

Review article

Temperature-sensitivity of water stable isotopes in global precipitation: a groundwater noble gas-based, paleoclimate perspective

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Author contributions: Conceptualization: DVB; Data Curation: DVB, AL; Funding Acquisition: DVB; Investigation: DVB, AL, CR, RR, AMS, WA, TM, PHB; Methodology: DVB, AL, CR, GF, CPH, RR, JZ, DG, AMS, WA, PHB; Project Administration: DVB; Resources: DVB, AL, CR, RR, AMS, SD, KK, KY, AC, HB, ANL; Validation: DVB; Visualization: DVB, AL, CR, GF, SD, KK, KY, AC, HB, ANL, PHB; Writing – original draft: DVB; Writing – review & editing: all authors

Data, code, and outputs: Bekaert et al. (2026) <https://doi.org/10.60520/IEDA/114589>

Submitted: 2025-12-22

Accepted: 2026-08-28

Published: 2026-10-19

Production editor:

Jessica Tierney

Handling editor:

Jordon Hemingway

Reviews:

Two anonymous reviewers

Copyediting:

Elliot Carter,

Marthe Klöcking

Water stable isotopes ($\delta^2\text{H}$ and $\delta^{18}\text{O}$) are widely used to reconstruct past hydrological changes and paleoclimate variability. A key limitation, however, is that fossil isotope records reflect not only local temperature but also changes in circulation, moisture sources, and hydrological processes along air-mass trajectories. Although modern monthly and interannual $\delta^2\text{H}$ – $\delta^{18}\text{O}$ –temperature relationships have been proposed to calibrate paleoclimate reconstructions, their applicability to longer timescales, such as glacial–interglacial transitions, remains uncertain. Here, we use fossil groundwaters worldwide as direct archives of past annual temperature and $\delta^2\text{H}$ – $\delta^{18}\text{O}$ co-variations over the last ~35,000 years. Our global compilation of water stable isotope and noble gas temperature records (746 samples from 39 locations) shows that, while significant relationships occur in several regions, European fossil groundwaters exhibit the most consistent $\delta^{18}\text{O}$ –temperature correlations. In Western Europe, $\delta^{18}\text{O}$ variations during deglaciation were primarily temperature-driven, allowing $\delta^{18}\text{O}$ to be used as a paleothermometer on glacial–interglacial timescales. In contrast, weaker correlations in North America, Asia, and Africa likely reflect stronger influences from atmospheric circulation, convection, and/or continental moisture recycling. For Western Europe, the observed relationships are consistent with a simple Rayleigh distillation model, although sites further inland seem to reflect more complex atmospheric processes. The mean European fossil groundwater-based $\delta^{18}\text{O}$ –temperature slope, corrected for ice-volume effects since the LGM, is $0.48 \pm 0.21 \text{ ‰ } ^\circ\text{C}^{-1}$ (1 SD), comparable to modern monthly relationships for precipitation ($0.29 \pm 0.07 \text{ ‰ } ^\circ\text{C}^{-1}$, 1 SD) but lower than modern interannual estimates ($1.04 \pm 0.30 \text{ ‰ } ^\circ\text{C}^{-1}$, 1 SD). These results provide robust constraints for interpreting isotope-based climate archives and evaluating isotope-enabled climate models.

1 Introduction

The use of water stable isotopes (WSI) as paleoclimate tracers (e.g., from speleothem isotope records; McDermott, 2004) is based on modern observations of how climate variability and changes in hydrological processes affect the stable isotope composition of contemporary precipitation. The annual mean $\delta^{18}\text{O}$ and $\delta^2\text{H}$ (also denoted δD) values of modern precipitation (where:

$$\delta X = [(R_X^{\text{sample}} - R_X^{\text{standard}})/R_X^{\text{standard}}] \times 1000 \quad (1)$$

with R_X representing a given isotopic ratio of an element X) have long been recognized to correlate with the Mean Annual Air Temperature (MAAT; Dansgaard, 1964; Rozanski et al., 1992). For mid- and high-latitude coastal regions, a spatial temperature sensitivity of WSI in precipitation has been derived ($\delta^{18}\text{O} \sim 0.69 \times \text{MAAT} - 13.6 \text{ ‰}$, $\delta^2\text{H} \sim 5.6 \times \text{MAAT} - 100 \text{ ‰}$, with MAAT given in $^{\circ}\text{C}$) and reproduced by simple Rayleigh distillation-based modelling (Dansgaard, 1964). By considering an isolated air mass undergoing isobaric cooling, with rain forming in isotopic equilibrium with the vapor between 0 and 20°C , theoretically derived slopes of the $\delta^{18}\text{O}$ –MAAT relationship indeed vary between 0.72 and $0.52 \text{ ‰ } ^{\circ}\text{C}^{-1}$, which encompass the observed relationship of $\sim 0.69 \text{ ‰ } ^{\circ}\text{C}^{-1}$ (Dansgaard, 1964). This empirical relationship between $\delta^{18}\text{O}$ and surface air temperature can hence be taken as a surrogate for the temporal slope at a given site. However, this conventional approach yields significant uncertainties, which have been estimated to be -10 to $+30 \text{ ‰}$ over glacial-interglacial transitions for central Antarctic ice cores (Jouzel et al., 2003).

At the regional scale, the temporal temperature sensitivity of WSI in precipitation is moreover not spatially homogeneous. Previous studies compiling modern precipitation data from North America and Europe have established distinct linear equations relating the $\delta^{18}\text{O}$ (and $\delta^2\text{H}$) of modern precipitation to air temperature variations, depending on the meteorological station considered (Rozanski et al., 1992). These correlations may also change with the period of integration, with notable discrepancies between correlations observed at the daily, seasonal, interannual or longer time scales (Rozanski et al., 1992). The hypotheses that we aim to test in this article are that (i) modern correlations are applicable to glacial-interglacial timescales (Pfahl and Sodemann, 2014), and (ii) other sources of WSI variations (i.e., moisture source and fractionation effects) are negligible or accounted for. If these hypotheses are valid, then modern estimates of the temperature sensitivity of WSI variations can be used to convert $\delta^{18}\text{O}$ – $\delta^2\text{H}$ variations archived in the geological record into paleotemperature signals. In particular, WSI variations of paleowaters recorded in speleothem calcite represent high-resolution paleoclimate archives (e.g., Fairchild et al., 2001) that can be precisely dated using the U/Th method (Richards and Dorale, 2003) or laminae counting with up to annual time resolution (Frisia et al., 2003). Although speleothem carbonate (e.g., for oxygen) and fluid inclusion (e.g., for hydrogen and oxygen) analyses provide key insight into the rate and

timing of terrestrial paleoclimate changes during the late Quaternary (e.g., Wang et al., 2001; Fleitmann et al., 2003; McDermott, 2004; Yuan et al., 2004; Zhang et al., 2018; Affolter et al., 2019), the deconvolution of $\delta^{18}\text{O}$ signals into individual climatic components (e.g., temperature) remains challenging and sometimes controversial.

One could imagine that the relationship between inter-annual changes in $\delta^{18}\text{O}$ in precipitation and MAAT at a given location would be most appropriate for paleoclimate reconstructions based on water isotope records. However, the magnitude of yearly variations in modern precipitation is often too small to offer the dynamic range required to precisely constrain the long-term temperature sensitivity of WSI (Rozanski et al., 1992). Hence, because they cover a wider range of different climate states, the $\delta^{18}\text{O}$ –temperature (T) coefficients derived from seasonal hydroclimate cycles have been considered for paleoclimate applications (e.g., Affolter et al., 2019). However, many processes and parameters control the WSI composition of precipitation (during evaporation, condensation, freezing, transport). Although these forcings have varied since the last glacial period, they are not reflected in the seasonal fluctuations of the modern hydrological cycle. For example, changes in the global $\delta^{18}\text{O}$ of the ocean (Schrag et al., 2002) or the relative humidity over land undoubtedly modulated the composition of continental precipitation during the last deglacial warming (arguably throughout all deglacial events of the Pleistocene). Additionally, a recent global re-analysis of long-term (60 year) temporal correlations between temperature and precipitation $\delta^{18}\text{O}$ suggested that temperature effects on WSI in precipitation may be better explained by non-linear trends, reflecting the multiple effects of climate dynamics and hydrological processes at regional and global scales (e.g., continental recycling seasonality, convective processes, atmosphere-ocean interactions; Vystavna et al., 2021). Thus, over millennial timescales, WSI variations in precipitation likely reflect an integrated hydroclimatic memory of numerous cumulative hydrological processes that cannot be captured by modern precipitation analyses. These caveats highlight the need for an alternative approach to better constrain the temperature sensitivity of WSI temporal variations in precipitation, with the overarching goal of better reconstructing past climate variability.

The approach proposed in this study relies on a compilation of co-variations between Mean Annual Surface Temperature (MAST) derived from dissolved noble gas concentrations in groundwater (hereafter NGT), and the stable water isotope (WSI) composition of paleoprecipitation, as recorded in Late Pleistocene to Holocene-age groundwater. While NGTs are rigorously interpreted as MAST, they are often compared to MAAT assuming the air–surface offset represents a small, systematic offset whereby MAST exceeds MAAT globally by $\sim 2^{\circ}\text{C}$ (Seltzer et al., 2021), such that temporal changes in MAST and MAAT are approximately equal. Here, the large dataset of NGT and WSI is used to constrain the regional temperature sensitivity of WSI temporal variations in global precipitation over the last deglaciation. Our compilation includes samples from North

and South America, Europe, Africa, and Asia. For each region, we assess the strength of WSI–NGT correlations. As many regions do not exhibit significant correlations, we examine potential causes of WSI–NGT decorrelation and evaluate the reliability of the WSI paleothermometer accordingly.

Because European records show the most robust WSI–NGT correlations, we focus on Europe to carry out a detailed comparison between fossil groundwater observations and modern monthly-averaged or interannual variations in precipitation from meteorological station data. We show that while paleocorrelations in Western Europe are broadly consistent with a Rayleigh distillation model of WSI evolution during the deglaciation, Central European fossil groundwater data require additional processes (e.g., potential changes in moisture source, continental recycling, deep convection) to be considered. Finally, we compare $\delta^{18}\text{O}$ –temperature slopes derived from fossil groundwater data with those simulated by General Circulation Models (GCMs) for the Last Glacial Maximum (LGM, 19–26 kyr ago), in order to evaluate the ability of climate models to reproduce observed slopes. The overarching goal of this study is to define robust transfer functions between temperature and $\delta^{18}\text{O}$ (and $\delta^2\text{H}$), which can be applied to develop more reliable paleotemperature reconstructions from WSI-based archives. An important advantage of WSI reconstructions is their higher temporal resolution and broader spatial coverage compared to NGT-based reconstructions.

2 Methods

2.1 Dissolved noble gases in fossil groundwater

One promising, yet overlooked, avenue to constrain the WSI–T correlations over millennial timescales is fossil groundwater (Rozanski et al., 1992; Loosli et al., 2001; Aeschbach-Hertig and Solomon, 2013). The term "fossil" is interpreted differently across various studies. In some contexts, it refers to water that surpasses a specific "absolute" age threshold, such as groundwater older than 1,000 years (Margat et al., 2006). In other contexts, particularly in semiarid and arid regions, it describes groundwater systems where current recharge is negligible compared to the overall size of the system (Issar et al., 1972; Tsur et al., 1989). Here, fossil groundwater refers to water that saturated the aquifer before the Holocene (Jasechko et al., 2017). Historically, the recognition that the stable isotope composition of water molecules in rainwater (i.e., $\delta^{18}\text{O}$ and $\delta^2\text{H}$) depends on several climatic factors, including air temperature, rain amount, evaporation processes, altitude, and latitude of precipitation (e.g., Craig, 1961; Dansgaard, 1964) paved the way for isotope hydrology and hydroclimatology. On one hand, the $\delta^{18}\text{O}$ value of groundwater mainly reflects that of precipitation (e.g., Zou et al., 2023), which is itself linearly dependent on the mean air temperature at mid and high latitudes (Dansgaard, 1964; Rozanski, 1985). On the other hand,

due to their temperature-dependent solubility in water and insensitivity to biological and chemical processes, dissolved atmospheric neon (Ne), argon (Ar), krypton (Kr), and xenon (Xe) concentrations in ancient groundwater have long been recognized as reliable proxies of past climates, with a physics-based, quantitative link to past MAST (e.g., Mazor, 1972; Stute et al., 1995; Seltzer et al., 2021). Today, about a third of Earth's land area holds LGM-aged (or older) groundwater (e.g., Befus et al., 2017). These ancient groundwater reservoirs with residence times of tens of millennia are contained within deep and confined aquifers that are marginally sensitive to modern climate fluctuations and undergo negligible mixing with younger groundwater parcels potentially affected by anthropogenic contamination (Foster and Chilton, 2003). Co-variations of groundwater stable isotopes and NGT over the Last Glacial Interglacial Transition (LGIT) represent a promising and underexploited way to constrain the sensitivity of the stable isotope thermometer (e.g., Rozanski et al., 1992; Loosli et al., 2001) for high resolution $\delta^{18}\text{O}$ -based, glacial-interglacial paleoclimate reconstructions in the continental realm (e.g., from speleothems).

In this contribution, we focus on fossil groundwater data worldwide that include paleoclimate reconstructions from the LGM (supporting data compilation of Bekaert et al., 2026, available at <https://doi.org/10.60520/IEDA/114589>). This period represents a crucial benchmark for climate model evaluation (e.g., for climate sensitivity; Harrison et al., 2014; Kageyama et al., 2021), as it is the most recent quasi-equilibrium climate state with dramatically different conditions (atmospheric circulation, global climate state, vegetation, glacial extension) from today, and with a plethora of climate records (e.g. Clark et al., 2009; Tierney et al., 2020; Seltzer et al., 2021). The exact timing of the LGM is not clearly established and may vary from one community to the other: while it is usually taken at 21 kyr ago for models and model intercomparison projects, the strict maximum in global ice volume is considered at ~26 kyr ago (Grant et al., 2014). In any case, this time period was characterized by a low atmospheric CO_2 concentration ($[\text{CO}_2]_{\text{atm}} \sim 185 \mu\text{L L}^{-1}$), which subsequently increased to pre-industrial levels ($\sim 280 \mu\text{L L}^{-1}$) in conjunction with changes in high latitude insolation and rising global temperatures (Shakun et al., 2012; Parrenin et al., 2013). For this reason, paleoclimatologists have long sought to refine estimates of climate changes between the LGM and Holocene, as a benchmark for climate models investigating the response of regional temperature and precipitation distributions to rising global CO_2 (e.g., Schmittner et al., 2011; Tierney et al., 2020; Legrain et al., 2023). Direct comparisons between continental paleoclimate proxy data and GCM outputs for the LGM represent a propitious way to consistently evaluate climate models and their ability to capture the heterogeneous nature of past and future continental hydroclimates (Braconnot et al., 2012), as well as the underlying atmospheric processes, notably the continentality effects and the elevation-dependent warming (e.g. Legrain et al., 2023).

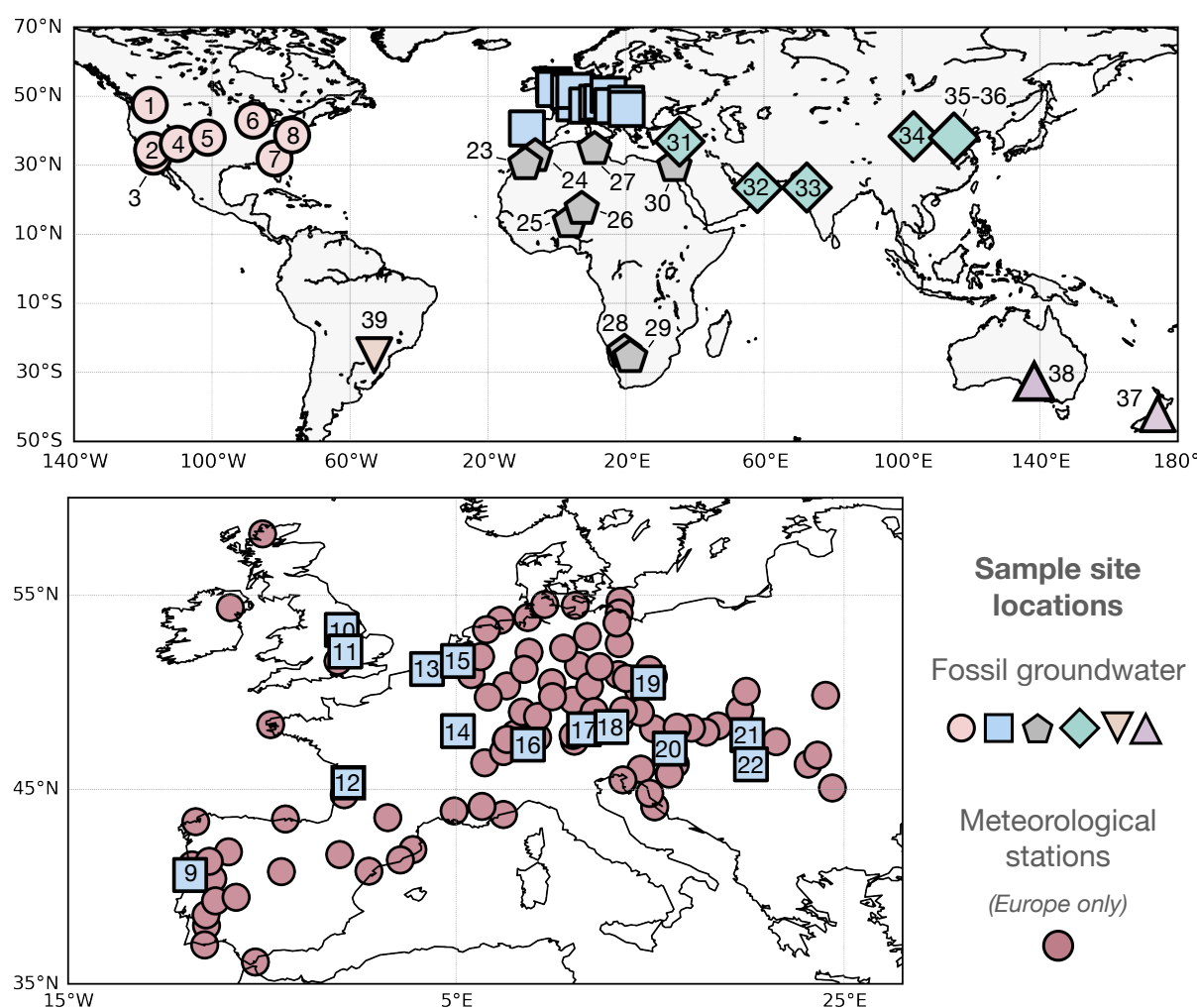


Figure 1. Global fossil groundwater and European meteorological station locations. Numbers refer to those used in data compilation (Bekaert et al., 2026) for information about site location and data characteristics. Note that sample sites are grouped by continent for simplicity, but this does not imply climatic uniformity within each group. There are, of course, substantial differences between climate dynamics in, for example, Turkey and China (Asia), or Botswana and Tunisia (Africa).

2.2 Fossil groundwater compilation

All the fossil groundwater locations used in this study are listed in the data compilation (Bekaert et al., 2026). We strived to consider all available data that contained ^{14}C dating, WSI and dissolved noble gas data, with no a priori quality control or minimum amount of data. Our compilation includes 39 fossil groundwater records: 8 from North America, 14 from Europe, 8 from Africa, 6 from Asia, 2 from Oceania, and 1 from Brazil. For the latter, no water stable isotope data could be found for the seminal Stute et al. (1995) study, and so we only use data from Kirchheim et al. (2021). When possible, NGT values were all recomputed following the approach by Seltzer et al. (2021), using the closed-system equilibration (CE) model (Aeschbach-Hertig et al., 2000). This model accounts for water table fluctuations that entrain bubbles of soil air, which partially dissolve under elevated hydrostatic pressure at solubility equilibrium. Only a few studies were not processed with the CE model, for instance, if alternative models were found to be more appropriate (e.g., the unfractionated air model

in (Aeschbach-Hertig and Solomon, 2013) or if raw noble gas data were unavailable (e.g., Andrews et al., 1985). For studies lacking raw data, NGT and WSI compositions were extracted using graph digitizer software to retrieve original data from published figures. All uncertainties reported in this study are 1 standard deviation (SD).

The reconstructed NGT, ^{14}C ages, and WSI data for each location are provided in the data compilation (Bekaert et al., 2026). To calculate the $\delta^{18}\text{O}$ -NGT and $\delta^2\text{H}$ -NGT ($\text{‰}^\circ\text{C}^{-1}$) relationships, we used the modified error-weighted least squares algorithm (model 1) by York et al. (2004) implemented in Isoplot-R (Vermeesch, 2018), which accounts for analytical uncertainties, assuming that the scatter between data points is solely due to analytical error (Vermeesch, 2018). For datasets without reported uncertainties, we assumed typical dissolved noble gas concentration uncertainties of $\pm 2\%$, and errors of ± 0.1 per mil (‰) and ± 1 ‰ for $\delta^{18}\text{O}$ and $\delta^2\text{H}$, respectively (1 SD). Given the inherent challenges of using ^{14}C in dissolved inorganic carbon for absolute groundwater dating (e.g., Han and Plummer, 2016), the ^{14}C age ranges

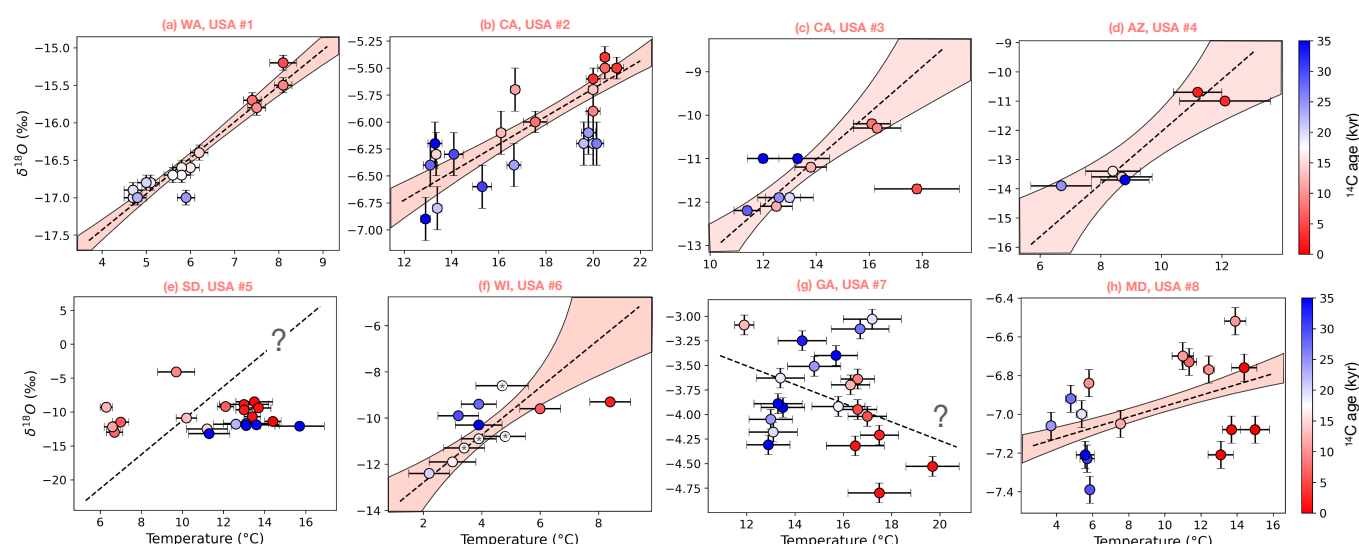


Figure 2. $\delta^{18}\text{O}$ –Temperature (NGT) reconstructions from USA fossil groundwaters. Marker colours refer to ^{14}C ages, with blue shades representing the last glacial period. Data without available ^{14}C ages are marked with an asterisk. As described in Section 2.2, given the inherent challenges of using ^{14}C in dissolved inorganic carbon for absolute groundwater dating (Han and Plummer, 2016), ^{14}C ages should be considered indicative only as these were taken directly from the original studies and, therefore not uniformly processed/calibrated. Numbers given as # correspond to site locations given on Figure 1. Dashed lines and associated envelopes represent error-weighted least squares fitting through the data using “model 1” by York et al. (2004) implemented in Isoplot-R (Vermeesch, 2018). Interrogation mark in panels (e) and (g) means no linear correlation was found. Reported uncertainties are 1 SD.

provided in the data compilation (Bekaert et al., 2026) should be considered indicative only. Additionally, these ^{14}C age data were not uniformly processed and were taken directly from the original studies: no ^{14}C age recalibration was applied, as (i) ^{14}C ages of fossil groundwater should only be regarded as indicative (with 15 to 45 kyr-old groundwater typically representing potential LGM samples; Seltzer et al., 2021), and (ii) we are anyway only interested in temporal evolutions since the last glacial period and especially the past ~ 35 kyr, so the exact timing of climate variations is not relevant. Finally, some datasets containing dissolved noble gas concentrations and WSI data were excluded due to identified issues with the hydrological context of the study (e.g., mixing between distinct water bodies, as noted by Visser et al., 2022, in Nevada, USA).

2.3 Modern European precipitation data

All modern European precipitation data referenced in this study (from 94 meteorological stations; Figure 1) were obtained from the WISER (Water Isotope System for Data Analysis, Visualization, and Electronic Retrieval) portal of the Global Network of Isotopes in Precipitation (GNIP), managed by the International Atomic Energy Agency and the World Meteorological Organization (data compilation available from Bekaert et al., 2026). These data consist of average air temperature and precipitation WSI data from the modern climatological period (i.e., the past ~ 60 years), including monthly and long-term annual abundance-weighted $\delta^2\text{H}$ and $\delta^{18}\text{O}$ means. Note that the modern climatological period is distinct from pre-industrial (PI) conditions (year ~ 1850) considered in climate models. For some meteorological stations, the datasets may only comprise a couple

years of data. The whole dataset is provided in the data compilation (Bekaert et al., 2026).

2.4 General Circulation Models (GCM)

We evaluated medium-to-high resolution isotope-enabled LGM simulations outputs from 5 GCMs: LMDZ4 (Risi et al., 2010), ECHAM6-wiso (Cauquoin et al., 2023), IsoGSM (Lan et al., 2023), MIROC5-iso (Kino et al., 2024) and GISS (Roberts et al., 2011). To do so, we used LGM and pre-industrial (PI, corresponding to the climate conditions at 1850 AD) global simulations and computed the slopes of the $\delta^{18}\text{O}$ –MAST relationships as:

$$\text{Slope}(\text{lat}, \text{lon}) = \frac{\text{MA } \delta^{18}\text{O}(\text{lat}, \text{lon})_{\text{LGM}} - \text{MA } \delta^{18}\text{O}(\text{lat}, \text{lon})_{\text{PI}}}{\text{MAST } \delta^{18}\text{O}(\text{lat}, \text{lon})_{\text{LGM}} - \text{MAST } \delta^{18}\text{O}(\text{lat}, \text{lon})_{\text{PI}}} \quad (2)$$

where (lat, lon) refers to the coordinates of each grid cell, MAST is the Mean Annual Surface Temperature, MA $\delta^{18}\text{O}$ is the Mean Annual $\delta^{18}\text{O}$. We interpolated the global map of slopes using linear interpolation, increasing the total number of grid cells by a factor of 2. From this interpolated map, we extracted slope values from the 4 grid cells closest to each groundwater sample location. We then computed the average slope and associated standard deviation from the selected grid cells and compared these results with the actual $\delta^{18}\text{O}$ –MAST correlations derived from groundwater data. Below, we provide a short description of each climate model used in this study.

2.4.1 LMDZ4

The LMDZ4 climate model is a version of the LMDZ (Laboratoire de Météorologie Dynamique Zoom) GCM which

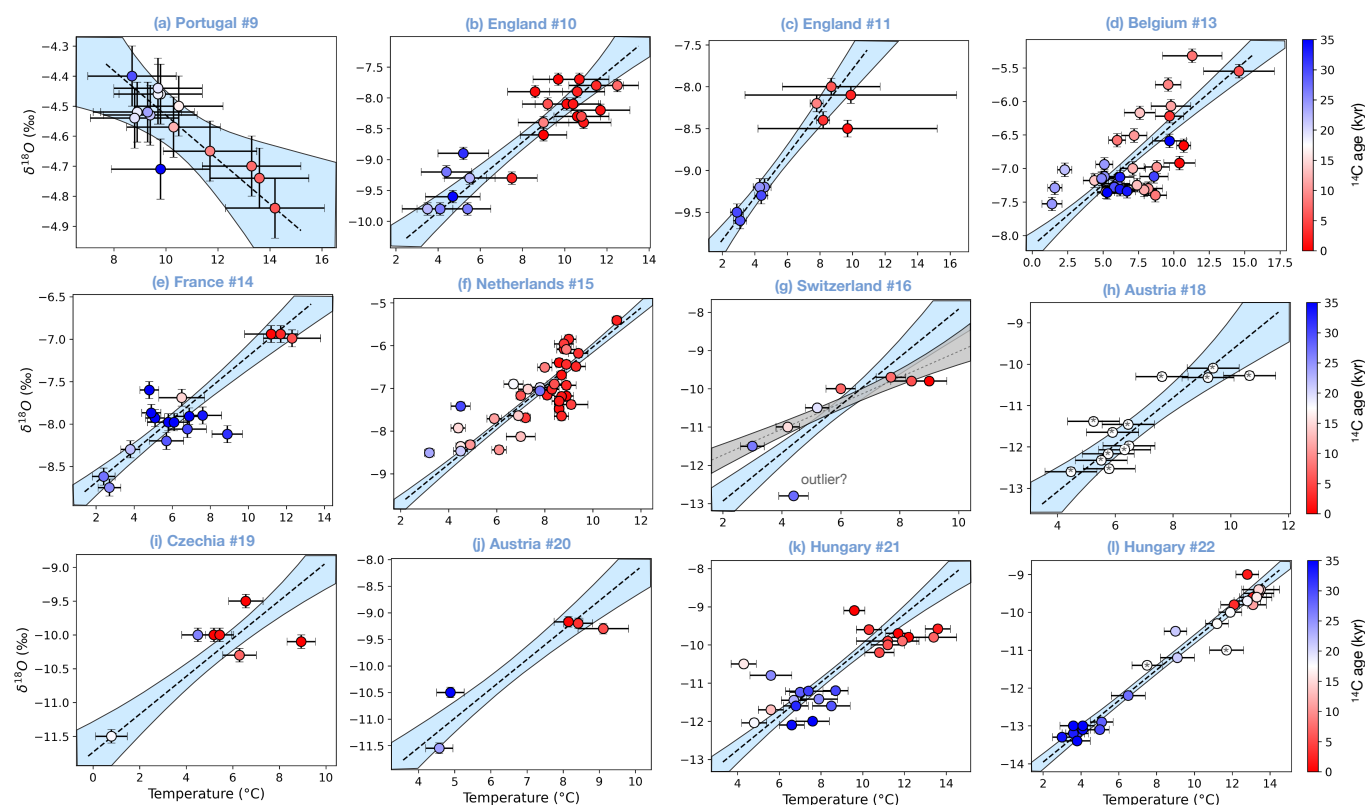


Figure 3. $\delta^{18}\text{O}$ –Temperature (NGT) reconstructions from European fossil groundwaters. Marker colours refer to ^{14}C ages, with blue shades representing the last glacial period. Data without available ^{14}C ages are marked with an asterisk. As described in Section 2.2, given the inherent challenges of using ^{14}C in dissolved inorganic carbon for absolute groundwater dating (Han and Plummer, 2016), ^{14}C ages should be considered indicative only as these were taken directly from the original studies and, therefore not uniformly processed/calibrated. The grey correlation on panel (g) shows the regression through the data once the potential outlier has been excluded. Numbers given as # correspond to site locations given on Figure 1. Dashed lines and associated envelopes represent error-weighted least squares fitting through the data using “model 1” by York et al. (2004) implemented in Isoplot-R (Vermeesch, 2018). Interrogation mark in panels (e) and (g) means no linear correlation was found. Reported uncertainties are 1 SD.

contributes to the IPSL Earth system model, developed by the Institut Pierre-Simon Laplace (IPSL). The dynamical equations are discretized in a latitude-longitude grid, with a standard resolution of $2.5^\circ \times 3.75^\circ$ and 19 vertical levels, with each grid cell divided into four subsurfaces: ocean, land, ice sheet and sea ice. The physical package is described by Hourdin et al. (2006). Details about the model and how isotopic processes are prescribed in LMDZ4 are provided in Risi et al. (2010).

2.4.2 ECHAM6-wiso

ECHAM6 is the sixth version of the ECHAM atmospheric general circulation model developed by the Max Planck Institute for Meteorology (Stevens et al., 2013). The ECHAM6-wiso version includes water isotopes, as described by Cauquoin et al. (2019) and Cauquoin and Werner (2021). It runs as an atmosphere-only model (uncoupled from an ocean GCM), with prescribed Sea Surface Temperatures (SST) and sea ice from AMIP and GLOMAP (Paul et al., 2021) datasets for PI and LGM conditions, respectively. The model resolution is approximately $1.9^\circ \times 1.9^\circ$ horizontal resolution and 47 vertical levels. Additional details about the model and how isotopic

processes are prescribed in ECHAM6-wiso are provided by Cauquoin et al. (2023).

2.4.3 IsoGSM

IsoGSM is an isotope-enabled version of the Global Spectral Model (GSM) developed by NOAA’s National Centers for Environmental Prediction. SST and sea ice are extracted from the sixth version of the Model for Interdisciplinary Research on Climate (MIROC6) that takes part in the Coupled Model Intercomparison Project Phase 6 and has demonstrated a general improvement in terms of the LGM sea ice extent (Vadsaria et al., 2021). The model resolution is approximately $1.8^\circ \times 1.8^\circ$ horizontal resolution and 28 vertical levels. Additional information about the model is provided in Lan et al. (2023).

2.4.4 MIROC5-iso

MIROC5-iso is built on the MIROC5 model (Model for Interdisciplinary Research on Climate version 5) developed through a collaboration between the Atmosphere and Ocean Research Institute (AORI) at the University of Tokyo, the Japan Agency for Marine-Earth Science and Technology (JAMSTEC), and the National Institute for Environmental

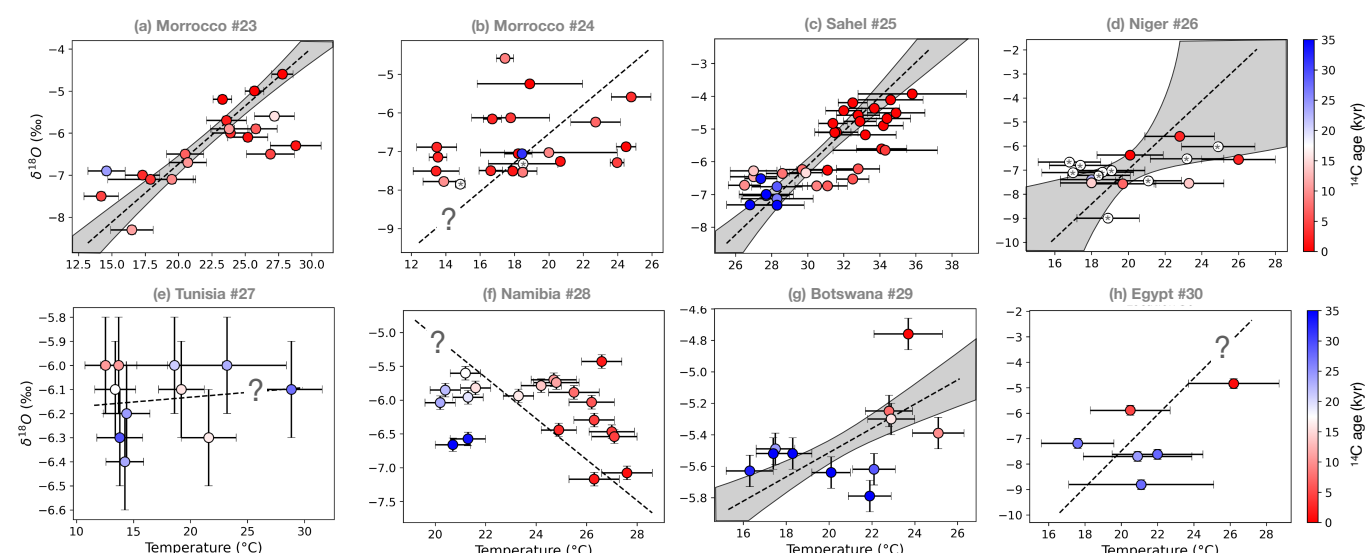


Figure 4. $\delta^{18}\text{O}$ –Temperature (NGT) reconstructions from African fossil groundwaters. Marker colors refer to ^{14}C ages, with blue shades representing the last glacial period. Data without available ^{14}C ages are marked with an asterisk. As described in Section 2.2, given the inherent challenges of using ^{14}C in dissolved inorganic carbon for absolute groundwater dating (Han and Plummer, 2016), ^{14}C ages should be considered indicative only as these were taken directly from the original studies and, therefore not uniformly processed/calibrated. Numbers given as # correspond to site locations given on Figure 1. Dashed lines and associated envelopes represent error-weighted least squares fitting through the data using “model 1” by York et al. (2004) implemented in Isoplot-R (Vermeesch, 2018). Interrogation mark in panels (e) and (g) means no linear correlation was found. Reported uncertainties are 1 SD

Studies (NIES). Version MIROC5-iso, where water isotopes in the atmosphere and land surface parts were implemented by Okazaki and Yoshimura (2017, 2019) was used in this study. The model resolution is approximately $2.8^\circ \times 2.8^\circ$ horizontal resolution and 40 vertical levels. It runs as an atmosphere-only model (uncoupled from an ocean GCM), with prescribed SSTs and sea ice from observational datasets and paleo reconstructions (specifically, Paul et al., 2021), making it suitable for isotope-enabled paleoclimate simulations. Additional information about the model is provided in Kino et al. (2024).

2.4.5 GISS-E2.1

The GISS-E2.1 model refers to a series of climate models developed by the NASA Goddard Institute for Space Studies, in collaboration with Columbia University (Roberts et al., 2011). Here, we used outputs from simulations conducted using the atmospheric component of the GISS-E2.1 GCM, with a horizontal resolution of approximately 2° latitude $\times$ 2.5° longitude resolution and 40 vertical levels (Kelley et al., 2020). The implementation of water stable isotopes in GISS-E2.1 was based on previous versions of the GISS model (e.g., Schmidt et al., 2007). Additional information about the model is provided in Zhang et al. (2025).

2.4.6 Last Glacial Maximum reanalysis

The climate model used in Osman et al. (2021) the water isotope-enabled Community Earth System Model, versions 1.2 and 1.3 (iCESM1.2 and iCESM1.3). iCESM explicitly simulates the transport and transformation of stable water isotopes in all of the component models and has been shown

to reproduce key features of climate and isotopes in present day and paleoclimate observations (Brady et al., 2019). All of the iCESM simulations have a horizontal resolution of 1.9° latitude $\times$ 2.5° longitude in the atmosphere and land, and a nominal 1° resolution in the ocean. The paleoclimate data assimilation method incorporates an offline ensemble square root Kalman filter approach, following the methodology of Tierney et al. (2020). Additional information about the model the “Last Glacial Maximum reanalysis” (LGMR) used in this study is provided in Osman et al. (2021).

3 Results

3.1 Overview of the data compilation

The $\delta^{18}\text{O}$ –NGT and $\delta^2\text{H}$ –NGT ($\text{‰ } ^\circ\text{C}^{-1}$) relationships derived from these fossil groundwater data are summarized in the data compilation (Bekaert et al., 2026), with $\delta^{18}\text{O}$ –NGT correlations illustrated in Figures 2–5. Note that similar correlations would be obtained when plotting and $\delta^2\text{H}$ –NGT. Among the 39 fossil groundwater records compiled in this study, 32 exhibit a positive correlation between reconstructed NGT and WSI compositions, consistent with lower $\delta^2\text{H}$ – $\delta^{18}\text{O}$ in precipitation during colder climates. Excellent correlation coefficients are found in both North America and Europe (R^2 up to 0.91 and 0.94, respectively), consistent with a temperature-driven change of WSI compositions since the LGM. While European and North American fossil groundwaters generally exhibit significant correlation coefficients, indicating that long-term temperature sensitivities of WSI variations in precipitation can be derived, other regions show fewer fossil records

with statistically significant correlations. In these regions, additional processes may influence noble gas and/or WSI systematics of regional precipitation.

3.2 Water stable isotope systematics

In $\delta^2\text{H}$ versus $\delta^{18}\text{O}$ space, the fossil groundwater data generally plot close to the Global Meteoric Water Line (GMWL; $7.96 \times \delta^{18}\text{O} + 8.86$; Araguás-Araguás et al., 2000, Figures 6–9), confirming that the WSI compositions of fossil groundwater primarily reflect the composition of past precipitation. However, several records show significant deviations from the GMWL, likely due to secondary processes, such as rock-water interactions, evaporation – from the source (ocean/sea) and at later stages between the time of precipitation and groundwater recharge – and/or mixing between different groundwater bodies (see Section 4.1). Records affected by such significant deviations of WSI from the GMWL (e.g., Germany; Winter et al., 2024, Fig. 6) were not considered for interpretation of WSI–NGT correlations. For each paleoclimate record, we also calculated the time evolution of deviations from the GMWL (referred to as deuterium excess, or D-excess = $\delta^2\text{H} - 7.96 \times \delta^{18}\text{O}$; Dansgaard, 1964). For most locations, we found no clear pattern of D-excess evolution over the past ~35 kyr (Figures 6–7).

3.3 Quality of the NGT-WSI correlations

When comparing the Mean Square Weighted Deviation (MSWD) or correlation coefficients (R^2 ; Figure 8) of $\delta^{18}\text{O}$ –MAST correlations worldwide, it appears that the regressions in Europe are generally of higher quality (i.e., with MSWD and R^2 values closer to 1) than in other regions (i.e., North America, Africa, Asia). Good correlations are observed in some regions of North America (e.g., Zhu and Kipfer, 2010; Seltzer et al., 2025) with MSWD values of 0.7 and 1.5, and R^2 values of 0.85 and 0.92, respectively. However, the qualities of most other North American regressions are comparable to those for Africa and Asia, with a lower quality than what is typically observed in Europe. This suggests that Europe is the region where $\delta^{18}\text{O}$ variations due to factors other than temperature have the lowest impact on the $\delta^{18}\text{O}$ variations observed during deglacial warming. In other words, our results suggest that temperature exerts a dominant control on the stable isotope composition of European precipitation, while other processes may contribute to WSI variations during deglacial warming in other regions.

3.4 Modern European precipitation systematics

For meteorological stations with at least 10 years of WSI and temperature data available, we compared the temperature sensitivities of $\delta^{18}\text{O}$ variations derived from monthly and annual averages. For the 20 stations shown in Figure 9, we found that the slopes of the $\delta^{18}\text{O}$ –surface air temperature correlations derived from monthly average data are consistently smaller (ranging from 0.17 to 0.42 ‰ °C^{−1}) than those derived from interannual averages (ranging from 0.59 to 1.75 ‰ °C^{−1}, consistent with Gat et al., 2001; Kohn and Welker, 2005). These slopes are also lower than

those derived from the spatial relationship between annual averages of precipitation $\delta^{18}\text{O}$ and surface air temperature for high-latitude stations (~0.69 ‰ °C^{−1}; Dansgaard, 1964; Rozanski et al., 1992). While some of the observed variability likely reflects geographical gradients, we found no clear relationship between the slopes of the monthly mean-based $\delta^{18}\text{O}$ –temperature correlations and the latitude or longitude of the stations (Figure 9). Although $\delta^{18}\text{O}$ –T correlations based on annual meteorological data (averaging 1.04 ± 0.30 ‰ °C^{−1}; 1 SD) for the 20 stations considered here) are considered most appropriate for paleoclimatic reconstructions based on isotope records (Rozanski et al., 1992), they appear more variable than those based on monthly averages (0.29 ± 0.07 ‰ °C^{−1}; 1 SD). These results emphasize the significant limitations associated with using modern meteorological station data to derive robust $\delta^{18}\text{O}$ –T correlations for paleoclimatic reconstructions: the choice of the timescale of modern calibrations may introduce a significant yet unrecognized bias in the reconstructed paleotemperature.

4 Discussion

4.1 Potential sources of $\delta^{18}\text{O}$ –/ $\delta^2\text{H}$ –NGT decorrelations

Water stable isotopes and noble gases are related to temperature via distinct physical processes (or mechanisms). While groundwater noble gas concentrations primarily reflect mean annual temperature at the water table during recharge, groundwater $\delta^{18}\text{O}$ and $\delta^2\text{H}$ reflect the composition of local precipitation, which is itself dependent on temperature and affected by a wide range of atmospheric processes (Galewsky et al., 2016), and subsequent processes in the soil. Below, we summarize the processes that may affect the WSI and noble gas systematics of fossil groundwater to evaluate potential variations of the $\delta^{18}\text{O}$ –/ $\delta^2\text{H}$ –NGT relationships in the fossil groundwater record. Note that potential changes to the original groundwater WSI composition resulting from water–rock interactions are generally only appreciable in extremely old groundwater (e.g., millions of years old or older) or in high-temperature hydrothermal systems (e.g., Karolytė et al., 2017; Jasechko, 2019). As these conditions are not relevant to this study, this process is not discussed further.

4.1.1 Water stable isotope systematics

4.1.1.1 Water vapor formation and atmospheric circulation. The variability of WSI in regional precipitation is primarily linked to changes in the conditions of water vapor formation and transmission from oceanic to continental regions. Crucially, many parameters (WSI composition and salinity of the ocean, WSI composition of ambient vapor, relative humidity of the air, diffusion and/or mixing at the water–air interface, temperature of both phases) control the WSI composition ($\delta^{18}\text{O}$, $\delta^2\text{H}$, but also D-excess) of atmospheric moisture generated by evaporation of the surface ocean (e.g., Cappa et al., 2003; Gonfiantini et al., 2018). Accordingly, D-excess variations in paleoclimatic

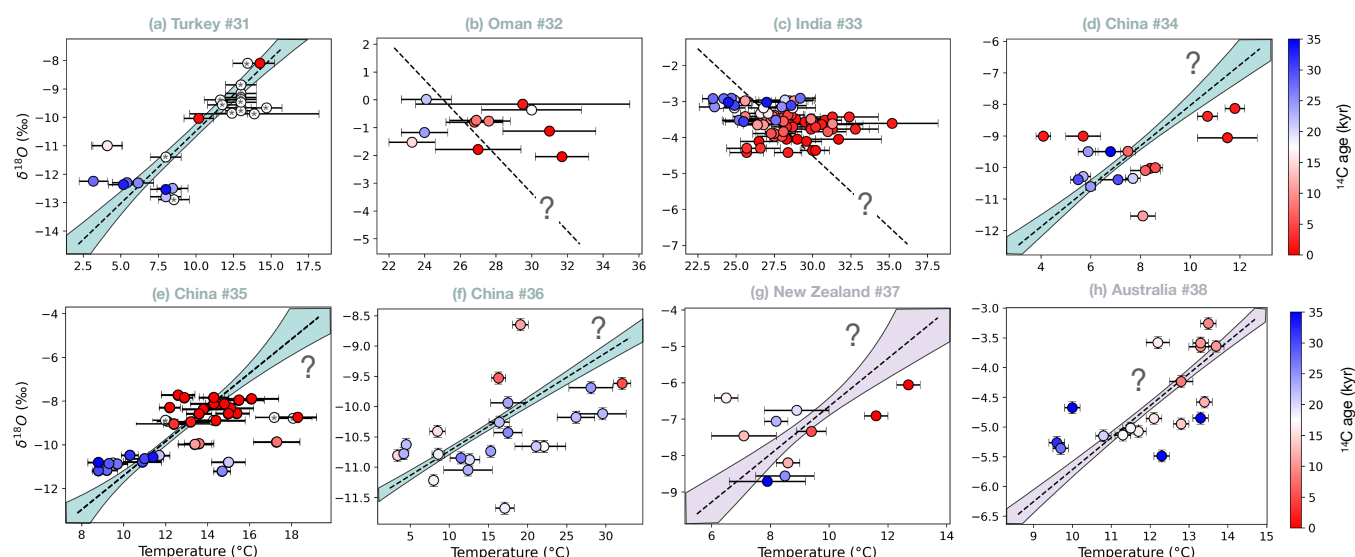


Figure 5. $\delta^{18}\text{O}$ –Temperature (NGT) reconstructions from fossil groundwaters in Asia, New Zealand and Australia. $\delta^{18}\text{O}$ –Temperature (NGT) reconstructions from aquifers in New Zealand (g) and Brazil (h) are also provided. Note that three different datasets from China are presented in panels (d–f). Marker colours refer to ^{14}C ages, with blue shades representing the last glacial period. Data without available ^{14}C ages are marked with an asterisk. As described in Section 2.2, given the inherent challenges of using ^{14}C in dissolved inorganic carbon for absolute groundwater dating (Han and Plummer, 2016), ^{14}C ages should be considered indicative only as these were taken directly from the original studies and, therefore, not uniformly processed/calibrated. Numbers given as # correspond to site locations given on Figure 1. Dashed lines and associated envelopes represent error-weighted least squares fitting through the data using “model 1” by York et al. (2004) implemented in Isoplot-R (Vermeesch, 2018). Interestingly, removing old (^{14}C -depleted, in blue) and “noisy” datapoints for panel (h) would decrease the MSWD of the correlation from 17 to 9, and cause a slight change in the reported correlation, from $\delta^{18}\text{O} = (0.54 \pm 0.06) \times \text{NGT} - (11.1 \pm 0.7)$ (see data compilation; Bekaert et al., 2026) to $\delta^{18}\text{O} = (0.65 \pm 0.08) \times \text{NGT} - (12.5 \pm 1.0)$. Interrogation mark in panels (e), (f), and (g) mean no linear correlation was found. Reported uncertainties are 1 SD.

proxy archives (e.g., ice cores) are often interpreted as reflecting past changes in atmospheric moisture source conditions (e.g., Vimeux et al., 1999; Jouzel et al., 2007), including past temperature and relative humidity (Pfahl and Sodemann, 2014). The lack of significant correlation between D-excess and NGT or age appears consistent with a stable moisture source over the LGIT (Fig. 10).

During the LGM, global oceanic circulation (Rahmstorf, 2002) as well as global and regional atmospheric circulation patterns (e.g., Laîné et al., 2009; Löffverström et al., 2014) were different from today. This time period was also characterized by colder and drier conditions (less precipitation is expected under glacial conditions because there is less moisture in the air), as well as significantly lower sea levels (–130 m relative to present-day, e.g., Huybrechts, 2002). Shifts in atmospheric circulation patterns since the LGM have been suggested in many regions (e.g., Asia) by both paleoclimatic data (e.g., Tibetan plateau; Zhu et al., 2015) and numerical simulations (e.g., Yanase and Abe-Ouchi, 2007). These atmospheric circulation changes imply potential changes in both (i) moisture sources (although our data suggest no significant shift in D-excess recorded in fossil groundwater (Figure 6–7) and (ii) hydrological processes along trajectories (in particular changes in convective activity). Both may lead to significant alterations in WSI composition of precipitation, independently from any change in regional temperature.

Additionally, topographical changes associated with the evolution of the Laurentide Ice Sheet in North America are thought to have affected global atmospheric circulation patterns, including a northward shift of North Pacific storm tracks (e.g., Cook and Held, 1988; Amaya et al., 2022) with consequences for regional climates in Central Asia and Europe (e.g., Löffverström et al., 2014). The pronounced effect of the Laurentide Ice Sheet on atmospheric circulation patterns above North America may be one source of the observed $\delta^{18}\text{O}$ –MAST disruptions in this region (e.g., Figure 2e). For Europe, the generally good $\delta^{18}\text{O}$ –MAST correlations suggest that, despite the presence of the British and Fennoscandian ice sheets, potential changes in global atmospheric patterns had only a secondary influence on the WSI composition of precipitation.

4.1.1.2 Atmospheric processes governing the WSI composition of atmospheric water vapor. Many atmospheric processes govern the WSI composition of atmospheric water vapor (e.g., evaporation from the sea surface, shallow and deep convection, large-scale advection and mixing, detrainment, cloud condensation, partial rain re-evaporation, continental recycling). A comprehensive review of these processes is beyond the scope of this study; readers are therefore referred to existing syntheses in the literature (e.g., Galewsky et al., 2016). Their relative influence varies across a range of spatial scales and vastly depends on regional

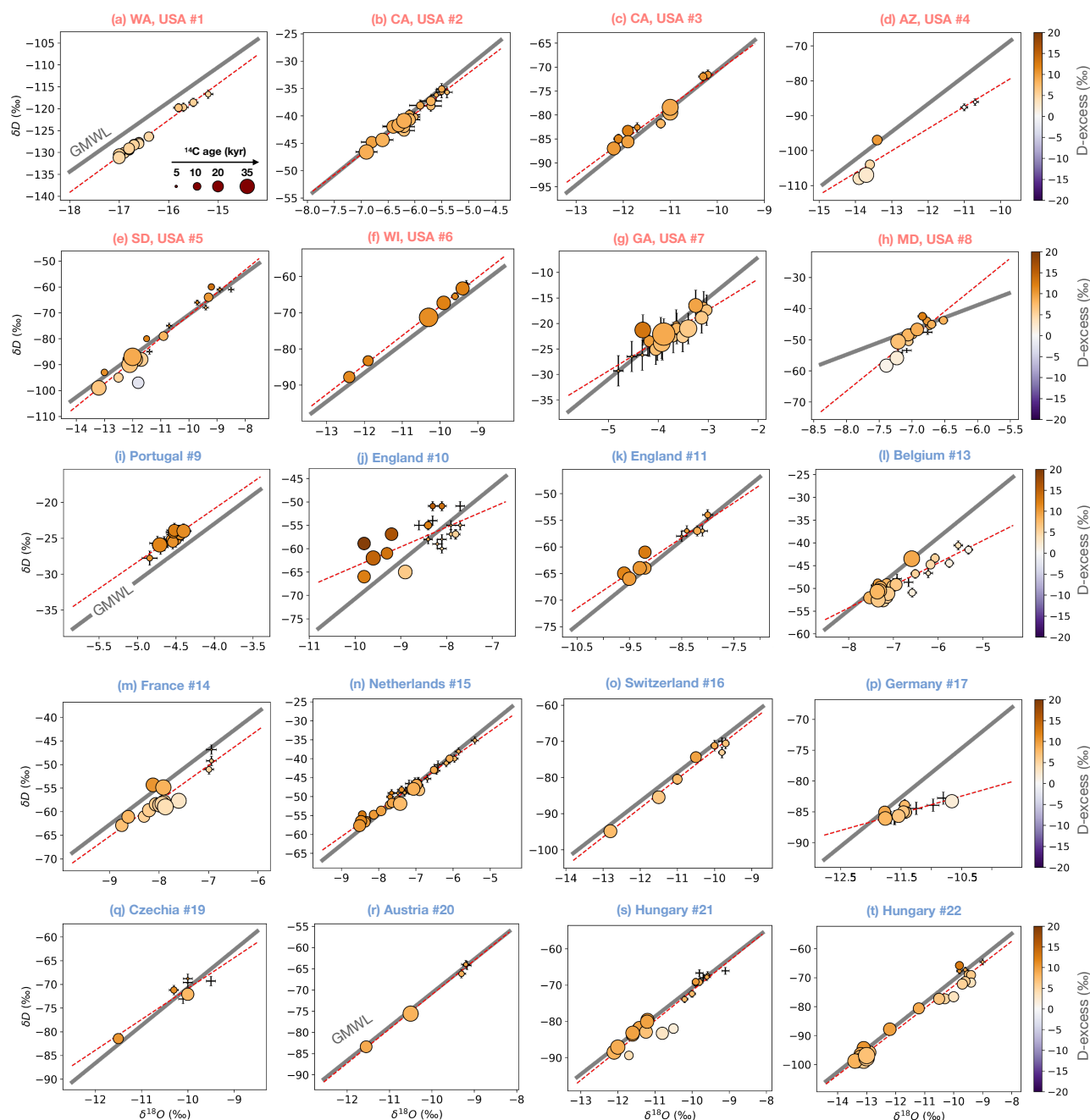


Figure 6. $\delta^{18}\text{O}$ – $\delta^2\text{H}$ data from USA and Europe fossil groundwaters. Marker colors refer to D-excess, representing deviations from reference slope ($= \delta^2\text{H} - 7.96 \times \delta^{18}\text{O}$; Araguás-Araguás et al., 2000). The Global Meteoric Water Line (GMWL, given by $\delta^2\text{H} = 7.96 \times \delta^{18}\text{O} + 8.86$; Araguás-Araguás et al., 2000) is shown in grey. The dashed red lines represent the best fits through the data. As shown in panel (a), marker size is proportional to ^{14}C age, which broadly span the past 35 kyr. As described in Section 2.2, given the inherent challenges of using ^{14}C in dissolved inorganic carbon for absolute groundwater dating (Han and Plummer, 2016), ^{14}C ages should be considered indicative only as these were taken directly from the original studies and, therefore, not uniformly processed/calibrated. Numbers given as # correspond to site locations given on Figure 1.

atmospheric and surface conditions. As these processes do not solely or linearly depend on temperature, potential changes in their relative contributions to controlling the WSI of water vapor could represent a significant source of $\delta^{18}\text{O}$ – $\delta^2\text{H}$ -NGT disruptions over the LGIT.

In the tropics, for example, vertical transport processes dominate over horizontal movement, and so precipitation primarily results from deep convection and large-scale subsid-

ence. The historical finding of empirical correlation between high monthly precipitation amounts and low monthly $\delta^{18}\text{O}$ at many tropical stations (Dansgaard, 1964) paved the way for utilizing tropical $\delta^{18}\text{O}$ as a proxy for local “precipitation amount”, in both modern and paleoclimate studies. Following this relationship, increased or decreased paleo-precipitation has been inferred from lower or higher paleo- $\delta^{18}\text{O}$ values, respectively (e.g., Wang et al., 2001; Schefuß

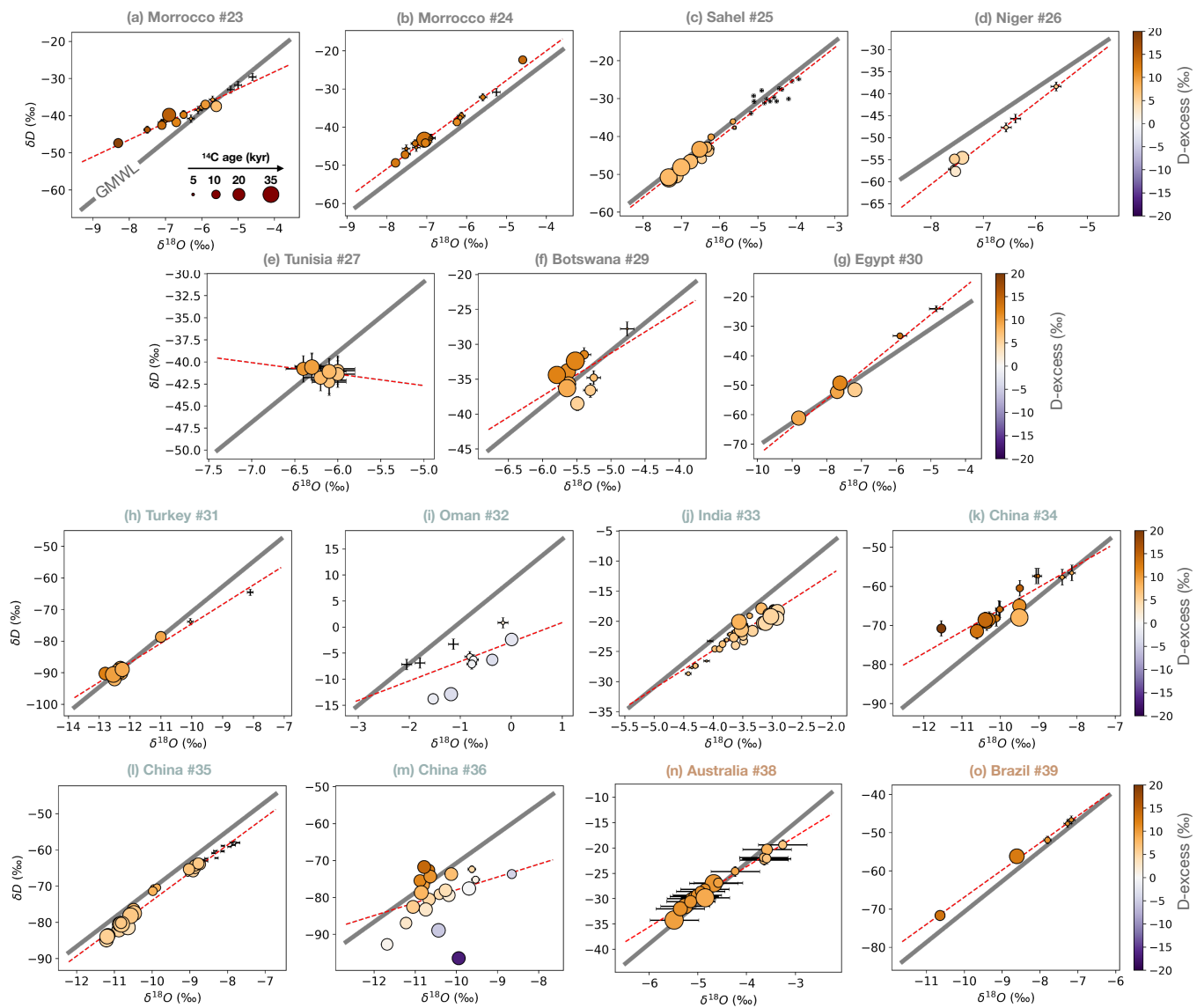


Figure 7. $\delta^{18}\text{O}$ – $\delta^2\text{H}$ data from fossil groundwaters in Africa, Asia, Australia and Brazil. Marker colors refer to D-excess, representing deviations from reference slope ($= \delta^2\text{H} - 7.96 \times \delta^{18}\text{O}$; Araguás-Araguás et al., 2000). The Global Meteoric Water Line (GMWL, given by $\delta^2\text{H} = 7.96 \times \delta^{18}\text{O} + 8.86$; Araguás-Araguás et al., 2000) is shown in grey. The dashed red lines represent the best fits through the data. As shown in panel (a), marker size is proportional to ^{14}C age, which broadly span the past 35 kyr. As described in Section 2.2, given the inherent challenges of using ^{14}C in dissolved inorganic carbon for absolute groundwater dating (Han and Plummer, 2016), ^{14}C ages should be considered indicative only as these were taken directly from the original studies and, therefore, not uniformly processed/calibrated. Numbers given as # correspond to site locations given on Figure 1.

et al., 2005; Wang et al., 2017). However, as discussed by Konecky et al. (2019), this approach yields variable agreement with qualitative inferences of wetter versus drier conditions derived from independent moisture-sensitive or from modern-day observations.

At higher latitudes, horizontal atmospheric processes become more prevalent in governing the WSI composition of water vapor and precipitation. In such regions, air masses migrating from the oceans to the continents preferentially lose heavy water stable isotopes by precipitation, leading to a progressive relative heavy isotope depletion of the remaining vapor reservoir following Rayleigh distillation theory, whereby:

$$R_{\text{vap}} = R_0 \times f^{\alpha_{\text{eq}}(T)-1} \quad (3)$$

with R_{vap} the isotope ratio of the vapor phase, R_0 the initial isotope ratio of the vapor phase (after evaporation above the ocean), f the fraction of water remaining in the vapor phase ($f = 1$ before condensation starts, $f = 0$ when condensation is complete), and $\alpha_{\text{eq}}(T)$ is the temperature-dependent equilibrium isotope fractionation factor (Majoube, 1971). The temperature-dependent increase in cumulative precipitation amount results in lower $\delta^{18}\text{O}$ – $\delta^2\text{H}$ of water vapor (Field et al., 2010; Krklec and Domínguez-Villar, 2014; Krklec et al., 2018), producing lower $\delta^{18}\text{O}$ of subsequent precipitation (Araguás-Araguás et al., 2000; Risi et al., 2008a; Breitenbach et al., 2010). As air masses move in-

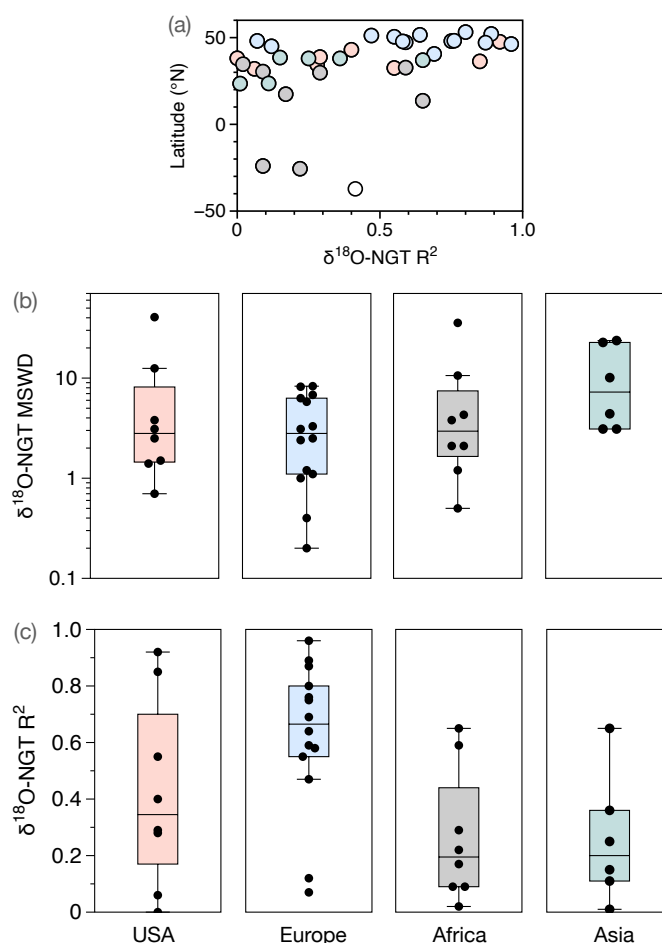


Figure 8. Box plot comparing $\delta^{18}\text{O}$ –Temperature (NGT) correlation coefficients (R^2) and MSWD (shown using a log₁₀ scale) from fossil groundwater worldwide. Boxplots show interquartile ranges. European correlations are generally more robust than correlations from the USA, Africa and Asia. The white circle in panel (a) corresponds to the dataset from Australia (#38). Most data also suggest a linear correlation between R^2 and latitude.

land, evapotranspiration also contribute substantial amounts of water vapor through a process known as continental recycling, which enriches the overlying water vapor in heavy isotopes along cloud trajectories (e.g., Eltahir and Bras, 1996; Trenberth, 1999; Burde and Zangvil, 2001; Bisselink and Dolman, 2008; Risi et al., 2013; Shi et al., 2022). This recycled moisture partly offsets the relative heavy isotope depletion caused by Rayleigh distillation during transport (Winnick et al., 2014). Similar to temperature-driven effects, the strength of the continental effect can vary depending on the relative roles of eddy diffusion and large-scale advection in transporting water vapor (Winnick et al., 2014). The fraction of local precipitation originating from continental recycling thus increases with continentality (i.e., distance from the coastline; van der Ent et al., 2010). In Eurasia, for instance, where the main moisture flux is mostly westerly, this fraction increases from ~30 % in Western Europe up to 90–100 % over Siberia, Mongolia, and China (e.g., Bosilovich, 2002; Yoshimura et al., 2004; Dirmeyer et al., 2009). The heavy isotope enrichment (i.e., higher δD – $\delta^{18}\text{O}$) of the water vapor phase during continental recycling may be attenuated in case of significant bare soil evaporation (Section 4.1.1.3). In general, however, fractionation processes at the land surface only have a secondary

effect on precipitation isotopic composition compared to that of continental recycling (Haese et al., 2013; Risi et al., 2016).

4.1.1.3 Evaporation and soil water evolution. Bare soil evaporation is known to cause significant heavy WSI enrichment (i.e., higher δD – $\delta^{18}\text{O}$) of residual soil water due to equilibrium and kinetic fractionation (lighter water molecules are favored during evaporation due to their higher diffusivity, so the remaining water is progressively enriched in ^{18}O and D by distillation, e.g., Mathieu and Bariac, 1996; Braud et al., 2009). Bare soil evaporation is a local effect that mostly affects the uppermost part of the soil column in temperate regions (i.e., a few tens of centimeters maximum; e.g., Horita et al., 2008). In Mediterranean and arid regions, the impact of bare soil evaporation is observed to greater depths of up 0.5 m and 3 m, respectively (e.g., Sprenger et al., 2016). Since the WSI composition of groundwater is not directly inherited from precipitation but from soil water that can have undergone variable extents of bare soil evaporation, there is potential for soil water residence time heterogeneities and mixing to cause significant WSI variations across the soil column (Wackerbarth et al., 2012; Genty et al., 2014; Zhang et al., 2023). However, in most cases, seasonal deviations from the annual weighted

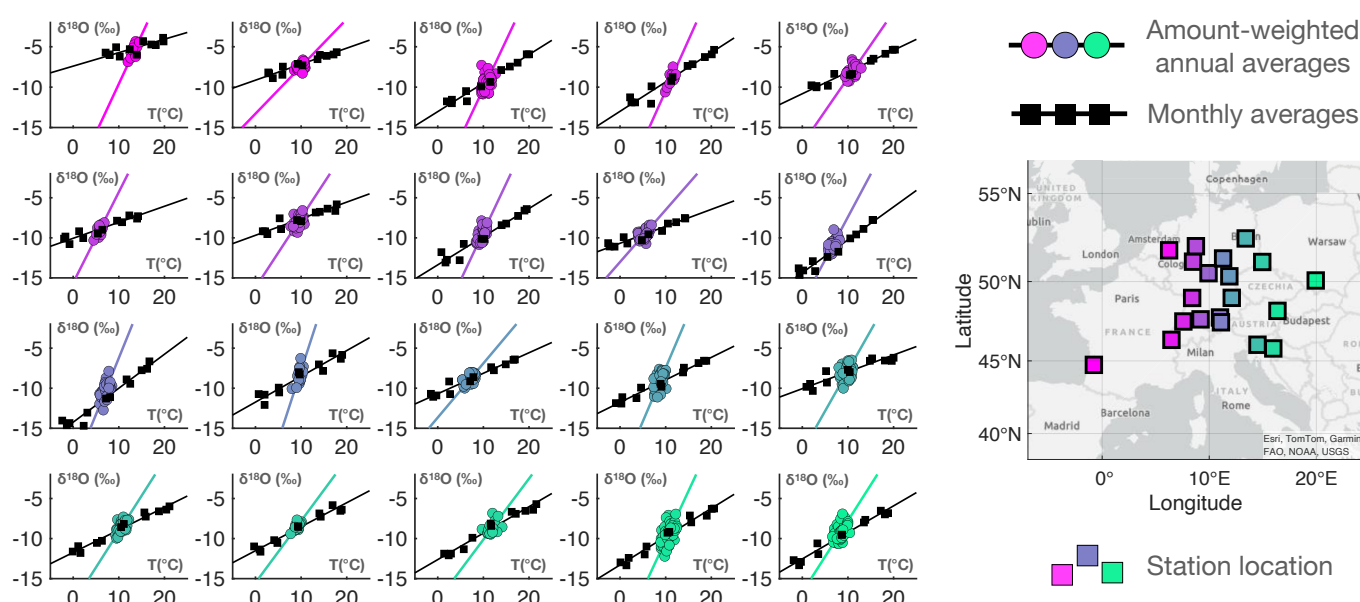


Figure 9. Comparison of average $\delta^{18}\text{O}$ –surface air temperature correlations derived from monthly mean data (black squares) and interannual amount-weighted $\delta^{18}\text{O}$ –surface air temperature correlations for 20 meteorological stations across Europe (longitude-based color coding). The slopes derived from monthly mean data are consistently smaller than those obtained from interannual averages. Note, however, that monthly means incorporate both the seasonal cycle and an interannual component arising from the integration of mean $\delta^{18}\text{O}$ –temperature relationships for a given month over multiple years. Because interannual variations produce larger $\delta^{18}\text{O}$ changes relative to temperature (as illustrated by the steep slopes of interannual $\delta^{18}\text{O}$ –temperature relationships), monthly means do not represent a purely seasonal signal.

average $\delta^{18}\text{O}$ of precipitation are exponentially dampened with depth (Allison et al., 1983; Yonge et al., 1985; Williams and Fowler, 2002; Riechelmann et al., 2017). The vertical and seasonal variations of soil water $\delta^{18}\text{O}$ critically depend on how precipitation water infiltrates the soil (Gaziz and Feng, 2004). For instance, groundwater recharge in arid regions commonly takes place by rapid, deep and focused infiltration during intermittent flood events (e.g., Scanlon et al., 2006), which minimize near-surface evaporation and enables preservation of (paleo) climate signals (e.g., Yokochi et al., 2019; Ram et al., 2024). While most other processes controlling the fate of surface water (e.g., runoff, percolation through soil, storage as soil water or groundwater, root water uptake, and transpiration) do not induce significant WSI variation, they can ultimately affect soil water and groundwater compositions if infiltration (i.e., = precipitation – evapotranspiration – runoff) varies seasonally. However, the combined effects of multiple soil water infiltration routes and flow dynamics (mixing, residence time) are specific to each hydrological system, and it is beyond the scope of this study to explore the isotopic consequences of vadose zone processes for each location.

Here, we consider that the WSI composition of soil water recharging as groundwater at depth mostly represents the average annual (or multi-year) composition of precipitation (which in reality may include snowfall, rainfall, and snowmelt). This is a reasonable assumption for temperate regions because they are characterized by rapid precipitation infiltration and minimal exposure to evaporation processes in the subsurface. The assumption that, in many cases,

subsurface processes only marginally affect the water stable isotope composition of deep soil water is supported by (i) the close correspondence between the $\delta^{18}\text{O}$ of late Holocene groundwater and modern regional surface waters and precipitation (e.g., Jasechko et al., 2015), as well as (ii) direct comparisons between the isotopic composition of precipitation and cave drip water in Villars cave based on 20 years' monitoring in southern France (Genty et al., 2014; Zhang et al., 2023). Pronounced kinetic fractionation during bare soil evaporation tends to produce higher D-excess in the evolving vapor, while lowering D-excess in the residual fluid; under sufficiently strong kinetic effects, the residual fluid can even develop negative D-excess values (Mathieu and Bariac, 1996; Risi et al., 2016; Graf et al., 2019). Here, the apparent stability of D-excess recorded in fossil groundwater is therefore also most consistent with minimal isotopic effects of potential changes in bare soil evaporation during deglacial warming. Therefore, we consider that, in most instances, changes in groundwater WSI faithfully represent changes in local precipitation WSI.

4.1.1.4 Altitudinal control on the WSI composition of precipitation. Precipitation typically forms in clouds that are far above the surface (typically within the 500 m to 3000 m – potentially up to 5000 m – elevation range). The temperature at which condensation occurs in the cloud primarily depends on the altitude of the cloud base, which is mainly determined by the near-surface relative humidity. The altitudinal thermal gradient (lapse rate) in the sub-cloud layer is typically close to the dry adiabatic lapse rate (i.e., $\sim 9.8^\circ\text{C km}^{-1}$; e.g., Barry and Chorley, 2009) rather than a

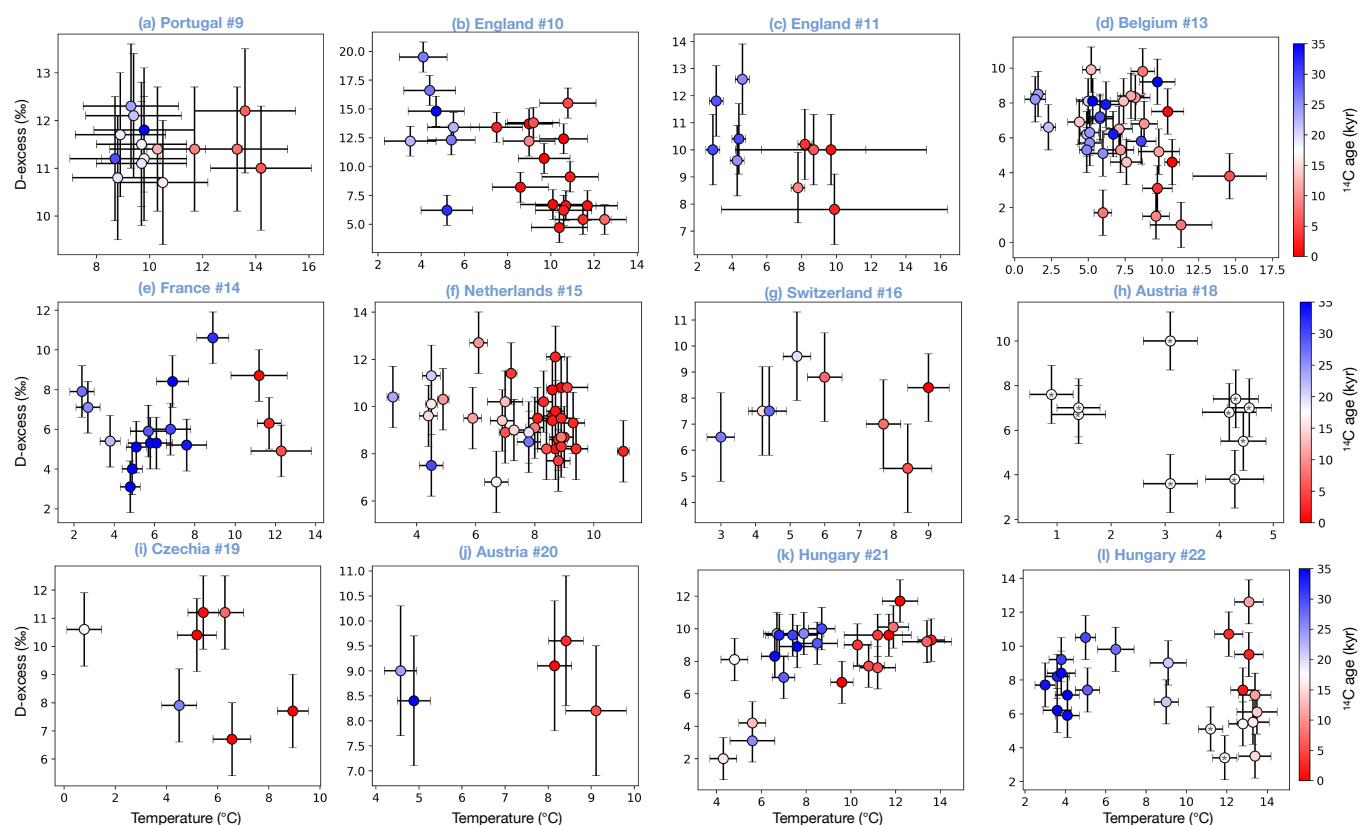


Figure 10. D-excess–Temperature (NGT) reconstructions from European fossil groundwaters. Marker colors refer to ^{14}C ages, with blue shades representing the last glacial period. Data without available ^{14}C ages are marked with an asterisk. As described in Section 2.2, given the inherent challenges of using ^{14}C in dissolved inorganic carbon for absolute groundwater dating (Han and Plummer, 2016), ^{14}C ages should be considered indicative only as these were taken directly from the original studies and, therefore, not uniformly processed/calibrated. Numbers given as # correspond to site locations given on Figure 1. Reported uncertainties are 1 SD. Similar plots for USA, African and Asia regions are provided as part of the [Supplementary Material](#) (Figs. S1–S3).

moist adiabat. How this lapse rate evolved during deglacial warming remains poorly understood, but several climatic factors are thought to have influenced atmospheric lapse rates in the past (e.g., seasonality or moisture; Qixiang et al., 2018; Legrain et al., 2023, respectively, for free tropospheric lapse rates). Various complex phenomena linked to the history of rainfall from the cloud base are also thought to potentially affect the WSI composition of precipitation reaching the ground (e.g., Dansgaard, 1964; Stewart, 1975; Gat, 1996; Lachniet, 2009; Pang et al., 2011). However, due to rapid isotopic re-equilibration between water vapor and surface liquid water, the WSI composition of precipitation is mostly controlled by the vapor composition of the atmospheric boundary layer below the cloud base. Risi et al. (2010) compared the degree of equilibration between vapor and precipitation at a few meteorological stations (Vienna, Manaus, and Ankara, from 1979 to 2007) and found that low-level vapor isotopic compositions were well reproduced by predictions for a liquid in equilibrium with the local vapor. This is consistent with recent findings by Graf et al. (2019), who emphasized the prominent role of below-cloud processes in modulating isotope signals in surface rain. Thus, past changes in lapse rate (e.g., Legrain et al., 2023) should not significantly perturb WSI records in fossil groundwater. One potential

issue is post-condensation processes such as rain evaporation (e.g. Risi et al., 2020), which can have a large impact on the precipitation composition, especially in arid regions (e.g., Lee and Fung, 2008; Zhu et al., 2021). For the sake of simplicity, we will consider here that rain-vapor equilibrium is the dominant process controlling the ultimate composition of precipitation (Deshpande et al., 2010; Tremoy et al., 2014; Conroy et al., 2016; Vimeux and Risi, 2021).

4.1.2 Noble gas systematics

4.1.2.1 “Excess air” and degassing. Bubble entrainment and partial dissolution during recharge often bring “additional” dissolved noble gases (referred to as “excess air” and often represented by neon supersaturation, “ ΔNe ”), which must be accounted for using an inverse model of Ne, Ar, Kr, and Xe concentrations to compute accurate NGTs (Aeschbach-Hertig and Solomon, 2013). Although the exact mechanisms of bubble dissolution and groundwater excess air generation remain uncertain, ΔNe has been proposed as a potential qualitative tracer of short-term (e.g., seasonal) water table depth fluctuations during recharge (Ingram et al., 2007). Several models exist in the literature for the translation of noble gas concentrations into recharge temperatures (NGTs), with different physical interpretations

for the excess air input leading to variable NGT reconstructions depending on the chosen model. Seltzer et al. (2021) evaluated leading noble gas models of excess air correction by comparing Late Holocene NGTs to ERA5-Land 1981–2019 MASTs. They found a negligible offset between these two datasets when using the closed-system equilibration (CE) model (Aeschbach-Hertig et al., 2000), providing strong support for the accuracy of this model to depict noble gas dissolution in most hydrological systems. This model was then applied for a uniform re-evaluation of previously published NGTs (see Methods, Section 2.2), providing confidence that the reconstructed NGTs were not biased by inappropriate excess air correction.

Degassing processes (within the aquifers or during sampling) may also artificially lower dissolved noble gas concentrations and generate anomalously high noble gas temperature signals. This potential bias has been observed and addressed in several publications (e.g., Wang et al., 2015). For example, in the Ledo-Paniselian Aquifer (Belgium), Walraevens et al. (2020) showed that the geochemical compositions of several groundwater samples recharged shortly after the LGM are compatible with specific methanogenic recharge conditions associated with thawing permafrost. The samples upstream from (i.e., younger than) the permafrost-related recharge gap show evidence of both degassing (anomalously high NGT) and evaporation ($\delta^{18}\text{O}$ and $\delta^2\text{H}$ systematics), resulting in stark deviations from the Global Meteoric Water Line (GMWL; Walraevens et al., 2020). Importantly, the vast majority of the groundwater samples considered here do not seem to have experienced such conditions of thawing permafrost (Figures 6–7). Note that, the CE model is capable to describe degassed groundwater samples, enabling the estimation of NGTs also in such cases, although degassing adds an additional uncertainty (Aeschbach-Hertig et al., 2008).

4.1.2.2 Mixing and re-equilibration. An additional complication can arise from mixing between distinct water bodies (e.g., Visser et al., 2022), which can have different salinities (e.g., brines) that affect dissolved noble gas concentrations and paleotemperature estimates (e.g., Gerber et al., 2017). Undesirable mixing between groundwater bodies (e.g., mixing between fresh and saline water, or between young and fossil groundwater) can be driven by several hydrogeological conditions, including loss or thinning of confining layers, changes in preferential flow paths in karst systems, changes in hydraulic gradients (e.g., due to pumping or irrigation), or occurrence of faults and fractures acting as conduits between aquifers that are otherwise hydraulically separated. Complex hydrological systems like the Pahute Mesa (Nevada, USA) also show evidence of decoupling from ambient conditions during recharge due to noble gas temperature (NGT) resetting, as indicated by warm (modern-day-like) NGTs associated with cold (late Pleistocene-like) $\delta^{18}\text{O}$ signatures. This can occur if portions of the fossil groundwater flow path become (at least partially) unconfined and allow for gas exchange with soil air connected to the surface. The studies for which

potential issues related to mixing or re-equilibration have been suggested by the authors (e.g., karstic regions) were not included in this compilation to avoid misinterpretation of paleoclimate signals.

4.1.2.3 Snow cover and vegetation changes. Because seasonal temperature variations diminish with soil depth, noble gas temperatures recorded at the water table tend to closely reflect the mean annual ground surface temperature (MAST; Aeschbach-Hertig and Solomon, 2013). In low-latitude, snow-free regions, the difference between the water table temperature and the mean annual air temperature (MAAT) is typically lower than $\sim 0.2^\circ\text{C}$ (e.g., Cey, 2009). At higher latitudes, however, decoupling of MAAT and MAST can occur due to variations in snow cover or vegetation, which may create discrepancies between the temperature controlling dissolved noble gas concentrations during recharge (MAST) and that controlling the WSI composition of local precipitation (MAAT). For example, Stute and Sonntag (1992) showed that Central Europe groundwater formed in forests exhibit noble gas temperatures that are 2.2°C lower than the groundwater formed in fields or meadows. In contrast, direct solar heating of the bare land surface, a common feature in arid regions, can lead to a few $^\circ\text{C}$ higher MAST than MAAT (e.g., Aeschbach-Hertig and Solomon, 2013). Quantifying potential effects of snow or vegetation cover changes on NGT reconstructions is often extremely challenging and not considered. Although this may not be correct for regions near boundaries of snow, permafrost and vegetation zones, the usual assumption is therefore that any potential discrepancies between MAST and MAAT were either preserved during deglacial warming or were relatively minor compared to the magnitude of observed temperature changes.

Broad global patterns of vegetation change from the last glacial period (grassland) to the Holocene (forest) could have increased continental recycling and thus potentially increased the $\delta^{18}\text{O}$ of water vapor (Prentice et al., 2000; Harrison and Prentice, 2003; Bowen et al., 2019). In addition, the transition from grassland to forest could have reduced the contribution of bare soil evaporation to continental recycling and thus slightly increased the $\delta^{18}\text{O}$ of recycled moisture. However, this process is unlikely to have significantly affected the composition of water vapor above continental masses or groundwater (see Section 4.1.1.3).

4.1.2.4 Seasonality. In monsoonal mountain regions, for example, groundwater NGTs may reflect soil temperatures over the ~ 3 month monsoon period rather than the MAST (Müller et al., 2024). This implies that NGT differences between the late Holocene and the LGM in regions with distinct wet and dry seasons may be attributable to shifts in precipitation regime rather than changes in air temperature. Seasonality in NGTs has previously been observed in several mountainous fractured rock aquifers, including cold regions where snowmelt-driven recharge can lead to NGTs below mean annual air temperature (e.g., Manning and Solomon, 2003; Masbruch et al., 2012), and tropical volcanic islands where fast recharge through fractured basalt preserves wet-

season temperature signals (Warrier et al., 2012; Niu et al., 2017).

Changes in seasonality during the LGM compared to the late Holocene, both in terms of temperature and precipitation amounts, may significantly affect the slopes of the $\delta^{18}\text{O}$ –NGT relationships (Caley et al., 2014; Ford et al., 2015). For instance, a lower contribution of winter precipitation to annual totals during the LGM would bias the weighted annual mean $\delta^{18}\text{O}$ of fossil groundwaters towards the higher $\delta^{18}\text{O}$ of summer precipitation, thus leading to weaker $\delta^{18}\text{O}$ –NGT slopes. Conversely, an enhanced contribution of winter precipitation during the LGM may yield to greater $\delta^{18}\text{O}$ –NGT slopes. However, our understanding of potential changes in seasonality – both in terms of temperature and precipitation amounts – since the LGM remains limited, which hinders robust correction of observed $\delta^{18}\text{O}$ –NGT slopes for potential seasonality changes.

4.2 Paleoclimate information from fossil groundwater data

4.2.1 *The tropics and Central Asia: a lack of significant correlation*

The lack of significant WSI–T correlations in fossil groundwater from regions such as the tropics and Central Asia provides conclusive evidence that isotope-based paleotemperature reconstructions are not applicable in these regions of the globe. The multiplicity of processes that can disturb the relationship between WSI and NGT recorded in fossil groundwater (Section 4.1) makes it complicated to simply attribute the lack of significant correlation to one or several processes specifically. However, there exists a large corpus of articles showing that, in tropical regions, $\delta^{18}\text{O}$ is negatively correlated to the precipitation amount (Partin et al., 2007). This has been demonstrated for Asia (e.g., Wang et al., 2001; Johnson and Ingram, 2004; Pausata et al., 2011; Lee et al., 2012; He et al., 2015), South America (Cruz et al., 2005; Vimeux et al., 2005; Cheng et al., 2012) and Africa (Risi et al., 2008b; Collins et al., 2013). Some regions, such as China, have climates influenced by both winter and summer monsoon systems as well as mid-latitude westerlies, and their relative importance changes by season and region. In general, our results are consistent with the observation that in many tropical regions, WSI–NGT relationships collapse, and that $\delta^{18}\text{O}$ should rather be interpreted in terms of precipitation amount in these regions.

4.2.2 *Modelling of $\delta^{18}\text{O}$ –T correlations with Rayleigh distillation*

For a given region, potential changes in moisture sources or in hydrological processes along air mass trajectories could directly impact $\delta^{18}\text{O}$ and $\delta^2\text{H}$ without necessarily reflecting a temperature change, e.g. atmospheric convection (e.g., Vimeux et al., 2005) or continental recycling (e.g., Trenberth, 1999). The observation of robust $\delta^{18}\text{O}$ –T correlations, where they occur, implies that, in many regions, most potential sources of WSI–noble gas disruption were not sufficient to significantly alter the correlation over the LGIT. Below, we consider the case of Western Europe,

where the most robust WSI–NGT correlations are observed. Hydroclimates in this region are vastly influenced by the westerlies over the North Atlantic, which represent the primary source of moisture transport to Europe, especially in winter (e.g., Hu et al., 2022).

Over the past decades, several approaches have been developed to model WSI fractionation during ocean evaporation, differing mainly in how they represent kinetic effects and air–sea exchange. Formulations based on the so-called Craig–Gordon framework (Craig and Gordon, 1965; Gonfiantini et al., 2018) combine equilibrium fractionation at the water surface with kinetic fractionation during diffusion through the viscous sublayer, and account for ambient humidity and isotopic composition of atmospheric vapor (Merlivat and Jouzel, 1979). Because the water–air exchanges operate far from isotopic equilibrium and water transport occurs across multiple scales, multiple studies have emphasized the importance of coupling microscale fractionation processes with large-scale moisture transport (including continental water recycling; Koster et al., 1993) to explain isotope variations in atmospheric water vapor and precipitation.

As described in Section 4.1.1.2, one of the main processes controlling the evolution of precipitation WSI compositions over Europe is the cumulative, temperature-dependent evolution of water vapor isotope compositions via Rayleigh distillation. Theoretical slopes of the $\delta^{18}\text{O}$ –MAAT relationship, derived by considering an isolated air mass undergoing isobaric cooling with rain forming in isotopic equilibrium with the vapor between 0 and 20 °C, vary between 0.52 and 0.72 ‰ °C^{−1}, encompassing the observed interannual value of ~0.69 ‰ °C^{−1} (Dansgaard, 1964). For Western and Central Europe, the slopes of the $\delta^{18}\text{O}$ –NGT relationships are much smaller, around 0.19 ‰ °C^{−1} (e.g., 0.1826 ± 0.0260 for France, 0.195 ± 0.166 for Belgium, 0.1863 ± 0.0327 for Czechia; see data compilation, Bekaert et al., 2026). To explore the physical mechanisms leading to these values, we consider a simplified model of Rayleigh distillation (after Stern and Blisniuk, 2002) in continental Europe, ignoring factors such as mixing between different air masses, cloud water–vapor isotope exchange, sub-cloud droplet evaporation, and supersaturation. We assume that newly formed atmospheric moisture above the ocean is continuously transported away by atmospheric circulation and does not remain in equilibrium with the sea surface (open Rayleigh model). As discussed below, we include both equilibrium and kinetic fractionation under unsaturated conditions, making isotopic fractionation effectively sensitive to relative humidity (Merlivat and Jouzel, 1979).

We assume that changes in the WSI composition of European precipitation over the deglaciation were primarily driven by rising temperatures of both the source of moisture (ocean, “0”) and local sites (“s”) on the continent. For illustration purposes, we consider the NGT–WSI record derived by Bekaert et al. (2023) from the Paris Basin in France (panel (e), Figure 3; Figure 10). LGM and present-day temperatures on this local site (T_S) are set as the minimum and late Holocene NGT recorded in the

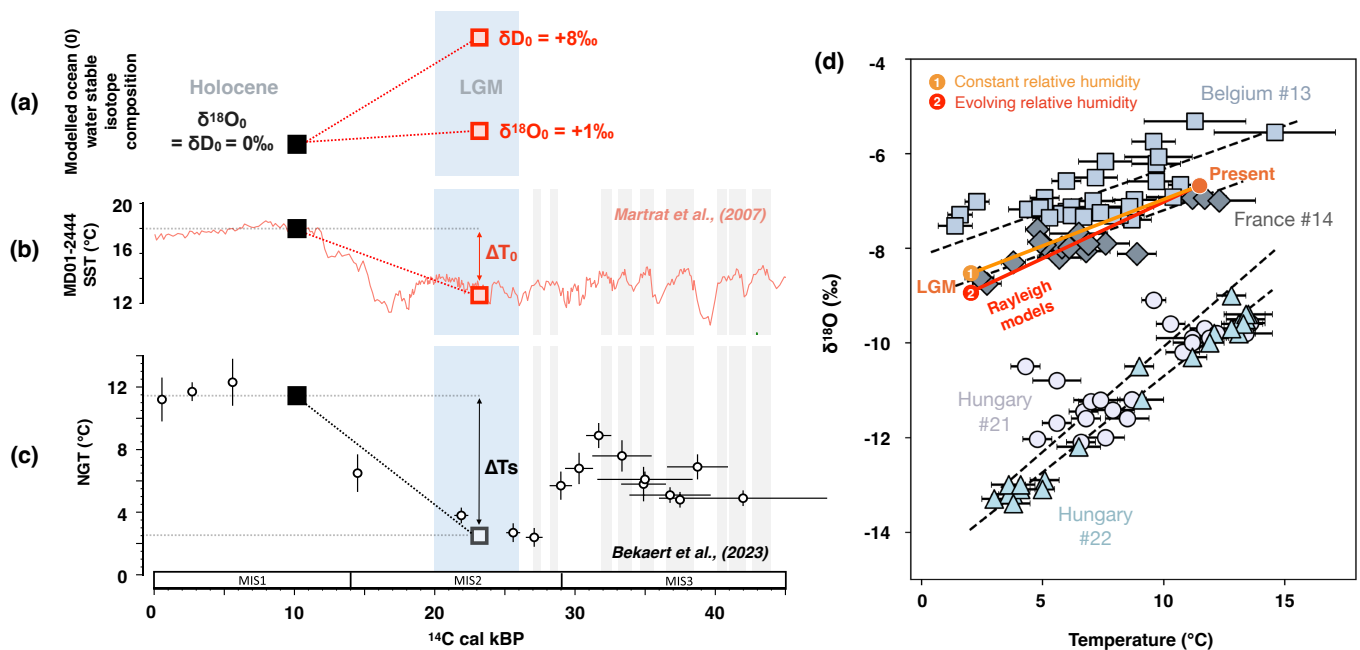


Figure 11. Parameterization of the Rayleigh distillation model (a-c) and comparison with the $\delta^{18}\text{O}$ –Temperature (NGT) reconstructions from fossil groundwaters in France (Bekaert et al., 2023, #14), Belgium (Blaser et al., 2010, #13) and Hungary (from the Great Hungarian Plain; Stute and Deak, 1989, #21) and Pannonian Basin (Varsányi et al., 2011, #22) (d). (a) LGM ocean water stable isotope data are from Schrag et al. (2002). Displayed Sea Surface Temperature (SST) from the MD01-2444 core of the Iberian margin (b) and Noble Gas Temperature (NGT) from the Eastern Paris Basin (c) are taken from Martrat et al. (2007) and Bekaert et al. (2023). The $\delta^{18}\text{O}$ –Temperature (NGT) correlation from the Albian aquifer in France ($y = (0.186 \pm 0.033)x + (-9.060 \pm 0.211)$; Bekaert et al., 2023) is compatible with expectations from Rayleigh distillation for scenarios of constant ($y = 0.1935x - 8.8366$ (1)) and evolving ($y = 0.2345x - 9.3118$ (2)); following Chadwick et al., 2016) relative humidity (d). Although the strong agreement among nearby locations reinforces the reliability of these reconstructions, additional processes than Rayleigh distillation are required to explain the data from more continental domains, such as Hungary. Numbers given as # correspond to site locations given on Figure 1. Reported uncertainties are 1 SD.

fossil groundwater, respectively (for simplicity, we consider MAAT = MAST). Over the deglaciation, the ocean temperature (T_0) is increased following the Sea Surface Temperature (SST) profile recorded in sediment core MD01-2444 off the coast of Portugal ($\Delta T_0 = 5^\circ\text{C}$ Martrat et al., 2007, Figure 10). Due to the intrinsic limitations of the ^{14}C dating approach (e.g., Geyh, 2000), the relative chronology of ocean versus land warming is unknown. We therefore assume that temperatures in both realms evolve synchronously over time. The WSI composition of the ocean is conjointly decreased from $\delta^{18}\text{O}_0 = +1\text{‰}$ and $\delta^2\text{H}_0 = 8\text{‰}$ during the glacial period (Schrag et al., 2002) to $\delta^{18}\text{O}_0 = \delta^2\text{H}_0 = 0\text{‰}$ in the Holocene (Figure 10). For simplicity, the larger LGM–Holocene $\delta^{18}\text{O}$ – $\delta^2\text{H}$ gap for the Mediterranean Sea (compared to the Atlantic Ocean, e.g., Grant et al., 2012) is not considered. We assume no significant change in the source of atmospheric moisture since the last glacial period, consistent with the observed absence of significant changes in D-excess (Figure 6).

As discussed in Section 4.1.1.1, ocean evaporation is strongly influenced by relative humidity because the humidity gradient between the saturated ocean surface and the overlying atmosphere controls both the magnitude of the evaporative flux and kinetic isotope fractionation during ocean evaporation. Under low relative humidity,

the kinetic contribution to isotope fractionation becomes more important, reflecting enhanced evaporation through the air boundary layer. As relative humidity increases, the kinetic contribution diminishes and isotope fractionation approaches equilibrium conditions. Here, we considered two end-member scenarios: (i) stable relative humidity above the ocean ($\sim 80\%$) and land ($\sim 70\%$) surfaces across the LGIT, with modern-like values supported by global hourly near-surface relative humidity since 1940 (ERA5; Hersbach et al., 2025) and assumptions from previous isotope-modelling studies (e.g., Gat, 1996; Gat, 2010). However, because relative humidity measures atmospheric moisture relative to the saturation capacity of the air, colder air during the LGM could have held much less water vapor than during the Holocene, implying that relative humidity could have been higher despite a lower absolute atmospheric moisture content. To model this, we considered (ii) evolving relative humidities following the formalism of Chadwick et al. (2016), where fractional changes in specific humidity over land are equal to fractional changes in specific humidity over ocean. The specific humidity over the continent, noted q_{land} , corresponds to the ratio of the mass of water vapor to the total mass of air (dry air and water vapor combined). By definition, $\text{RH}_{\text{land}}(\text{LGM}) \sim q_{\text{land}}(\text{LGM})/q_{\text{sat,land}}(\text{LGM})$, where $q_{\text{sat,land}}(\text{LGM})$ is the saturation specific humidity

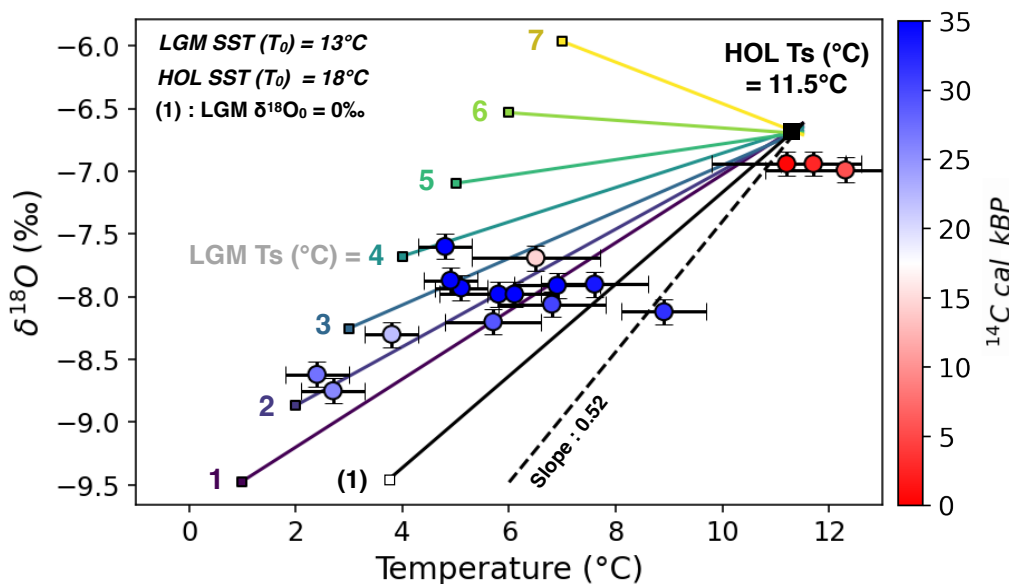


Figure 12. Comparison between outputs of the Rayleigh distillation model and $\delta^{18}\text{O}$ –Temperature (NGT) correlation from fossil groundwaters in France (Bekaert et al. 2023, #14 on Figure 1) (Bekaert et al., 2023, #14 on Figure 1). The colors of the symbols correspond to ^{14}C ages provided on the color bar. Lines 1 to 7 represent the different correlations expected for LGM $T_0 = 13^\circ\text{C}$, Holocene $T_0 = 18^\circ\text{C}$ (Martrat et al., 2007, Figure 10), Holocene $T_S = 11.5^\circ\text{C}$ and LGM T_S given in degree Celsius by the corresponding number (from LGM $T_S = 1^\circ\text{C}$ to 7°C). Line (1) represents the case where LGM $\delta^{18}\text{O}_0$ is set to 0 ‰ rather than +1 ‰, with an assumed LGM T_S of 1°C . Ocean and land relative humidities are assumed to evolve across the LGIT following the formalism of Chadwick et al. (2016, Figure 10). The black dashed line shows a slope of 0.52 for reference (Dansgaard, 1964). Reported uncertainties are 1 SD.

(i.e., the specific humidity of the atmosphere when it is saturated). By combining the fact that (i) $q_{\text{sat,land}}(\text{LGM})$ and $q_{\text{sat,ocean}}(\text{LGM})$ are directly prescribed by the model ocean and land temperature, and (ii) the relative humidity over the oceans (RH_{ocean}) changes by $0.54\% \text{ K}^{-1}$ of SST (Seltzer et al., 2023), our model computes the changes in q_{land} for every temperature step over the deglaciation by assuming that it decreases by the same fraction as q_{ocean} . This permits following the evolution of RH_{land} over the course of the LGIT, which we use to model the potential effect of bare soil evaporation on groundwater composition, following the formalism of Risi et al. (2016).

Our model considers that, at any point of the deglaciation history (i.e., from the LGM to the Holocene), atmospheric moisture forms above the Atlantic Ocean and progresses eastwards towards Europe. It undergoes isobaric cooling and contributes to European precipitation once saturation is reached. When the temperature reaches the T_S set by the NGT record, precipitation contributes to the recharge of the aquifer. Thus, both the WSI of “local precipitation” ($\delta^{18}\text{O}_S$ and $\delta^2\text{H}_S$) and “local temperature” are recorded in groundwater, at each step of the deglaciation.

The slope of the $\delta^{18}\text{O}$ – T correlation modelled with this approach is significantly smaller than the $0.72\% \text{ } ^\circ\text{C}^{-1}$ range previously derived by Dansgaard (1964), but consistent with reconstructions from groundwater data (Bekaert et al., 2023, Figure 12). Increasing the LGM T_S further decreases the slopes of the modelled $\delta^{18}\text{O}$ – T correlations, until negative correlations are generated for LGM $T_S \geq 6^\circ\text{C}$ (i.e., when ΔT_S becomes smaller than ΔT_0 ; Figure 12). The best fit to the data is obtained by assuming

a LGM $T_S \sim 2^\circ\text{C}$ (i.e., $\Delta T_S = 11.5 - 2 = 9.5^\circ\text{C}$), as derived from noble gas systematics ($y = 0.2345x - 9.3118$; Figure 12). Assuming LGM $\delta^{18}\text{O}_0 = +1\text{ ‰}$ with no change in relative humidity over the LGIT ($\text{RH}_{\text{ocean}} = 0.80$), we find that the initial $\delta^{18}\text{O}$ of ocean moisture decreases from -14.9 ‰ to -15.5 ‰ . In the case of evolving relative humidity (RH_{ocean} increases from 0.77 to 0.80), the initial $\delta^{18}\text{O}$ of ocean moisture moderately increases from -15.7 ‰ to -15.5 ‰ . These slight changes in the initial $\delta^{18}\text{O}$ of ocean moisture slightly affect the expectations from Rayleigh distillation, although both scenarios are consistent with the $\delta^{18}\text{O}$ –NGT correlation from the Albien aquifer in France (Figure 11). An experiment assuming LGM $\delta^{18}\text{O}_0 = 0\text{ ‰}$ (i.e., no change in the WSI composition of the source over the deglaciation) and LGM $T_S = 1^\circ\text{C}$ shows that temporal changes in the WSI of the source ($\delta^{18}\text{O}_0$ and $\delta^2\text{H}_0$) driven by global ice volume changes (Lambeck et al., 2014) further contribute to lowering the slopes of the modelled $\delta^{18}\text{O}$ – T correlations (see (1) on Figure 12). These considerations allow readily explaining why the slopes of the $\delta^{18}\text{O}$ – T correlations in fossil groundwater are usually smaller than expected from standard Rayleigh distillation predictions (Dansgaard, 1964). The prominent effect of varying T_0 in concert with local temperature T_S can be directly expressed from the Rayleigh equations. Assuming that only f varies in the Rayleigh distillation model (Eq. 3), one derives:

$$\frac{d[\delta^{18}\text{O}]}{dT} \propto \frac{d \ln(f)}{dT} \times \left(1 - \frac{dT_0}{dT_S}\right) \quad (4)$$

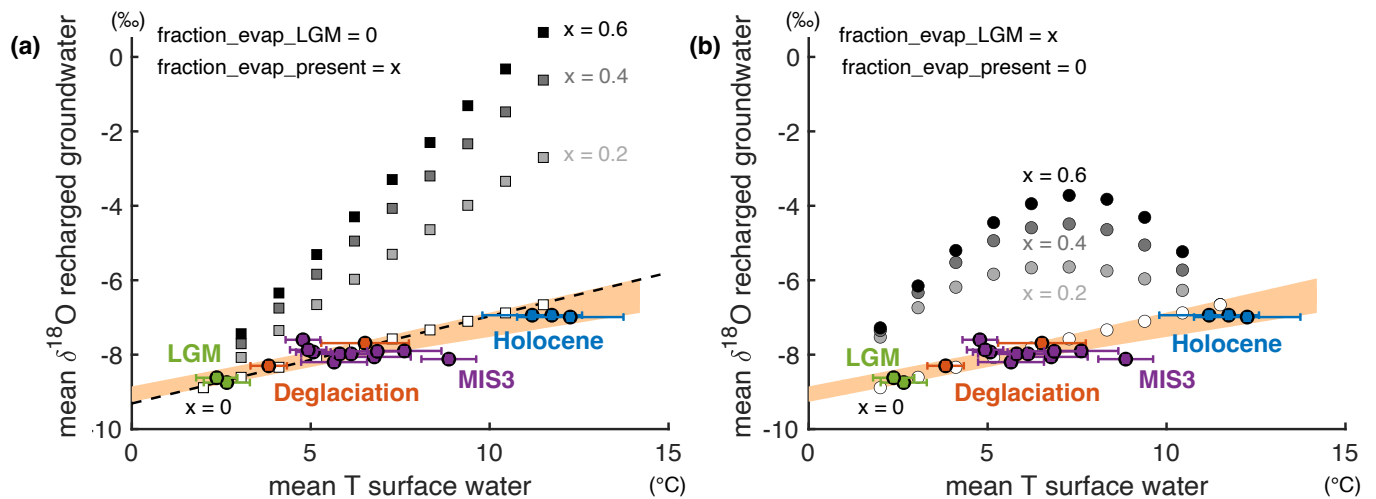


Figure 13. Simulated effect of bare soil evaporation on groundwater WSI composition and its impact on $\delta^{18}\text{O}$ –NGT correlations, compared to the $\delta^{18}\text{O}$ –Temperature (NGT) correlation from fossil groundwaters in France (Bekaert et al., 2023, #14 on Figure 1). The effect of bare soil evaporation was simulated using the parameterization of Risi et al. (2016), with `fraction_evap_LGM` and `fraction_evap_present` referring to the fraction of surface groundwater undergoing evaporation (i.e., $1 - \text{fraction_evap}$ contributes to groundwater recharge) at the LGM and present day, respectively. Increasing or decreasing `fraction_evap` over the LGIT produces distinct patterns of $\delta^{18}\text{O}$ –NGT co-evolution due to the competing controls of relative humidity (Chadwick et al., 2016) and `fraction_evap` over the LGIT. The large deviations from the $\delta^{18}\text{O}$ –NGT correlation observed in the Albien aquifer (France; Bekaert et al., 2023) (France; Bekaert et al., 2023) in scenarios involving significant present-day (a) or LGM (b) bare soil evaporation suggest that bare soil evaporation had only a minor, if any, impact on the WSI composition of fossil groundwater.

This indicates that, in the extreme case where T_0 and T_S vary in the same proportion, $\frac{d[\delta^{18}\text{O}]}{dT}$ becomes close to 0. When dT_S is greater than dT_0 (lines 1 to 5 on Figure 12), $\frac{dT_0}{dT_S} < 1$ and $\frac{d[\delta^{18}\text{O}]}{dT}$ (i.e., the slope of the $\delta^{18}\text{O}$ – T correlation) is positive. When dT_S is smaller than dT_0 (lines 6 and 7 on Figure 12), $\frac{dT_0}{dT_S} > 1$ and $\frac{d[\delta^{18}\text{O}]}{dT}$ is negative. These results support the concept that changes in the WSI composition of Western European precipitation over the deglaciation were primarily driven by rising temperatures of both the source of moisture (ocean) and local sites on the continent, with only minimal change in moisture origin (as independently constrained by the consistency of D-excess in most records; Figure 6). Additional processes than Rayleigh distillation may be required to explain the data from other regions of Europe. For example, the pronounced influence of a mediterranean climate in Southern Europe may explain the observation of a negative (rather than positive for Central and Northern Europe) WSI–NGT correlation from fossil groundwater in Portugal (Figure 3a). In more continental domains, such as Hungary (Figure 11d), other processes such as continental recycling may become prevalent, including heavy isotope enrichment (i.e., higher δD – $\delta^{18}\text{O}$) of residual soil water due to bare soil evaporation. Using the parameterization of Risi et al. (2016), we find that significant bare soil evaporation would move the expected Rayleigh distillation trends away from $\delta^{18}\text{O}$ –NGT correlation from the Albien aquifer in France (Figure 13; Risi et al., 2016). This approach is, however, poorly constrained as the effect of evaporative isotope enrichment depends jointly on the fraction of soil water lost to

evaporation and local relative humidity. When evaporation increases toward warmer conditions, the increasing evaporated fraction and stronger kinetic fractionation associated with lower relative humidity (RH_{land} evolving from 0.95 to 0.7, from LGM to present-day; Chadwick et al., 2016) act together, producing an approximately linear $\delta^{18}\text{O}$ –NGT relationship (Figure 13). Conversely, when evaporation is greater under colder conditions, the decreasing evaporated fraction and increasing kinetic fractionation toward warmer, lower-humidity conditions exert opposing effects, resulting in a curved, approximately parabolic $\delta^{18}\text{O}$ –NGT trajectory (Figure 13). These results are generally consistent with a limited impact of bare soil evaporation during water infiltration on WSI records in groundwater (Allison et al., 1983; Genty et al., 2014; Jasechko et al., 2015; Zhang et al., 2023).

4.2.3 Modern meteorological data: a critical analysis

Modern precipitation and temperature records from meteorological stations provide a valuable, spatially distributed dataset for examining the relationship between water stable isotopes (WSI) and temperature. As discussed in Section 4.2.3, previous studies have variably suggested that either seasonal (Siegenthaler and Oeschger, 1980) or interannual (e.g., Rozanski et al., 1992) co-variations between WSI in precipitation and local temperature may be most appropriate for paleoclimatic reconstructions based on isotope records. While both metrics are supposed to approximate how strongly condensation temperature controls isotopic fractionation during precipitation formation, the extent to which these modern relationships can be

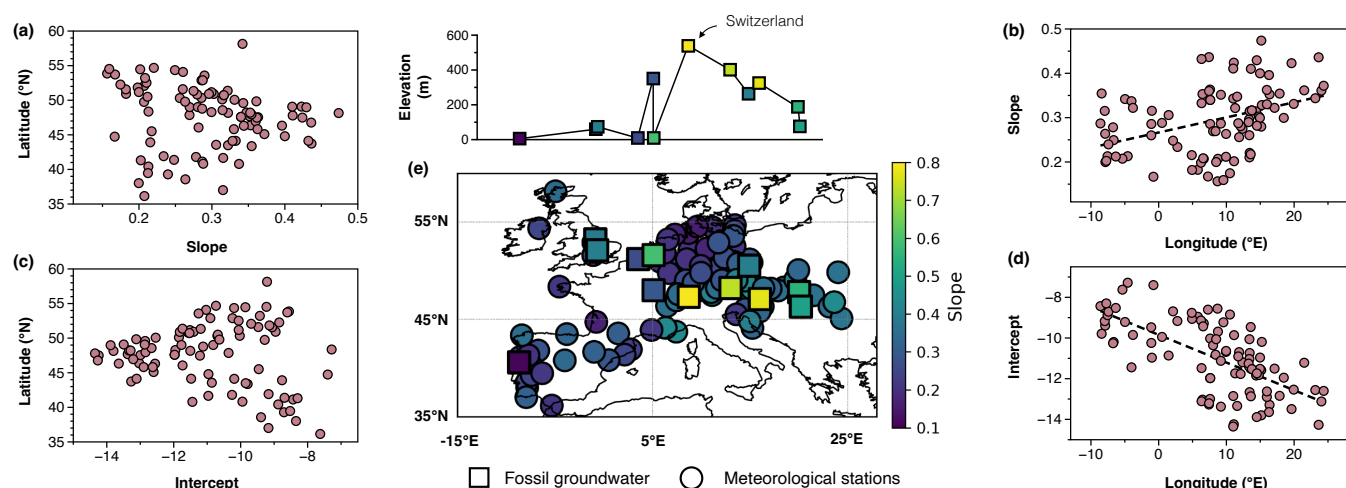


Figure 14. Compilation of modern European precipitation data from the WISER portal of the Global Network of Isotopes in Precipitation (GNIP). Slopes (a,b) and intercept (c,d) of $\delta^{18}\text{O}$ –surface air temperature correlations derived from monthly average data are represented as a function of latitude (a,c) and longitude (b,d). The locations of each meteorological station (Figure 1) are reported on panel (e) as circles, with the locations of fossil groundwater shown as squares. The color coding of both markers corresponds to the slope of the $\delta^{18}\text{O}$ – T correlations. Data from meteorological stations are slopes derived from monthly averages. Fossil groundwater slopes have been corrected for global ice-volume-related seawater $\delta^{18}\text{O}$ changes over the deglaciation. Indicative elevations of the fossil groundwater recharge areas are provided on top of panel (e) for reference, as a function of longitude.

extrapolated to longer timescales (e.g., glacial–interglacial periods) remains uncertain. Specifically, why the slopes of $\delta^{18}\text{O}$ –temperature correlations derived from monthly mean precipitation data are consistently smaller (average slope of $0.29 \pm 0.07 \text{‰ } ^\circ\text{C}^{-1}$) than those obtained from interannual averages (average slope of $1.04 \pm 0.30 \text{‰ } ^\circ\text{C}^{-1}$; Figure 9) is a key question that warrants further discussion.

First, monthly data arguably mix non-temperature controls with temperature, which dampens the apparent $\delta^{18}\text{O}$ – T sensitivity, whereas interannual averages filter those effects out, revealing a stronger temperature signal (e.g., temperature changes within a given season). To this extent, seasonal changes in atmospheric circulation patterns, moisture sources and transport pathways and/or condensation heights, may affect $\delta^{18}\text{O}$ with no proportional surface temperature change. For example, moisture transport to the Alps exhibits pronounced seasonality, shifting from an oceanic regime dominated by North Atlantic sources in winter to a more continental regime with substantial contributions from Central European land areas in summer (Sodemann and Zubler, 2010). Such month-specific atmospheric circulation regimes and seasonal moisture-source shifts should be broadly filtered out by the interannual averaging. More specifically, enhanced ^{18}O -depletion associated with heavy precipitation may occur during warm periods, thereby directly flattening the $\delta^{18}\text{O}$ – T slope. This “amount effect”, arguably more stable on interannual timescales, may also more drastically decouple $\delta^{18}\text{O}$ from temperature in seasonal data.

In detail, monthly WSI and temperature averages do not represent a purely seasonal signal: they span climatologically distinct regimes across the year while also incorporating both the seasonal cycle and an interannual component arising from the integration of mean $\delta^{18}\text{O}$ – T relationships

for a given month over multiple years. As such, they are unlikely to represent reliable $\delta^{18}\text{O}$ – T transfer functions for isotope-based paleoclimate reconstructions. Because interannual regressions span the same regime with different mean states, they are more likely to isolate robust thermodynamic controls on isotopic fractionation. However, they are not immune to temperature-unrelated changes in the WSI composition of precipitation. For example, the North Atlantic Oscillation (NAO) is known to directly influence precipitation $\delta^{18}\text{O}$ across Europe by affecting air temperature and steering the westerly winds that bring moisture to Europe during boreal winters (e.g., Comas-Bru et al., 2016). Paleoclimate studies have even attempted to use temporal variations in $\delta^{18}\text{O}$ recorded in natural archives from central Europe to reconstruct past NAO states over the Holocene (e.g., Baldini et al., 2008). The multiplicity of factors controlling the $\delta^{18}\text{O}$ composition of precipitation may thus ultimately require the use of non-linear regression to best capture time-varying relationships between $\delta^{18}\text{O}$ in precipitation and temperature (Vystavna et al., 2021). Previous analysis of global maps of precipitation isotopic seasonality also indicated spatially patterned variation in isotope/climate related to continentality in temperature-dominated regions (e.g., Europe; Figure 14b) and meridional trends in amount-dominated areas (Bowen, 2008), pointing to complex climatological controls on isotope seasonality.

Last but not least, one should bear in mind that precipitation isotope compositions primarily reflect temperature during precipitation events, which can be significantly different than time-averaged temperatures (Kohn and Welker, 2005). As precipitation amounts are rarely distributed uniformly throughout a particular time interval, systematic differences between average climatological temperatures and condensation temperatures during precipitation formation

may consistently bias WSI–T relationships that rely on time-averaged temperatures (Kohn and Welker, 2005).

4.2.4 Comparison of fossil groundwater and meteorological data

At first glance, the steeper slopes and lower intercepts of $\delta^{18}\text{O}$ –T correlations derived from fossil groundwater in Hungary compared to Western Europe (France and Belgium; Figure 10d) appears consistent with monthly-average modern precipitation data suggesting broadly increasing slopes ($R^2 = 0.15$) and decreasing intercepts ($R^2 = 0.44$) with longitude across Europe (Figure 14bd). However, utilizing $\delta^{18}\text{O}$ –T correlations derived from modern meteorological stations for isotope-based paleoclimate reconstructions requires several assumptions regarding atmospheric circulation patterns and source effects on WSI compositions. Due to global ice-volume-related changes in seawater composition, the $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of the ocean during the LGM (maximum ice-sheet extent) were approximately +1‰ and +8‰, respectively, relative to present-day values (Schrage et al., 2002). These differences must be considered when comparing modern and fossil data or when applying meteorological data to isotope-based paleoclimate reconstructions. In addition, changes in atmospheric circulation since the glacial period may have affected WSI compositions of precipitation, independently of temperature. For example, during the last glacial period, North Atlantic atmospheric circulation differed from that of today (e.g., Laine et al., 2009), largely due to topographic changes caused by extensive ice sheets in North America and Europe (Pausata et al., 2011). How past changes in the NAO influenced the evolution of precipitation $\delta^{18}\text{O}$ since the LGM (Luetscher et al., 2015) and throughout the deglaciation remains poorly understood. Critically, when using WSI–T correlations derived from fossil groundwater data for isotope-based paleoclimate reconstructions, no assumptions regarding oceanic WSI composition (Schrage et al., 2002) or past changes in atmospheric circulation patterns are required, as these factors are inherently incorporated into the proxy.

Here, the slopes of the WSI–T correlations derived from European fossil groundwater data (m) are corrected for global ice-volume-related seawater $\delta^{18}\text{O}$ and $\delta^2\text{H}$ changes during the LGM ($m_{\text{corr.}}$) following:

$$m_{\text{corr.}} = \frac{(m \times \max_{\text{NGT}} + c) - (m \times \min_{\text{NGT}} + c - \delta^{18}\text{O}_{\text{oc. LGM}})}{\max_{\text{NGT}} - \min_{\text{NGT}}} \quad (5)$$

whereby $\min_{\text{NGT}}$ and $\max_{\text{NGT}}$ represent the minimum and maximum NGT values derived for the considered record, c is the intercept of the uncorrected WSI–T correlation, and $\delta^{18}\text{O}_{\text{oc. LGM}}$ is the oceanic $\delta^{18}\text{O}$ during the LGM (for δD , $\delta\text{D}_{\text{oc. LGM}}$ is used analogously; see data compilation, Bekaert et al., 2026). These slopes are then compared with the correlations from modern precipitation data and local temperature measurements. This comparison is notably justified given the prominent role of temperature changes in governing WSI changes in European precipitation over the

LGIT (Section 4.2.2). The average slope of the corrected WSI–MAST regressions from European fossil groundwater (excluding two locations with particularly low correlation coefficients) is $0.48 \pm 0.21 \text{‰} \text{°C}^{-1}$ (1 SD; Figure 14), compatible within error with the monthly average-based correlations from modern precipitation data ($0.29 \pm 0.07 \text{‰} \text{°C}^{-1}$) but markedly lower than the average slope of interannual $\delta^{18}\text{O}$ –T correlations ($1.04 \pm 0.30 \text{‰} \text{°C}^{-1}$).

More specifically, the question remains as to which meteorological station(s) should be considered representative of the local or regional $\delta^{18}\text{O}$ –T sensitivity to be used for paleoclimate reconstructions. Estimates for the temperature sensitivity of WSI in precipitation from the 94 meteorological stations across Europe vary from ~ 0.16 up to $0.47 \text{‰} \text{°C}^{-1}$ (Figure 14). A $\delta^{18}\text{O}$ variation of +1‰ in an isotope record could thus potentially translate into reconstructed paleotemperatures ranging from +6.3°C down to +2.1°C, respectively. Most of this variability arguably reflects spatial heterogeneities in the temperature sensitivity of WSI in precipitation, as suggested by the longitudinal gradient of $\delta^{18}\text{O}$ –surface air temperature correlations across Europe (Figure 14b). Considering the average slope of the corrected WSI–MAST regressions from European fossil groundwater (i.e., $0.48 \pm 0.21 \text{‰} \text{°C}^{-1}$), a $\delta^{18}\text{O}$ variation of +1‰ would translate into reconstructed paleotemperatures of $2.1^{+1.6}_{-0.6} \text{°C}$. However, this average temperature sensitivity of paleoprecipitation WSI to paleotemperature changes should also be taken with caution, as it neglects local and regional effects such as continentality or elevation. Interestingly, three corrected WSI–MAST regressions from European fossil groundwater (1 from Switzerland, 2 from Austria) appear significantly higher (in the range $0.7\text{--}0.8 \text{‰} \text{°C}^{-1}$) than suggested from monthly averages of modern precipitation data (Figure 14). These three locations are also located at higher elevation than most other sites, potentially suggesting a control of elevation on the temperature-sensitivity of WSI in precipitation (Figure 14).

4.2.5 Implications for isotope-based paleoclimate reconstructions

Affolter et al. (2019) reconstructed temperature variations in Central Europe over the past 14,000 years from the WSI compositions of fluid inclusions in speleothems from the Milandre Cave (Switzerland). They used a $\delta^{18}\text{O}$ –T transfer function ranging from $0.36 \text{‰} \text{°C}^{-1}$ to $0.48 \text{‰} \text{°C}^{-1}$ (a weighted mean seasonal transfer function derived from meteorological data at the Bern station, Switzerland), which is consistent with our average value of $0.48 \pm 0.21 \text{‰} \text{°C}^{-1}$, but lower than the transfer function of $0.79 \text{‰} \text{°C}^{-1}$ from Switzerland fossil groundwater, once corrected for global ice-volume-related seawater $\delta^{18}\text{O}$ changes. Importantly, however, as illustrated on Figure 3g, the noble gas record from fossil groundwater in Switzerland only comprises 8 data-points, with one clear outlier. Removing this outlier lowers the slope of the $\delta^{18}\text{O}$ –NGT correlation from 0.63 ± 0.15 down to 0.34 ± 0.07 (Figure 3g). Once corrected for global ice-volume-related seawater $\delta^{18}\text{O}$ changes, this would yield a $\delta^{18}\text{O}$ –NGT transfer function of $0.51 \text{‰} \text{°C}^{-1}$, comparable

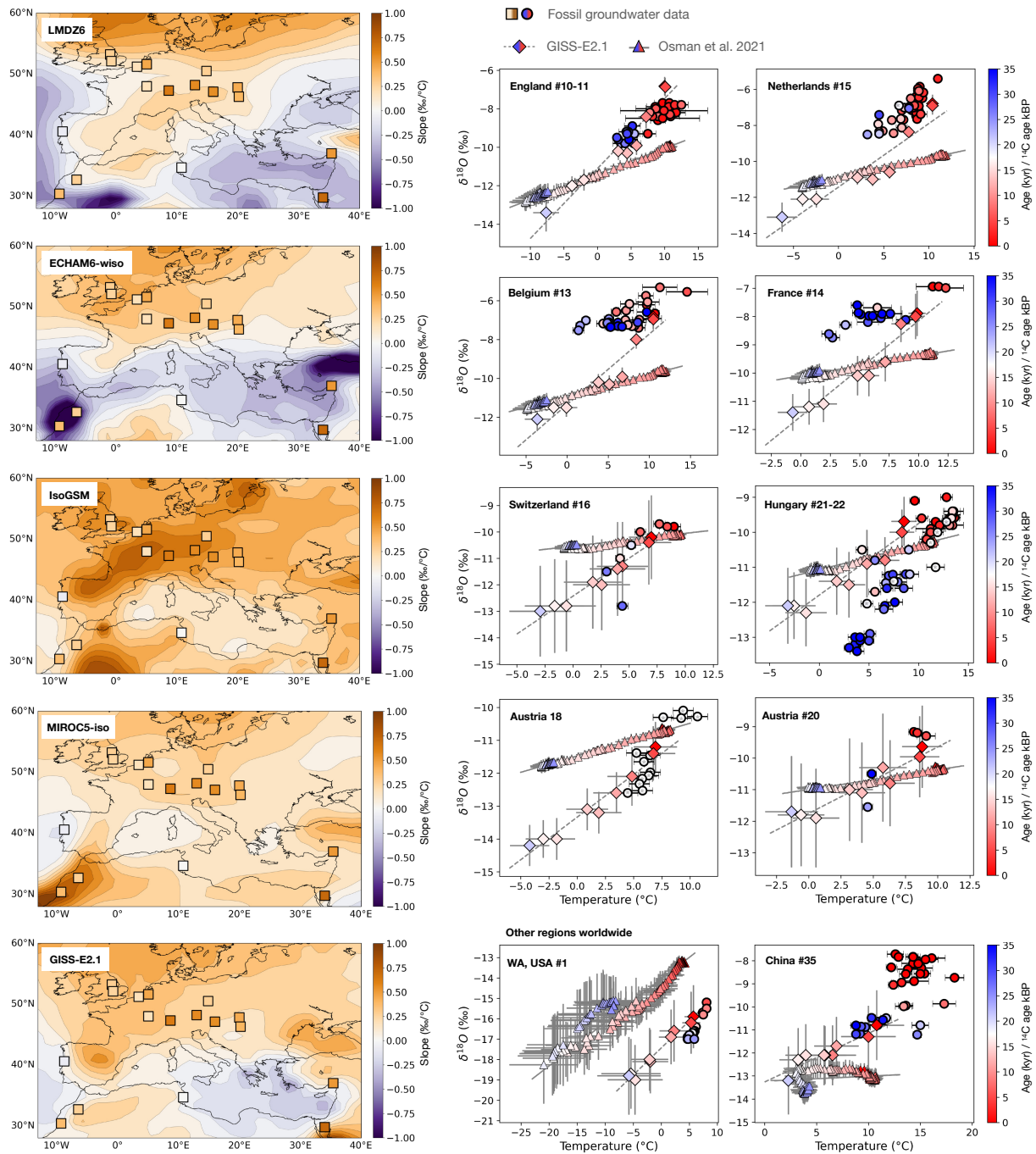


Figure 15. Comparison between GCM outputs and fossil groundwater data. Left side: Interpolated maps of $\delta^{18}\text{O}$ – T slopes from (top to bottom) LMDZ4 (Risi et al., 2010), ECHAM6-wiso (Cauquoin et al., 2023), IsoGSM (Lan et al., 2023), MIROC5-iso (Kino et al., 2024) and GISS-E2.1 (Roberts et al., 2011), compared with $\delta^{18}\text{O}$ – T slopes derived from fossil groundwater in this study. Right side: Comparison between $\delta^{18}\text{O}$ – T data derived from fossil groundwater and transitory simulation outputs from GISS-E2.1 (Roberts et al., 2011) and “Last Glacial Maximum reanalysis” (LGMR) (Osman et al., 2021). Marker colors refer to ^{14}C ages, with blue shades representing the last glacial period. Data without available ^{14}C ages are marked with an asterisk. As described in Section 2.2, given the inherent challenges of using ^{14}C in dissolved inorganic carbon for absolute groundwater dating (Han and Plummer, 2016), ^{14}C ages should be considered indicative only as these were taken directly from the original studies and, therefore, not uniformly processed/calibrated. Numbers given as # correspond to site locations given on Figure 1.

to the upper boundary ($0.48\text{‰}\text{°C}^{-1}$) of the range of $\delta^{18}\text{O}$ – T transfer functions used by Affolter et al. (2019) and compatible with our average value of $0.48 \pm 0.21\text{‰}\text{°C}^{-1}$ for Europe. This emphasizes the need for replicate analyses of groundwater samples in each paleoclimate reconstruction

study, as well as for maximizing the number of samples collected to evaluate potential artifacts.

The choice of one or several meteorological stations to derive $\delta^{18}\text{O}$ – T transfer functions may drastically change the outcomes of the climate reconstructions. In some cases,

this choice may also be somehow arbitrary, as there is no established maximal distance for a calibration site. What is the main source of uncertainty when adopting meteorological station data to derive $\delta^{18}\text{O}$ – T transfer functions for paleoclimate reconstructions? Is it the lack of constraints on the spatial gradients in the $\delta^{18}\text{O}$ – T transfer functions? Or is it the difference between the present-day and glacial-holocene slopes? Considering these uncertainties, to what extent do we need a local calibration, rather than a single transfer function for Europe? Which transfer functions should be used when the site of fossil groundwater is located in a region with limited access to meteorological data (e.g., England; Figure 14)? As outlined in Section 4.2.3, the question also remains as to whether $\delta^{18}\text{O}$ – T correlations based on monthly average or interannual variations should be used. That fossil data across glacial–interglacial transitions more closely resemble modern seasonal cycles than interannual variations in mean temperature and precipitation isotope compositions, as shown here, is somehow surprising given the line of arguments outlined in Section 4.2.3.

The quality of the $\delta^{18}\text{O}$ –MAST regressions derived from fossil groundwater data arguably provides an indication of whether $\delta^{18}\text{O}$ changes since the LGM have been primarily driven by local temperature (indicated by high-quality regressions) or by other processes (suggested by poor-quality regressions). In regions with no significant $\delta^{18}\text{O}$ –MAST regressions from fossil groundwater, paleotemperature reconstructions from WSI variations are likely not possible. For Europe, we argue that, when available, $\delta^{18}\text{O}$ – T correlations derived from European fossil groundwater data would be best appropriate for paleoclimate reconstructions, as they provide a direct record of the transfer function over the relevant timescale. At present, a physical interpretation of the intercepts in $\delta^{18}\text{O}$ – T relationships remains uncertain, and their significance in a paleoclimate context warrants further investigation.

To further evaluate these conclusions, the next section explores the consistency between fossil groundwater records and outputs from isotope enabled GCM simulations of the LGM, demonstrating that $\delta^{18}\text{O}$ – T correlations from fossil groundwater provide a valuable information for evaluating the accuracy of climate models that incorporate stable isotopes of precipitation.

4.3 Comparison with isotope-enabled model simulations of the LGM

Climate models aim to simulate the dynamics of the atmosphere, oceans, and land surfaces numerically. They simulate past and present climates and enable projections of future climate conditions. For about forty years (Joussau et al., 1984), water stable isotopes have been included into GCMs in order to both (i) support/guide the interpretation of WSI-based climate archives, and (ii) contribute to the evaluation of the representation of physical processes (e.g., Field et al., 2014). Today, many isotope-enabled atmospheric GCMs (Sections 2.4 and 4.3) treat H_2^{18}O and H_2^{16}O as prognostic tracers that are transported alongside H_2^{16}O through the hydrological cycle, using physically based

equilibrium and kinetic fractionation parameterizations to explicitly simulate isotopic fractionation during evaporation, condensation, cloud processes, and precipitation (e.g., Sturm et al., 2010; Werner et al., 2011). The incorporation of WSI into climate models allows direct process-based comparisons between simulated and proxy-derived isotope records, greatly improving our ability to investigate the processes governing the global hydrological cycle and to interpret isotope-based climate archives (e.g., Dee et al., 2023).

All isotope-enabled climate models explored here are consistent with positive $\delta^{18}\text{O}$ – T slopes over Europe over the deglaciation. However, no model strictly reproduces the spatial gradient of $\delta^{18}\text{O}$ – T slopes derived from fossil groundwater data. However, the LMDZ4 is the only model that simultaneously reproduces (i) the near-zero slope in Portugal, (ii) the consistent, low slopes ($\sim 0.2\text{‰}\text{°C}^{-1}$) in England, Belgium and France, and (iii) the greater slopes ($\sim 0.5\text{‰}\text{°C}^{-1}$) in Central Europe (Figure 15). While Iso-GSM clearly overestimates $\delta^{18}\text{O}$ – T slopes over Western Europe, MIROC5-iso yields satisfying results for Western Europe but fails to capture the effect of continentality (i.e., increasing slopes towards Central Europe). The effect of continentality is also not clearly observed in ECHAM6-wiso and GISS-E2.1 outputs, although ECHAM6-wiso generally provides a good match to fossil groundwater data. A comparison between fossil groundwater $\delta^{18}\text{O}$ – T data and transient climate simulation outputs indicates a significantly better match with GISS-E2.1 results (Roberts et al., 2011) than with outputs of the LGMR (Osman et al., 2021), for most sites in Europe and other regions worldwide (Figure 15). The earlier observation by Bekaert et al. (2023) that the slope of the $\delta^{18}\text{O}$ – T correlation from the LGMR presented in Osman et al. (2021) is significantly underestimated compared to French fossil groundwater constraints (and other climate model outputs) is confirmed at the global scale. These authors assimilated planktic foraminiferal oxygen isotope time series and compared output $\delta^{18}\text{O}$ of precipitation to independent ice core- and speleothem-derived $\delta^{18}\text{O}$ time series to evaluate model performances. While these authors found a good correspondence between simulated $\delta^{18}\text{O}$ and ice core proxy records, they did not use any terrestrial proxy data from Europe to evaluate their model. Our study demonstrates that the LGMR consistently underestimates the slopes of $\delta^{18}\text{O}$ – T correlations over the deglaciation in mid-latitude continental regions such as Europe, with the notable exception of the Pacific Northwestern United States (Seltzer et al., 2025).

Importantly, discrepancies between modelled and proxy-based $\delta^{18}\text{O}$ – T correlations may arise from biases in the simulated $\delta^{18}\text{O}$, temperature, or both. Additionally, modelled and proxy-based correlations may differ due to model resolution, if real-world variability occurs at finer spatial scale than simulated by the model. As shown in Figure 15, the slopes of $\delta^{18}\text{O}$ – T relationships between the LGM and PI may be similar when comparing GCM outputs with fossil groundwater data, even when the magnitudes of the corresponding changes ($\Delta\delta^{18}\text{O}_{\text{LGM-PI}}$ and $\Delta T_{\text{LGM-PI}}$)

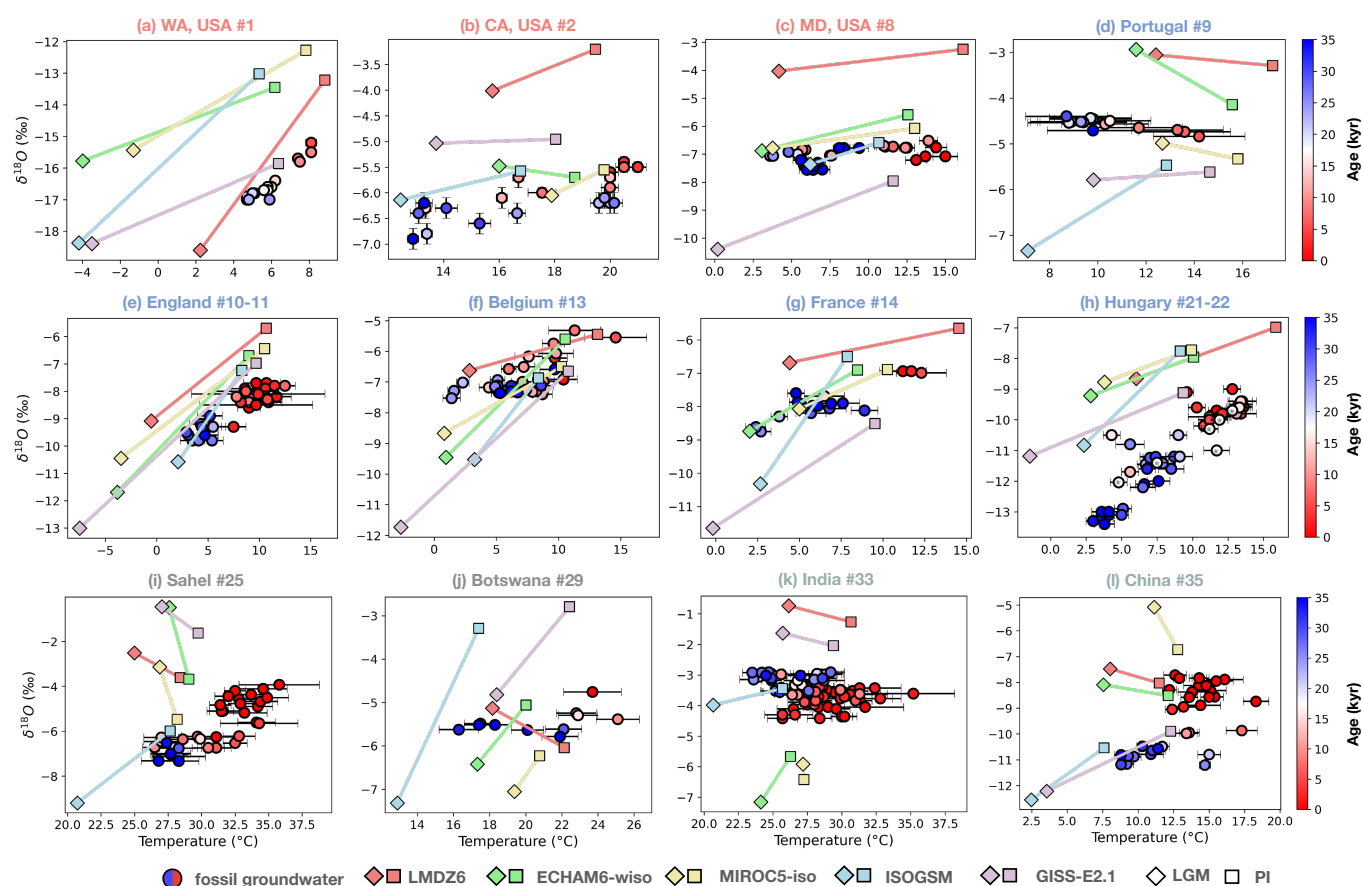


Figure 16. Comparison between GCM outputs and fossil groundwater data. LGM and PI $\delta^{18}\text{O}$ and temperature data from LMDZ4 (Risi et al., 2010), ECHAM6-wiso (Cauquoin et al., 2023), IsoGSM (Lan et al., 2023), MIROC5-iso (Kino et al., 2024) (Kino et al., 2024) and GISS-E2.1 (Roberts et al., 2011) are shown as diamond and square symbols. The data are taken from interpolated $\delta^{18}\text{O}$ and temperature maps at the location of each fossil groundwater site. Marker colors refer to ^{14}C ages, with blue shades representing the last glacial period. Data without available ^{14}C ages are marked with an asterisk. As described in Section 2.2, given the inherent challenges of using ^{14}C in dissolved inorganic carbon for absolute groundwater dating (Han and Plummer, 2016), ^{14}C ages should be considered indicative only as these were taken directly from the original studies and, therefore, not uniformly processed/calibrated. Numbers given as # correspond to site locations given on Figure 1. Reported uncertainties are 1 SD.

differ significantly (e.g., England). The magnitude of LGM cooling ($\Delta T_{\text{LGM-PI}}$) inferred from proxy data may be smaller (e.g., WA, USA), larger (e.g., CA, USA), or comparable (e.g., MD, USA; Portugal; France) to that simulated by GCMs. In some regions, particularly in Africa and Asia, large discrepancies exist among GCM outputs, with no satisfactory match to fossil groundwater data (Figure 16). These considerations underscore the need for comprehensive and exhaustive future compilations of proxy data, alongside systematic comparisons with GCM outputs, to evaluate the performance of climate models in simulating past (and future) climates.

5 Conclusion and perspectives

This study presents the first comprehensive compilation of fossil groundwater WSI and NGT records spanning the past ~35 kyr. We show that, in different regions worldwide (particularly in Europe), groundwater-based WSI compositions of precipitation and MAST co-evolved over the LGIT, revealing a primary control of regional temperature

on WSI compositions. In other regions, however (including temperate and arid basins over North America, Asia and Africa), the poor-quality of $\delta^{18}\text{O}$ –MAST regressions suggests that processes other than regional temperature influence WSI systematics (e.g., precipitation amounts). We argue that only regions presenting a clear dependency of WSI compositions on temperature may allow for isotope-based paleoclimate reconstructions from WSI variations in natural archives. In such regions, the use of the temperature sensitivity of WSI derived from fossil groundwater arguably represents the most reliable estimate of the WSI–T transfer functions. From European fossil groundwater, we find an average slope of the WSI–MAST regressions (corrected for global ice-volume-related seawater changes since the LGM) of $0.48 \pm 0.21 \text{ ‰ } ^\circ\text{C}^{-1}$ (1 SD). This value is compatible with the average slope of $\delta^{18}\text{O}$ –T correlations derived from monthly-averaged data of modern precipitation ($0.29 \pm 0.07 \text{ ‰ } ^\circ\text{C}^{-1}$; 1 SD), but lower than the average slope derived from interannual variations ($1.04 \pm 0.30 \text{ ‰ } ^\circ\text{C}^{-1}$; 1 SD).

The incorporation of stable water isotopes into atmospheric general circulation models offers a unique opportunity to explore the cause(s) of temporal and spatial slopes of $\delta^{18}\text{O}$ –MAST correlations. Assuming that only climate-related processes (such as atmospheric circulation, temperature, and seasonality) may disrupt the $\delta^{18}\text{O}$ –T relationships over deglaciation, the absence of statistically significant $\delta^{18}\text{O}$ –MAST correlations also offers valuable information for testing whether atmospheric models are able to capture processes other than Rayleigh distillation. Future work should explore potential causes of these $\delta^{18}\text{O}$ –T disruptions and attribute them to relevant climate processes (e.g., changes in moisture sources due to alterations in atmospheric circulation patterns, changes in seasonality, and/or oceanic evaporation-recharge processes). This may also require exploring complementary sources of local and regional WSI–T information, including cave drip water, for which a progressive $\delta^{18}\text{O}$ increase has been for example reported in France over the past 20 years ($\sim 0.50\text{‰}\text{°C}^{-1}$) and attributed to global warming (Genty et al., 2014; Zhang et al., 2023).

Future studies could also complement available insights from WSI–T reconstructions with proxies of groundwater paleorecharge dynamics (Wang et al., 2001; Johnson and Ingram, 2004; Cruz et al., 2005; Vimeux et al., 2005; Partin et al., 2007; Risi et al., 2008a; Pausata et al., 2011; Cheng et al., 2012; Lee et al., 2012; Collins et al., 2013; He et al., 2015). For example, excess air (ΔNe), which quantifies bubble entrainment and dissolution during recharge (Ingram et al., 2007; Aeschbach-Hertig and Solomon, 2013), has been proposed as a proxy for past water table depth fluctuations, reflecting changes in climatic conditions (e.g., humidity) and recharge dynamics (e.g., strong intermittent recharge events associated with a highly seasonal or millennial timescale recharge pulses; Ram et al., 2022). Another potential tool is ultrahigh precision Ar–Kr–Xe isotope systematics, which primarily reflect depth-proportional signals set by gravitational settling in soil air at the time of recharge and provide a unique quantitative proxy for past water table depth reconstructions (Seltzer et al., 2019, 2025). Associating these proxies of precipitation evolution during e.g. deglacial warming (e.g., high ΔNe and rising water table = increasing precipitation) with WSI systematics in regions where WSI variations are thought to primarily reflect changes in precipitation amounts represents a promising avenue to better document past hydroclimate transitions worldwide. Such efforts may also provide an opportunity to improve paleoclimate reconstructions based on dissolved noble gas systematics – for example, by enabling the evaluation of potential effects associated with geothermal heating variations, particularly in deep aquifers ($\geq 50\text{ m}$).

Finally, this study presents a first attempt to evaluate WSI–T correlations derived from fossil groundwater data using isotope-enabled climate models. Overall, we find substantial disagreement between proxy data and climate model results, as well as considerable spread among the models themselves. Understanding why climate models fail

to capture glacial–interglacial WSI–T relationships warrants further investigation. Critically, the fidelity of simulated continental temperatures depends strongly on the prescribed sea surface temperatures (SSTs). Because atmospheric circulation – such as the position of the westerly jet and the strength of the monsoon – is strongly influenced by imposed SSTs, this likely contributes to site-to-site differences in the simulated WSI–T slopes. While it is clear that model results exhibit large scatter, the underlying causes remain to be fully explored. Testing different LGM SST patterns (e.g., following Kino et al., 2024) could represent a promising avenue for quantifying the influence of prescribed SSTs on simulated WSI–T relationships.

Acknowledgements

This research was funded by the French Agence Nationale de la Recherche, Grant ANR-22-CPJ2-0005-01 to D.V.B. We thank the members of the “Cycle, Atmosphères, Climats” theme at CRPG for fruitful discussions that helped the interpretations of this study.

Data, code, and outputs availability

All modern European precipitation data obtained from the WISER (Water Isotope System for Data Analysis, Visualization, and Electronic Retrieval) portal of the Global Network of Isotopes in Precipitation (GNIP) are available in the supporting data compilation of Bekaert et al. (2026, <https://doi.org/10.60520/IEDA/114589>). This online repository also contains all the noble gas temperature and water stable isotope data compiled from the literature, along with corresponding references and correlation parameters derived in this study. Climate model output data are directly available through corresponding public repositories: ECHAM6-wiso; LMDZ4; IsoGSM; MIROC5-iso; GISS-E2.1; LGMR. Main text figures are available for download in the online version of this article.

Competing interests

The authors declare no competing interests.

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