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Speleothem record from Pentadactylos cave (Cyprus): new insights into climatic variations during MIS 6 and MIS 5 in the Eastern Mediterranean

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18 1- Introduction

The Eastern Mediterranean (EM) is located in a climatically sensitive region since small changes in regional atmospheric circulation may have significant and rapid changes in rainfall amount and effective infiltration (Bar-Matthews et al., 2003; Cheng et al., 2015). Whereas climatic changes in this region are reasonably known regarding spatial and temporal resolution for the Holocene (e.g. Frumkin et al., 1999, 2000; Verheyden et al., 2008a; Ayalon et al., 2002, 2013; Nehme et al., 2015; 2018; Cheng et al., 2015, Rowe et al., 2019, Keinan et al., 2019; Bar-Matthews et al, 2019), climatic information is more limited for the Last Glacial and Last Interglacial (Marine Isotope Stage- MIS 5e) and gets particularly poor prior to MIS 5e.

19 In this context, speleothems (e.g. stalagmites) are currently considered as the most suitable terrestrial
20 archives for establishing high resolution proxy time-series in paleoclimate research (Genty et al., 2003;
21 Cheng et al., 2012; Fairchild and Baker, 2012). Beyond qualitative trends, speleothems can also provide
22 quantitative estimates of paleoclimate data (temperature and rainfall amount) either through cave-specific
23 proxy calibration (Tremaine et al., 2011) or through the isotopic analysis of fluid inclusions trapped
24 within the carbonate speleothems at the time of their formation (Matthews et al., 2000; Verheyden et al.,
25 2008b; Vonhof et al., 2006; Dublyansky and Spötl, 2009). Moreover, the application of clumped isotopes
26 on associated calcite provides an additional control on the cave temperatures and its variation at the time
27 of speleothem formation (Affek et al., 2008; Meckler et al., 2015; Kluge et al., 2020).

28
29 This contribution presents a new stalagmite stable isotope data ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) from Pentadactylos cave
30 (Cyprus), which provides a high-resolution and well-dated record for the Northern Levant region covering
31 the time period from 175 to 163 ka, 141 to 132 ka and 127 to 110 ka. The combined measurements of
32 calcite $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ and fluid inclusions isotope values (δD ; $\delta^{18}\text{O}_w$), together with carbonate clumped
33 isotopes (Δ_{47}), allow the estimation of the amplitude of the temperature rise between the end of the
34 Penultimate Glacial and the Last Interglacial optimum. This record also allows to quantitatively
35 understand the variations of the calcite $\delta^{18}\text{O}_c$ values through time. The different controls on the oxygen
36 isotopic composition of the calcite ($\delta^{18}\text{O}_c$) during MIS 6 and MIS 5 (Mediterranean sapropels, rainfall
37 amount, atmospheric circulation) will also be discussed in the Eastern Mediterranean (EM) context.

38 39 2- State of the art

40 2.1. The last Glacial (G)-Interglacial (IG) in the Mediterranean region

41 The last G-IG cycle covers Marine Isotope Stages (MIS) 6 (185-135 ka; Ehlers and Gibbard, 2007) and 5
42 (130-75 ka). During Marine Isotope Substage 6e (ca.179–165 ka; Margari et al., 2010), sea-level was –40
43 to –60 m relative to present (Thompson and Goldstein, 2006). The Mediterranean basin experienced

44 overall wetter conditions than previously thought (Chedadi and Rossignol-Strick, 1995), concurrent with
45 glacial conditions in northern latitude regions (Ehlers et al., 2011). The overall wet conditions are the
46 result of enhanced ITCZ activity through an intensified monsoon activity at the Nile Headwaters and
47 concurrent with enhanced rainfall over the entire basin (Kallel et al., 2000). Although sapropel layers, an
48 organic-rich sediment, in the EM basin are generally deposited during cycles of extreme wet periods
49 (Emeis et al., 2003; Ziegler et al., 2010), the atypical sapropel S6 during the glacial MIS6 is a result of
50 freshwater input, mostly from the Nile discharge, inducing lower $\delta^{18}\text{O}$ values in the seawater and
51 consequently in EM rainwater and speleothems.

52
53 Marine cores revealed the existence of D-O-like climatic oscillations related to shifts in atmospheric
54 circulation during the early and mid-MIS 6 (~178–150 ka; Margari et al., 2010; Stocker and Johnsen,
55 2003; Barker et al., 2011). These oscillations, sensitively mark some of the terrestrial records in the
56 Central and EM basins reflecting millennial and sub-millennial rainfall and/or temperature variability
57 (Tzedakis et al., 2009; Roucoux et al., 2011; Regattieri et al., 2014; Nehme et al., 2018). The late-MIS 6
58 (ca. 150-135 ka) was generally colder with a reduction of millennial variability as the climate reached
59 more stable maximum glacial conditions (Bar-Matthews et al., 2003; Regattieri et al., 2014). This relative
60 stability ended with the deglaciation (Termination - TII, ca. 135-130 ka), which is well-recorded in south-
61 Levant speleothems (Bar-Matthews et al., 2003). Mediterranean marine cores suggest a temperature
62 change of roughly 12°C between MIS 6 and MIS 5 (Emeis et al., 2003). Villars cave speleothems in
63 North-Western Europe indicate a similar temperature difference for this deglaciation of about 13.2 to
64 $14.6 \pm 2.6^\circ\text{C}$ (1σ) (Wainer et al., 2011). To date, no terrestrial temperature estimate is available for the
65 penultimate G/IG cycle (MIS 6 to MIS 5e) in the Levant region.

66
67 The succeeding interglacial (MIS-5e) period (129-116 ka; Dutton and Lambeck, 2012; Masson-Delmotte
68 et al., 2013) is generally characterized by a minimal ice volume, with a sea-level of 4-6 m higher than
69 present (Kopp et al., 2009, Waelbroeck et al. 2002) and a global mean surface temperature of 1-2°C
70 above the Holocene reference (Otto-Bliesner et al., 2013). The Mediterranean basin experienced warm
71 conditions during the MIS 5e period with the deposition of the sapropel S5 (Rohling et al., 2002, 2015;
72 Scrivner et al., 2004) and enhanced rainfall (Kallel et al., 2000). The linkage between sapropel deposition
73 (marine environment) and the Levant terrestrial records is expressed in the similar timing of the deposits
74 with an increase in speleothem deposition (growth rate) (Ayalon et al., 2002; Bar-Matthews et al., 2003;
75 Nehme et al., 2015) occurring simultaneously with low $\delta^{18}\text{O}$ values registered in speleothems,
76 demonstrating that both land and sea systems simultaneously experienced enhanced rainfall and fresh
77 water input during this period (Bar-Matthews et al., 2019).

78
79 In the EM, the LIG (Eemian) started around 128 ka within age uncertainties (Bar-Matthews et al., 2003;
80 Tzedakis et al., 2004, 2009). The related wet phase is supported by a general increase in speleothem
81 growth rates, also observed in the Central (Drysedale, et al. 2009; Regattieri et al., 2014; Columbu et al.,
82 2017) and Western Mediterranean (Dimitriu et al., 2018; Columbu et al., 2019). Except for the Dim
83 speleothem, Southern Turkey (Rowe et al., 2020), which partially grew during the deglaciation and the
84 LIG onset until 128 ka, no highly resolved terrestrial record covers the LIG period and its demise in the
85 Northern Levant. The wet/warm conditions along the Levant seems to be relatively stable in general as
86 shown in Peqin and Soreq speleothem records (128-126 ka; Bar-Matthews et al., 2003, 2019) as well as
87 in Kanaan cave, Lebanon (Nehme et al., 2015). To check for possible regional discrepancies, related to
88 differences in location, atmospheric circulation patterns, local effect (orography, rainfall gradient) or
89 differential influence of sapropel events on terrestrial records regarding the LIG onset, length and demise
90 as well as the climatic variations during the last glacial MIS 6, it is crucial to investigate new records. The
91 new paleoclimatic data will expand the spatial resolution of the available paleoclimatic information in the
92 EM, in particular in the Northern Levant basin.

93 94 **2.2. Speleothem proxies in general and in the Mediterranean**

95 Useful proxies in the speleothems archive for paleoclimate reconstruction includes elemental ratios, $\delta^{13}\text{C}$
96 and $\delta^{18}\text{O}$ values, growth rate assessments, fluid inclusions isotopes, clumped isotopes, and others
97 (Fairchild and Baker, 2012). Speleothem growth can be a proxy by itself and has three significant
98 determinants: i) calcite saturation (Ca^{2+} and HCO_3^- -concentrations) depending on CO_2 degassing, ii)
99 temperature and iii) water supply rate (Baker et al., 1998; Dreybrodt, 1988; Genty et al., 2001). Rainfall
100 amount and temperature have an important influence on soil biological activity and consequently on CO_2
101 production driving the dissolution and precipitation processes. Therefore, warm and humid conditions are
102 generally seen as ideal for speleothem deposition (Dreybrodt, 1988; Baker and Smart, 1995; Genty et al.,
103 2006). Warm and humid conditions will increase the growth rate of speleothems. Speleothem diameter
104 can also be used to infer climatic conditions because it is related to the dripwater supply, and thus to the
105 effective infiltration, and to the saturation state (Dreybrodt, 1988, 1999). Large stalagmite diameters are
106 related to wetter periods or periods with higher calcite saturation of the water. A minimal diameter relates
107 to drier periods or periods with minimal calcite saturation.

108
109 Carbonate $\delta^{18}\text{O}$ values are amongst the most often used speleothems proxies. Unraveling the factors
110 controlling the $\delta^{18}\text{O}$ signal in speleothems, however, can be complex (McDermott, 2004; Fairchild et al.,
111 2006; Lachniet, 2009). In the Levant, more negative speleothem $\delta^{18}\text{O}$ values are associated with wetter
112 conditions, while less negative speleothem $\delta^{18}\text{O}$ is generally ascribed to drier periods with lower rainfall
113 amounts (Ayalon et al., 2002; Verheyden et al., 2008a; Cheng et al., 2015; Nehme et al. 2015, 2018; Bar-
114 Matthews et al., 2019). Important variations in $\delta^{18}\text{O}$ values have also been linked to changes in the source
115 of the water vapour and/or changes in storm trajectories (Frumkin et al., 1999; McGarry et al., 2004;
116 Kolodny et al., 2005).

117
118 Generally, variations in $\delta^{13}\text{C}$ values can be the result of several factors: soil biogenic CO_2 , a mix between
119 C3/C4 vegetation, reduced or increased soil vegetation, increased residence time and carbonate
120 dissolution, change in soil respiration rate, and/or seasonality. In the EM, $\delta^{13}\text{C}_{\text{calcite}}$ variations are mainly
121 linked to changes in soil biogenic CO_2 through changes in local effective hydrologic conditions (Frumkin
122 et al., 2000; Bar-Matthews et al., 2003; Verheyden et al., 2008a; Cheng et al., 2015).

123
124 Quantitative estimates of temperature and/or vegetation changes are generally difficult due to the
125 existence of non-equilibrium deposition processes in at least some speleothems influencing the isotopic
126 composition (Mickler et al., 2006; Affek et al., 2014). However, under continuous humid conditions
127 speleothems are more likely to be deposited close to equilibrium, whereas and during drier periods, non-
128 equilibrium formation may be significant (Verheyden et al., 2008c, Lachniet et al., 2009).

129 Another possibility for reconstruction of paleotemperature changes from speleothems is given by clumped
130 isotope analysis (Affek et al., 2008; Affek, 2012; Meckler et al., 2014). Carbonate clumped isotopes refer
131 to heavy isotopologues that include two or more rare isotopes. They are quantified with the Δ value
132 ($=R/R_{\text{stochastic}}$) that describes their overabundance relative to the stochastic distribution (Eiler, 2011).
133 Speleothem studies using clumped isotopes can be subdivided in two groups: i) direct application as
134 absolute or relative temperature proxy in case of equilibrium mineral formation (Daeron et al., 2019) or if
135 disequilibrium stays constant (Affek et al., 2008), and ii) for detection and correction of variable
136 disequilibrium (Wainer et al., 2011; Kluge and Affek, 2012; Affek et al., 2014).

137 Fluid inclusions in speleothems as remnants of ancient dripwater record the water isotopic composition
138 from which they form. Since dripwater is closely related to the surface rainwater (Genty et al., 2014;
139 Affolter et al., 2015), isotopic composition may be used to recover ancient rainwater isotopic composition
140 if calcite contains sufficient water, i.e. $>0.3 \mu\text{l}/\text{analysis}$. McGarry et al. (2004) studied fluid inclusion
141 isotope values in three caves in Israel/Palestine during the last 140 ka and found correspondences to the
142 global meteoric water line and the Mediterranean meteoric water line in varying climate stages.

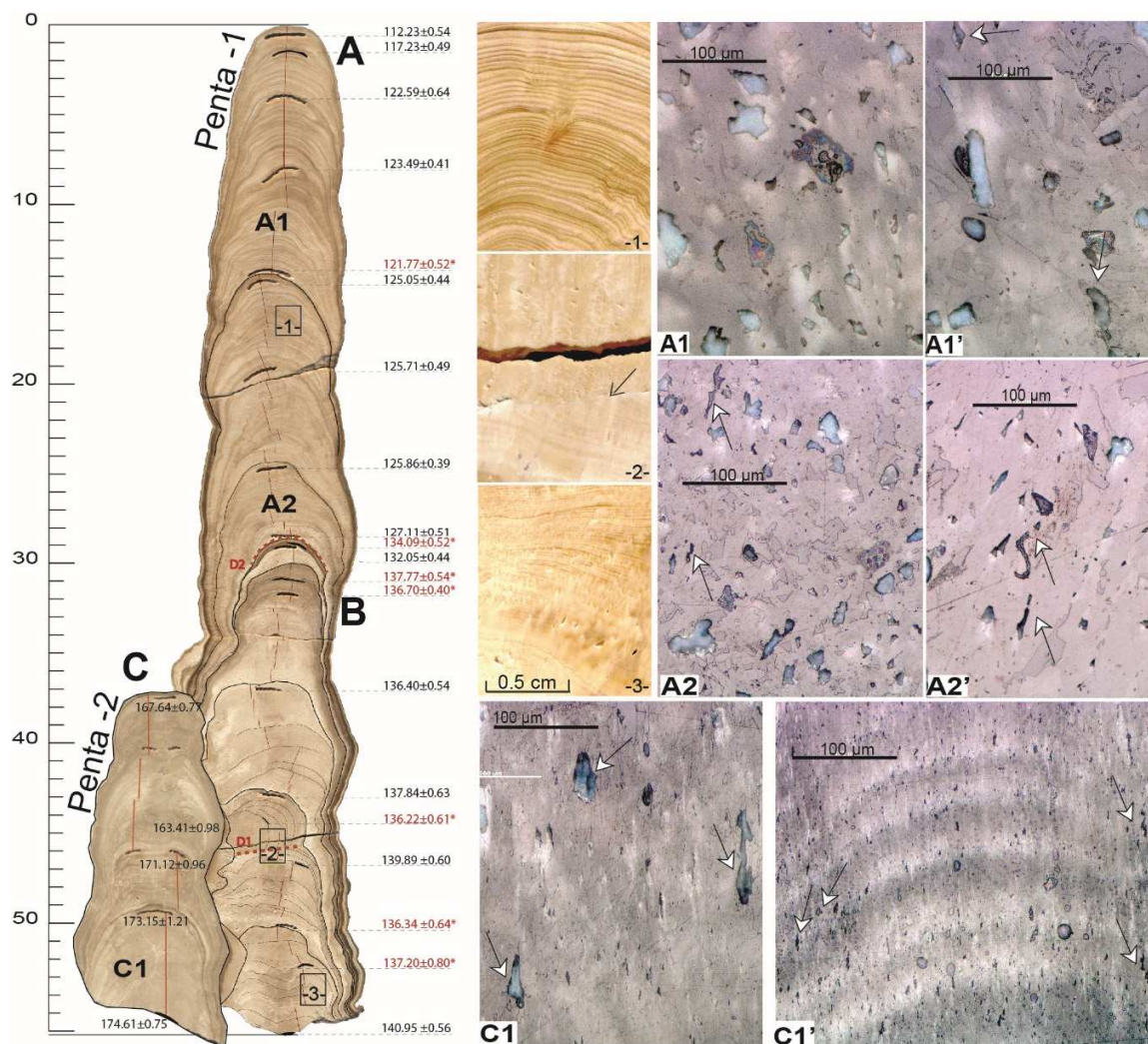
143 **3- Current climatic settings of the Eastern Mediterranean**

144 The EM climate is currently influenced by the Atlantic westerlies, which bring in moist winds associated
 145 with extra-tropical cyclones. These originate in the Atlantic and track east across the Mediterranean Sea,
 146 forming a series of sub-synoptic low-pressure systems (Alpert et al., 2005). In winter, outbreaks of cold
 147 air plunging south over the relatively warm Mediterranean Sea enhance cyclogenesis and create the
 148 Cyprus Low. These low-pressure systems drive moist air onshore, generating intense orographic rainfall
 149 across the Levant coast. The duration, intensity, and tracks of these storm systems strongly influence the
 150 amount of rainfall in this region. On the island of Cyprus, the climate is warm temperate in the western
 151 part and arid in its eastern side according to the Köppen–Geiger Climate Classification (Kottek et al.,
 152 2006). Mean monthly temperatures range from 21°C along the coast to 15°C at the Kyrenia Range. The
 153 climate is seasonal, with relatively wet winters (November to March) and dry, hot summers. The annual
 154 precipitation reaches 600 mm along the Kyrenia Mountains at 1000 m above sea-level (asl) and up to
 155 1000 mm on the Troodos (1952 m asl) summit (Cailhol et al., 2019). Both mountain ranges (Fig. 1B and
 156 C) receive orographic-driven rainfall mostly from northern and northwestern winds. Despite its northern
 157 situation in the EM, local climate is drier than the North-Levant coast where annual rainfall varies from
 158 700 to 1400 mm on Mount-Lebanon (Nehme et al., 2019).



159
 160 **Figure 1.** The Mediterranean basin and site settings. A: Location of Pentadactylos Cave and other Mediterranean palaeoclimatic
 161 records spanning Marine Oxygen Isotope Stage (MIS) 6 and MIS 5. Speleothems: Villars (Wainer et al., 2011), Crovassa Azzurra
 162 (Columbu et al., 2019), Campanet (Dimitru et al., 2018), Tana Che Urla (TCU) (Regattieri et al., 2014), Argenterola (Bard et
 163 al., 2002), Dim (Rowe et al., 2020), Peqiin (Bar-Matthews et al., 2003), Soreq (Ayalon et al., 2002), Kanaan (Nehme et al., 2015,
 164 2018). Lacustrine records: Dead-Sea/Lake Lisan (Kiro et al., 2017), the Yammouneh paleolake (Gasse et al., 2015), Lake Van
 165 (Litt et al., 2014), and the Ioaninna (Roucoux et al., 2011) and Tenaghi-Philippon (Tzedakis et al., 2003, 2009) paleo lakes.
 166 Marine cores: M40-71 (Emeis et al., 2003), ODP 968 (Ziegler et al., 2010), ODP 967 (Chedadi and Rossignol-Strick, 1995;
 167 Emeis et al., 2003; Scrivner et al., 2004), ODP 977/A (Martrat et al., 2004, 2014). B: photo of the Pentadactylos peaks and the
 168 white arrow showing the entrance to the Cave. C: map of the annual weighted mean precipitation of Cyprus (Interwies and
 169 Görnitz, 2012).

171 **4- Cave site and sample description**
 172 The Pentadactylos Cave (35°17'22.13"N, 33°28'18.49"E) is located at 800 m asl in the central part of the
 173 Kyrenia Range (1023 m altitude, at the Northern coast of Cyprus (Fig. 1B). This mountain range with a
 174 West-East direction faces the northern Levant sea-basin and include newly discovered caves developed
 175 within Triassic to Cretaceous dolomitic limestone (Cailhol et al., 2019). The cave of 200 m depth
 176 comprises few detrital deposits, non-active speleothems (draperies) and stalagmites on fallen blocks. The
 177 two adjacent Penta stalagmite samples were retrieved from a 2 m-height chamber at 50 m depth from the
 178 entrance and with. ca 200 m of limestone overburden. Relative humidity was not measured in
 179 Pentadactylos cave, but was found to be between 88.6% closest to the entrance and up to 99.9% in the
 180 deepest cave parts of Smoky Cave, another cave in the Kyrenia range at similar altitude and with similar
 181 morphology (Cailhol et al., 2019). The Kyrenia Range is covered with a typical Mediterranean sparse
 182 'garrigue' forest (i.e. pine shrubs) with sufficient soil thickness to ensure soil biological activity.



183
 184 **Figure 2.** Petrographic sections of the speleothem Penta-1 and Penta-2 showing from left to right: an image scan of the
 185 speleothem, the stable isotopes sampling track (red line), the location of the U–Th datings, location of photos 1, 2 and 3 showing
 186 the structure of the calcite laminae at 15 cm (photo 1), 46 cm (photo 2) and 53 cm (photo 3) depths. Locations of A1, A2 and C1
 187 microscope images along the speleothem growth axis, are thin section photos with an indication of the identified large fluid
 188 inclusions (arrows). Discontinuities D1 and D2 as well as ages considered outliers are marked in red.
 189

190 Stalagmites Penta-1 and Penta-2 were cut along their growth axis and petrographically studied. The
 191 Penta-1 sample is 56.1 cm long and comprises two petrographic segments A and B (Fig. 2). Segment A,
 192 i.e. the upper part of Penta-1, is 30 cm long with a regular diameter of 6 to 7 cm wide. The segment
 193 comprises a laminated structure of alternating white and brown laminae with a porous texture, a
 194 dominance of columnar fabric and no post-crystallization disturbances. Segment B, i.e. the basal part of
 195 stalagmite Penta-1, is 26.5 cm long and has a diameter of ~7 cm at the basal part to ~4 cm at the top. It
 196 displays a laminated structure with alternating white and brown laminae at its top part and some post-
 197 crystallization disturbances characterized by microcrystalline calcite and dissolution horizons, especially
 198 in the basal part (Fig. 2-photo 3). Between 43 and 46 cm depth, perturbations in the sedimentary structure
 199 are due to the interlinked growth of both stalagmites, i.e. part of the side of the Penta 2 stalagmite is
 200 imprinted in the Penta 1 stalagmite and leads to circular structures (Fig.2 - photo 2) in the longitudinal
 201 section. Penta-2 (segment C) is 28 cm long, has a regular diameter of 4-5 cm and consists of a laminated
 202 structure of alternating white and brown laminae with a porous texture. Columnar fabrics are dominating
 203 with no post-crystallization disturbances. Thin sections of 80 to 100 μm thickness of the base of segments

204 A and C show a high concentration of fluid inclusions (Fig. 2-photos A1; A2). In segment A, these
 205 inclusions are up to 40 μm long and are situated between columnar calcite crystals. In Segment C, fluid
 206 inclusions of more than 50 μm length are clearly identified with a higher concentration of inclusions
 207 along the stalagmite flanks (Fig.2-photo C1).
 208

209 Two discontinuities are observed along the growth axis of Penta-1. The discontinuity D1 at 46 cm depth
 210 is a break in the stalagmite (Fig.2, Photo 2). There, a side part of the Penta-2 stalagmite is imprinted in the
 211 Penta-1 stalagmite with re-dissolved calcite and no clear laminated structure (Fig. 2-photo 2). Another
 212 discontinuity D2 is clear at 28 cm depth with a brownish dark lamina and contrasting the lamination in
 213 part A2 of Penta-1 (Fig. 2). Discontinuity D2 is marked by a higher detrital content (Table 1).
 214

215 5-Methods

216 5.1. U-series dating

217 25 ages were determined at Xi'an Jiaotong University (China), using the procedures for uranium and
 218 thorium chemical separation and purification described in Edwards et al. (1987) and Cheng et al. (2013;
 219 2016). 100-150 mg powdered calcite samples were collected with a dental drill from 24 levels along or
 220 close to the growth axis of the speleothem, except for one sample drilled off the growth axis (Pentad-17).
 221

222 **Table 1.** Uranium-Thorium dating results. The uncertainty is given as 2σ error. U decay constants: $\lambda^{238} = 1.55125 \times 10^{-10}$ (Jaffey
 223 et al., 1971) and $\lambda^{234} = 2.82206 \times 10^{-6}$ (Cheng et al., 2013). Th decay constant: $\lambda^{230} = 9.1705 \times 10^{-6}$ (Cheng et al.,
 224 2013). $\delta^{234}\text{U} = ([^{234}\text{U}/^{238}\text{U}]_{\text{activity}} - 1) \times 1000$. $\delta^{234}\text{U}_{\text{initial}}$ was calculated based on ^{230}Th age (T),
 225 i.e., $\delta^{234}\text{U}_{\text{initial}} = \delta^{234}\text{U}_{\text{measured}} \times e^{\lambda^{234} \times T}$. Corrected ^{230}Th ages assume the initial $^{230}\text{Th}/^{232}\text{Th}$ atomic ratio of $4.4 \pm 2.2 \times 10^{-6}$
 226 corresponding to a material at secular equilibrium with the bulk earth $^{232}\text{Th}/^{238}\text{U}$ value of 3.8. The $^{230}\text{Th}/^{232}\text{Th}$ error is arbitrarily
 227 assumed to be 50%. B.P. stands for "Before Present" where the "Present" is defined as the year 1950 A.D. Samples with (*)
 228 were considered as outliers.

Sample Number	Depth mm	^{238}U (ppb)	^{232}Th (ppt)	$^{230}\text{Th}/^{232}\text{Th}$ (atomic $\times 10^{-6}$)	$\delta^{234}\text{U}$ (measured)	$^{230}\text{Th}/^{238}\text{U}$ (activity)	^{230}Th Age (yr) (uncorrected)	^{230}Th Age (yr) (corrected)	$\delta^{234}\text{U}_{\text{initial}}$ (corrected)	^{230}Th Age (yr BP) (corrected)
Pentad-14	4	134.1 \pm 0.1	1359 \pm 27	1231 \pm 25	153.3 \pm 1.2	0.7568 \pm 0.0018	112539 \pm 513	112294 \pm 540	210 \pm 2	112226 \pm 540
Pentad-24a	14	179.0 \pm 0.2	1970 \pm 39	1125 \pm 23	120.1 \pm 1.3	0.7511 \pm 0.0013	117575 \pm 448	117301 \pm 487	167 \pm 2	117232 \pm 487
Pentad-13	38	162.7 \pm 0.2	2802 \pm 56	765 \pm 15	156.6 \pm 1.3	0.7991 \pm 0.0018	123067 \pm 576	122656 \pm 643	221 \pm 2	122588 \pm 643
Pentad-12	78	161.6 \pm 0.1	614 \pm 12	3501 \pm 71	164.8 \pm 1.1	0.8063 \pm 0.0012	123250 \pm 410	123160 \pm 414	233 \pm 2	123092 \pm 414
Pentad-11*	135	158.4 \pm 0.2	798 \pm 16	2651 \pm 53	175.0 \pm 1.2	0.8095 \pm 0.0017	121964 \pm 520	121846 \pm 526	247 \pm 2	121778 \pm 526
Pentad-10	141	167.5 \pm 0.1	692 \pm 14	3277 \pm 66	175.2 \pm 1.2	0.8216 \pm 0.0013	125211 \pm 432	125114 \pm 437	249 \pm 2	125046 \pm 437
Pentad-09	189	165.5 \pm 0.1	696 \pm 14	3187 \pm 65	161.0 \pm 1.2	0.8127 \pm 0.0014	125879 \pm 482	125779 \pm 487	230 \pm 2	125711 \pm 487
Pentad-08	243	176.6 \pm 0.2	782 \pm 16	3083 \pm 63	180.4 \pm 1.2	0.8287 \pm 0.0022	126034 \pm 663	125931 \pm 667	257 \pm 2	125863 \pm 667
Pentad-07	283	136.9 \pm 0.1	1000 \pm 20	1795 \pm 36	133.0 \pm 1.1	0.7955 \pm 0.0014	127353 \pm 500	127174 \pm 515	190 \pm 2	127106 \pm 515
Pentad-19*	287	290.0 \pm 0.3	337 \pm 7	12758 \pm 267	229.7 \pm 1.3	0.8982 \pm 0.0015	134185 \pm 522	134159 \pm 522	335 \pm 2	134091 \pm 522
Pentad-06	289	286.7 \pm 0.2	257 \pm 5	16743 \pm 356	251.9 \pm 1.3	0.9097 \pm 0.0012	132338 \pm 437	132319 \pm 437	366 \pm 2	132251 \pm 437
Pentad-18*	307	347.9 \pm 0.4	693 \pm 14	7759 \pm 158	260.6 \pm 1.4	0.9372 \pm 0.0017	137880 \pm 582	137838 \pm 582	385 \pm 2	137770 \pm 582
Pentad-05*	315	304.6 \pm 0.2	391 \pm 8	11853 \pm 240	247.0 \pm 1.2	0.9218 \pm 0.0010	136798 \pm 404	136770 \pm 405	363 \pm 2	136702 \pm 405
Pentad-04	368	139.1 \pm 0.1	459 \pm 9	4351 \pm 89	186.0 \pm 1.2	0.8699 \pm 0.0015	136548 \pm 535	136473 \pm 538	273 \pm 2	136405 \pm 538
Pentad-03	428	184.4 \pm 0.1	881 \pm 18	2954 \pm 60	163.2 \pm 1.2	0.8556 \pm 0.0018	138024 \pm 627	137912 \pm 631	241 \pm 2	137844 \pm 631
Pentad-17*		119.6 \pm 0.1	239 \pm 5	7005 \pm 150	163.8 \pm 1.3	0.8506 \pm 0.0017	136336 \pm 616	136289 \pm 616	241 \pm 2	136221 \pm 616
Pentad-16	474	161.1 \pm 0.1	147 \pm 3	15670 \pm 354	170.2 \pm 1.3	0.8677 \pm 0.0016	139989 \pm 602	139967 \pm 602	253 \pm 2	139899 \pm 602
Pentad-02*	509	114.8 \pm 0.1	2462 \pm 49	659 \pm 13	165.9 \pm 1.2	0.8570 \pm 0.0021	137784 \pm 724	137281 \pm 804	244 \pm 2	137213 \pm 804
Pentad-15*	522	133.0 \pm 0.1	745 \pm 15	2554 \pm 52	183.7 \pm 1.3	0.8679 \pm 0.0018	136541 \pm 639	136412 \pm 644	270 \pm 2	136344 \pm 644
Pentad-01	557	144.8 \pm 0.1	125 \pm 3	17239 \pm 431	210.9 \pm 1.3	0.9059 \pm 0.0015	141037 \pm 564	141017 \pm 564	314 \pm 2	140949 \pm 564
Pentad-23	546	195.6 \pm 0.2	289 \pm 6	10551 \pm 216	182.3 \pm 1.2	0.9468 \pm 0.0017	163519 \pm 768	163485 \pm 768	289 \pm 2	163416 \pm 768
Pentad-22	571	190.8 \pm 0.2	248 \pm 5	11620 \pm 240	137.9 \pm 1.2	0.9158 \pm 0.0021	167748 \pm 981	167716 \pm 980	221 \pm 2	167647 \pm 980
Pentad-21	625	239.5 \pm 0.2	441 \pm 9	8036 \pm 163	109.5 \pm 1.2	0.8968 \pm 0.0018	171239 \pm 961	171194 \pm 961	177 \pm 2	171125 \pm 961
Pentad-20	656	150.2 \pm 0.2	963 \pm 19	2474 \pm 50	172.5 \pm 1.6	0.9624 \pm 0.0024	173368 \pm 1209	173222 \pm 1212	281 \pm 3	173153 \pm 1212
Pentad-00	711	109.0 \pm 0.1	167 \pm 4	10485 \pm 224	181.6 \pm 1.2	0.9744 \pm 0.0014	174720 \pm 751	174685 \pm 751	297 \pm 2	174617 \pm 751

229 The data were obtained on a Thermo Neptune Plus multicollector inductively coupled plasma mass
 230 spectrometers (MC-ICP-MS) following procedures modified from Cheng et al. (2013). Ages are
 231 calculated from time of analysis (2018) with an uncertainty at the 2σ level, typically between 400 and
 232 1000 a (Table 1). The stratigraphic uncertainties are less than 1.5 mm. The age model was constructed
 233 using the Constructing Proxy Records from Age models (COPRA) routine (Breitenbach et al., 2012),
 234 based on polynomial interpolation and 5000 Monte Carlo simulations for each segment of the stalagmite
 235 (Fig. 3). The polynomial function generates a monotonously positive depth-age slope (Breitenbach et al.,
 236 2012) with the assumption that all dates should be in chronological order (within errors). The median and
 237
 238

239 2.5% and 97.5% confidence intervals are based on 5000 Monte Carlo simulations. COPRA handles
240 multiple hiatuses, as long as each is bracketed by at least two dates, which is the case for this sample.

241 **5.2. Calcite and water stable isotopes measurements**

242 739 speleothem calcite samples with 125 duplicates (92% replicability) were sampled along the growth
243 axis (Fig. 2) for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ measurements and were drilled along the speleothem central axis with a
244 0.5-1 mm resolution using a Merchantek Micromill mounted on a Leica microscope. Between every
245 sample, the drill bit and sampling surface were cleaned with compressed air. The samples were analyzed
246 together with modern calcite samples using either a Nu Carb carbonate device coupled to a Nu
247 Perspective MS at the Vrije Universiteit Brussel or a Thermo Gas Bench II connected to a Thermo Delta
248 Advantage mass spectrometer in continuous flow mode at Karlsruhe Institute of Technology (Germany).
249 Analytical uncertainties were better than 0.1‰ (1 σ) for oxygen and 0.05‰ (1 σ) for carbon on both
250 instruments. All $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values are calibrated against Vienna Pee-Dee Belemnite (V-PDB) and are
251 reported in permil (‰) relative to VPDB. The stalagmite was tested for isotopic equilibrium conditions
252 by carrying out Hendy tests (Hendy, 1971) at twelve different heights along the speleothem axis
253 (Appendix S3).

254 The Pentadactylos Cave showed insufficient active water seepage to sample during the fieldwork.
255 Therefore, nine seepage and water samples from nearby caves and springs were collected in December
256 2018 for $\delta^{18}\text{O}$ and δD measurements in hermetically sealed glass bottles. Isotope measurements were
257 performed at the Vrije Universiteit Brussel on a Picarro L2130-i analyzer using the cavity ring-down
258 spectroscopy (CRDS) technique (Van Geldern and Barth Johannes, 2012). All values are reported in
259 permil (‰) relative to Vienna standard mean ocean water (VSMOW). Analytical uncertainties (1 σ) were
260 better than 1‰ for δD and 0.1‰ for $\delta^{18}\text{O}$.

262 **5.3. Fluid inclusions stable isotope (H-O) analyses**

263 21 calcite samples, from 9 levels in Penta-1 and Penta-2 were measured using a custom-built extraction
264 line connected to a (Picarro L2130i) analyzer using cavity ring down spectroscopy (WS-CRDS). This
265 technique allows to simultaneously measure hydrogen and oxygen isotopes for minute water amounts
266 released from calcite. The extraction line follows the design of Affolter et al. (2014) and is described in
267 detail in Weißbach, 2020). In brief, the calcite samples were hydraulically crushed and the released fluid
268 inclusion water instantly vaporized in the heated extraction system and transferred to the analyzer.
269 Reference water injections with known $\delta^{18}\text{O}$ and δD values were used for calibration and quality control.
270 μl -sized glass capillaries were used for high-precision water amount calibration (on the order of 10^{-8}l) and
271 isotopic control. The overall precision of fluid inclusion water analyses is 0.5‰ for $\delta^{18}\text{O}$ and 1.5‰ for
272 $\delta^2\text{H}$ if fluid water amounts are 0.5-1 μl and 0.25‰ for $\delta^{18}\text{O}$ and 1‰ for $\delta^2\text{H}$ for fluid water amounts >1 μl
273 (Weißbach, 2020).

275 **5.4. Clumped Isotopes**

276 Clumped isotope measurements were performed on aliquots of the powder (>100 mg each) produced
277 during the fluid inclusion analysis of samples 7b, 6b, 4b and 3 (Table 2). The sample treatment follows
278 Weise and Kluge (2020). Calcite samples of ca. 3 mg were individually reacted with 105% phosphoric
279 acid (~1 ml / sample) in a stirred glass reaction vessel for 10 min at 90°C. The emerging CO_2 was
280 continuously collected with a liquid- N_2 cooled trap and subsequently cryogenically cleaned (separation of
281 volatile gases from this trap at liquid nitrogen temperature). Subsequently, water was separated from the
282 remaining gas using a dry-ice ethanol cooled glass trap. The water-free CO_2 gas was then passively
283 passed through silver wool and another trap filled with Porapak Q and held at -35 °C (transit time 33
284 min). The cleaned CO_2 gas was analyzed immediately or within a few hours.

285 Mass spectrometric measurements were performed on a Thermo Scientific MAT 253 Plus with a baseline
286 monitoring cup on m/z 47.5 and $10^{13}\Omega$ resistors on m/z 47-49. The analysis protocol consisted of 8
287 acquisitions with 10 cycles each and an integration time for each cycle of 26 s. Each acquisition included
288 a peak center, background measurements and an automatic bellows pressure adjustment aimed at a 6 V

289 signal at mass 44. The first acquisition additionally included a recording of the sample m/z 18 (water
290 vapor residual) and m/z 40 signal (Ar – indicator for air remainder). For each cycle the baseline signal on
291 m/z 47.5 was measured simultaneously to the actual sample and reference gas analysis on m/z 44-49. For
292 pressure-baseline (PBL) correction high-voltage peak scans were manually taken at the beginning and/or
293 end of a measurement run. The sample gas was measured against an in-house reference gas standard
294 (Oberlahnstein: $\delta^{13}\text{C} = -4.42 \text{‰ VPDB}$, $\delta^{18}\text{O} = -9.79 \text{‰ VPDB}$). For establishing an absolute reference
295 frame and for interlaboratory comparability, we regularly analyzed community-wide distributed
296 carbonates (ETH1-4, Meckler et al., 2014), Carrara Marble and equilibrated gases (5°C, 25°C, 90°C).
297 All data were evaluated using a custom-written R script that includes an automatic PBL correction, taking
298 into account that the signal on m/z 47 is additionally influenced by a negative background potentially
299 induced by secondary electrons and broadening of the m/z 44 peak (He et al., 2012; Bernasconi et al.,
300 2013; Fiebig et al., 2015). The empirical transfer function (ETF) was determined based on PBL-corrected
301 Δ_{47} values and uses carbonate standards and equilibrated gases with agreed Δ_{47} values as reference
302 (Dennis et al., 2011). Updated isotope parameters following Daëron et al. (2016) were used for the
303 evaluation.
304

305 **5.5. Data correction clumped isotopes and fluid inclusions**

306 *a) Clumped isotopes*

307 Speleothem calcite isotopes may be influenced by disequilibrium effects. Δ_{47} values are sensitive
308 recorders of kinetic or disequilibrium effects and can be used to assess its magnitude via the difference
309 from the equilibrium Δ_{47} value that is solely determined by temperature (Kluge and Affek, 2012).
310 Disequilibrium in Δ_{47} and $\delta^{18}\text{O}$ seems to be linked and leads to a co-variation that is at least consistent at
311 specific cave sites or experimental setups (Daëron et al., 2011; Kluge et al., 2013). Deviations to smaller
312 Δ_{47} values and apparently too high temperatures indicate disequilibrium due to evolution of the mineral-
313 precipitating solution on top of the stalagmite (Hansen et al., 2019) related to prior carbonate precipitation
314 and/or CO_2 degassing (Guo, 2020). The knowledge of the disequilibrium in Δ_{47} , thus, provides also the
315 disequilibrium in calcite $\delta^{18}\text{O}$ via the Δ_{47} - $\delta^{18}\text{O}$ co-variation slope and allows for its correction (Appendix
316 S1), complementing thus Hendy tests (Daëron et al., 2011; Kluge and Affek, 2012; Kluge et al., 2013).
317 The Δ_{47} - $\delta^{18}\text{O}$ correction as outlined in Kluge et al. (2013) for paleo-drip water $\delta^{18}\text{O}$ determination is used
318 based on independent paleo-temperature estimates from Emeis et al. (2003) for the EM Sea surface and
319 scaled to the cave location. Together with the Δ_{47} - $\delta^{18}\text{O}$ co-variation slope of -0.047‰‰ of Kluge et al.
320 (2013), the observed disequilibrium in Δ_{47} is transferred into calcite $\delta^{18}\text{O}$ offsets.
321

322 *b) Fluid inclusion isotope ratios*

323 Fluid inclusions are robust archives for paleowater and can be used to calculate the calcite formation
324 temperature from the fractionation $^{18}\epsilon_{\text{calcite-water}}$. We determined the temperature from $^{18}\epsilon_{\text{calcite-water}}$ with the
325 kinetic-corrected calcite $\delta^{18}\text{O}$ (using Δ_{47} ; Supplementary S1) and the fluid inclusion $\delta^{18}\text{O}$, applying the
326 experimental calibration of Kim and O'Neil (1997) for $^{18}\epsilon_{\text{calcite-water}}$. However, for some samples additional
327 effects need to be considered. Seasonal or temporary drying of the stalagmite surface can significantly
328 increase water $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values before full fluid inclusion close-off. The fluid $\delta^{18}\text{O}$ value could
329 therefore be out of equilibrium with the surrounding calcite that was precipitated earlier from the non-
330 evaporated fluid. Evaporative influence becomes visible if isotope ratios significantly deviate from the
331 Meteoric Water Line (following an evaporation line towards higher $\delta^{18}\text{O}$; Weißbach, 2020;
332 Supplementary S3). We assessed the measured fluid inclusion $\delta^{18}\text{O}$ values for evaporative effects by
333 comparison with the Mediterranean Meteoric Water Line (MMWL). Note that the MMWL as reference
334 line for the infiltrating karst water may not be fully adequate during glacial periods. McGarry et al. (2004)
335 noted a shift towards the GWML in their fluid inclusion isotope study for some glacial samples. Some of
336 the glacial samples in our study are close to the GWML. Note, however, that the related $^{18}\epsilon_{\text{calcite-water}}$ of
337 those samples correspond to an unrealistically high nominal temperature (Table 4), indicating that effects
338 other than a shift in the meteoric water line (e.g., evaporation; Weißbach, 2020), may explain the
339 deviation from the MMWL. Thus, our approach of using the MMWL as reference is a first-order

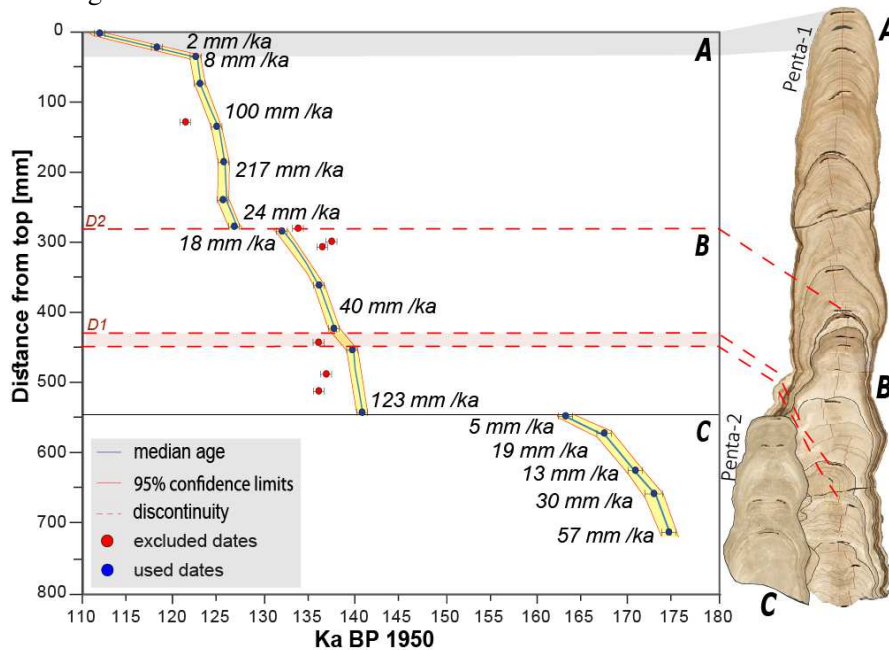
340 approach for separating fluid inclusions with likely undisturbed meteoric signal from samples influenced
 341 by evaporation.

342

343 6- Results

344 6.1. Chronology

345 25 Uranium/Thorium ages were obtained from the stalagmite Penta-1 (Parts A and B) and Penta-2 (Part
 346 C) (Table 1). The age distribution based on the COPRA age model indicates that the stalagmite Penta 2
 347 grew from ca. 174.3 ± 0.6 to 163.4 ± 0.9 ka with a constantly decreasing growth rate (Fig 3). Penta-1
 348 grew from 140.9 ± 0.4 to 112.2 ± 0.5 ka including two discontinuities. One noteworthy hiatus or period
 349 with a much lower growth rate from 132.2 ± 0.4 to 127.1 ± 0.5 , corresponds to discontinuity D2. The
 350 discontinuity D1 covering the period 139.9 ± 0.6 to 137.8 ± 0.6 ka was not investigated further due to
 351 stratigraphical disturbances between the 2 adjacent stalagmites. In the upper part of Penta-1 (Segment A),
 352 dated between ~ 127 and ~ 112 ka, the age model, that was constructed with 8 ages (Fig. 3) displays two
 353 distinct periods of relatively constant growth rates: an upper part from 40 mm to top with a lower growth
 354 rate of less than 8 mm/ka and the part from 270 to 40 mm with a growth rate of 24 to 217 mm/ka. The
 355 lower part of Penta-1 (segment B) covers from 140.9 ± 0.5 ka to 132.2 ± 0.4 up to the second
 356 discontinuity D2. All ages in Penta-2 (segment C) are in good stratigraphic order and were used in the
 357 calculation of the age model.



358

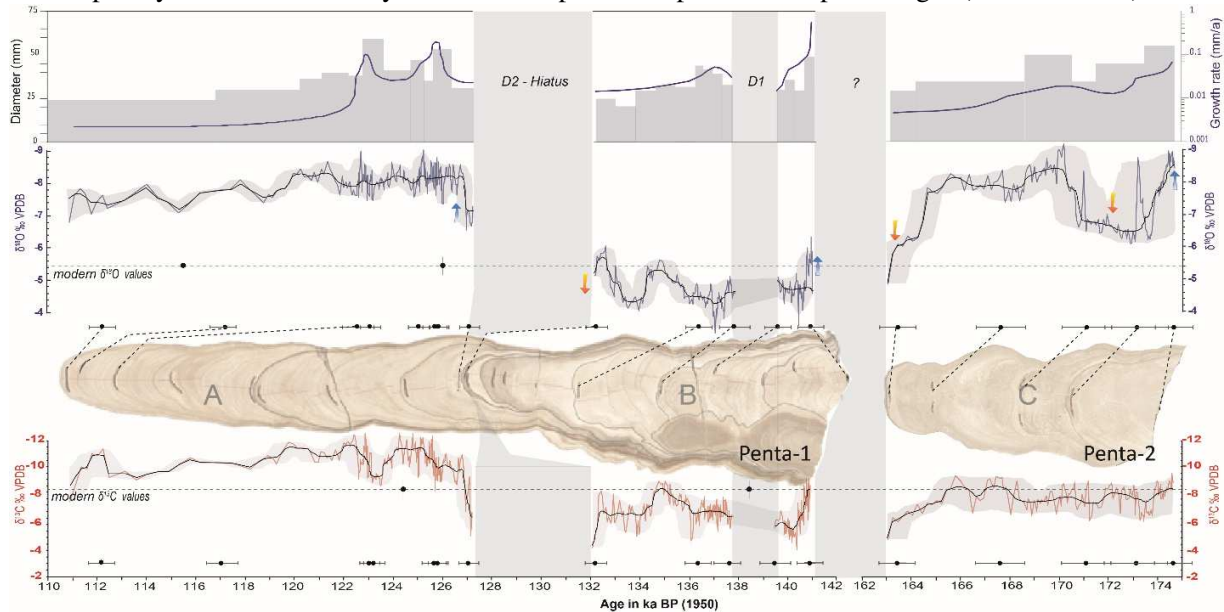
359 **Figure 3.** Age model of the *Pentadactylos* record. Growth rate of the stalagmite is given next to the age model using COPRA
 360 polynomial statistics model between two consecutive dates.

361

362 6.2. Stable isotopic composition of calcite

363 The $\delta^{18}\text{O}$ (-3.3‰ to -9.2‰) and $\delta^{13}\text{C}$ values (-4.2‰ to -12.6‰) (Fig. 4) of the *Pentadactylos* stalagmite
 364 roughly display distinct signatures corresponding to the three segments A, B and C as determined based
 365 on the sedimentological profile with segments A (from ~ 110 to ~ 127 ka) and C (from ~ 163 to ~ 174 ka)
 366 showing relatively lower $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values than segment B (from 132 ka to 141 ka). The overall
 367 resolution of the isotope record is multi-annual (3 to 9 a/mm) to multi-decadal (15 to 60 a/mm), except for
 368 the top most part (from ~ 122 to ~ 112 ka), which is at lower resolution (270 a/mm). Secondary centennial
 369 negative $\delta^{18}\text{O}$ excursions of 2 to 3‰, as for example at ~ 170 ka or ~ 165 ka (Fig. 4) are observed, mainly
 370 in segment C, while $\delta^{13}\text{C}$ values are varying rather smoothly on a millennial scale without important
 371 centennial excursions.

372 The $\delta^{13}\text{C}$ values generally co-vary, at millennial scale, with the $\delta^{18}\text{O}$ values. Growth rate generally vary
 373 simultaneously with the $\delta^{18}\text{O}$ but especially with the $\delta^{13}\text{C}$ values (Fig. 4). A high growth rate generally
 374 corresponds to more negative $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values. Noteworthy are the two growth rate peaks at ~126
 375 and ~123 ka occurring together with more negative $\delta^{18}\text{C}$ values. After ~122 ka, the growth rate decreases
 376 drastically to the top of segment A coeval with a slowly increasing trend in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values.
 377 Hendy tests (Hendy, 1971) performed along twelve different growth layers show a variation of less than
 378 0.5‰ for $\delta^{18}\text{O}$ values and not more than 1‰ for $\delta^{13}\text{C}$ values, without clear covariation (Appendix S3). A
 379 potential exception could be the samples at 3 cm and 33 cm from top. Thus, kinetic isotope effect cannot
 380 be completely excluded but likely have not overprinted important isotopic changes (cf. section 4.4).



381 **Figure 4.** $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ profiles of micro-drilled samples along the growth axis of Pentadactylos Cave stalagmite (Cyprus)
 382 reported with their radiometric age. The original isotopic data are presented with their moving average (black line) to clearly
 383 show the general trend in isotopic variations from 163 to 174 ka (Penta-2) and from 112 to 141 ka (Penta-1). The period from
 384 138 to 140 ka is not displayed as no isotopic measurements were made for this period. The isotopic changes are displayed
 385 relative to V-PDB (Vienna Pee Dee belemnite) together with growth rate (mm/a) and stalagmite diameter (mm). The modern
 386 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values represent calcite samples retrieved under active seepage water from the cave. Blue arrows show wet phases
 387 and yellow arrows refer to dry phases.
 388
 389

390 6.3. Stable isotopic composition of modern water and calcite fluid inclusions

391 The average fluid inclusion value is -4.9‰ for $\delta^{18}\text{O}_w$ and -26.5‰ for $\delta^2\text{H}_w$. Replicate analysis shows
 392 consistent results at each level, except for the Penta 1-1 samples (56.1 cm from top, 140.9 ka). During
 393 MIS6e, $\delta^{18}\text{O}_w$ varies between $\sim -6.0\text{‰}$ and -1.5‰ , with least negative values at the end of MIS6 (140.9
 394 ka). At 137.1 ka, more negative values are set around -4.0‰ , leading up to the most negative values
 395 during MIS5e: -4.8‰ at 127.0 ka and reaching -8.6‰ at 126.7 ka and -6.8‰ at 125.6 ka. The $\delta^2\text{H}_w$ values
 396 vary between -12‰ and -30‰ during MIS6 and TII, whereas $\delta^2\text{H}_w$ values shift to $\sim -45.0\text{‰}$ at ~ 126 ka
 397 (Table 2). On a comparative basis, present cave and spring water were sampled in December 2018 along
 398 the Kyrenian mountain range with -6.6‰ for $\delta^{18}\text{O}_w$ and -32.9‰ for $\delta^2\text{H}_w$. There is a general trend towards
 399 more negative $\delta^{18}\text{O}_w$ in spring and drip water with increasing altitude from the northern coast to the
 400 Kyrenia summits. From the 21 fluid inclusions samples measured, 4 results are considered outliers. Fi-
 401 Penta1_1c; Fi_Penta1_3b and Fi_Penta1_8a gave insufficient water (below $0.3\mu\text{l/g}$; Table 2) for reliable
 402 analysis. Fi-Penta1_1a is related to calcite that was taken at the outer flank of the stalagmite and may
 403 therefore be biased due to evaporative enrichment along the flow path.

404 **Table 2.** Pentadactylos fluid inclusions samples in chronological order (old to young) in comparison with modern cave drip and
 405 spring water (bottom table). Measurements with an (*) are considered outliers.

Sample ID	cm/top	U-Th age (ka)	$\delta^2\text{H}_w$ [‰]	1 σ error [‰]	$\delta^{18}\text{O}_w$ [‰]	1 σ error [‰]	volume [μl]	mass [g]	Water amount [$\mu\text{l/g}$]
Fluid inclusions									
F1_Penta2_2a	17.5	174.5	-24.39	1.5	-5.43	0.5	2.26	0.41	5.52
F1_Penta2_2b			-26.14	1.5	-5.46	0.5	0.83	0.41	2.02
F1_Penta2_2t			-27.10	1.5	-5.91	0.5	1.14	0.30	3.79
F1_Penta2_4a	10	170.4	-29.81	1.5	-5.97	0.5	0.74	0.31	2.37
F1_Penta2_4b			-27.61	1.5	-5.67	0.5	1.26	0.34	3.70
F1_Penta2_5a	1	163.9	-16.88	1.5	-3.72	0.5	1.45	0.30	4.84
F1_Penta2_5b			-17.12	1.5	-3.86	0.5	0.84	0.24	3.48
F1_Penta1_1c*	56.1	140.9	-19.74	1.5	-3.77	0.5	0.23	0.30	0.78
F1_Penta1_1a*			-20.57	1.5	-6.19	0.5	0.77	0.54	1.43
F1_Penta1_1b			-12.64	1.5	-1.79	0.5	0.78	0.48	1.62
F1_Penta1_1d			-11.36	1.5	-1.61	0.5	0.69	0.47	1.46
F1_Penta1_1t			-20.94	1.5	-4.02	0.5	0.43	0.30	1.42
F1_Penta1_6a	40.5	137.1	-20.54	1.5	-3.60	0.5	0.31	0.28	1.11
F1_Penta1_6b			-21.39	1.5	-4.50	0.5	0.68	0.50	1.37
F1_Penta1_9a	27.2	126.9	-25.43	1.5	-4.83	0.5	0.65	0.31	2.11
F1_Penta1_9c			-24.59	1.5	-4.78	0.5	0.77	0.30	2.57
F1_Penta1_7b	26	126.7	-45.45	1.5	-9.04	0.5	0.64	0.30	2.12
F1_Penta1_7d			-38.86	1.5	-8.29	0.5	0.58	0.33	1.75
F1_Penta1_3b*	14	125.6	-48.47	1.5	-4.86	0.5	0.17	0.31	0.55
F1_Penta1_3t			-43.91	1.5	-6.81	0.5	0.31	0.31	1.00
F1_Penta1_8a*	3	117	-43.99	1.5	-4.32	0.5	0.27	0.32	0.84
Modern cave & spring waters									
		<i>altitude (m asl)</i>							
Smoky cave (mix pool & drip)		756	-20.29	1	-5.24	0.1			
Hot cave_1 (drip)		611	-36.49	1	-7.31	0.1			
Hot cave_2 (drip)		611	-39.51	1	-7.58	0.1			
Kalga spring		505	-33.97	1	-6.67	0.1			
Karagaac spring		289	-32.52	1	-6.51	0.1			
K.Evenkous spring			-35.30	1	-7.08	0.1			
Lepta spring (in-situ)		91	-34.77	1	-6.66	0.1			
Sulfur spring		25	-27.77	1	-5.28	0.1			

406 6.4. Clumped isotopes

407 Clumped isotope analysis of the four samples yielded apparent temperatures of 24-29°C (using the Δ_{47} -T
408 relationship of Kluge et al., 2015) that are well above the current cave temperature of 15°C (Table 3),
409 suggesting significant kinetic effects. Offsets of Δ_{47} -related temperatures of 10-15°C are within the range
410 of previous studies (e.g., Daëron et al., 2011; Kluge and Affek, 2012; Kluge et al., 2013; Affek et al.,
411 2014). Following the method outlined in section 5.5, the correction for the kinetic contribution is
412 completed with an effect on the calcite $\delta^{18}\text{O}$ value of +0.4-0.9 ‰. The kinetic correction of the calcite
413 $\delta^{18}\text{O}$ value is relatively insensitive to the cave paleo-temperature estimates. A deviation of 3°C would
414 change the kinetic $\delta^{18}\text{O}$ correction by a maximum of 0.2‰.

415 Drip water $\delta^{18}\text{O}$ values that are related to the formation of the corresponding growth layer were calculated
416 from the corrected calcite $\delta^{18}\text{O}$ and an independent temperature estimate. These were scaled to cave
417 temperatures from Sea Surface temperature (SST) of Emeis et al., (2003): 14, 9, 13 and 16.7°C for 126.7,
418 137, 170.4, and 125.6 ka, respectively. The reconstructed water $\delta^{18}\text{O}$ values range from -6.6 to -9.4 ‰
419 (Table 3) using the oxygen isotope fractionation factor of Kim and O'Neil (1997). These values are
420 similar or more negative than the modern dripwater values (Table 2).

422 6.5 Paleotemperatures

423 Paleotemperature calculation is based on the fractionation $^{18}\epsilon_{\text{calcite-water}}$ using the kinetic-corrected calcite
424 $\delta^{18}\text{O}$ (using Δ_{47}) and the fluid inclusion $\delta^{18}\text{O}$ (section 5.5 and Appendix S1). $^{18}\epsilon_{\text{calcite-water}}$ is then converted
425 into temperature applying the experimental calibration of Kim and O'Neil (1997). Calculated
426 temperatures from $^{18}\epsilon_{\text{calcite-water}}$ range from $14.7 \pm 2.7^\circ\text{C}$ to $30.6 \pm 2.7^\circ\text{C}$ (Table 3). Warmest temperatures
427 are obtained for ~170 ka and ~126 ka, while coldest values are obtained for ~127-141 ka. In contrast to
428 mid- and high latitudes, inclusion $\delta^{18}\text{O}_w$ and δD_w cannot be used for direct temperature calculation in the

429 Mediterranean due to an inconsistent or only weak relationship of mean annual rainfall $\delta^{18}\text{O}$ and
 430 temperature.

431
 432 **Table 3:** *Pentadactylos clumped isotope samples in chronological order (old to young). Calcite powder was retrieved from the*
 433 *remaining powder of the fluid inclusion analysis. The apparent T is based on the Δ_{47} -T calibration of Kluge et al. (2015). n refers*
 434 *to the number of replicates measured per sample. 4b refers to Penta 2, 6b,7b and 3 to Penta 1. The paleotemperature related to*
 435 *each sample was calculated based on $^{18}\text{E}_{\text{calcite-fluid}}$ following Kim and O'Neil (1997), including corrections for kinetics based on*
 436 *the clumped isotope (see section 5.5 and appendix S1). Uncertainties are based on analytical uncertainties ($\delta^{13}\text{C}$, $\delta^{18}\text{O}$, Δ_{47}) or*
 437 *Gaussian error propagation (calculated dripwater $\delta^{18}\text{O}$ and reconstructed paleotemperature).*

Sample ID	cm from top	U/Th age (ka)	$\delta^{13}\text{C}_{\text{cc}}$ (‰)	$\delta^{18}\text{O}_{\text{cc}}$ (‰)	Δ_{47} (‰)	Apparent Δ_{47} -T (°C)	Calc $\delta^{18}\text{O}_{\text{drip}}$ (‰)	T (°C) from $^{18}\text{E}_{\text{calcite-water}}$	n
4b	10 - Penta 2	170.4	-7.1 ± 0.1	-7.7 ± 0.1	0.700 ± 0.007	24 ± 2	-8.5 ± 0.3	25.5 ± 2.8	3
6b	40.5 - Penta 1	137.1	-7.18 ± 0.02	-4.68 ± 0.04	0.701 ± 0.011	23.5 ± 4	-6.6 ± 0.4	17.2 ± 2.7	2
7b	26 - Penta 1	126.7	-9.5 ± 0.1	-8.5 ± 0.1	0.686 ± 0.017	29 ± 6	-9.4 ± 0.5	14.7 ± 2.4	3
3	14 - Penta 1	125.6	-10.67 ± 0.03	-8.6 ± 0.2	0.699 ± 0.004	24 ± 1	-8.4 ± 0.3	Offset from MMWL	3

438
 439 **7- Discussion**
 440 **7.1. Age model**
 441 COPRA succeeded to produce a coherent age model, although in total seven points out of twenty-five
 442 were considered as outliers. A few minor age inversions in growth phases with porous stalagmite parts
 443 (Table 1) may be explained by a minor degree of open system behavior. In contrast, there are no
 444 indications of detrital Th. Petrographical observation shows that section B (base of Penta 1) is the most
 445 porous part of the whole record and where most of the age inversions occurs (Fig. 2). Porous calcite
 446 structures are known to be subject of U–Th open-system behavior (Richards and Dorale, 2003; Borsato et
 447 al., 2005) *via* dissolution and recrystallization processes of the primary carbonate (Frisia, 2015),
 448 remobilization, or post-depositional filling of pores (Schwarcz, 1989), resulting in out-of-sequence ages.
 449 Dating points Pentad-2, -5, -14 and Pentad-18 were therefore excluded from the age model. Pentad-19
 450 and 17 are closest to the D2 and D1 respectively, showing stratigraphic perturbations with porous
 451 structures and were therefore discarded.

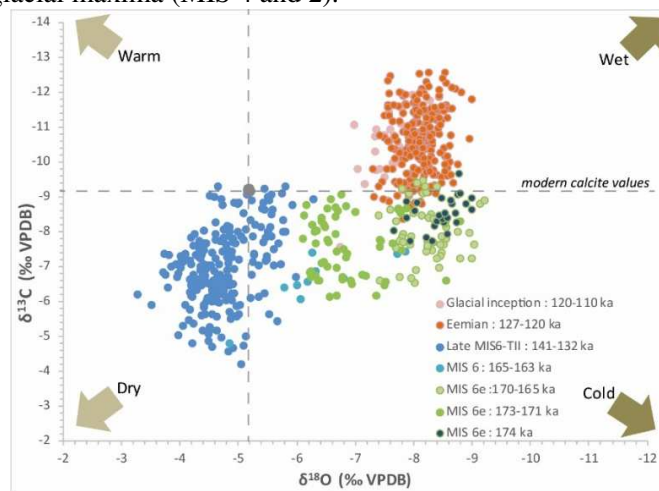
452 The age model reveals a rather continuous speleothem deposition with two hiatuses in agreement with the
 453 observed discontinuities. Two growth stops are noted from ~163 to ~141 ka and from ~132 to ~128 ka.
 454 The first growth stop covering the mid-to late penultimate glacial, although consistent generally with a
 455 dry period, corresponds to a time gap between the two adjacent stalagmites Penta 1 and Penta-2. The stop
 456 in growth may be related to a real climatic deterioration or to a local perturbation in the water routing
 457 finally leading to the deposition of the second stalagmite adjacent to the first one. Without replication by a
 458 second stalagmite to confirm the cause for this growth stop, one must remain cautious on the possible
 459 reasons for this growth interruption. The second growth stop in Penta-1 stalagmite, from ~132 to ~128 ka,
 460 corresponds to the discontinuity D2 and is considered as hiatus or period with a much lower growth rate.
 461 The discontinuity D1 from 139 to 137 ka is clearly due to the stratigraphical disturbances of the interface
 462 between the 2 adjacent stalagmites and is therefore not considered as a hiatus.

464 7.2. General climatic conditions of the last Glacial/Interglacial cycle in Cyprus

465 The isotope values of both stalagmites (Fig. 5) roughly suggest three ‘climate-related regimes: i) less
 466 negative $\delta^{13}\text{C}$ (~-8‰) and negative $\delta^{18}\text{O}$ values (~-8‰) during the early-penultimate Glaciation (with
 467 exception for 173-170 ka and after 165 ka), ii) comparatively less negative $\delta^{13}\text{C}$ (~-6‰) and $\delta^{18}\text{O}$ values
 468 (~-5‰) during the late-penultimate Glaciation and iii) negative $\delta^{13}\text{C}$ (~-11‰) and $\delta^{18}\text{O}$ values (~-8‰)
 469 during the last interglacial period.

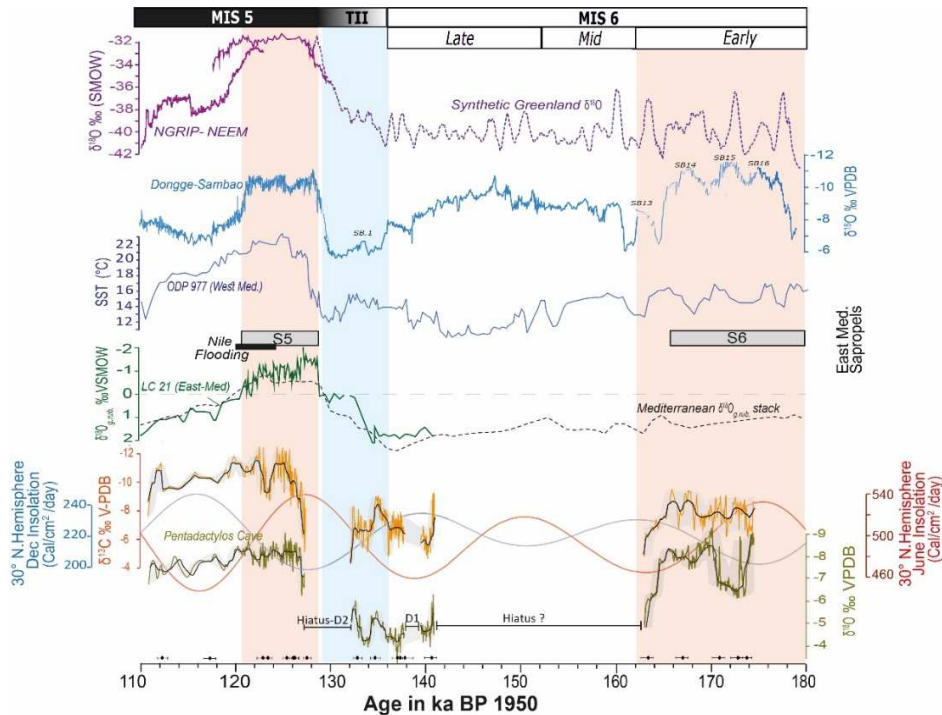
470 During the early-penultimate Glacial, despite large $\delta^{18}\text{O}$ variations (~-9.0‰ to ~-5.0‰), $\delta^{13}\text{C}$ values
 471 remained almost constant around -8‰ (Fig.6), suggesting a Mediterranean vegetation of C3 mixed with
 472 C4 type, or the dominance of a C3 type vegetation under water stress due to colder temperatures and/or
 473 reduced rainfall as previously suggested for speleothems growth during glacial cycles (Bar-Matthews et
 474 al., 2019). However, the quite negative $\delta^{18}\text{O}$ values (avg. -8‰) suggest rather wet conditions, supported

475 by a higher growth rate at the onset of Penta-2 growth. The stalagmite grew continuously until 163 ka,
 476 with sufficiently humid conditions to maintain drip water flow in the epikarst zone. After 165 ka, growth
 477 rate and stalagmite diameter decreased drastically along with less negative $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values,
 478 indicating unfavourable conditions for calcite precipitation. Therefore, the Early-MIS 6 glacial conditions
 479 seemed unstable but less severe in Cyprus than the full glacial conditions of the MIS 3/4 period in the
 480 Levant (Bar-Matthews et al., 2003; 2019). Speleothem studies from the Soreq, Peqiin, Tzavoa, and
 481 Jerusalem Caves and recently from Kanaan cave, Lebanon (Nehme et al., 2018) demonstrated that the
 482 climate dynamics from early-penultimate glacial in the EM were unstable with large $\delta^{18}\text{O}$ variations.
 483 Continuous millennial-scale variabilities extended to the mid-penultimate Glacial and were attributed to
 484 D-O like events (Barker et al., 2009) (Fig.6). Climate conditions were relatively cold during the MIS6e,
 485 but more humid than the last glacial MIS 3 (Ayalon et al., 2002; Bar-Matthews et al., 2003; Vaks et al.,
 486 2006). These findings contradict with those of Cheddadi and Rossignol-Strick (1995), who argued that
 487 conditions in the EM during early MIS 6 were very cold and arid, with precipitation below 200 mm,
 488 similar to the last two glacial maxima (MIS 4 and 2).



489
 490 **Figure 5.** cross plot of the stable isotope values from Pentadactylos Cave covering the early-MIS 6, the late-MIS 6-TII, the MIS
 491 5e (Eemian), the glacial Inception and modern (late-Holocene) periods. VPDB, Vienna Pee Dee belemnite.
 492

493 From 141 to 134 ka, $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values reached the most positive values (Fig. 5) for the entire
 494 Pentadactylos record. The average $\delta^{13}\text{C}$ value of -6‰ is close to values reached in Soreq and Peqiin caves
 495 during the last glacial, with calculated cold temperature of $14\text{-}16^\circ\text{C}$ throughout the MIS 4-2 (Affek et al.,
 496 2008). Less negative $\delta^{13}\text{C}$ in Pentadactylos, potentially caused by a decrease in soil CO_2 concentration, is
 497 accompanied by a lower growth rate suggesting lower rainfall amounts or colder temperatures. Similar
 498 conditions are noted in the Southern Levant with similar cold and arid climatic conditions in Soreq,
 499 Peqiin and Jerusalem records even though with a continuous speleothem growth throughout the entire TII.
 500



501
502 **Figure 6.** Marine and global records from 110 to 176 ka excluding the period from 142 to 162 ka (white column) and compared
503 to the *Pentadactylos* $\delta^{18}\text{O}$ profile (this study). Marine and global records comprise $\delta^{18}\text{O}$ Greenland ice core (NGRIP, 2004 ;
504 NEEM, 2013) with the synthetic $\delta^{18}\text{O}$ record (Barker et al., 2010), Sambaog-Dongge $\delta^{18}\text{O}$ profile in central China (Wang et al.,
505 2008, Cheng et al., 2016), SST ($^{\circ}\text{C}$) in ODP 977 marine core (Martrat et al. 2004, 2014), $\delta^{18}\text{O}_{g,rub}$ in LC21 marine core (Grant et
506 al., 2012, 2016) and Mediterranean $\delta^{18}\text{O}_{g,rub}$ stack (Wang et al., 2010), Sapropels S6 and S5 (Ziegler et al., 2010) and the July
507 insolation at 30°N (Berger and Loutre, 1991).
508

509 The climate conditions during the last Interglacial in Cyprus mark a return to warm and wet conditions
510 with negative $\delta^{18}\text{O}$ (-8‰) and $\delta^{13}\text{C}$ (-11‰) values starting from ~ 128 ka (Fig. 6) and supported by the
511 highest growth rates (217 mm/ka) related to a high effective and sustained infiltration in the epikarst. The
512 $\delta^{13}\text{C}$ values of $\sim -11\text{‰}$ indicate a significant contribution of soil biogenic CO_2 and thus relatively humid
513 (temperate to warm) conditions. Wet and warm climate conditions were recorded in most of the
514 speleothems of that time period in the Levant basin (Frumkin et al., 2000; Bar-Matthews et al., 2003,
515 Nehme et al., 2015). However, some discrepancies attributed to local and/or regional conditions arise
516 between the records and will be discussed hereafter in section 7.3.
517

518 7.3. Climate and $\delta^{18}\text{O}$ dynamics from the *Pentadactylos* record

519 *The penultimate glacial MIS6 conditions, in Cyprus: regional vs local conditions*

520 Between 185 and 135 ka BP, $\delta^{18}\text{O}$ values vary between $\sim -8\text{‰}$ and $\sim -3\text{‰}$ in the *Pentadactylos* record
521 while $\delta^{13}\text{C}$ remains fairly constant between $\sim -8\text{‰}$ and $\sim -6\text{‰}$ up to ~ 165 ka. At 175.6 ka, $\delta^{18}\text{O}$ values
522 reached -8.8‰ (Fig. 7A) with a moderate growth rate (57 mm/ka), a large stalagmite diameter and the
523 most negative $\delta^{13}\text{C}$ signal ($\sim -9.6\text{‰}$) of MIS 6, in phase with the maximum 30°N June insolation peak
524 (Berger and Loutre, 1991) and a high SST stand (Martrat et al., 2014) (Fig.6). The onset of Penta 2
525 growth suggests favorable conditions for speleothem growth with active effective recharge, higher drip
526 rate and sufficient contribution of soil biogenic CO_2 . Similar conditions are noted in Soreq with more
527 negative $\delta^{18}\text{O}$ values from 177 to 174.5 ka (Ayalon et al., 2002). An arboreal pollen (AP%) peak in the
528 Ioannina Basin (Greece) (Roucoux et al., 2011), further supports basin-wide wet conditions (Fig.7A).
529 From 174 to 171 ka, *Pentadactylos* shows a positive shift in $\delta^{18}\text{O}$ values for ~ 4 ka, indicating a brief
530 return to drier conditions, coinciding with a positive $\delta^{18}\text{O}$ shift at ~ 174 ka in the Soreq record, albeit
531 longer (3 ka in Penta record vs 8 ka in Soreq record) than in *Pentadactylos*. From ~ 170 - 165 ka, a more

532 negative $\delta^{18}\text{O}$ value corresponds to the maximum 30° June insolation. It occurs together with a moderate
533 growth rate (19 mm/ka) and is in phase with a gradual increase in the AP% in the Ioannina Basin
534 (Greece). In the southern Levant, the Soreq record displays negative $\delta^{18}\text{O}$ values at ~166 ka, slightly later
535 than in Cyprus or Greece. On larger and hemispheric scale, the negative Pentadactylos $\delta^{18}\text{O}$ values fall
536 within the large wet intervals identified in Argentarola, Italy (Bard et al., 2002) and Sanbao, China (Wang
537 et al., 2008) (Fig.7A).

538
539 The MIS6e period was known in the EM to be cold and wet (Emeis et al., 2003), during which Sapropel
540 layer S6 is deposited from 178.5 to 165.5 ka (Ziegler et al., 2010). The driver of negative $\delta^{18}\text{O}$ values in
541 speleothems would be ascribed to the Mediterranean depleted water source but also to some extent, to
542 enhanced pluvial conditions (amount affect) on land as shown in pollen assemblages in Mediterranean
543 marine records (Cheddadi and Rossignol-Strick, 1995) as well as in the Ioaninna (Roucoux et al., 2011),
544 Greece and Van records (Litt et al., 2014) in North-Eastern Turkey (Fig 7A). An alternative explanation
545 of negative $\delta^{18}\text{O}$ values only by temperature changes is rather unlikely, as the observed changes of
546 approximately 2.5‰ at the wet peaks would require a change in temperature of 10-12.5°C, which seems
547 unreasonably high.

548
549 Minor discrepancies between the $\delta^{18}\text{O}$ signal in Pentadactylos and other speleothems in the Levant,
550 especially for the 170-165 ka period, lays probably in the local pluvial conditions, which were more
551 intense in the Kyrenia mountains and comparable to pluvial conditions at the Ioannina and Tenaghi-
552 Philippon basins. The persistent and negative $\delta^{13}\text{C}$ values from ~175 to 165 ka in Cyprus indicate
553 however a continuous but moderate contribution of soil biogenic CO_2 . Speleothem records from the
554 Levant coast (e.g. Soreq, Peqiin, Kanaan) located at lower altitude (<600 m asl) showed $\delta^{13}\text{C}$ values more
555 negative than -8‰ for the same period. The $\delta^{13}\text{C}$ values exceeding -8‰ in Pentadactylos record can be
556 ascribed to the cave watershed altitude (>800 m asl) showing the altitudinal gradient effect on $\delta^{13}\text{C}_{\text{calcite}}$
557 (Johnston et al., 2015). After 165.7 ka, a significant change in oxygen and carbon values, growth rate (5
558 mm/ka) and stalagmite diameter point towards unfavorable growth conditions until a growth stop at 163
559 ka along with a clear deterioration of the vegetation cover.

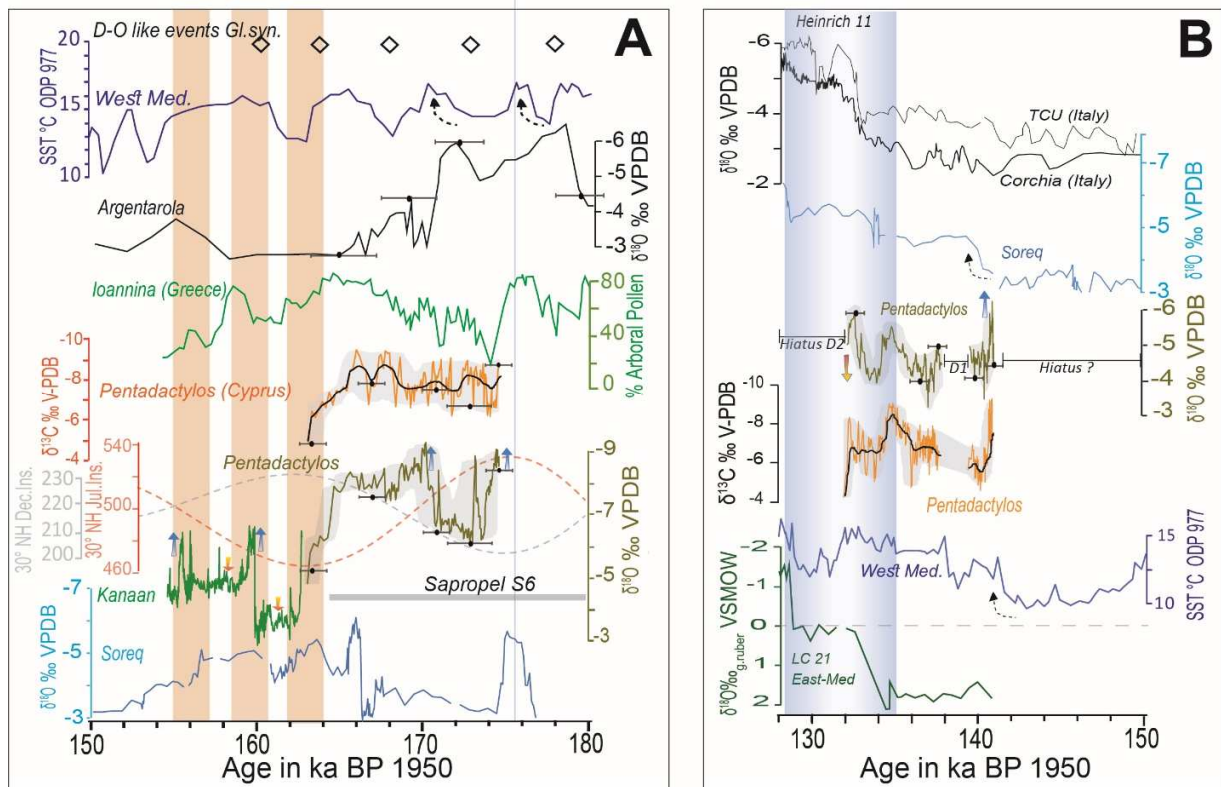


Figure 7. Paleoclimate records for the penultimate glacial in the Mediterranean basin. **A)** comparison of the paleoclimate records between 150 and 180 ka showing the Pentadactylos record from 163 to 175 ka (this study) and the U/Th ages with 2σ uncertainty, D-O like events from the Gl.synthetic record (Barker et al., 2009), SST ($^{\circ}$ C) in ODP 977 marine core (Martrat et al., 2014) the July insolation at 30° N (Berger and Loutre, 1991), $\delta^{18}\text{O}$ profile of Argentarola speleothem (Bard et al., 2002), the arboreal pollen (AP) % in the Ionnina basin (Roucoux et al., 2011), Kanaan $\delta^{18}\text{O}$ profile (Nehme et al., 2018), Sapropel S6 (Ziegler et al., 2010) and Soreq $\delta^{18}\text{O}$ profile (Ayalon et al., 2002; Bar-Matthews et al., 2019). Orange rectangles indicate the Chinese interstadials. **B)** Comparison of the paleoclimate records between 150 and 130 ka showing the Pentadactylos record from 141 to 132 ka (this study) and the U/Th ages with 2σ uncertainty, TCU (Regattieri et al., 2014) and Corchia (Drysdale et al., 2004, 2009) $\delta^{18}\text{O}$ profile, Soreq $\delta^{18}\text{O}$ profile (Ayalon et al., 2002; Bar-Matthews et al., 2019), SST ($^{\circ}$ C) in ODP 977 marine core (Martrat et al., 2014) and $\delta^{18}\text{O}_{\text{g,rub}}$ in LC21 marine core (Grant et al., 2012). Blue arrows show wetting phase and yellow arrows refer to dry phases.

The late-penultimate glacial in the Pentadactylos record (Penta 1) includes the period from 141.01 ± 0.56 to 132.25 ± 0.44 ka and part of the Termination (TII), marked by the most less negative $\delta^{18}\text{O}$ (avg: -4.6‰) and $\delta^{13}\text{C}$ (avg: -7.0‰) values of the entire record (Fig. 6). Overall, climatic conditions were colder (Table 4) and drier than during the early-penultimate period and the record shows a relatively stable maximum glacial state (Fig. 7B). A high growth rate (123 mm/ka) suggests however favorable humid conditions at that time marking a short shift to more effective recharge with the reactivation of calcite growth in Pentadactylos after a long hiatus from 163 to 141 ka. Noteworthy is also the good agreement of Pentadactylos $\delta^{18}\text{O}$ with the small negative $\delta^{18}\text{O}$ offset in Soreq at ~ 140 ka. Nonetheless, mild climate conditions prevailed after 140 ka, with lower growth rate (18 mm/ka) and low contribution of soil biogenic CO_2 in Pentadactylos (i.e., less negative $\delta^{13}\text{C}$), marking therefore a drier period consistent with an SST lower than 15°C (Martrat et al., 2004) in the Mediterranean Sea and low insolation (Fig. 6). Inland speleothem records show similar climatic trends with less negative $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in TCU, Corchia (Italy) and the Soreq record (Southern Levant).

After 136 ka, two $\delta^{18}\text{O}$ negative excursion centered at 134.5 ka and 132.5 ka are noted in the Pentadactylos record. Although the signal at 134.5 ka is not observed or resolved in other records (Fig. 7B), the $\delta^{18}\text{O}$ negative peak centered at 132.5 ka agrees within uncertainty with a clear $\delta^{18}\text{O}_{\text{g,rub}}$ shift in

590 LC21 EM marine record (tuned with Soreq chronology) and is concordant with $\delta^{18}\text{O}$ negative shift in
591 TCU (Regattieri et al., 2014) and Corchia records. This $\delta^{18}\text{O}_{\text{g.rub}}^{-}$ negative shift in the Mediterranean
592 basin, prior to the onset of Sapropel S5 (128-120 ka) (Ziegler et al., 2010) has been related to freshwater
593 input to the Mediterranean basin (Rodríguez-Sanz et al., 2017), derived from the North Atlantic source
594 and combined with meltwater discharge from the surrounding mountain ranges (Grant et al., 2016). A
595 source-water control on $\delta^{18}\text{O}_{\text{speleo}}^{-}$ in southern Italy by the North Atlantic freshwater was also confirmed
596 by Marino et al., (2015) and Tzedakis et al., (2018) for the 135-130 ka period. The $\delta^{18}\text{O}$ negative
597 excursions at 134.5 ka and 132.5 ka in Pentadactylos is therefore potentially unrelated to changes in
598 precipitation in Cyprus. At the contrary, a hiatus linked to D2 occurring from 132 to ~128 ka indicates
599 less favourable conditions for speleothem deposition. This period of non-growth is in agreement with the
600 period of lower SST values in the W. Med. marine record and with a positive $\delta^{18}\text{O}$ excursion in the TCU
601 (Regattieri et al., 2014) and Dim (Rowe et al., 2020) speleothem record, attributed to a C28 cold event
602 (Oppo et al., 2001) at ~129 ka. Indeed, the non-growth period in the Cypriot speleothem falls within the
603 Heinrich 11 (Heinrich, 1988; McManus et al., 1998), suggesting general cold/dry climate conditions
604 concordant with reduced moisture over the Mediterranean basin.
605

606 *The G/IG shift in Northern Levant*

607 The last G/IG shift in the EM basin is well-studied in the marine cores as well as in continental records in
608 the southern Levant. In the EM basin, marine $\delta^{18}\text{O}_{\text{g.rub}}^{-}$ curves (Emeis et al., 2003; Kallel et al., 2000,
609 Rohling et al., 2002; Grant et al., 2012, 2016) display a well-defined orbital modulation of G/IG MIS
610 cycles. Although the marine record follows the global pattern, the $\delta^{18}\text{O}_{\text{g.rub}}^{-}$ is characterized by larger
611 variations, reaching up to 4.5‰ for the MIS6/5 transition. This large shift reflects the amplification of
612 climate change effects in the EM, due to the contribution of significant inflow of ^{18}O -depleted water
613 during the Sapropel S5 event (128-121 ka ; Ziegler et al., 2010), together with enhanced pluvial
614 conditions in the entire EM basin (Kallel et al., 2000). The effect of the source-water control on $\delta^{18}\text{O}_{\text{calcite}}^{-}$,
615 was largely demonstrated in Southern Levant speleothems for the last G/IG cycle (Bar-Matthews et al.,
616 2019).

617 The MIS6/5 transition is prominently recorded as a ~5.7‰ shift affecting $\delta^{18}\text{O}_{\text{calcite}}^{-}$ in the Pentadactylos
618 record. This shift is close to Peqin and Soreq oxygen shifts (5.5 and 5.7‰ respectively) and larger than
619 the amplitude of the $\delta^{18}\text{O}_{\text{g.rub}}^{-}$ change (source) of the Mediterranean Sea water (Grant et al., 2012) for the
620 last G/IG cycle (Fig. 6). Both Levant and Cypriot speleothem records show a common source of moisture
621 mostly coming from the Mediterranean Levant at the LIG onset. However, the last G/IG amplitude shift
622 in continental records, mostly accounting for the source effect, embed other factors controlling the shift in
623 the $\delta^{18}\text{O}_{\text{calcite}}^{-}$, such as temperature change, rainfall amount, sea-land distance and sea-level. If the seawater
624 $\delta^{18}\text{O}_{\text{g.rub}}^{-}$ shift (4.5‰) accounts for the source effect (Grant et al., 2012) for the MIS 6/5 transition, the
625 remaining 1.2‰ difference would be ascribed for temperature and rainfall amount change in the $\delta^{18}\text{O}_{\text{calcite}}^{-}$
626 signal of Pentadactylos (Table 4).
627

628 *The wet and warm Eemian in Cyprus*

629 The last interglacial in the Pentadactylos stalagmite is recorded at high-resolution from ~127 to ~123 ka.
630 The stalagmite re-started its growth at 127.1 ± 0.5 ka when interglacial conditions were fully attained at
631 the maximum insolation during the Eemian (LIG) period (Fig. 8). A clear negative shift in both $\delta^{18}\text{O}$ (avg.
632 -8.03‰) and $\delta^{13}\text{C}$ (avg. -10.6‰) curves seems to show a clear vegetation recovery of C3 type with a high
633 contribution of soil biogenic CO_2 , combined with enhanced pluvial conditions enough to reactivate
634 effective recharge in a high-altitude epikarst basin (1000 m asl) and to induce the onset of calcite
635 precipitation (growth 24 to 217 mm/ka). The Pentadactylos record shows the highest growth rate (217
636 mm/ka) at ~126-125 ka. From ~127 to ~123 ka, the record seems to show stable conditions with both
637 negative $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ isotope values and high growth rate, indicating favorable warm and wet conditions
638 for speleothem growth, together with developed vegetation cover along the Kyrenia range (Fig. 4).
639 On the regional scale, these warm and wet conditions in the Kyrenia range fall within longer wet/warm
640 periods identified in the Yammounh lake record (Develle et al., 2011), the lake Van lacustrine sequence

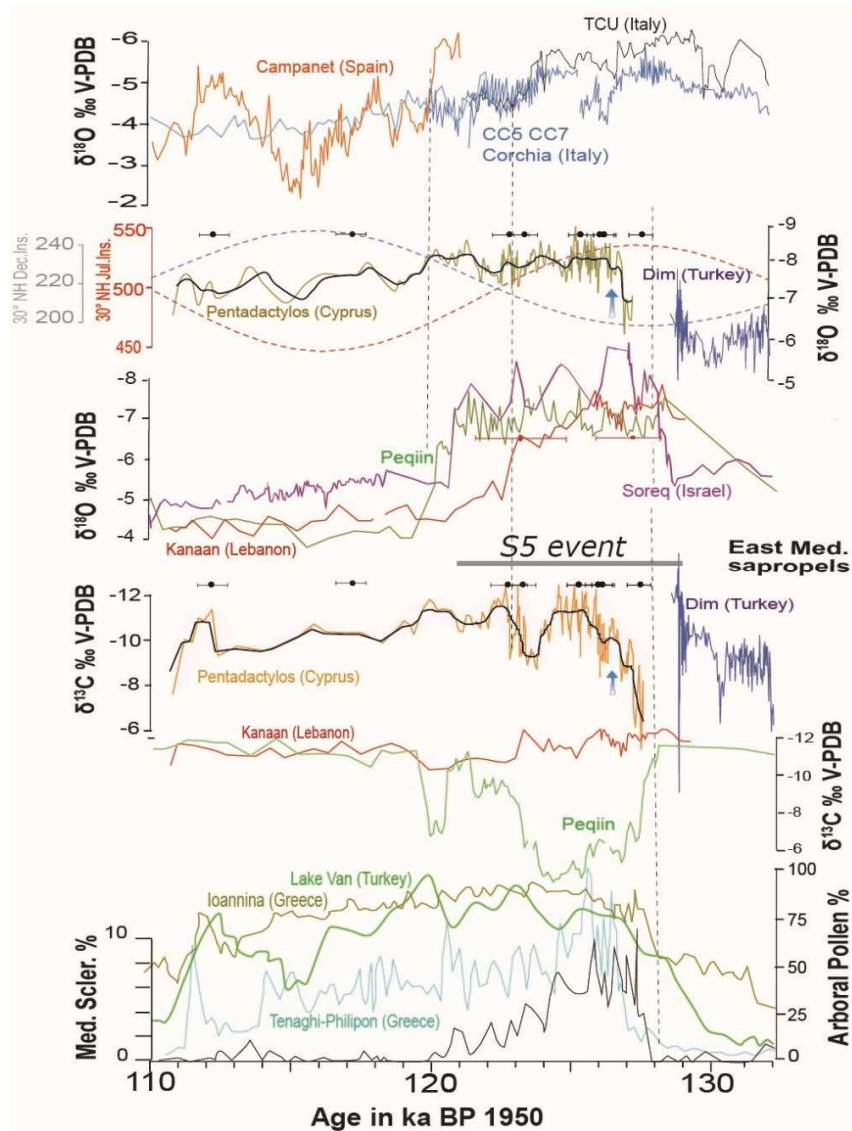
641 (Litt et al., 2014), the pollen lacustrine record in southeastern Greece and in Soreq, Peqiin (Ayalon et al.,
642 2002; Bar-Matthews et al., 2003) and Kanaan speleothem records along the Levant coast. The pollen
643 records in the EM basin shows a sharp arboreal pollen increase at the onset of Interglacial 5e (Fig. 8) in
644 agreement with a negative shift in the $\delta^{13}\text{C}$ signal of Pentadactylos record. The $\delta^{18}\text{O}$ negative shift is also
645 synchronous, within uncertainty, with the Tenaghi-Philipon (Tzedakis et al., 2003, 2009) and Ioaninna
646 pollen records (Roucoux et al., 2011) in Greece, albeit less clear with lake Van, located further East. The
647 particular increase of Sclerophylls pollen in the Greek record, characteristic of Mediterranean woodlands
648 and scrub, indicate a clear regional response to a climate wetting and warming phase.
649

650 At ~124-123 ka, a sharp drop in the $\delta^{13}\text{C}$ signal lasted 1 ka and coincides with slightly less negative $\delta^{18}\text{O}$
651 values (Fig. 8). This early change in the stable isotope signal could be a response to enhanced seasonality
652 or a modification in moisture source or an early climate deterioration signal with a lower precipitation
653 amount above the Levant basin. This period with low humidity is imprinted in pollen records with a drop
654 in AP % in the Tenaghi-Philippon and Van basins (Fig. 8) at the same time interval, implying thus a
655 regional response to an early climate deterioration.
656

657 *The Glacial inception in Cyprus vs regional records*

658 The moderate change after 123 ka, with only slightly less negative $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values contrasts with
659 most other speleothem records (Fig.8). Speleothem records along the Levant coast shows drastically less
660 negative $\delta^{18}\text{O}$ values after 123 ka, and was interpreted as i) a change towards drier conditions at the onset
661 of the glacial inception period and a reduced contribution of the direct Mediterranean moisture source
662 (Ayalon et al., 2002; Bar-Matthews et al., 2003; Nehme et al., 2015) over the central Levant coast or ii)
663 changes in atmospheric circulation pattern (Kolodny et al., 2005; Lisker et al., 2010). The Kanaan
664 isotopic trend, follows within larger chronological uncertainty the sharp $\delta^{18}\text{O}$ drop of Soreq and Peqiin
665 records, reaching significantly less negative values (~4‰) at 122 ka, while the oxygen isotope signal in
666 Cyprus maintains relatively constant negative values after 123 ka, approximately close to ~-7‰ until 110
667 ka (Fig 8). This clear difference in the stable isotope curves after 123 ka indicate that the Pentadactylos
668 record is closer in amplitude and trend with Corchia and TCU records in northern Italy than the Levant
669 records (Fig 8) when entering the glacial inception.
670

671 At the LIG demise, the Pentadactylos record suggests less favourable conditions of growth (8 to 2 mm/
672 ka) and less effective recharge as seen in the drastic growth rate drop after 123 ka (Fig.4). Drier or colder
673 climate conditions might have prevailed from 123 to 110 ka. However, the isotopic values trend in
674 Pentadactylos show a less clear change. Indeed, even if a reduction in the rainfall amount is considered
675 after 123 ka, the $\delta^{18}\text{O}$ trend in Cyprus exhibits a clear offset (Fig. 8) from the $\delta^{18}\text{O}_{\text{g.rub}}$ (source) of the
676 Mediterranean Sea water (Grant et al., 2016), and does not mimic the $\delta^{18}\text{O}$ signal of the Levant
677 speleothems, which are mostly source-driven. Pluvial conditions might have been maintained at the LIG
678 demise as shown in the low $\delta^{18}\text{O}$ signal by possible distal moisture sources (e.g. North Atlantic), as
679 proposed in studies on rainfall trajectories in the EM realm (Aouad-Rizk et al., 2005; Argiriou and
680 Lykoutis, 2005), with a longer trajectory bringing rainfall with a lighter isotopic signal over the island and
681 extensively over the Northern Levant basin. If the hydrological change at the LIG demise (the glacial
682 inception) between Cyprus and the Levant coast holds true, then the northern Levant basin would be more
683 open to distal and northern rainfall sources under a strong NAO index as noted in the large-scale $\Delta\delta^{18}\text{O}_\text{c}$
684 variability of the Yammouneh record (Northern Lebanon) during glacial intervals (Gasse et al., 2014) as
685 well as in the Jeita record during the LGM period (Cheng et al., 2015).



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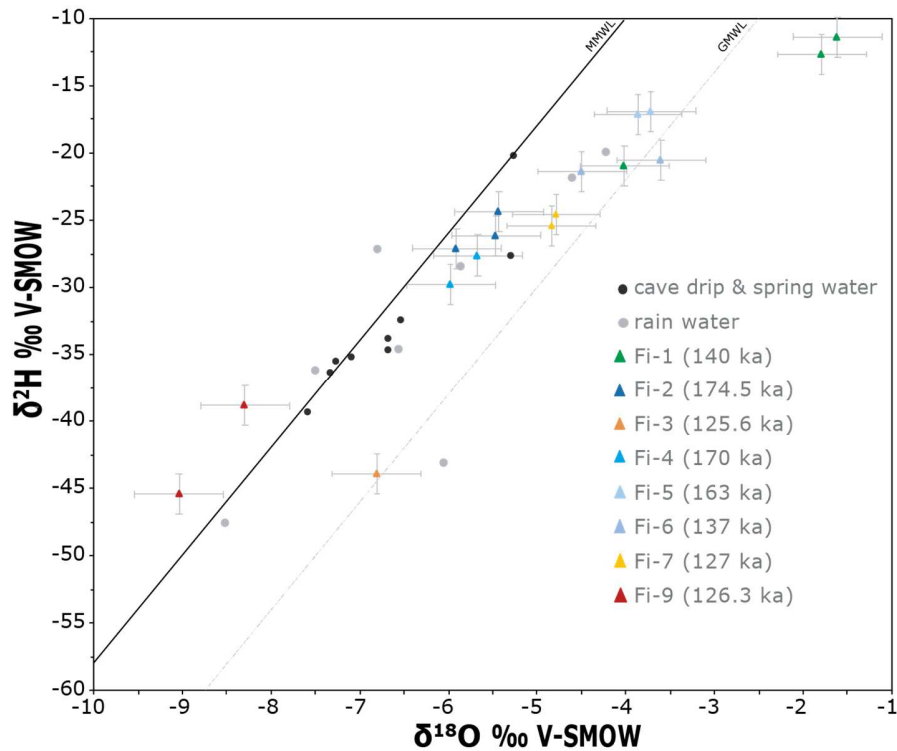
Figure 8. Paleoclimate records for the penultimate glacial and last Interglacial records in the Mediterranean basin. Continental paleoclimate records from 110 to 134 ka showing TCU and Crochia $\delta^{18}\text{O}$ profile (Regattieri et al., 2014; Drysdale et al., 2004, 2009) in northern Italy and Campanet $\delta^{18}\text{O}$ profile in Mallorca (Dimitriu et al., 2018), Dim $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ profile, Turkey (Rowe et al., 2020), Peqiin and Soreq $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ profile (Ayalon et al., 2002, Bar-Matthews et al., 2003), Kanaan $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ profile (Nehme et al., 2018), the arboreal pollen (AP) % in Van lake (Litt et al., 2014) and Tenaghi-Philippou (Tzedakis et al., 2004, 2009) and particularly the Mediterranean Sclerophylls pollen in the Ionina (Tzedakis et al., 2003).

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7.4. Estimates of past temperature and rainfall shifts from MIS 6 to MIS 5

Interpretation of fluid inclusion δD and $\delta^{18}\text{O}$

The changes in fluid inclusion δD and $\delta^{18}\text{O}$ along the current meteoric waterline with time (Fig. 9) are likely caused by the interplay of three main factors: i) amount effect, ii) source effect and iii) evaporation on the stalagmite top prior to final fluid inclusion close-off. Focusing first on i), the influence of the rainfall amount is exemplified by the most negative δD and $\delta^{18}\text{O}$ values at 126.3 and 125.6 ka. This time period has been found to be very wet in the EM (Bar-Matthews et al., 2003; McGarry et al., 2004) pointing towards amount-effect driven changes in the dripwater signal, which results in most negative isotope values in both the Pentadactylos and the Soreq/Peqiin/Ma'ale Efrayim data set. Note that the difference of last Interglacial $\delta^{18}\text{O}$ values relative to modern drip water $\delta^{18}\text{O}$ is comparable for Pentadactylos and Israel (up to -2 ‰).



705
706 **Figure 9.** $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of 17 fluid inclusion samples from *Pentadactylos* stalagmite along with present-day dripwater and
707 stream samples (this study) from the Kyrenia Range and rain water (IAEA/WMO, 2019) plotted on the MMWL (Gat et al., 2003)
708 and GMWL (Rozanski et al., 1993).
709

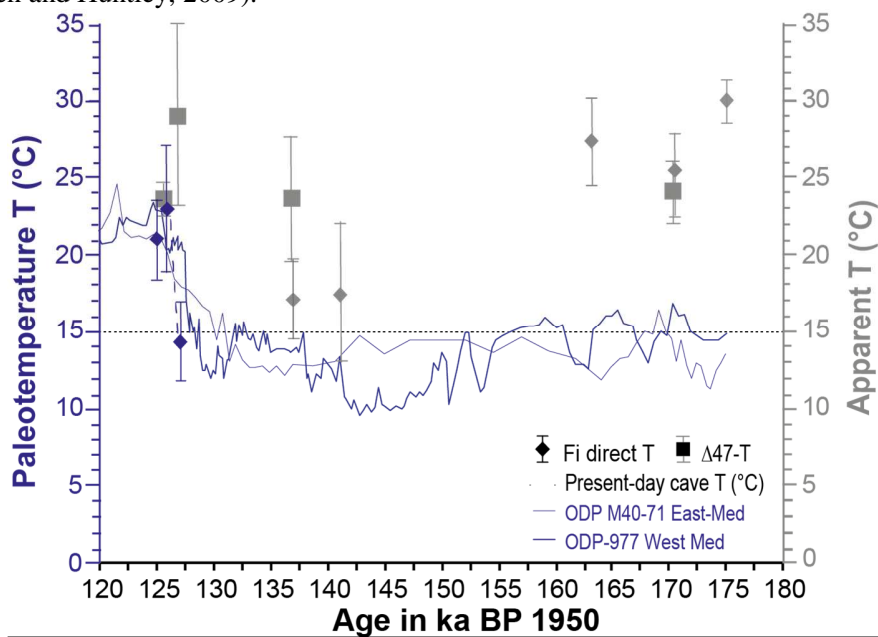
710 Positive $\delta^{18}\text{O}$ offsets from the MMWL/GMWL and less negative δD values compared to today, may
711 indicate an influence of evaporation processes. The most positive δD and $\delta^{18}\text{O}$ values, with significant
712 offsets from the MMWL or GMWL (Fi-1) suggests an evaporative enrichment in soil and epikarst (Vaks
713 et al., 2003) or significant inner-cave evaporation, both processes taking place preferentially during dry
714 periods. The shifts observed in fluid inclusion $\delta^{18}\text{O}$ are up to +6 ‰ relative to modern cave and spring
715 water $\delta^{18}\text{O}$ (Table 2) and by far exceed those from equivalent records from Israel for the same time
716 periods (maximum shift: +2 ‰; McGarry et al., 2004). There, glacial samples provided less negative fluid
717 inclusion $\delta^{18}\text{O}$ values, closer to the GMWL and reflecting a reduced d-excess that may indicate a higher
718 relative humidity. The $\delta^{18}\text{O}$ values of McGarry et al. (2004) were calculated from measured fluid
719 inclusion δD and therefore do not reflect the typically strong evaporation effect on $\delta^{18}\text{O}$. This can explain
720 the difference to our data set where the positive shift in the measured $\delta^{18}\text{O}$ values away from the MMWL
721 cannot solely be explained by a change in the water source isotope values, but also requires an
722 evaporative component (Appendix S3).
723

724 Although the amount effect seems to have had a significant influence on the *Pentadactylos* fluid inclusion
725 $\delta^{18}\text{O}$, a determination of paleotemperatures, based on the temperature dependence of the fractionation $^{18}\alpha$
726 between drip water and calcite is attempted. The temperature dependence of $^{18}\alpha(\text{CaCO}_3\text{-H}_2\text{O})$ is known
727 from laboratory experiments or empirical studies and can be used as thermometer. Calculations using the
728 measured fluid inclusion $\delta^{18}\text{O}$ and surrounding calcite $\delta^{18}\text{O}$ values lead to reasonable temperatures in the
729 range between 14 and 30°C with typical uncertainties of 1-4°C (Fig. 10). However, calculated
730 temperatures for 140 and 160-175 ka appear to be unreasonable, being similar or higher than Interglacial
731 values. An increased influence of evaporative effects during glacial periods as explained previously, may
732 be a likely explanation for this discrepancy. As demonstrated by Verheyden et al. (2008c), low drip rates
733 enable evaporation in caves with increased $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of the subsequent deposited calcite. Moreover,
734 comparing the ϵ -based temperatures with SSTs (Fig.10) shows that the most reliable reconstructed

735 temperatures correspond to known wet periods with the lowest speleothems $\delta^{18}\text{O}$ values. Therefore, our
 736 estimate for the G-IG temperature shift has to be seen as a minimum estimate, with interglacial
 737 temperatures close to current values and larger deviations for the glacial samples.
 738

739 *Regional comparison of G-IG temperature shifts*

740 Besides the G-IG temperature shift derived from calcite and fluid inclusion $\delta^{18}\text{O}$, the difference in
 741 apparent temperature derived from clumped isotopes, also yields a minimum temperature shift of ca. 9°C
 742 which is of the same order as from the $^{18}\epsilon_{\text{calcite-water}}$ estimate but smaller than independent estimates from
 743 Villars cave speleothems, SW France (Wainer et al., 2011). The authors indicate a shift of $14 \pm 2.7^\circ\text{C}$ for
 744 the MIS 6 - MIS 5e transition. Noble gas fluid inclusion analyses from a western Germany stalagmite
 745 yield $9 \pm 5^\circ\text{C}$ (1σ) as a lower bound for temperature rise across TII (Scholz et al., 2009). Other
 746 reconstructed temperature data, such as from pollen records from Monticchio Maar in Italy indicate a
 747 similar shift with a maximum amplitude of 12°C between the end of MIS 6 and the Last Interglacial
 748 optimum (Allen and Huntley, 2009).



749 **Figure 10.** Paleotemperatures based on the fractionation $^{18}\alpha(\text{CaCO}_3\text{-H}_2\text{O})$ from fluid inclusion values (in blue) before 130 ka
 750 with the apparent T from clumped isotopes analysis (grey squares) based on the $\Delta_{47}\text{-T}$ calibration of Kluge et al. (2015).
 751 Temperatures are plotted against the present-day cave temperature and the SST (blue lines) from the nearby marine cores (M40-
 752 71 from Emeis et al., (2003) and ODP-977 from Martrat et al., (2014). Fluid inclusion T that are assumed to be influenced by
 753 evaporation are shown as small grey triangles.
 754
 755

756 *Rainfall amount estimates*

757 Based on the inverse relationship between rainfall $\delta^{18}\text{O}$ and precipitation amount in modern observations
 758 (IAEA/WMO, 2019; Gat et al., 2003; El-Asrag, 2005; Dirican et al., 2005; Nehme et al., 2019) in the EM,
 759 a rainfall increase of ca. 290 mm/a is estimated to relate to a decrease in mean annual rainfall $\delta^{18}\text{O}$ of 1‰
 760 (Appendix S2). This value is comparable to the observation of Bar-Matthews et al. (2003) with ca. 200
 761 mm/a per 1‰ at the Soreq cave site. The quantification leads to annual rainfall reconstructions of
 762 between <400 and ca. 1200 mm for Cyprus, with peak rainfall during the last interglacial optimum (table
 763 4). Around 170 ka rainfall amounts seem to be <400mm. Rainfall reconstructions for the Last Glacial
 764 Period are challenging and not quantitatively attempted here, because little is known on the rainfall $\delta^{18}\text{O}$ -
 765 amount relationship beyond the interglacial conditions.
 766

767 **Table 4.** Temperature T(°C) from $^{18}\epsilon_{\text{calcite-fluid}}$ after correction for Δ_{47} offset (2nd column) and rainfall amount estimates (3rd column)
 768 following an EM relationship between mean annual rainfall $\delta^{18}\text{O}$ and amount (Appendix S2). Data were scaled based on modern

769 day rainfall values in Cyprus. *Elevated or high apparent temperatures indicate that an evaporative effect has increased the fluid
 770 $\delta^{18}\text{O}$ value (*)an evaporative influence may exist. No rainfall amount has been estimated for the full glacial period due to an
 771 unknown rainfall amount- $\delta^{18}\text{O}$ relationship.

Age in ka BP	Apparent T (°C)	Estimated precipitation amount (mm)
125.6	20.8 ± 2.5	650 ± 170
126	23.2 ± 4.2	1190 ± 150
127	14.7 ± 2.7	-
137.1	(*)17.2 ± 2.7	-
141	(*)17.6 ± 4.1	-
163.9	*27.6 ± 2.7	-
170.4	*25.5 ± 2.8	370 ± 160
174.5	*30.2 ± 1.4	300 ± 160

772
 773 **8- Conclusion**

774 The Pentadactylos stalagmite record from North-Eastern Cyprus gives new insights into climatic
 775 variations during MIS 6 and 5 for the Northern Levant, a region with sparse records spanning the last
 776 G/IG cycle. This study suggests moderately temperate and wet conditions in the early MIS6. These
 777 conditions prevailed between probably before ~175, to ~163 ka with possibly a short return to dry
 778 conditions at ~174ka. The wet period is followed by a stalagmite growth stop, coinciding with the coldest
 779 and driest period of MIS 6, from 160 to 141 ka. After 141 ka, slightly more favorable conditions prevail
 780 as indicated by slow stalagmite growth. A stop in growth between 132 and 128 ka suggests a short cold
 781 and dry period contemporaneous with a Heinrich event.

782 The onset of the Eemian and along the whole LIG and glacial inception reveal similarities and
 783 discrepancies between the Eastern-Mediterranean (EM) records regarding their sensitivity to regional
 784 climate. The Pentadactylos $\delta^{18}\text{O}$ record was driven mostly by the source effect (Sapropel 5) at the onset of
 785 the Eemian, similarly to the Central and Southern Levant sites. Stable conditions during MIS 5e were
 786 however rather short in Cyprus compared to Soreq-Peqiin caves located further South. At ~122 ka, a
 787 regional climate deterioration is visible in all sites but with different behaviour and isotopic shifts across
 788 the glacial inception. A possible change in the regional circulation patterns under a higher NAO index
 789 make the northern-Levant more open to distal and northern rainfall sources as noted clearly in the
 790 Pentadactylos record situated further North. Other sites along the more Southern coastal Levant would be
 791 englobed in the North Atlantic System but remain less influenced than Cyprus, due to the distal locations
 792 of these sites.

793 The estimations of temperature and rainfall amount in Cyprus indicate a temperature shift of ca.9°C at the
 794 MIS 6-5 transition and a drastic change in rainfall amount between MIS 6 and 5 (large shift in the
 795 measured and calculated fluid inclusion $\delta^{18}\text{O}$ value - see Table 2 and 3). The paleotemperature difference
 796 estimated from Pentadactylos cave compares well to similar changes in SST for MIS 6 and 5 and
 797 confirms the notable marine influence on Cyprus.
 798

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816 **10-References**

817 Affek H.P., Bar-Matthews M., Ayalon A., Matthews A., Eiler J.M., 2008. Glacial/interglacial temperature variations in Soreq
818 cave speleothems as recorded by ‘clumped isotope’ thermometry. *Geochim. Cosmochim. Ac.*, 72(22), 5351-5360.

819 Affek H.P., Matthews A., Ayalon A., Bar-Matthews M., Burstyn Y., Zaarur S., Zilberman T., 2014. Accounting for kinetic
820 isotope effects in Soreq Cave (Israel) speleothems. *Geochim. Cosmochim. Ac.*, 143, 303-318.

821 Affolter S., Fleitmann D., Leuenberger M., 2014. New-on-line method for water isotope analysis of speleothem fluid inclusions
822 using laser absorption spectroscopy (WS-CRDS). *Clim. Past*, 10, 1291-1304.

823 Affolter S., Häuselmann A.D., Fleitmann D., Häuselmann P., Leuenberger M., 2015. Triple isotope (δD , $\delta^{17}O$, $\delta^{18}O$) study on
824 precipitation, drip water and speleothem fluid inclusions for a W. Central European cave (Switzerland), *Quat. Sci. Rev.* 127, 73-
825 89

826 Allen, J.R.M. and Huntley, B., 2009. Last Interglacial palaeovegetation, palaeoenvironments and chronology: a new record from
827 Lago Grande di Monticchio, southern Italy. *Quat. Sci. Rev.* 28, 1521e1538.

828 Alpert P., Price C., Krichak S.O., Ziv B., Saaroni H., Osetinsky I., Kishcha P., 2005. Tropical teleconnections to the
829 Mediterranean climate and weather. *Advances in Geosciences*, 2, 157-160.

830 Aouad-Rizk A., Job J.L., Najem W., Travi Y., Blavoux B., Gourcy L., 2005. Oxygen-18 and deuterium contents over Mount
831 Lebanon related to air mass trajectories and local parameters. IAEA, Isotopic composition of precipitation in the Mediterranean
832 Basin in relation to air circulation patterns and climate, 75-82.

833 Argiriou, A.A. and Lykoutis S., 2005. Stable isotopes in rainfall over Greece: results of the 2000–2003 measurement campaign.
834 IAEA, Isotopic composition of the precipitation in the Mediterranean Basin in relation to air circulation patterns and climate,
835 Vienna, Austria, 83-98.

836 Ayalon A., Bar-Matthews M., Kaufman A., 2002. Climatic conditions during marine isotopic stage 6 in the Eastern
837 Mediterranean region as evident from the isotopic composition of speleothems: Soreq Cave, Israel. *Geology* 30, 303–306.

838 Ayalon A., Bar-Matthews M., Frumkin A., Matthews A., 2013. Last Glacial warm events on Mount Hermon: the southern
839 extension of the Alpine karst range in the east Mediterranean, *Quat. Sci. Rev.*, 59, 43–56.

840 Baker A. and Smart P.L., 1995. Recent flowstone growth rates: field measurements in comparison to theoretical predictions,
841 *Chem. Geol.*, 122, 121–128.

842 Baker A., Genty D., Dreybrodt W., Barnes W.L., Mockler N.J., Grapes J., 1998. Testing theoretically predicted stalagmite
843 growth rate with recent annually laminated samples implication for past stalagmite deposition. *Geochim. Cosmochim. Acta* 62,
844 393–404

845 Bard E., Antonioli F., Silenzi S., 2002. Sea-level during the penultimate interglacial period based on a submerged stalagmite
846 from Argentarola Cave (Italy). *Earth Planet Sci. Lett.*, 196, 135–146.

847 Barker S., Knorr G., Edwards R.L., Parrenin F., Putnam A.E., Skinner L.C., Wolff E., Ziegler M., 2011. 800,000 years of abrupt
848 climate variability. *Science* 334, 347–351.

849 Berger A. and Loutre M.F., 1991. Insolation values for the climate of the last 10 million years. *Quat. Sci. Rev.*, 10, 297–317.

850 Bar-Matthews M., Ayalon A., Gilmour M., Matthews M., Hawkesworth C., 2003. Sea-land isotopic relationships from
851 planktonic foraminifera and speleothems in the Eastern Mediterranean region and their implications for paleorainfall during
852 interglacial interval. *Geochim. Cosmochim. Acta*, 67, 3181–3199.

853 Bar-Matthews M., Keinen J., Ayalon A., 2019. Hydro-climate research of the late quaternary of the Eastern Mediterranean-
854 Levant region based on speleothems research. A review. *Quat. Sci. Rev.*, 221, 105872.

855 Benetti M., Reverdin G., Pierre C., Merlivat L., Risi C., Steen-Larsen H.C., Vimeux F., 2014. Deuterium excess in marine water
856 vapor: Dependency on relative humidity and surface wind speed during evaporation. *J. Geophys. Res. Atmos.*, 119, 584–593.

857 Bernasconi S. M., Hu B., Wacker U., Fiebig J., Breitenbach S.F.M., Rutz T., 2013. Background effects on Faraday collectors for
858 clumped isotope measurements. *Rapid Commun. Mass Spectrom.* 27, 603–612.

859 Borsato A., Quinif Y., Bini A., Dublyansky Y., 2005. Open system alpine speleothems: implications for U-series dating and
860 paleoclimate reconstructions. *Studi Trentini Di Scienze Naturali, Acta Geologica* 80, 71–83.

861 Breitenbach S.F.M., Rehfeld K., Goswami B., Baldini J.U.L., Ridley H.E., Kennett D.J., Prufer K.M., et al., 2012. Constructing
862 proxy records from age models (COPRA). *Clim. Past*, 8, 1765–1779.

863 Cailhol D., Audra P., Nehme C., Nader F., Garašić M., Heresanu V., Gucl S., Charalambidou I., Cheng H., Edwards R. L.,
864 2019. The contribution of condensation-corrosion in the morphological evolution of caves in semi-arid regions: preliminary
865 investigations in the Kyrenia range, Cyprus. *Acta Carsologica*, 48/1, 9-33.

866 Cheddadi R. and Rossignol-Strick M., 1995. Eastern Mediterranean Quaternary paleoclimates from pollen and isotope records of
867 marine cores in the Nile cone area, *Paleoceanography*, 10, 291–300.

868 Cheng H., Zhang P.Z., Spötl C., Edwards R.L., Cai Y.J., Zhang D.Z., Sang W.C., Tan M., An Z.S., 2012. The climatic cyclicity
869 in semiarid-arid central Asia over the past 500,000 years. *Geophys. Res. Lett.*, 39, L01705.

870 Cheng H., Edwards R.L., Shen C.C., Polyak V.J., Asmerom Y., Woodhead J., Hellstrom J., 2013. Improvements in ^{230}Th dating,
871 ^{230}Th and ^{234}U half-life values, and U–Th isotopic measurements by multi-collector inductively coupled plasma mass
872 spectrometry. *Earth Planet Sci. Lett.*, 371, 82–91.

873 Cheng, H., Sinha, A., Verheyden, S., Nader, F.H., Li, X.L., Zhang, P.Z., Yin, J.J., Yi, L., Peng, Y.B., Rao, Z.G. and Ning, Y.F.,
874 2015. The climate variability in northern Levant over the past 20,000 years. *Geophys. Res. Lett.*, 42(20), 8641-8650.

875 Cheng H., Edwards R.L., Sinha A., Spötl C., Yi L., Chen S., Kelly M., Kathayat G., Wang X., Li X., Kong X., Wang Y., Ning
876 Y., Zhang H., 2016. The Asian monsoon over the past 640,000 years and ice age terminations. *Nature* 534, 640–646.
877 Columbu A., Drysdale R., Capron E., Woodhead J., De Waele J., Sanna L., Hellstrom J. Bajo P., 2017. Early last glacial intra-
878 interstadial climate variability recorded in a Sardinian speleothem. *Quat. Sc. Rev.*, 169, 391–397.
879 Columbu A., Spötl C., De Waele J., Yu T.L., Shen C.C., and Gázquez F., 2019. A long record of MIS 7 and MIS 5 climate and
880 environment from a western Mediterranean speleothem (SW Sardinia, Italy). *Quat. Sci. Rev.*, 220, 230–243.
881 Constantinou, G., 1995. Geological Map of Cyprus, 1/250.000.- Geological Survey of Cyprus.
882 Daëron M., Guo W., Eiler J., Genty D., Blamart D., Boch, R., Drysdale R.N., Maire R., Wainer K., Zanchetta G., 2011.¹³C-¹⁸O
883 clumping in speleothems: observations from natural caves & precipitation experiments, *Geochim. Cosmochim. Acta*, 75, 3303–
884 3317.
885 Daëron M., Blamart D., Peral M., Affek H.P., 2016. Absolute isotopic abundance ratios and the accuracy of Δ_{47} measurements.
886 *Chem. Geol.* 442, 83–96.
887 Dennis K.J., Affek H.P., Passey B.H., Schrag D.P., Eiler J.W., 2011. Defining an absolute reference frame for ‘clumped’ isotope
888 studies of CO₂. *Geochim. Cosmochim. Acta* 75, 7117–7131.
889 Develle A.L., Gasse F., Vidal L., Williamson D., Demory F., Van Campo E., Ghaleb B., Thouveny N., 2011. A 250ka
890 sedimentary record from a small karstic lake in the Northern Levant (Yammoûneh, Lebanon): Paleoclimatic implications,
891 *Palaeogeogr. Palaeoclim., Palaeoecol.*, 305, 10–27, 2011
892 Dumitru O.A., Onac B.P., Polyak V.J., Wynn J.G., Asmerom Y., Fornós J.J., 2018. Climate variability in the western
893 Mediterranean between 121 and 67 ka derived from a Mallorcan speleothem record. *Palaeogeogr. Palaeoclim., Palaeoecol.*, 506,
894 128–138.
895 Dirican A., Unal S., Acar Y., Demircan M., 2005. The temporal and seasonal variation of δ 2H and δ 18O in atmospheric water
896 vapour and precipitation from Ankara, Turkey in relation to air mass trajectories at Med. Basin. In: *Isotopic composition of*
897 *precipitation in the Mediterranean Basin in relation to air circulation patterns and climate.* IAEA-TECDOC, 1453: 191–214.
898 Dutton A. and Lambeck K., 2012. Ice volume and sea level during the last interglacial, *Science*, 337, 216–219.
899 Dreybrodt, W., 1988. *Processes in Karst Systems*. Springer, New York.
900 Dreybrodt W., 1999. Chemical kinetics, speleothem growth and climate. *Boreas* 28, 347–356.
901 Drysdale R.N., Zanchetta G., Hellstrom J.C., Fallick A.E., Zhao J.X., Isola I., Bruschi G., 2004. Palaeoclimatic implications of
902 the growth history and stable isotope ($\delta^{18}\text{O}$; $\delta^{13}\text{C}$) geochemistry of a Middle to Late Pleistocene stalagmite from central western
903 Italy. *Earth Planet Sci. Lett.*, 227, 215–229.
904 Drysdale R., Hellstrom J., Zanchetta G., Fallick A.E., Sánchez Goñi M.F., Couchoud I., McDonald J., Maas, R., Lohmann, G.,
905 and Isola I., 2009. Evidence for obliquity forcing of glacial Termination II, *Science*, 325, 1527–1531.
906 Dublyansky Y.V. and Spötl C., 2009. $\delta^{18}\text{O}$ & $\delta^2\text{H}$ of water from inclusions in minerals: design of a new crushing system and on-
907 line continuous-flow isotope ratio mass spectrometric analysis. *Rapid Commun Mass Spectrom*, 23(17), 2605–2613.
908 Duplessy J.C., Labeyrie L., Arnold M., Paterne M., Duprat J., Van Weering T.C.E., 1992. Changes in surface salinity of the
909 North Atlantic Ocean during the last deglaciation. *Nature*, 358 (6386), 485.
910 Edwards R.L., Chen J.H., Ku T.L., Wasserburg G.J., 1987. Precise timing of the last interglacial period from mass spectrometric
911 determination of ²³⁰Th in corals. *Science* 236, 1547–1553.
912 Ehlers J., Gibbard P.L., 2007. The extent and chronology of Cenozoic global glaciation. *Quaternary International* 164, 6–20.
913 El-Asrag A.M., 2005. Effect of synoptic and climatic situations on fractionation of stable isotopes in rainwater over Egypt and
914 east Mediterranean. In: *Isotopic composition of precipitation in the Mediterranean Basin in relation to air circulation patterns and*
915 *climate.* IAEA TECDOC, 1453: 51–73.
916 Emeis K.C., Schulz H., Struc U., Rossignol-Strick M., Erlenkeuser H., Howell M.W., Kroon D., 2003. Eastern Mediterranean
917 surface water temperatures and $\delta^{18}\text{O}$ during deposition of sapropels in the late Quaternary. *Paleoceanography* 18, 1005–1029.
918 Fairchild I.J., Smith C.L., Baker A., Fuller L., Spötl C., Matthey D., McDermott F., 2006. Modification and preservation of
919 environmental signals in speleothems. *Earth Sci. Rev.* 75, 105–153.
920 Fairchild, I.J. and Baker, A., 2012. *Speleothem science: from process to past environments*, Vol. 3. John Wiley & Sons, 416
921 pages.
922 Fiebig J., Hofmann S., Löffler N., Lüdecke T., Methner K., and Wacker U., 2015. Slight pressure imbalances can affect accuracy
923 and precision of dual inlet-based clumped isotope analysis, *Isotopes in Environmental and Health Studies*.
924 Frisia S., 2015. Microstratigraphic logging of calcite fabrics in speleothems as tool for paleoclimate studies. *Int. J. Speleol.* 44, 1–
925 16.
926 Frumkin A., Ford D.C., Schwarcz H.P., 1999. Continental oxygen isotopic record of the last 170,000 years in Jerusalem, *Quat.*
927 *Res.*, 51, 317–327.
928 Frumkin A., Ford D.C., Schwarcz H., 2000. Paleoclimate and vegetation of the Last Glacial cycles in Jerusalem from a
929 speleothem record. *Global Biogeochem. Cycles* 14, 863–870.
930 Gat J.R., Klein B., Kushnir Y., Roether W., Wernli H., Yam R., Shemesh A., 2003. Isotope composition of air moisture over the
931 Mediterranean Sea: an index of the air–sea interaction pattern. *Tellus*, 55(5), 953–965.
932 Gasse F., Vidal L., Van Campo E., Demory F., Develle A.L., Tachikawa K., Thouveny N., 2014. Hydroclimatic changes in
933 northern Levant over the past 400,000 years. *Quat. Sci. Rev.*, 111, 1–8.
934 Grant K.M., Rohling E.J., Bar-Matthews M., Ayalon A., Medina-Elizalde M., Bronk-Ramsey C., Satow C., Roberts A.P., 2012.
935 Rapid coupling between ice volume and polar temperature over the past 150 ka, *Nature* 491, 744–747.
936 Grant K.M., Grimm R., Mikolajewicz U., Marino G., Ziegler M., Rohling E.J., 2016. The timing of Mediterranean sapropel
937 deposition relative to insolation, sea-level and African monsoon changes. *Quat. Sci. Rev.*, 140, 125–141.

938 Genty D., Baker A., Vokal B., 2001. Intra- and inter-annual growth rate of modern stalagmites. *Chem. Geol.*, 176, 191–212.

939 Genty D., Blamart D., Ouahdi R., Gilmour M., 2003. Precise dating of Dansgaard–Oeschger climate oscillations in western

940 Europe from stalagmite data. *Nature* 421, 833–837.

941 Genty D., Blamart D., Ghaleb B., Plagnes V., Causse C., Bakalowicz M., Zouari K., et al., 2006. Timing and dynamics of the last

942 deglaciation from European and North African $\delta^{13}\text{C}$ stalagmite profiles-comparison with Chinese and South Hemisphere

943 stalagmites. *Quat. Sci. Rev.*, 25, 2118–2142.

944 Genty D., Labuhn I., Hoffmann G., Danis P.A., Mestre O., Bourges F., Wainer K., Massault M., Van Exter S., Régnier E.,

945 Orengo Ph., Falourd S., Minster B., 2014. Rainfall and cave water isotopic relationships in two South-France sites. *Geochim.*

946 *Cosmochim. Acta*, 131, 323-343.

947 Guo W., 2020. Kinetic clumped isotope fractionation in the DIC-H₂O-CO₂ system: patterns, controls, and implications. *Geochim.*

948 *Cosmochim. Acta*, 268, 230-257.

949 Hansen M., Scholz D., Schöne B.R., Spötl C., 2019. Simulating speleothem growth in the laboratory: Determination of the stable

950 isotope fractionation ($\delta^{13}\text{C}$, $\delta^{18}\text{O}$) between H₂O, DIC and CaCO₃. *Chem. Geol.*, 509, 20-44.

951 He B., Olack G.A., Colman A.S., 2012. Pressure baseline correction and high-precision CO₂ clumped-isotope (Δ_{47})

952 measurements in bellows and micro-volume modes. *Rapid Commun Mass Spectrom.* 26, 2837–2853.

953 Hendy C.H., 1971. The isotopic geochemistry of speleothems – I, The calculation of the effects of different modes of formation

954 on the isotopic composition of speleothems and their applicability as palaeoclimatic indicators, *Geochim. Cosmochim. Acta*, 35,

955 801– 824.

956 Heinrich H., 1988. Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130,000 years.

957 *Quat. Res.*, 29(2), 142-152.

958 Huntington K.W., Eiler J.M., Affek H.P., Guo W., Bonifacie M., Yeung L.Y., Thiagarajan N., Passey B., Tripathi A., Daëron M.,

959 Came R., 2009. Methods and limitations of ‘clumped’ CO₂ isotope (Δ_{47}) analysis by gas-source isotope ratio mass spectrometry.

960 *J. Mass Spectrom.* 44, 1318–1329.

961 IAEA/WMO, 2019. Global Network of Isotopes in Precipitation. The GNIP Database. Accessible at: <http://www.iaea.org/water>

962 Interwies E., Görlitz S., 2012. The role of water pricing and water allocation in agriculture in delivering sustainable water use in

963 Europe – CASE STUDY CYPRUS. InterSus - Sustainability Services. European Commission Project n° 11589. 35 pages.

964 Jaffey A.H., Flynn K.F., Glendenin L.E., Bentley W.C., Essling A.M., 1971. Precision measurement of half-lives and specific

965 activities of ²³⁵U and ²³⁸U. *Phys. Rev. C*4, 1889–1906.

966 Kallel N., Duplessy J.C., Labeyrie L., Fontugne M., Paterne M., Montacer M., 2000. Mediterranean pluvial periods and sapropel

967 formation over the last 200 000 years. *Palaeogeol., Palaeoclim., Palaeoecol.* 157 (1), 45–58

968 Keinan J., Bar-Matthews M., Ayalon A., Zilberman T., Agnon A., Frumkin A., 2019. Paleoclimatology of the Levant from

969 Zalmon Cave speleothems, the northern Jordan Valley, Israel. *Quat. Sc. Rev.*, 220, 142-153.

970 Kim S.T. and O’Neil J.R., 1997. Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates. *Geochim.,*

971 *Cosmochim. Acta*, 61(16), 3461-3475.

972 Kiro Y., Goldstein S.L., Garcia-Veigas J., Levy E., Kushnir Y., Stein M., Lazar B., 2017. Relationships between lake-level

973 changes & water & salt budgets in the Dead-Sea during extreme aridities in the East. *Mediterr. Earth Planet Sci. Lett.*, 464, 211-

974 226.

975 Kluge T. and Affek H.P., 2012. Quantifying kinetic fractionation in Bunker Cave speleothems using Δ_{47} . *Quat. Sci. Rev.* 49, 82–

976 94.

977 Kluge T., Affek H.P., Marx T., Aeschbach-Hertig W., Riechelmann D.F.C., Scholz D., Riechelmann, S., Immenhauser, A.,

978 Richter, D.K., Fohlmeister, J., Wackerbarth, A., Mangini, A., Spötl, C., 2013. Reconstruction of drip-water $\delta^{18}\text{O}$ based on calcite

979 oxygen and clumped isotopes of speleothems from Bunker Cave (Germany). *Clim. Past*, 9(1), 377–391

980 Kluge T., John C.M., Jourdan A.L., Davis S., Crawshaw J., 2015. Laboratory calibration of the calcium carbonate clumped

981 isotope thermometer in the 25–250°C temperature range. *Geochim. Cosmochim. Acta*, 157, 213-227.

982 Kluge T., Münster T.S., Frank N., Eiche E., Mertz-Kraus R., Scholz D., Finné M., Unkel I., 2020. A 4000-year long Late

983 Holocene climate record from Hermes Cave (Peloponnese, Greece). *Clim. Past Discuss.*, doi.org/10.5194/cp-2020-47, in review.

984 Kroon D., Alexander I., Little M., Lourens L.J., Matthewson A., Robertson A.H., Sakamoto T., 1998. Oxygen isotope and

985 sapropel stratigraphy in the eastern Mediterranean during the last 3.2 Ma. *Proceedings of the Ocean Drilling Program, Scientific*

986 *Results*, 160, 14.

987 Kopp R.E., Simons F.J., Mitrovica J.X., Maloof A.C., Oppenheimer M., 2009. Probabilistic assessment of sea level during the

988 last interglacial stage. *Nature*, 462, 863–867.

989 Kottek M., Grieser J., Beck C., Rudolf B., Rubel F., 2006. World map of the Köppen-Geiger climate classification updated.

990 *Meteorologische Zeitschrift*, 15(3), 259-263.

991 Kolodny Y., Stein M., Machlus M., 2005. Sea–rain–lake relation in the Last Glacial East Mediterranean revealed by $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ in

992 Lake Lisan aragonites. *Geochim. Cosmochim. Acta* 69, 4045–4060.

993 Lachniet, M.S., 2009. Climatic and environmental controls on speleothem $\delta^{18}\text{O}$ values. *Quat. Sci. Rev.*, 28(5-6), 412-432.

994 Lachniet M.S., Bernal J.P., Asmerom Y., Polyak V., 2012. Uranium loss and aragonite–calcite age discordance in a calcitized

995 aragonite stalagmite. *Quat. Geochronology*, 14, 26–37.

996 Lisker S., Vaks A., Bar-Matthews M., 2010. Late Pleistocene palaeoclimatic and palaeoenvironmental reconstruction of the Dead

997 Sea area (Israel), based on speleothems and cave stromatolites. *Quat. Sci. Rev.*, 29, 1201–1211.

998 Litt T., Pickarski N., Heumann G., Stockhecke M., Tzedakis P.C., 2014. A 600,000 year-long continental pollen record from

999 Lake Van, eastern Anatolia (Turkey). *Quat. Sci. Rev.*, 104, 30–41.

1000 Margari V., Skinner L.C., Tzedakis P.C., Ganopolski A., Vautravers M., Shackleton N.J., 2010. The nature of millennial-scale
1001 climate variability during the past two glacial periods. *Nature Geoscience* 3, 127–131.

1002 Martrat B., Grimalt J.O., Lopez-Martinez C., Chaco I., Sierro F.J., Flores J.A., Zahn R., Canals M., Jason H.C., Hodell D.A.,
1003 2004. Abrupt temperature changes in the Western Mediterranean over the past 250,000 years. *Science* 306, 1762–1765.

1004 Martrat B., Jimenez-Amat P., Zahn, R., Grimalt J. O., 2014. Similarities and dissimilarities between the last two deglaciations
1005 and interglaciations in the North Atlantic region. *Quat. Sci. Rev.*, 99, 122-134.

1006 Masson-Delmotte V., Schulz M., Abe-Ouchi A., Beer J., Ganopolski A., González Rouco J.F., Osborn T., 2013. Information
1007 from paleoclimate archives. 383-464

1008 Matthews A., Ayalon A., Bar-Matthews M., 2000. D/H ratios of fluid inclusions of Soreq cave (Israel) speleothems as a guide to
1009 the Eastern Mediterranean Meteoric Line relationships in the last 120 ka. *Chem. Geol.*, 166(3-4), 183-191.

1010 McGarry S., Bar-Matthews M., Matthews A., Vaks A., Schilman B., Ayalon A., 2004. Constraints on hydrological and paleo-
1011 temperature variations in the Eastern Mediterranean region in the last 140 ka given by the δD values of speleothem fluid
1012 inclusions, *Quat. Sci. Rev.*, 23, 919–934.

1013 McManus J.F., Oppo D.W., Cullen J.L., 1999. A 0.5-Million-Year record of millennial scale climate variability in the North
1014 Atlantic. *Science* 283, 971-975.

1015 Meckler A.N., Ziegler M., Millan M.I., Breitenbach S.F.M., Bernasconi S.M., 2014. Long-term performance of the Kiel
1016 carbonate device with a new correction scheme for clumped isotope measurements. *Rapid Commun. Mass Spectrom.* 28, 1705-
1017 1715.

1018 Meckler, A.N., Affolter, S., Dublyansky, Y., Krüger, Y., Vogel, N., Bernasconi, S.M., Frenz, M., Kipfer, R., Leuenberger, M.,
1019 Spötl, C., Carolin, S., Cobb, K.M., Moermann, J., Adkins, J.F., Fleitmann, D., 2015. Glacial-interglacial temperature change in
1020 the tropical West Pacific: A comparison of stalagmite-based paleo-thermometers. *Quat. Sci. Rev.*, 127 (1), 90-116.

1021 Nehme C., Verheyden S., Noble S. R., Farrant A.R., Sahy D., Hellstrom J., Delannoy J.J., Claeys P., 2015. Reconstruction of
1022 MIS 5 climate in the central Levant using a stalagmite from Kanaan Cave, Lebanon. *Clim. Past*, 11(12), 1785-1799.

1023 Nehme C., Verheyden S., Breitenbach S.F., Gillikin D.P., Verheyden A., Cheng H., Noble S., Farrant A., Sahy, D., Salem Gh,
1024 Ph. Claeys, 2018. Climate dynamics during the penultimate glacial period recorded in a speleothem from Kanaan Cave, Lebanon
1025 (central Levant). *Quat. Res.*, 90(1), 10-25.

1026 Nehme C., Verheyden S., Nader F. H., Adjizian-Gerard J., Genty D., De Bont K., Claeys, P., 2019. Cave dripwater isotopic
1027 signals related to the altitudinal gradient of Mount-Lebanon: implication for speleothem studies. *Int. J. Speleol.*, 48(1), 8.

1028 NEEM, 2013. Eemian interglacial reconstructed from a Greenland folded ice core, *Nature* 493, 489–494.

1029 North Greenland Ice Core Project members, 2004. High-resolution climate record of Northern Hemisphere climate extending into
1030 the Last Interglacial period. *Nature* 431, 147–151.

1031 Oppo D.W., Keigwin L.D., McManus J.F., 2001. Persistent suborbital climate variability in MIS5 and Termination II.
1032 *Paleoceanography* 16, 280–292.

1033 Otto-Bliesner B.L., Rosenbloom N., Stone E.J., McKay N.P., Lunt D.J., Brady E.C., Overpeck J.T., 2013. How warm was the
1034 last interglacial? New model–data comparisons. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and
1035 Engineering Sciences*, 371(2001), 20130097.

1036 Regattieri E., Zanchetta G., Drysdale R.N., Isola I., Hellstrom J.C., Roncioni A., 2014. A continuous stable isotope record from
1037 the penultimate glacial maximum to the Last Interglacial (159–121 ka) from Tana Che Urla Cave (Apuan Alps, central Italy).
1038 *Quat. Res.*, 82, 450–461.

1039 Richards D.A. and Dorale J.A., 2003. Uranium-series chronology and environmental applications of speleothems. *Rev. in
1040 Mineral. and Geochem.* 52, 407–460.

1041 Rodríguez-Sanz L., Bernasconi S.M., Marino G., Heslop D., Mueller I.A., Fernandez A., Grant K.M., Rohling E.J., 2017.
1042 Penultimate deglacial warming across the Mediterranean Sea revealed by clumped isotopes in foraminifera. *Sci. Rep.* 7, 16572.

1043 Rohling E.J., Cane T.R., Cooke S., Sprovieri M., Bouloubassi I., Emeis K.C., Schiebel R., Kroon D., Jorissen F.J., Lorre A.,
1044 Kemp A.E.S., 2002. African monsoon variability during the previous interglacial maximum, *Earth Planet Sci. Lett.*, 202, 61– 75.

1045 Rohling E.J., Marino G., Grant K.M., 2015. Mediterranean climate and oceanography, and the periodic development of anoxic
1046 events (sapropels), *Earth Sci. Rev.*, 143, 62–97.

1047 Roucoux K.H., Tzedakis P.C., Lawson I.T., Margari V., 2011. Vegetation history of the penultimate glacial period (marine
1048 isotope stage 6) at Ioannina, north-west Greece. *J. of Quat. Sci.* 26, 616–626.

1049 Rowe P.J., Wickens L.B., Sahy D., Marca A.D., Peckover E., Noble S., Özkul M., Baykara M.O., Millar I.L., Andrews J.E.,
1050 2020. Multi-proxy speleothem record of climate instability during the early last interglacial in southern Turkey. *Palaeogeogr.
1051 Palaeoclim., Palaeoecol.* 538, 109422.

1052 Rozanski K., Araguas L., Gonfiantini R., 1993. Isotopic patterns in modern global precipitation, in: *Climate Change in
1053 Continental Isotopic Record*, Geophysical Monograph Series, Washington DC, AGU, 78, 1–37.

1054 Schwarcz, H.P., 1989. Uranium series dating of Quaternary deposits. *Quat. Int.*, 1, 7–17.

1055 Scholz, D., Kluge, T., Marx, T., Spötl, C., Schröder-Ritzraun, A., Riechelmann, D., Aeschbach-Hertig, W., Mangini, A., Richter,
1056 D.K., Niggemann, S., 2009. Temporal gradient between temperature and the oxygen isotope value of precipitation in the reay
1057 Eemian reconstructed from a sepleothem in western Germany. *Geophys. Res.* 11 abstract.

1058 Scrivner A.E., Vance D., and Rohling E. J., 2004. New neodymium isotope data quantify Nile involvement in Mediterranean
1059 anoxic episodes, *Geology*, 32, 565–568, 2004.

1060 Stocker T.F. and Johnsen S.J., 2003. A minimum thermodynamic model for the bipolar seesaw. *Paleoceanography* 18, 1087.

1061 Thompson W.G. and Goldstein S.L., 2006. A radiometric calibration of the SPECMAP timescale. *Quat. Sci. Rev.*, 25, 3207-
1062 3215.

1063 Tremaine D.M., Froelich P.N., Wang Y., 2011. Speleothem calcite formed in situ: Modern calibration of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$
1064 paleoclimate proxies in a continuously-monitored natural cave system. *Geochim. Cosmochim. Acta* 75, 4929–4950.
1065 Tzedakis P.C., Frogley M.R., Heaton T.H.E., 2003. Last interglacial conditions in southern Europe: evidence from Ioannina,
1066 northwest Greece. *Glob. Planet. Change* 36, 157–170.
1067 Tzedakis P.C., Roucoux K.H., de Abreu L., Shackleton N.J., 2004. The duration of forest stages in southern Europe and
1068 interglacial climate variability. *Science* 306, 2231–2235.
1069 Tzedakis P.C., Pälike H., Roucoux K.H., de Abreu L., 2009. Atmospheric methane, southern European vegetation and low-mid
1070 latitude links on orbital and millennial timescales. *Earth Planet Sci. Lett.*, 277, 307–317.
1071 Uemura R., Y. Matsui K. Yoshimura H. Motoyama, Yoshida N., 2008. Evidence of deuterium excess in water vapor as an
1072 indicator of ocean surface conditions, *J. Geophys. Res.*, 113, D19114,
1073 Uemura R., Kina Y., Shen C.-C., Omine K., 2020. Experimental evaluation of oxygen isotopic exchange between inclusion water
1074 and host calcite in speleothems. *Clim. Past*, 16, 17-27.
1075 Van Geldern R. and Barth Johannes A.C., 2012. Optimization of instrument setup and post-run corrections for oxygen and
1076 hydrogen stable isotope measurements of water by isotope ratio infrared spectroscopy. *Limnology Oceanography Methods* 10,
1077 1024–1036.
1078 Verheyden S., Nader F.H., Cheng H.J., Edwards L.R., Swennen R., 2008a. Paleoclimate reconstruction in the Levant region from
1079 the geochemistry of a Holocene stalagmite from the Jeita cave, Lebanon. *Quat. Res.* 70(3), 368-381.
1080 Verheyden S., Genty D., Cattani O., van Breukelen M.R., 2008b Water release patterns of heated speleothem calcite and
1081 hydrogen isotope composition of fluid inclusions. *Chem. Geol.*, 247(1-2), 266-281.
1082 Verheyden S., Genty D., Deflandre G., Quinif Y., Keppens E., 2008c. Monitoring climatological, hydrological and geochemical
1083 parameters in the Pere Noel cave (Belgium): implication for the interpretation of speleothem isotopic and geochemical time-
1084 series. *Int. J. of Speleol.*, 37(3), 221-234.
1085 Vonhof H.B., van Breukelen M.R., Postma O., Rowe P. J., Atkinson T.C., Kroon D., 2006. A continuous-flow crushing device
1086 for on-line $\delta^2\text{H}$ analysis of fluid inclusion water in speleothems. *Rapid Commun. Mass Spectrom.*, 20(17), 2553-2558.
1087 Waelbroeck C., Labeyrie L., Michel E., Duplessy J.C., McManus J.F., Lambeck K., Labracherie M., 2002. Sea-level and deep-
1088 water temperature changes derived from benthic foraminifera isotopic records. *Quat. Sci. Rev.*, 21(1-3), 295-305.
1089 Wainer K., Genty D., Daeron M., Bar-Matthews M., Vonhof H., Dublyansky Y., Pons- Branchu E., 2011. Speleothem record of
1090 the last 180ka in Villars cave (France): investigation of a large $\delta^{18}\text{O}$ shift between MIS6 and MIS5. *Quat. Sci. Rev.*, 30, 130–146.
1091 Wang Y.J., Cheng H., Edwards R.L., Kong X.G., Shao X., Chen S., Wu J.Y., Jiang X.Y., Wang X.F., An Z.S., 2008. Millennial
1092 and orbital-scale changes in the East Asian monsoon over the past 224,000 years. *Nature* 451, 1090–1093.
1093 Wang P., Jun T., Lourens L.J., 2010. Obscuring of long eccentricity cyclicity in Pleistocene oceanic carbon isotope records. *Earth
1094 Planet Sci. Lett.*, 290, 3-4: 319-330.
1095 Weise A. and Kluge T., 2020. Isotope exchange rates in dissolved inorganic carbon between 40°C & 90°C. *Geochim.
1096 Cosmochim. Acta*, 268, 56-72.
1097 Weißbach, T., 2020. Spectroscopic isotope ratio analysis on speleothem fluid inclusions - analytics and paleoclimatic case
1098 studies. PhD thesis, Heidelberg University, 225 pages. <https://doi.org/10.11588/heidok.00028559>
1099 Ziegler M., Tuenter E., Lourens L.J., 2010. The precession phase of the boreal summer monsoon as viewed from the eastern
1100 Mediterranean (ODP Site 968). *Quat. Sci. Rev.*, 29, 1481–1490.