

Interplay of North Atlantic Freshening and Deep Convection During the Last Deglaciation Constrained by Iberian Speleothems

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Abstract. The last deglaciation featured abrupt climate shifts driven by interactions among Earth system components, notably retreating ice sheets and meltwater input. While globally detected, the magnitude, timing, and sequence of North Atlantic ~~source-meltwater~~ events remain uncertain. We present a Uranium-Thorium-dated stalagmite from northwestern Iberia spanning 24–12 ka BP, capturing both the impact of North Atlantic meltwater on surface ocean chemistry and regional air temperature changes. Our record reveals primarily gradual meltwater inflow during the Last Glacial Maximum and early deglaciation (about 20.8–18.2 ka BP), followed by abrupt increases during Heinrich Stadial 1. ~~An abrupt cooling lags~~ The first abrupt cooling is decoupled from the first meltwater pulse ~~by ca. 850 years, unlike~~, appearing around 810 years later— unlike the timing observed for later pulses. This evolving relationship between meltwater and cooling provides new constraints on the changing sensitivity of deep Atlantic convection to meltwater input throughout the deglaciation.

10 1 Introduction

The last deglaciation started in the Northern Hemisphere (NH) around 19 thousand years before present (ka BP), and marks the transition from the Last Glacial Maximum (LGM), lasting from approximately 23 to 18 ka before present (BP), into the Holocene (Clark et al., 2012). During the LGM, the NH was covered by large ice sheet complexes (Fig. 1a) covering much of North America and Eurasia (Stokes et al., 2012; Hughes et al., 2016). An increase in boreal summer insolation triggered their retreat and during their subsequent deglaciation there were episodically large amounts of meltwater (MW) and icebergs delivered into the North Atlantic (e.g., Toucanne et al. (2015); Hodell et al. (2017); Lin et al. (2021); Zhou and McManus (2024)).

It is well established, both conceptually (Denton et al., 2010; Alvarez-Solas and Ramstein, 2011; Menviel et al., 2020) and by modeling studies (~~Jackson et al., 2015; ?; Romé et al., 2022~~) (Jackson et al., 2015; Klockmann et al., 2018; Romé et al., 2022), that MW contributions can cause an abrupt weakening of the Atlantic Meridional Overturning Circulation (AMOC) – the zonal integral of the surface and deep currents in the Atlantic - if the Atlantic is conditioned for instability. Such weakening, in turn,

would lead to hemispheric-scale cooling (~~warming~~) in the north (~~south~~), [warming in the south](#) and trigger atmospheric CO₂ rise (Broecker et al., 2010). While there is evidence for several episodes of AMOC weakening and NH cooling during the deglaciation (~~Mcmanus et al., 2004; Shakun et al., 2012; ?~~) ([Mcmanus et al., 2004; Shakun et al., 2012; Ng et al., 2018](#)), linking these
25 to specific melt discharge events remains challenging. Uncertainties in ice geometry, even at the scale of total volume and global distribution (Stokes et al., 2015), propagate into the derived meltwater histories mostly used for transient deglacial simulations (Bethke et al., 2012; Sherriff-Tadano et al., 2018; Kapsch et al., 2022; Snoll et al., 2024). Furthermore, direct evidence for meltwater from palaeoclimatic records is sparse and retrieving accurate radiocarbon ages from marine cores on such short centennial time scales is hindered by the increased stratification during weak AMOC states (Peck et al., 2006; Peck,
30 2017). Generally, time scales relevant for melt events range from a few years to a few centuries, a time range in which the North Atlantic ocean and its adjoint ice sheets are not able to equilibrate (Wunsch and Heimbach (2008), Fig. 1c), emphasizing the need for more regional records to reconstruct their sequence and impact on the Earth system. Even more so, as modulations of AMOC strength depend on background climate (Zhu et al., 2015), and can occur directly forced by freshwater (~~Zhu et al., 2017a; Snoll et al., 2024~~) ([Zhu et al., 2017b; Snoll et al., 2024](#)), spontaneously (Romé et al., 2022; Armstrong et al.,
35 2023) or delayed (Alvarez-Solas and Ramstein, 2011; Romé et al., 2024). This adds an additional layer of complexity to the synchronization of regional melt events with global records.

To move forward our understanding of the relationship between melt-induced freshening and AMOC strength, we are building upon a direct freshening mechanism by detecting the ¹⁸O-depleted MW in speleothem $\delta^{18}\text{O}$ proximal to the melt water source regions, an effect previously documented in the Northwest Iberian Speleothem Archive (NISA) (Stoll et al., 2022)
40 and model simulations (~~Zhu et al., 2017a~~) ([Zhu et al., 2017b](#)). Here, we provide a new decadal-resolved isotopic record from NISA, 24-12 ka BP, to constrain the timing of MW addition and elucidate the relationship between melt-induced freshening, North Atlantic temperature changes and AMOC strength throughout the entire last deglaciation. Given the coastal location of our site and prevailing atmospheric conditions induced by Earth's rotation, the direct effect of $\delta^{18}\text{O}$ -depleted MW on the surface ocean is captured by $\delta^{18}\text{O}$ in the speleothem, allowing us to infer the history of NH MW at high temporal resolution.
45 In addition, we provide a coeval record of relative temperature from speleothem $\delta^{13}\text{C}$, which we have corrected for in-cave fractionation effects ($\delta^{13}\text{C}_{\text{init}}$). Since both stable isotopes are measured on the exact same samples, this allows a direct and high-resolution study of the temporal relationship between ~~NH-MW-ocean-in-flux~~ [the influx of NH MW to the ocean](#) and temperature change in the North Atlantic realm.

2 Material and Methods

50 2.1 Cave site, sample, preparation, and sampling procedure

Modern precipitation at the cave site is dominated by cool season rainfall, but many drips remain active also during the summer period of lower precipitation-evapotranspiration. Speleothem Glas has been selected for this detailed study because of the absence of any signs for non-continuous growth and for featuring a homogenous trace element structure, also when diverging from the main growth axis (SI Appendix, Fig ~~S1-S2~~, [A1](#)). We focus on the 4.2 cm interval of Glas bracketed by a

55 growth discontinuity on the older part and significant decrease in growth rate on the younger boundary (Fig. 2). Due to the slow
growth rate and the presence of abrupt climate changes over the early last deglaciation, the sampling strategy for U/Th dating
has been optimised by continuously micro-milling the sample with a 50 µm trench along growth layers, ~~measure-measuring~~
stable isotopes on these individual samples, before combining the remaining powder of around 10 samples to retrieve one date.
This approach is superior when aiming for high temporal resolution in a slow-growing stalagmite because it eliminates possible
60 mismatches when aligning the ~~mid-midpoint~~ of a dating hole to the isotope track. Samples were collected by a trench using the
ESI NewWave microdrill ~~and the trench was programmed~~. ~~The trench drill track was pre-programmed~~ to follow growth layers
within ~~Fiji~~ the Image Processing Software Fiji (Schindelin et al., 2012), based on prior analysis of microscopic and confocal
laser scanning microscopy. The new trench was aligned to an older 100 µm trench for quality control and aligning trace element
measurements (SI Appendix, Fig. S7).

65 2.2 U-Th dating and age model

Despite small sample sizes (20-40 mg), good analytical errors of less than 100 years have been achieved in each sample. U-
Th samples were analyzed with a ThermoFisher Neptune plus Multicollector Inductively Coupled Plasma Mass Spectrometer
(MC-ICP-MS) at Xi'an Jiaotong University, following the methodology and decay constants reported in Cheng et al. (2013),
see ~~Tab. S1. The age model used in this study~~ SI Appendix Tab. A1.

70 The age-depth model (Fig. 2) has been computed ~~with BChron (Haslett and Parnell, 2008), and has been optimised for per~~
the algorithm of Haslett and Parnell (2008), using the package BChron with 50'000 iterations, equal outlier probability of 0.1
for each sample and by explicitly providing the dating sample thickness as an additional constraint. The age uncertainties stated
in the paper are the 95% credible interval (CI) output of this statistical model. The chosen author age model that is used
for further analysis in this work is selected from the full ensemble of plausible posterior BChron age models as one of the
75 posterior realisations featuring the highest correlation with the Sr/Mg ratio, while also exhibiting a comparably small root
mean square error (RMSE) compared to the BChron median age model. An extended comparison plot between the differences
in the resulting growth rates available in the supplementary (SI Appendix, Fig. A2). The author age-depth model remains
within the 95% CI of the BChron model and correlates to Sr/Mg ratio with a pearson correlation coefficient of 0.6 (BChron
Median: 0.15). We consider this age model superior to the statistical median model as it allows to use Sr/Mg to inform on any
80 short term growth rate variations that are not resolved by the coarser U/Th sampling increment. This assessment is based on
prior studies documenting that Sr/Mg from this cave system correlates well with growth due to the growth rate dependence of
Sr partitioning (Sliwinski et al., 2023). Further, we investigated whether changes in the delivery of Mg from sea spray could
break this relationship by monitoring the delivery of sea aerosols to the cave system (Kost and Stoll, 2023): median modern
dripwater Na concentrations in the interior cave sectors where Glas was collected are 5 ppm, thus for the minimum stalagmite
85 Mg/Ca at 9 mmol/mol (D_{Mg} of 0.0225) this Mg from marine aerosols would constitute 3 to 6% of the dripwater Mg for open
system dissolution in the range of 600 to 8000 ppmv CO_2 , leading to variations in the drip water Mg/Ca and Sr/Mg of about
3% between glacial and interglacial endmembers, a very small variation compared to the measured Sr/Mg range of 0.002 -
0.006 (a 40% increase) in this sample.

2.3 Stable isotope and trace element ratios measurements

90 Stable isotope ratios were measured at ETH Zurich with a Thermo-Finnegan Delta V Plus coupled to Gas Bench II (Breitenbach
et al., 2015). Carbonate powder for trace element analysis were dissolved in 2% HNO₃ and analyzed for trace element to
calcium ratios employing an Agilent 8800 QQQ ICP-MS at ETH. Samples and standards were run at Ca concentrations of 400
ppm. Calibration used matrix-matched standards prepared from single element standard solutions to cover the range of trace
element to Ca ratios found in stalagmite samples. Our calibration standard composition accounts for trace element impurities in
95 the Ca standard, which were determined using standard additions. Calibration was conducted offline using the intensity ratio
method described by De Villiers et al. (2002). Analytical drifts were corrected with standards run after every 10 unknowns.
Precision on Mg/Ca and Sr/Ca ratios is < 2%(1σ).

2.4 Correction for in-cave fractionation

Prior calcite precipitation (PCP) occurs when parts of the initial dripwater solution already supersaturate and precipitate due
100 to the exchange with cave air before reaching the stalagmite. PCP can alter the δ¹³C signature significantly, by preferentially
removing ¹²C from the remaining dripwater solution (Rudzka et al., 2011; Fohlmeister et al., 2020). To focus on the temperature
signal preserved in soil CO₂ and consequently δ¹³C, the record was corrected for PCP by using the Mg/Ca ratio as a guide
line for the amount of degassing occurring, following the methodology of Stoll et al. (2023), with details being described in
SI appendix. In contrast, in most instances, the actual bedrock δ¹³C is of minor relevance, as the system can be assumed to be
105 rather open and therefore the CO₂ in the soil overwhelms the rock signature (Lechleitner et al., 2021).

2.5 ~~Breakpoint determination~~ Event nomenclature, midpoint and duration estimation

~~We constructed a piecewise linear model to determine the breakpoints using a smoothed 200-year rolling mean time series
with the R package 'segmented'. The algorithm uses Taylor expansion to iteratively approach the best estimate, starting from
initial breakpoint guesses and providing standard errors for each breakpoint (Muggeo, 2003). For this study, To estimate the~~
110 timing and rate of the key transitions discussed in the manuscript, we propagated age and measurement uncertainties using a
Monte Carlo errors-in-variables framework. For each of the 6000 age-depth posterior realizations derived from the Bayesian
age model (BChron, Haslett and Parnell (2008)), we extracted a start and end age corresponding to the ~~plotted uncertainty~~
~~is given as~~ two bracketing depths of a given transition. Measurement uncertainty in δ¹⁸O (±0.1‰, 1σ) was incorporated by
adding Gaussian noise to the ~~integrated 2σ uncertainty of the breakpoint estimation and the age model ensemble.~~ observed
115 endpoint values in each realization. For the author age model and each posterior draw i, we computed the midpoint as
age_{mid,i} ≡ (age_{end} + age_{start})/2 and the rate of change by computing the slope β_i = (δ¹⁸O_{end,i} - δ¹⁸O_{start,i})/(age_{end,i} - age_{start,i}).
Duration of an event was estimated similarly by subtracting the end from the start age in all posterior realisations. Uncertainty
estimates (95% CI) were obtained directly from the empirical distributions of these quantities across all posterior realizations,
while the event names are derived from the author age model. To assess robustness to the exact choice of bracketing depths, we
120 repeated the analysis using adjacent samples surrounding each transition. The resulting midpoint ages and durations remained

within the 95% CI of the original estimates, indicating that the inferred timing and rates are not sensitive to small variations when selecting the depth interval. Further, the estimates of the author age model also remain within the 95% CI for all values, which gives further support that the author age model is consistent with all data and not an outlier realisation. The summarized results are given in Tab. 1 and are visualised in Fig. A3.

125 2.6 Climate model experiments and NISA Melt Source Contribution Index

We reemploy an already published HadCM3 model simulation (Romé et al., 2022) to examine the progression of a freshwater anomaly from different source origins and under different AMOC states to the NISA location, ~~the model is described in the Appendix.~~ The original simulation includes 10,000 model years and captures a glacial climate state with an AMOC that oscillates between a strong (relatively ²‘warm’ climate) and a weak (relatively ²‘cold’ climate) state, with about 1,500 year
130 periodicity, triggered by a constant meltwater flux corresponding to a reconstructed ice sheet history at 17.8 ka BP (GLAC-1D, Ivanovic et al. (2016)). ~~The,~~ with more detailed model description in the Appendix. To study the effect of an AMOC decline on the regional climate above the cave, we have computed the seasonal averages of surface air temperature and precipitation - ~~evaporation~~evapotranspiration, plotted in Fig. 1c-d, ~~were computed within for~~ the region confined by ~~an a~~ red border in Fig. 1b. To distinguish between a strong and a weak state AMOC within the simulation, we followed the methodology of Romé et al.
135 (2025), where the years that fulfilled each condition were selected based on changes in the mixed layer depth (MLD) in key convection sites (Irminger and GIN Seas). In these simulations, Irminger Sea MLD is in the range 17-25 m during the weak AMOC state, deepening to 25-65 m during the strong AMOC state. GIN seas MLD is 27-40 m and 65-105 m, respectively. From the full simulation, 2,367 years fulfilled the ‘weak’ condition (average AMOC strength at 26.5°N: 8 Sv), and 1,488 years fulfilled the ²‘strong’ condition (16 Sv), coined ²‘zonal’ in Romé et al. (2022).

140 Segments of the simulation were re-run for our study, adding a set of passive ocean tracers tagging the different likely meltwater source regions to assess the potential of each region to influence the isotope signal at our study site. The results are ~~shown~~presented as a qualitative NISA Melt Source ~~Sensitivity-Contribution~~ Index (Fig. 1e). ~~To~~We calculate the index ~~we use the-by scaling the model’s surface-ocean dye anomalies after 200–300 years of continuous injection (Fig. 1b) using~~ estimated moisture uptake at ~~our site from a hysplit~~ the site and provide a percentage estimate of three key delivery regions to the full
145 signal at NISA expected for a strong AMOC mode. The uptake estimates are derived from a HYSPLIT analysis based on rainfall in-at the NISA location El Pindal from 2015-2016 (González, 2019)to scale the uptake from the model’s surface-ocean dye anomalies (Fig. 1b). The index, then calculated as a percentage of the mean total arriving meltwater at the study site during a strong AMOC state after 200-300 years of continuous injection, gives a relative estimate of the importance and the variability of the signal in the surface ocean during 2015–2016 (González, 2019). The index is not tied to a specific meltwater scenario;
150 instead, it provides a relative estimate of how efficiently meltwater from a given northern ice sheet sector can reach our study site under varying AMOC circulation modes. The constant meltwater flux that was applied in the underlying simulations has the consequence that lag times between meltwater injection and AMOC decline can not be assessed for these simulations. However, both, AMOC theory (e.g., Barker and Knorr (2021)) and GCM model results (e.g., Romé et al. (2025)), suggest that the response of AMOC to a specific meltwater forcing is likely climate and ocean background state dependent and includes

155 the possibility of a delayed response. For example, a recent high resolution GCM model found AMOC collapse occurred 1750
model years after the onset of a gradually increasing North Atlantic freshwater forcing (Van Westen et al., 2024). As we argue
here, NISA speleothem data provides a direct pathway to investigate such lag times by comparing the two proxy time series
 $\delta^{18}O$ and $\delta^{13}C_{init}$, thus contributing to the understanding of AMOC strength sensitivity to a meltwater forcing under evolving
background conditions.

160 3 Results and Discussion

3.1 Sensitivity of NISA to surface ocean freshening and cooling

At NISA cave locations (Fig. 1), precipitation is mainly controlled by westerly winds and the passage of Atlantic fronts while the Cantabrian mountain ridge reduces the influence of water vapor transport from the south (Moreno et al., 2021), making rainfall from the proximal North Atlantic the dominant moisture source. Over the penultimate and the last deglaciation, speleothem $\delta^{18}O_{NISA}$ closely tracks the long term trend in $\delta^{18}O$ of seawater ($\delta^{18}O_{sw}$) reconstructed from marine archives from the eastern North Atlantic (Stoll et al., 2022). This strong response to the moisture source composition occurs because in this setting, the temperature-dependent fractionation of $\delta^{18}O$ into calcite is of a similar magnitude but opposite sign to the temperature dependence of $\delta^{18}O$ in precipitation (McDermott et al., 2006; Stoll et al., 2015), which mitigates additional speleothem $\delta^{18}O$ variability due to temperature changes. Additionally, due to the coastal cave location, effects on $\delta^{18}O$ due to changes in altitude or prior rainout are minimized.

On a centennial time scale, a freshwater anomaly originating from a certain ice sheet sector will not spread evenly across latitudes and depths. This may amplify or attenuate the impact of a given meltwater flux on the proxy archive. To test the dependency of source origin and AMOC background state, the distribution of meltwater from different NH ice sheet sectors has been tracked by adding conservative dye tracers to the simulations of Romé et al. (2022). We compute the relative intensity of the tracer in the proximal sector of the modern NISA moisture source region (Fig. 1b). The sensitivity of the tracer to the source region and AMOC state is expressed as the NISA Melt Source ~~Sensitivity-Contribution~~ Index (Methods, SI Appendix). The Melt Source ~~Sensitivity-Contribution~~ Index demonstrates that for a given meltwater flux (Fig. 1b, e), the signal from ice sheet meltwater reaching the NISA sites is higher when AMOC is in a weak state than in a strong state, regardless of the melt source region. This is because a stronger AMOC more rapidly redistributes meltwater throughout the global ocean. Furthermore, we can identify that under the weak AMOC state, the tracked NISA source region more strongly accumulates meltwater of Eurasian origin delivered to the GIN ~~seas (of Eurasian origin)~~ Seas and the Eastern North Atlantic, whereas meltwater discharged by the Laurentide ice sheet to the western North Atlantic more strongly accumulates in the tracked region under strong AMOC conditions (Fig.1b). For both, weak and strong AMOC, meltwater anomalies originating from the Arctic will already largely propagate to the subsurface waters and have limited impact on the surface North Atlantic $\delta^{18}O_{sw}$.

185 ~~Northwestern Iberia is also particularly sensitive to AMOC-driven changes in temperature (global maps in SI Appendix, Fig. S3). A switch from a strong to weak AMOC state in the early deglaciation drives surface air temperature changes of up to $-5^{\circ}C$ in NW Iberia for all seasons in the same simulations (Fig. 1e). In contrast, the moisture availability within the region~~

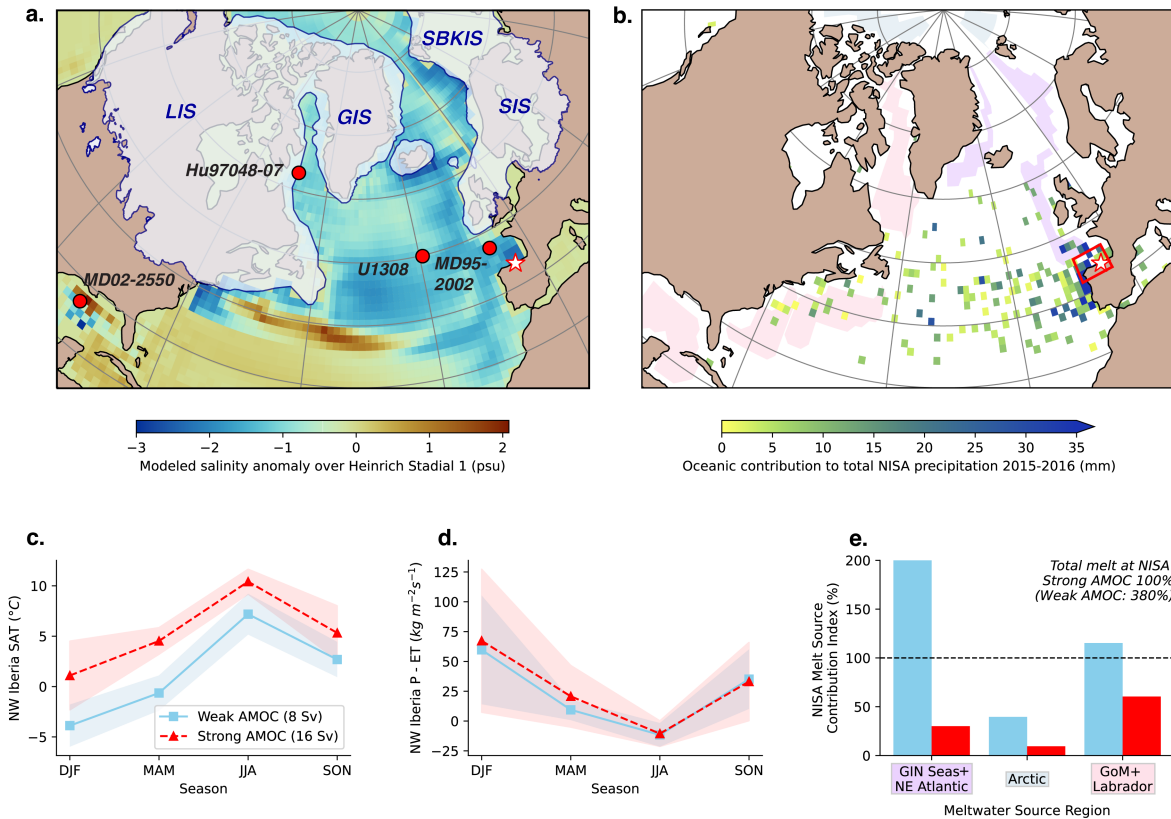


Figure 1. Overview of sites referenced in this study, the expected meltwater signal originating from different ocean basins, and the expected environmental changes related to an AMOC decline. **(a.)** A Map including the study site NISA (red-rimmed white star) and other sediment cores used in this study for proxy comparison (red circles), namely MD95-2002 (Toucanne et al., 2015), U1308 (Hodell et al., 2017), Hu97048-07 (Rashid et al., 2012), and MD02-2550 (Wickert et al., 2023). The maximum LGM ice sheet extent (Batchelor et al., 2019) (grey, blue border) is illustrated for the main northern ice sheet sectors: Laurentide (LIS), Greenland (GIS) and the European Ice sheet (EIS) sectors: Svalbard-Barents-Kara (SBKIS), Scandinavian (SIS) and British-Isles (BIIS). Ocean colouring illustrates the simulated mean annual sea surface salinity anomaly expected for Heinrich Stadial 1 (Ivanovic et al., 2018) due to meltwater input and associated AMOC decline. **(b.)** A Map showing the oceanic moisture source areas from a Hysplit analysis on precipitation monitoring data at NISA cave sites (red-rimmed white star) from 2015-2016 (González, 2019). The colored regions denote three composite Northern hemisphere ice sheet's meltwater source regions: the GIN Seas and northeastern North Atlantic (purple), the Arctic Ocean (petrol light green), and the Gulf of Mexico and Labrador Sea (orange). These regions were further used to compute the regional NISA Melt Source Sensitivity-Contribution Indexes in (e.). **(c.,d.)** Based on a simulation simulating regional meltwater discharge at 17.8 ka, re-employed from Romé et al. (2022), the expected difference at our NISA study site (averaged over the red rectangular region) between different AMOC states for (c.) surface air temperature and (d.) moisture availability (precipitation - evapotranspiration; P-ET). **(e.)** Barplot shows the NISA Melt Source Sensitivity-Contribution Index (SI Appendix, Methods); a relative mean contribution estimate of how much meltwater reaches the NISA location from each of the three regions after 200-300 years of continuous injection during a weak (blue) and strong (red) AMOC state, based on a simulation using meltwater input at 17.8 ka BP (Romé et al., 2022).

~~remains similar across such transitions (Fig. 1d). In~~ our record, $\delta^{13}\text{C}_{\text{init}}$ provides a qualitative measure of regional surface air temperature. This is because the $\delta^{13}\text{C}$ of cave dripwater is initially set by the exchange of infiltrating water with CO_2 in the soil and epikarst. In moisture-replete regions such as NW Iberia, warmer temperatures stimulate higher heterotrophic and autotrophic respiration rates and raise soil CO_2 (Romero-Mujalli et al., 2019). Because respired CO_2 is much lower in $\delta^{13}\text{C}$ than atmospheric CO_2 , warm periods of high soil CO_2 lead to a more negative $\delta^{13}\text{C}$ signature in cave dripwater. While in-cave processes of degassing and prior calcite precipitation (PCP) can in some cases subsequently modify the $\delta^{13}\text{C}$ signature in cave dripwater (Mühlinghaus et al., 2007; Mickler et al., 2019), recent approaches using Mg/Ca as an independent PCP indicator allow calculation of the initial $\delta^{13}\text{C}_{\text{init}}$ defined by the soil and vegetation processes from the measured $\delta^{13}\text{C}$ (Stoll et al., 2023; Lechleitner et al., 2021). ~~We thus apply this technique to derive the temperature sensitive~~ (parameters are given in SI Appendix). ~~The relative temperature indicator $\delta^{13}\text{C}_{\text{init}}$ (Methods, SI Appendix) is likely to be more sensitive in a colder climate, as then the overall soil CO_2 is lower and $\delta^{13}\text{C}$ of soil CO_2 follows the Keeling mixing line (Kost et al., 2022).~~

The temperature sensitive indicator $\delta^{13}\text{C}_{\text{init}}$ is critical for this work, because Northwestern Iberia is also particularly sensitive to AMOC-driven changes in temperature (global maps in SI Appendix, Fig. S3). A switch from a strong to weak AMOC state in the early deglaciation drives surface air temperature changes of up to -5°C in NW Iberia for all seasons in our simulations (Fig. 1c), a finding in agreement with previous studies highlighting the Eastern North Atlantic sensitivity to AMOC strength variation in the context of millennial-scale variability (Zhang et al., 2015). In contrast, the moisture availability within the region remains similar across such transitions (Fig. 1d). Over the entire last deglaciation other processes besides AMOC variability may have influenced North Iberian surface temperatures such as changes in arctic sea ice extent and export (Condrón et al., 2020), atmospheric circulation (Ludwig et al., 2016; Bellomo et al., 2023) as well as further changes in global climate feedbacks and insolation (e.g., Alley and Clark (1999)). Thus, attributing the well-dated NISA temperature changes to a change in AMOC strength follows a combined assessment with other proxy archives indicating variation in AMOC strength in the studied interval (Shakun et al., 2012; Ng et al., 2018; Pöppelmeier et al., 2023).

210 3.2 The deglacial North Atlantic climate record of Glas

The stalagmite named ‘Glas’ originates from the NISA cave ‘La Vallina’ (4.8067°W , 43.4100°N), located 4 km from the modern coastline. Twenty-five new U/Th dates, with analytical errors of less than 100 years, constrain the chronology of Glas between 24 and 12 ka; and potential variations in growth rates are assessed through comparison with Sr/Mg measurements (Fig. 2, Methods, SI Appendix, Tab. S1, Fig. S1). Annual growth rate averages $4\ \mu\text{m}$, but drops to $1\ \mu\text{m}$ around 13 ka. By using combined aliquots of the same sample powder for Uranium-Thorium dating as have been used for high-resolution stable isotope analysis, a fully consistent age model with minimized depth uncertainties is attained for the sequence of isotopic events despite the slow growth rate. The slow growth rates and the lack of seasonal laminations in Glas suggest that the speleothem has been growing from a matrix-fed drip, which, at the depth of the cave passage ($>20\ \text{m}$), may have continued dripping throughout the year. At the depth of the cave passage, the seasonal soil temperature cycle is muted, while overall temperatures are expected to remain above freezing even under a weak AMOC state (e.g., model results in Tapia et al. (2025), Fig. 8). Featuring this improved age model, Glas supersedes the prior NISA stack for the last deglaciation, but the main features of Glas have been

Table 1. Overview of the abrupt salinity and temperature transitions in stalagmite Glas, which are discussed further in this work. All events are named by their midpoint and their type, where ‘S’ and ‘T’ indicate variation in surface ocean salinity (related to freshwater input) and abrupt change in regional temperatures, respectively. The first value in the columns Midpoint, Duration and Slope indicate the value derived from the author age model, while the 95% CI from the BChron ensemble is given in brackets. The sample depth informs about the start and end point chosen to characterise the transition. This analysis was repeated using adjacent samples surrounding each transition, resulting in midpoint ages and durations remaining within the 95% CI of the original estimates. Computation of midpoints and their uncertainties are visualised in SI Appendix Fig. S3.

| Event | Midpoint (ka) | Duration (yr) | Slope ($\% \text{ yr}^{-1}$) | Indicator | Sample Depth (mm) |
|--------------------------|----------------------------|-----------------------|---------------------------------|---|----------------------|
| <u>23.4k_S</u> | <u>23.44 (23.33–24.05)</u> | <u>559 (341–1755)</u> | <u>0.0013 (0.0004–0.0024)</u> | <u>↑ $\delta^{18}\text{O}$ (re-salinization)</u> | <u>37.2 - 39.2</u> |
| <u>17.8k_S</u> | <u>17.84 (17.82–18.03)</u> | <u>313 (109–474)</u> | <u>-0.0025 (-0.0072–0.0014)</u> | <u>↓ $\delta^{18}\text{O}$ (freshening)</u> | <u>20.2 - 20.95</u> |
| <u>16.4k_S</u> | <u>16.44 (16.14–16.55)</u> | <u>68 (20–200)</u> | <u>-0.0076 (-0.0276–0.0022)</u> | <u>↓ $\delta^{18}\text{O}$ (freshening)</u> | <u>15.35 - 15.5</u> |
| <u>16.1k_S</u> | <u>16.07 (15.90–16.38)</u> | <u>136 (37–182)</u> | <u>-0.0060 (-0.0226–0.0036)</u> | <u>↓ $\delta^{18}\text{O}$ (freshening)</u> | <u>14.45 - 14.75</u> |
| <u>15.3k_S</u> | <u>15.27 (15.14–15.36)</u> | <u>221 (128–493)</u> | <u>-0.0039 (-0.0070–0.0015)</u> | <u>↓ $\delta^{18}\text{O}$ (freshening)</u> | <u>12.2 - 12.8</u> |
| <u>14.7k_S</u> | <u>14.68 (14.54–14.81)</u> | <u>113 (38–187)</u> | <u>0.0059 (0.0029–0.0182)</u> | <u>↑ $\delta^{18}\text{O}$ (re-salinization)</u> | <u>11.1 - 11.45</u> |
| <u>23.6k_T</u> | <u>23.64 (23.49–24.32)</u> | <u>348 (68–1643)</u> | <u>-0.0088 (-0.0455–0.0019)</u> | <u>↓ $\delta^{13}\text{C}_{init}$ (warming)</u> | <u>10.70 - 11.50</u> |
| <u>17.0k_T</u> | <u>17.03 (17.00–17.39)</u> | <u>148 (90–528)</u> | <u>0.0057 (0.0015–0.0094)</u> | <u>↑ $\delta^{13}\text{C}_{init}$ (cooling)</u> | <u>17.75 - 18.30</u> |
| <u>14.6k_T</u> | <u>14.62 (14.53–14.76)</u> | <u>260 (103–359)</u> | <u>-0.0129 (-0.0324–0.0093)</u> | <u>↓ $\delta^{13}\text{C}_{init}$ (warming)</u> | <u>38.20 - 39.60</u> |

validated in other NISA records (SI Appendix, Fig. S2)–A4). The record features a several abrupt transitions within $\delta^{18}\text{O}$ and $\delta^{13}\text{C}_{init}$, which are summarized in Tab. 1.

3.3 North Atlantic surface ocean freshening history

225 3.3.1 LGM evolution

Our new record from Glas captures the end of Heinrich Stadial 2 (HS2) with a trend to more saline conditions in the North Atlantic indicated by the shift to a heavier $\delta^{18}\text{O}$ signature ~~and~~ (event 23.4k_S, see Tab. 1 and Fig. 3), and the coincident rapid recovery to warmer temperatures evident by a shift to more negative $\delta^{13}\text{C}_{init}$ (Fig. 3 event 23.6k_T). The observed warming likely reflects a strengthening of AMOC after HS2, which results in a more efficient dispersal/removal of $\delta^{18}\text{O}$ anomalies from the surface ocean ~~and is in accordance with the previously discussed~~, a process highlighted by our modeling results for which the NISA melt source ~~sensitivity index~~ Contribution Index is dampened if the AMOC is stronger for all meltwater source origin sites (Fig. 1e). Our chronology is consistent with the proposition that the HS2 phase of Southern BIIS retreat and associated enhanced MW flux to the ocean ended around 23.5 ka, aligning with the Southern European Ice sheet (EIS) events found in sediment cores (Toucanne et al., 2015). The ice sheet model GLAC-1D (Ivanovic et al., 2016), which has been used to force transient deglacial climate simulations in a number of studies previously (Snoll et al., 2024), suggests a slightly later

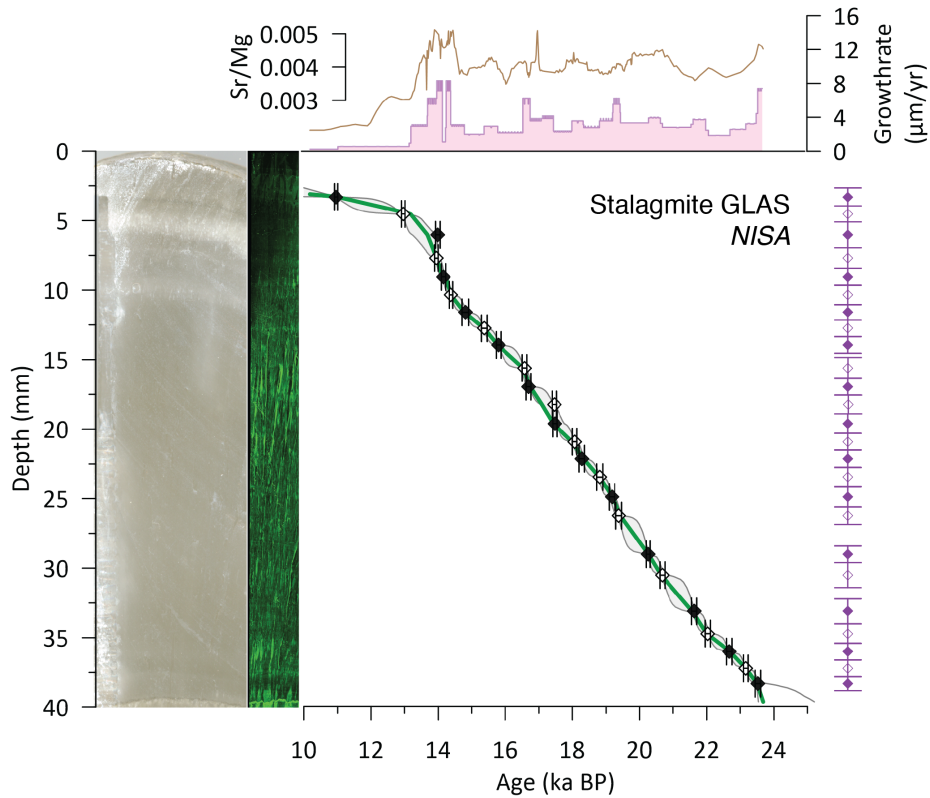


Figure 2. Age model of NISA Stalagmite Glas. Main panel from left to right: scan of the slab used for analysis (grey shading); the corresponding Confocal laser scanning microscopy image (green shading); and the constructed BChron age model (green line) used in all further analysis (Methods), based on the U-Th ages (yellow dots/black diamonds, bounded by horizontal errorbars denoting the 2σ uncertainty); and, on the right, actual depth of the combined, original samples used in this study (purple vertical errorbar). Open and filled symbols were chosen to simplify retrieving the width of a specific dating hole by comparing directly with the purple vertical errorbar indicators. Note that our sampling procedure ensures no shift between the assigned depths for dating and isotope track. In the top panel, growth rate is calculated based on the BChron age model (pink fill), and we use Sr/Mg ratio (brown line) as an additional indicator for changes in growth rate to confirm the age model from the U/Th dating sequence. The pearson correlation coefficient between Sr/Mg and growth rate of the author age model is 0.6.

timing of peak meltwater delivery from the Southern Laurentide ice sheet (LIS) via the Labrador Sea ~~+GoM outlets and the~~ Gulf of Mexico as well as some meltwater discharging into the Arctic region (Fig. 3a-b.; extended discussion and ice sheet mass loss maps in SI Appendix, Fig. S5). Thus, the new Glas record suggests that either the GLAC-1D model shows these elevated melt events around 500 years too late, or despite a persistently high meltwater discharge, AMOC has restrengthened and successfully dissipated the surface $\delta^{18}\text{O}$ anomaly from a LIS meltwater pulse postdating HS2.

A subsequent episode of melting from 22.5 to 21.5 ka is suggested by a small, gradual decrease in $\delta^{18}\text{O}$ of -0.3‰ , which returns to previous levels between 21.5 and 20.5 ka. Sustained freshening begins around 20.5 ka, marked by a gradual decrease of -0.5‰ $\delta^{18}\text{O}$ over 2 ky (Fig. 3). This progressive freshening confirms the presence of MW in the eastern North Atlantic, for which a source could be melt from the Scandinavian ice-sheet (SIS) as previously suggested by Nd isotope provenance of detrital minerals delivered into the Bay of Biscay (Toucanne et al., 2015).

3.3.2 Abrupt freshwater ~~onset~~ increase at 18.04-17.80 ka

An abrupt freshening (-0.58‰ $\delta^{18}\text{O}$ in ~~382^{+164}_{-94} years~~) starts at $18.04^{+0.16}_{-0.09}$ kaBP, 313^{+161}_{-204} years, event 17.8ks in Tab. 1) occurs centered around its midpoint at 17.8 ka BP and which we interpret as the onset of Heinrich stadial 1; ~~HS1a in Fig 3~~, with other regional Iberian speleothem records also indicating negative $\delta^{18}\text{O}$ shift around this time (~~SI Appendix OST2, Bernal-Wormull et al. (2021)~~). This freshening may reflect accelerated ice sheet melting in various regions (Fig. 3c). For example, there is evidence of elevated freshwater drainage from the Southern Laurentide to the Gulf of Mexico (Wickert et al., 2023), and reconstructions of EIS evolution (Hughes et al., 2016; Clark et al., 2022) suggest a strong likelihood that SIS and BIIS separated by 18 ka BP. The river-routed GLAC-1D ice model time series suggests discharge mainly into the Arctic, but since our dye tracking indicates that freshwater discharged into the Arctic produces only a weak signal in our speleothem record (Fig. 1e), the strong and abrupt freshening signal recorded by Glas at ~~18.04 kaBP rather indicates 17.8 ka BP rather suggests~~ that at least part of the ~~melt from Svalbard-Barents-Kara ice sheet (SBKIS) retreat was associated melt from the overall EIS retreat, visualised in the maps in Fig. 3a, were~~ delivered to the GIN Seas and ~~/or down the Channel River Eastern North Atlantic directly~~. This route would be consistent with the high terrestrial input through the channel river recorded at that time (Toucanne et al., 2015). Although the NISA Melt Source Contribution Index (Fig. 1e) shows that the elevated freshening levels could alternatively be explained by a simultaneous weakening of AMOC amplifying the signal in the North Atlantic surface ocean, on the identical chronology of GLAS we find no evidence of a major AMOC weakening at this time (see section 3.4.2).

The phase of elevated freshening is directly followed by a period of sustained low $\delta^{18}\text{O}$ (~~here HS1b~~), indicating that the sources and sinks of a MW-induced $\delta^{18}\text{O}$ anomaly were in balance over the next 1500 years. A negative $\delta^{18}\text{O}$ value from a freshwater pulse can be subsequently sustained despite lowered meltwater delivery if AMOC is weak, especially if the MW discharge region (i.e. GIN ~~seas~~ Seas and NE Atlantic) is close to our study site (Fig. 1b). In this case, sustained low $\delta^{18}\text{O}$ values in a weak AMOC state could potentially also mask limited ice sheet regrowth, as suggested in Wickert et al. (2023). Alternatively, marine records from the Hudson Strait indicate elevated meltwater delivery to the Atlantic (Rashid et al., 2012;

Denton et al., 2021), which could provide an additional source to sustain the Glas $\delta^{18}\text{O}$ anomaly provided that the AMOC
270 remains relatively vigorous.

3.3.3 Transient freshening from 16.5 ka through the Bølling-Allerød

A first smaller ~~transient event with onset at $16.5^{+0.09}_{-0.31}$ ka~~ excursion with a midpoint at 16.4 ka BP and a change of -0.5%
 $\delta^{18}\text{O}$, is followed by two larger events ~~at $16.22^{+0.24}_{-0.13}$ (HS1e in Fig. 3e) and $15.44^{+0.19}_{-0.25}$ ka~~ with transition midpoints at 16.1 and
275 ~~15.3 ka~~ BP with a shift around -0.79% $\delta^{18}\text{O}$ each ~~(Tab. 1)~~. All events are short-lived and dissipate from the subtropical
North Atlantic in a few hundred years. The latter two events coincide with the radiocarbon age estimates for two distinct
central North Atlantic ice rafted debris (IRD) peaks (Fig. 3c), which have been interpreted as two episodes of enhanced iceberg
calving from the Laurentide Ice Sheet related to Heinrich Event 1 (Hodell et al., 2017). Such a twin-peak structure has also been
achieved with simulations considering dynamic meltwater re-routing, which produces a second wave of enhanced Laurentide
iceberg calving around 1,000 years after the first episode (Ziemen et al., 2019). In addition, all events feature an abrupt onset
280 of the negative $\delta^{18}\text{O}$ anomaly but a gradual dissipation of the anomaly, which is consistent with a binge-purge mechanism
suggested for Heinrich events (MacAyeal, 1993). Alternatively, the timing of the later ~~$15.44^{+0.19}_{-0.25}$ ka~~ freshwater ~~15.3 ka BP~~
~~freshening~~ pulse is within age uncertainty of the most likely timing for Cordillerean-Laurentide (CIS-LIS) ice saddle collapse
($15.226^{+1}_{-0.81}$ ka; Norris et al. (2022)).

~~By 14.5 ka, the~~ At the onset of the Bølling-Allerød (BA), in our record centered at $14.68^{+0.13}_{-0.14}$ ka (Tab. 1), the subpolar North
285 Atlantic $\delta^{18}\text{O}$ had relaxed back-relaxes back quickly to values characterizing the plateau ~~of HS1b. However, the following the~~
freshening at 17.8 ka BP. This decline is analagous to the end of HS2 (23.4ka), where in both instances the AMOC likely
re-strengthened, promoting the accelerated removal of the surface ocean anomalies as predicted by our model (Fig. 1e). The
remainder of the Glas record (until 12.5 ka) exhibits higher variability in $\delta^{18}\text{O}$, with numerous high amplitude, short lived
freshening pulses. ~~We propose that this~~ While some of the variability might also be caused by the decrease in growth rate
290 at that point, we propose that the continued ice sheet shrinkage (Fig. 3a) causes a greater vulnerability to climate change,
creating more reactive ice sheets, and that together with a stronger AMOC (compared to the earlier deglaciation) ~~this~~ explains
the observed higher variability in $\delta^{18}\text{O}$. Surprisingly, the rapid ~~, large (14–18 m) sea-level rise termed meltwater pulse and~~
large (14–18 m) sea-level rise termed Meltwater Pulse 1A (MWP-1A), thought to occur slightly after have occurred around
the start of the BA ~~around at approximately~~ 14.6 ka (Deschamps et al., 2012; Gregoire et al., 2016), is not manifest as a
295 salient appears only as a minor freshening anomaly in our record. Potentially, ~~the amplitude of Meltwater pulse 1A is Northern~~
Hemisphere ice-sheet melt during MWP-1A was lower than previously ~~described, has suggested, with the total sea-level rise~~
instead explained by a larger contribution from the ~~southern hemisphere and/or part of this sea level rise is reflected in the~~
meltwater pulse we detected at 15.44 ka BP. Alternatively, West Antarctic Ice Sheet (Coonin et al., 2025). Further, consistent
with our modeling (Fig. 1e), a more vigorous AMOC may have efficiently downwelled ~~the~~ surface freshwater anomalies,
300 ~~so that freshening is not detected in the surface North Atlantic moisture source area of Glassuch that even strong Northern~~
Hemisphere freshening would lead to only short-lived and more muted surface ocean anomalies in the North Atlantic.

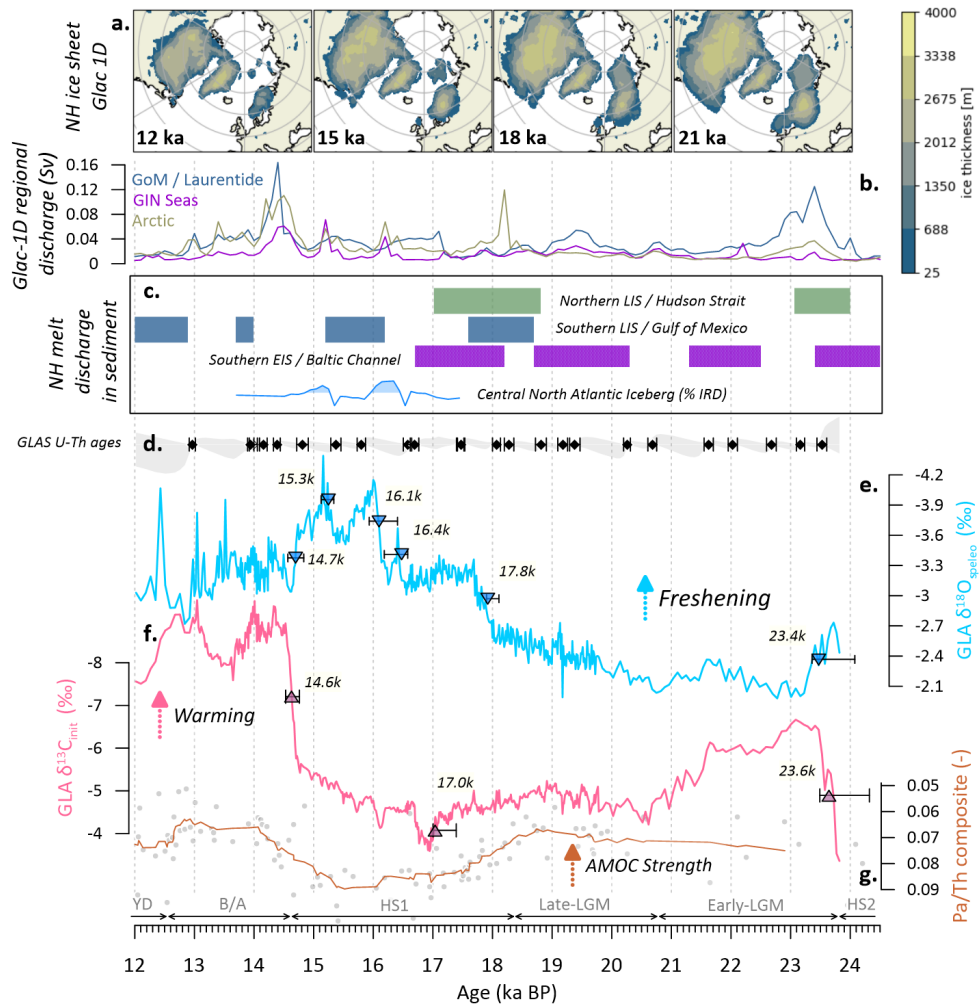


Figure 3. Speleothem record Glas and its relationship to last deglaciation ice sheet history and AMOC strength. **(a)** Snapshots of GLAC-1D ice sheet history at key intervals (Ivanovic et al., 2016). **(b)** River-routed MW delivered to the three key ocean regions identified in Fig. 1, computed in (Romé et al., 2022) from the GLAC-1D ice sheet reconstruction. **(c)** Simplified history of proposed MW discharge events from different Northern Hemisphere ice sheet sectors in sedimentary archives, broadly: the Northern Laurentide (Denton et al., 2021; Rashid et al., 2012), Southern Laurentide (Wickert et al., 2023), Channel River discharge (Toucanne et al., 2015), and distally transported iceberg rafted debris (IRD) concentration at site U1308 (Hodell et al., 2017); simplified (SI Appendix, Fig. S4). The new data of Glas (this study) is plotted in **(d)** the position of U/Th ages and (black diamonds), their associated analytical age uncertainty (black error bars) and schematic representation of resulting age model uncertainty (grey shading); **(e)** $\delta^{18}\text{O}_{\text{speleo}}$, used as a proxy for surface ocean freshening, and **(f)** $\delta^{13}\text{C}_{\text{init}}$, used as a proxy for surface air temperature above the cave. The triangles in both Glas time series mark the location of the midpoints of abrupt events discussed in this manuscript, with the nomenclature introduced in Tab. 1. **(g)** 9-point average composite Pa/Th ratio as a qualitative indicator inferred to respond to AMOC strength indicator (Ng et al., 2018). The brown fill reflects the 2- σ standard error of the composite and the grey dots the individual composite data points.

3.4 Iberian temperatures and relationship to freshening

3.4.1 AMOC insensitive to gradual freshening during the LGM

The new $\delta^{13}C_{init}$ record from Glas provides a direct speleothem chronology for a relative regional temperature evolution in the North Atlantic region (Fig. 3f). The Glas record highlights the warmer temperatures of the early LGM, followed by a significant cooling 21.5-20.5 ka BP to values sustained through the late LGM. This cooling is likely a global cooling, as it follows closely the timing of the LGM sea level lowstand (?), ~~whereas the AMOC does not show significant variations over this time based on Pa/Th ratio Fig. 3g~~ (Yokoyama et al., 2018). . During the two periods of very gradual freshening of the early and late LGM intervals, the stable temperatures indicated by $\delta^{13}C_{init}$ suggests that AMOC is too stable and/or the MW forcing is too gradual and insufficient to disrupt convection at this time.

3.4.2 Delayed AMOC response to HS1 freshening

The rather sustained temperatures over the late LGM in northern Iberia are followed by a rapid decline to coldest temperatures during the earliest part of HS1. ~~This cooling falls within the broad period of decreased AMOC strength diagnosed from Previous studies have interpreted the increased Pa/Th composite ratio ; ratio in marine sediment cores as an indicator of weakened AMOC during HS1 (Fig.3g), with but distal changes in surface productivity and boundary scavenging are now hypothesized to drive a significant fraction of this signal (Missiaen et al., 2020; Scheen et al., 2025). Additionally, the composite Pa/Th record by Ng et al. (2018) integrates results from multiple sediment cores all with greater uncertainty on age models than the Glas chronology, so the rate and chronology of circulation and associated biogeochemical cycle changes cannot be precisely inferred from the composite Pa/Th proxy record. However, a substantial weakening of AMOC over HS1 remains plausible considering also the signals detected in other proxy systems (Shakun et al., 2012; Pöppelmeier et al., 2023). Further, during the Glas record featuring a greater chronological precision and resolution. During the~~ early stages of the last deglaciation, temperature changes are typically not manifest in Greenland ice core (NGRIP) $\delta^{18}O$ (Fig. ??a) because a decrease in winter rainfall changes the overall annual $\delta^{18}O$ signature and cancels the temperature effect (He et al., 2021c). Critically, ~~Glas suggest gradual cooling temperatures starting around 19 ka BP, however,~~ compared to the ~~transient freshening starting at 18.04 ka BP described abrupt freshening 17.8k_S (Tab. 1),~~ the $\delta^{13}C_{init}$ indicates that regional surface air temperature initially remained stable, but ~~then~~ significantly cooled only ~~around 850-810⁺¹⁵³₋₃₇₅ years later, with an onset at 17.17^{+0.34}_{-0.12} kaBP: around 17.0 ka BP (event 17.0k_T, Tab. 1).~~ Because both isotope records derive from the same powders, this reflects a genuine lag ~~in response time between the two series.~~ As has been shown by model studies (Romé et al., 2022; Van Westen et al., 2024), the abrupt temperature change within the European realm and at NISA cave locations is a key characteristic of a transition into a weak AMOC state (see Fig. 1c). We, accordingly, interpret the Glas record such that, even though subtropical North Atlantic surface freshening peaked ~~at 18.04 kaBP, over the 17.8k_S event (Tab. 1) and the circulation strength might have weakened in concert with the temperature decline after 19 ka BP, the~~ most effective AMOC weakening did occur ~~around 17.17 kaBP and the coldest later, at 17.03^{+0.36}_{-0.03} ka BP (17.0k_T, Tab. 1), where the coldest deglacial temperatures at our site for the last deglaciation were reached only by 16.87^{+0.31}_{-0.16} ka BP. Further, while the composite Pa/Th record is smoothed and points towards less abrupt decrease in AMOC~~

335 ~~strength, individual cores contributing to the composite rather feature more abrupt changes as well (?) were reached. In Glas~~
~~our sampling technique prioritizes a tight chronological control by splitting power for dating and isotope proxies, but this may~~
~~underestimate the magnitude of the cooling by smoothing across the layers covering the most extreme cold period.~~

~~Although in~~ In many cases the weakening of AMOC scales directly and almost immediately with freshwater forcing (Snoll
et al., 2024), ~~however,~~ this is not necessarily true if the forcing is not strong enough, not well located, or the background state
340 of the ocean is not (yet) sensitive enough (Van Westen et al., 2024; Romé et al., 2025). We propose that ~~the~~ delayed cooling
after the ~~freshwater input with onset at 18.04 ka~~ ~~first abrupt freshwater input (17.8k_S, Tab. 1)~~ reflects the need to build up
a positive subsurface salinity anomaly or/and a required CO₂ level to cross an AMOC threshold and trigger its weakening.
The required delay is expected to depend on the meltwater forcing properties, internal ocean structure and ice sheet geometry
(Kapsch et al., 2022; Romé et al., 2025). Additionally, already before the rapid decline, $\delta^{13}\text{C}_{\text{init}}$ shows early signs of a prior
345 slight weakening, which would be consistent with a mechanism involving the build-up of subsurface anomalies (Alvarez-Solas
and Ramstein, 2011). The period of most extreme cooling was brief and recovered back to near late-LGM levels without any
evident change in North Atlantic surface freshwater conditions. ~~Alternatively or in addition, the weakening could have been~~
~~triggered directly by additional local meltwater close to the convection centres, e.g. in the Norwegian Seas.~~

The sequence of the Glas record and its comparison to radiocarbon dated IRD events (Hodell et al., 2017) both suggest that
350 cooling has preceded the large discharge of icebergs into the North Atlantic associated with Heinrich Event 1 by a bit more
than 800 years ~~-(events 16.1k_S, 15.3k_S; Tab. 1)~~. A similar sequence of events with a lag between initial surface ocean cooling
by planktic Mg/Ca and peak IRD abundance was recorded within a single core on the Iberian Margin (Skinner and Shackleton,
2006) ~~as well as in %NPS to IRD abundance in core ODP983 (Barker et al., 2015).~~

3.4.3 Temperature response to transient freshwater from 16.5 ka onward

355 Coinciding with the ~~midpoint in the~~ ~~negative~~ shift in $\delta^{18}\text{O}$ ~~, at 16.14 \pm 0.28 kaBP (HS1e in Fig 3e),~~ ~~at 16.1 ka BP (Tab. 1)~~ is
a very short-lived cooling, for which the precise duration and magnitude cannot be well constrained due to very slow growth
rates. Notably, a similarly short-lived cooling also accompanied the largest freshwater pulse of the Penultimate Deglaciation
(MWPTII-B around 134.25 kaBP, ~~(Stoll et al., 2022)~~ ~~BP, Stoll et al. (2022)~~). Such brief episodes could link to the presence
of faster, potentially atmospheric, pathways, reorganizing global circulation patterns (Markle et al., 2017; Fohlmeister et al.,
360 2023), ~~as well as the expression of a direct AMOC response due to more freshwater forcing (Snoll et al., 2024) or simply North~~
~~Atlantic sea ice expansion in response to freshening~~. In sharp contrast, no temperature response is present for the ~~freshwater~~
~~pulse at 15.44 ka (HS1d)~~ ~~subsequent freshening (15.3k_S, Tab. 1)~~, which we hypothesize indicates that this pulse was not able to
alter AMOC strength any further.

3.4.4 ~~Role of MW~~ ~~Meltwater forcing and AMOC restrengthening~~ in the Bølling-Allerød

365 The progressive decay of the MW anomaly in HS1d ~~between 15 and 14.7 ka following the negative peak after event 15.3k_S,~~
suggests that the anomaly in the eastern North Atlantic was diminishing already before the ~~abrupt onset of~~ the Bølling-Allerød
(BA) warming, which is recorded in Glas ~~at 14.77-14.47 kaBP,~~ ~~$\delta^{13}\text{C}_{\text{init}}$ at 14.62 \pm 0.14 ka BP (14.6k_T, Tab. 1),~~ a timing

indistinguishable from that in NGRIP (North Greenland Ice Core Project members, 2004))(Fig??), see Fig. ??). The abrupt BA warming coincided with a subsequent steep ~~reduction in freshwater in the North Atlantic, which may reflect increase in~~ Glas $\delta^{18}O$, which might be related to a decrease in freshwater input in the surface ocean North Atlantic but could also reflect a decrease in surface ocean accumulation due to the accelerated advection and downwelling due to AMOC strengthening as suggested by the NISA melt source contribution index (Fig. 1b-e) and as previously hypothesized in section 3.3.3. While having less freshwater in the North Atlantic may have facilitated the resumption of AMOC at the BA onset, the Glas record suggests a surface freshwater anomaly comparable to the end of the prior freshening event (HS1e) ~~17.8ks~~, where there was no evidence of AMOC restrengthening. Thus, instead of changes in NH MW, the initial trigger for BA warming might be the crossing of a AMOC sensitivity threshold related to the evolving climate and ocean boundary conditions or located within the Southern Hemisphere cryosphere.

Within BA, the $\delta^{13}C_{init}$ record indicates two warm phases separated by a cooling initiating around ~~14ka~~ 14 ka BP, potentially consistent with the onset of the Older Dryas (e.g., as seen in NGRIP (North Greenland Ice Core Project members, 2004)). ~~However, while the ice core resolves here more detail (Buizert et al., 2014). Here, the slowed growth rate in Glas complicates resolution of the precise chronology of~~ makes the Glas record potentially more smoothed and the age model more uncertain—complicating precise chronological constraints for this cooling event. ~~Nevertheless, during the early BA temperature seems to follow the,~~ and we therefore discourage detailed comparison using the Glas record during this younger interval. We note that temperature broadly tracks changes in $\delta^{18}O$ closely, which would be, consistent with AMOC strength following MW input such responding to meltwater input as in most hosing simulations (Snoll et al., 2024), ~~whereas no clear impact on temperature is present during the early BA (14.7 - 13.7 ka BP, approximately). In contrast, no clear temperature response is evident during the later BA ,suggesting a change towards lower sensitivity to MW (13.7 - 12.6 ka BP, approximately), suggesting a shift toward reduced sensitivity to meltwater~~ inputs.

3.5 Impacts Comparison with impacts beyond the North Atlantic Realm

We evaluate the relationship of North Atlantic abrupt freshening and temperature change events to other aspects of the global climate system recorded in archives possessing an absolute layer counted ice core or speleothem U-Th chronology (~~extended discussion in SI Appendix, Fig. S6). We focus on the comparisons of four prominent change points in Glas (Methods, Fig. ??).~~ Because the expected ocean circulation changes during HS1 complicate assumptions about ocean reservoir ages and thus limiting the accuracy of radiocarbon chronologies in sediment cores (Skinner et al., 2014), we do not include marine archives in the comparison presented here. The following analysis concentrates on the abrupt transitions that have been described previously in the manuscript (Tab. 1).

3.5.1 Propagation of the signal to North Atlantic-adjacent archives

In other published Iberian speleothems covering the same time, namely OST2 (Bernal-Wormull et al., 2021) and MAAT (Pérez-Mejías et al., 2019), due to the respective cave systems, the $\delta^{18}O$ signal is likely to be affected not only by the meltwater signal (as recorded in

Figure 4. Abrupt events over the last deglaciation recorded in Glas and their temporal relationship to published proxy data featuring highly resolved independent age control and a relationship to AMOC strength. Glas (a., this study) is compared to (b.) the North Atlantic is shown by (a.) the Greenland ice core on GCC05 timescale (North Greenland Ice Core Project members, 2004) and b. the alpine speleothem record NALPS (Luetscher et al., 2015), (c.) and compared to Glas (c.-d., this study). The transition midpoints, summarized in Tab. 1 are displayed throughout the entire figure as upward and downward facing triangles indicating abrupt changes in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}_{\text{init}}$, respectively. Glas is further compared to (e.) CO_2 from West Antarctica ice core (Marcott et al., 2014) and, (d.)(f.), Chinese Speleothem record YT from Hulu Cave recording changes in the East Asian Summer Monsoon (EASM) (Zhang et al., 2014). Red shadings mark three identified transient freshening events (E1-E3), blue shading the temperature events (E1,E4). Blue and pink square symbols The concurrent shifts in (a.) mark the change points in Glas South American Summer Monsoon and found drying over Puerto Rico are shown through segmented linear regression the Brazilian speleothem record PX-7 (Strikis et al., 2015) and Larga cave record PR-LA-1 (Warken et al., 2020). The errorbars denote of the combined 2σ uncertainty from transition midpoints indicate the full BChron age model ensemble and the change point detection 95% CI.

400 Glas but also additional influence from changes in temperature and moisture availability (extended discussion and records shown in Appendix).

Over the Northern hemisphere, specifically over Europe and Greenland, cooler temperatures can be expected with a weakening of AMOC. The onset of such a cooling, described as a shift in storm track and local rainfall patterns, has been detected in the NALPS record (Luetscher et al., 2015). NALPS (Fig. ??b), in many instances, follows the structure of the Greenland ice core NGRIP (Fig. ??a): the indicated freshening events E1 (E1 freshening: 18.04 – 17.65 kaBP, as has been shown by Luetscher et al. (2015) and Li et al. (2021). However, here we would like to emphasize that the onset of HS1, around 18 ka BP, is more pronounced in the NALPS record than in NGRIP and we suggest two factors contributing. Firstly, as previously noted, winter rainfall is expected to decrease with a AMOC weakening over Greenland, which is a process that is likely mitigating the impact of temperature change on the $\delta^{18}\text{O}$ recorded in the Greenland ice core (He et al., 2021b). Secondly, the sudden shift recorded in NALPS is a shift towards more negative $\delta^{18}\text{O}$, which, in the context of this work and modeling evidence presented by Zhu et al. (2017b), could also be amplified through a change in moisture source isotopic composition, caused by the addition of meltwater to the North Atlantic surface ocean. Further, in NGRIP, the start of GS-2.1a has been placed at 17.43 ± 0.33 ka BP, E1 temperature: 17.18 – 16.88 kaBP), E2 which is in agreement with the pronounced cooling in Glas at $17.03^{+0.36}_{-0.03}$ ka BP (16.22–15.99 kaBP) and E3 (15.44–15.19 kaBP) and the indicated abrupt warming event E4 (14.77–14.47 kaBP) 17.0k_T) when respecting the associated age uncertainties. In all those records the onset of the BA is pronounced and with a timing indistinguishable from the one in Glas, emphasizing the synchronized structure of this transition further.

More evidence for a concurrent re-organization of the North Atlantic ocean and atmosphere circulation patterns over HS1 stems from Brazil speleothems (Fig. ??h), which record the southward shift of ITCZ precipitation during AMOC weakenings. Brazil speleothem PX-7 from Paixão cave suggests two episodes of intensification of the South American monsoon system from 18.1 to 16.66 ka BP, and from 16.11–14.69 ka BP (Strikis et al., 2015). While broadly consistent with the Glas record, this

425 first intensification is slightly earlier than the abrupt signal in Glas. We hypothesize, that this difference could also be related to feedbacks not only with the ocean but also the northern hemisphere ice sheet thickness altering global monsoon systems (see e.g., Fig.3 in Gao et al. (2020)), resulting in a shift in the recorded timing between PX-7 and Glas. A similar structure with two phases of drier conditions, similarly related to hydroclimate re-organization, has also been found in the Puerto Rico speleothem PR-LA-1 (Fig. ??g), with a reported onset at approximately 17.5 ka BP (Warken et al., 2020).

3.5.2 Monsoon and the Westerly Jet over Asia

Isotope-enabled models suggest that millennial scale positive anomalies in the $\delta^{18}\text{O}$ in rainfall and speleothems from the East Asian summer monsoon domain (EASM) reflect reduced intensity of convection and rainfall over the Indian summer monsoon and southward-shifted westerly jet enhancing southern EASM rainfall and reducing northern EASM rainfall (He et al., 2021a). Across the 18.04 ka event (E1f), in the The EASM record from Hulu Cave (Fig.??d) features a positive shift in $\delta^{18}\text{O}$ suggests from 17.8-17.5 ka suggesting a southward shift of the westerly jet concurrent or slightly lagging the first abrupt North Atlantic freshening found in Glas (17.8ka_S). Thus, the shift in the westerly jet may have preceded the most abrupt AMOC reduction (E1f), but could instead be related to the large changes European ice sheet sectors subsequent abrupt cooling in Glas (17.0ka_T), which is a further indicator that beside the ocean feedbacks with the cryosphere, namely changes in SBKIS and the separation of SIS and BIIS (Hughes et al., 2016), might have been affecting the atmospheric circulation already prior to the extreme AMOC weakening. The following transient freshening at 16.22 ka(E2) is manifest in a concurrent and even facilitate the extreme cooling and AMOC weakening around 17.0 ka BP. Alternatively, already earlier slight weakening of AMOC, as suggested by the gradual cooling in Glas from 19 ka BP onwards, might have been sufficient in order to alter the EASM rainfall patterns.

440 Centered around 16.1 ka BP, the freshening signal and cooling recorded in Glas is concurrent with the abrupt positive $\delta^{18}\text{O}$ shift indicating a rapid southward shift of the Westerly jet in phase with the North Atlantic freshening in that interval. The recovery of the jet position is coincident with the relaxation of the freshwater anomaly. The 15.44 kaMW-pulse (E3) freshening recorded at 15.3 ka BP, in contrast, has no corresponding changepoint in the EASM record, similar to the previously discussed Iberian record OST2. This may reflect that it was another pulse of freshwater into an already weak AMOC, which triggered no further AMOC weakening or further associated shifts in atmospheric patterns cooling. Such an interpretation also strengthens our prior hypothesis based on the NISA melt source contribution index that the surface freshening anomaly originates from a source that makes it well visible in Glas, but less effective in further disrupting AMOC, such as meltwater from the Gulf of Mexico (Fig. 1be), consistent with the most likely timing of CIS-LIS breakup (Norris et al., 2022). In addition, any further effect of freshwater on the Westerly Jet may have been compensated for by the atmospheric effects of changes in LIS height (Ullman et al., 2014). At the onset of the BA event, the rapid southward-northward shift of the Westerly Jet is synchronous with the abrupt warming, confirming previous interpretations of a coherent global response to AMOC reinvigoration (Pöppelmeier et al., 2023). Overall, our findings imply that while millennial scale variations in the $\delta^{18}\text{O}$ from speleothems in the EASM do share many features with the North Atlantic realm, it is not an exact copy of its ocean dynamics and caution is needed when tuning the centennial to millennial scale features of North Atlantic records directly to an EASM signal.

455 3.5.3 Links between AMOC and atmospheric CO₂ rise

The onset of the first rapid deglacial atmospheric rise in CO₂ at ~~18.1 ka~~ 18.1 ± 0.2 ka BP (Fig. ??~~be~~) has been hypothesized to be ~~"broadly coincident"~~ "broadly coincident" with the reduction in AMOC strength (Marcott et al., 2014), but hitherto, this has been difficult to evaluate since the NGRIP ice core does not mark well the HS1 cooling. The results from Glas indicate that the first centuries of rapid CO₂ rise, like the shift in the EASM, began with, or slightly after the freshening ~~but before the onset~~ of abrupt cooling. This suggests, and the rise might be conditioned by a gradual AMOC weakening following 19 ka BP but the timing is distinct from the abrupt cooling recorded in Glas at 17.0 ka BP. This sequence of events should be further studied in more non-bioturbated high resolution North Atlantic records, but we thus hypothesize here that the first centennial carbon cycle response ~~is could rather be~~ linked to rapid atmospheric reorganization ~~which may predate the main AMOC weakening,~~ e.g., related to changes in ice sheet or also to more southern influences (Trombini et al., 2025) and those predate the abrupt North Atlantic ocean re-organization at 17.0 ka BP. Thereafter, the millennial scale CO₂ increase corresponds to the coldest period in NW Iberia, interpreted to reflect the weakest AMOC (Fig. ??~~c~~). ~~Around 16.2 ka,~~ Further, both the small precursor freshening at 16.4 ka BP and the abrupt centennial larger event at 16.1 ka BP are visible as abrupt CO₂ rise is coincident with the abrupt freshening (E2) and southward shift of the Westerly jet manifest in the EASM, suggesting an abrupt synchronous atmospheric reorganization. Our data suggest a brief cooling at the same time, which could be an expression of a very brief abrupt AMOC weakening or also North Atlantic sea ice expansion in response to freshening. After this step, there is a long plateau of concentration, but despite the lack of a AMOC response in Glas rise, with the latter confirming the global nature of the ongoing changes around 16.1 ka BP. Interestingly, there is a small CO₂ excursion at the ~~freshening event of 15.44 ka (E3).~~ It following freshening event 15.4 ka, which might hint to concurrent processes occurring in Antarctica (e.g., see Li et al. (2023)). In the aftermath of the synchronized rise with the onset of BA around 14.7 ka BP, it has been suggested that the plateau may reflect a balance between ocean CO₂ emission and uptake of CO₂ by regrowth of the terrestrial biosphere (Marcott et al., 2014), which may be consistent with continued reduction of the NH ice sheet area evidenced by persistent MW pulses. Alternatively, the CO₂ content of waters ventilating in the Southern Ocean may be stabilizing through this later part of HS1 (Yu et al., 2022).

4 Conclusions

480 The new Northern Iberian stalagmite record from Glas resolves the timing of North Atlantic surface freshening anomalies from melting NH ice sheets during the last deglaciation (24 -12 ka BP). After the LGM, a gradual freshening is followed by a series of transient freshening events with ~~onsets at 18.04 (E1), 16.22 (E2) and 15.44 kaBP (E3)~~ midpoints at 17.84^{+0.19}_{-0.02}, 16.44^{+0.09}_{-0.3}, 16.07^{+0.29}_{-0.17}, and 15.27^{+0.09}_{-0.13} ka BP. Many of these freshening pulses are present in river-routed time series of the GLAC-1D ice sheet model and meltwater-proxies from sediment cores, but the new Glas record provides a coherent, deglacial history with accurate U-Th dating. Additionally our analysis, our combined analysis with model output suggests that for ~~event E~~ the first abrupt freshening around 17.8 ka BP, stalagmite and proxy time series are most consistent with European ice sheet origin but more southerly meltwater routing for the earliest event than derived from river-routed GLAC-1D estimation. With the onset

of Bølling-Allerød, in Glas recorded around 14.7 ka BP, the subpolar North Atlantic re-salinizes according to our data. The comparison with the modeled distribution data elucidates that this rapid decline of a negative surface ocean $\delta^{18}\text{O}$ anomaly is likely linked to the concurrent AMOC re-strengthening at this point. Critically, the new speleothem record shows ~~no North Atlantic freshwater impact~~ only a very minor North Atlantic freshening coincident with the classically cited onset of MWP 1a ~~at around 14.6 ka BP, suggesting either that the most rapid pulse to the North Atlantic event happened nearly 1000 years earlier,~~ BP. We hypothesize based on our modeling that the minor signal could be caused by the very vigorous AMOC and rapid advection out of the surface, and/or that the ~~freshwater was rapidly advected out of the surface North Atlantic by strong~~ AMOC melt event occurred in multiple steps, involving also a more significant contribution from the southern hemisphere.

Because this record also resolves regional temperature changes resulting from changes in AMOC strength, the record elucidates the evolution of AMOC sensitivity to local meltwater perturbations over the critical period of the last deglaciation. ~~While some Slight gradual cooling, thus potentially early AMOC weakening, occurs after 19 ka BP. However, while some of the following~~ transient freshening events (e.g. 16.22 ka BP), namely the one at 16.1 ka BP, trigger immediate albeit brief intensification of the cooling suggestive of ~~further direct~~ rapid AMOC weakening, ~~there is a 850 years lag in the response of AMOC to the initial freshening at 18.04 ka. The freshening event at 15.44 ka~~ the initial abrupt freshening is de-coupled from the most abrupt cooling, with the latter emerging around $17.03^{+0.36}_{-0.03}$ ka BP, this is 810^{+153}_{-375} years later. In contrast, the freshening event around 15.3 ka BP does not trigger further ~~AMOC-weakening~~ cooling nor can it be seen in global archives, thus likely did not further affect the AMOC. These varied responses of AMOC to meltwater in our new record ~~reveal~~ re-enforce the view that there is not ~~a~~ ubiquitous response of AMOC to freshening in the North Atlantic ~~but it is instead background state dependent~~. Yet, the similar timing of early meltwater pulses and GLAC-1D suggests that model difficulties simulating AMOC weakenings with GLAC-1D freshwater forcing may reflect challenges in simulating evolving AMOC sensitivity to freshening rather than erroneous timing of the meltwater forcing. Taken together, these results suggest non-linear dynamics within the coupled ice-ocean-climate system which invite ~~for~~ advanced transient AMOC sensitivity studies and need to be further evaluated with precise model-data comparisons.

Code and data availability. Details of the new U-Th data presented in this study is reported in SI Appendix, Table S1. All meta, geochemical and age data for speleothem Glas is further available in SISALv3 submission format (Kaushal et al., 2024) in the ETH Research collection (<https://doi.org/10.3929/ethz-b-000726747>), and will be added to the SISAL database with their next release. The underlying model simulation data is available from the NERC Centre for Environmental Data Analysis repository (Rome et al., 2022). The analysis script to compute the transition midpoints is available at https://github.com/lrnrd/midpoints_EnsembleComputation.git (last access: 3.3.2026)

Appendix A: Supplementary Materials

A1 The Glas sample and its connection to the NISA speleothem archive

Stalagmite Glas originates from the cave ²⁴La Vallina' in Northwest Iberia, which is one of the key caves in the Northwest Iberian Speleothem Archive (NISA) (Stoll et al., 2022). Glas is a very compact and translucent sample (Fig. A1). Glas has
520 been selected for this detailed study because of the absence of any signs for non-continuous growth and for featuring a homogenous trace element structure, also when diverging from the main growth axis. Other NISA speleothems were revisited in the scope of this study, namely Candela and Laura (Fig. A4). Laura confirms the presence of a double-peak freshening structure before Bølling-Allerød warming occurs (data from Stoll et al. (2022)). A set of new dates on Candela using the procedure outlined in this study, confirms the presence of a shift around 18.04 ka BP, but has also indicated condensed/ceased
525 growth from 17.8 ka to 15.6 ka BP (Tab. A1).

A2 Simulated Temperature and Precipitation response to a change in simulated AMOC state

A number of studies have previously simulated the effect of changes in AMOC on global climate (e.g., Jackson et al. (2015); Van Westen et al. (2024)) and the most dominant features such as cooling of the Northern hemisphere and large-scale precipitation changes, such as a weakening of Asian summer monsoon, are robust across models and form the theoretical background
530 to interpret global archives in the context of AMOC strength. In this study, a quasi-idealised simulation is re-employed from Romé et al. (2022) featuring a glacial climate state with an oscillating AMOC strength (period ca. 1500 years, see methods main manuscript), which is triggered by a constant meltwater (MW) flux of 0.084 Sv corresponding to the GLAC-1D ice sheet history at 17.8 ka BP. From the two strong states defined in Romé et al. (2022), here we selected the more stable *zonal* mode as the strong mode. When comparing periods of strong and a weak state, the simulation does show the expected global climatic
535 patterns for a AMOC slowdown, with a temperature reduction of up to 10 K the North Atlantic realm in winter, a southward shift of the ITCZ and significant changes in winter rainfall patterns (Fig. A5).

A3 Meltwater time series from sediment cores

In main text Fig. 3, previously published MW time series are presented as simplified blocks for a better visual overview. For completeness, the underlying time series are presented here in Fig. A6. Note that the event classification in both, Toucanne et al.
540 (2015) and Wickert et al. (2023) (~~dataset: Wickert (2023)~~), are taken from the original publication, respective the associated dataset (Wickert, 2023). For the Baffin Slope record (Rashid et al., 2012), melt events were set in this study based on the prominent negative $\delta^{18}O$ shifts (shading in Fig. A6).

A4 Extended comparison with ice sheet reconstruction GLAC-1D and the derived discharge time series

The GLAC-1D ice sheet model (Ivanovic et al., 2016) was used to derive river-routed regional MW time series for three
545 key regions (Fig.A7a), which were introduced previously (Romé et al. (2022)), see the supplementary information for the

methodology). The three regions are 1.) the GIN Seas (pink shading in Fig. A7a), 2.) Arctic Ocean (green shading), and 3.) Labrador Sea+Gulf of Mexico (GoM) (purple shading). Comparison between the discharge time series derived from GLAC-1D, sediment cores and from the Glas record show significant differences in timing and location of expected discharge, as featured in Fig. 3 in the main text. These differences likely expose the still existing uncertainties in reconstructing the MW discharge.

For GLAC-1D, the estimated source ice sheet sector for MW pulses ending up in the three regions used in this study (Fig. A7a), is exposed by plotting ice sheet mass loss maps over short time snippets (Fig. A7b-c). However, although this is informative, the timings have to be taken with caution since GLAC-1D is a model, not an ice sheet reconstruction. For example, the calculation of river-routed MW crucially depends on the ice sheet height and global distribution expected at LGM, which also remains challenging to constrain from data.

Yet, by considering the difference maps in Fig. A7c, for end of Heinrich-Stadial 2 (HS2), strong Southern LIS melt has driven the peak visible in the Laurentide+GoM region and some of Northern LIS/CIS melt also discharges into the Arctic region. Further, some MW that could be visible in Glas maybe also originates from the Southern edge of BIIS. This aligns with Southern BIIS discharge also being registered by Toucanne et al. (2015), as outlined in the main text Fig.3. In Early-LGM, MW discharges into GIN seas, related to thinning of the ice sheet border between BIIS and SIS, according to GLAC-1D. This slight freshening, close to the Glas site, is consistent and visible as a small excursion in our record. The gradual increase in Late-LGM is reflected in stalagmite Glas, sediment cores and GLAC-1D. For GLAC-1D, most of the discharge seems to originate from southern BIIS, SIS and also Southern part of Laurentide ice sheet. Over this period, proximal MW into Bay of Biscay seems plausible but also some of the signal could originate from the Southern Laurentide melt. Conversely, the MW sediment cores do not show evidence of MW over this time into the GoM (Wickert et al., 2023). From 18.5-18 ka, GLAC-1D models a rapid melt of Svalbard-Barents-Kara ice sheet (SBKIS), which is dumped into Arctic, which we believed might have at least partially discharged into GIN Seas, or routed to the channel river. In addition, the collapse of Hudson strait, often seen to be connected to enhanced ice berg discharge (HE1), seems to occur from 17.4 k-16.9 ka, reflected in the peak in the purple region Labrador+GoM. Nothing changes during this time in Glas, and sediment cores, so it might be plausible that this peak could be shifted, i.e. either collapse was later, or melt discharge delayed. The interval 16.4-16 ka shows here mostly the collapse of SBKIS, and is also the only time with ice sheet mass loss in Greenland. This exposes a certain contrast to the general idea to have HE1 then, supposedly sourced from LIS. GLAC-1D 15.4-15 ka shows a peak in GIN seas, originating from SIS and potentially related to Glas 15.4 ka BP. With the beginning of BA, GLAC-1D shows a large retreat of all NH ice sheets, to accomodate for sea level changes, and associated with MWP1a. Interestingly, neither Glas nor any of the discharge proxy time series in this study show a big MW signal there.

A5 ~~Extended comparison with~~ Complementary discussion of other global independently dated archives Iberian speleothems

~~An extended compilation of global records can be found in Fig. ??.~~ A noteworthy record from another regional cave from Northern Iberian Peninsula is OST2 from Ostolo cave (Bernal-Wormull et al., 2021) ~~. This~~ and MAAT from Meravelles cave

580 [\(Pérez-Mejías et al., 2021\) \(Fig. A8\)](#). Ostolo cave features higher elevation and has nearly double the rainfall compared to La Vallina, indicative of a strong orographic enhancement. Therefore, beside OST2 also showing a shift towards more negative $\delta^{18}O$ likely associated to a changed isotopic signature of the surface ocean moisture source around 18 and 16.1 ka, also the cold air temperature is described to cause negative shifts in $\delta^{18}O$ of the stalagmite, an effect superimposed on the changes in the $\delta^{18}O_{sw}$ seen in the coastal caves. [Consequently, while some meltwater signals may be manifest, such as the short-lived excursion to negative \$\delta^{18}O\$ values in OST2, corresponding to to the Glas event 16.1k_S \(Bernal-Wormull et al., 2021\), temperature and hydrological processes amplify the signal in OST2. The onset of BA in OST2, reported as 14.9 ka BP \(reported age uncertainty \$\pm 80\$ years\) predates both the estimates of the warming in Glas \(14.6k_T, with the midpoint at 14.62 \$^{+0.14}_{-0.09}\$ ka BP\) and the onset in Greenland ice cores \(start at 14.64 \$\pm\$ 0.19 ka BP; Rasmussen et al. \(2014\), Fig. ??a\).](#) In contrast, in mediterranean caves, such as Meravelles cave (MAAT_{7SA}), the onset of HS1 is synchronous with a shift to more positive $\delta^{18}O$, which

585

590 in this setting has been interpreted to indicate a decrease in rainfall and resulting deficit in moisture balance consistent with regional downscaled models of precipitation for the region during weakened AMOC. Additionally, stalagmite MAAT shows also an abrupt decrease in temperature, visible in $\delta^{13}C$ (Pérez-Mejías et al., 2021).

Appendix B: Methods

B1 Alignment of the old and new trench

595 Due to the slow growth rate and the presence of abrupt climate changes over the early last deglaciation the sampling strategy for U/Th dating has been optimised by continuously micro-milling the sample with a 50 μ m trench along growth layers, measure stable isotopes on these individual samples before combining the remaining powder of about 10 samples to one date. Trace elements and stable isotopes were also measured on a parallel trench and results from the two trenches were aligned using QAnalySeries (Kotov and Pálike, 2018). The two trenches are plotted on top of each other in [Fig. A9](#). In all other figures

600 oxygen isotope results are plotted from the same trench as U/Th dates, and $\delta^{13}C_{init}$ are plotted using $\delta^{13}C$ from the U/Th trench corrected with trace elements from the parallel trench.

B2 Details about $\delta^{13}C_{init}$ correction

Through the correction of the $\delta^{13}C$ ratio for in-cave process of Prior calcite precipitation (PCP), $\delta^{13}C_{init}$ becomes an improved qualitative indicator for temperature above the cave. We follow the procedure of Stoll et al. (2023): the initial Mg/Ca is

605 assumed to be the minimum Mg/Ca ratio over the entire dataset and is used to compute the remaining calcite fraction (fCa) over time. Specifically, the equation to calculate the remaining calcite fraction is $fCa(t) = \frac{\min(Mg/Ca)}{(Mg/Ca(t))}$. Based on previous sensitivity experiments for a glacial Northwest Iberian climate and the La Vallina cave setting (Lechleitner et al., 2021; Stoll et al., 2023), the degassing slope fCa_{slope} is set to -8 ‰, the bedrock scaling factor B to 1, and the attenuation factor AF also to 1 for the entire last deglaciation. The resulting equation to calculate the PCP-corrected $\delta^{13}C_{init}$ is $\delta^{13}C_{init}(t) =$

610 $d13C(t) - fCa_{slope} * \log(fCa(t))$.

B3 Details about age model and changepoint analysis

The age model was constructed from 25 previously unpublished U-Th ages (SI Tab. A1) using the Bayesian chronology model BChron (Haslett and Parnell, 2008). In a first iteration, an outlier probability of 0.1 was assumed for each date, and the actual thickness of the combined sample layers was given as a prior to the model. In a second iteration and based on the initial results, the outlier probabilities on the two dates GLA-Z 14.85 and GLA-X 17.55 were increased to 0.4 and 0.8, respectively. From the ensemble of plausible age models within the 95% confidence interval, we have finally selected the age model that features the highest correlation between computed growth rate and Sr/Mg ratio as the main age model. This was because in this cave setting, the Sr/Mg ratio has been shown to correlate with changes in growth rate (Sliwinski and Stoll, 2021). For the breakpoints, the changepoints and their age estimates respecting the full 95% uncertainty derived from the piecewise linear model are summarized in Table ?? and ?? The comparison between the measured $\delta^{13}\text{C}$ and the corrected $\delta^{13}\text{C}_{\text{init}}$ time series is shown in Fig. A9.

B3 Details about the model simulation

The underlying simulation is the experiment 'tfgbj', previously published and made publicly available in Rome et al. (2022). The simulation was completed using the HadCM3 atmosphere-ocean general circulation model in the BRIDGE (Bristol Research Initiative for the Dynamic Global Environment group) version (Valdes et al., 2017). The model consists of 19 atmosphere layers ($2.5^\circ \times 3.75^\circ$) (Pope et al., 2000), coupled to an ocean model with 20 layers ($1.25^\circ \times 1.25^\circ$) (Gordon et al., 2001), and includes the Moses 2.1 land model and TRIFFID dynamic vegetation model (Cox, 2001). The passive advective tracers were added and run for 500 years in two selected periods, in which the AMOC was strong or weak, respectively.

| Sample Number | 238U (ppb) | 232Th (ppt) | 230Th / 232Th (atomic x 10 ⁻⁶) | 4234U* (measured) | 230Th / 238U (activity) | 230Th Age (yr) (uncorrected) | 230Th Age (yr) (corrected) | 4234U/Initial** (corrected) | 230Th Age (yr BP)** (corrected) | Depth (corrected mm) |
|---------------|------------|-------------|--|-------------------|-------------------------|------------------------------|----------------------------|-----------------------------|---------------------------------|----------------------|
| GLA-I 2.05 | 782.3 ±0.9 | 720 ±18 | 1968 ±50 | 137.3 ±1.3 | 0.1098 ±0.0004 | 11050 ±44 | 11027 ±47 | 142 ±1 | 10956 ±47 | 3.300 |
| GLA-G 3.95 | 837.0 ±1.4 | 77 ±10 | 23137 ±3088 | 137.4 ±1.4 | 0.1285 ±0.0004 | 13038 ±51 | 13036 ±52 | 143 ±1 | 12965 ±52 | 4.525 |
| GLA-F-5.1 | 598.7 ±0.8 | 1696 ±37 | 807 ±18 | 138.3 ±1.3 | 0.1387 ±0.0004 | 14131 ±46 | 14059 ±68 | 144 ±1 | 13988 ±68 | 6.025 |
| GLA-E 6.95 | 796.5 ±0.9 | 126 ±14 | 14401 ±1661 | 139.3 ±1.3 | 0.1378 ±0.0005 | 14020 ±53 | 14016 ±53 | 145 ±1 | 13945 ±53 | 7.700 |
| GLA-D-8.45 | 540.9 ±0.6 | 68 ±14 | 18337 ±3652 | 140.5 ±1.3 | 0.1399 ±0.0006 | 14230 ±64 | 14227 ±64 | 146 ±1 | 14156 ±64 | 9.050 |
| GLA-C 9.65 | 543.4 ±0.5 | 115 ±12 | 11043 ±1140 | 142.3 ±1.3 | 0.1422 ±0.0005 | 14460 ±57 | 14454 ±57 | 148 ±1 | 14383 ±57 | 10.350 |
| GLA-I 11.05 | 375.4 ±0.5 | 154 ±14 | 5899 ±549 | 146.3 ±1.8 | 0.1467 ±0.0009 | 14894 ±97 | 14884 ±97 | 153 ±2 | 14813 ±97 | 11.600 |
| GLA-B 12.15 | 526.6 ±0.7 | 78 ±19 | 16913 ±4146 | 148.9 ±1.7 | 0.1522 ±0.0008 | 15447 ±85 | 15444 ±85 | 156 ±2 | 15373 ±85 | 12.750 |
| GLA-A-13.35 | 386.5 ±0.4 | 144 ±11 | 6905 ±520 | 151.5 ±1.3 | 0.1565 ±0.0006 | 15880 ±69 | 15871 ±70 | 158 ±1 | 15800 ±70 | 13.950 |
| GLA-Z 14.85 | 490.4 ±0.4 | 118 ±13 | 11210 ±1239 | 154.0 ±1.3 | 0.1639 ±0.0006 | 16651 ±70 | 16645 ±70 | 161 ±1 | 16574 ±70 | 15.600 |
| GLA-Y 16.35 | 543.9 ±0.6 | 328 ±15 | 4517 ±201 | 154.2 ±1.4 | 0.1651 ±0.0006 | 16778 ±71 | 16763 ±72 | 162 ±1 | 16692 ±72 | 16.950 |
| GLA-X 17.55 | 667.7 ±0.7 | 88 ±13 | 21551 ±3173 | 155.1 ±1.3 | 0.1721 ±0.0005 | 17532 ±62 | 17529 ±62 | 163 ±1 | 17458 ±62 | 18.225 |
| GLA-W 18.9 | 735.8 ±0.8 | 62 ±11 | 33805 ±5878 | 155.0 ±1.3 | 0.1722 ±0.0005 | 17547 ±62 | 17545 ±62 | 163 ±1 | 17474 ±62 | 19.600 |
| GLA-V 20.3 | 860.4 ±1.0 | 89 ±17 | 28385 ±5490 | 154.8 ±1.3 | 0.1776 ±0.0006 | 18141 ±67 | 18138 ±67 | 163 ±1 | 18067 ±67 | 20.900 |
| GLA-U 21.5 | 490.0 ±0.5 | 27 ±12 | 53786 ±23214 | 154.0 ±1.3 | 0.1793 ±0.0007 | 18351 ±80 | 18349 ±80 | 162 ±1 | 18278 ±80 | 22.125 |
| GLA-T-22.75 | 613.0 ±0.8 | 74 ±13 | 25043 ±4499 | 151.4 ±1.6 | 0.1837 ±0.0008 | 18886 ±91 | 18883 ±91 | 160 ±2 | 18812 ±91 | 23.450 |
| GLA-S-24.15 | 514.0 ±0.6 | 112 ±10 | 14153 ±1239 | 151.5 ±1.3 | 0.1870 ±0.0007 | 19253 ±82 | 19248 ±82 | 160 ±1 | 19177 ±82 | 24.875 |
| GLA-R 25.6 | 445.1 ±0.5 | 46 ±10 | 30053 ±6324 | 153.2 ±1.5 | 0.1891 ±0.0007 | 19455 ±87 | 19452 ±87 | 162 ±2 | 19381 ±87 | 26.225 |
| GLA-P 28.4 | 451.1 ±0.4 | 21 ±7 | 70030 ±23585 | 149.6 ±1.3 | 0.1963 ±0.0005 | 20331 ±57 | 20330 ±57 | 158 ±1 | 20256 ±57 | 29.000 |
| GLA-O 29.6 | 454.5 ±0.5 | 12 ±5 | 125285 ±47103 | 151.4 ±1.3 | 0.2003 ±0.0006 | 20755 ±74 | 20754 ±74 | 161 ±1 | 20680 ±74 | 30.500 |
| GLA-N 32.2 | 408.2 ±0.6 | 22 ±4 | 62430 ±11976 | 150.8 ±1.5 | 0.2085 ±0.0006 | 21701 ±77 | 21700 ±77 | 160 ±2 | 21626 ±77 | 33.100 |
| GLA-M 34.0 | 470.1 ±0.6 | 24 ±5 | 68338 ±13962 | 149.3 ±1.3 | 0.2117 ±0.0006 | 22106 ±77 | 22104 ±77 | 159 ±1 | 22030 ±77 | 34.700 |
| GLA-LA 35.44 | 426.8 ±0.5 | 64 ±5 | 23970 ±2014 | 148.6 ±1.2 | 0.2172 ±0.0007 | 22755 ±81 | 22752 ±81 | 158 ±1 | 22678 ±81 | 36.000 |
| GLA-AB 36.6 | 410.3 ±0.5 | 24 ±6 | 62653 ±16661 | 147.4 ±1.3 | 0.2212 ±0.0005 | 23243 ±69 | 23241 ±69 | 157 ±1 | 23167 ±69 | 37.200 |
| GLA-AC 37.8 | 451.7 ±0.5 | 60 ±5 | 27912 ±2419 | 148.9 ±1.3 | 0.2245 ±0.0007 | 23601 ±82 | 23598 ±82 | 159 ±1 | 23524 ±82 | 38.300 |
| CAN-C+2.06 | 197.3 ±0.5 | 456 ±15 | 1343 ±45 | 220.9 ±2.8 | 0.1884 ±0.0013 | 18024 ±144 | 17969 ±149 | 243 ±3 | 17895 ±149 | 8.3 |
| CAN-G 10.18 | 215.6 ±0.2 | 431 ±10 | 1563 ±38 | 225.9 ±1.3 | 0.1894 ±0.0006 | 18210 ±63 | 18163 ±71 | 238 ±1 | 18089 ±71 | 3.9 |
| CAN-M+18.58 | 187.8 ±0.4 | 221 ±6 | 2678 ±75 | 216.9 ±2.2 | 0.1908 ±0.0008 | 18499 ±95 | 18471 ±97 | 228 ±2 | 18397 ±97 | 5.3 |

Table A1. (Table previous page.) **New Uranium and thorium isotopic compositions and 230Th ages for Glas (GLA) and Candela (CAN) samples analyzed by ICP-MS.** U decay constants: $\lambda_{238} = 1.55125 \times 10^{-10}$ (Jaffey et al., 1971) and $\lambda_{234} = 2.82206 \times 10^{-6}$ (Cheng et al., 2013). Th decay constant: $\lambda_{230} = 9.1705 \times 10^{-6}$ (Cheng et al., 2013). $*d_{234}U = ([^{234}U/^{238}U]_{activity} - 1) \times 1000$. $**d_{234}U_{initial}$ was calculated based on ^{230}Th age (T), i.e., $d_{234}U_{initial} = d_{234}U_{measured} \times e^{\lambda_{234} \times T}$. Corrected ^{230}Th ages assume the initial $^{230}Th/^{232}Th$ atomic ratio of $4.4 \pm 2.2 \times 10^{-6}$. Those are the values for a material at secular equilibrium, with the bulk earth $^{232}Th/^{238}U$ value of 3.8. The errors are arbitrarily assumed to be 50%. $***B.P.$ stands for “Before Present” where the “Present” is defined as the year 1950 A.D. The depth noted in the last column is the mid-depth of all isotope sample trenches combined for the U-Th dating.

| | Age (yr BP) | Age 2σ negative | Age 2σ positive | $\delta^{18}O$ | 11382 | 275 | 799 | -3.0156 | 14422 | 68 | 189 | -3.3065 | 14600 | 131 | 162 | -3.1283 | |
|-----|-------------|------------------------|------------------------|----------------|------------------------|------------------------|-----------------------|---------|---------|---------|---------|--|---------|---------|---------|--|--|
| 630 | 14788 | 83 | 145 | -3.6251 | 15185 | 254 | 110 | -3.9327 | 15436 | 85 | 193 | -3.5451 | 15993 | 136 | 335 | -4.0323 | |
| | 94 | 288 | -3.2054 | 18036 | 29 | 164 | -2.6244 | 20559 | 203 | 169 | -2.1724 | 20740 | 82 | 474 | -2.0584 | 21701 | |
| | 81 | 248 | -2.3183 | 22293 | 330 | 400 | -2.1495 | 23038 | 206 | 102 | -2.0073 | Estimated Breakpoints for . Ages are given in years before present (yr BP). The uncertainty of the break point estimate integrates the uncertainty from estimating the breakpoint and from the full BChron age model ensemble. | | | | | |
| 635 | | | | Age (yr BP) | Age 2σ negative | Age 2σ positive | $\delta^{13}C_{init}$ | 12185 | 725 | 463 | -7.5998 | 12604 | 835 | 222 | -8.9948 | 13061 | |
| | | | | 106 | 302 | -8.8653 | 13387 | 310 | 288 | -7.9635 | 14469 | 62 | 195 | -9.1089 | 14767 | 104 | |
| | | | | 125 | -5.5894 | 16143 | 245 | 284 | -4.4712 | 16623 | 194 | 86 | -4.6218 | 16876 | 116 | 311 | |
| | | | | -3.7553 | 17177 | 120 | 343 | -4.4369 | 18967 | 144 | 144 | -5.0006 | 20609 | 141 | 138 | -4.4498 | |
| | | | | 21788 | 108 | 193 | -5.9801 | 22592 | 282 | 89 | -5.9012 | 23038 | 198 | 94 | -6.8444 | Estimated Breakpoints for . Ages are given in years before present (yr BP). The 2σ uncertainty of the break point estimate integrates the uncertainty from estimating the breakpoint and from the full BChron age model ensemble. | |
| 640 | | | | | | | | | | | | | | | | | |



Figure A1. Scan of Stalagmite Glas before cutting a slab for the new trench.

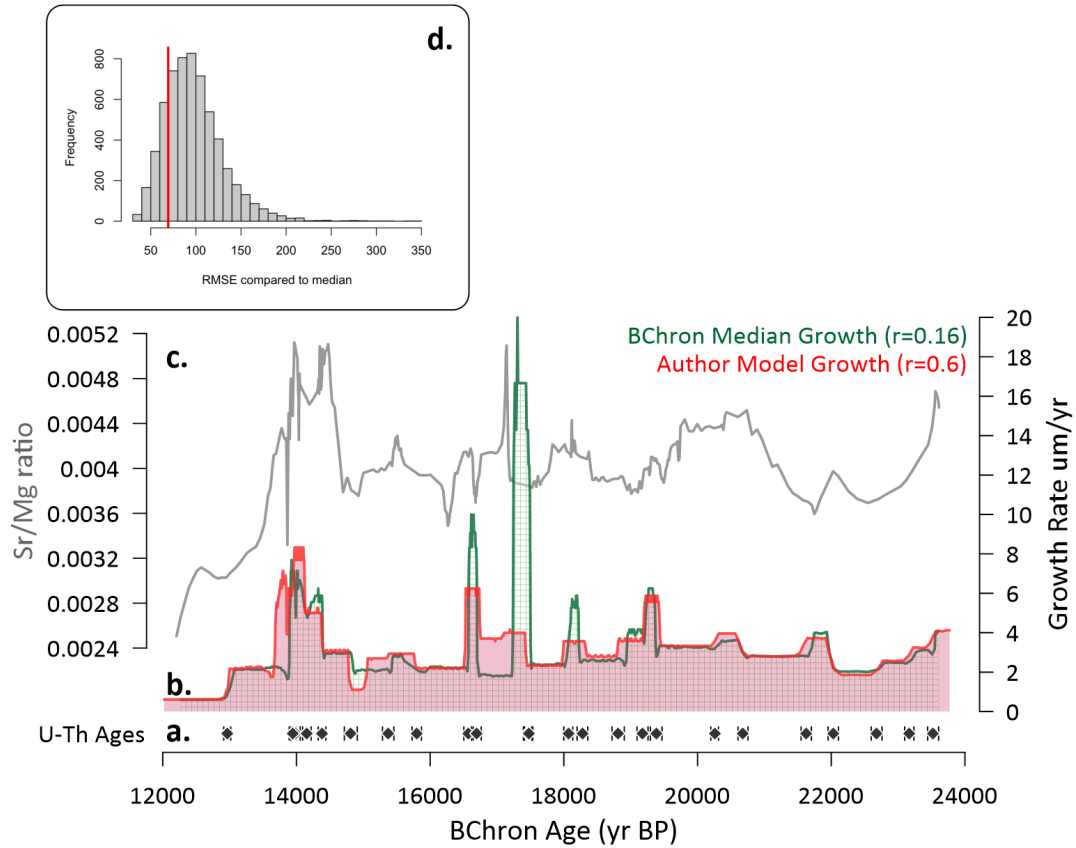


Figure A2. Comparison of growth rates from the pointwise median BChron age model to the posterior realization selected as author age model. The age model has been calculated using 26 U-Th ages (a.), and the respective growth rates have been computed using a centered differences. The goal was to improve the correlation with the Sr/Mg ratio measured in Glas (c.). Correlation coefficients were computed for the 6000 posteriors from the BChron age models and a realisation showing high correlation with Sr/Mg ratio, as well as reasonable growth rates and a low RMSE, thus high similarity, compared to the median BChron (d.).

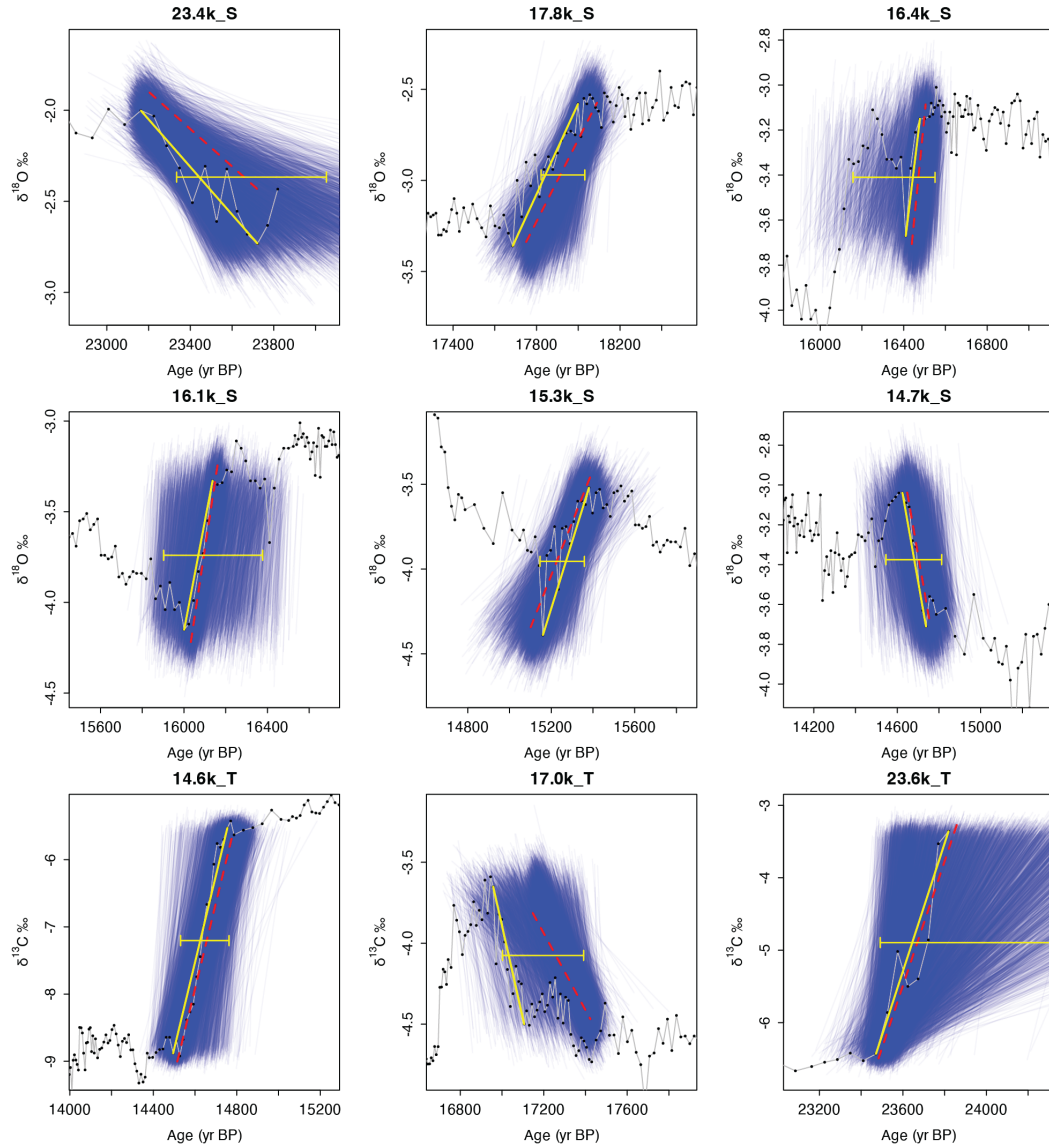


Figure A3. Visualization of the transition mid point derivation from computation with the full BChron posterior age ensemble, data is summarized in Tab. 1. The subtitles for the individual panels are according to the nomenclature in Tab. 1. In each panel, blue lines connect the two age estimates at the relevant sample depths, while the yellow line is the estimate from the author age model and the red dashed line is the BChron median estimate. The horizontal errorbar denotes the 95 % Credible Interval (CI) of the full ensemble, representing the uncertainty reported throughout the manuscript. The isotope data plotted on the author age model is shown as a grey line.

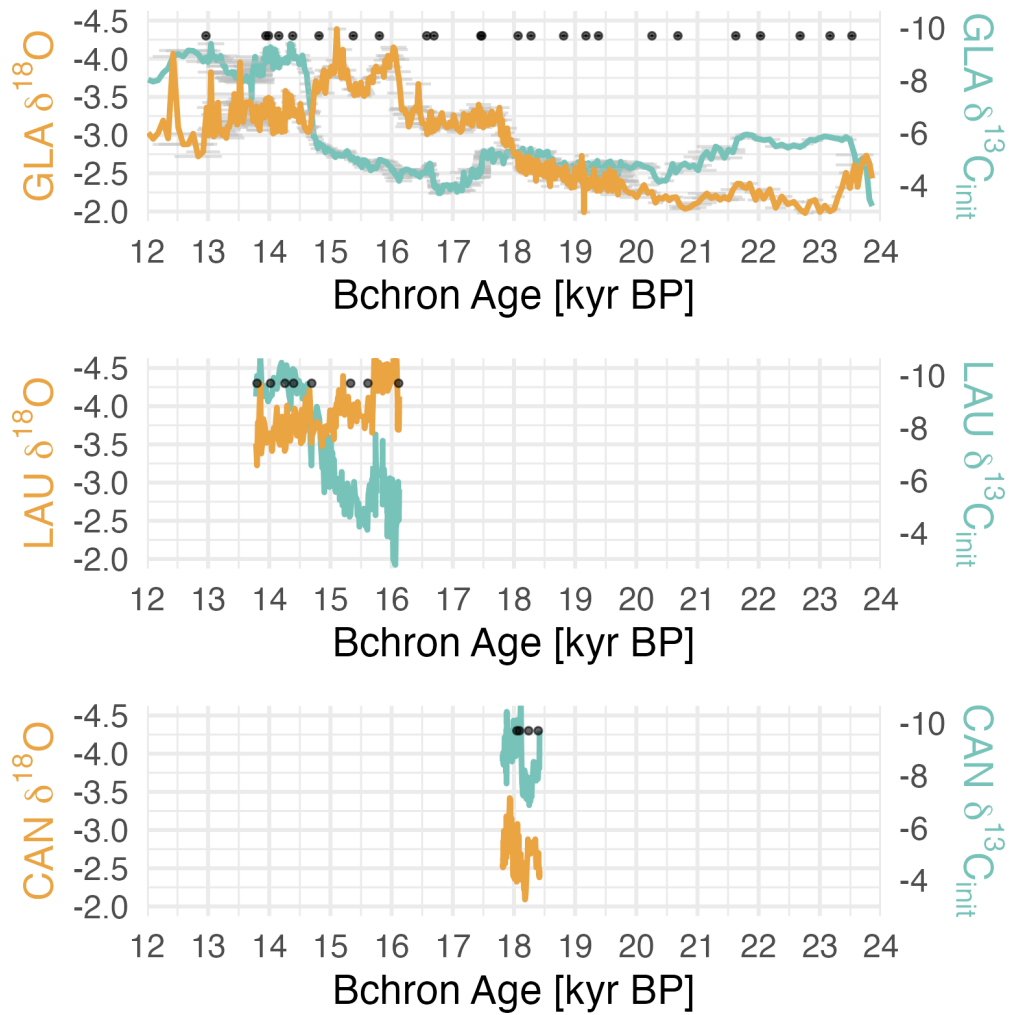


Figure A4. Comparison with NISA stalagmites covering segments of the same period: Laura (LAU) and Candela (CAN). Ages of Laura have been previously published in Stoll et al. (2022), ages of Candela are ages on a new trench and are ~~published~~ [provided](#) here in Tab. A1.

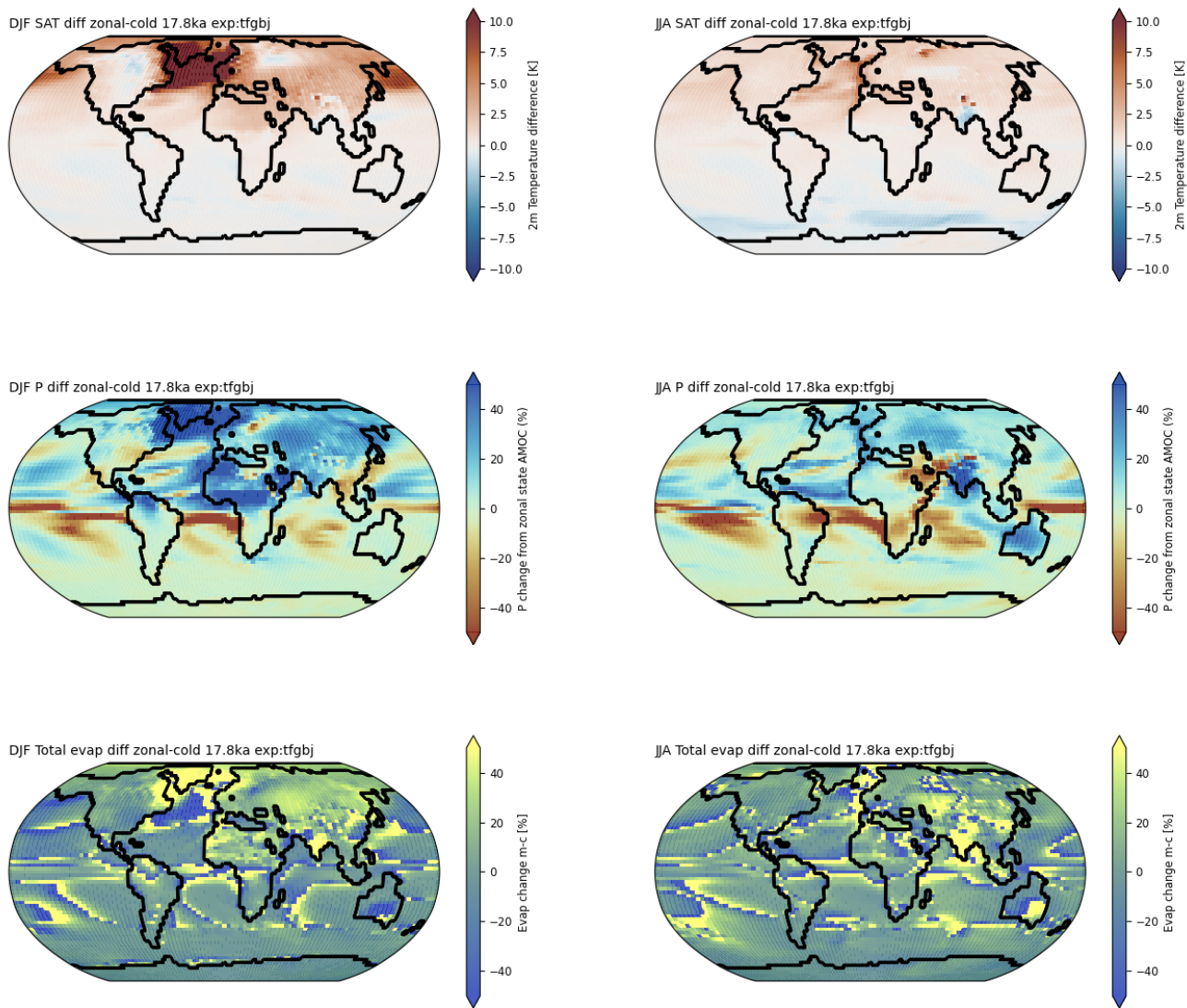


Figure A5. Global maps of expected environmental changes due to AMOC weakening. Results from the simulation described in Romé et al. (2022), snapshot 17.8 ka. The toprow shows differences (strong-weak AMOC) in surface air temperature (SAT) on the left for winter (DJF) and on the right for summer (JJA). The middle row shows the same analysis for differences in Precipitation (P), and the bottom for total evaporation.

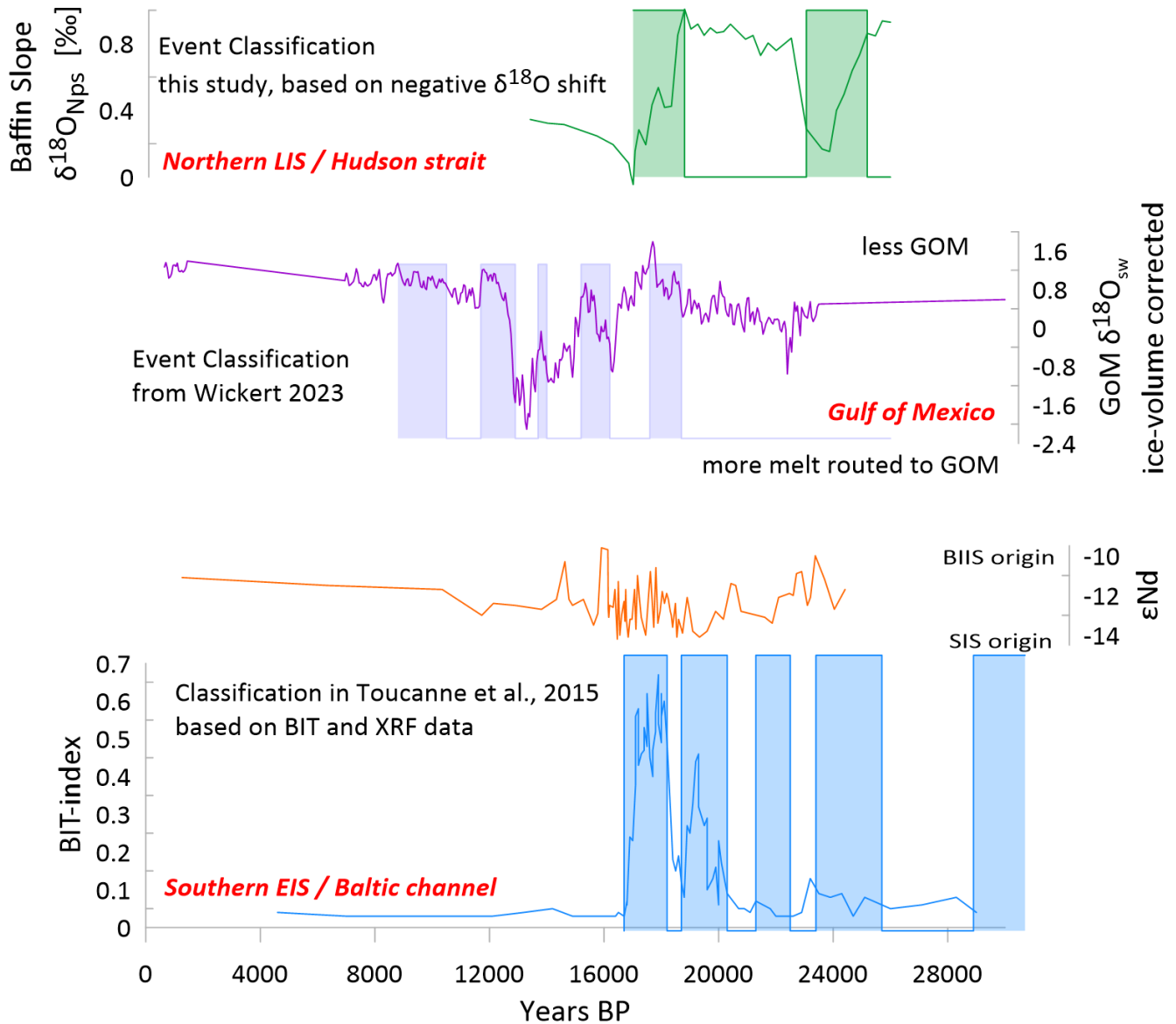


Figure A6. Previously published meltwater discharge time series in sediment cores: Baffin slope $\delta^{18}O$ (Rashid et al., 2012)(green), Gulf of Mexico (Wickert et al., 2023)(purple), and from the Bay of Biscay (Toucanne et al., 2015)(orange,blue). Shaded areas are the bulk events featured in the main manuscript Fig. 2-3c. Classification of the events is taken from the original publication for the Gulf of Mexico and the Bay of Biscay. For the Baffin slope record we have defined the events based on the large shifts in $\delta^{18}O$ in the here plotted time series. All data is plotted on the age model of the original publication which may feature several ky of uncertainty.

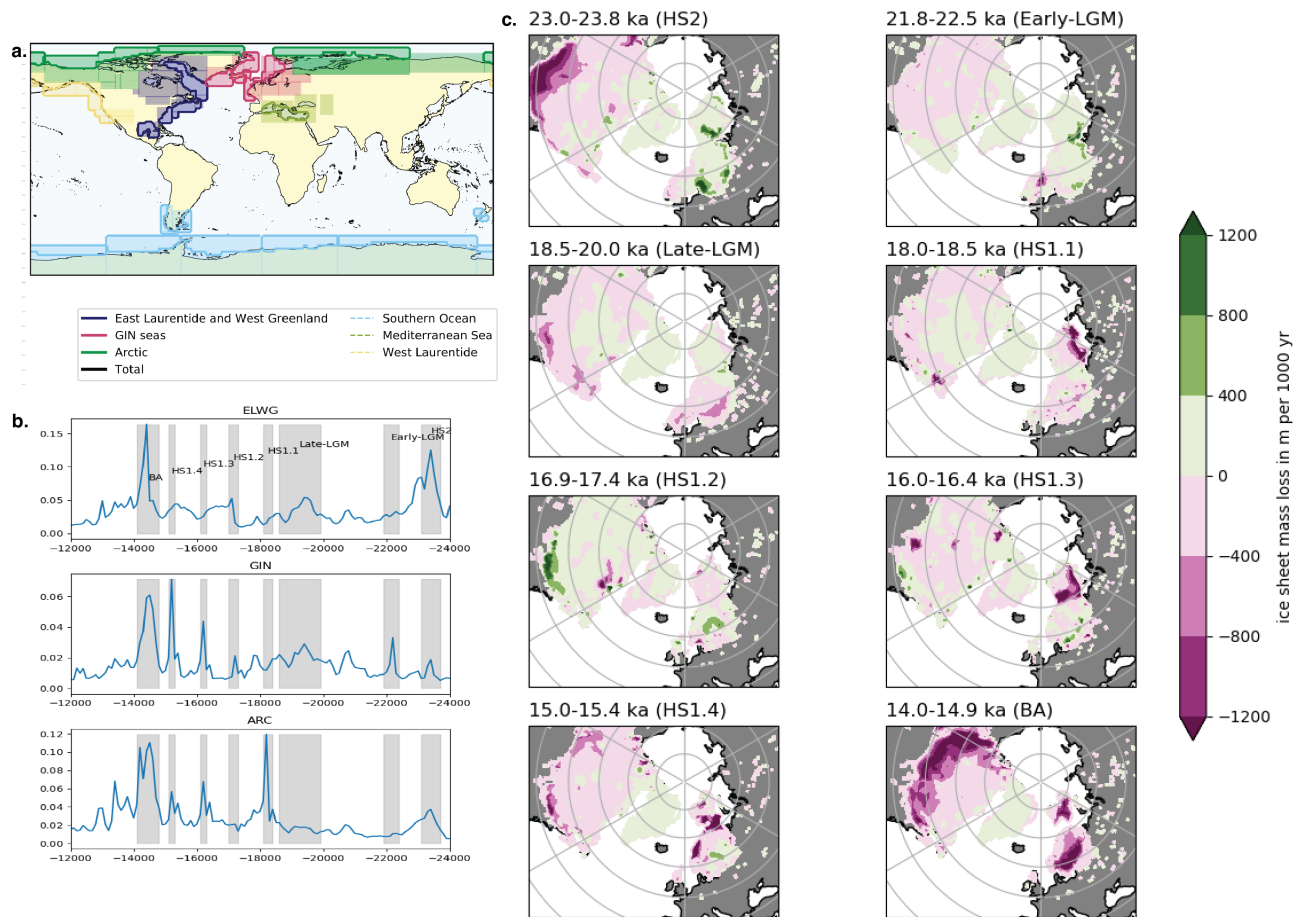


Figure A7. GLAC-ID Ice sheet mass loss over selected time snapshots. a) shows a global map including the three regions used in this study East Laurentide and West Greenland (ELWG), Arctic (ARC) and GIN seas, replotted from Romé et al. (2022). In this study, we renamed the ELWG after the drainage ocean bassins and called it Labrador+GoM. b) Detail line plot of the three river-routed time series ELWG, ARC and GIN seas. The grey shadings indicate the time snippets defining name, and start/stop year for the maps in c). c) ice sheet mass loss maps over selected periods with significant peaks in the derived meltwater time series. The ice sheet mass loss has been converted into m per 1000 yr for easier comparison between the panels.

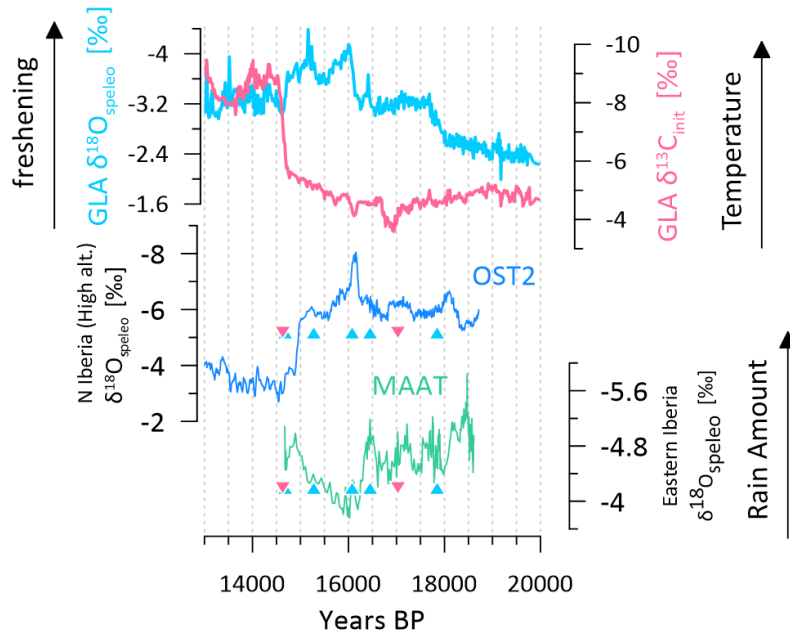


Figure A8. Extended comparison with global Iberian records. All speleothem data The record of Glas is extracted from SISALv3 (Kaushal et al., 2024) compared to OST2 (Bernal-Wormull et al., 2021) and the Antarctic ice core data from Marcott et al. (2014) MAAT (Pérez-Mejías et al., 2021).

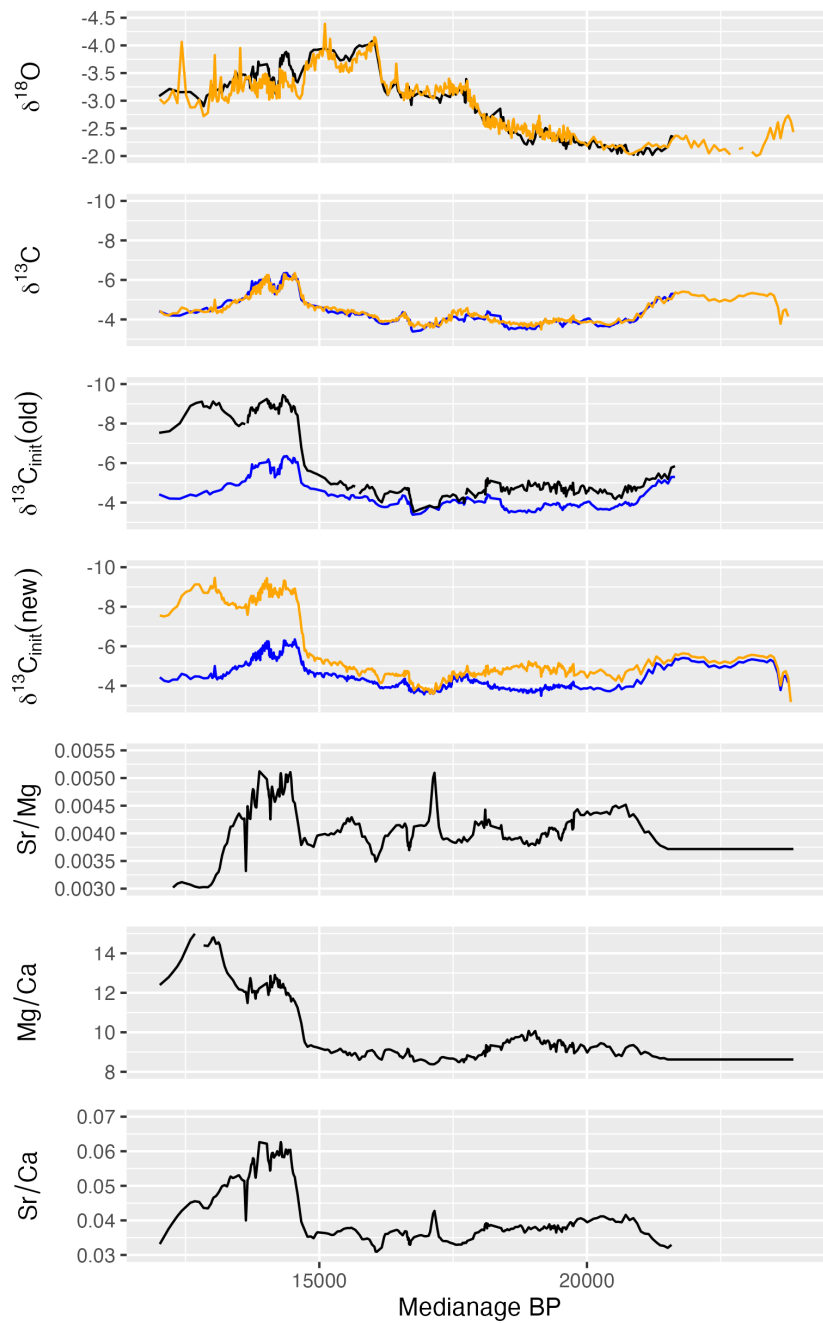


Figure A9. Comparison between a prior low resolution trench (black), that was aligned to the new data (orange) with the help of Qanalyseries (Kotov and Pälike, 2018). Trace element data of the old trench was used to correct $\delta^{13}C$ in the new trench. The blue line indicates the measured $\delta^{13}C$ before applying the degassing correction.

Author contributions. The study was designed by H.M.S., R.I. and L.E. L.E. wrote the paper with substantial scientific contributions from R.I, L.G., A.L.C.H., and H.M.S. Uran-Thorium dating was conducted by C.P-M and H.C. Fieldwork campaigns were organized by H.M.S. Sample preparation and geochemical labwork was carried out by L.E. and ETH Zürich technical staff. GCM dye tracer modeling was conducted by L.E., R.I and University of Leeds technical staff. All authors contributed to reading and editing the text.

645 *Competing interests.* There are no competing interests to declare.

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