17. An animation of the dike can be viewed at http://pangea.stanford.edu/~segall/izu.html.
18. The centroid depth is the average depth weighted by the magnitude of the dike opening.
19. D. D. Pollard, P. Delaney, W. A. Duffield, E. T. Endo, A. T. Okamura, *Tectonophysics* **94**, 541 (1983).
20. Y. Okada, E. Yamamoto, T. Ohkubo, in preparation.
21. A. M. Rubin and D. Gillard, *J. Geophys. Res.* **103**, 10017 (1998).
22. D. Gillard, A. M. Rubin, P. Okubo, *Nature* **384**, 343 (1996); D. P. Schaff, G. Beroza, B. E. Shaw, *Geophys. Res. Lett.* **25**, 4549 (1998).
23. We thank the Geographical Survey Institute of Japan and the National Research Institute for Earth Science and Disaster Prevention of Japan for access to data and P. Delaney, Y. Okada, B. Ellsworth, G. Beroza, T. Dixon, and an anonymous reviewer for helpful comments. Supported by a grant from NASA's Solid Earth and Natural Hazards program to Stanford University.

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Abrupt Climate Change at the End of the Last Glacial Period Inferred from Trapped Air in Polar Ice

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The last glacial period was terminated by an abrupt warming event in the North Atlantic ~15,000 years before the present, and warming events of similar age have been reported from low latitudes. Understanding the mechanism of this termination requires that the precise relative timing of abrupt climate warming in the tropics versus the North Atlantic be known. Nitrogen and argon isotopes in trapped air in Greenland ice show that the Greenland Summit warmed 9 ± 3°C over a period of several decades, beginning 14,672 years ago. Atmospheric methane concentrations rose abruptly over a ~50-year period and began their increase 20 to 30 years after the onset of the abrupt Greenland warming. These data suggest that tropical climate became warmer or wetter (or both) ~20 to 80 years after the onset of Greenland warming, supporting a North Atlantic rather than a tropical trigger for the climate event.

Evidence for extremely abrupt changes in Earth's climate has come principally from the annually layered Greenland ice cores (*1*–*7*), although the tropics collect most solar radiation and are central to Earth's heat and water vapor budgets (*8*). Accordingly, tropical climate records are critical for understanding abrupt climate events. Especially important are those records that reveal cause-and-effect relations by giving the precise relative timing of changes in the tropics and the high latitudes. However, high-resolution tropical climate records are few in number and for the most part equivocal on the existence and timing of abrupt climate events.

Evidence from Bolivian ice cores at 18°S shows that tropical climates warmed rapidly about 15 thousand years (ky) B.P. (before present, where present is 1950 A.D.) (*9*). This warming is roughly synchronous with prominent warming at 14.67 ky B.P., known as the Bølling Transition, seen in the Greenland Summit ice cores (*1*, *2*, *5*), European pollen records (*10*), and many North Atlantic sediment records (*11*, *12*). Other recent studies have documented low-latitude rapid climate shifts at about this time in the Santa Barbara basin (*13*), Arabian Sea (*14*), and the Cariaco basin of Venezuela (*15*); however, dating uncertainties of a century or more limit the precision with which the timing of these events and their putative Arctic counterpart may be compared.

Atmospheric methane concentrations also increased abruptly at this time, as inferred from measurements of trapped air bubbles in the Greenland Summit ice cores (*16*, *17*) and Antarctic ice cores (*18*, *19*). This increase is thought to have been caused by an increase in wetland extent and temperature, as wetlands were the principal source of methane in the preindustrial period (*17*) and changes in sink strength are thought to have been small (*20*). Tropical wetlands have been proposed as a major contributor to the Bølling methane increase, because ice sheets covered the primary extratropical methane source areas at this time (*17*, *21*). This hypothesis is supported by the fact that the difference between Greenland and Antarctic methane concentrations (referred to as the interpolar gradient) underwent a relatively small increase despite the ~25% jump in concentration (*21*, *22*). When inverted for source distribution with a three-box atmospheric transport model, methane records from Greenland and Antarctica imply a predominantly low-latitude source (*23*). Because the atmosphere is well mixed (mixing time of ~1 year) relative to methane's lifetime in the atmosphere (~10 years) (*24*) and wetland methane emissions are broadly correlated with temperature and precipitation, methane concentration integrates climate information over the entire tropical region. Knowledge of the precise timing of the abrupt methane increase relative to the Greenland warming should thus place constraints on the relative timing of tropical and Arctic abrupt climate change.

Two problems have prevented relating changes in temperature and methane in the ice core record. First, the age of the air trapped in bubbles is less than the age of the enclosing ice because air is occluded at some depth below the surface of the ice sheet in the bubble close-off region (*25*, *26*). This gas age–ice age difference is not known well for past times, making uncertain the relative timing of temperature (which is obtained from the ice matrix) and atmospheric gas changes (*17*, *26*). Second, the temperature inferred from the $^{18}O/^{16}O$ ratio of the ice ($\delta^{18}O_{ice}$) is uncertain because factors other than local mean annual temperature may affect this ratio, such as the seasonality of precipitation at the ice core site (*27*) and the temperature and proximity of the water vapor source (*28*, *29*). Borehole temperature calibrations of the $\delta^{18}O_{ice}$ paleothermometer have demonstrated that the modern spatial calibration underestimates the glacial cooling by a factor of two (*30*). These issues have raised the question of whether the abrupt increases in Greenland $\delta^{18}O_{ice}$ represent isotopic artifacts rather than local temperature changes as commonly inferred (*29*, *31*) and have created uncertainty about the magnitude of temperature change they may indicate.

Here we address these problems with measurements of isotopes of nitrogen and argon gas trapped in air bubbles in the GISP2 (Greenland Ice Sheet Project 2) ice core. Bubble $^{15}N/^{14}N$ and $^{40}Ar/^{36}Ar$ record a signal of rapid temperature change at the surface of the ice sheet (*7*). The combination of nitrogen and argon isotope measurements provides a direct estimate of the magnitude of the temperature increase, based on laboratory calibration of the isotope fractionation due to thermal diffusion. Methane and nitrogen diffuse downward through the snow layer at nearly the same speed (*32*) and are trapped together in the bubbles, making a precise comparison of the timing of atmospheric methane change and local temperature change possible (*7*).

The porous and partially consolidated layer of snow on top of polar ice sheets (known as firn) contains air that mixes by molecular diffusion with the overlying atmosphere (*25*). The isotopes of gases in this stagnant column of air

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separate as a result of temperature gradients and gravitational settling (*7*, *33*) (air in the free troposphere does not separate in this manner because of vigorous convective mixing). A gas mixture subjected to a temperature gradient will unmix slightly, with heavier gases generally moving toward colder regions (*34*, *35*). The magnitude of the steady state effect is given by $\delta = \Omega \Delta T$, where δ is the fractional deviation of the isotope ratio of the cold gas parcel from the warm gas parcel, ΔT is the temperature difference between the two gas parcels, and Ω is an empirical quasi-constant we call the "thermal diffusion sensitivity" (*36*). We measured Ω in a known temperature gradient in the laboratory and found preliminary values of $\Omega = +0.0145$ per mil K^{-1} for $\delta^{15}N$ and $+0.036$ per mil K^{-1} for $\delta^{40}Ar$ at a mean temperature of $-43°C$ (estimated error of $\pm 3\%$; future work may reduce this error). For example, a 10 K temperature difference will result in a $+0.145$ per mil enrichment in $\delta^{15}N$ in the cold gas relative to the warm gas. For comparison, our 1σ measurement precision (defined as the reproducibility of replicate analyses of ice cut from the same depth) is ± 0.007 per mil for $\delta^{15}N$ and ± 0.03 per mil for $\delta^{40}Ar$, corresponding to ± 0.5 and ± 1 K, respectively.

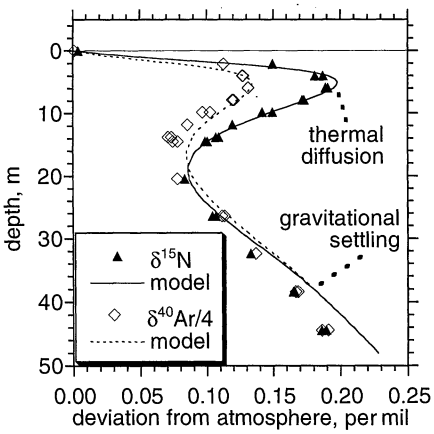

Fig. 1. Depth profile of nitrogen and argon isotope ratios in air withdrawn from the snowpack at Siple Dome, Antarctica, on 14 January 1998, $\delta^{15}N$ is defined as $[(^{15}N/^{14}N_{sample})/(^{15}N/^{14}N_{atmosphere}) - 1] \times 10^3$ and is expressed in units of per mil. $\delta^{40}Ar$ is the corresponding quantity for the $^{40}Ar/^{36}Ar$ ratio and is divided by 4 to facilitate comparison with $\delta^{15}N$. The seasonal cycle of temperature at the surface creates a strong temperature gradient in the top few meters, causing thermal fractionation. Because the profile shown was taken in summer, temperature decreases with depth. Thermal diffusion thus drives a transient "pulse" of heavy isotopes down into the colder firn, which mixes by diffusion with the underlying gas. Note the "pulse" of enriched $\delta^{15}N$ and $\delta^{40}Ar/4$ centered at about 6-m depth, superimposed on the linear downward increase due to gravitational settling. The good agreement of $\delta^{15}N$ with $\delta^{40}Ar/4$ at depths >20 m confirms that gravity is the only process affecting the deep firn at present at this site, whereas the weaker $\delta^{40}Ar/4$ response in the "pulse" is the fingerprint of thermal diffusion.

Gravitational settling (*33*) is given by $\delta = [e^{\Delta mgz/RT} - 1] \times 10^3$, where Δm is mass difference, g is gravitational acceleration, z is depth, R is the gas constant, and T is temperature. For example, a 92-m-deep stagnant air column at 226 K will be enriched by $+0.48$ per mil in $\delta^{15}N$ relative to the top of the column (in our case, the atmosphere). Gravitational effects may be separated from thermal effects by comparing $\delta^{15}N$ with $\delta^{40}Ar/4$, because the latter is only ~60% as sensitive to thermal fractionation as $\delta^{15}N$. Gravitational settling scales with the mass difference (4 for $^{40}Ar/^{36}Ar$), so gravity affects $\delta^{15}N$ and $\delta^{40}Ar/4$ equally (*33*). Measurements in modern firn air confirm the different sensitivities of nitrogen and argon isotopes to thermal diffusion (Fig. 1).

After an abrupt climate warming, a temperature gradient will persist in the firn for several hundred years (the thermal equilibration time of the firn) and will thermally fractionate the entire air column (*7*). Gases diffuse about 10 times as fast as heat in polar firn (*37*), so the isotopic signal penetrates to the bottom of the firn long before the temperature equilibrates, in about a decade (*7*). The bubbles thus record a signal of the climate event as an abrupt increase in $\delta^{15}N$, slightly smoothed by the diffusion process in the firn (*38*), followed by a gradual decrease in $\delta^{15}N$ over several hundred years as the firn becomes isothermal once again.

Two prominent abrupt increases in GISP2 $\delta^{18}O_{ice}$ occurred during the last deglaciation, at 14.67 and 11.64 ky B.P. on the layer-counted time scale, and were accompanied by abrupt methane increases and $\delta^{15}N$ excursions (Fig. 2). $\delta^{15}N$ may be thought of as a temperature-change indicator that is only sensitive to rapid temperature change (*7*). The $\delta^{15}N$ signal thus confirms that the $\delta^{18}O_{ice}$ increases were indeed abrupt warming events and not isotopic artifacts. The event at 14.67 ky B.P. separates the last glacial period from the Bølling warm interval and is known as the Bølling Transition (*5*) (in the ice core record, this interval is also referred to as Interstadial 1). The cold period immediately preceding the transition is referred to as the Oldest Dryas (*5*). The gas time scale presented in Fig. 2 is approximate and is based on model estimates of the gas age–ice age difference that may be in error by several hundred years (*39*), so the slight apparent lead of $\delta^{15}N$ over $\delta^{18}O_{ice}$ is not significant. A detailed expansion of the record from 14.5 to 14.9 ky B.P., covering the Bølling Transition, is shown in Fig. 3. In contrast to Fig. 2, the gas ages shown are deduced on the basis of our inference of a thermal diffusion signature in the gas isotopes (*40*). Individual replicate analyses of $\delta^{15}N$ and $\delta^{40}Ar/4$ in pieces of ice cut from the same depth in the ice core are shown. Means of replicate methane analyses are given. Analytical techniques for gas isotopes and methane are described elsewhere (*41*).

$\delta^{15}N$ rises across the transition from baseline Oldest Dryas values of $+0.48$ per mil to peak values of $+0.63$ per mil (Fig. 3). Methane appears to rise over a ~50-year period from 14.65 to 14.60 ky B.P. $\delta^{15}N$ first rises clearly beyond the envelope of Oldest Dryas variability at 14.672 ky B.P. (1821.16-m depth). We infer that this point marks the onset of the warming

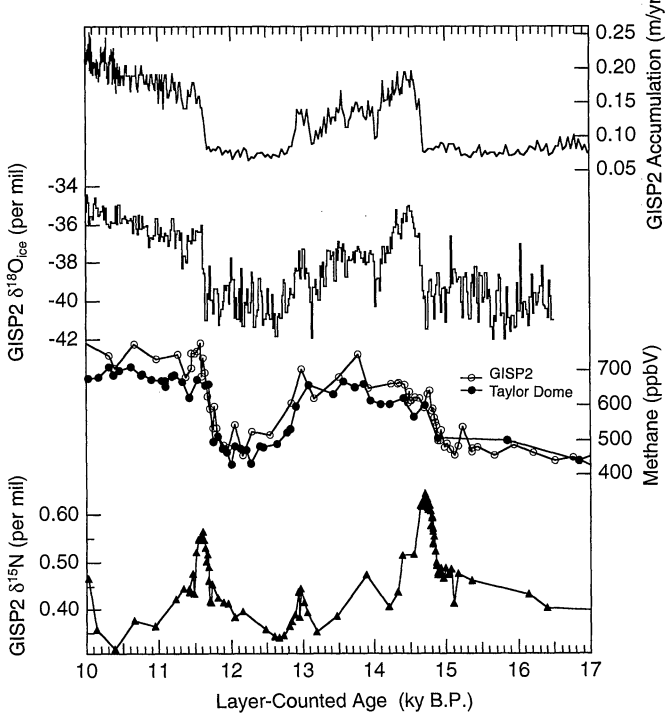

Fig. 2. GISP2 accumulation (*51*), oxygen isotopes (*5*), methane, and nitrogen isotopes of air in bubbles during the last deglaciation, with Taylor Dome (Antarctica) methane shown for comparison (*21*). Low-resolution nitrogen data are from (*54*).

and may correspond to the 1.7 per mil steplike increase in $\delta^{18}O_{ice}$ (Fig. 3) and the abrupt doubling of snow accumulation (*5*) at this time. However, methane concentration remains within its range of Oldest Dryas variability [470 to 510 parts per billion by volume (ppbv)] in this ice sample and in the subsequent sample at 14.655 ky B.P. (Table 1). Methane first exceeds this range at 14.645 ky B.P. (1820.17 m), ~27 years after the corresponding change in $\delta^{15}N$. We infer that methane began rising at some time during the 10-year interval between the latter two samples and thus conclude that atmospheric methane concentration started rising ~20 to 30 years after the onset of the Greenland warming. Annual layers are discernable in this interval (*3*), making these estimates robust. Absolute time scale uncertainty (~2%) (*42*) does not affect these conclusions, because they are based on comparisons of gas isotopes and methane from the same depths.

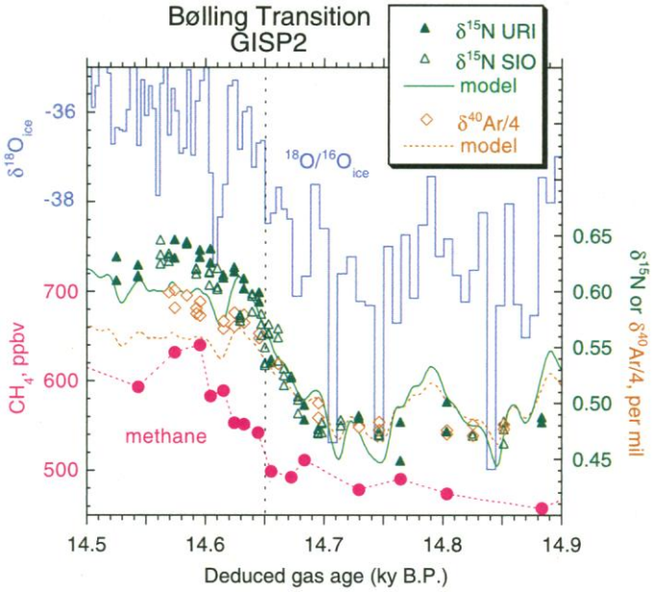

Fig. 3. The 400-year window encompassing the Bølling Transition, showing high-resolution measurements of oxygen isotopes of ice (*5*) and nitrogen, argon, and methane measurements made on trapped air bubbles. The gas age is deduced by assuming that the abrupt change in nitrogen isotopes marked by the dashed vertical line corresponds to the shift in oxygen isotopes at 14.65 ky B.P. Nitrogen isotope measurements made at URI and the Scripps Institution of Oceanography (SIO) appear to have a ~0.02 per mil offset in the age range 14.59 to 14.64 ky B.P., which we do not understand. The mismatch between model and observed gas isotopes in the range 14.55 to 14.62 ky B.P. is likely due to gravitational enrichment from transient firn thickening brought about by the increase in snow accumulation, which was not included in the model.

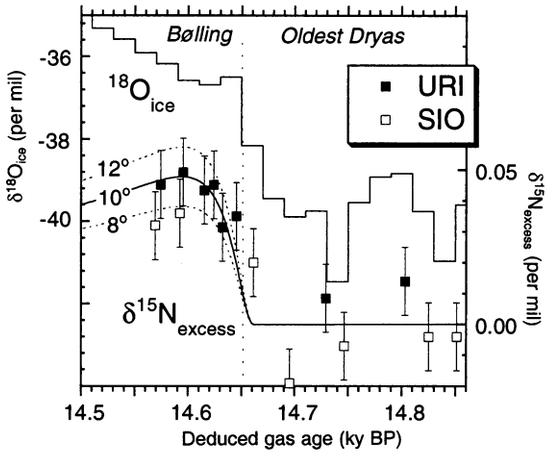

Fig. 4. Plot of measured $\delta^{15}N_{excess} = \delta^{15}N - \delta^{40}Ar/4$, which is not affected by gravitational settling, compared with model results forced by 8°, 10°, and 12°C "step function" increases in surface temperature. Error bars show quadrature sum of pooled RMS deviations of nitrogen and argon replicate measurements. Measurements made at URI and SIO appear to have a ~0.01 per mil systematic offset, equivalent to 2°C, possibly due to a calibration error. We adopt an estimate of 9 ± 3°C for the magnitude of the abrupt warming.

Table 1. Depths and deduced gas ages of inferred Bølling Transition.

Mid depth (m)	Deduced gas age (ky B.P.)	$\delta^{15}N$* (per mil)	CH_4† (ppbv)	Layer counting§ age of ice (ky B.P.)	Deduced Δage (years)
1821.16	14.672	0.524 ± 0.001	492 ± 11		
1820.55	14.655	0.538 ± 0.002	499 ± 11		
1820.36	14.650			15.513	863
1820.17	14.645	0.596 ± 0.006	542 ± 15		

*Mean and standard deviation of duplicate analyses of separate pieces of ice from the same depth. †Mean and standard deviation of replicate analyses of separate pieces of ice from the same depth (n = 4 for 1821.16 m and 1820.55 m and n = 3 for 1820.17 m). §See (*42*).

The separation between $\delta^{15}N$ and $\delta^{40}Ar/4$ values in the Bølling interval (Fig. 3) confirms that thermal diffusion rather than gravity is responsible for the $\delta^{15}N$ anomaly and implies that air temperatures warmed by 9 ± 3°C by 14.60 ky B.P. (Fig. 4). $\delta^{18}O_{ice}$ increased by 3.4 per mil across the transition (difference between the means of the 60-year period after 14.65 ky B.P. and the 100-year period before 14.67 ky B.P.) (*5*), implying an oxygen isotope temperature sensitivity (α) across the transition of 0.38 per mil K^{-1} (+0.19 or −0.09 per mil K^{-1}). This is close to the borehole temperature-calibrated value of 0.33 per mil K^{-1} (*30*), providing independent verification of the borehole calibration.

Climate-driven changes in methane emissions could have preceded the observed atmospheric methane concentration change by no more than a decade, the approximate lifetime of methane in the atmosphere (*24*). Interpretation of the methane rise is complicated by evidence for two methane source regions during the transition. Part of the increase in methane concentration from the Last Glacial Maximum to the Bølling has been attributed to boreal sources (*21–23*). During the ~50-year interval in which methane concentrations were rising, we cannot know the interpolar gradient because of time scale uncertainties (arising from the fact that the rapid transition in methane concentration is used to synchronize Antarctica and Greenland ice core time scales). Thus, no source attributions based on the interpolar gradient are possible within this period. For example, the abrupt increase in methane concentration between 14.655 and 14.645 ky B.P. (Fig. 3) cannot be firmly attributed to a tropical climate event, because it is possible that this was the result of the boreal portion of the source increase. However, some part of the entire methane concentration increase from 14.655 to 14.595 ky B.P. must have been tropical in origin; otherwise, the interpolar gradient would have been steeper than observed just after 14.595 ky B.P. (Fig. 2) (*21–23*, *43*). Thus, we infer that there must have been a change in tropical temperatures or precipitation (or both) at some time within the interval 14.655 to 14.595 ky B.P. This brackets the possible time of the tropical climate event to within ~20 to 80 years after the onset of the Greenland warming.

Methane production might not respond instantaneously to environmental change, however. For example, it might take several decades for water tables to rise and fill previously drained wetlands, or sustained organic matter production might be required to

create the anoxia necessary for methane production. At least some immediate response of methane production to increased tropical temperature or precipitation is expected and probably inescapable, though. Soils become waterlogged and plant growth responds to precipitation on subannual time scales, and methane emissions from modern wetlands respond to changes in climate conditions on similar time scales (*44*, *21*). Thus, we take the absence of any substantial methane increase before 14.65 ky B.P. as an indication that the tropical climate change probably lagged Greenland warming by at least ~20 years.

The implication that high-latitude North Atlantic warming led tropical climate by several decades may provide a test for proposed mechanisms of abrupt climate events. Two alternative mechanisms may be imagined. In one scenario (the "North Atlantic trigger"), North Atlantic thermohaline overturning abruptly deepened and intensified, drawing heat poleward and warming Greenland (*45*), which warmed the tropics by some as yet poorly understood mechanism. Increased poleward heat transport might have reduced the Northern Hemisphere equator-to-pole temperature gradient, decreasing the driving force for the winds (*46*). Slackening of the winds would have reduced oceanic upwelling, wind mixing, evaporative cooling, and the gyre circulation, leading to an abrupt warming of tropical and subtropical sea surface temperature (SST). Much evidence points to a sudden drop in wind speed at the time of the Bølling Transition, including records from laminated sediments in the Cariaco basin of Venezuela (*15*) and dust particle size and concentration in GISP2 ice (*47*). Also, warmer North Atlantic SST has been shown in atmospheric model studies to produce strengthening of the Asian monsoon (*27*, *48*), which may have warmed the tropics. In a second, speculative scenario (the "tropical trigger"), the tropics warmed first, perhaps because the temperature of upwelling water in the tropics crossed a threshold that caused the tropical ocean-atmosphere system to abruptly reorganize into a warmer state. In this scenario, the deepening of North Atlantic thermohaline overturning and warming of Greenland would be responses to tropical warming rather than primary causes. Our phasing data would appear to favor a North Atlantic trigger.

References and Notes

1. W. Dansgaard, J. W. C. White, S. J. Johnsen, *Nature* **339**, 532 (1989); S. J. Johnsen *et al.*, *Nature* **359**, 311 (1992).
2. P. M. Grootes *et al.*, *Nature* **366**, 552 (1993).
3. R. B. Alley *et al.*, *Nature* **362**, 527 (1993).
4. K. C. Taylor *et al.*, *Nature* **361**, 432 (1993).
5. M. Stuiver, P. M. Grootes, T. F. Braziunas, *Quat. Res.* **44**, 341 (1995).
6. K. C. Taylor *et al.*, *Science* **278**, 825 (1997).
7. J. P. Severinghaus *et al.*, *Nature* **391**, 141 (1998).
8. T. J. Crowley and G. R. North, *Paleoclimatology* (Oxford Univ. Press, New York, 1991).
9. L. G. Thompson *et al.*, *Science* **282**, 1858 (1998).
10. J. L. Debeaulieu and M. Reille, *Quat. Sci. Rev.* **11**, 431 (1992).
11. N. Koc and E. Jansen, *Geology* **22**, 523 (1994).
12. G. C. Bond and R. Lotti, *Science* **267**, 1005 (1995).
13. R. J. Behl and J. P. Kennett, *Nature* **379**, 243 (1996).
14. H. Schulz, U. vonRad, H. Erlenkeuser, *Nature* **393**, 54 (1998).
15. K. A. Hughen *et al.*, *Nature* **380**, 51 (1996).
16. E. J. Brook, T. Sowers, J. Orchardo, *Science* **273**, 1087 (1996).
17. J. Chappellaz *et al.*, *Nature* **366**, 443 (1993).
18. E. J. Steig *et al.*, *Science* **282**, 92 (1998).
19. T. Blunier *et al.*, *Geophys. Res. Lett.* **24**, 2683 (1997).
20. P. Martinerie, G. P. Brasseur, C. Granier, *J. Geophys. Res.* **100**, 14291 (1995).
21. E. J. Brook, J. P. Severinghaus, S. Harder, M. Bender, in *Mechanisms of Global Climate Change at Millennial Time Scales*, *Geophysical Monograph Ser.*, vol. 112 (American Geophysical Union, Washington, DC, 1999), pp. 165–176.
22. A. Dallenbach *et al.*, *Geophys. Res. Lett.*, in press.
23. Inversions of interpolar gradient data indicate that the methane source was ~70% tropical after the Bølling Transition (*21*, *22*). The data from Dallenbach *et al.* (*22*) most closely constrain the change in methane source at the Bølling Transition. Their results indicate that tropical sources increased by ~40 Tg year^{-1} and boreal sources increased by ~30 Tg year^{-1}. The total methane source immediately after the Bølling Transition was ~175 to 185 Tg year^{-1} (*21*, *22*). Brook *et al.* (*21*) reported a larger increase in tropical source at this time, based on data that indicate a greater contribution from boreal sources before the Bølling Transition. Those data, from the Taylor Dome ice core, did not include the time period immediately before the Bølling warming (the Older Dryas), where it appears that the contribution from boreal sources was lower.
24. R. G. Prinn *et al.*, *Science* **269**, 187 (1995).
25. J. Schwander, B. Stauffer, A. Sigg, *Ann. Glaciol.* **10**, 141 (1988).
26. T. A. Sowers, M. L. Bender, D. Raynaud, *J. Geophys. Res.* **94**, 5137 (1989).
27. P. J. Fawcett, A. M. Ágústsdóttir, R. B. Alley, C. A. Shuman, *Paleoceanography* **12**, 23 (1997).
28. E. A. Boyle, *Geophys. Res. Lett.* **24**, 273 (1997).
29. C. D. Charles, D. Rind, J. Jouzel, R. D. Koster, R. G. Fairbanks, *Science* **263**, 508 (1994).
30. K. M. Cuffey *et al.*, *Science* **270**, 455 (1995); D. Dahl-Jensen *et al.*, *Science* **282**, 268 (1998).
31. R. G. Fairbanks, *Paleoceanography* **5**, 937 (1990).
32. CH_4 diffuses 7% faster than $^{15}N/^{14}N$, so the methane signal will traverse the firn in about 1 year less than the nitrogen signal, but we neglect this difference.
33. F. A. Lindemann and F. W. Aston, *Philos. Mag. Ser. 6* **37**, 523 (1919); H. Craig, Y. Horibe, T. Sowers, *Science* **242**, 1675 (1988); J. Schwander, in *The Environmental Record in Glaciers and Ice Sheets*, H. Oeschger and C. C. Langway Jr., Eds. (Wiley, New York, 1989), pp. 53–67.
34. S. Chapman and F. W. Dootson, *Philos. Mag.* **33**, 248 (1917).
35. K. E. Grew and T. L. Ibbs, *Thermal Diffusion in Gases* (Cambridge Univ. Press, Cambridge, 1952).
36. In the physical chemistry literature, a slightly different parameter is used to describe the magnitude of the thermal diffusion effect, called the "thermal diffusion factor" and denoted α_T, defined by $\alpha_T = \ln(R_c/R_h)/\ln(T_h/T_c)$, where R_c and T_c are the isotope ratio and absolute temperature, respectively, of the cold gas parcel and R_h and T_h are the isotope ratio and absolute temperature, respectively, of the hot gas parcel (*35*). For convenience, we chose to use Ω rather than α_T because the latter is very temperature dependent and, fortuitously, the former is nearly independent of temperature in the temperature range of interest for ice core studies. To convert α_T to Ω, one may use the relation Ω (per mil K^{-1}) $\cong \alpha_T \times 10^3/T_{average}$, where $T_{average} = T_hT_c\ln(T_h/T_c)/(T_h-T_c)$.
37. W. S. B. Paterson, *The Physics of Glaciers* (Pergamon, Tarrytown, NY, 1969).
38. The amount of smoothing can be calculated from the age distribution of the gases in the firn, which has a width of about a decade (*25*). Thus, the smoothing imparted by diffusion in the firn has roughly the same impact as a 10-year running average. Additional smoothing may result from the fact that the bubbles do not all close off at the same time, but the importance of this effect in GISP2 is currently unknown.
39. J. Schwander *et al.*, *J. Geophys. Res.* **102**, 19483 (1997).
40. Note the abrupt increase in $\delta^{15}N$ [University of Rhode Island (URI) values] from +0.54 to +0.60 per mil at 14.65 ky B.P. that separates an earlier set of lower values from a later set of higher values. To establish the gas chronology presented in Fig. 3, we assumed that this abrupt change in $\delta^{15}N$ corresponds to the steplike change in $\delta^{18}O_{ice}$ at 14.65 ky B.P. and used this marker as a tie point. To extend gas ages to surrounding data, we used the model-predicted gas age–ice age differences of Schwander *et al.* (*39*), adjusted to be consistent with the assumption that the major $\delta^{15}N$ shift occurred at 14.65 ky B.P. This was accomplished by adding 208 years to all model Δage values and 4.08 m to all model depth values. Finally, a firn heat and gas diffusion model was used to predict $\delta^{15}N$ and $\delta^{40}Ar/4$, in a forward-modeling sense, from the $\delta^{18}O_{ice}$ data (*49*). The good correspondence of certain details of the model with the $\delta^{15}N$ and $\delta^{40}Ar/4$ data reassures us that the deduced gas age is accurate (Fig. 3). For example, the small dip in model $\delta^{15}N$ and $\delta^{40}Ar/4$ at 14.61 ky B.P. appears in the data, as does the slight plateau in model $\delta^{15}N$ at 14.67 ky B.P. We emphasize that the uncertainty of the gas ages so assigned increases with distance from the tie point at 14.65 ky B.P. The main conclusion of this paper does not, however, depend on the accuracy of the gas age scale, as it comes from comparing $\delta^{15}N$ and CH_4 in samples from the same depth.
41. The analytical procedure for ice core $\delta^{15}N$ and $\delta^{40}Ar/4$ was as given in (*26*, *7*), except that $\delta^{40}Ar$ was measured by means of a wet extraction method in which the ice sample was melted and the gases were transferred to a water trap through a capillary, in effect using water vapor as a carrier gas. The melt was agitated with a glass-covered magnetic stirrer bar to enhance gas exchange and warmed to 40°C in a water bath during the 15-min extraction. The dried gas was frozen cryogenically at 10 K in a vessel and then gettered as per the usual procedure (*7*). Comparisons of this wet extraction method with the melt-refreeze method followed in (*7*) showed no substantial difference between the methods. Another important procedural difference was that the $\delta^{15}N$ and $\delta^{40}Ar$ samples were cut in vertical 10-cm-long pieces, effectively averaging over five annual layers as each annual layer is about 2 cm thick. Replicate samples were cut from a position horizontally adjacent in the ice core. This was done because of the recognition that annual layering exists in ice core gases, which may explain the poorer reproducibility of $\delta^{15}N$ in (*7*), where the ice samples were cut without considering annual layering. The analytical procedure for methane was as given in (*16*). Methane points (Fig. 3) are the means of three or four replicate analyses of ice from the same depth in the core. Firn air was sampled from tubes placed in the firn 1 year earlier. Analyses (Fig. 1) of $\delta^{40}Ar$ were done by gettering 40 cm^3 standard temperature and pressure of air and then analyzing the residual noble gas. $\delta^{15}N$ was analyzed as in (*50*). Pooled root mean square (RMS) deviation of flask pairs from the same depth was 0.002 per mil for both $\delta^{15}N$ and $\delta^{40}Ar/4$.
42. D. A. Meese *et al.*, *J. Geophys. Res.* **102**, 26411 (1997).
43. One potential criticism of this argument arises from uncertainty in the chronology of the Taylor Dome ice core. The timing of the methane rise in Taylor Dome ice is independently dated to no better than ±300 years. Thus, the methane rise observed in Taylor Dome arguably could have occurred 300 years after the rise in GISP2, a scenario we dub the "Boreal burst." During the interval between the boreal and austral methane increases, the interpolar gradient would have been equal to the difference between Bølling and Oldest Dryas concentrations, or 25%. To maintain this exceptionally steep gradient, atmospheric transport must have been substantially reduced relative to

today or the austral sink radically enhanced, starting at the time of the boreal methane rise and fortuitously stopping at the time of the austral methane rise. Thus, we argue that the Boreal burst scenario is extremely unlikely.

44. S. Frolking and P. Crill, *Global Biogeochem. Cycles* **8**, 385 (1994).
45. W. S. Broecker and G. H. Denton, *Geochim. Cosmochim. Acta* **53**, 2465 (1989); S. Manabe and R. J. Stouffer, *Nature* **378**, 165 (1995); S. Rahmstorf, *Nature* **378**, 145 (1995).
46. A. M. Ágústsdóttir, R. B. Alley, D. Pollard, W. H. Peterson, *Geophys. Res. Lett.* **26**, 1333 (1999).
47. P. A. Mayewski *et al.*, *Science* **263**, 1747 (1994).
48. S. W. Hostetler, P. U. Clark, P. J. Bartlein, A. C. Mix, N. J. Pisias, *J. Geophys. Res.* **104**, 3947 (1999). We speculate that abrupt warming in the North Atlantic basin may have stimulated the Asian monsoon by the decrease in continental albedo associated with reduced snow cover and greening of desert in downwind Eurasia.

49. $\delta^{18}O_{ice}$ was converted to temperature for the purpose of forcing the nitrogen and argon isotope model by means of the relation T (degrees Celsius) = $(\delta^{18}O_{ice} - 1.3 \text{ per mil})/(0.384 \text{ per mil } K^{-1}) + 60$. The 1.3 per mil was subtracted from glacial $\delta^{18}O_{ice}$ to account for the glacial seawater $\delta^{18}O$ change due to ice volume. The coefficient 0.384 and constant of 60 were obtained by forcing the equation to predict a temperature of −47.5°C during the Oldest Dryas (*51*) and a mean for the last 1000 years of −31°C from respective $\delta^{18}O_{ice}$ values of −40 per mil and −35.04 per mil (*5*). The firn heat and gas model consists of a separate firn heat model adapted from (*52*) that drives a firn gas diffusion model. The gas model is based on the diffusion equation, modified to include gravitational settling and thermal diffusion:

$$\frac{\partial C}{\partial t} = \frac{\partial}{\partial z}\left\{ D(z,T)\left[\frac{\partial C}{\partial z} - \frac{\Delta m g}{RT} + \Omega \frac{dT}{dz}\right]\right\}$$

where C is isotope delta value (for example, $\delta^{15}N$), t is time, D is molecular diffusivity of gas in porous snow, and the other symbols are as defined in the text. The model is discretized on a 5-m grid, with a time step of 8 days, with diffusivities calculated from an empirical diffusivity-density relation based on observed CO_2 concentrations in South Pole firn air after (*50*). The density profile was modeled from (*53*). To assign gas ages, we assumed a mean gas age at the time of bubble closure of 15 years. The heat model was forced at the top boundary condition (that is, the surface) with temperature calculated from $\delta^{18}O_{ice}$. The bottom boundary of the heat model was held at 1000 m to account for the thermal inertia of the ice. The bottom boundary of the gas model is impermeable and held constant at 95-m depth, the firn depth inferred from the Herron-Langway model (*53*) with an accumulation rate of 0.07 m of ice per year and a temperature of −47.5°C (*51*).

50. M. Battle *et al.*, *Nature* **383**, 231 (1996).
51. K. Cuffey and G. Clow, *J. Geophys. Res.* **102**, 26383 (1997).
52. R. B. Alley and B. R. Koci, *Ann. Glaciol.* **14**, 6 (1990).
53. M. M. Herron and C. C. Langway, *J. Glaciol.* **25**, 373 (1980).
54. M. Bender *et al.*, *Nature* **372**, 663 (1994).

55. We thank M. Bender for helpful discussions and for hosting us in his laboratory at the University of Rhode Island, where many of these measurements were made; J. Orchardo for technical assistance; T. Ellis for the firn air $\delta^{15}N$ analyses; B. Luz for developing the $^{40}Ar/^{36}Ar$ technique for air samples; L. Tuttle and S. Cowburn for methane measurements at Washington State University; A. Grachev for the thermal diffusion sensitivity measurements; D. Mastroianni for the $\delta^{15}N$ measurements at SIO; and R. Alley, T. Sowers, R. Keeling, D. Jacob, L. Talley, and S. Harder for helpful discussions. This work was supported by NSF grants OPP 9725305 (J.P.S.) and OPP 9725918 (E.J.B.).

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16°C Rapid Temperature Variation in Central Greenland 70,000 Years Ago

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Variations in the $^{29}N_2/^{28}N_2$ ratio of air bubbles trapped in polar ice cores and their relation to variations of the $^{18}O/^{16}O$ of the ice allow past surface temperature variations and ice age–gas age differences to be determined. High-resolution measurements of $^{29}N_2/^{28}N_2$ in Dansgaard-Oeschger event 19 (around 70,000 years before the present) in ice from Central Greenland show that at the beginning of the event, the ice age–gas age difference was 1090 ± 100 years. With the use of a combined firn densification, temperature, and gas diffusion model, the $\delta^{18}O_{ice}$-temperature coefficient α was determined to be 0.42 ± 0.05 per mil per kelvin. This coefficient implies a mean surface temperature change of 16.0 kelvin (between 14.3 and 18.1 kelvin), which differs substantially from values derived from borehole temperatures and modern spatial $\delta^{18}O_{ice}$–surface temperature correlations.

Dansgaard-Oeschger (DO) events are characterized by rapid increases in the stable oxygen isotope composition of the ice [$\delta^{18}O_{ice}$ (*1*)] in ice cores of the last glacial period to high values that persist for several hundred to several thousand years. When these increases were discovered [in Camp Century ice (*2*)], it was not clear whether they reflected drastic changes in Arctic (or even Northern Hemispheric) climate, local effects, or stratigraphic disturbances in the core. Their climatic relevance was shown by their widespread geographic extent (*3*), and they are now interpreted as warmer interstadial periods. Several attempts have been made to estimate these temperature variations, with the contemporary surface temperature–$\delta^{18}O_{ice}$ correlation, borehole temperature measurements, and differences between gas age and ice age (Δage) based on $\delta^{18}O_{ice}$-methane and $\delta^{15}N$-$\delta^{18}O_{ice}$ correlations. The correlation of modern mean annual surface temperature and the mean $\delta^{18}O_{ice}$ of precipitation at various sites in Greenland (spatial correlation) leads to a coefficient, α (= $\Delta\delta^{18}O_{ice}/\Delta T$, where T is temperature), of 0.67 per mil/K (*4*). Deconvolutions of borehole temperatures show that α may vary over climatic cycles, with similar values during the Holocene (α = 0.53 to 0.67 per mil/K) and different values for the last glacial termination (α = 0.3 to 0.33 per mil/K) (*5*, *6*), although no analogous conclusions can be drawn for the relatively short DO events. Studying the GRIP (Greenland Ice Core Project) and GISP2 (Greenland Ice Sheet Project 2) methane and $\delta^{18}O_{ice}$ records, Schwander *et al.* suggested (on the basis of the calculated Δages) an α of 0.4 to 0.5 per mil/K for the DO events between 40,000 and 20,000 years before the present (yr B.P.) (*7*). Using the same approach with nitrogen isotope ratios, Severinghaus *et al.* (*8*) obtained a value of about 0.30 per mil/K for the transition at the end of the Younger Dryas cold period, in agreement with the borehole temperature studies.

Here we present a reconstruction of the surface temperature change in Central Greenland during DO 19 (70,000 yr BP), based on high-resolution measurements of nitrogen isotopes on the GRIP ice core. This reconstruction provides us with an estimate of the relation between temperature and $\delta^{18}O_{ice}$ for a DO event. We chose DO 19 because it is one of the larger events during the glacial, with a $\delta^{18}O_{ice}$ change of +6.7 per mil (uncertainty of 0.05 per mil). If the present-day spatial relation (α = 0.67 per mil/K) is applied to the $\delta^{18}O_{ice}$ change, a temperature shift of 10°C is expected. With the glacial-Holocene borehole sensitivity (0.3 to 0.33 per mil/K), this $\delta^{18}O_{ice}$ change would correspond to a far larger 20° to 22°C temperature shift.

Because the isotope ratio of atmospheric nitrogen ($\delta^{15}N$) is constant over several hundred thousand years (*9*), changes of this ratio in the air trapped in the ice reflect processes that occur in the firn. These are gravitational enrichment of the heavier molecules at the bottom of the firn column, which is proportional to the height of the diffusive firn column (*10*, *11*), and thermal diffusion effects due to temperature gradients between the surface and the bottom of the firn column, which force the heavier molecules generally to migrate toward the colder end of the firn column. The fractional deviation of $\delta^{15}N$ due to thermal diffusion in equilibrium is given by (*12*)

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References and Notes

[6] **The Holocene-Younger Dryas Transition Recorded at Summit, Greenland**

K. C. Taylor; P. A. Mayewski; R. B. Alley; E. J. Brook; A. J. Gow; P. M. Grootes; D. A. Meese; E. S. Saltzman; J. P. Severinghaus; M. S. Twickler; J. W. C. White; S. Whitlow; G. A. Zielinski

Science, New Series, Vol. 278, No. 5339. (Oct. 31, 1997), pp. 825-827.

Stable URL:

http://links.jstor.org/sici?sici=0036-8075%2819971031%293%3A278%3A5339%3C825%3ATHDTRA%3E2.0.CO%3B2-Z

[9] **A 25,000-Year Tropical Climate History from Bolivian Ice Cores**

L. G. Thompson; M. E. Davis; E. Mosley-Thompson; T. A. Sowers; K. A. Henderson; V. S. Zagorodnov; P.-N. Lin; V. N. Mikhalenko; R. K. Campen; J. F. Bolzan; J. Cole-Dai; B. Francou

Science, New Series, Vol. 282, No. 5395. (Dec. 4, 1998), pp. 1858-1864.

Stable URL:

http://links.jstor.org/sici?sici=0036-8075%2819981204%293%3A282%3A5395%3C1858%3AA2TCHF%3E2.0.CO%3B2-I

[12] **Iceberg Discharges Into the North Atlantic on Millennial Time Scales During the Last Glaciation**

Gerard C. Bond; Rusty Lotti

Science, New Series, Vol. 267, No. 5200. (Feb. 17, 1995), pp. 1005-1010.

Stable URL:

http://links.jstor.org/sici?sici=0036-8075%2819950217%293%3A267%3A5200%3C1005%3AIDITNA%3E2.0.CO%3B2-0

LINKED CITATIONS

[16] **Rapid Variations in Atmospheric Methane Concentration During the Past 110,000 Years**

Edward J. Brook; Todd Sowers; Joe Orchardo

Science, New Series, Vol. 273, No. 5278. (Aug. 23, 1996), pp. 1087-1091.

Stable URL:

http://links.jstor.org/sici?sici=0036-8075%2819960823%293%3A273%3A5278%3C1087%3ARVIAMC%3E2.0.CO%3B2-1

[18] **Synchronous Climate Changes in Antarctica and the North Atlantic**

E. J. Steig; E. J. Brook; J. W. C. White; C. M. Sucher; M. L. Bender; S. J. Lehman; D. L. Morse; E. D. Waddington; G. D. Clow

Science, New Series, Vol. 282, No. 5386. (Oct. 2, 1998), pp. 92-95.

Stable URL:

http://links.jstor.org/sici?sici=0036-8075%2819981002%293%3A282%3A5386%3C92%3ASCCIAA%3E2.0.CO%3B2-Z

[24] **Atmospheric Trends and Lifetime of CH₃CCl₃ and Global OH Concentrations**

R. G. Prinn; R. F. Weiss; B. R. Miller; J. Huang; F. N. Alyea; D. M. Cunnold; P. J. Fraser; D. E. Hartley; P. G. Simmonds

Science, New Series, Vol. 269, No. 5221. (Jul. 14, 1995), pp. 187-192.

Stable URL:

http://links.jstor.org/sici?sici=0036-8075%2819950714%293%3A269%3A5221%3C187%3AATALOC%3E2.0.CO%3B2-I

[29] **Glacial-Interglacial Changes in Moisture Sources for Greenland: Influences on the Ice Core Record of Climate**

C. D. Charles; D. Rind; J. Jouzel; R. D. Koster; R. G. Fairbanks

Science, New Series, Vol. 263, No. 5146. (Jan. 28, 1994), pp. 508-511.

Stable URL:

http://links.jstor.org/sici?sici=0036-8075%2819940128%293%3A263%3A5146%3C508%3AGCIMSF%3E2.0.CO%3B2-0

[30] **Large Arctic Temperature Change at the Wisconsin-Holocene Glacial Transition**

Kurt M. Cuffey; Gary D. Clow; Richard B. Alley; Minze Stuiver; Edwin D. Waddington; Richard W. Saltus

Science, New Series, Vol. 270, No. 5235. (Oct. 20, 1995), pp. 455-458.

Stable URL:

http://links.jstor.org/sici?sici=0036-8075%2819951020%293%3A270%3A5235%3C455%3ALATCAT%3E2.0.CO%3B2-J

LINKED CITATIONS

[30] **Past Temperatures Directly from the Greenland Ice Sheet**

D. Dahl-Jensen; K. Mosegaard; N. Gundestrup; G. D. Clow; S. J. Johnsen; A. W. Hansen; N. Balling

Science, New Series, Vol. 282, No. 5387. (Oct. 9, 1998), pp. 268-271.

Stable URL:

http://links.jstor.org/sici?sici=0036-8075%2819981009%293%3A282%3A5387%3C268%3APTDFTG%3E2.0.CO%3B2-X

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H. Craig; Y. Horibe; T. Sowers

Science, New Series, Vol. 242, No. 4886. (Dec. 23, 1988), pp. 1675-1678.

Stable URL:

http://links.jstor.org/sici?sici=0036-8075%2819881223%293%3A242%3A4886%3C1675%3AGSOGAI%3E2.0.CO%3B2-T

[47] **Changes in Atmospheric Circulation and Ocean Ice Cover over the North Atlantic During the Last 41,000 Years**

P. A. Mayewski; L. D. Meeker; S. Whitlow; M. S. Twickler; M. C. Morrison; P. Bloomfield; G. C. Bond; R. B. Alley; A. J. Gow; P. M. Grootes; D. A. Meese; M. Ram; K. C. Taylor; W. Wumkes

Science, New Series, Vol. 263, No. 5154. (Mar. 25, 1994), pp. 1747-1751.

Stable URL:

http://links.jstor.org/sici?sici=0036-8075%2819940325%293%3A263%3A5154%3C1747%3ACIACAO%3E2.0.CO%3B2-Q