Surface and Deep Ocean Interactions During the Cold Climate Event 8200 Years Ago
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Other mechanisms that may contribute to the presence of this peak include two- and three-magnon non–momentum-conserving processes that originate from scattering from defects (25, 33, 34). The linewidth originating from the latter process is peaked at intermediate q. Using parameters derived from comparison with data on RbMnF$_3$, one can estimate its contribution in MnF$_2$ to be two orders of magnitude smaller than the data (34).

The challenge to theory posed by the temperature- and momentum-dependent peaks in the magnon linewidth in MnF$_2$ should stimulate new activity in the field of spin-wave decay mechanisms. High-resolution lifetime measurements over the full Brillouin zone in a relatively simple antiferromagnet such as MnF$_2$ permit detailed evaluation of proposed processes, which should provide a basis for addressing such interactions in more complex magnetic systems.

References and Notes
7. Materials and methods are available as supporting material on Science Online.
11. We use a notation in which the momentum Q transferred by the neutron is Q = G + q, where q is the momentum transfer within the Brillouin zone centered at the reciprocal lattice vector G. These quantities are expressed in reciprocal lattice units (r.l.u.). For instance, Q = (HKL), with Q$_{hkl}$ = H (2π/λ), and q = (hkl), with q$_{hkl}$ = (2π/λ).
13. The conditions for application of the results of Harris et al. in various intervals of temperature and energy are quite restrictive due to the presence of strict inequalities, and the resulting analytical expressions for the linewidth are valid only in extremely limited regions, if at all. Here, we have treated these conditions as if they involved simple inequalities, and we have plotted the four solutions that apply to different temperature regions together.
19. In contrast, in earlier work Cottam and Stinchcombe suggest that in the general case, the linewidth contribution from four-magnon scattering dominates at low temperatures (14).
20. A suggestion of nonmonotonic behavior in the linewidth as a function of q is contained in (14). For the case in which the anisotropy takes the form of an anisotropy field $H_a$, the linewidth peaks at an intermediate value of q and then declines as it approaches the Brillouin zone boundary at q = 0.5 r.l.u. However, the estimated value of q at which the peak occurs would be only ~0.008 r.l.u. for MnF$_2$. The behavior of the linewidth at large q for the case of anisotropic exchange was not evaluated.
21. Wooley and White (25) predict a peak as a function of q in MnF$_2$ for the linewidth due to a four-magnon scattering process in a relatively small magnetic field between 2 and 3 K. The peak becomes less pronounced and shifts to higher q as the temperature increases to 3 K (from 0.30 r.l.u. at 2 K to 0.37 r.l.u. at 3 K). At a slightly higher field, only a simple maximum is present between 4 and 10 K for a four-magnon scattering process (34). The authors did not incorporate anisotropy in their calculations; later work by the same authors showed that the inclusion of anisotropy results in a decrease in the calculated linewidth of roughly an order of magnitude for q ~ 0 in a large magnetic field (35).
24. The effect of the curvature of the magnon dispersion on the spin-echo resolution is largest at q ~ 0, and largest at low temperature, where the spin-wave stiffness is strongest. Thus, curvature not properly taken into account in the analytical correction calculation would produce an apparent linewidth. However, this spurious effect would be peaked at q ~ 0 and is therefore unlikely to be responsible for the observed peak.
36. We thank G. Schmidt of the Crystal Growth Facility of the Cornell Center for Materials Research for the loan of a MnF$_2$ crystal of excellent quality, R. Henes and J. Major for the γ-ray diffractionmetry measurements, J. Peters for cryogenic assistance, G. Khaliullin and R. K. Kremer for illuminating discussions, P. Aynajian for participation in some of the calibration measurements, and R. Noack for technical assistance.

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Fig. S1
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Surface and Deep Ocean Interactions During the Cold Climate Event 8200 Years Ago
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Evidence from a North Atlantic deep-sea sediment core reveals that the largest climatic perturbation in our present interglacial, the 8200-year event, is marked by two distinct cooling events in the subpolar North Atlantic at 8490 and 8290 years ago. An associated reduction in deep flow speed provides evidence of a significant change to a major downwelling limb of the Atlantic meridional overturning circulation. The existence of a distinct surface freshening signal during these events strongly suggests that the sequenced surface and deep ocean changes were forced by pulsed meltwater outbursts from a multistage final drainage of the proglacial lakes associated with the decaying Laurentide Ice Sheet margin.

Changes in the mode of operation of the Atlantic meridional overturning circulation (MOC) are thought to be an important driver of rapid, large-scale climate change (1, 2) and are widely believed to be linked to changes in freshwater forcing (3).

The climate of the present interglacial, which began about 10,000 years ago, is remarkably stable when compared to the last glacial period (4) with the exception of a single, large climatic excursion ~8200 years before the present (yr B.P.) (5), commonly referred to as the 8.2 ky event. It has been suggested that the short-lived cooling episode captured in the Greenland ice core records is one element of a broader climate signal that has an interhemispheric signature (6), but the magnitude and the sequence of changes surrounding this climate event are not always straightforward to decipher (7). The decay of the Laurentide Ice Sheet and the catastrophic final drainage of the large proglacial Lakes Agassiz and Ojibway at ~8470 yr B.P. (8) preceded the 8.2 ky event, so the precise relation between the lake outburst, the oceanic response, and the cooling observed in the wider circumpolar-Atlantic region remains unresolved (6). However, the similarity in the timing of these two events, coupled with model-derived suggestions regarding the sen-
sitivity of the MOC to freshwater inputs (9), have led to a widely held view that a meltwater-induced alteration to the ocean circulation may explain the trends observed in paleoclimate records. Despite recent efforts to detect MOC changes in paleoceanographic archives (10, 11), clear and conclusive evidence has yet to be reported (6, 7).

We analyzed deep-sea sediment core MD99-2251 recovered from the southern limb of the Gardar Drift in the subpolar North Atlantic (57º26.87'N, 27º54.47'W; 2620-m water depth), where the interaction of Iceland Scotland Overflow Water (ISOW) with the underlying sea-floor topography results in sediment focusing and exceptionally enhanced sediment accumulation rates. Calibrated accelerator mass spectrometry (AMS) 14C dates that have been converted to calendrical ages by using the CALIB program (12) indicate a mean sediment accumulation rate of ~110 cm ky⁻¹ through the studied time interval of 9200 to 7200 years ago. Our sampling resolution over this interval is <20 years, with each sample representing an integrated signal of 6 to 10 years. Consequently, it has been possible to produce regional climate proxy records of unprecedented detail that reveal unambiguous information about the nature of subcentennial scale events and the relative phasing of surface and deep ocean changes. Because our climate proxy reconstructions are coregistered signals from within a single sediment core, the observed sequence of events and temporal offsets are robust features independent of correlation problems (7).

Variations in the relative abundance of the polar foraminifer Neogloboquadrina pachyderma sinitral (%) coiling are routinely used to determine past positional changes of cold surface waters (4, 13). The marked decline in N. pachyderma s. abundances from >60% to <2% at ~1600 cm in core MD99-2251 defines the northward retreat of polar waters at the onset of the Holocene (Fig. 1). Low abundances of the polar foraminifer (mean of 1.4%) persist throughout the Holocene with the exception of a marked excursion centered at ~1220 cm. This episode marks the cold climatic extreme in the Holocene, with faunal-derived transfer function estimates of summer sea surface temperature (SST) ~2°C colder than at present. The N. pachyderma s. percent abundance maximum is dated to 8290 yr B.P. in our record, in good agreement with the age of the pronounced cold excursion identified in the δ18O-derived atmospheric temperature record from the Greenland Ice Sheet Project 2 (GISP2) ice core (5, 14). Although age estimates for the cold extreme differ by ~100 years in the oceanic and ice proxy records, this difference is not significant compared with the combined dating uncertainties (12, 15), and we consider the surface ocean and atmospheric cooling events to be synchronous. This interpretation is supported by the structural similarity across the temperature excursion, i.e., the rates of change associated with the cooling into and warming after the event together with the near-identical duration for the event of ~70 years in both archives (Fig. 2).

Both of our surface ocean proxies, % N. pachyderma s. and the δ18O composition of the planktonic foraminifer Globigerina bulloides, reveal the existence of a separate, earlier climate event centered at 8490 yr B.P. (Fig. 2). This indicates that the prominent 8.2 ky cooling was not an isolated event but may be the culmination of a change initiated ~200 years earlier. These two marked cooling episodes occur within an interval of reduced SSTs that existed from ~8900 to 8000 yr B.P. (Fig. 3). This multicentennial cool phase, which is broadly consistent with the recent identification of a 400- to 600-year climate anomaly preceding the 8.2 ky event (7), signifies a longer-term change in the pattern of ocean heat transport and may reflect the cumulative impact of outflow from melting ice sheets (16). The dust supply to Greenland, inferred from variations in potassium content in the GISP2 ice core (17), suggests a similar pattern of polar atmospheric reorganization over this interval.

The abrupt initial cooling at 8490 yr B.P., which lasted ~80 years according to the % N. pachyderma s. data, was coincident with a ~0.6 per mil (‰) depletion in planktonic foraminiferal δ18O values (Fig. 2). The magnitude of this δ18O excursion increases further once temperature and ice volume effects are incorporated (Fig. 3). The resulting shift in δ18Oseawater composition can only readily be explained by a substantial freshening of the surface ocean. This perturbation corresponds to a salinity reduction of 1 to 1.3, assuming a freshwater input with an end-member δ18O composition of between ~35‰ and ~20‰ (18). The changes in surface ocean proxy records 8490 years ago in the subpolar North Atlantic have no large-scale counterpart within the GISP2 temperature data but are hinted at in other paleoclimate proxy records (17, 19), and, significantly, the timing of these coregistered signals is indistinguishable from the timing of the catastrophic meltwater release from Lakes Agassiz and Ojibway at ~8470 yr B.P. (8). This discharge event, whose combined volume has been estimated at 163,000 km³ [or 5.2 × 10⁷ m³ s⁻¹ (Sv) if released in one year (16)], is the only identified potential freshwater source of sufficient magnitude to explain the presence of the pronounced δ18Oseawater signal at the distal, open ocean location of core MD99-2251.

The more prominent cooling episode, associated with a further SST decrease of ~1.2°C centered at 8290 yr B.P. (Fig. 3), occurs ~200 years after the initial lake outburst and surface ocean cooling event. This event, correlative with the 8.2 ky event observed in Greenland ice, has a wider oceanic imprint, and is registered elsewhere both in the North Atlantic (4, 20) and Norwegian Sea (13, 21). Notably, the initiation of this SST cooling phase is synchronous with a second and less pronounced freshening of the surface ocean. This signal may be related to a later, smaller freshwater outburst from the western Agassiz

![Fig. 1. The % abundance (>150 μm) of the polar foraminifer N. pachyderma s. coiling through the Holocene section of core MD99-2251 (note reversed axis). N. pachyderma s. % traces the rapid retreat of polar waters, marking the onset of the Holocene warming, at ~1600 cm and captures a significant, brief readvance of polar waters at ~1220 cm. The core chronology has its basis in 23 accelerator mass spectrometry (AMS) 14C dates that have been converted to calendrical ages by using the CALIB program (12). Position and calibrated ages of AMS 14C dates are indicated along depth axis (black triangles) (12). The shaded area highlights the study interval shown in Figs. 2 and 3 (i.e., 9200 to 7200 years ago). Mean sediment accumulation rate through this interval is ~110 cm ky⁻¹.](https://www.sciencemag.org/content/sci/312/5773/1930.full)
Fig. 2. Surface ocean climate proxy records in core MD99-2251 compared to the GISP2 ice core data over the interval 9200 to 7200 years ago. (A) N. pachyderma s. % abundance data and (B) the GISP2 temperature (δ18Oice) record (24). Dating errors around the 8.2 ky event are shown for each time series. The lower and upper 2σ limits on the calibrated AMS 14C date for the cooling event in core MD99-2251 (8290 yr B.P. at 1220-cm depth) are 8180 and 8340 yr B.P., respectively. The GISP2 time scale has an estimated error of up to 2% on the basis of annual layer counting (25); this equates to a maximum dating uncertainty of ±164 years for 8200 years ago. (C) GISP2 δ18Oice record (red) overlain on the N. pachyderma s. % abundance data (blue) in MD99-2251, with the age scale of the GISP2 δ18Oice offset by 100 years (i.e., showing the interval 9100 to 7100 ice core years). The application of a constant age offset of ±100 years, which is well within the dating uncertainties of either one of the two records, yields a good agreement between the cold extremes represented in each proxy record. (D) planktonic foraminiferal stable isotope (δ18Ocalcite) data measured for the species G. bulloides. Diamond marker and error bar indicates the timing of Lakes Agassiz and Ojibway outburst at ~8470 yr B.P. (8) and is associated with the first major shift in MD99-2251 proxy data (indicated by older shaded bar). Younger shaded bar highlights the largest cooling in the MD99-2251 proxy data, correlated with the 8.2 ky event.
meaningful if the prevailing MOC configurations were similar for both events. Whatever the origin of the ISOW flow signal, it is clear that we observe a significant change in the initial stages of the MOC. These findings lend support to models that predict a reduction in MOC intensity during the 8.2-ky event (6, 9, 24).

Although two minima in solar output inferred from cosmogenic $^{14}$C production records (29) may correlate with the key meltwater events and SST minima that we identify at 8490 and 8290 yr B.P., there is no clear relation between oceanic proxy records and solar variability through the remainder of the study interval. Hence, any causal linkage would appear to be dependent on other factors, such as the preconditioning of the North Atlantic by enhanced meltwater input. Rather, our coregistered paleoclimatic records most likely demonstrate a cause and effect relationship: The 8.2 ky event was forced by the freshwater anomaly associated with the pulsed drainage of Lakes Agassiz and Ojibway, which caused extensive cooling and freshening in the North Atlantic Ocean, and this in turn can be directly linked to alterations in the MOC. Although the duration of the meltwater release is likely to be on the order of years (16) and the surface ocean perturbations seen in the subpolar North Atlantic are on the order of decades, the associated change in the deep ocean flow lasted ~400 years. Such information on the sensitivity and response of the climate system to a disturbance of the MOC during interglacial times is crucial and relevant to our possible future, because many global climate models suggest that an interruption of the MOC is a likely outcome of future climate change (30).

References and Notes
5. R. B. Alley et al., Geology 25, 483 (1997).
12. Materials and methods are available as supporting material on Science Online.
32. We thank G. Bianchi, Cardifc, for running the mass spectrometer and H. Medley, Cardifc, and S. Bennett, UEA, for providing invaluable laboratory assistance. Financial support was provided by the Natural Environment Research Council (NERC) and NERC Radiocarbon Laboratory. Data are available from World Data Center-A for Paleoclimatology (www.ncdc.noaa.gov/paleo/data.html).

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Materials and Methods
Table S1
References
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Fig. 3. Surface and deep ocean linkages over the interval from 9200 to 7200 years ago. (A) N. pachyderma s. % abundance data, (B) summer SST, (C) surface ocean stable isotope ($\delta^{18}$O$_{\text{seawater}}$) composition, and (D) SS mean grain size. SST values (error ± 1°C) were estimated using the modern analog technique and are derived from full planktonic foraminiferal assemblage counts (~400 individuals) using the >150-µm size fraction (22). The depletion in $\delta^{18}$O$_{\text{seawater}}$ values at 8490 and 8330 yr B.P. correspond to reductions in surface ocean salinity of 1 to 1.3 and 0.5 to 0.7, respectively, assuming a freshwater end-member $\delta^{18}$O composition of between −35‰ and −20‰ (28). The SS is a paleocurrent flow speed proxy (double arrow), where a higher mean reflects stronger flow of the depositing current and vice versa (23). Shaded bars are as in Fig. 2 and indicate the two pulses of cooling and freshening of surface ocean conditions that appear to correlate with a larger (3.6 Sv) meltwater outburst at ~7700 $^{14}$C yr B.P. and a second, smaller meltwater discharge (1.6 Sv) dated at ~7600 $^{14}$C yr B.P. (16).