



1 **Hydroclimatic instability during the late Roman Warm Period–early Dark Ages Cold** 2 **Period transition in the eastern Mediterranean**

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8 **Abstract**

9 Here we present a stalagmite-based annual-resolved hydroclimate record, with a mean sampling
10 resolution of 1.1 yr, spanning 154–529 CE from Börtlücke Cave, western Türkiye, encompassing
11 the latter part of the Roman Warm Period (RWP) and the transition into the early Dark Ages Cold
12 Period. During the late RWP, elevated Mediterranean Sea surface temperature (SST) likely
13 enhanced cyclone activity and moisture transport, promoting persistently humid conditions. The
14 subsequent transition from 290 to 430 CE was characterized by pronounced multi-decadal
15 hydroclimatic variability, likely reflecting the combined influences of Mediterranean SSTs, large-
16 scale atmospheric circulation associated with the North Atlantic Oscillation, and solar variability.
17 The reconstructed humid conditions broadly coincide with the period of prosperity and territorial
18 expansion of the Roman Empire, whereas the following increase in hydroclimatic instability may
19 have amplified the environmental stresses during the decline of the Western Roman Empire. Our
20 findings highlight the complex interplay among Mediterranean SST, atmospheric circulation,
21 volcanic eruption, and solar forcings in shaping the eastern Mediterranean hydroclimate through
22 the early Common Era.

23 **1. Introduction**

24 Climate variability has affected ancient civilizations and human societies (Cullen et al., 2000; Shi
25 et al., 2022; Wanner et al., 2022). For example, favorable hydroclimatic conditions during the
26 Roman Empire most likely enhanced agricultural production through increased water availability,
27 contributing to societal prosperity and resilience (McCormick et al., 2012). Intensified
28 hydroclimatic variability (Büntgen et al., 2011) and/or prolonged droughts (Hu et al., 2022) likely
29 intensified environmental stresses associated with the decline of the Western Roman Empire,
30 which ultimately collapsed in 476 CE (Norman et al., 2025). Understanding the range and drivers
31 of past climate variability is particularly important in the context of ongoing anthropogenic-
32 enhanced climate change (Rawat et al. 2024; Smith et al. 2025), as it provides a long-term
33 perspective on climate dynamics and helps to improve projections of future environmental change.

34 The Roman Warm Period (RWP), generally considered to span from ~200 BCE–400 CE
35 (Ljungqvist, 2010; Helama et al., 2017; Shi et al., 2022) represents a prominent warm phase of the



last two millennia, characterized by generally warm and relatively humid conditions across much of Europe and the Mediterranean region (Ljungqvist, 2010; Hu et al., 2022; Shi et al., 2022). The RWP has commonly been linked to a combination of reduced volcanic activity, relatively elevated solar irradiance, and favorable ocean–atmosphere circulation conditions that promoted warmer climatic conditions across many regions of the Northern Hemisphere (Büntgen et al. 2020; Margaritelli et al. 2020; Shi et al., 2022). In contrast, the Dark Ages Cold Period (DACP), commonly dated to ~400–765 CE (Ljungqvist, 2010; Helama et al., 2017; Shi et al., 2022) is generally regarded as a cooling interval (Ljungqvist, 2010; Helama et al., 2017; Bartolomé et al., 2024), primarily attributed to clustered volcanic eruptions, with additional contributions from reduced solar activity, sea-ice–ocean feedbacks, and changes in regional atmosphere–ocean circulation (Büntgen et al., 2016; Helama et al., 2017; Waltgenbach et al., 2021).

The timing and magnitude of both the RWP and the DACP vary considerably among regions, reflecting possible differences in the response of climate systems to external and internal forcings. For instance, the RWP is generally dated to 1–250 CE in the Northern Hemisphere (Shi et al., 2022), 150 BCE–270 CE in Spain (Gil García et al., 2007), and 200 BCE–400 CE in Italy (Hu et al., 2022). Likewise, the DACP spans approximately 400–765 CE in the Northern Hemisphere (Shi et al., 2022) and 450–850 CE in Spain (Bartolomé et al., 2024). These differences highlight the spatial heterogeneity in hydroclimatic evolution across Europe and the Mediterranean and emphasize the necessity of high-resolution, absolutely dated regional paleoclimate records to better constrain the timing, duration, and mechanisms of these climatic transitions.

This study presents an annually resolved hydroclimate reconstruction for the eastern Mediterranean from 154 to 529 CE (1796–1421 yr BP, relative to 1950 CE), encompassing the latter part of the RWP and the transition toward the early DACP. The reconstruction is based on speleothem $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ time series constrained by a robust U-Th chronology from Börtlüce Cave in western Türkiye, providing a reliable temporal framework for assessing the relative influences of regional and large-scale forcings on the eastern Mediterranean hydroclimatic variability.

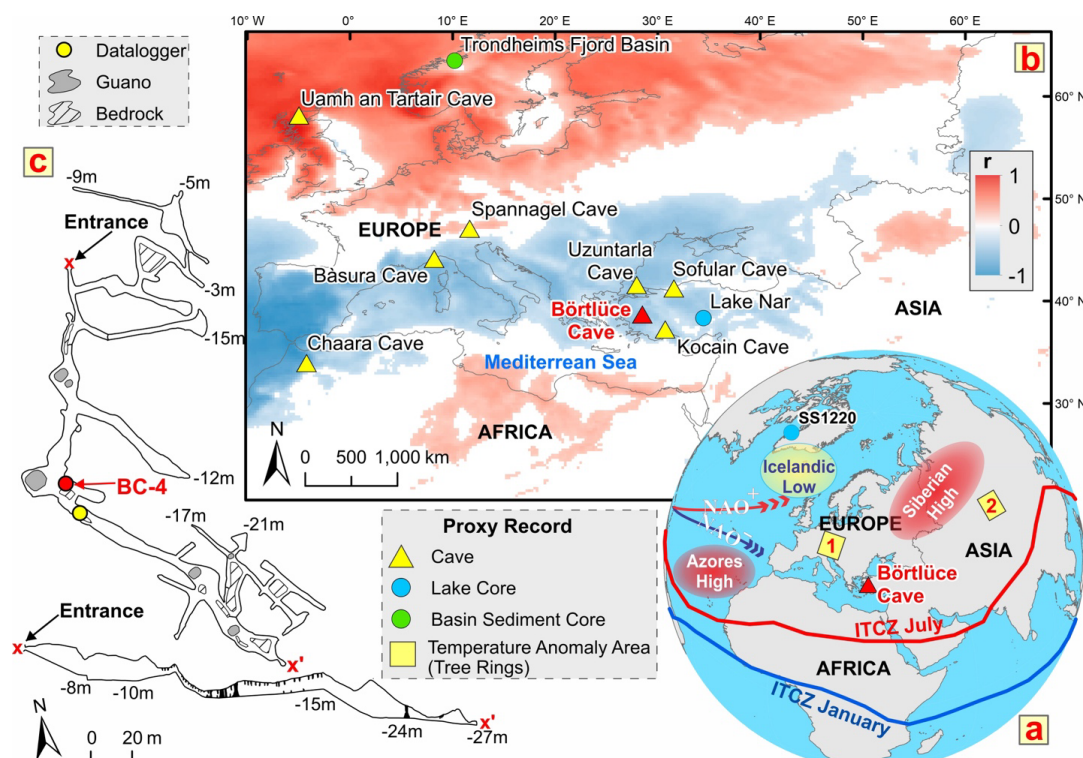
2. Cave settings and sample

Börtlüce Cave is located in Kula District of Manisa Province, western Türkiye (38° 41.624'N, 28° 30.031'E; 459 m above sea level) and developed in Paleozoic marbles overlying Precambrian metamorphic formations (Fig. 1a-b). The study region is situated within a climatic transition zone between polar and tropical atmospheric circulation systems (Fig. 1a). Hydroclimatic variability in the region is controlled by a number of large-scale climate drivers, including the Icelandic Low, Azores High, the Siberian High, changes in the strength and position of the westerlies associated with the North Atlantic Oscillation (NAO) (Türkes and Erat, 2003; Sezen and Partal, 2017; Baltaci et al., 2025), and latitudinal shifts of the Intertropical Convergence Zone (ITCZ).



72 The typical Mediterranean climate patterns with warm, dry summers and cold, wet winters, are
 73 dominant. The vegetation above and in Kula District consists of Mediterranean climate-adapted
 74 woodland, mainly deciduous *Quercus*, *Pinus brutia* and herbs (Akbaş, 2020). The instrumental
 75 observations from the nearest meteorological station in the Salihli District, located 38 km
 76 southwest of the cave, indicate that the total annual precipitation is 486 ± 180 (2σ , hereafter) mm
 77 (1960-2020 CE), with 45% occurring in winter, followed by 28% in spring and 21% in autumn,
 78 whereas summer contributes only about 6%. Mean annual temperature is 16.3 ± 2.3 °C, increasing
 79 from 7.1 ± 2.5 °C in winter to 25.9 ± 3.3 °C in summer (Fig. S1a).

80 Börtlücke Cave with an only narrow entrance, 1.0 m in diameter is developed mainly horizontally
 81 with a main passage of 575 m and several side passages, 10s-100s m in length (Fig. 1c). A candle-
 82 shape stalagmite BC-4 was collected 80 m from the entrance (Fig. 1c). The cave data logger,
 83 located 10 m next to the stalagmite sampling site (Fig. 1c), displays high relative humidity of 99.7
 84 $\pm 1.2\%$ and stable monthly temperature of 15.5 ± 0.15 from August, 2023 to July, 2024 (Fig. S2c-
 85 d).



86 **Figure 1.** Overview of the Börtlücke Cave (Kula, Manisa) and regional atmospheric circulations. **a)** General
 87 atmospheric circulation systems influencing climate settings in the circum-Mediterranean realm. The solid
 88 red and light-blue curves represent the averaged modern positions of the Intertropical Convergence Zone
 89 (ITCZ) in July and January, respectively (Cheng et al., 2012). The red and blue arrows indicate the wind
 90



91 directions of the westerlies associated with the positive (NAO⁺) and negative (NAO⁻) phases of the North
 92 Atlantic Oscillation (NAO), respectively. The yellow rectangles outlined in black indicated tree ring-based
 93 summer temperature anomalies from European Alps (1) and Russian Altai (2) (Büntgen et al., 2016). **b)**
 94 The location of the Börtlüce Cave (red triangle outlined in black) and cited proxy records on spatial
 95 correlation distribution map of the NAO and boreal winter precipitation (December–February) between
 96 1940–2024 ($p \leq 0.05$) (NAO index from NOAA, <https://www.ncei.noaa.gov/access/monitoring/nao/>;
 97 precipitation data from the ERA5 monthly averaged single-level dataset, Hersbach et al., 2023). **c)** The plan
 98 of the Börtlüce Cave (MTA, 2011), the location of studied stalagmite BC-4 (red circle), and monitoring
 99 data logger (yellow circle) in the cave. “x” and “x” respectively mark the entrance and end of this cave.

100 3. Material and Methods

101 All procedures, from subsampling to instrumental analyses, for stalagmite BC-4 were conducted
 102 at the High-Precision Mass Spectrometry and Environment Change Laboratory (HISPEC),
 103 Department of Geosciences, National Taiwan University. For U-Th dating, 14 subsamples, 15–60
 104 mg each, were drilled at selected horizons along the main growth axis to provide an age model
 105 (Fig. S2a). The wet chemistry was conducted on class-100 clean benches in a class-10,000 clean
 106 room (Shen et al., 2008) and U-Th isotopic compositions were measured using a Thermo Fisher
 107 Neptune multi-collector inductively coupled plasma mass spectrometer (Shen et al., 2012).
 108 Analytical uncertainties are reported as two-sigma (2σ) errors of the mean. Age calculations were
 109 based on half-lives of ²³⁴U in Hu et al. (2025) and others available in Cheng et al. (2013). Corrected
 110 ages with 2σ uncertainty are reported as years before present (BP), where present is defined as
 111 1950 CE, converted to calendar years BCE/CE throughout the manuscript (Table S1). Detailed
 112 analytical procedures including chemical preparation and instrument are described by Shen et al.
 113 (2002, 2012) and Cheng et al. (2013).

114 A total of 348 powdered subsamples, 50–100 μg each, were drilled on a class-100 clean bench for
 115 stable isotope analysis at 0.5 mm intervals along the growth axis of stalagmite BC-4 (Fig. S2a). In
 116 addition, 15 coeval powdered subsamples were collected from three individual growth layers for
 117 the Hendy test (Hendy, 1971) in order to evaluate possible kinetic isotope fractionation during
 118 deposition. Oxygen and carbon isotope compositions were measured using a Thermo Scientific
 119 Delta V Plus isotope ratio mass spectrometer (IRMS) coupled to an automated GasBench II
 120 carbonate preparation system. All $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values are reported in per mil (‰) relative to the
 121 Vienna Pee Dee Belemnite (VPDB) standard. The reproducibility of the measurements was better
 122 than $\pm 0.074\text{‰}$ for $\delta^{13}\text{C}$ and $\pm 0.14\text{‰}$ for $\delta^{18}\text{O}$ at one-sigma level (1σ).

123 4. Results and Discussion

124 4.1. Chronology and BC-4 $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ records

125 Determined U-Th isotopic compositions and contents are listed in Table S1. Stalagmite BC-4
 126 deposited from 7477 ± 423 BCE at a depth of 195 mm to 522 ± 7 yr CE at 8 mm (Fig. S2). A major
 127 hiatus is observed between 175–171 mm, where ages shift abruptly from 3150 ± 307 BCE to 163



± 13 CE (Fig. S2, Table S1). The 11 U-Th ages, constrained by low age uncertainties of ± 7–44 years (2σ) for the upper 175-mm section are in a stratigraphic order and used in this study (Table S1). StalAge algorithm (Scholz and Hoffmann, 2011) was used to build an age with 2-sigma uncertainties of ± 13–34 years at 154–370 CE, and ± 5–25 at 370–529 CE for this section (Fig. S2). This upper section exhibits growth rates ranging from 0.2–1.5 mm yr⁻¹ and an exceptional high rate of 3.3 mm yr⁻¹ between 438 ± 10 and 442 ± 10 CE (Fig. 3d).

The Hendy Test (Hendy, 1971) results on the three layers (Fig. S2a) indicate that intra-lamina δ¹⁸O variation was limited, ranging from ± 0.16 to ± 0.29‰ (1σ) (Fig. S3a). The correlations between coeval δ¹⁸O and δ¹³C values vary, with correlation coefficients of $r = -0.04$ ($p = 0.95$), $r = 0.49$ ($p = 0.41$), and $r = 0.78$ ($p = 0.12$) for Layers 1, 2, and 3, respectively, which are statistically insignificant (Fig. S3b-d). These lines of evidence and stable cave temperature and high humidity (Fig. S1b) provide no evidence for strong kinetic isotope fractionation (McDermott, 2004; Dorale and Liu, 2009) and support the paleoclimatic interpretation of the BC-4 record. The minor correlation between two δ¹⁸O and δ¹³C time series of this section with a correlation coefficient of $r = 0.43$ ($p < 0.001$), suggests that the variations of two proxies could be affected by the same or similar forcing(s) (Lachniet, 2009).

Based on the chronology (Fig. S2b), the average temporal resolution is 1.1 years (Fig. 2e) per stable isotope data point for both stalagmite BC-4 δ¹⁸O and δ¹³C time series (Fig. 2b-c), providing hydroclimate changes during this 375-yr interval from 154 CE to 529 CE. The δ¹⁸O values range from -6.3 to -4.4‰, with a mean value of -5.5‰ and the δ¹³C values from -8.9 to -5.6‰, with a mean value of -7.3‰ (Fig. 2b-c). After a decreasing from -5.3‰ at 154 CE to -6.2‰ at 175 CE, δ¹⁸O displays relatively stable values until 290 CE, with an average of -5.9‰ (-6.3 to -5.2‰), whereas δ¹³C values showing more fluctuations ranging from -8.4 to -6.4‰ with an average of -7.3‰. From 295 to 430 CE, both isotope records show multi-decadal variability with amplitudes of 1.0-1.6‰ for δ¹⁸O and 1.5-2.4‰ for δ¹³C. Subsequently, a sharp negative excursion at 437–442 CE with the highest growth rate of 3.3 mm yr⁻¹, with δ¹⁸O falling by ~1.3‰ and δ¹³C declining by ~2.0‰. After this excursion, both δ¹⁸O and δ¹³C gradually shifted towards low values until ~484 CE from -5.3 and -7.2‰ to -6.2 and -8.6‰, respectively, followed by another gradual increase in δ¹⁸O and δ¹³C shifts until 509 CE up to -4.8 ‰ and -5.9‰, respectively, and then a subsequent 2-decadal decrease to -5.6‰ for δ¹⁸O and -7.2‰ for δ¹³C at 529 CE (Fig. 2b-c), the end of the record.

Stalagmite BC-4 δ¹⁸O record is interpreted primarily as a proxy for effective moisture (precipitation minus evaporation, P-E), with low (high) δ¹⁸O values indicate relatively wet (dry) conditions (Bar-Matthews et al., 1997; Ünal-İmer et al., 2015; Baykara et al., 2026). The δ¹³C record predominantly reflects effective moisture through soil biological activity, vegetation productivity, and karst processes, with low (high) δ¹³C values indicating enhanced (reduced) soil productivity under relatively wet (dry) conditions (Ünal-İmer et al., 2015; Baykara et al., 2026). Detailed climatic interpretation is given in Text S1.

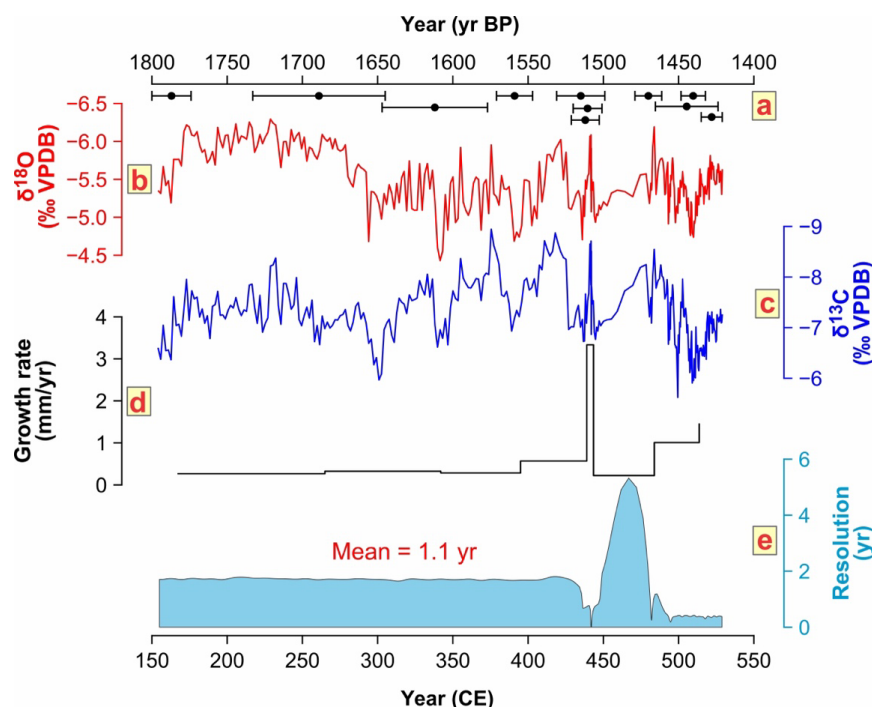


Figure 2. Stalagmite BC-4 proxy records from Börtlücke Cave. **a)** U-Th ages (black points with horizontal 2σ error bars). **b)** $\delta^{18}\text{O}$ (red) and **c)** $\delta^{13}\text{C}$ (blue) time series. **d)** Growth rate. **e)** Temporal resolution of every data point (mean = 1.1 years).

4.2. Hydroclimate dynamics in the Eastern Mediterranean region

The BC-4 $\delta^{18}\text{O}$, after a decrease by $\sim 1\text{‰}$ from 154 to 170 CE, is characterized with its lowest values of $\sim 6\text{‰}$ during 170–290 CE, accompanied by a similar pattern of $\delta^{13}\text{C}$ values (Fig. 3k–l), suggesting relatively wet climate conditions in the region. Similarly, tree ring-based summer temperature reconstructions from the European Alps and the Russian Altai (Fig. 3b–c) indicate a general warming period during 175–290 CE, except for around 5-years negative excursions at 200 and 270 CE. Multi-decadal-resolved proxy records in the Mediterranean, such as Sofular Cave (NW Türkiye), Kocain Cave (SW Türkiye) and Bäsura Cave (N Italy), also reflect such a wet climate interval between 170–290 CE (Fig. S4). However, the southwestern Mediterranean rainfall index from Chaara Cave (N Morocco) indicates dry conditions between 200 and 290 CE (Fig. 3j), while the Spannagel Cave record (W Austria) reflects cold conditions between 154 and 290 CE, except for a brief 15-year warm interval between 260 and 275 CE (Fig. 3i). Similarly, the Uamh an Tartair Cave record (NW Scotland) indicates a gradual shift toward drier conditions from 154 to 290 CE (Fig. 3h). The spatial heterogeneity among these records suggests that hydroclimate across the Afro-European Mediterranean realm was most likely governed by interacting multiple forcings rather than a single driver.



186 During this period between 154 and 290 CE, two large unidentified volcanic eruptions (Fig. 3n)
187 took place at 169 CE and 266 CE with a radiative forcing of -11 and -15 W m^{-2} , respectively (Sigl
188 et al. 2015; Toohey and Sigl, 2017). Previous studies (Robock, 2000; Sigl et al. 2015) have shown
189 that large volcanic eruptions can inject sulfate aerosols into the stratosphere, leading to Earth's
190 surface cooling through the reflection of incoming solar radiation while simultaneously warming
191 the stratosphere through the absorption of both solar and terrestrial radiation. The resulting
192 radiative perturbations can subsequently alter atmospheric circulation patterns, affecting the
193 regional climate conditions for up to a decade. Such volcanic eruptions generally reduce global
194 precipitation through negative radiative forcing while simultaneously generate pronounced
195 regional hydroclimatic anomalies by modifying regional atmospheric circulation (Robock, 2000;
196 Iles et al., 2013). In particular, some arid and semi-arid regions may experience enhanced
197 precipitation following major volcanic eruptions (Zuo et al., 2019; D'Agostino et al., 2026).

198 The unidentified volcanic eruption recorded in bipolar ice-core sulfate records at 169 CE (Sigl et
199 al., 2015) does not coincide with reduced sunspot activity or solar irradiance, whereas the 266 CE
200 eruption occurred during a period of reduced solar activity (Fig. 3d-e). Temperature
201 reconstructions indicate cooling following both eruptions, with the strongest cooling response after
202 169 CE recorded in the Russian Altai up to $2 \text{ }^{\circ}\text{C}$ (Fig. 3c) and after 266 CE in the European Alps
203 with a cooling of $1.5 \text{ }^{\circ}\text{C}$ (Fig. 3b). The BC-4 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records exhibit a brief negative
204 excursion of $\sim 0.5\text{‰}$ between 170 and 175 CE (± 13 years, 95% confidence interval), but show no
205 distinguishable response after the 266 CE eruption (Fig. 3k-l). The Uamh an Tartair Cave,
206 Spannagel Cave, and the rainfall index of Chaara Cave records show no clear response to either
207 eruption (Fig. 3h-j). In contrast, the Uzuntarla Cave (NW Türkiye) record exhibits a $\sim 1.0\text{‰}$
208 decrease in $\delta^{13}\text{C}$ after the 169 CE eruption and a $\sim 0.5\text{‰}$ increase after the 266 CE eruption with
209 large age uncertainties (± 65 and ± 64 years, 95% confidence interval, respectively). Overall, the
210 sparse chronological control in the records, based on only one or two U-Th age determinations,
211 precludes a confident assessment of the synchronicity between the observed proxy responses and
212 the volcanic events.

213 One of the major drivers affecting the regional hydroclimate variability is the NAO, the primary
214 mode of winter atmospheric circulation over the North Atlantic, defined by fluctuations in the
215 pressure gradient between the Azores High and the Icelandic Low (Hurrell et al., 2003). During
216 the positive (negative) phase of the NAO, an enhanced (reduced) pressure gradient strengthens
217 (weakens) the westerlies and shifts the North Atlantic storm track northward (southward), leading
218 to increased (decreased) winter precipitation over northern Europe and generally drier (wetter)
219 conditions across the Mediterranean region (Visbeck et al., 2001; Türkeş and Erlat, 2003).

220 Between 154–290 CE, the decadal-resolution Norwegian fjord record (Fig. 3f) indicates generally
221 weakly positive NAO conditions, suggesting relatively modest westerly circulation, whereas the
222 Greenland NAO reconstruction with a multi-decadal resolution (Fig. 3g) shows a broad trend
223 toward a positive NAO phase, implying strengthened westerly circulation. The centennial



224 decreasing growth rate recorded in Uamh an Tartair Cave (Fig. 3h) implies a transition from dry
225 to wet conditions associated with increasing in the strength of NAO (Baker et al., 2015). The
226 western Mediterranean rainfall index (Fig. 3j) indicates low values between 180–290 CE,
227 suggesting that dry conditions mainly associated with the positive NAO phase (Ait Brahim et al.,
228 2019). By contrast, BC-4 isotope records point to sustained humid conditions in the northeastern
229 Mediterranean. These opposite hydroclimatic patterns suggest that the NAO variability alone
230 cannot explain the hydroclimatic evolution across the Mediterranean during 154–290 CE. Hu et
231 al. (2022) proposed that the Mediterranean Sea warming may have provided an additional regional
232 mechanism for increased moisture availability in the surrounding areas. This interpretation is
233 consistent with modern climatological studies showing that warm Mediterranean Sea enhances
234 evaporation and atmospheric moisture supply, thereby increasing precipitation over surrounding
235 regions (Mariotti, 2010; Gimeno et al., 2012).

236 Modern climate-model simulations suggest that elevated Mediterranean Sea surface temperatures
237 (SST) can enhance precipitation over Türkiye (Bozkurt and Sen, 2011). The stacked Mediterranean
238 SST anomaly record suggests warm sea surface conditions for the period of 150–190 CE (Fig.
239 3m). High Mediterranean SST may have strengthened the sea–land thermal gradient, promoting
240 cyclone trajectories from the Aegean Sea toward western Türkiye and enhancing regional
241 precipitation (Karaca et al., 2000).

242 During 290–430 CE, BC-4 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records exhibit three distinct multi-decadal variations
243 (Fig. 3k-l). Each variation is characterized by a gradual 30–40-year decrease, 0.5‰ in $\delta^{18}\text{O}$ and
244 2‰ in $\delta^{13}\text{C}$ (Fig. 3k-l), indicating a slow return to humid conditions, followed by an abrupt decadal
245 isotopic enrichment indicative of short-lived dry episodes. Multi-decadal-to-centennial
246 hydroclimatic variability is also observed in sub-decadal-resolved proxy records of Uamh an
247 Tartair and Spannagel (Fig. 3h-i) and Uzuntarla Caves (Fig. S4g). However, the limited
248 chronological control of these records, based on only one or two U–Th age determinations with
249 uncertainties of ± 20 –300 years, precludes the reliability of direct comparisons with the Börklüce
250 Cave record. In contrast, sub-decadal-to-multidecadal resolved records of Sofular Cave in
251 northwestern Türkiye (Fig. S4c), Kocain Cave in southwestern Türkiye (Fig. S4d), Lake Nar in
252 central Türkiye (Fig. S4e), and Bäsura Cave in northern Italy (Fig. S4f) preserve only gradual
253 centennial-scale transitions toward aridity (Fig. S4). The stacked Mediterranean SST anomaly
254 shows a gradual decline from 340 to 430 CE (Fig. 3m), probably reflecting reduced cyclonic
255 activity and moisture transport (Hu et al., 2022). Temperature reconstructions from the Northern
256 Hemisphere (Fig. 3a) and the European Alps (Fig. 3b) also display multi-decadal fluctuations
257 during 290–430 CE. The absence of large volcanic eruption recorded during this period (Fig. 3n)
258 suggests volcanic forcing is not likely to be responsible for the sustained hydroclimatic variability.
259 However, solar activity records (Fig. 3d–e) exhibit sub-centennial- to centennial-scale variability.
260 We propose that solar forcing may have interacted with Mediterranean SSTs and large-scale
261 atmospheric circulation to modulate the pronounced multi-decadal climatic variability observed
262 during the transition from the late RWP to the early DACP.



263 The Greenland NAO reconstruction (Fig. 3g) indicates relatively strong westerlies at 290–320 CE,
 264 followed by a weakening from 320 to 400 CE and a subsequent recovery toward 430 CE. Instead,
 265 Norwegian NAO reconstruction (Fig. 3f) exhibits noticeable multi-decadal variability at this
 266 Transition interval. Together with variability in the solar activity, the European Alps temperature
 267 anomalies, and the NAO-sensitive Uamh an Tartair and Spannagel Cave records (Fig. 3), these
 268 observations suggest that atmospheric circulation remained highly dynamic throughout 290–430
 269 CE.

270 Wavelet coherence analysis, using the Morlet wavelet (Torrence and Compo, 1998; Grinsted et al.,
 271 2004), reveals intermittent but significant coherence between the BC-4 isotope records and
 272 regional temperature reconstructions, NAO records, and European speleothem proxies,
 273 predominantly at interannual-to-decadal periodicities (~4–20 yr) (Figs. S5–S6). Significant
 274 coherence is observed particularly during the ~290–430 CE interval, although its timing, strength,
 275 and phase relationships vary among the individual proxy records. Due to the chronological
 276 uncertainties and differences in temporal resolution among the proxy records, the phase
 277 relationships at these relatively short timescales cannot be robustly resolved. Therefore, the
 278 wavelet coherence results are interpreted only as evidence of intermittent shared variability, rather
 279 than precise synchronicity, phase relationships, or lead–lag behavior. The occurrence of significant
 280 coherence during parts of ~290–430 CE suggests time-varying associations between Börtlüce
 281 Cave hydroclimate and regional climate variability; however, differences in proxy sensitivity to
 282 climate, temporal resolution, and chronological uncertainty prevent attribution of individual
 283 variations to specific climate drivers.

284 The Tierra Blanca Joven (TBJ) volcanic eruption of the Ilopango Caldera (El Salvador) occurred
 285 at 431 ± 2 CE (Smith et al., 2020). The eruption is estimated to have produced a radiative forcing
 286 of approximately -10 W m^{-2} (Fig. 3n) and likely to have induced short-term global cooling of ~0.5
 287 °C lasting several years (Smith et al., 2020). The coincidence with reduced solar activity (Fig. 3d–
 288 e) raises the possibility that changes in regional atmospheric circulation were driven by the
 289 combined influence of volcanic and solar forcing. Years after the eruption, the Börtlüce Cave
 290 record with a solid age interval (± 8 years, 95% confidence interval), accompanied by three U-Th
 291 ages (Fig. 3, Supplementary Data) exhibits a pronounced negative excursion of $> 1\text{‰}$ in both $\delta^{18}\text{O}$
 292 and $\delta^{13}\text{C}$ values between 437 and 442 CE (Fig. 3k–l), corresponding to a sharp increase in the
 293 stalagmite growth rate from 0.6 to 3.3 mm/yr (Fig. 2d), indicating enhanced effective moisture and
 294 precipitation. The close temporal correspondence suggests that the TBJ volcanic eruption may
 295 have contributed to this hydroclimatic anomaly.

296 After the negative isotope excursion at 436–440 CE, both Börtlüce BC-4 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values
 297 rapidly increased, indicating a shift toward drier hydroclimatic conditions (Fig. 3k–l), followed by
 298 a 44-year decrease in the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values until 484 CE, which suggests a gradual return to
 299 wet conditions. Subsequently both isotope records increased to 509 CE, reflecting a 29-yr trend
 300 toward relatively dry conditions (Fig. 3k–l). This trend is interrupted by a brief negative excursion



301 at ~503 CE before both records become progressively decrease again, marking a subsequent ~20-
302 year interval of relatively wetter conditions that persisted until the end of the record at 529 CE
303 (Fig. 3k-l).

304 There is a broad agreement between the European Alps summer temperature reconstruction (Fig.
305 3b) and the BC-4 isotope records (Fig. 3k-l). Between 430 and 480 CE, the Greenland NAO index
306 (Fig. 3g) indicates a gradual transition from a positive to a negative NAO phase, followed by
307 persistently negative conditions until ~530 CE. In contrast, the Norwegian NAO index (Fig. 3f)
308 shows a general positive NAO phase during 430–530 CE and the western Mediterranean Rainfall
309 Index (Fig. 3j) exhibits an opposite pattern to the BC-4 record, with wetter (drier) conditions
310 associated with negative (positive) NAO phases corresponding to drier (wetter) conditions at
311 Börtlüce Cave. However, the limited chronological control and relatively large age uncertainties
312 of NAO reconstruction records (Fig. 3f-g) and western Mediterranean Rainfall Index (Fig. 3j)
313 prevent a robust evaluation of their temporal correspondence with the BC-4 record. During 480–
314 530 CE, the stacked Mediterranean SST gradually decreased (Fig. 3m), probably reducing regional
315 cyclone activity, corresponding to the BC-4 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ -inferred drying trend (Fig. 3k-l). These
316 observations suggest that hydroclimate in western Türkiye during 430–530 CE was primarily
317 influenced by variations in large-scale atmospheric circulation together with declining
318 Mediterranean SSTs.

319 BC-4 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records indicate that the hydroclimatic optimum between 154–290 CE,
320 corresponding to the late RWP, after which a BC-4-inferred transition interval at 290–430 CE was
321 characterized by pronounced multi-decadal-scale hydroclimatic variability, rather than a gradual
322 shift toward arid conditions, preceding the onset of the early DACP (Fig. 3k-l). This interval of
323 increasing climatic instability broadly coincides with the shift from the prosperity and territorial
324 expansion of the Roman Empire to the decline of the Western Roman Empire, culminating in its
325 collapse in 476 CE (Norman et al., 2025). We hypothesize that increasing climatic variability,
326 rather than persistent drought alone, could intensify the environmental stresses faced by Late
327 Antique societies. While the climatic deterioration alone cannot account for the decline of the
328 empire, hydroclimatic instability may have exacerbated preexisting political, economic, and social
329 vulnerabilities (Marx et al., 2018; Norman et al., 2025). The close temporal correspondence
330 between the TBJ eruption (431 ± 2 CE; Smith et al., 2020) and the onset of the early DACP
331 warrants further investigation to determine whether a causal link exists.

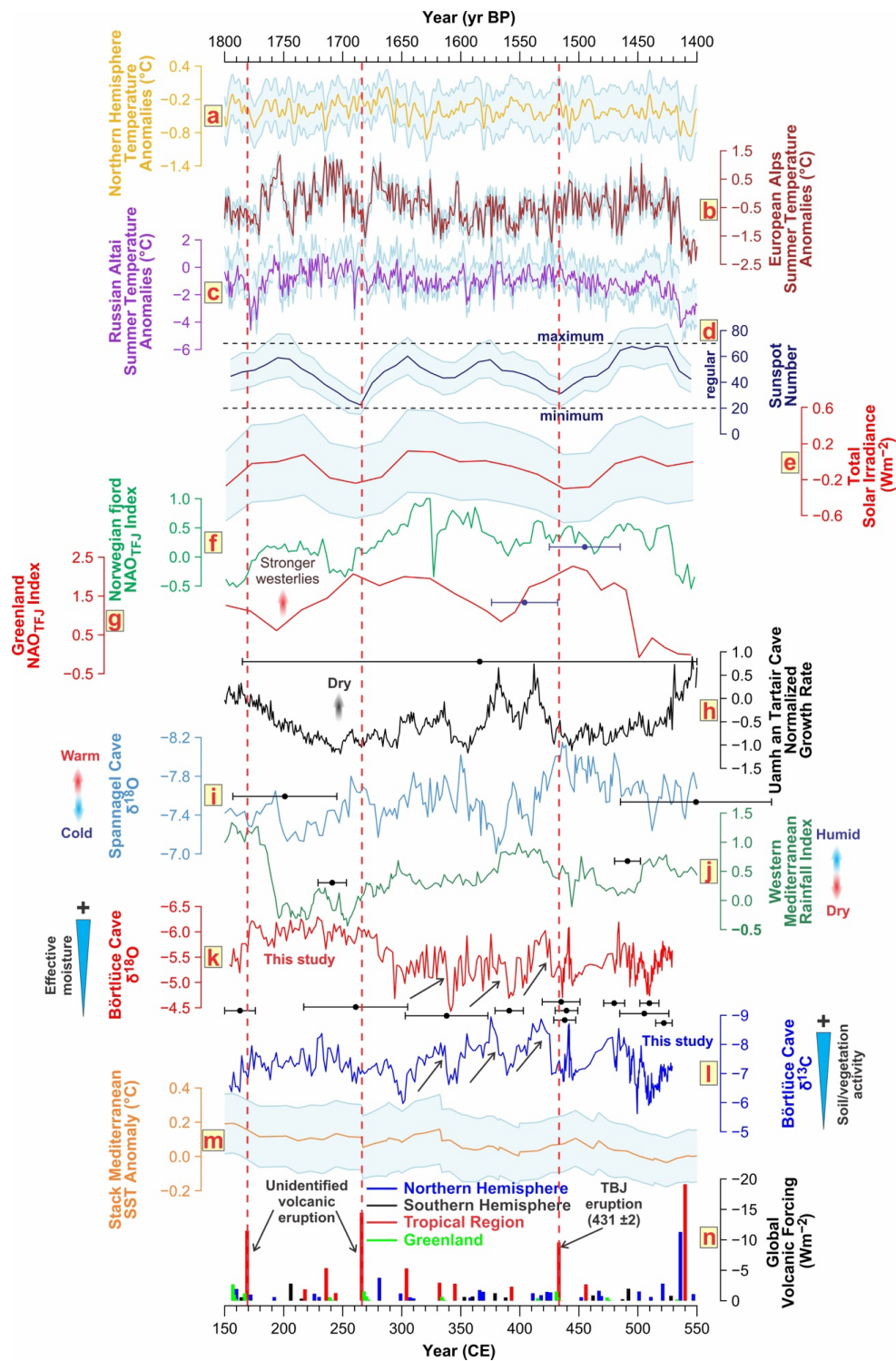




Figure 3. Comparison of stalagmite BC-4 isotope record with regional paleoclimate records and external climate forcings. **a)** Northern Hemisphere temperature anomalies (Moberg et al., 2005). **b)** Tree-ring reconstructed European Alps summer temperature anomalies (Büntgen et al., 2016). **c)** Tree-ring reconstructed Russian Altai summer temperature anomalies (Büntgen et al., 2016). **d)** Sunspot numbers, horizontal dashed lines stand for the three suggested modes: Grand minimum, Regular and Grand maximum modes (Usoskin et al. 2014). **e)** Cosmogenic radionuclide-inferred total solar irradiance reconstruction (Steinhilber et al., 2009). **f)** Marine sediment reconstruction of the NAO from Trondheimsfjord Seaward Basin, Norway (Faust et al. 2016). **g)** Lake sediment reconstruction of the NAO from southwestern Greenland (Olsen et al., 2012). **h)** Normalized growth rate from Uamh an Tartair Cave, northwest Scotland (Baker et al., 2015). **i)** Composite $\delta^{18}\text{O}$ record from Spannagel Cave, western Austria (Fohlmeister et al., 2013). **j)** Speleothem proxy-inferred Western Mediterranean rainfall index from Chaara Cave, Morocco (Ait Brahim et al., 2019). **k)** Stalagmite BC-4 $\delta^{18}\text{O}$ and **l)** $\delta^{13}\text{C}$ records from Börtlüce Cave (this study). **m)** Stack of Mediterranean Sea Surface Temperature (SST) anomalies, reconstructed from 54 marine records from 44 locations across the Mediterranean Sea (Marriner et al., 2022). **n)** Global volcanic forcing data from northern, southern hemispheres and tropical regions (Sigl et al., 2015), together with the Greenland volcanic sulfate record from the GISP2 ice core (Mayewski et al., 1997; Kobashi et al., 2013). Vertical dashed red lines represent volcanic eruptions: the Tierra Blanca Joven eruption (TBJ) of Ilopango, El Salvador in 431 ± 2 CE (Smith et al. 2020) and two unidentified volcanic eruptions in 169 and 266 CE (Sigl et al. 2015; Toohey and Sigl, 2017). Horizontal black error bars indicate U-Th ages with 2σ errors and horizontal dark blue error bars illustrate calibrated/uncalibrated radiocarbon (^{14}C) ages. Light blue shaded envelopes indicate the 95% confidence intervals for the proxy records where uncertainty estimates are available. The back arrows indicate distinct fluctuations in BC-4 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records.

5. Conclusions

Speleothem records from the Börtlüce Cave in western Türkiye exhibit hydroclimatic changes from the late RWP, through the transition period, into the early DACP between 154 and 529 CE. Higher SSTs in the Mediterranean likely increased cyclone activity and moisture transport, resulting in more winter precipitation and more effective moisture in the late RWP (154-290 CE). The reconstructed humid conditions for the late RWP generally overlap with the period of prosperity and territorial expansion of the Roman Empire. The multi-decadal hydroclimatic variability across the transition to the early DACP could have been largely driven by changes in the NAO and associated with the shifts in westerly circulation. This climatic instability after the RWP, rather than a long-term aridification trend, coincides with the decline of the Western Roman Empire, culminating in its collapse in 476 CE. Our findings indicate a complex interaction between Mediterranean SST, NAO variability, solar and volcanic forcing in controlling the complicated regional hydroclimate over the eastern Mediterranean. It also points to the need for more absolutely dated paleoclimate records to resolve spatial hydroclimatic heterogeneity, to better constrain interactions between external and internal forcings and to improve projections of future hydroclimatic changes in the Mediterranean realm.

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383 **Supplementary file**

384 Supplementary files are provided separately.

385 **Data availability**

386 All speleothem data presented in this study will be made available after the acceptance of the
387 manuscript.

388 **Competing interests**

389 The contact author has declared that none of the authors has any competing interests

390 **Author contributions**

391 M.K. developed the conceptual framework, conducted sample preparation and stable isotope
392 analyses, U-Th chemistry, performed data analysis and visualization, prepared the figures,
393 interpreted the results, and wrote the original draft of the manuscript. E.Ü.İ. reviewed and edited
394 the manuscript. M.K., C.-Y. H. and C.-C. S. performed the U-Th measurements. C.-C.S supervised
395 the study, contributed to the conceptual development, and reviewed and edited the manuscript. All
396 authors approved the final version of the manuscript.

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