

Supporting Information for
**Interstadial diversity of East Asian summer monsoon linked to changes of the
Northern westerlies**

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Supplementary Notes

1.1 Speleothem $\delta^{18}\text{O}$ interpretation

Speleothem $\delta^{18}\text{O}$ ($\delta^{18}\text{O}_c$) inherits the atmospheric precipitation oxygen isotope composition ($\delta^{18}\text{O}_p$) via cave drip water, usually exhibiting similar patterns on millennial-to-orbital timescales from Northeast to Southwest China¹⁻⁵. During the pre-monsoon (March-April-May) season, the Asian summer monsoon (ASM) region receives precipitation from proximal sources, and thus precipitation is characterized by higher values of $\delta^{18}\text{O}_p$ (Supplementary Fig. 11a–c). During the summer monsoon season from June to September, moisture from remote sources increases (Fig. 1d; Supplementary Fig. 12b), resulting in lower $\delta^{18}\text{O}_p$ ^{3,6} (Supplementary Fig. 11a–c). Modern cave monitoring work in the monsoonal region also demonstrates seasonal isotopic variability in cave drip water $\delta^{18}\text{O}$ (e.g., ref. 7,8). On interannual timescale, a lower $\delta^{18}\text{O}_c$ indicates a stronger monsoon intensity, with a strengthened large-scale monsoonal circulation and a northward displacement of tropical rain belt^{3,9}, supported by modeling simulations¹⁰. These observations imply that the $\delta^{18}\text{O}_c$ records from the Indian summer monsoon (ISM) and East Asian summer monsoon (EASM) region on multiple timescales can be broadly interpreted in a framework of overall changes in the upstream convection, moisture source dynamics, and overall monsoon intensity/circulation^{1,3,4,10-15}.

The South American summer monsoon (SASM) season is in austral summer (December to February). During the SASM season, stronger low-level winds transfer

moisture from the tropical Atlantic across the Amazon basin, as far as southern South America¹⁶. At centennial-to-orbital timescales, lower (higher) Paraíso $\delta^{18}\text{O}_\text{c}$ is generally considered as a proxy for stronger (weaker) SASM intensity (overall monsoon circulation) and associated rainfall^{1,17}.

1.2 Greenland ice-core records and the GICC05 chronology

The Greenland ice-core $\delta^{18}\text{O}$ records utilized in this study include NGRIP, GISP2 and GRIP records¹⁸⁻²². Greenland ice-core $[\text{Ca}^{2+}]$ records include NGRIP, GISP2, GRIP and NEEM records²³⁻²⁶. We refer to the Greenland ice-core records ($\delta^{18}\text{O}$ and $[\text{Ca}^{2+}]$) on the GICC05 chronology²⁷⁻³⁰ with ages reported in thousands of years before 1950 CE (ka BP) or in years before 1950 CE (a BP). We take the reported “Maximum Counting Error” as 2σ (95%) age uncertainty and use the latest nomenclature and timing of interstadials/stadials from ref. 19.

1.3 Uncertainty of the U-Th-based Greenland ice-core chronology

Our study established three tie points between Greenland ice-core and speleothem records (Supplementary Fig. 3; Supplementary Table 1). We combined the maximum tie-point uncertainty in the SN23-1 record (110 years; Supplementary Table 1), the maximum tie-point uncertainty in the ice-core records (110 years; Supplementary Table 1), and the uncertainty regarding the causal correlation between Greenland ice-core $[\text{Ca}^{2+}]$ and speleothem $\delta^{18}\text{O}$ [estimated to 20 years; following ref. 31], resulting in a combined uncertainty of 160 years (2σ). It is worth noting that the tie point uncertainty of the SN23-1 record is already a combined uncertainty that includes the age model uncertainty and the change point uncertainty (Supplementary Table 1; see Methods).

1.4 Correlation of Antarctic ice-core gas chronology to the improved Greenland chronology

During the last glacial period, the atmospheric CH_4 was primarily associated with tropical and subtropical wetland emissions^{32,33}. Greenland temperature (as reflected by ice-core $\delta^{15}\text{N}$) leads CH_4 concentrations by 4.5^{+21}_{-24} years during the onset of the Bølling transition³⁴, while other studies suggest that $\delta^{15}\text{N}$ -based Greenland temperature leads

CH₄ increases by an average of 56 ± 38 years or 25–70 years over several DO events^{35,36}. To combine these scenarios, Buizert et al.³⁷ proposed a 25-year lag between the midpoints in the West Antarctic Ice Sheet Divide ice core (WDC) CH₄ and the North Greenland Ice-core Project (NGRIP) $\delta^{18}\text{O}$ transitions, allowing for a direct alignment of WDC CH₄ and NGRIP $\delta^{18}\text{O}$. Based on a comparison of the NGRIP-based GICC05 ice-core chronology with the U-Th-dated Hulu cave speleothem record, a scaling of GICC05 was proposed, and therefore, $1.0063 \times \text{GICC05}$ was applied to the WDC gas record over the interval 67.8–31.2 ka³⁷. This scaling yields a ca. +350-year shift at 55 ka BP for GICC05, which agrees with the timescale proposed in this study (+230-year shift) within uncertainty.

The match points between WDC CH₄ and NGRIP $\delta^{18}\text{O}$ on their original chronologies (WD2014 and GICC05, respectively) are shown in [Supplementary Table 2](#). In order to obtain a common chronology and given that GICC05 has been shifted by +230 years in this study, the WD2014 gas chronology should be adjusted accordingly during our studied time interval. Notably, the time lag between the midpoints in the WDC CH₄ and NGRIP $\delta^{18}\text{O}$ transitions has recently been revised to be 29 years³⁸. Taken these factors into consideration, we calculated the new match point ages in WDC CH₄ and NGRIP $\delta^{18}\text{O}$ ([Supplementary Table 2](#); [Supplementary Fig. 3e](#)), thus obtaining the adjusted Antarctic ice-core gas timescale in our studied time interval and are provided in [Supplementary Data 3](#).

1.5 Marine sediment records

Cariaco Basin and Arabian Sea sediment reflectance records were on the GICC05 chronology³⁹. In view of the improvement of the GICC05 chronology in our study, we shifted these records by +230 years ([Supplementary Fig. 4g, h](#)). For the North Atlantic Pa/Th record, the chronology remains unchanged ([Fig. 2a](#)).

1.6 Hemispheric imprints of other centennial-scale events

Another pair of centennial-scale events, DO-14d–e, has also been captured by several proxy records ([Supplementary Fig. 4d–j](#)) with an interhemispheric “monsoon seesaw pattern” ([Supplementary Fig. 4e, f, i](#)) in the context of ITCZ shifts northward

and then southward ([Supplementary Fig. 4h](#)). Intriguingly, a discrepancy among proxy records is apparent between different regions in terms of relative values between the cool phase of interstadial 14d and stadial 15.1. After the warm phase of DO-14e, the cool phase of DO-14d in Greenland ice-core $\delta^{18}\text{O}$ ([Supplementary Fig. 4d](#)) and $[\text{Ca}^{2+}]$ ([Supplementary Fig. 3b, c](#)), ASM speleothem ([Supplementary Fig. 4e, f](#)) and Arabian Sea reflectance ([Supplementary Fig. 4g](#)) records reached a level that is notably higher than Stadial 15.1 but lower than the adjacent interstadials. In contrast, the Cariaco reflectance record and the SASM speleothem record, derived from the tropical Atlantic sector ([Supplementary Fig. 1](#)) reached a level during the cool phase of 14d equivalent to stadial 15.1 value ([Supplementary Fig. 4h, i](#)). This difference may provide a clue for the potential sources of atmospheric CH_4 . During the last glacial period, atmospheric CH_4 was primarily associated with tropical and subtropical wetland emissions³²⁻³⁴, but it remains difficult to ascertain the exact sources of CH_4 emission. Proxy record comparison shows that the CH_4 change is quite different from proxy records in the tropical Atlantic sector ([Supplementary Fig. 4h-j](#)), which suggests that the CH_4 emission from tropical Atlantic wetlands might not be the major control or source of atmospheric CH_4 , at least for this time period.

Supplementary Tables

Supplementary Table 1 Age and combined uncertainties for speleothem and Greenland tie points

Speleothem record	Search time window	Age model uncertainty (years)	Change point uncertainty (years)	Age and combined uncertainty (a BP)	Passed the sensitivity tests? Yes (Y)/No (N)
SN23-1 $\delta^{18}\text{O}$	55,000–55,250	100	20	55,140 $\pm$ 100	Y
SN23-1 $\delta^{18}\text{O}$	55,600–55,750	90	50	55,660 $\pm$ 100	Y
SN23-1 $\delta^{18}\text{O}$	55,850–56,200	110	20	55,970 $\pm$ 110	Y
Below are age and combined uncertainties for Greenland ice-core [Ca^{2+}] tie points					
Tie-points in speleothem and uncertainty (a BP)	Ice cores	Search time window	Tie points in ice-cores and uncertainty (a BP)	Combined timing for ice-core tie-points	
55,140 $\pm$ 100	NGRIP		55,110 $\pm$ 30		
	GRIP	54,930–55,350	55,120 $\pm$ 40	55,119 $\pm$ 90	
	GISP2		55,100 $\pm$ 70		
	NEEM		55,130 $\pm$ 30		
55,660 $\pm$ 100	NGRIP	55,570–55,750	55,620 $\pm$ 10		
	GISP2	55,580–55,760	55,620 $\pm$ 10	55,623 $\pm$ 20	
	NEEM	55,590–55,730	55,630 $\pm$ 10		
55,970 $\pm$ 110	NGRIP		56,030 $\pm$ 40		
	GRIP	55,900–56,250	56,060 $\pm$ 30	56,025 $\pm$ 110	
	GISP2		55,980 $\pm$ 30		
	NEEM		56,080 $\pm$ 90		

In cases where the plus and minus errors are asymmetrical, the maximum errors were taken. The combined uncertainty is the square root of the age model uncertainty and change point uncertainty ([Methods](#)).

The combined timing is the weighted average age calculated based on IsoplotR software⁴⁰, and their uncertainties are obtained via quadratically combining the tie point uncertainties from the four ice cores.

All errors are 2σ , and we refer to Greenland and speleothem ages in years before 1950 CE (a BP). The errors that are less than 10 years are listed as 10 years. Greenland ice-core records are on the GICC05+230-year chronology.

Supplementary Table 2 Ages for new match point in NGRIP $\delta^{18}\text{O}$ and WDC CH_4

NGRIP match point on GICC05	WDC CH_4 match point on WD2014	NGRIP match point age on GICC05+230	Targeted WDC match point age	WD2014 gas age shift
54,164	54,480	54,394	54,365	−115
54,850	55,170	55,080	55,051	−119
54,940	55,261	55,170	55,141	−120
55,369	55,693	55,599	55,570	−123
55,737	56,063	55,967	55,938	−125
56,555	56,887	56,785	56,756	−131

NGRIP and WDC match point age on their original chronologies are from ref. 37.

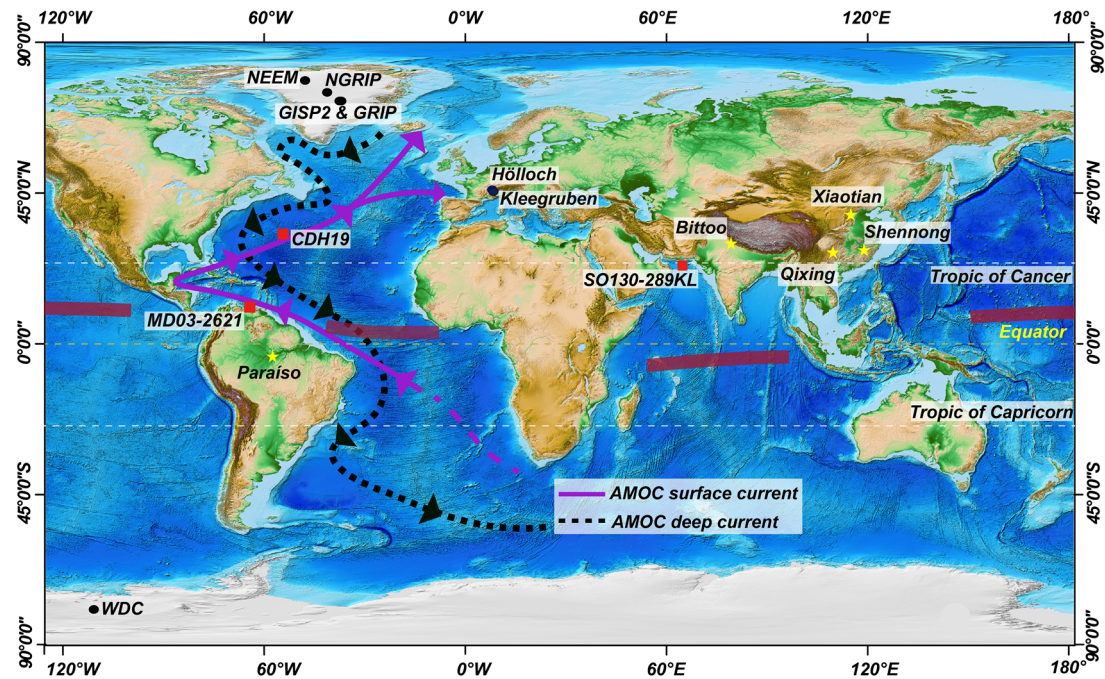
The targeted WDC match point age is equal to the value in the third column minus 29 (Supplementary Note 1.4).

Supplementary Table 3 Asian summer monsoon domain speleothem sample dating information

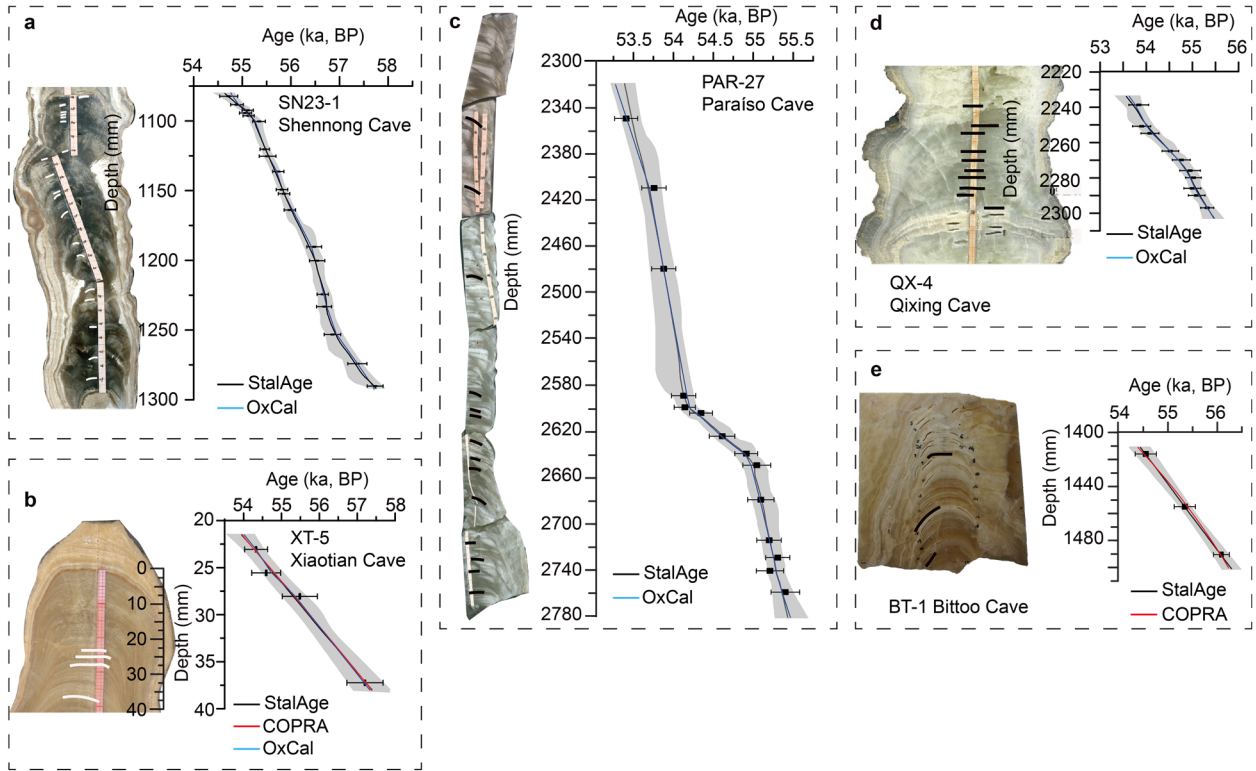
Speleothem name.	Growth length (mm)	Growth rate	Isotope resolution (mm)	Dating resolution (mm)	Average dating error (years)
SN23-1 (this study)	178	0.077 mm/a	0.8	10	140
YX-55 ⁴¹	25	0.011 mm/a	0.7	15	290
YX-51 ⁴¹	71.5	0.03 mm/a	0.7	17	210
Wu-26 ⁴²	150	0.063 mm/a	1.5	180	200
BT-2 ⁴³	69.5	0.029 mm/a	0.5	12	350

The growth length corresponds to 57–54.65 ka.

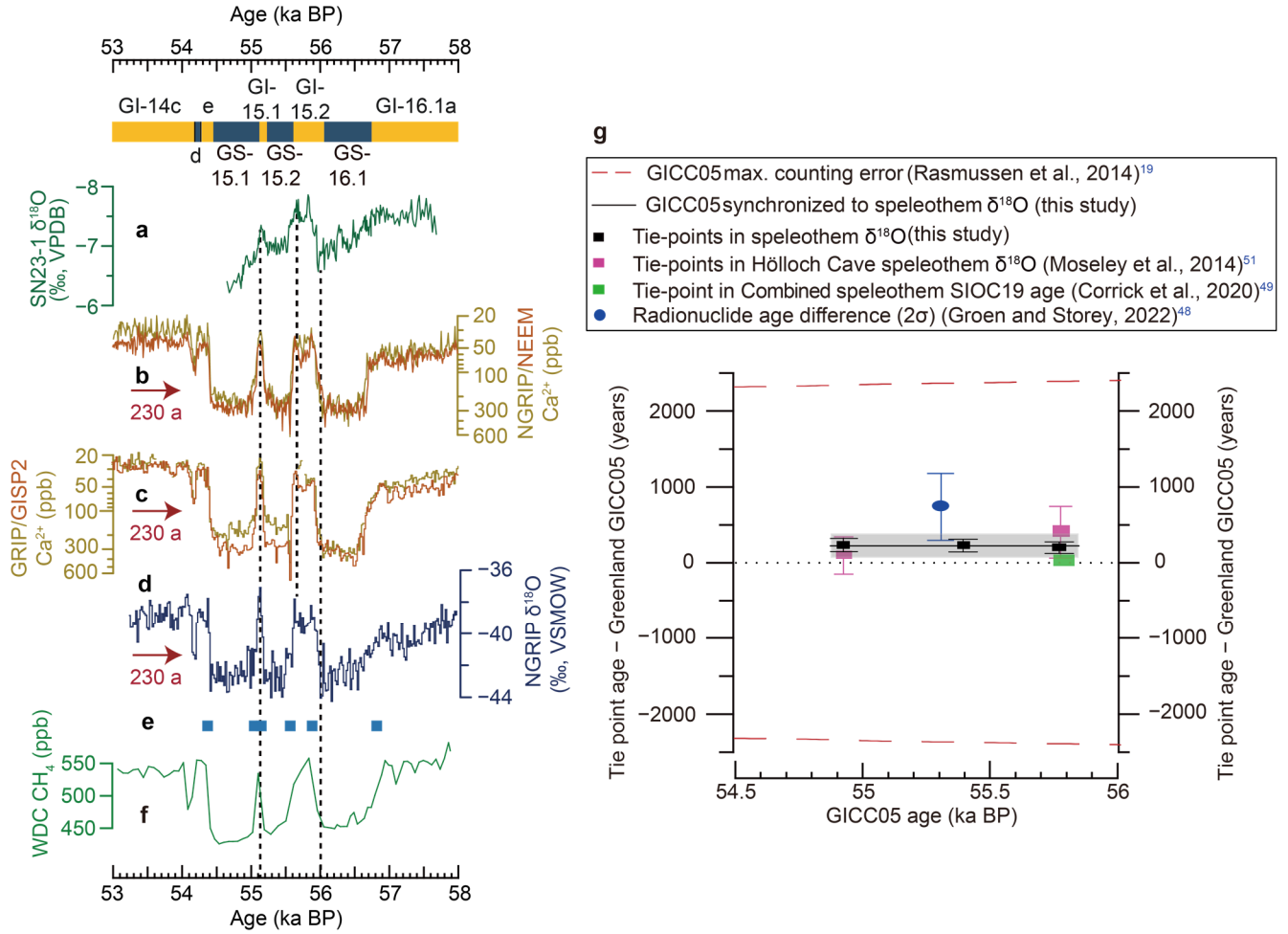
Supplementary Figures

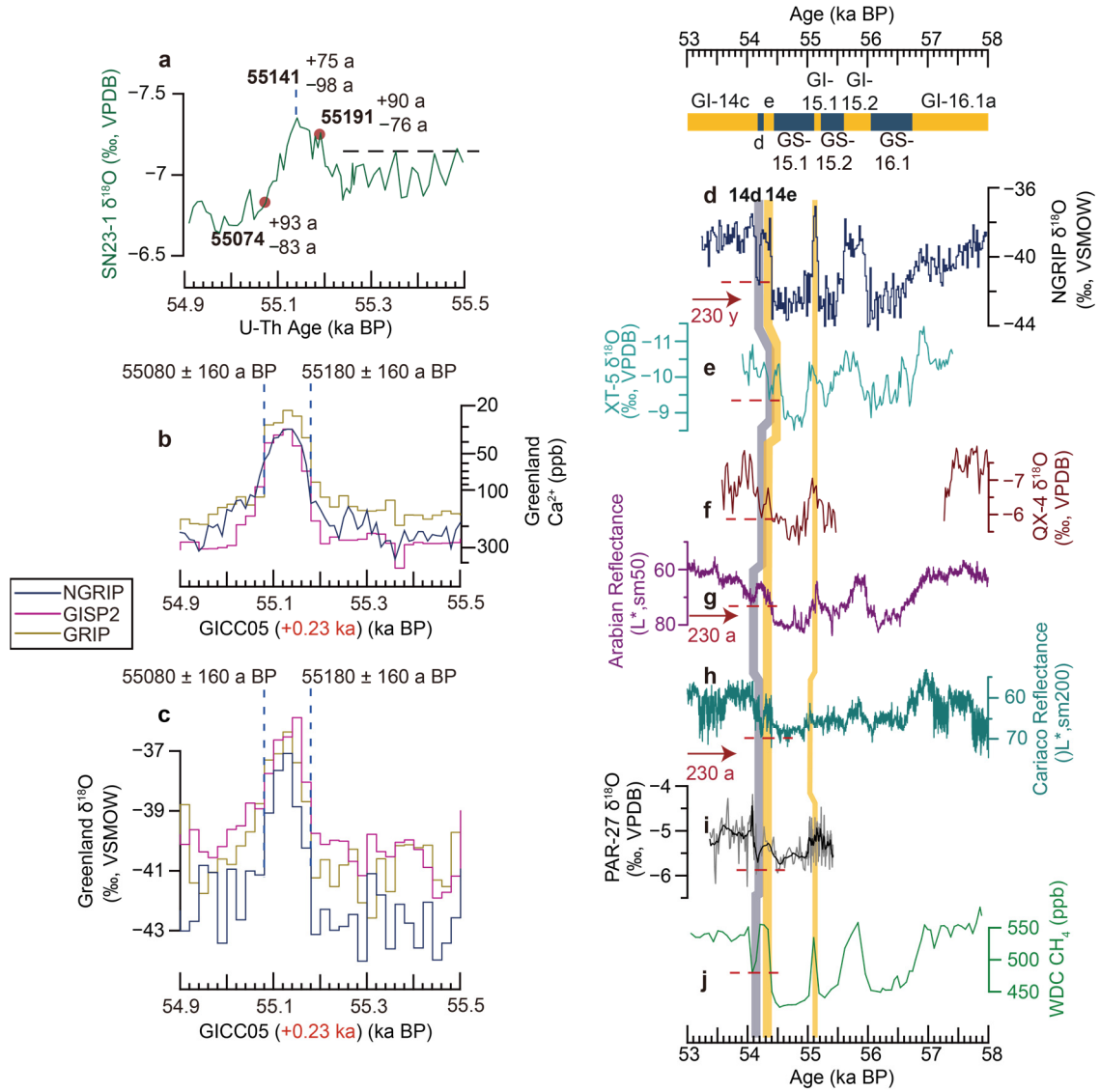


Supplementary Fig. 1. Proxy record locations. Yellow stars show locations of the caves presented in this study. Black dots indicate Greenland and Antarctic ice-core locations and other cave locations discussed in the main text. Red squares delineate marine sediment core locations. Red bars within the tropics indicate the annual-mean position of the Intertropical Convergence Zone (ITCZ) over the oceans⁴⁴. The map was obtained from <https://www.ngdc.noaa.gov/mgg/global/>. AMOC: Atlantic Meridional Overturning Circulation.

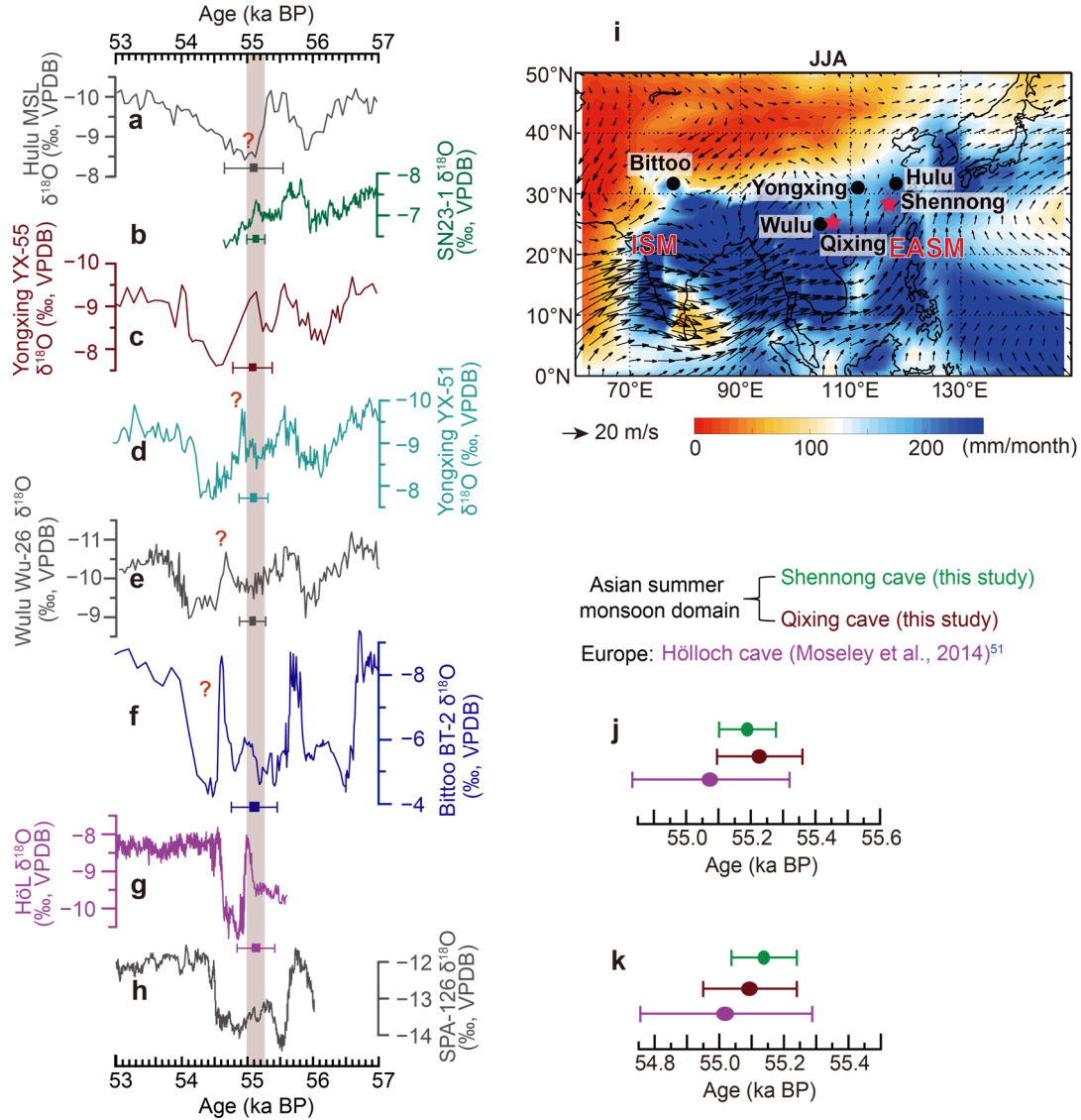


Supplementary Fig. 2. Age models and slabbed images of speleothems. (a)–(e) Scan images and age models of the speleothem SN23-1 (1,075–1,288 mm), XT-5 (the upper 40 mm), PAR-27 (2,270–2,780 mm), QX-4 (2,220–2,330 mm) and BT-1 (1,410–1,500 mm). The white/black bars indicate positions of subsamples for U-Th dating ([Supplementary Data 1](#)). The curves show modeled mean values using different age-modeling algorithms (color-coded)^{45–47}, the gray bands depict the 95.4% confidence interval using StalAge algorithm⁴⁵. Error bars on U-Th dates represent 2 σ analytical error.



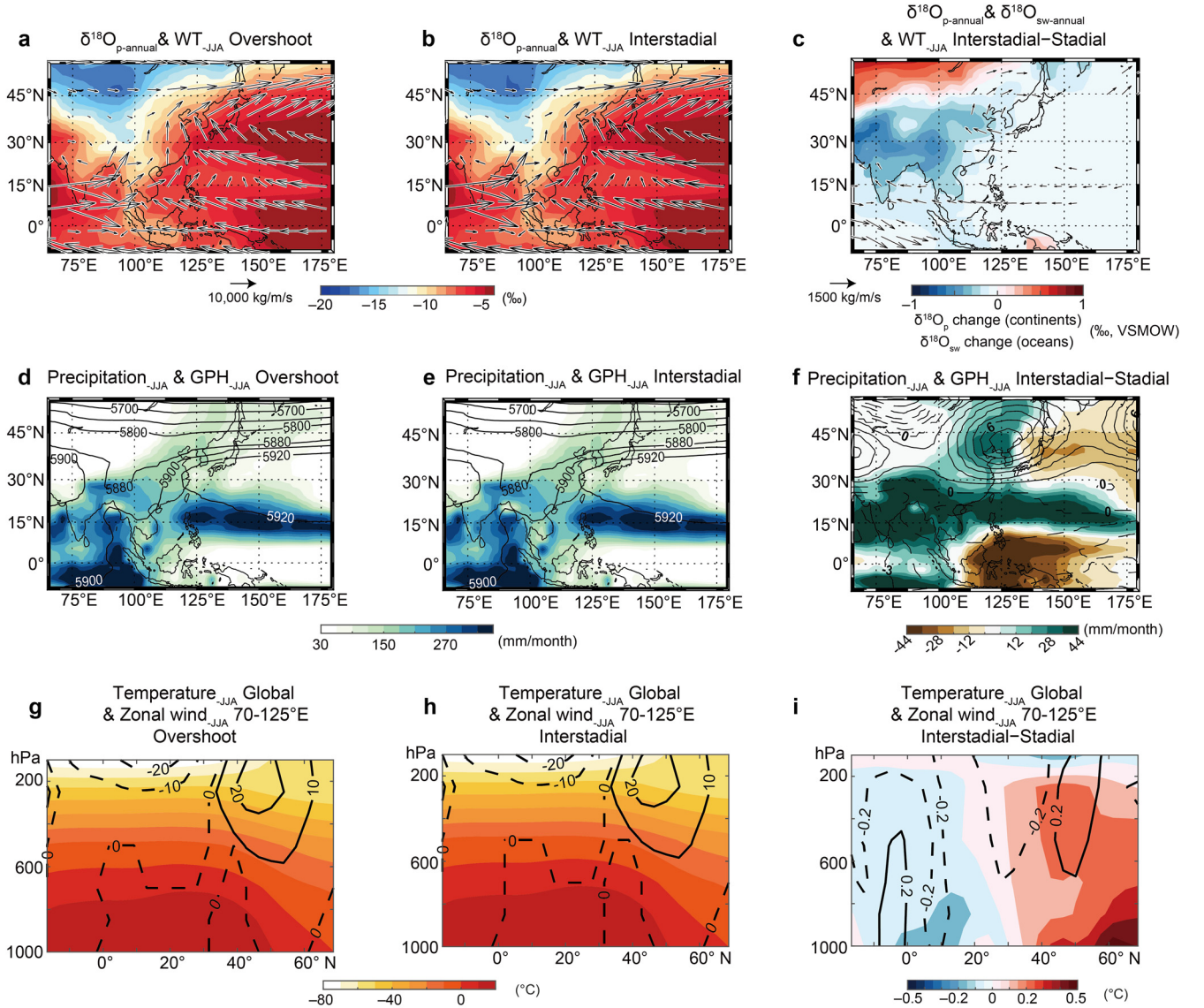


indicate the signal level within the cool phase 14d. Cave and core locations are shown in [Supplementary Fig. 1](#).



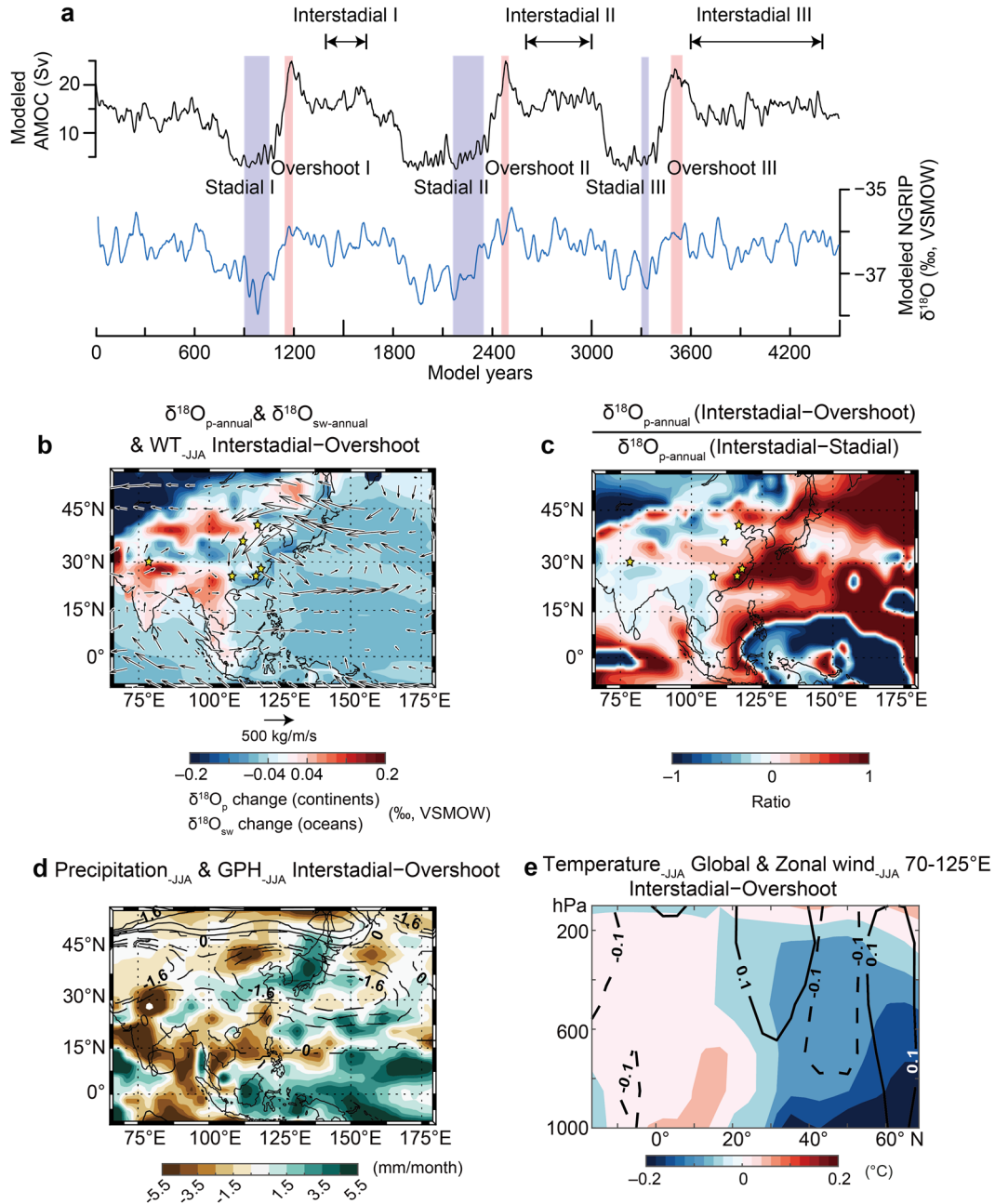
Supplementary Fig. 5. Comparison between Asian summer monsoon domain and European speleothem records. (a)–(f) Asian summer monsoon domain speleothem $\delta^{18}\text{O}$ records from Hulu cave⁵⁰, Shennong cave (this study), Yongxing cave⁴¹, Wulu cave⁴² in China and Bittoo cave in India⁴³, respectively. (g) and (h) European speleothem records from Hölloch cave⁵¹ and Klee gruben cave⁵², respectively. The vertical bar indicates the timing of DO15.1 as constrained by our SN23-1 record. Error bars in (a)–(g) show typical age model uncertainties (2σ) for each record (color-coded). (i) Asian summer monsoon domain cave locations. Background is the June-July-August (JJA) monthly spatial rainfall amount (color scale, 1980–2010 CE) and 850 hPa low-level wind (black arrows), data obtained from the Global Precipitation Climatology Project (GPCP) (<https://www.esrl.noaa.gov/psd/data/gridded/data.gpcp.html>). Red stars show locations for the caves presented in this study, while the black dots indicate locations of other caves mentioned in this study. For European cave locations refer to [Supplementary Fig. 1](#). (j) and (k) Comparison of the DO-15.1 timing between different regions. (j) Comparison of the DO-15.1 onset, where error bars correspond to different cave records (color-coded). SN23-1: $55,191 \pm 90$ a BP, QX-4: $55,225 \pm 135$ a BP, Höl: $55,068 \pm 250$ a

BP. **(k)** same as **(j)** but at the peak of DO-15.1. SN23-1: $55,141 \pm 100$ a BP, QX-4: $55,095 \pm 150$ a BP, Höl: $55,018 \pm 275$ a BP. Note that error bars in **(j)** and **(k)** are combined uncertainties that include the age-model uncertainty as well as change point uncertainty.



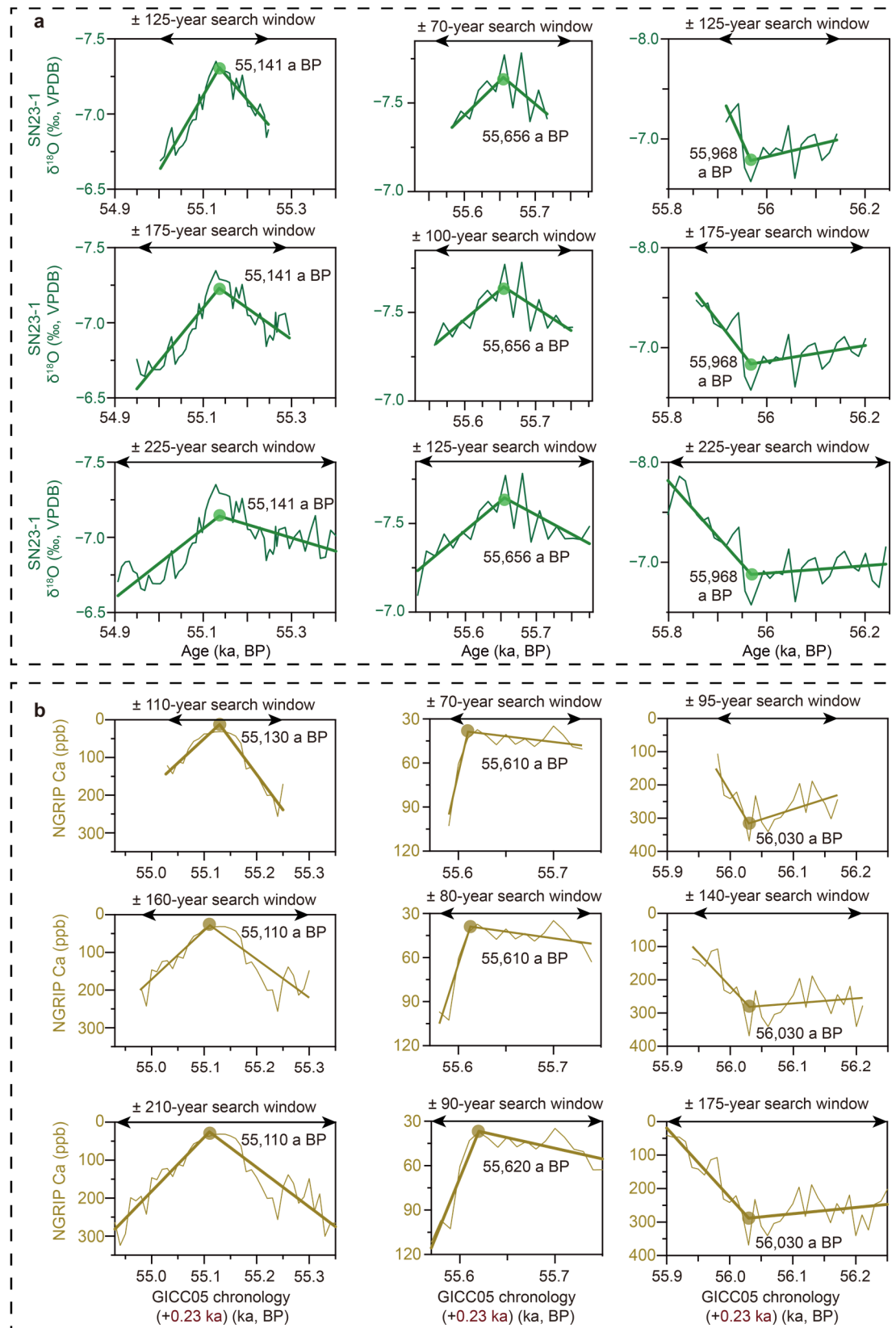
Supplementary Fig. 6. Simulated hydroclimate changes during DO warmings. (a) Composites of changes in mean annual $\delta^{18}\text{O}_p$ over continents and mean annual $\delta^{18}\text{O}_{sw}$ of sea-surface water (shaded, units: ‰) and water-vapor transport (WT) (vectors, units: kg/m/s) during short (simulated Overshoot phases) DO warmings. (b) Same as (a), but for long (simulated Interstitial phases) DO warmings. (c) Same as (a), but for differences between long interstitials and stadials (Interstitial minus Stadial phases). (d)–(f) are as same as (a)–(c), respectively, but for summer precipitation (shading, units: mm/month) and 500 hPa geopotential height (lines). (g)–(i) are as same as (a)–(c), respectively, but for globally zonal-averaged summer temperature (shadings, °C) and zonally averaged summer wind between 70° and 125° E (lines). The time periods of Stadial, Overshoot and Interstitial phases refer to Fig. 4a. Note that solid lines indicate positive values while dashed lines indicate negative values. DO: Dansgaard-Oeschger. GPH: Geopotential Height.

Results based on alternative definition of “interstadial” excluding the prior “overshoot” phase



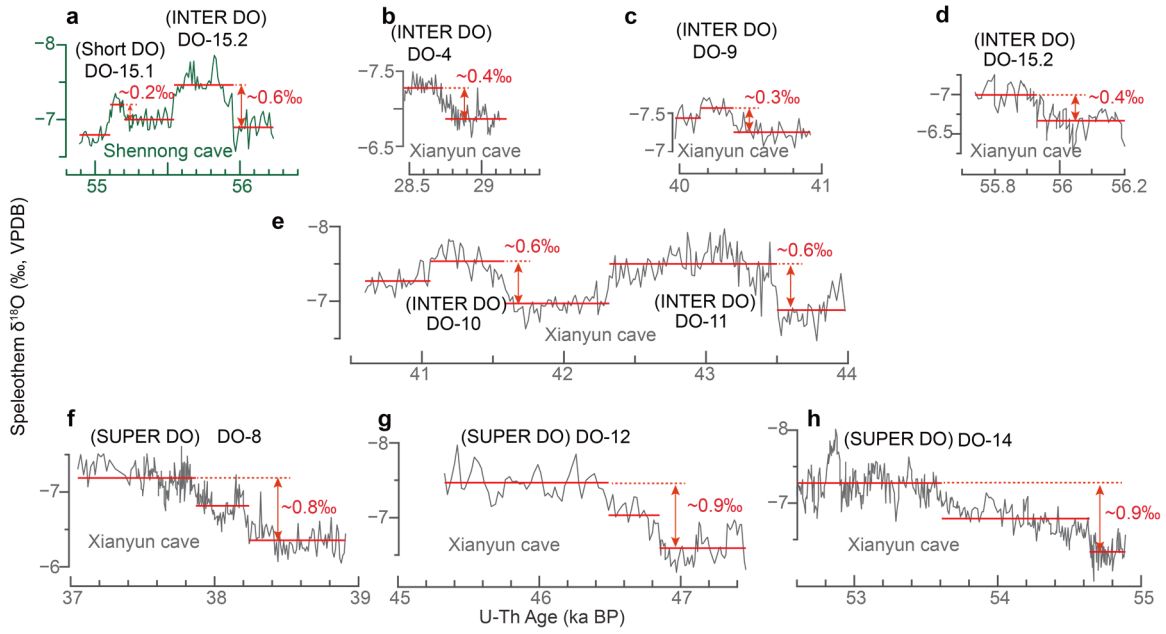
Supplementary Fig. 7. The results based on alternative definition of “interstadial” excluding the prior “overshoot” phase. (a) Modeled AMOC index and the NGRIP mean annual $\delta^{18}\text{O}_p$ (as same as in Fig. 4a). We classify three AMOC phases: Stadal phases, Overshoot phases and Interstadial phases. The Stadal and Overshoot phases are as same as in Fig. 4a, while the Interstadial phases used here do not include the “Overshoot” phases (Methods). (b) Composites of differences in mean annual $\delta^{18}\text{O}_p$ over continents and mean annual $\delta^{18}\text{O}$ of sea-surface water ($\delta^{18}\text{O}_{sw}$) over oceans (shaded, units: ‰) and vertical integrated water-vapor transport (WT) (vectors, units: kg/m/s) between long and short interstadials (Interstadial minus Overshoot phases). (c) The proportion of $\delta^{18}\text{O}_p$ changes between long and short interstadials (Interstadial minus Overshoot phases) in the long (Interstadial minus Stadal phases) DO warming events. Bittou, Qixing, Xiaotian, Shennong and Xinglong cave locations are indicated by the yellow stars (refer to Fig. 1d). (d) Simulated differences in precipitation (shading, units:

mm/month) and 500 hPa geopotential height (lines) between long and short interstadials (Interstadial minus Overshoot phases). The intervals between each line correspond to $\frac{8}{15}$. (e) is as same as in (d), but for globally zonal-averaged summer temperature (shadings) and zonal-averaged wind between 70° and 125° E (lines). Note that solid lines indicate positive values while dashed lines indicate negative values. DO: Dansgaard-Oeschger. AMOC: Atlantic Meridional Overturning Circulation.

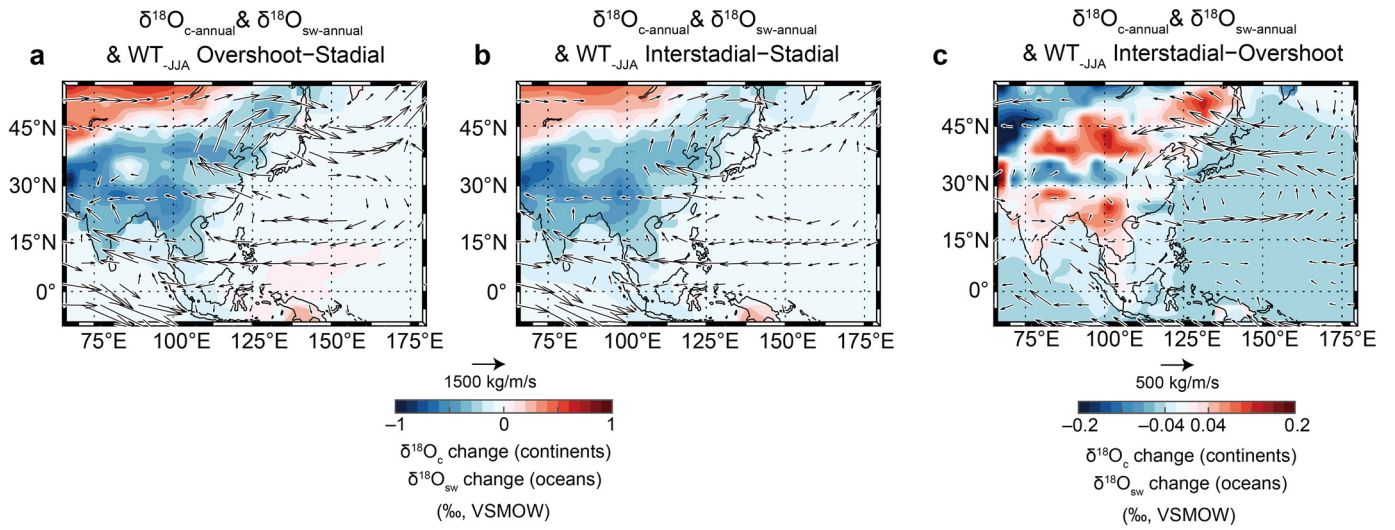


Supplementary Fig. 8. Sensitivity tests of speleothem and ice-core records. The bold ramps are defined by the BREAKFIT algorithm⁵³ over three search windows of different widths (Methods). The change points are shown and the selected search windows are indicated by the horizontal black arrows.

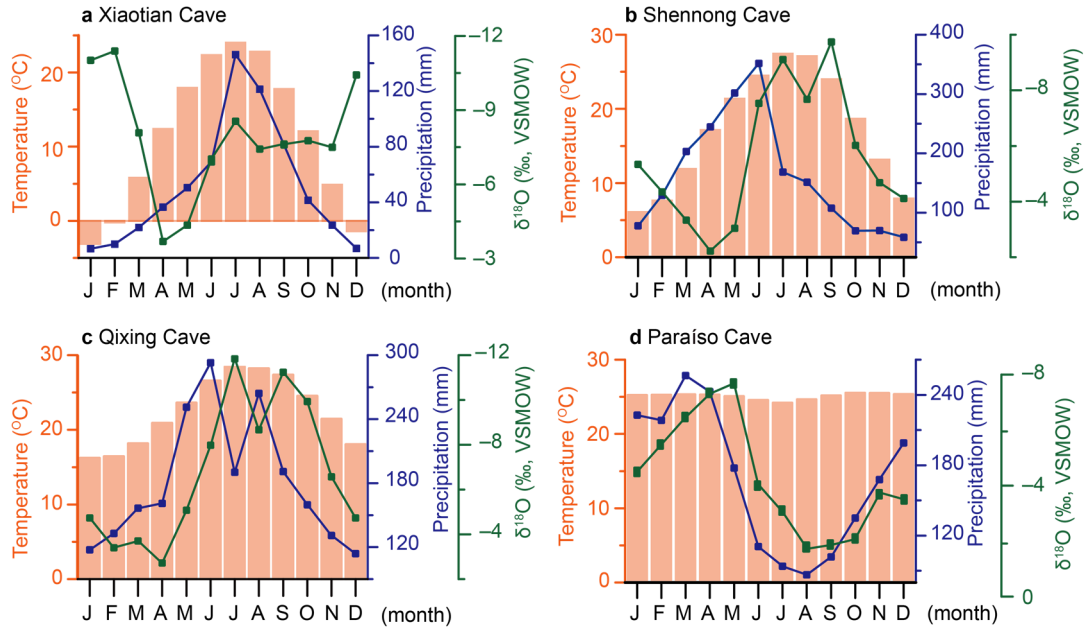
Southeast China



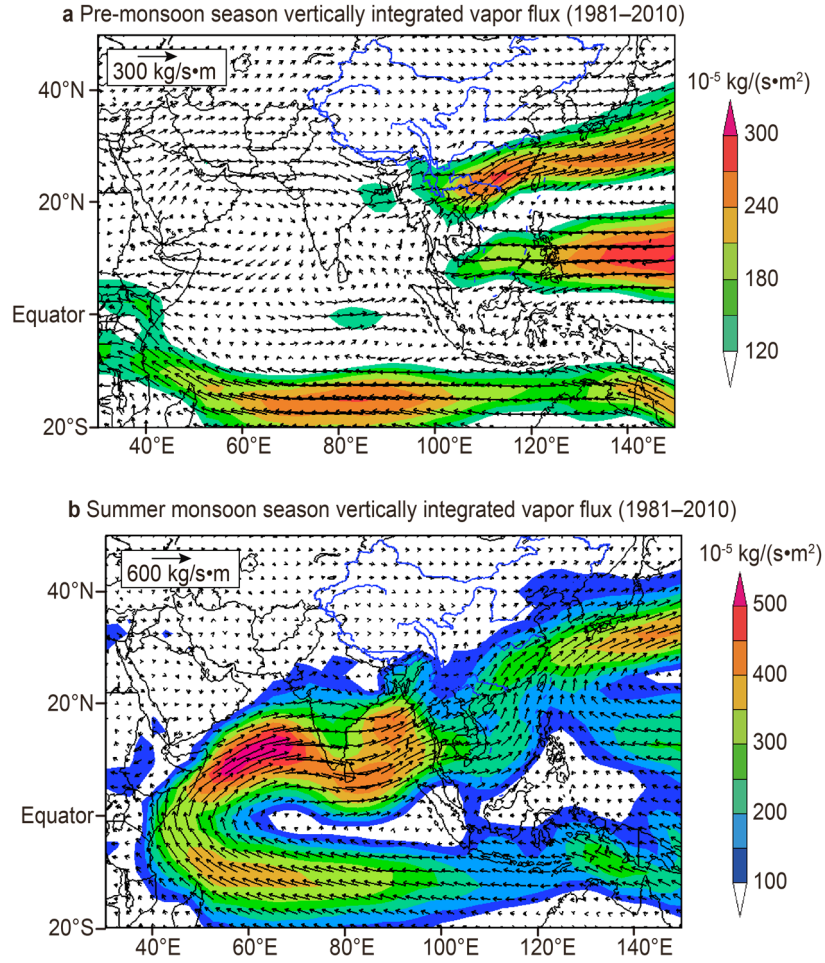
Supplementary Fig. 9. The event amplitude calculated by “Mean-fitting” algorithm. Speleothem $\delta^{18}\text{O}$ records from (a) Shennong cave (this study) and (b)–(h) Xianyun cave^{54–58} from Southeast China, respectively. (i)–(l) Indian summer monsoon (ISM) domain speleothem records from Bittou cave (this study and ref.⁴³). The “Mean fitting” algorithm is described in Methods. The vertical red lines show the pre-transition and post-transition levels given by the algorithm; the double-sided arrows illustrate the amplitude. DO: Dansgaard-Oeschger. SUPER DO: Super-long DO; INTER DO: Intermediate-long DO (refer to Fig. 3).



Supplementary Fig. 10. Simulated speleothem and sea-surface water $\delta^{18}\text{O}$ and water-vapor transport during long and short DO warming events. (a)–(c) Composites of changes in annual mean cave speleothem $\delta^{18}\text{O}$ ($\delta^{18}\text{O}_c$) over continents and annual mean $\delta^{18}\text{O}$ of sea-surface water ($\delta^{18}\text{O}_{sw}$) over oceans (shaded, units: ‰) (Methods) and water-vapor transport (WT) (vectors, units: kg/m/s) during simulated short (Overshoot minus Stadial phases) and long (Interstadial minus Stadial phases) DO warming events, and the difference between long and short interstadials (Interstadial minus Overshoot phases). Note that the color scales in (a) and (b) are different from (c). The time periods of Overshoot and Interstadial phases refer to Fig. 4a.



Supplementary Fig. 11. Climatology of the cave sites. Yellow columns and blue lines in (a)–(d) illustrate the average temperature and precipitation over different cave sites, respectively. Green line and dots in (a)–(d) indicate precipitation $\delta^{18}\text{O}$ ($\delta^{18}\text{O}_p$) from Shijiazhuang meteorological station during 1985–2003 CE, $\delta^{18}\text{O}_p$ from Changsha meteorological station during 1988–1992 CE, $\delta^{18}\text{O}_p$ from Guiyang meteorological station during 1988–1992 CE and $\delta^{18}\text{O}_p$ from Manaus during 1998–2010 CE¹⁷, respectively. Temperature and precipitation data for 1951–2020 CE are from Climate Research Unit Time Series version 4.06 (CRU-TS4.06) gridded observation dataset, which has a horizontal resolution of $0.5^\circ \times 0.5^\circ$ ⁵⁹. $\delta^{18}\text{O}_p$ data are from the International Atomic Energy Agency (IAEA) Global Network of Isotopes in Precipitation (GNIP) database (<http://www.iaea.org>).



Supplementary Fig. 12. Vertically integrated vapor flux in the Asian summer monsoon domain. (a) Pre-monsoon season (March 15 to May 15), and **(b)** monsoon season (June 1 to July 15) during 1981–2010 CE. The background figure was generated using climate data obtained from the Nation Climate Centre of China (http://cmdp.ncc-cma.net/Monitoring/cn_stratosphere_interaction.php?cat=water) and was further modified in this study.

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