The rate and consequences of future high latitude ice sheet retreat remain a major concern given ongoing anthropogenic warming. Here, new precisely dated stalagmite data from NW Iberia provide the first direct, high-resolution records of periods of rapid melting of Northern Hemisphere ice sheets during the penultimate deglaciation. These records reveal the penultimate deglaciation initiated with rapid century-scale meltwater pulses which subsequently trigger abrupt coolings of air temperature in NW Iberia consistent with freshwater-induced AMOC slowdowns. The first of these AMOC slowdowns, 600-year duration, was shorter than Heinrich 1 of the last deglaciation. Although similar insolation forcing initiated the last two deglaciations, the more rapid and sustained rate of freshening in the eastern North Atlantic penultimate deglaciation likely reflects a larger volume of ice stored in the marine-based Eurasian Ice sheet during the penultimate glacial in contrast to the land-based ice sheet on North America as during the last glacial.
During glacial terminations, retreating ice sheets release large meltwater fluxes into the ocean. This leads to rising sea level and can initiate strong climate feedbacks when sufficiently large meltwater fluxes reach regions of deepwater formation in the North Atlantic. These ocean-atmosphere interactions, together with associated ocean CO₂ release, operate as strong and rapid amplifiers of the original orbital induced insolation change. The rapid retreat of ice sheets which yields high meltwater fluxes, can be caused by marine ice sheet instability, ice sheet saddle collapse and marine ice cliff failure\(^{12}\). While the timing and associated deglacial feedbacks have been extensively studied for the last deglaciation (Termination I, TI)\(^1\), knowledge of the insolation thresholds, rate of ice sheet melting, and feedback sequence of previous terminations is more limited due to the lack of direct absolute chronology for ice retreat and deglaciation warming. Some studies suggest that the sequence of millennial feedbacks may be different in previous terminations\(^3\). The penultimate deglaciation, Termination II (TII), is of particular interest because orbital boundary conditions were different, and the termination was followed by an interglacial with a \(+1.2\) to \(5.3\) m sea level highstand - suggesting both Greenland and Antarctic ice sheets retreated further than during TII\(^6\), despite similar highs in atmospheric CO₂\(^7\). To fully understand the last interglacial highstand, clear knowledge of the ice sheets melting time and retreat rates during TII is needed\(^8,9\).

Water stored in high latitude ice sheets has a \(\delta^{18}O\) lower than the mean ocean, meaning the deglacial melting of the Northern Hemisphere (NH) ice sheets lowers the \(\delta^{18}O\) of the surface ocean (\(\delta^{18}O_{\text{sw}}\)). For regions proximal to release of glacial meltwater, such as the North Atlantic, the rate of \(\delta^{18}O_{\text{sw}}\) depletion and freshening can exceed the global average\(^10,11\), which has been useful in diagnosing meltwater routes during TII\(^12\). We propose that the evolution of the North Atlantic \(\delta^{18}O_{\text{sw}}\) can be recorded in coastal European speleothems through the transfer of the \(\delta^{18}O_{\text{sw}}\) signal of the ocean moisture source to the \(\delta^{18}O\) of rainfall, and thereby the \(\delta^{18}O\) of dripwater\(^15\), from which the speleothem is formed. We report on the North Iberian Speleothem Archive (NISA) from caves within 10 km of the Atlantic coast (Fig. 1) at coastal elevation (<70 m above sea level), directly adjacent to the main moisture source region in the eastern North Atlantic. We employ speleothems spanning the last 25 ky to evaluate the \(\delta^{18}O\) and \(\delta^{13}C\) proxy relationships against independent records of regional temperature and \(\delta^{18}O_{\text{sw}}\). The analysis of speleothems covering TI and TII from the same caves and same proxy indicators provides, by the first time, the opportunity to unambiguously compare the timing and rate of the last two deglaciations on an absolute 230Th chronology, with rates further refined by annual layer counting in TII. We show that the NISA record provides a unique opportunity to directly link the impact of rapid freshening of the North Atlantic on regional atmospheric temperatures, using the same archive, in order to test mechanisms for deglacial feedbacks.

**Results and discussion**

**Controls on the \(\delta^{18}O\) and \(\delta^{13}C\) of NISA speleothems.** Using six stalagmites spanning the last 25 ky to generate a composite splice speleothem \(\delta^{18}O\) record (Methods; Supplementary Figs. 1 and 2), we test the relationship between the speleothem \(\delta^{18}O\) and independent \(\delta^{18}O_{\text{sw}}\) estimates from foraminifera in North Atlantic marine sediment cores (Supplementary Fig. 3).

We find that throughout the last 25 ka, where both marine and speleothem records rely on independent absolute chronologies, \(\delta^{18}O_{\text{NISA}}\) exhibits an unusually close correlation with the \(\delta^{18}O_{\text{sw}}\) in the eastern North Atlantic Ocean. The \(\delta^{18}O_{\text{NISA}}\) is most strongly correlated (\(r^2\) of 0.91) with the \(\delta^{18}O_{\text{sw}}\) on the Irish Margin\(^16\) and a low standard error indicates a consistent relationship throughout the deglaciation despite the potential for variation in atmospheric and surface ocean boundary conditions (Fig. 2, Supplementary Table 1). Strong correlation is also attained for the \(\delta^{18}O_{\text{sw}}\) of the S. Iberian Margin\(^17\) and W. Iberian Margin\(^18\) (Fig. 2, Supplementary Table 1), as expected given the southward flowing eastern boundary current from the Irish margin along the Iberian margin and via the Atlantic Jet surface water flow through the Straits of Gibraltar\(^19\). The latter regions \(\delta^{18}O_{\text{sw}}\) are also closely correlated with the Irish margin \(\delta^{18}O_{\text{sw}}\) (Supplementary Table 2) and show no systematic changes in the \(\delta^{18}O_{\text{sw}}\) gradients (Supplementary Fig. 3). A HådCM3 model of the early phases of the TI deglaciation, simulating meltwater based on ICE6G ice sheet evolution and drainage routing (meltwater ~45% from EIS and 55% from the NAIS), yields similar spatial pattern of meltwater-induced salinity anomaly. Low salinity is concentrated in the North Atlantic north of 40°N, including the Irish Margin\(^16\) and extending further southward due to eastern boundary current\(^20\) along the Iberian Peninsula, NW coast of Africa, and into the Alboran Sea (Fig. 1). The strong relationship between \(\delta^{18}O_{\text{NISA}}\) and the proximal surface ocean \(\delta^{18}O_{\text{sw}}\) is not observed in Mediterranean region speleothems over TI (Supplementary Table 3), likely due to additional hydrological effects\(^21-23\).

Several factors may contribute to the coherence between \(\delta^{18}O_{\text{NISA}}\) and the \(\delta^{18}O_{\text{sw}}\) of the proximal ocean. Our analysis of all precipitation events in Northern Spain at a station ~100 km west of the cave site in 2015–2016, shows that the majority of rainfall events in this region derive from oceanic moisture uptake in the easternmost North Atlantic ocean proximal to the Iberian Peninsula, especially north of Iberia (Fig. 2d). The high east-west topographic barrier of the Pyrenees and Cantabrian mountains of Northern Spain likely contributes both to the modern dominance of this northern proximal ocean moisture source and the stability of this moisture source over time. The location of caves at coastal elevation, directly adjacent to the main Atlantic moisture source region, minimizes the isotopic distillation between the moisture source area and the cave site. The imprint of such distillation on speleothem \(\delta^{18}O\) from European caves increases with their increasing inland distance and elevation\(^24\), and temperature modulates this distillation so that temperature rather than the \(\delta^{18}O_{\text{sw}}\) of moisture source, becomes the dominant influence on \(\delta^{18}O_{\text{NISA}}\) in central Europe\(^25,26\). In our NW Iberian cave locations, rainfall monitoring shows that the slight decrease in rainfall \(\delta^{18}O\) with decreasing temperature\(^27\) appears of similar magnitude but opposite sign as the temperature-dependent fractionation between dripwater and calcite in the cave, leaving the \(\delta^{18}O_{\text{sw}}\) of the Atlantic moisture source as the principal variable expressed in the stalagmites. We propose that the \(\delta^{18}O_{\text{sw}}\)\(\delta^{18}O_{\text{NISA}}\) relationship identified in TI remained stable over TII.

Additionally, we confirm that over TII, the \(\delta^{13}C_{\text{NISA}}\) in stalagmites from our caves varies inversely with regional temperature records from marine archives (Methods, Fig. 3; Supplementary Fig. 4), matching millennial variations in SST despite potential deviations in marine \(14^C\) chronology due to variation in surface ocean reservoir ages\(^28\) and the potential for alkeneons of differing production ages to be deposited together\(^29\). Similar temporal correlation between temperature and \(\delta^{13}C\) during marine isotope stages 3 and 4 has been observed in speleothems from the Atlantic coastal region of France\(^30\). We conclude that the main trends are attributable to the carbon isotope signature acquired through rain equilibration with soil gas and bedrock dissolution, rather than in-cave processes such as prior calcite precipitation (Methods). Multiproxy process modeling suggests that this correlation arises because higher temperature strongly increases vegetation productivity in this biogeographic regime, enhancing soil CO₂ production and
oversupply of CO₂ to karst waters, both factors which produce more negative δ¹³C in speleothems (δ¹³Cspeleo). We infer that the specific slope of the temperature vs δ¹³CNISA relationships may be specific to a given cave systems and host lithology, and may be sensitive to the degree of smoothing of millennial-scale changes which is affected by the speleothem growth rate and the years aggregated in each drilling increment.

NISA speleothem records spanning TII. In the same coastal cave systems, we document the evolution of δ¹⁸ONISA and δ¹³CNISA over TII as indicators of δ¹⁸Osw and temperature, respectively. Three ²³⁰Th–dated NISA stalagmites replicate the main features of the TII deglaciation between 135 and 129 ka (Fig. 4a). Annual countable fluorescent banding in stalagmite Garth (Supplementary Figs. 5 and 6) additionally provides a precise estimation of the rate of the main δ¹⁸ONISA transitions. The record from stalagmite Garth extends to 112 ka BP and provides further context for the deglaciation, albeit with slowed growth and lower resolution between 127.5 and 122.5 ka. We assess the potential effect of in-cave processes such as PCP on the isotope records to focus on the most robust proxy trends (Methods, Supplementary Figs. 7–10). Given the correlation between δ¹⁸ONISA and marine microfossil records of δ¹⁸Osw from west and south Iberia in TI (Fig. 1), we tune marine sediment age models to speleothem chronology by synchronizing the major freshening associated with deglacial ice melting and some key temperature events (Methods, Supplementary Figs. 11–15).

δ¹³CNISA records the stepped deglacial warming from 134 to 128 ka, punctuated by millennial cooling events, while δ¹⁸ONISA records the deglacial freshening of the eastern North Atlantic beginning 135.7 ka (Fig. 4a). The subsequent descent into the last glacial cycle is also reflected in the increase in δ¹⁸ONISA after 122 ka and persistent cooling in δ¹³CNISA after 118 ka which suggest temperatures similar to the PGM by 112 ka. The annual layer counted chronology in many sections of stalagmite Garth resolves significant centennial to millennial scale variations in both the rate of eastern North Atlantic freshening (δ¹⁸ONISA) and its relationship with regional warming (δ¹³CNISA) (Fig. 4b).

Two large freshening pulses characterize the TII onset (MWPTII-A and MWPTII-B, Fig. 4b). Following previous studies, we infer that the addition of meltwater is the main driver of rapid deglacial freshening and declines in the δ¹⁸Osw in the eastern North Atlantic. Both abrupt freshening pulses began during a period of rapid warming, then, after several decades of rapid freshening, temperatures cooled rapidly, consistent with freshening-induced AMOC slowdown (Fig. 4b). Yet despite cooling, the freshening rate remained high for nearly a century during both events. Following the first period of rapid meltwater release (MWPTII-A), a local freshwater δ¹⁸O anomaly was
maintained in the eastern North Atlantic for a duration of ~600 yrs before the rate of warming increased and over several centuries the surface ocean δ18O anomaly was reduced (Fig. 4b). We infer that slowed meltwater flux and re-invigorated AMOC dissipated the δ18O anomaly through dilution and mixing of meltwater throughout the global ocean.

Following the second and most intense negative δ18O shift (MWPTII-B) and AMOC slowdown, a North Atlantic freshwater δ18O anomaly persisted. Unusually, for ~3000 years, the δ18Oapia remained lower than the final interglacial state. The maintenance of a freshwater δ18O anomaly for so long in the surface eastern North Atlantic requires addition of meltwater at rates greater than it can be distributed through mixing with the global ocean. Thereafter, starting around 131 ka, the North Atlantic freshwater δ18O anomaly diminishes rapidly, which suggests that the rate of addition of meltwater into the eastern North Atlantic had slowed sufficiently so that ocean circulation homogenized the deglacial meltwater anomaly to the global ocean average. This decline coincides with a rapid decrease in IRD delivery in the Northeast Atlantic and Labrador Sea (Fig. 4b, Supplementary Fig. 12). The rate of decrease in δ18O of benthic foraminifera (δ18Oapia) on the Iberian margin also slowed significantly at 131 ka (Figs. 4 and 5), consistent with slowed melt rate.

We detect late phases of melting between 130 and 129 ka. A rapid negative shift in δ18Oapia around 129.7 ka, during regional warming, likely reflects acceleration of meltwater addition. Notably, the continued meltwater flux through 129.3 ka appears to coincide with the deposition of a distinctive sediment layer on the Labrador margin33 which occurred at ~129.3 ka on our chronology (Supplementary Fig. 13), a depositional event attributed to a final North American Ice Sheet (NAIS) flood outburst, analogous to the 8.2 ka event following TI. The 8.2 ka event initiated when ~90% of North American ice had melted34, so if the sediment layer at 129.3 ka is analogous33, it suggests that a similarly high fraction of melting of PGM NAIS ice was likewise complete by this time.

On the W Iberian margin, our speleothem based chronology for TII indicates a much faster rate of depletion in δ18Oapia during TII than during TI, suggesting that deep waters in the North Atlantic underwent freshening much more rapidly in TII than during TI (Fig. 5). The chronology of the global marine δ18Oapia stack35, used as index of glacial ice volume to estimate rates of deglaciation and as inputs for model calculations36 is based on tuning to NH summer insolation, assuming a comparable phase relationship to TI. δ18Oapia curves have also previously been used to estimate sea level change during TII compared to TI. However, our chronology suggests that the rapid TII deglaciation precedes the peak in high latitude NH summer insolation and the existing chronology35 probably yields minimum estimates of rates of deglaciation. The speleothem chronology for the δ18Oapia on the W Iberian margin provides an alternate estimate of the rate of sea level change and deglaciation (Fig. 5). This likely represents the upper estimate of the rate of sea level rise, duration, and amplitude of freshwater forcing compared to those employed in recent models using original chronology7. Our independent tuned chronology for δ18Oapia-based sea level estimates are closest to the oldest chronology (95% upper CI) for the Red Sea curve38,39, and our chronology implies the potential for an earlier and even more concentrated meltwater pulse than estimated from the derivative of the Red Sea sea level curve (Fig. 5)40.

Similarities and contrasts between the meltwater anomaly during TI and TII deglaciation. Our speleothem absolute chronology confirms that both TI and TII deglaciations initiated at a similar threshold of caloric summer insolation (5.8 GJ m⁻²)
Yet, the speleothem chronology for $\delta^{18}O_{\text{benthic}}$ reveals attainment of interglacial values within ~5 ky of the TII deglacial onset, compared to nearly 9 ky in the TI deglaciation. Notably, the development of the eastern North Atlantic $\delta^{18}O_{\text{sw}}$ anomaly during TII is markedly different from that of TI (Fig. 6). During TII, $\delta^{18}O_{\text{sw}}$ remained more negative than the final interglacial state for nearly 3000 years, a situation which never occurred in TI. During TI, the regional surface ocean $\delta^{18}O_{\text{sw}}$ anomaly, calculated relative to the benthic $\delta^{18}O_{\text{calcite}}$ on the Iberian margin ($\delta^{18}O_{\text{NISA-BC}}$; Methods), remains below $-1\%$ only between 16.4 and 15.2 ka (1.2 ky duration), whereas during TII it remains below $-1\%$ from 134.7 to 131.2 (3.5 ky duration; Fig. 6).

During TI, this period of greatest regional $\delta^{18}O_{\text{sw}}$ anomaly coincides with the interval of fast melt rate of the Eurasian ice sheet according to recent sea level reconstructions (Fig. 7; Supplementary Fig. 16). One interpretation is that EIS meltwater leads to a greater negative $\delta^{18}O_{\text{sw}}$ anomaly in the eastern north Atlantic compared to NAIS meltwater. Southward advection of surface waters from the Nordic seas along the European continental margin is simulated during periods of retreat of the northern and western EIS (Fig. 6).

Intermediate complexity models comparing identical freshwater forcing in different outlets infer a greater salinity anomaly in the eastern North Atlantic due to Eurasian meltwater compared to North American meltwater routes through Hudson Bay, Gulf of Mexico, or Gulf of St. Lawrence. This is because for these NAIS routes, the salinity (and $\delta^{18}O_{\text{sw}}$ anomaly) would be diluted during advection across the Atlantic by the North Atlantic Drift. These model simulations, albeit limited, suggest the potential for higher amplitude $\delta^{18}O_{\text{sw}}$ anomaly in the eastern North Atlantic during EIS melt than NAIS melt, given similar $\delta^{18}O_{\text{ice}}$ as assumed in previous studies.

Therefore, one explanation for the longer duration surface ocean $\delta^{18}O_{\text{sw}}$ anomaly during TII is a larger proportion of EIS-derived meltwater in TII compared to TI.

In support of this explanation, Eurasian PGM deposits are found well beyond the limit of ice extent from the LGM, with the largest expansion in the Barents-Kara sectors of northern Russia (Fig. 1). However, the sparse direct chronology of ice positions does not distinguish if these advances were synchronous across all sectors to yield at a single time during the penultimate glacial
significantly larger EIS, or if, as in the last glacial cycle, sectors were out of phase\textsuperscript{44,45}. The potential for a larger PGM EIS ice sheet has been explored in dynamical ice sheet models with total EIS ice volumes of up to 70 m sea-level equivalent, nearly triple that of the LGM EIS\textsuperscript{46}. Because of similar estimated PGM and LGM sea level and benthic $\delta^{18}$O, this simulated large PGM EIS likely coexisted with a smaller than LGM NAIS\textsuperscript{16}. The ice sheet reconstruction featuring a large TII EIS\textsuperscript{46} encompasses a 2.5–3.5x greater volume of marine-grounded (grounded below paleo sea level) EIS ice at the PGM, compared to the LGM (Methods) with
thickest ice in the Fennoscandian sector. Because of isostatic loading and higher relative sea level caused by self-gravitation, the larger PGM EIS would have been even more sensitive to ocean forcing and marine ice sheet instability which could have contributed to a rapid retreat. Although the onset of modeled TII EIS retreat differs slightly from our record, its duration is broadly consistent with the here documented duration of eastern North Atlantic δ18O anomaly.

An alternative explanation for the greater freshening in the eastern North Atlantic during TII is that LGM-like ice sheet configuration melted faster in TII than in TI. However, we consider this scenario less likely because it would require faster melting of the portions of NAIS which were longest lived during TII, and we have no evidence of a stronger forcing during TII. An Arctic ice shelf, neutrally buoyant but potentially containing a freshwater equivalent to 10 m s.l.e. is proposed to have existed at some time during the penultimate glacial cycle but the timing of formation and collapse remains controversial. Based on our records, we cannot confirm nor rule out disintegration of such an ice shelf as a contributor to freshening at the onset of TII, leading the retreat of Northern sectors of the EIS. However, the modest magnitude of Arctic shelf ice volume suggests that this cannot be the only factor responsible for differences in the meltwater anomaly and rate of change in δ18O_benthic and sea level rise over TII (Figs. 5 and 6).

In addition, models suggest that the intensity and duration of the regional δ18O associated meltwater anomaly are also increased by a slowing of the rate of dilution of the meltwater signal, primarily because slowed overturning circulation reduces the vertical dissipation rate. During TI, the greatest regional freshening anomaly (16.4 to 15.2 ka) also coincides with a period of AMOC reduction inferred from proxies such as Pa/Th (Fig. 7, Supplementary Fig. 16). For this time interval, the relative abundance of N. pachyderma sinistral and C37:4 alkenone are likely to be sensitive to the strong winter season climatic impacts of AMOC slowdown such as the relative abundance of Neogloboquadrina pachyderma sinistral or relative abundance of C37:4 alkenone. These climatic indicators suggest a less extreme AMOC slowdown and sea ice response to stratification during HE11 in TII than in HE1 during TI (Fig. 6, Supplementary Fig. 9). The duration of HE11 and HE1 are comparable. Thus, available evidence would not support a difference in physical circulation as the only cause for the contrasting surface ocean eastern North Atlantic δ18O fresh anomaly. Although there are no direct quantitative proxies for AMOC intensity in TI or TII, some mid-latitude proxy records are likely to be sensitive to the strong winter season climatic impacts of AMOC slowdown such as the relative abundance of Neogloboquadrina pachyderma sinistral or relative abundance of C37:4 alkenone. These climatic indicators suggest a less extreme AMOC slowdown and sea ice response to stratification during HE11 in TII than in HE1 during TI (Fig. 6, Supplementary Fig. 9). The duration of HE11 and HE1 are comparable. Thus, available evidence would not support a difference in physical circulation as the only cause for the contrasting surface ocean eastern North Atlantic δ18O fresh anomaly in TII vs TII.

Based on these considerations, we suggest that our record of freshening is most consistent with a PGM EIS much larger than its LGM counterpart, and is compatible with the larger EIS volume scenarios of 60 m to 71 m sea level equivalent based on glacial-isostatic modelling. Further ice sheet modeling and assimilation of near field geophysical data are required to evaluate this interpretation.
Causes and feedbacks from meltwater-induced AMOC disruption. Our new, highly resolved record of TII reveals that, as in TI, the deglacial sequence was characterized by a series of millennial-scale variations not previously resolved in records of sea level rise$^{35,37-40}$. We find evidence for multiple phases of AMOC reduction and reinvigoration in TII which provide new clues to the mechanisms of AMOC instability during glacial terminations. It has been debated, whether meltwater release causes AMOC reduction, or if it is the subsurface ocean warming during AMOC shutdown that causes accelerated ice sheet collapse and meltwater release$^{53}$. Here, our new annually laminated records show that enhanced freshwater addition during brief
periods of rapid warming led, and therefore likely caused, AMOC reductions early in TII.

AMOC recovery has most often been attributed to a slowing of the rate of meltwater addition. Alternatively, recent model experiments suggest that AMOC can also recover, despite sustained meltwater flux, if there is a change in the boundary conditions which set the threshold for weakened AMOC. For example, AMOC may have recovered after HE1 in TII during the Bolling-Allerød because the rapid CO2 rise during HE1 raised the forcing threshold required to maintain weak AMOC5,62. TII was previously proposed to lack such a mid-termination AMOC recovery. Yet, we identify a recovery after the first AMOC reduction of TII (Fig. 4b). This event occurred earlier in the deglaciation (e.g. δ18O benthic on the Iberian margin was 4.45 % during the TII recovery compared to 4.25% at the start of the Bolling-Allerød) (Fig. 6). One possibility is that AMOC recoveries at TII may have been driven not by evolving thresholds in AMOC sensitivity but rather by strong temporal variations in meltwater forcing. This short duration of the first TII AMOC reduction (600 years vs ~3000 years for HE1 in TII) may reflect the operation of a negative feedback of AMOC slowdown on TII melting rate, if the main location of early TII melting occurred in an area sensitive to AMOC-induced cooling.

On the other hand, if a change in boundary conditions is required for AMOC recovery, then our results in TII present a paradox, because according to the current available ice core chronology, the onset of the first AMOC recovery occurred when CO2 had risen only to 207 ppmv67, much less than the 245 ppmv attained by the onset of the Bolling-Allerød (Fig. 8). Even by the end of HE11, according to AICC2012 chronology, pCO2 had only reached 245 ppmv, much less than the 260 ppm reached by the end of the Younger Dryas. This ice core CO2 chronology would suggest a very different sensitivity of AMOC to CO2 between TII and TII. This circumstance could reflect other initial boundary conditions, such as a lower glacial height of the North American ice sheet or different initial initial PGM thermohaline regime63, which conditioned the AMOC threshold.

It is also possible that ice core records may underestimate the rate of TII CO2 rise due to the differing precision of age models based on annual layer counting for TII64, and age models based on age interpolation between two age control points at 121 and 135 ka tuning ice core O2/N2 ratios to local summer insolation65. The chronology of the main TII CO2 rise can be further explored from an available marine CO2 proxy record based on planktic foraminifera δ13C from the western Caribbean66. This record indicates a large initial CO2 rise synchronous with a surface ocean freshening recorded by the same planktic foraminifera67. Although tuning this site to our speleothem chronology has some uncertainty (Methods, Supplementary Fig. 17), and proxy CO2 estimates are less precise than direct measurements in ice cores, a conservative correlation suggests that a larger portion of the deglacial CO2 rise may have occurred rapidly, coincident with the onset of rapid EIS melting and the second AMOC slowdown (Fig. 7). An early and rapid CO2 rise, introducing greenhouse-forced warming, is also consistent with the progressive warming above glacial background levels occurring after MWP-TIIB in our air temperature record, which would have maintained the melting of NH ice sheets during the HE11 AMOC slowdown (Fig. 8). Thus, diagnosing the links between rate of ice melting, AMOC recovery, and deep ocean carbon storage during TII may require reassessment of the rate of CO2 rise.

Overall, the first direct absolute age constraints on the timing and rate of TII freshwater release to the North Atlantic are consistent with a large EIS, which was prone to rapid retreat due to its enlarged marine-based boundaries. These observations highlight the control that glacial boundary conditions exert, by shaping ice sheet anatomy, over the rates of ice melting and also the nature of the subsequent interglacial68. The new records also provide the first evidence for discrete centennial scale meltwater pulses indicative of phases of accelerated ice sheet failure during TII and a detailed view of their coupling with cooling consistent with AMOC reductions. The role of internal ice processes and climate feedbacks associated with the retreat of large marine-based northern hemisphere ice sheets should be further investigated with coupled ocean-climate-ice sheet models.

Methods
Identification of final moisture source exchange via backtracking. Moisture uptake regions of 104 rain events which were sampled <100 km west of the cave during 2015 and 2016, were calculated using backward-trajectory analysis with 1 h temporal resolution, performed using the Hybrid Single-Particle Lagrangian Integrated Trajectory (HYSPLIT) Model (Version 4.8)69 and following a similar methodology as previous studies70. Moisture uptake regions along 850 hPa, 700 hPa and 500 hPa were identified using a specific humidity threshold of 0.5 g/kg increment in at least 6 h, using meteorological data provided by backward-trajectories. Results are shown in Fig. 2d.

Stalagmite samples. Stalagmite samples were analyzed from the following previously described71 NW Iberian caves: Pendal Cave (stalagmites Candela, Laura), La Vallina Cave (Garth, Gael, Galia, Gloria), and Cueva Rosa (Neith, Alicia). These stalagmites cover TII, Greenland Stadial 22 (GS22), and TII27,71,72. Additional growth phases have been studied in the stalagmite Candela since earlier publication72, and we report additional dates and higher resolution geochemical data. In Candela, A ~10 cm long basal growth phase (CANB) has been sampled along its main growth axis and redated, contrasting with strongly off-axis and condensed sampling of this phase in earlier work. Additionally, within the Bolling-Allerød, a transient 1.5 cm shift in growth axis has been resampled along its expanded central growth axis for 2.5 cm, expanding by 5-fold the record previously sampled in a lateral portion at only 0.5 cm thickness.

Measurement of seasonal fluorescent banding for growth rate evaluation in key intervals. To account for variation in growth rates25 between 230Th dates, for intervals of rapid freshening in TII and TII, we refine age models by estimating stalagmite growth rates from the width of growth bands defined by fluorescent
layers\textsuperscript{74} which have been described to result from seasonal delivery and/or incorporation of fluorescent organic substances in dripwater\textsuperscript{75}. Stalagmite slabs were mounted in low-viscosity Laromin C260 epoxy (cycloaliphatic diamine polymer), ground flat with SiC paper, and then polished with a diamond suspension to 0.25 µm in preparation for Confocal laser scanning microscopy (CLSM) and laser ablation ICP-MS. Imaging by CLSM was performed at the Scientific Center for Optical and Electron Microscopy (ScopeM) at ETH Hönggerberg using an Olympus Fluoview 3000. Here, a series of overlapping images were obtained under 100–200x magnification using an incident wavelength of 488 nm and measuring fluorescence in a 490–590 nm window. Images were obtained at 1024 × 1024 or 2048 × 2048 resolution, corresponding to 1.2 to 0.6 µm/pixel (respectively) under 100x magnification. Images were processed using Fiji/ImageJ, where the exact distances between growth layers were recorded on an absolute length scale and later cross-referenced to laser ablation and drilling (isotope, trace element) data. Confocal images were counted in stalagmite Garth section from 177.5 to 269.5 mm from the tip. Growth layer thickness was used to
is used to constrain growth rates (green line in growth rate curve in Supplementary Fig. 1a) and to identify the point at which growth becomes most strongly condensed. For example, the age model is pinned at a new MC age of 18.0 ka ± 78 year 230Th date (at reference level 4.5 mm), and advanced upwards ~90 years using widths of fluorescent growth layers, constrained by 230Th growth rate. Growth layers are not countable in the following 7.7 mm of stalagmite (brown line in Supplementary Fig. 1a). Consequently, in this section two tie points with the geochemical variations in stalagmite Laura (red points in Supplementary Fig. 1a) were used to constrain the Candela age model, and growth rate was varied from 10 to 1 μm/year inversely with Candela δ13C, with a 630 year negligible deposition (<0.2 μm/year) interval at the end of the δ13C maximum. The precision of the age model is lowest between 17.85 and 16.1 ka, and improves by 16 ka due to a 15.4 ka 230Th date and the 16.1 and 14.5 ka tie points with Laura. Following these tie points, the age model until the hiatus at 11.6 ka is established from 230Th tie points, with interpolation between dated points further constrained by counting of 600 annual layers (14.5 to 13.9 ka) during the Bølling-Allerød. This set of δ13C stalometh records chronicles the changes across TI in higher detail than previous records from the region, with the following average absolute 95% CI age uncertainties 20–18 ka (Candela, +/- 300 years); 18–16 ka (Candela, +/- 500 years), 16–14 ka (Laura, +/- 150 years); 14–12.5 ka (Candela +/- 260 years).

Age models for stalagmites spanning TI and the Bchron uncertainty windows, are illustrated in Supplementary Figs. 5 and 6. For TI, Garth provides anchor chronology due to its high U content and growth rate and presence of annual fluorescent growth banding; Gael also provides an independent chronology albeit of lesser resolution. For Neth, whose very low U content complicates a precise age model, we develop an age model by tuning main features of isotope record related to that of Garth within the uncertainties of the Bchron model. Neil is therefore presented to indicate the reproducibility of the main features of the isotope records shown in Garth, not as independent verification of precise event chronology. The age model in Garth was developed with the aid of laminae counting (Supplementary Fig. 5).

Stable isotope analysis and TI stable isotope splice. New stable isotope determinations were measured at Scientific and Technologic Centers from the University of Barcelona with a Thermo-Finnegan MAT-252 coupled to a CarboKiel-II and at ETH Zürich with a Thermo-Finnegan Delta V Plus coupled to Gas Bench II43 and are reported in % relative to the Vienna PDB Standard. Analytical uncertainties for calibration with two in-house carbonate standards and NBS-19 and NBS-18 international standard and were 0.08 %0 for both isotopes. Samples for isotopes were micromilled at 0.1 to 0.5 mm increments, or drilled at 1 to 5 mm increments, from the central axis of stalagmites. Stalagmite δ13C0 and δ18O were screened for the coupled effects of CO2 degassing and prior calcite precipitation, as detailed below. A spliced record of δ13C and δ18O is compiled from stalagmites spanning 25–5 ka BP. Where multiple stalagmites grow synchronously, we employ for the splice, the Bchron with best constrained chronology and the stalagmite in which there is the least variation in Mg/Caindex over the included time interval. The goal of this approach is to limit the potential influence of PCP on the inferred trends in δ13C and δ18O.

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**Fig. 8 Evolution of atmospheric pCO2 during TI and TII.**

**a.** The evolution of ice core pCO2 for TI64 and TII7 versus benthic δ18O on the Iberian margin; benthic δ18O tuned to speleothem chronology for TII. Symbols are colored according to the C37:4 alkenone abundance in the S. Iberian margin58 (dark blue C37:4 > 6%; light blue C37:4 <2%) as an indicator of timing and intensity of HE climate impact. Horizontal gray lines emphasize CO2 levels reached following abrupt rises during stadials in TI (HE1 and Y.D.), and horizontal dark green lines highlight CO2 levels attained at the end of stadials in TII (HE11 and HE11').

**b.** Comparison of the TII temporal evolution of CO2 on ice core chronology (gray curve)7 with pCO2 estimated from marine proxies (pink, vertical bars show 95% CI on CO2 estimate)66 on speleothem chronology as discussed in Methods (horizontal bars show +/− 750 year uncertainty on age). The TII marine proxy pCO2 estimates from between 138 and 128 ka are shown with small black squares in a), with error bars omitted for clarity.

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be an additional component of variation in δ13CNISA due to in-cave modification of dripwater δ13C due to CO2 degassing and PCP. Monitoring suggests that secondary factors such as changes in Mg partitioning, non-bedrock Mg sources, and variations in the congruency of bedrock dissolution due to variable water-rock contact times exert a limited influence on speleothem Mg/Ca in most of our studied stalagmites. We refrain from use of the Mg/Ca index, and interpretation of isotope results, for the <9 ka portions of Pindal Cave stalagmites in which the cave position on the sea cliff generates a strong marine aerosol Mg/Ca influence. Based on our finding that in stalagmites spanning TI, the main glacial/interglacial δ13CNISA contrasts and main transitions do not coincide with unidirectional changes in Mg/Ca_mgcalc, we conclude these are not artefacts of varying PCP.

In TII, the Mg/Ca_mgcalc is used to ascertain trends which are not influenced by PCP. In stalagmite Garth, the principal transition to strongly negative δ13CNISA 136–130 ka occurs within a very narrow range in Mg/Ca (4.7 ± 0.37 (1 s) mmol/mol) (Supplementary Figs. 7, 8), confirming this δ13CNISA-transition is not driven by variable PCP. However, Mg/Ca in Garth rises abruptly after 129 ka and between 127 and 122 ka, the coincidence of maximum Mg/Ca ratios and strongly condensed growth indicated by 230Th ages is most likely explained by enhanced PCP (Supplementary Fig. 10). Between 127 and 122 ka, δ13CNISA may be shifted to more positive values than would be attained if PCP values were comparable to other sections of the stalagmite and therefore, PCP may be attenuating the magnitude of change in δ13CNISA during peak MIS6e warmth in this stalagmite. Consequently, we rely on other stalagmite record for this interval. Stalagmite Gael, shows overall more constant Mg/Ca_mgcalc than Garth (Supplementary Fig. 8), suggesting that relative trends in δ13CNISA between early termination (134–130 ka) and peak MIS 5e (127–123 ka) are less influenced by PCP in Gael. At the same time, the trend in δ13CNISA at the termination onset may be amplified in Gael by decrease in PCP between 135 and 134 ka. In stalagmite Neith, Mg/Ca and δ13CNISA correlate strongly within glacial and termination time windows (Supplementary Fig. 8) and although a shift in mean δ13CNISA between glacial and interglacial is evident, PCP may be the dominant source of temporal variation in δ13CNISA within each of these time windows, so we abstain from interpreting the time series of δ13CNISA in Neith. We propose stalagmite Garth, with high and relatively stable Mg/Ca_mgcalc and most precise chronology, to present the most reliable δ13CNISA record, except during its condensed interval between 128 and 122 ka; for this latter period, we splice in the temporal variation observed in stalagmite Gael so that long term trends can be accurately assessed.

We also use the Mg/Ca index to evaluate the potential effect of varying PCP on δ18O, since stalagmite δ18O may be isotopically enriched when forming from dripwaters which have experienced significant PCP at rates exceeding those of oxygen isotope equilibrium between dissolved inorganic carbon and water, as observed in laboratory studies56,58. The magnitude of this effect in cave settings could be buffered by the large water reservoir exchanging with DIC especially if carbonic anhydrase from soil and cave microbial populations catalyzes exchange. Using previously published δ18ONISA records from 100 to 80 ka57, we find that a lower Mg/Ca_mgcalc coincides with stalagmite sections offset to higher δ18ONISA, relative to overall stalagmites, suggesting this index is useful to identify PCP-related influences on δ18ONISA in these cave settings (Supplementary Fig. 7). In stalagmites spanning TI, the major temporal trends in our δ18ONISA records do not correlate with Mg/Ca_mgcalc in the intervals of each record selected for the splice (Supplementary Fig. 2). In the spliced record overall, there is no systematic correlation between Mg/Caindex and the decrease in δ18O. Therefore there is no evidence the δ18O trends are artefacts of temporally variable PCP.

During TII, we evaluate whether deglacial δ18ONISA changes coincide with changes in cave-influences (such as PCP) using crossopt selected from isotope records with Marcia Ca (Supplementary Fig. 8). We color code data by age. This analysis confirms that in Garth, there is no correlation between Mg/Ca and δ18OA of the full transition of δ18ONISA from 136 to 132 ka occurs when the speleothem maintains a very narrow range in Mg/Ca (4.8 ± 0.35 (1 s) mmol/mol), suggesting that variation in PCP is not a significant influence on the trend in δ18ONISA in this time interval. Over the termination there is also no correlation between growth rate and δ18ONISA in Garth, the stalagmite for which growth rate is most precisely resolved due to annual layer counting (Supplementary Fig. 8). Although the timing and features of the δ18ONISA are similar in stalagmite Gael, its δ18ONISA is on average 1% higher compared to stalagmite Garth or Neith. Gael has a lower average Mg/Ca_mgcalc than the other two stalagmites, suggestive of greater PCP which could lead to offset of the absolute δ18ONISA for stalagmite Gael. Additionally in Gael, the shift to slightly lower mean value of Mg/Ca midway through the termination onset coincides with a reduced offset in δ18ONISA between Gaal and Garth (Supplementary Fig. 9), amplifying the δ18ONISA transition in Gael between 135 and 134 ka by about 0.5%. In Neith, there is no correlation between δ18ONISA and Mg/Ca, and for Gael and Neith, growth rates are less precisely constrained but there is no consistent correlation between average growth rate and δ18ONISA. Therefore, we employ stalagmite Garth, with high and relatively stable Mg/Ca_mgcalc and most precise chronology, as our core δ18ONISA record, except during its condensed interval between 128 and 122 ka, when the period of most significant PCP may increase the δ18ONISA compared to other time intervals in that stalagmite. During the interval of condensed growth in Garth, we splice in the temporal variation observed in stalagmite Gael, plotting them on a separate y-axis so that long term trends can be accurately assessed. We employ the record from stalagmite Gael to estimate the eastern North Atlantic surface ocean δ18Osw anomaly relative to the Iberian margin δ18Osw (BC; Methods).

**Estimation of δ18Osw relationship with δ18ONISA over TI and calculation of surface ocean anomaly.** The δ18O of surface seawater (δ18Osw) is commonly estimated from coupled Mg/Ca and δ18O of planktic foraminifera59. We compare published marine δ18Osw with δ18ONISA over the time interval from 25 to 5 ka BP in fixed 500 year time bins (Fig. 2, Supplementary Fig. 3). The marine records are based on original 14C chronology which for the S Iberian margin assumes constant
Tuning of Northern Caribbean marine age models for Iberian Margin to speleothem chronology. We use ODP 999 using its δ¹⁸O record during the termination and using its δ¹⁸Oplanktic record during the interglacial. δ¹⁸Oplanktic tuning of TII is based on observed lags during TI. During TI, the δ¹⁸Ow of ODP 999 is based on 14 C age model, updated to INTCAL 2013 with constant 400 year reservoir age. During TI the W Iberian margin δ¹⁸Oplanktic is on chronology tuned to GRIP and Hulu. We observe that depletion of δ¹⁸Ow at ODP 999 lags that of W Iberian margin δ¹⁸Oplanktic by an average of 1100 years (Supplementary Fig. 17a). This lag is consistent with timescales of ocean mixing required to distribute the ice volume. Therefore the same lag is used on TI to align the δ¹⁸Ow and W Iberian margin δ¹⁸Oplanktic and assign chronology to ODP 999. We do not use the same δ¹⁸Oplanktic for tuning during the termination because during the phase of meltwater addition, δ¹⁸Oplanktic would be expected to have contrasting temperature components on the Iberian margin (strong cooling linked to AMOC shutdown) and western Caribbean (no cooling from AMOC shutdown). However, during the interglacial, these two sub-tropical locations may be expected to have similar evolution of SST and δ¹⁸Ow. Therefore we use the combined signal of δ¹⁸Oplanktic to serve directly for correlation during MIS 5e (Supplementary Fig. 17b).

Estimation of terrestrial and marine components of the PGM EIS. We approximate the ‘marine’ proportion of the TI and T2 Eurasian ice sheet using the ice sheet reconstructions δ¹⁸OIS, and an ice history that is the same as ICE-5G for T2. The sea level history and glacial isostatic adjustment modelling is from 6, using the same ice sheet histories. ‘Marine’ ice is defined here simply as the volume of ice that is grounded below the reconstructed sea level.

Reporting summary. Further information on research design is available in the Nature Research Reporting Summary linked to this article.

Data availability. The absolute U-Series and Radiocarbon dates are provided in the Supplementary Information. The Speleothem stable isotope data and interpolated ages are submitted to the SISAL database, and additionally are available from the ETH Research data archive at doi: 10.3929/ethz-b-00054925.

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