The "usual" assumption in Greenland is that accumulation increases

$\approx 4\%/°C$ warming (Johnsen et al., 1989; Oerlemans, 1991), which is equivalent to assuming that precipitation is extracted from throughout the troposphere proportional to the saturation vapor pressure. Huybrechts, Letreguilly and Reeh, (1991) used a slightly higher sensitivity of 5.3%/°C. Higher sensitivities up to 10%/°C might be possible if precipitation were extracted solely from the base of the troposphere (Kapsner et al., 1995).

The warming of 5–10°C at the end of the Younger Dryas occurred with a doubling of snow accumulation. Taking the 5–10°C warming at face value, and using 4% increase per degree warming, would yield an accumulation-rate increase of 22–48%. Sensitivity of 7% per degree and the maximum warming of 10°C would be required to explain the doubling of accumulation. Because the temperature change applies to the surface, and free-tropospheric changes are likely to have been only about 2/3 of the surface change (Cuffey et al., 1995), this indicates that it is difficult or impossible to explain the accumulation-rate change solely based on warming (Kapsner et al., 1995). "Storminess" apparently increased in central Greenland as the Younger Dryas ended.

At the same time, some broad measure of windiness decreased, based on the observed reduction in grain size of particulates in the ice and of reconstructed atmospheric concentrations of wind-blown materials (Alley et al., 1995a; Zielinski and Mershon, 1997). A likely inference is that the storm track shifted closer to central Greenland when the Younger Dryas ended, so that Greenland storminess increased even as the strength of the core of the storm track decreased (Kapsner et al., 1995). As discussed below, such a storm-track shift is supported by atmospheric modeling (Fawcett et al., 1997).

The calibration of the ice-isotopic paleothermometer, derived by comparing shifts in ice-isotopic ratios to independent estimates of temperature change, is itself a paleoclimatic indicator (Alley et al., 1999). By assuming that space can be substituted for time in studying a system, Dansgaard (1964) estimated that a change of about 0.7 per mil in mean-annual isotopic ratio corresponds to a temperature change of 1°C. Data from Greenland indicate that major climate changes actually exhibited a dependence of about half this or slightly less (Cuffey et al., 1995; Cuffey and Clow, 1997).

Multiple causes for this large difference may exist, and two likely were important: changes in source temperature, and changes in snowfall seasonality. First, because the isotopic thermometer measures the temperature difference between moisture source and precipitation site, the correlated changes that likely occurred in temperature at source and site would cause isotopic changes to underestimate temperature changes, as observed (Boyle, 1997).

Much or all of the signal is explainable by changes in the seasonality of precipitation in central Greenland (Fawcett et al., 1997; Jouzel et al., 1997; Krinner et al., 1997). As discussed below, changes in oceanic heat transport are strongly implicated in the Younger Dryas cooling. The energy balance of the high-latitude North Atlantic region is dominated by sunshine in the summer, with oceanic processes much more important in the winter. Hence, reducing oceanic heat transport is expected to cause cooling primarily in the winter.

Through a combination of reduced saturation vapor pressure plus southward shift of the storm track to follow a more-southerly sea-ice edge, oceanic wintertime cooling would be expected to be accompanied by wintertime cooling and precipitation reduction in central Greenland. Ice isotopes record temperature during snowfall, and so would not reflect extreme wintertime cooling if accompanied by wintertime drying (Fawcett et al., 1997). Atmospheric modeling supports this explanation (Fawcett et al., 1997; Krinner et al., 1997).

8. The end of the Younger Dryas

The rapidity of the termination of the Younger Dryas is of great interest. Estimated rates of change have increased as resolution of available paleoclimatic records has improved. The Greenland ice-core records offer an exceptionally clear view of the termination, which was indeed abrupt.

The most complete analysis of the termination in Greenland records was developed by Taylor et al. (1997). They identified three "steps", each lasting approximately 5 yr, and spread over about 40 yr. However, the first step did not move out of the band of Younger Dryas variability, and so one can argue whether it is significant. Most of the change in most of the variables occurred in the second step.

Some of the changes, such as that in snow accumulation (Alley et al., 1993), can be interpreted as having been even more dramatic. The record appears to indicate a "flicker" (Taylor et al., 1993), followed by slightly more than doubling of accumulation in three years, with most of that change in a single year (Fig. 3). Given the possibility of snowdrifting, and of mis-identification of a year, one cannot insist on this interpretation, but even the possibility of a 1-yr end of the Younger Dryas is provocative (also see Hammer, 1997).

The gas-isotopic data of Severinghaus et al. (1998) are also important here. They document large warming in Greenland on a decadal time scale, and show that methane increase, which probably indicates widespread increase in wetland area, lagged Greenland warming by one sample (0–30 yr). When combined with the near-synchrony of changes in wind-blown materials, this

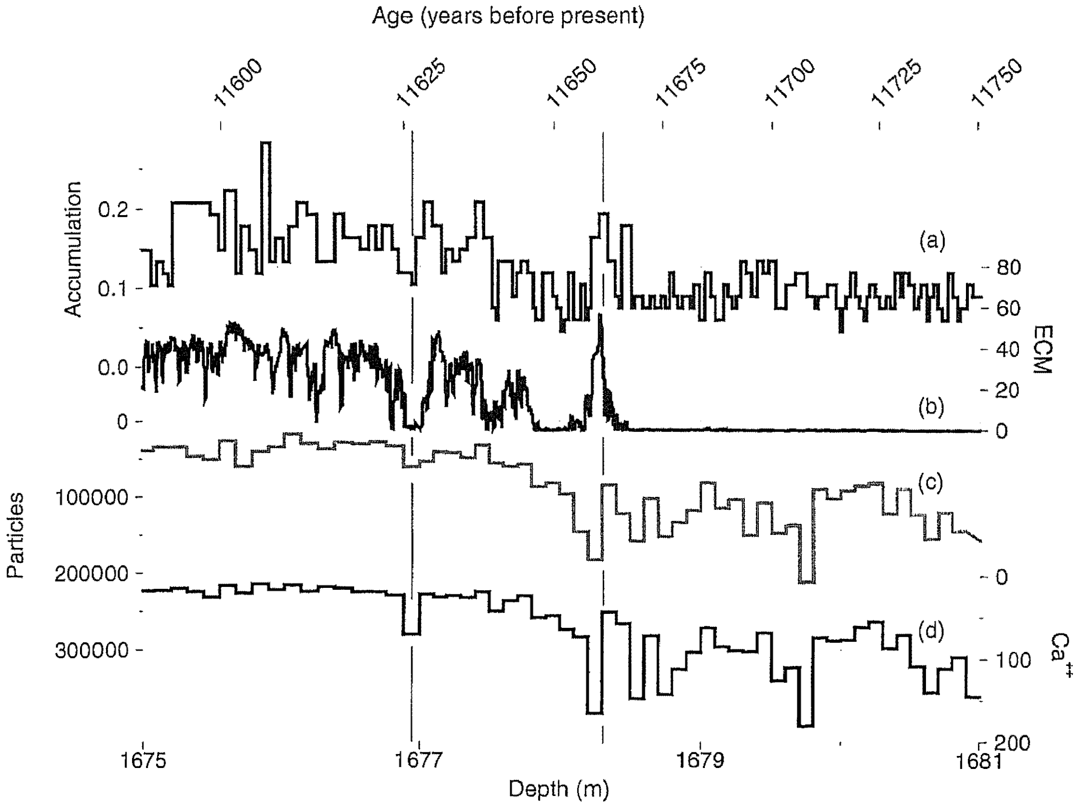

Fig. 3. The end of the Younger Dryas in the GISP2 ice core from central Greenland. Shown are selected curves from Taylor et al. (1997), illustrating the possible "flickering" behavior near the transition: (a) snow accumulation in m ice/year (Alley et al., 1993); (b) electrical conductivity (ECM), current in microamps (Taylor et al., 1992, 1993, 1997); (c) the insoluble-particulate concentration in number per ml (Zielinski and Mershon, 1997); (d) the soluble-calcium concentration in ppb (Mayewski et al., 1997). The main step at the end of the Younger Dryas falls between 1677 and 1678 m depth, between the two probable "flickers" indicated by the vertical lines at about 1676.9 and 1678.2 m depth.

strengthens the argument for atmospheric transmission of an abrupt signal, as noted above.

The work of Taylor et al. (1993) showed that there is a tendency for the climate system to "flicker" near major transitions such as the end of the Younger Dryas (Fig. 3). In "flickers", some variable oscillates between two levels (one characteristic of a cold stadial, and the other characteristic of a warm interstadial) over years before stabilizing at one of the levels. Intriguingly, increased "noise" possibly corresponding to "flickers" can be modeled near transitions (e.g., Stocker and Wright, 1996).

Taylor et al. (1997) interpreted the detailed phasing of different paleoclimatic indicators at the Younger Dryas termination (cf. Dansgaard et al., 1989), and suggested that they found evidence of mid-latitude changes leading high-latitude changes by about 15 yr, although with the clear lower-latitude indicator, methane, lagging Greenland warming. Further studies of other abrupt transitions are in progress to assess the generality of the results (Wolff et al., 1998), and to learn whether enhanced "noise" near transitions might have contributed to the observed signal.

Detailed laser-light-scattering profiles of dust concentration in the GRIP ice core by Hammer (1997) (cf. Ram and Koenig, 1997) show seasonal cycles with "winter" highs and "summer" lows. Hammer (1997) found that the end of the Younger Dryas occurred in 1–2 yr in the winter records, but 20–30 yr in summer records.

Noteworthy in Fig. 1 is that many of the ice-core climate records of the Younger Dryas exhibit a weak trend. Colder, lower-accumulation, dustier conditions occurred near the onset of the event, but became slightly warmer, wetter, and less dusty before the large, step change at the end of the Younger Dryas.

9. Synthesis

A full review of the Younger Dryas is beyond the scope of this paper. Broecker and Denton (1989) and Peteet (1993) provide introductions, and some material is covered by Alley and Clark (1999). A brief overview is that, within the dating uncertainties, the Younger Dryas involved:

- reduction or southward shift in sites of formation of North Atlantic deepwater formation, and probably reduction in cross-equatorial flow of warm surface waters to the North Atlantic (Broecker et al., 1988; Broecker et al., 1990; Lehman and Keigwin, 1992; Keigwin and Lehman, 1994);
- reduction in African and Asian monsoon strength, causing drying (Street-Perrott and Perrott, 1990; Gasse and Van Campo, 1994; Schulz et al., 1998);
- cooling centered on the North Atlantic but extending well beyond, and possibly encompassing most of the globe (Denton and Hendy, 1994, but cf. Singer, Shulmeister and McLea, 1998; Steig et al., 1998); but
- warming centered on the South Atlantic and including much, but not all, of Antarctica (Bender et al., 1994; Blunier et al., 1998; Alley and Clark, 1999).

A plausible story is that the Younger Dryas began when something (likely, although not certainly, diversion of meltwaters from the Laurentide ice sheet of North America from drainage southward in the Mississippi River to drainage eastward in the St. Lawrence River; Broecker et al., 1988) caused the density of North Atlantic surface waters to decrease. This caused a reduction in sinking of waters in the North Atlantic and cross-equatorial flow to feed that sinking, producing high-latitude cooling but leaving more heat in the South Atlantic to produce the Antarctic warming (Stocker et al., 1992; Crowley, 1992). Atmospheric effects of the colder North Atlantic included reduced monsoon strength, and steeper temperature gradients causing stronger winds (e.g., Fawcett et al., 1997; Ágústsdóttir et al., 1999). Feedbacks associated with the monsoonal changes (Hostetler et al., 1999) and changes in wind-driven oceanic mixing (Hughen et al., 1996; Hughen et al., 1998) then transmitted the signal farther into the southern hemisphere. Warming then may have occurred after re-diversion of meltwater drainage allowed density increase of North Atlantic surface waters, leading to abrupt re-initiation of sinking.

Atmospheric modeling with reduced North Atlantic oceanic heat transport to high-latitude seas produces anomaly patterns that are in many ways similar to those observed for the Younger Dryas (e.g., Rind et al., 1986; Fawcett et al., 1997; Renssen, 1997; Ágústsdóttir, 1998; Ágústsdóttir et al., 1999). For central Greenland, Fawcett et al. (1997) found that cooling the Nordic Seas with Younger Dryas boundary conditions produced a strengthening and southward shift in storm tracks such that central Greenland became less stormy, and a cooling and drying that were amplified in the winter, in excellent agreement with ice-core observations. Model-data matches across much of the northern hemisphere are also good.

However, observed anomalies far from the North Atlantic often are somewhat larger than expected based on modeled atmospheric response to North Atlantic cooling. Possible explanations include model weaknesses, forcing in addition to North Atlantic cooling, or even data weaknesses. A possible (partial?) explanation is suggested by the observations in the Cariaco Basin (Hughen et al., 1996), and the model results of Ágústsdóttir et al. (1999) and Hostetler et al. (1999). The Cariaco Basin records show enhanced trade-wind-driven upwelling during the Younger Dryas interval, and Ágústsdóttir et al. (1999) modeled this response in the Cariaco Basin and other subtropical and tropical regions owing to imposed North Atlantic cooling. Upwelled tropical and subtropical waters are colder than surface waters, so enhanced upwelling would tend to cool the overlying atmosphere, with water vapor and other feedbacks amplifying the cooling. Hostetler et al. (1999) show that northern cooling (modeled for Heinrich events) affects cross-equatorial flow in monsoonal regions, also tending to transmit signals farther from the North Atlantic. However, it is not yet clear whether a North Atlantic cooling by itself can explain the full observed anomaly pattern of the Younger Dryas.

The slow moderation of Greenland conditions during the Younger Dryas is suggestive of an ocean–atmosphere system that operates in broad modes or bands separated by gaps (cf. Alley and Clark, 1999). Such behavior is shown diagrammatically by Stocker and Wright (1996). Many explanations are possible. Perhaps the slow moderation of Greenland conditions during the Younger Dryas represents slow increase in intermediate-water formation in the mid-latitude North Atlantic in response to salt build-up in the low-latitude Atlantic, prior to the abrupt initiation of high-latitude North Atlantic deepwater formation, perhaps in response to re-diversion of Laurentide meltwater away from the St. Lawrence (Broecker et al., 1988).

Although the Younger Dryas is plausibly related to meltwater diversion on North America (Broecker et al., 1988), as noted above, the Younger Dryas is one of many similar abrupt climate changes (e.g., Johnsen et al., 1992; Mayewski et al., 1997), and it is difficult to imagine that all have similar causes (reviewed by Alley and Clark, 1999). Learning the relation of the Younger Dryas to these other events will be important in gaining a predictive understanding of the climate system. Forced oscillations, free oscillations of the climate system, or some intermediate behavior such as stochastic resonance (cf. Gammaitoni et al., 1998) may be involved; much exciting research remains to be done.

10. Summary

Understanding abrupt climate changes is of considerable interest. Some models of future greenhouse warming produce Younger-Dryas-like reductions in North Atlantic oceanic overturning owing to increased precipitation or ice melt at high northern latitudes (e.g., Stocker and Schmittner, 1997). The modeled response to a Younger Dryas reduction in ocean heat transport can be compared to reconstructed paleoclimatic anomalies to learn how well a model performs, and thus to place likely error limits on the climate anomalies a climate model projects for possible great reductions in North Atlantic overturning in the future.

Such comparisons require high-time-resolution, well-calibrated proxy records from widespread regions of the Earth, including marine, lacustrine, tropical-ice, and terrestrial records. Polar ice cores can yield subannually resolved records over tens of thousands of years. In favorable places on ice sheets, site-specific calibrations allow confident reconstruction of snow accumulation, surface temperature, and atmospheric composition, together with reasonably confident reconstruction of atmospheric concentrations of aerosols. Together with confident reconstruction of past atmospheric composition from trapped bubbles, these provide an excellent history of climate. Because wind-blown materials originated beyond the ice sheets, and methane and perhaps other trace gases record widespread processes, the ice cores allow determination of the synchrony of events over large regions with much better time resolution than is possible by direct comparison of different records, although without detailed information on just which regions were involved. The ice cores show that the end of the Younger Dryas interval involved: 5–10°C warming and a doubling of snow accumulation in central Greenland; a large drop in wind-blown materials, indicating reduced wind speed and other changes in distant source regions or between source regions and Greenland; and a large increase in methane, indicating expansion of global wetlands, probably including those of the tropics. Most of these changes occurred in less than a few decades, and possibly in less than a few years.

A simple picture emerging from these and other data is that the "normal" climate experienced by agricultural and industrial humans has been more stable in many or most regions than is typical of the climate system. Large, rapid, widespread changes were common in the pre-agricultural past, especially in regions near the North Atlantic, but apparently also in monsoonal regions affected by the North Atlantic, and likely elsewhere or even globally. Critically, the typically smaller (although still quite significant!) climate changes experienced by agricultural and industrial humans have had dramatic impacts on many of them (e.g., Thompson et al., 1988; Barlow et al., 1997; Sandweiss et al., 1999). Recurrence of a larger Younger Dryas type event is not impossible, and this possibility merits careful study.

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