

GLOBAL CLIMATE CHANGE

Synchronous timing of abrupt climate changes during the last glacial period

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Abrupt climate changes during the last glacial period have been detected in a global array of palaeoclimate records, but our understanding of their absolute timing and regional synchrony is incomplete. Our compilation of 63 published, independently dated speleothem records shows that abrupt warmings in Greenland were associated with synchronous climate changes across the Asian Monsoon, South American Monsoon, and European-Mediterranean regions that occurred within decades. Together with the demonstration of bipolar synchrony in atmospheric response, this provides independent evidence of synchronous high-latitude-to-tropical coupling of climate changes during these abrupt warmings. Our results provide a globally coherent framework with which to validate model simulations of abrupt climate change and to constrain ice-core chronologies.

Climate records from Greenland ice cores spanning the last glacial cycle (115,000 to 11,700 years ago) reveal a series of centennial- to millennial-scale cold-warm oscillations called Dansgaard-Oeschger (DO) events (1–3). In Greenland, each event commenced with a rapid transition to warm conditions [a “Greenland Interstadial” (GI)] followed by a gradual, then abrupt, return to a cold climate state [a “Greenland Stadial” (GS)] (2, 3). This pattern is widely accepted to be associated with changes in the strength of the Atlantic Meridional Overturning Circulation (AMOC), which regulates interhemispheric oceanic heat flux, as captured in the thermal oceanic bipolar seesaw theory (4–7). During the DO warm phase, a strong AMOC exports heat from the south and tropics to the high latitudes of the North Atlantic, leading to cooling of the global ocean north of the Antarctic Circumpolar Current (ACC) and reduced temperatures over Antarctica (6). At times of DO cooling, a weakened AMOC reduces northward ocean heat transport and results in heat accumulation in the Southern Ocean, leading to Antarctic warming. These changes are broadly reflected in ocean-sediment records across the Atlantic (8–10) as far south as the mid-latitude South Atlantic, where both

surface (4) and deep-ocean signals (11) show a similar abruptness and timing to those seen in the North Atlantic but of opposite sign in surface ocean temperature. Further poleward, the abruptness is dampened by the ACC, which influences heat flux into and out of the Southern Ocean (4–6). This ultimately leads to less abrupt air-temperature changes over Antarctica (12), whose onsets lag the Greenland counterparts by around 200 ± 100 years (13). The Antarctic changes are therefore symptomatic of a slow-oceanic response to the abrupt changes recorded north of the ACC (6, 14).

Changes to the cross-hemisphere ocean-temperature difference induced by switches in AMOC mode drive meridional shifts in the atmospheric mean state, particularly the position of the Intertropical Convergence Zone (ITCZ) (14, 15). This is most vividly observed in terrestrial monsoon records from both sides of the equator (16, 17) at times when Greenland climate switches abruptly from a stadial to an interstadial (called here an “interstadial onset”), and is supported by numerical climate model outputs (Fig. 1) (18–20). These higher-amplitude warming episodes correspond to air-temperature increases over the Greenland ice sheet of up to 16°C (21), typically in around 80 years. Such a strong high-latitude-to-tropical teleconnection indicates widespread atmospheric reorganization at the onset of interstadials. However, there is uncertainty over the exact phasing: Were climate responses synchronous between different monsoon regions, and between the wider tropical realm and Greenland? Recent reassessments of ice-core data (22, 23) suggest that atmospheric changes associated with abrupt warming and cooling in Greenland were transmitted more or less synchronously as far south as the Antarctic ice sheet, indicating rapid reorganization of atmospheric circulation. Such bipolar synchrony implies that mid-latitude and tropical regions also responded rapidly as Green-

land temperatures changed abruptly. This has been the assumption in previous studies (24, 25) but has yet to be rigorously tested. Answering these questions will deepen our understanding of the underlying dynamics of global climate teleconnections during abrupt climate changes, which serves as a basis to test climate models used for future projections (26).

Testing the large-scale synchrony of stadial-interstadial changes to augment the suggestion of bipolar synchrony is hampered by the limited availability of independently dated palaeoclimate time series with sufficiently constrained chronologies (2σ age uncertainties in the range of decades to centuries). The chronologies of many last-glacial records are aligned to other marine, ice-core, or speleothem age models under the assumption that abrupt events did occur synchronously between sites at least to within the resolution of the records in question (25, 27, 28). This prevents any independent assessments of potential regional leads and lags between these records (29). The current Greenland Ice Core Chronologies (GICC05 and its extension GICC05modelext) (30–32) provide the chronological framework for the latest event stratigraphy of abrupt climate changes (3), but published age uncertainties in the annual-layer counting accumulate with increasing age, reaching ± 2600 years at 60,000 years before 1950 [before the present (B.P.)]. Beyond this, age uncertainties of the flow-model extension to the chronology, GICC05modelext, are unquantified (32). Thus, any comparison of the timing of interstadials and stadials between ice cores and other archives becomes less certain through time, even though the incremental nature of the ice-core counting errors produces high-precision constraints on the time difference between successive events—that is, the duration of and spacing between events (fig. S1).

Speleothem records of interstadial onset

To test the global synchrony of these events, we investigated the timing of the abrupt interstadial onset for 25 major and 28 minor interstadial events (3, 33) using 63 published, high-resolution, and precisely dated speleothem records covering the last glacial period (Fig. 1 and table S1) (34). Speleothems (secondary mineral deposits found in caves) have great potential because they can be radiometrically dated with great accuracy with uranium-series methods to a precision of 0.1 to 1% (2σ) over last-glacial time scales [for example, (35)]. Numerous speleothem stable oxygen isotope ($\delta^{18}\text{O}$) records spanning the Northern Hemisphere (NH) mid-latitudes to the Southern Hemisphere (SH) subtropics capture stadial-interstadial transitions, particularly the more pronounced interstadial onsets, where local precipitation and/or temperature changes lead to prominent changes in $\delta^{18}\text{O}$ (17, 36, 37).

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We standardized all speleothem time series to avoid methodological bias in age calculations and depth-age model construction then identified the onset of each interstadial by adapting a technique applied to the ice-core record (fig. S2) (3, 34). Because of the nature of speleothem growth, variability is exhibited in the temporal span of each record, with the majority covering a few tens of thousands of years, although some are much shorter (fig. S3). Within a single record, the temporal resolution may vary greatly owing to changes in the growth rate of the speleothem. All records are dated directly by using uranium-thorium (U-Th) methods, and each time series is free of synchronization to any tuning target. The isotope records for all speleothems are shown in data file S3 and provided in data file S7. Of the 53 interstadials identified in the ice cores (3), 39 could be confidently resolved in at least two records in the speleothem dataset, with the overwhelming majority of records falling into either the Asian Summer Monsoon (ASM), the South American Monsoon (SAM), or the Europe-Mediterranean (EM) domains (Fig. 1) (34). The speleothem $\delta^{18}\text{O}$ in these three regions exhibits well-documented changes across interstadial onsets (Fig. 2), with patterns gen-

erally well reproduced in DO-type transient climate-model simulations (Fig. 1, background maps). In the ASM domain, warming in the North Atlantic is associated with a strengthening of both the Indian and East Asian Summer Monsoon subsystems, which produces decreased $\delta^{18}\text{O}$ values driven by variations in rainout, air-mass trajectories, and/or rainfall amounts (Fig. 1A) (36–38). The $\delta^{18}\text{O}$ in the SAM domain increases because of the same processes under a weakened monsoon (Fig. 1B) (39). This “monsoon-seesaw pattern” over interstadial onsets is consistent with a northward shift of the ITCZ (10, 17). In the EM region, the speleothem $\delta^{18}\text{O}$ response to interstadial onsets is slightly more complex. Sites situated around the Alps and western Turkey (Sofular) show increased isotopic values (40, 41) as the climate warms, which is consistent with the dominating temperature effect (Fig. 1C) on the $\delta^{18}\text{O}$ of local precipitation (42), whereas seasonal changes in moisture source and the source effect cause $\delta^{18}\text{O}$ values in northeastern Turkey (Karaca) to decrease (43). Around the Mediterranean, regional warming increases the amount of rainfall reaching cave sites (44), leading to decreasing $\delta^{18}\text{O}$ values owing to the rainfall amount effect (42). This effect

also occurs in the eastern Mediterranean, where it may be modified by changes in local seawater $\delta^{18}\text{O}$ composition related to higher Nile discharge (45). The abruptness of speleothem $\delta^{18}\text{O}$ change at the onset of the interstadials (Fig. 2) is, in most cases, comparable with what is observed in the ice-core record (41), when accounting for differences in temporal resolution, speleothem age-model uncertainties, and the smoothing of atmospheric signals in speleothems due to groundwater transport and mixing processes in the karst aquifer (46).

Testing the synchrony of interstadial onset

In restricting our comparison to the three regions best represented by the speleothem data, we first tested the intraregional synchrony of an interstadial onset, hypothesizing that abrupt changes within a single region should be recorded practically simultaneously by speleothems at each cave site. Second, we tested the synchrony between the three regions for each interstadial onset. We assessed the degree of intra- and interregional synchrony using the reduced χ^2 statistic, known in geochronology as the mean-square weighted deviation (MSWD) (47, 48), which tests whether a group of radiometric ages belong to a single population (34). Where we found statistical agreement between speleothem ages from the same region, we calculated the error-weighted mean (EWM) (34) and assigned this as the “regional age.” These regional ages were used to test the synchrony between regions and to derive a composite speleothem age for each interstadial onset in cases for which data from multiple regions were available. These age estimates for interstadial onsets form the Speleothem Interstadial Onset Compilation data set (SIOC19). We also assessed the extent to which an age estimate for the timing of an interstadial onset based on data from only one region could be used as a wider event-age indicator. We then compared the SIOC19 age estimates for each interstadial onset with their timing in the ice-core GICC05 and GICC05modelext chronologies (3). Last, we investigated the regional age offset over interstadial onsets for which two or more speleothem records from the same region were available.

For 34 of the 37 interstadials recorded in multiple speleothem records, there is strong intraregional agreement (within 2σ uncertainties), comprising cases in which one (nine interstadials) or multiple (25 interstadials) regions are represented (Fig. 2, fig. S4, and table S2). The three remaining onsets (into interstadials 12a, 14e, and 23.1) show disagreement within all of the represented regions [the MSWD lies outside the accepted range (34)], and we assigned an indeterminate result to these (table S2D). Three of the 34 transitions show disagreement within one or more of the regions represented (interstadials 4,

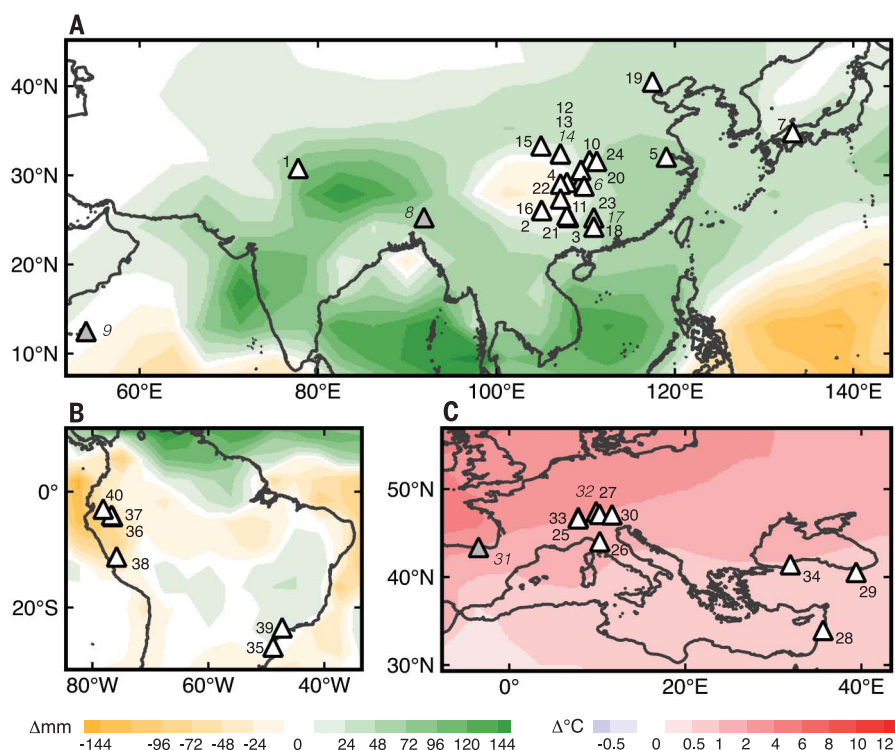
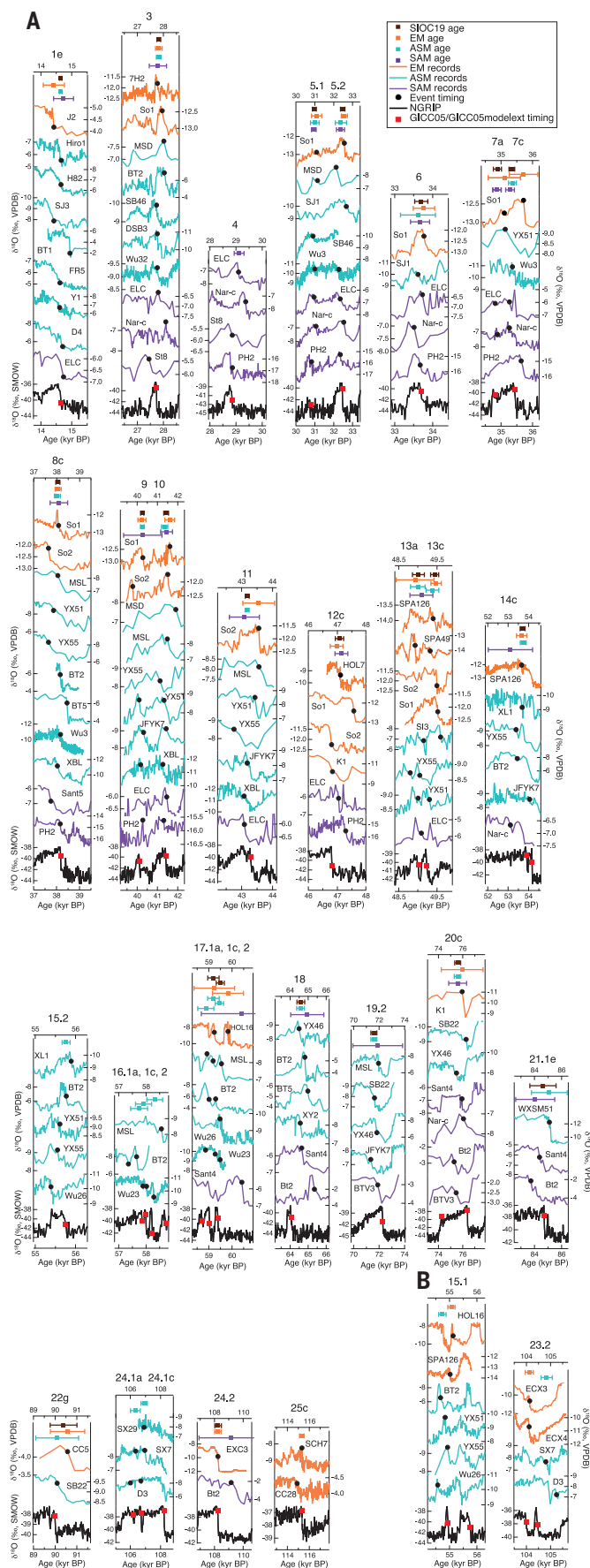


Fig. 1. Location of last glacial speleothem records included in the compilation. (A to C) Cave sites for records from the (A) Asian Summer Monsoon region, (B) South American Monsoon region, (C) Europe-Mediterranean region. Gray triangles and italicized font represent records that were used in the comparison but are not represented in the age estimates for the synchronous events shown in Table 1. Reference to the numbering is provided in table S1, which shows the complete list of speleothem records and citation to the original publication. Shaded backgrounds are composite anomalies of [(A) and (B)] annual mean precipitation and (C) surface air temperature between interstadial and stadial states in DO-type transient experiments (19, 20, 34, 52).

ASM and EM; 12c, ASM; and 24.2, ASM), so we excluded the data from the anomalous region from further (interregional) comparisons (table S2, A and B, region listed as “unresolved”). The strong overall intraregional agreement validates the assumption of intraregional synchrony. This enables us to treat the records from each region as a single population and combine them into a single regional age, based on the EWM, with a reduced uncertainty relative to the individual records. It is likely that statistically significant age differences within a region may be due to either sample and/or site-specific processes affecting the registration of the climate signal (for example, competing mechanisms driving local $\delta^{18}\text{O}$) or chronological issues such as undetected growth hiatuses, uranium leaching (49), and/or suboptimal positioning of U-Th age or stable isotope samples.

In comparing the regional EWM ages, we found strong evidence of synchronous onset for 23 of the 25 interstadials for which we were able to make interregional comparisons (Fig. 2; Table 1, entries denoted with an asterisk; fig. S4; and table S2). In the case of onsets younger than 40,000 years B.P., the difference in the mean age between any region represented by multiple speleothem records (thus giving greater confidence and narrowing the regional uncertainty) ranges from 3 to 72 years (table S2), with an uncertainty that spans zero. This indicates that the onset of interstadials is likely synchronous within decades. It also highlights that the major limiting factor in deriving more precise constraints of event synchrony is related to the magnitude of the U-Th dating errors of individual speleothem records rather than scatter in the ages themselves.

Fig. 2. Stable isotope records of individual speleothems for each interstadial onset. (A and B) Each speleothem record displayed has met screening and event identification criteria (34) and is color-coded according to region (EM, orange; ASM, blue; and SAM, purple). All records are plotted on the recalculated chronology (34). Plots are provided for interstadial onsets found to be (A) synchronous and (B) non-synchronous. (A) includes interstadials for which we have data from only one region. The North Greenland Ice Core Project (NGRIP) $\delta^{18}\text{O}$ series on the GICC05 or GICC05modelext chronology is plotted in black, with each event position (3) indicated with a red square. The corresponding event position in each speleothem record is indicated with a solid black circle (timing and uncertainty are provided in data file S1). Markers with error bars at the top of (A) and (B) indicate the age and 2σ uncertainty of the interregional age estimate (brown) and intraregional age estimates (EM, orange; ASM, blue; and SAM, purple). Lettering provides the name of each speleothem record. The full details and original reference for each record are provided in table S1. VPDB, Vienna Pee Dee belemnite; SMOW, standard mean ocean water; kyr, thousand years.



The onset of the two interstadials that are not recorded synchronously between regions are those to interstadial 15.1 (recorded 334 ± 185 years earlier in the EM compared with the ASM) and interstadial 23.2 (recorded 716 ± 287 years later in the EM compared with the ASM); in both cases, there were no high-quality records from the SAM region (Fig. 2B and table S2C). These anomalies could reflect a genuine asynchrony, implying that these interstadials are

of an unusual nature in the context of the other 23. However, such asynchrony may reflect varying sensitivity to these short-lived interstadials, which is also suggested by some structural difference to the very sharp peaks in Greenland. The regional lead-lag is not consistent, however, so does not suggest a systematic regionally asynchronous response to interstadials of this type.

Given that the great majority of interstadial onsets tested for interregional synchrony (23

out of 25) show chronological agreement, it is reasonable to assume that an EWM age estimate for an interstadial onset in which only a single region is represented (the case for nine interstadials) (Table 1, entries denoted with “+”, and table S2B) is also indicative of a transition’s broader timing. We tested this assumption by comparing the spacing of consecutive interstadial onsets in the speleothems with the corresponding annual-layer-counted intervals in the GICC05 ice-core chronology (Fig. 3) (3, 31, 50). As noted above, although the precision on the absolute ages of the GICC05 chronology decreases back in time, the interval between the onset of consecutive interstadials over the layer-counted section (0 to 60,000 years B.P.) can be estimated with relatively high precision [the average 2[σ] error for all consecutive intervals in this period is 5.0 ± 1.3 (1[σ]) (Fig. 3 and fig. S1). Interstadial duration estimates based on the SIOC19 age estimates (including those based on data from only one region) agree within uncertainty with those from the GICC05 chronology, with the exception of one interval out of 22 (interval 12c to 11, GICC05 estimate 3520 ± 176 years, SIOC19 3886 ± 158 years) (Fig. 3). Beyond the layer-counted section (>60,000 years B.P.), only two out of nine intervals show discrepancies outside of the uncertainty estimate of SIOC19 (interval 19.2 to 18, GICC05modelxt 8240 years, SIOC19 6998 ± 312 years; interval 18 to 17.2, GICC05modelxt 4660 years, SIOC19 5107 ± 273 years) (Fig. 3). Overall, however, the agreement between SIOC19 and the GICC05 and GICC05modelxt chronologies for the duration of consecutive interstadials validates the accuracy of the relative ages (within uncertainties) of the single-region estimates and thus their inclusion alongside the multiregional EWM ages. Nevertheless, we suggest that the timing estimates of interstadial onsets based on data from only a single region (denoted by † in Table 1) should be treated cautiously until they are verified by additional high-quality speleothem records from other regions.

The regional speleothem ages of interstadial onsets from three geographically diverse areas being likely synchronous to well within a century suggests that the timing should agree—within the same uncertainty—with corresponding events in Greenland. We next evaluated the GICC05 and GICC05modelxt chronologies by comparing speleothem and ice-core ages for the 32 interstadials in Table 1. The multiregion (23 interstadials) and single-region (nine interstadials) speleothem age estimates are all within 1.1% of the ice-core mean ages (Table 1 and fig. S6A). Three of the four largest offsets (20c, 19.2, and 18) are observed beyond the layer-counted interval (fig. S6A), for which ice-core chronological control is less secure owing to uncertainties associated with the glaciological modeling that underpins this section (GICC05modelxt) (32).

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Table 1. The timing of each interstadial onset in the combined speleothem records (SIOC19) and the corresponding ages in the Greenland ice-core chronology. The combined speleothem age is the error-weighted mean (and the corresponding 2σ uncertainty) of multiple individual speleothem records that have been grouped at a regional level (Asian Summer Monsoon, South American Monsoon, and Europe-Mediterranean regions). *n* denotes the total number of speleothem records contributing to the age. The corresponding GICC05/GICC05modelxt age (3) is shown, including the maximum counting error for the annual layer-counted section (back to 60,000 years B.P.). No errors are shown beyond 60,000 years B.P. because the uncertainty on this section of the chronology (GICC05modelxt) is not quantified (32). The age difference between the ice core and composite speleothem ages is given as both a percentage and absolute difference. The ages in GICC05 are from (3) and were identified in a 20-year resolution multicore ice-core dataset. All ages are reported in years B.P. (before 1950).

Interstadial number	Combined speleothem SIOC19 age	<i>n</i>	GICC05/GICC05modelxt age	Age difference (%)	Age difference (years)
1e	14,646 ± 34*	9	14,642 ± 186	−0.03	−4
3	27,822 ± 72*	10	27,730 ± 832	−0.33	−92
4	29,117 ± 189†	4	28,850 ± 898	−0.93	−267
5.1	30,980 ± 103*	7	30,790 ± 1024	−0.62	−190
5.2	32,477 ± 115*	7	32,450 ± 1132	−0.08	−27
6	33,699 ± 174*	5	33,690 ± 1212	−0.03	−9
7a	34,884 ± 130†	3	34,830 ± 1293	−0.16	−54
7c	35,342 ± 90*	6	35,430 ± 1321	0.25	88
8c	38,044 ± 88*	11	38,170 ± 1449	0.33	126
9	40,251 ± 81*	6	40,110 ± 1580	−0.35	−141
10	41,431 ± 119*	10	41,410 ± 1633	−0.05	−21
11	43,185 ± 64*	7	43,290 ± 1736	0.24	105
12c	47,071 ± 145†	6	46,810 ± 1912	−0.56	−261
13a	49,015 ± 154*	5	49,010 ± 2021	−0.01	−5
13c	49,439 ± 106†	7	49,230 ± 2031	−0.43	−209
14c	53,713 ± 124*	6	53,910 ± 2289	0.37	197
15.2	55,770 ± 98†	5	55,750 ± 2368	−0.04	−20
16.1a	57,746 ± 311†	2	57,870 ± 2492	0.21	124
16.1c	57,910 ± 293†	2	57,990 ± 2497	0.14	80
16.2	58,315 ± 288†	2	58,230 ± 2511	−0.15	−85
17.1a	58,870 ± 346†	2	58,730 ± 2540	−0.24	−140
17.1c	59,199 ± 241†	4	59,030 ± 2557	−0.29	−169
17.2	59,489 ± 173*	6	59,390 ± 2569	−0.17	−99
18	64,596 ± 211†	6	64,050	−0.85	−546
19.2	71,594 ± 230†	5	72,290	0.96	696
20c	75,583 ± 248*	7	76,390	1.06	807
21.1e	84,621 ± 971†	3	84,710	0.11	89
22g	90,386 ± 617†	2	89,990	−0.44	−396
24.1a	106,328 ± 325†	2	106,170	−0.15	−158
24.1c	106,918 ± 204†	3	106,700	−0.20	−218
24.2	108,254 ± 235†	2	108,230	−0.02	−24
25c	115,310 ± 209†	2	115,320	0.01	10

*All three regions are represented by the corresponding speleothem records.

†One region is represented by the corresponding speleothem records.

‡Two of the three regions are represented by the corresponding speleothem records.

We do not detect any systematic GICC05-to-SIOC19 age difference that could be linked to either background climate state or interstadial duration, amplitude, and rate of warming (fig. S6, C to F).

Because the interstadial onsets captured by the speleothems are sufficient in number and encompass the entire range of interstadial types for the last glacial period (for example, including longer-duration events and rebound-type events), they can therefore be regarded as a representative sample of the population of all last glacial interstadial onsets. On the basis of this assumption, we quantified the GICC05-to-SIOC19 age difference across all 23 events younger than 60,000 years using robust regression (GICC05 versus SIOC19 ages) (34). The regression yields a slope of 1.000 and a y intercept (mean age difference) of -48 ($-160/+240$) years (fig. S6B), providing compelling evidence that the speleothem and ice-core ages agree to within, at a maximum, a couple of centuries. These radiometric-age com-

parisons provide strong support for the accuracy of both the GICC05 and GICC05modelext chronologies throughout the duration of the last glacial period and suggest that the quoted uncertainties are too conservative. On the basis of our time points for the time period, we do not find strong support for GICC05 being too young over the period 45,000 to 15,000 years B.P. as is suggested by linking the GICC05 and U-Th time scales through the use of cosmogenic radionuclides (51), nor can we confirm the 0.63% counting bias correction found by correlation to Hulu cave U-Th dates during the dating of the West Antarctic Ice Sheet (WAIS) Divide ice core (24). Such differences may arise because of the methodological approach, including the choice of detrital-thorium correction and depth-age modeling (34). The effect of compiling multiple speleothem records is also likely to be substantial, in that the influence of potentially younger or older individual records is not as dominant. Our study also looks across the whole of the last

glacial period rather than subsections or the individual temporal span of selected records.

Although overall synchrony between individual monsoon regions, and between Europe and the monsoon regions, has been assumed [for example, (25, 41)], it has yet to be fully tested quantitatively. To examine this further, we undertook two steps. First, we computed the EWM (and 95% uncertainties) of the differences between the Asian and South American Monsoon speleothem age estimates for each interstadial onset, where a 0 mean would indicate perfect synchrony (table S3A). We included only the interstadial onsets in which both monsoon regions were represented by two or more speleothem records (the case for eight interstadials) on the grounds that multiple speleothem records for an onset provide a more robust age estimate. The results give an EWM age difference (ASM minus SAM) of 19 ± 113 years (MSWD = 0.26) [Fig. 4A, blue probability density function (PDF) curve, and table S3A]. This strong degree of synchrony suggests that the ASM and SAM regions can be treated as a single category to yield “monsoon” EWM ages (and 95% uncertainties). Second, we then determined the age difference between the EM and combined monsoon regions for the same onsets. There are six interstadials (3, 8c, 9, 10, 12c, and 13c) for which there are at least two records from the EM region and at least two monsoon records (regardless of their origin). The mean onset-age difference and uncertainty (EM minus Monsoon) is 25 ± 84 years (MSWD = 0.90) (Fig. 4A, red PDF curve, and table S3B). This implies that the 95th percentile EM lead and EM lag over the monsoon regions is 109 and 59 years, respectively.

Comparison with model output and implications

We then compared the regional phasing of the onset of interstadials as indicated by the speleothems with modeling simulations and previous research (12, 13, 15, 23) to attain an overall picture regarding global climate teleconnections during an abrupt interstadial onset associated with AMOC recovery. We used long-term North Atlantic hosing experiments (52), in which a ~ 0.2 sverdrup freshwater flux was released to the Ruddiman Belt for 1000 years to mimic a cold “stadial” climate state. As the freshwater flux is removed at the 501st model year (1st model year in Fig. 4), the AMOC starts increasing instantly and reaches its “interstadial” state within a century (Fig. 4). We acknowledge that other forcing factors (such as ice-sheet height and atmospheric CO_2) are also able to mimic DO-type abrupt AMOC changes (19, 20), and freshwater flux may not be a realistic forcing factor for all interstadials (for example, those that follow non-Heinrich stadials). However, freshwater forcing is the ideal surrogate to evaluate the role of AMOC changes on the

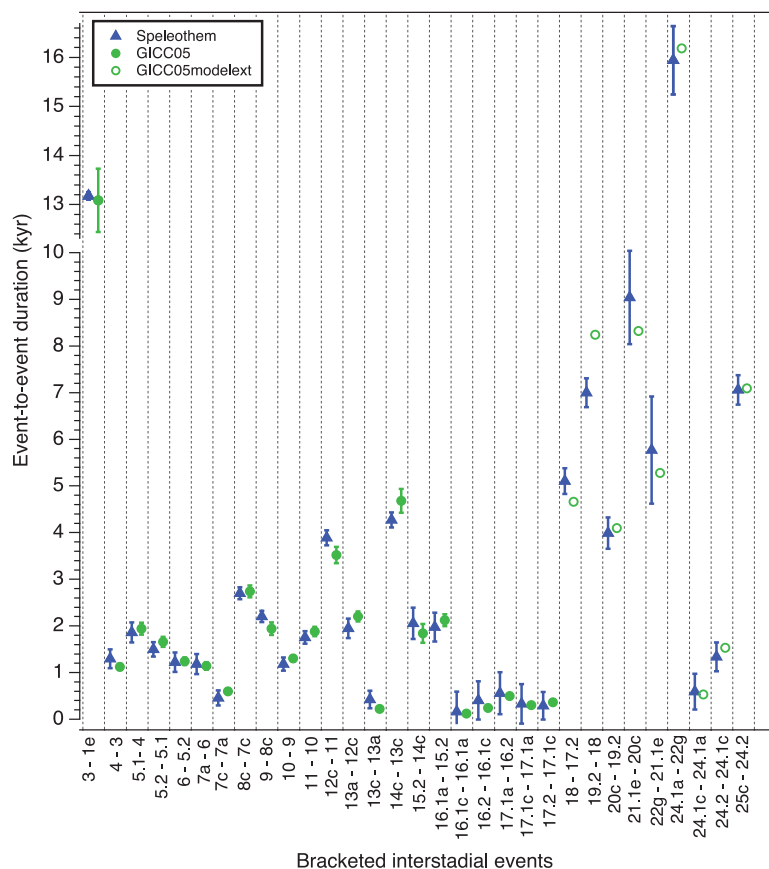
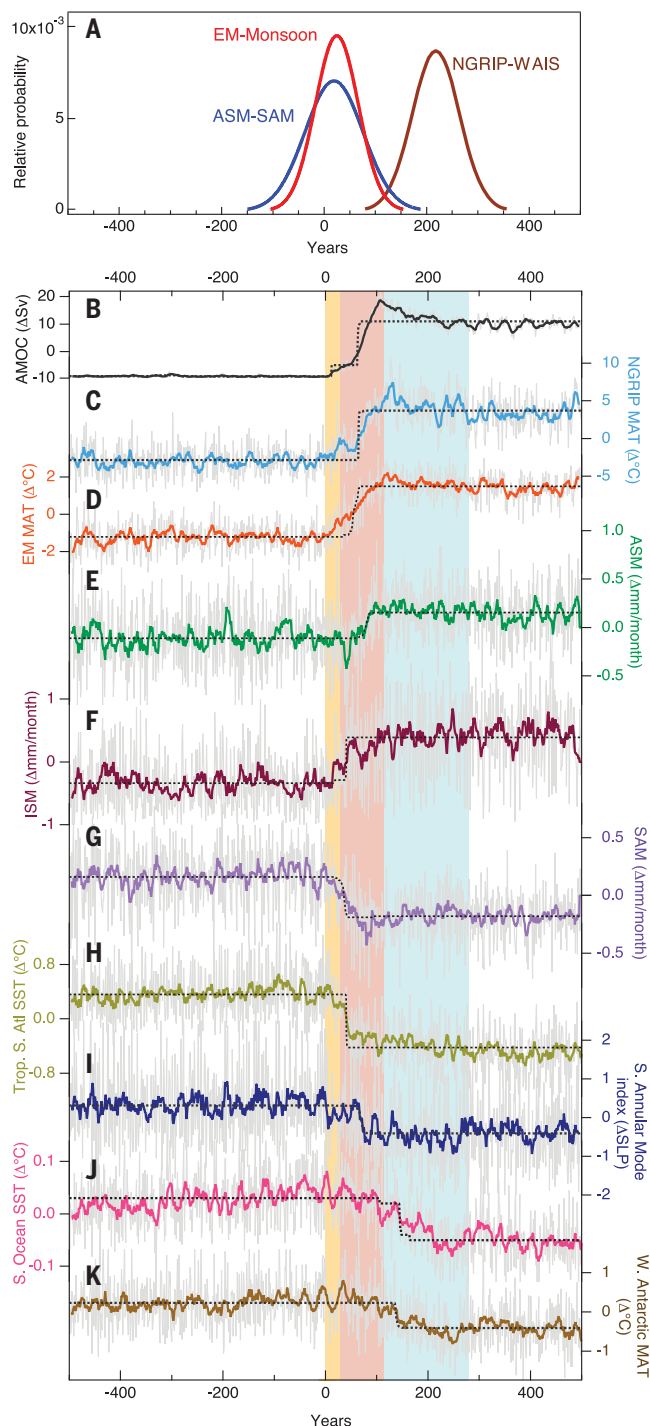


Fig. 3. Years between the onset of consecutive interstadials. The time interval between the onset of consecutive interstadials in the GICC05/GICC05modelext chronology is compared with the corresponding interval based on the SIOC19 ages. Error bars represent the 2σ age uncertainties on the interval. For the GICC05 chronology, this was calculated as the change in the accumulated layer-counting uncertainty between events (fig. S1); errors are not shown for events within the GICC05modelext section because they are not quantified (32). For the speleothems, the error bar is the uncertainty of the consecutive SIOC19 age estimates in quadrature. The timing between interstadial onsets is shown for those interstadials demonstrated to be synchronous in the speleothems, including estimates based on data from only one region.

Fig. 4. Spatial synchrony of climate changes during interstadial onsets. (A) PDFs of spatial age offsets between the two monsoon regions (ASM minus SAM, blue) and between Europe and both monsoon regions (EM minus Monsoon, red) based on composite speleothem data (table S3). Also shown (brown) is the previously determined offset between the WAIS ice core $\delta^{18}\text{O}$ and NGRIP $\delta^{18}\text{O}$ (13). (B to K) Simulated climate changes in a DO-type hosing simulation, each expressed as anomalies from the mean of the simulation (fig. S6 and table S4) (34). (B) AMOC index (defined as maximal meridional stream function at the water depth of 1500 to 3000 m north of 45°N in the North Atlantic). (C) Annual mean surface air temperature over the NGRIP drilling site (NGRIP MAT). (D) Annual mean surface air temperature over the Europe-Mediterranean region (30°N to 45°N, 25°E to 40°E) (EM MAT). [(E) to (G)] Mean annual precipitation over the (E) Eastern Asian (20°N to 30°N, 108°E to 120°E), (F) Indian (25°N to 35°N, 75°E to 85°E), and (G) South American regions (5°S to 10°S, 30°W to 75°W). (H) Annual mean sea-surface temperature (SST) in the tropical-subtropical South Atlantic (5°S to 30°S, 60°W to 10°E). (I) Southern Annular Mode index reflecting changes in sea-level pressure from 20°S to 90°S. (J) Annual mean SST in the Atlantic sector of the Southern Ocean (55°S to 75°S, 60°W to 10°E). (K) MAT over West Antarctica (75°S to 82°S, 90°W to 135°W) taken to be representative of the WAIS ice core site. Yellow shading indicates the period of fast atmospheric response in which the Atlantic ITCZ starts migrating toward the subtropical North Atlantic. Pink shading indicates the period of further strengthening of the AMOC and NH warming. Blue shading indicates the start of the AMOC interstadial mode. The gray series are raw model outputs, and the color series are 11-year running averages. The black dotted lines are model series derived from the application of a Bayesian, least-squares change-point analysis (58) to each raw time series. For (B) to (K), the x axis shows model years, with year 1 reset to the year of the removal of the freshwater flux (34). For (A), the year scale refers to the age offset for each PDF. The NGRIP-WAIS PDF curve is therefore in its approximate correct position with respect to the model-year scale in (B) to (K) (34).



in sea-level pressure from 20°S to 90°S. (J) Annual mean SST in the Atlantic sector of the Southern Ocean (55°S to 75°S, 60°W to 10°E). (K) MAT over West Antarctica (75°S to 82°S, 90°W to 135°W) taken to be representative of the WAIS ice core site. Yellow shading indicates the period of fast atmospheric response in which the Atlantic ITCZ starts migrating toward the subtropical North Atlantic. Pink shading indicates the period of further strengthening of the AMOC and NH warming. Blue shading indicates the start of the AMOC interstadial mode. The gray series are raw model outputs, and the color series are 11-year running averages. The black dotted lines are model series derived from the application of a Bayesian, least-squares change-point analysis (58) to each raw time series. For (B) to (K), the x axis shows model years, with year 1 reset to the year of the removal of the freshwater flux (34). For (A), the year scale refers to the age offset for each PDF. The NGRIP-WAIS PDF curve is therefore in its approximate correct position with respect to the model-year scale in (B) to (K) (34).

spatial phasing of climate changes (53) because changes in ice-sheet height and atmospheric CO_2 will exert additional regional and global climate responses that are not associated with AMOC changes (fig. S7) (19, 20, 34).

The climate response to AMOC reinvigoration expressed in the modeled regional climate anomalies is consistent with the observations from the regional array of speleothems (Fig. 4). Furthermore, the degree of synchrony shown by the speleothems between the ASM and SAM and between EM and the combined Monsoon agrees with the timing of modeled regional precipitation and temperature anomalies in the Monsoon and EM, respectively (Fig. 4). On the basis of this, we can summarize the pattern of climate changes during the interstadial onsets in three stages (Fig. 4, B to K). At the beginning, AMOC recovery (Fig. 4B) will lead to an initial warming anomaly in the northern North Atlantic (Fig. 4, C and D), causing thermal asymmetry between the North and South Atlantic. This would drive a fast atmospheric response in which the Atlantic ITCZ starts migrating toward the subtropical North Atlantic (Fig. 4, yellow shading). As the AMOC is strengthened further, the entire NH warms, pulling the ITCZ further northward globally (Fig. 4, E to H, pink shading). This strengthening and weakening of the ASM and SAM, respectively, leads to the observed respective decrease and increase in speleothem $\delta^{18}\text{O}$ records (37–39). Over Europe and the Mediterranean, warming drives increased speleothem $\delta^{18}\text{O}$ at sites where the local rainfall $\delta^{18}\text{O}$ is most sensitive to mean atmospheric temperature changes (40, 41) and lower speleothem $\delta^{18}\text{O}$ at sites where warmer ocean temperatures drive higher rainfall amounts (44). Meanwhile, Antarctica starts perceiving the northern signal by way of the atmospheric bridge (6, 22). This is evident in the model output of changes to the Southern Annular Mode index, a proxy for meridional changes in the position of the westerlies: As interstadial onset commences, and the ITCZ shifts northward, the index becomes more negative over a similar time frame to the ITCZ changes (Fig. 4I). This is consistent with a northward migration of the southern mid-latitude westerlies and supports the conceptual model proposed by (23) to explain the changes in the d -excess of Antarctic ice across the warming transitions. Once the AMOC reaches interstadial mode (Fig. 4, blue shading), atmospheric circulation assumes its interstadial mean state, leading to a cascade of processes that alters heat transport across the ACC, leading to an Antarctic temperature decrease (Fig. 4K) (6) that lags the AMOC recovery [and Greenland warming (Fig. 4C)] by approximately 200 ± 100 years (Fig. 4A, brown PDF curve) (13). The compilation of precise speleothem records now lends support to this pattern of changes

and provides crucial empirical evidence for the synchronous response to these changes in the mid-latitudes.

The SIOC19 age estimates (Table 1) constitute the most precise, absolute dating of interstadial onsets yet produced. Using the full range of interstadial types across the whole last glacial period, and incorporating as much speleothem information as possible from the target regions, we demonstrate that synchronous climate changes (within a century) occurred between Europe and the Asian and South American monsoon domains at the onset of interstadials. A similar level of synchrony is observed between Europe and both monsoon regions (Fig. 4A). Given the proximity of Greenland to EM, compared with the proximity of the EM to both monsoon domains, it is highly likely that such synchrony extends across the North Atlantic to Greenland. A recent comparison, based on four speleothem records, of interstadials in the latter part of the last glacial (<45,000 years B.P.) also shows the timing of the onset to be synchronous within uncertainties (57). Our findings are also consistent with a previous study of a single interstadial, where synchrony between tropical (using methane as a proxy) and Greenland temperatures took place within 24 years (54). Thus, our study strongly supports the long-held, but theretofore untested, assertions of practically synchronous high-latitude-to-tropical climate changes during abrupt interstadial onsets (24, 25, 28). Recent results from Antarctica suggest that changes in the source location of moisture reaching the ice sheet were abrupt and occurred within decades of DO warmings (and coolings) in Greenland (22), implying an inter polar atmospheric teleconnection. Our results provide crucial, independent supporting evidence of near-synchronous atmospheric propagation of abrupt climate changes during interstadial onsets that were global in scale by providing spatial detail between the NH mid-latitudes and the SH subtropics.

The prospect of future abrupt climate change due to anthropogenic forcing, and the importance of reliable climate-model predictions, has brought past abrupt changes, such as DO events, into sharp focus in recent years. Our study provides precise chronological information on the timing of climate changes during stadial-interstadial transitions, reveals the wide-

spread synchrony of their response, and provides radiometric validation and constraint for refining ice-core chronologies. The triggering mechanism for DO events (and the associated AMOC changes), however, remains an open question, and discriminating between external forcing (such as ice-sheet height, greenhouse gases, meltwater, and volcanism) (18–20, 55, 56) and “self-oscillation” mechanisms (57) as drivers of AMOC-mode transitions rests with future studies that take advantage of state-of-the-art Earth-system models and well-dated, high-resolution palaeoclimate records.

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SUPPLEMENTARY MATERIALS

science.sciencemag.org/content/369/6506/963/suppl/DC1
Materials and Methods
Figs. S1 to S8
Tables S1 to S4
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Data Files S1 to S7

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Synchronous timing of abrupt climate changes during the last glacial period

Ellen C. Corrick, Russell N. Drysdale, John C. Hellstrom, Emilie Capron, Sune Olander Rasmussen, Xu Zhang, Dominik Fleitmann, Isabelle Couchoud, and Eric Wolff

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All together then

Many geographically dispersed records from across the globe reveal the occurrence of abrupt climate changes, called interstadial events, during the last glacial period. These events appear to have happened at the same time, but the difficulty of determining absolute dates in many of the records have made that proposition difficult to prove. Corrick *et al.* present results from 63 precisely dated speleothems that confirm the synchrony of those interstadial events. Their results also provide a tool with which to validate model simulations of abrupt climate change and calibrate other time series such as ice-core chronologies.

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