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Coupled stalagmite – alluvial fan response to the 8.2 ka event and early Holocene palaeoclimate change in Greece

Peckover, E.N., Andrews, J.E.* , Leeder, M.R, Rowe, P.J., Marca, A., Sahy, D.¹, Noble, S.¹, Gawthorpe, R².

School of Environmental Sciences, University of East Anglia, Norwich NR4 7TJ, UK.

¹Geochronology and Tracers Facility, NERC Isotope Geosciences Laboratory, British Geological Survey, Keyworth, Nottingham NG12 5GG, UK.

²Basin and Reservoir Studies Group, Department of Earth Science, University of Bergen, Allegaten 41, 5020 Bergen, Norway.

*Corresponding author email j.andrews@uea.ac.uk

Running header: Terrestrial responses to 8.2 ka palaeoclimate event in Greece

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Abstract

We explore the expression of early Holocene climatic change in the terrestrial Mediterranean of southern Greece. A regional palaeoclimate record from stable isotope and trace element geochemical proxies in an early Holocene stalagmite is compared to the timing of palaeosol development on an early Holocene alluvial fan located less than 100 km from the stalagmite site. Palaeosol development records abandonment of the active part of the studied fan, and is dated using radiocarbon, allowing direct coupling with the climatic signal in the stalagmite. Stalagmite growth between ~12.4 ka, and 6.7 ka was largely coincident with the timing of sapropel 1 in the eastern Mediterranean, with conditions broadly wetter and warmer than the rest of the Holocene. However, $\delta^{13}\text{C}$ values in particular, record a number of more arid periods a short one between 9.2 and 9.1 ka and a longer event documenting episodic, dryness between ~8.8 and 8.2 ka. The widely documented northern hemisphere ‘8.2 ka event’ of cooler and drier

34 conditions has a rather muted $\delta^{18}\text{O}$ climatic signal in common with other stalagmite climate
35 records from the wider Mediterranean. The oldest alluvial fan palaeosols were developing by
36 ~ 9.5 ka, corresponding broadly with drying indicators in the speleothem record at ~ 9.2 ka and
37 a thick rubified palaeosol developed ~ 8.3 to 8.4 ka, indicating pedogenesis within dating error
38 the 8.2 ka event. The speleothem record of episodic dryness, combined with other regional
39 proxies for episodic convective summer rainfall in the period between ~ 8.8 and 8.2 ka, suggest
40 this part of the eastern Mediterranean changed its precipitation pattern from predominantly
41 winter frontal to summer convective. Palaeosol formation on the alluvial fan may have been an
42 allocyclic response to this change. It is plausible that fan-channel incision, driven by temporary
43 development of a ‘flashier’ summer rainfall regime isolated large areas of the fan surface
44 allowing onset of prolonged pedogenesis there.

45

46 **1. Introduction**

47 It is a consensus view today that $\sim 8.47 \pm 0.3$ ka, glacial lakes Agassiz and Ojibway released
48 meltwater into the north Atlantic causing surface water freshening (Barber et al. 1999; Clarke
49 et al. 2004; Alley and Ágústsdóttir 2005). The freshening caused slowdown of meridional
50 North Atlantic deep-water flow from c. 8.4 ka, with rapid deceleration and sea surface
51 temperature reductions at ~ 8.3 ka (Ellison et al. 2006). The event (“ 8.2 ka event”) resulted in
52 near global (Cheng et al., 2009) climatic cooling typically accompanied by drier, windier
53 conditions that increased susceptibility to forest fire (Alley et al. 1997). The meltwater pulse
54 that stimulated the 8.2 ka event is the last of up to 17 similar, albeit mostly smaller, pulses
55 identified in the early Holocene (Teller and Leverington 2004). In addition to the 8.2 ka event,
56 at least two more of these melt water pulses are thought to have produced cool and dry climatic
57 anomalies: one at ~ 9.2 ka (Fleitmann et al., 2008) and the other at ~ 11.4 ka (the Preboreal
58 Oscillation; Fisher et al., 2002). While all three climatic anomalies register clearly in Greenland
59 Ice cores (Vinther et al., 2006) the 8.2 ka event has the best-documented terrestrial climatic
60 expression.

61 In this study we were interested in understanding the expression of these three large, early
62 Holocene climatic anomalies in the terrestrial Mediterranean, specifically in Greece. We set
63 out to do this by constructing a palaeoclimate record from an early Holocene stalagmite (Figs
64 1 and 2). Stalagmites are well suited to this approach as they can be precisely dated using U

65 series, while petrographic fabrics, stable isotope and trace element geochemical proxies can
66 record environmental and climatic variability (e.g. Fairchild and Baker 2012).

67 Despite near-global climatic cooling caused by the “8.2 ka event” its expression is
68 surprisingly patchy, muted or absent in speleothems from wider Mediterranean regions (e.g.
69 Frumkin et al., 1994; Bar Matthews et al. 1999; Zanchetta et al., 2007; Verheyden et al. 2008).
70 It is thus important, where possible, to construct regionally specific, well-resolved speleothem
71 records that provide the best opportunity to attribute climatic effects on regional precipitation,
72 runoff and sediment yields, particularly seasonal distribution, magnitude and source.

73 We also had an unusual opportunity to compare the local well-dated stalagmite
74 palaeoclimate record with the sedimentary response of an early Holocene alluvial fan located
75 at Schinos, less than 100 km from the stalagmite site (Fig. 1). Radiocarbon chronology for the
76 fan, allowed, perhaps for the first time, quantitative analysis of centennial-scale fan response
77 to climatic drivers, taking such studies beyond the coarser temporal and climatic-change scales
78 seen, for example at the Pleistocene-Holocene transition where dated lake shorelines intersect
79 the distal parts of alluvial fans (e.g. Harvey et al., 1999; Garcia and Stokes 2006).

80 Our studied fan displays a number of well-developed palaeosols in its lower part: the
81 thickest and stratigraphically youngest palaeosol having an age of 7620 ± 40 radiocarbon years
82 (Leeder et al., 2002), showing that the fan sediments overlap in age with the early Holocene
83 stalagmite record. Palaeosols in alluvial fan sequences represent abandoned surfaces where
84 sedimentation has temporarily ceased (e.g. Talbot and Williams 1979; Ritter et al., 1995;
85 Reheis et al., 1996; Stokes et al., 2007; Ventra and Nichols 2014; Antinao et al., 2016). Such
86 abandonment may arise from *autocyclic* lobe switching when a currently active channel cuts
87 across the fan topographic gradient to jump sideways (avulse) during flood discharge. The new
88 locus of deposition robs sediment supply to the formerly active fan segment (e.g. Ventra and
89 Nichols 2014), allowing pedogenesis on the abandoned surface. Repeated avulsions and lobe
90 abandonment produce a patchwork fan stratigraphic architecture comprising local soil horizons
91 intercalated within alluvium. Our null hypothesis was thus that the fan palaeosols are
92 ‘autocyclic’ being randomly distributed in time and space and not related to climatic drivers as
93 recorded in the stalagmite.

94 If however, the age of the alluvial fan palaeosols corresponds with climatic events in the
95 speleothem the null hypothesis is challenged. *Allocyclic* forcing of alluvial fan sedimentation
96 is largely non-random, driven by the sensitivity of the entire catchment-fan system to the
97 balance between predominant deposition and the non-deposition. This sensitivity often arises

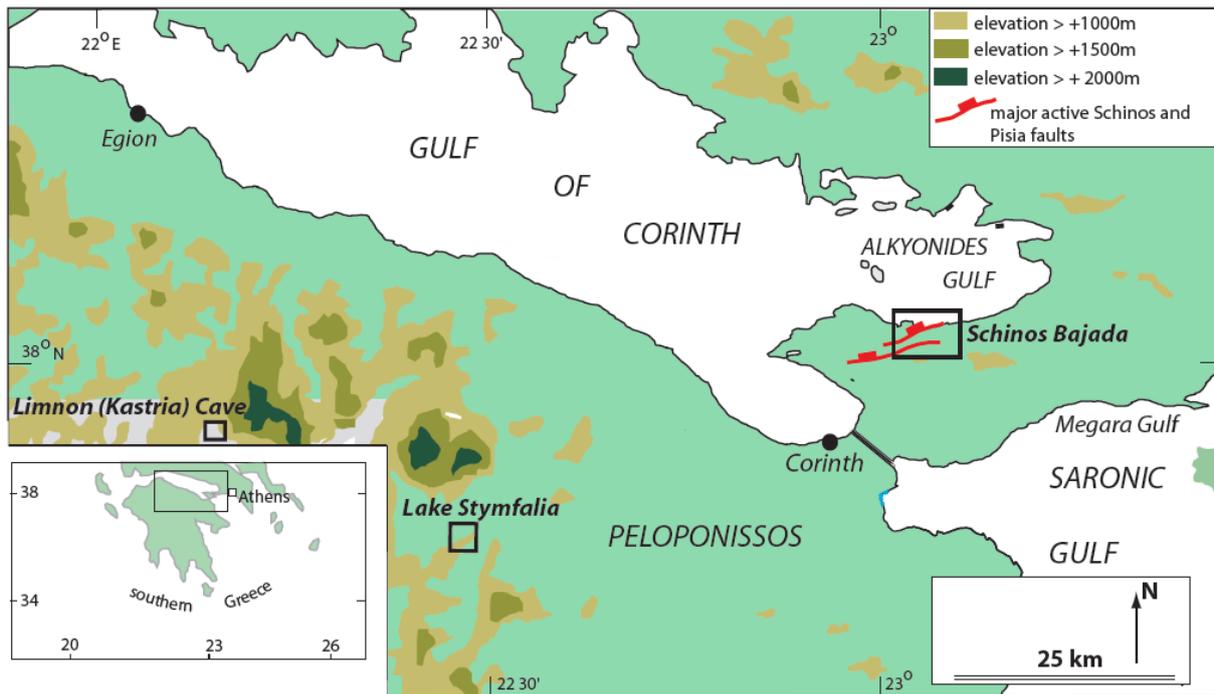
98 through changing hydrology-influenced variables such as seasonal water balance, magnitude
99 and rate of surface and sediment runoff and the density and type of vegetation (Leeder et al.
100 1998). The commonest signal of allocyclic change is from deposition to erosion, caused by fan
101 channel incision. This promotes development of soil horizons on sediment-starved interfluvies
102 left stranded above the flooding levels of the incised and eroding channels. Prolonged incision
103 is likely to be driven by climatic changes on centennial or longer timescales and as such,
104 palaeosol development should align with palaeoclimatic events recorded in the stalagmite. Of
105 course, changing gradients caused by (random) tectonic activity and changing base level may
106 also be important, although these are not necessarily a major factor on Quaternary timescales
107 (Ritter et al., 1995).

108 Our early Holocene stalagmite palaeoclimate record comes from Limnon Cave (37°57'
109 37.8" N 22°08' 24.9"E), 2 km north of Kastria village in the Peloponnese, some 90 km SE of
110 Patras (Figs 1 and 2). The climate record spans the first five thousand years of the Early
111 Holocene, broadly between 12 - 7 ka, significantly older than any published speleothem records
112 (e.g. Finné et al., 2015; Weiberg et al., 2016) from the Peloponnese. The stalagmite record is
113 directly related to dated palaeosol development on the Schinos alluvial fan located in western
114 Attica, some 80 km to the ENE (38°02' 53.7" N 23°02' 54.0" E; Fig. 1). Most important, the
115 time-period of our study precedes any significant impact from human activity on vegetation in
116 the region, which began around 7.0 ka (Weiberg et al., 2016).

117

118 **2. Early Holocene climate (Greece)**

119 Central Asian Holocene climate between 10-6 ka was generally warm and humid, linked
120 to onset of the early Holocene (11.5 ka) solar radiation maximum (Cheng et al., 2012). In the
121 Eastern Mediterranean, warm and humid conditions (Peyron et al. 2011; Cheng et al., 2015)
122 coincide with deposition of sapropel 1 (S1; 10.5 ka to 6.1 ka; Grant et al., 2016) a precession
123 minima phenomenon driven by wetter climate, increased river runoff, nutrification and near -
124 surface stratification (e.g. Mercone et al., 2001; Meyers and Arnaboldi 2008). On land, the
125 warm and humid conditions favoured growth of 'climatic optimum' Mediterranean mixed
126 forests. The best-resolved terrestrial palaeoclimate data in Greece comes from Tenaghi
127 Philippon (Peyron et al. 2011) in the north of the country. Here, pollen-based climate
128 reconstructions show a strongly seasonal (stronger than today) moist period from 9.5-7.8 ka
129 characterised by wet winters and dry summers.

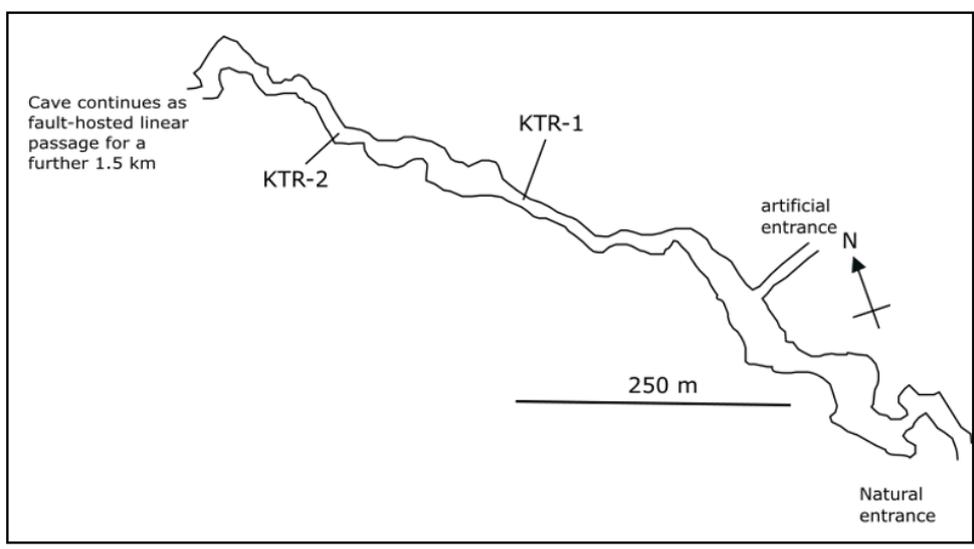
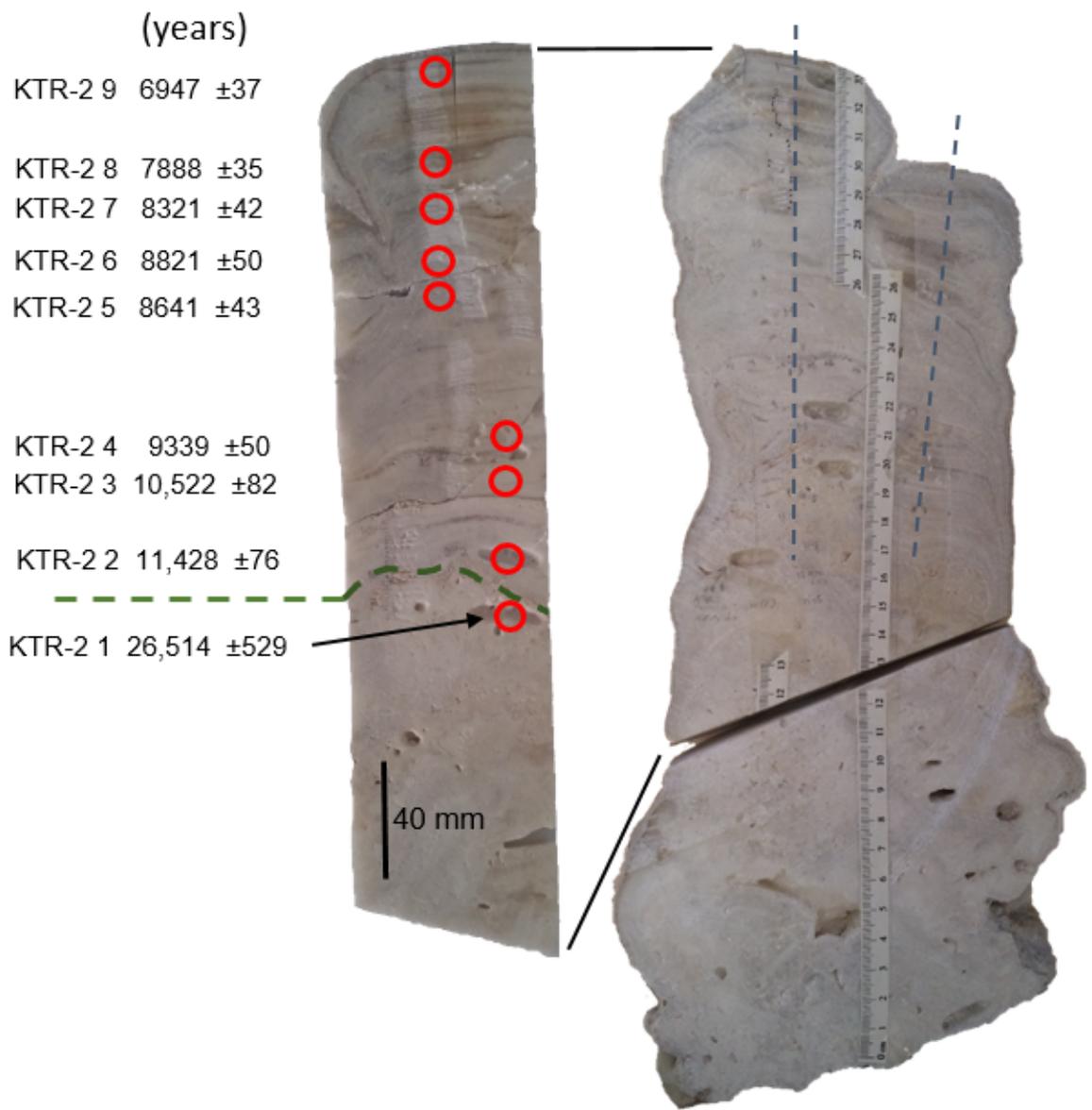


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132 Fig. 1. Main panel shows location of the Schinos Bajada, Limnon Cave and Lake Stymphalia
 133 in the North Peloponnese and Corinth isthmus area of Greece. Insets show wider national and
 134 regional context where A – Athens; P – Patras; TP - Tenaghi Phillipon.

135

136 These optimum conditions were interrupted at 9.2 and 8.2 ka (see above). The 8.2 ka
 137 northern hemisphere climate event is best-documented, dated to 8271 ± 113 years BP (Vinther
 138 et al., 2006) based on electrical conductivity measurements in the Greenland GISP2 ice core
 139 (see also Thomas et al., 2007). Atmospheric teleconnection from the cooled and freshened N
 140 Atlantic surface waters (see above) caused expansion of the northern hemisphere polar winter
 141 vorticity field (Siberian High; see Renssen et al. 2002; Rohling et al. 2002; Vellinga and Wood
 142 2002). Effects of pan-hemispheric cooling ~ 8.2 ka (Alley and Ágústsdóttir 2005) included
 143 reduced growth rates of Central European oaks (Spurk et al., 2002) and changes to deciduous
 144 tree populations (notably *Corylus*; Tinner and Lotter, 2001). Sediment cores in the Aegean Sea,
 145 record the 8.2 ka event superimposed on a broader same-sign climatic anomaly between 8.8-
 146 7.8 ka (Rohling and Pälike 2005; Marino et al. 2009). During this period, Aegean Sea S1
 147 deposition was interrupted as cold winter outbursts from the Siberian High led to surface
 148 cooling, renewed deep-water formation and temporary reversion to ‘normal’ oxygenated
 149 hemipelagic deposition (Kotthoff et al. 2008a). Onset of this interruption began at 8.5 ka



153 Fig. 2. Axial slab of twin stalagmite KTR-2 showing growth axes (blue dashed lines). All
154 samples were taken from the left stalagmite and U/Th sample positions (red circles are
155 accompanied by the dates. StalAge excluded sample KTR-2 6 from the age model (see text).
156 Green dashed line indicates the base of Holocene calcite at 127 mm. Map shows a cave plan
157 of the first 600 m of the cave with stalagmite sample positions. All cave water samples were
158 collected between these positions.

159
160 according to Triantaphyllou et al. (2009). The 8.2 ka climatic anomaly interrupted the
161 otherwise moist conditions and reversed the seasonality (dry winters and wet summers) in
162 pollen records at Tenaghi Philippon (Peyron et al. 2011). However, further south in the
163 Peloponnese, geochemical records in Lake Stymphalia show no clear evidence of
164 environmental perturbation at 8.2 ka (Heymann et al., 2013). From 7.8-5.0 ka, the Tenaghi
165 Philippon terrestrial record suggests lower overall precipitation and reduced seasonality
166 (Peyron et al. 2011) but there are no supporting records from southern Greece.

167

168 **3. Environmental setting**

169 *3.1. Limnon cave stalagmite*

170 Limnon Cave is hosted in Cretaceous Limestones of Mount Amolinitza (summit at
171 1420 m). This montane cave is about 2 km long and developed along a NW trending fault.
172 The natural cave entrance is 820 m above sea level, but in 1981, an artificial entrance was
173 opened for tourist access that began in 1990 (Fig. 2). The first prominent stalactite mass is
174 found 80 m from the natural entrance and the cave is actively wet with cave floor lake
175 development 280 m from the natural entrance. There are thirteen cave floor lakes of various
176 sizes in the following 520 m (Iliopouou-Georgudaki and Economidou 1991). Epikarst
177 thickness increases more or less linearly from the natural entrance to a maximum of 540 m
178 below the summit of Mount Amolinitza. Present day terra-rossa soil cover above the cave is
179 thin and patchy, mostly hosted in bedrock fissures. Vegetation is sparse Mediterranean
180 sclerophyllous scrub characterized by *Quercus coccifera* and *Phlomes fruticosa* (Iliopouou-
181 Georgudaki and Economidou 1991).

182 Today, mean cave air temperature varies between 12.5 °C (winter) and 14.5 °C
183 (summer) with relative humidity between 89% (winter) and 96% (summer; Iliopouou-
184 Georgudaki and Economidou 1991). In summer, air flows in through the entrances and exits
185 via roof fissures with flow velocities of 0.12-0.69 m s⁻¹ (0.5-2.0 m from the cave floor;
186 Iliopouou-Georgudaki and Economidou 1991). This flow reverses in winter. All of these

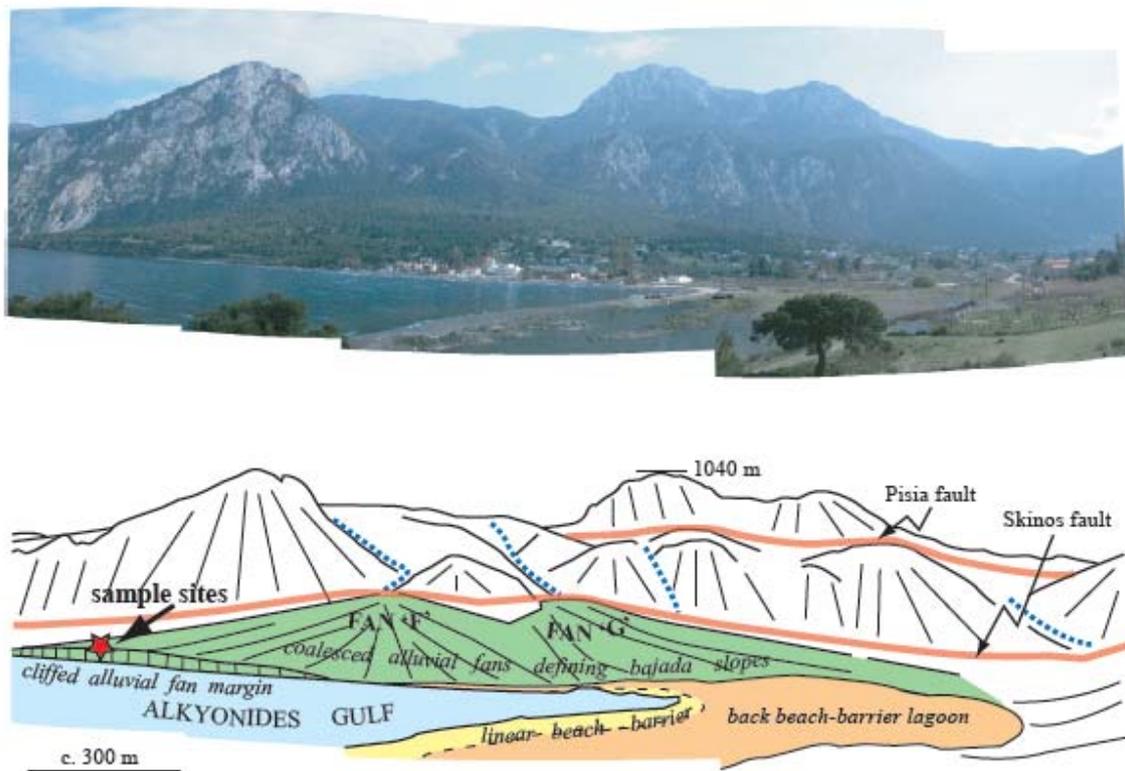
187 measured environmental parameters are clearly affected by the artificial entrance to some
188 degree and it is reasonable to assume that all of them were either lower (temperature range;
189 air flow) or higher (relative humidity) under natural conditions.

190 The closest data-rich IAEA and WMO weather stations are both coastal, at Patras
191 (2000-2014) and Athens (combined Hellinikon Airport (1960-1976) and Pendeli (2001-2014)
192 records). Modern rainfall distribution is strongly seasonal with >70% of precipitation falling
193 in autumn and winter months and <20 mm rainfall in summer months. Stalagmite extension
194 and isotopic compositions should thus respond largely to autumn and winter rainfall. In the
195 Limnon region annual rainfall is around 1200 mm per year (Flocas and Giles 1991) with an
196 annual rainfall relative intensity of 3.3 mm/h (between 1962 and 2002), among the highest
197 values in Greece (Kambezidis et al. 2010). Between 65 and 70 % of regional rainfall in the
198 Peloponnese originates from frontal depressions in the winter (Flocas and Giles 1991), such
199 that a summer convective signal, although present (Kambezidis et al. 2010), is not likely to be
200 much represented in stalagmite growth today.

201 3.2. *Schinus alluvial fans*

202 Opportunities for observing early Holocene sedimentary successions are rare over
203 much of the Aegean hinterland due to younger sedimentary cover. This holds along the
204 majority of the Corinth rift basin where the entry points of major drainages are marked by
205 depositional Holocene coastlines featuring prograding and aggrading fans. However, the
206 southern active-faulted margin to the Alkyonides Gulf in the easternmost rift is undergoing
207 coastal erosion and is bordered by an incised, hanging-wall coastal bajada. This range front
208 bajada comprises coalesced sea-cliffed alluvial fans, talus cones, coastal lagoons, marshlands
209 and beach/barrier spit and beach shorelines (Fig. 3; Leeder et al. 1991, 1998, 2002). The alluvial
210 fans are km-scale, coarse grained, stream-flow dominated, fed from drainage catchments
211 located in uplifting footwall-mountains of the Gerania Range with Mesozoic basement
212 comprising limestone, chert, and ophiolitic serpentinites. It is likely that a bajada system has
213 been present here for c. 2 Ma since initiation of the active coastal faults (Leeder et al. 2002).
214 The youngest lowstand bajada was drowned around 7 ka at the Holocene highstand and up to
215 150 m of coastal retreat may have occurred since. Sea-cliffing of the Holocene alluvial fans
216 began after the 7 ka highstand, amplified by $\sim 1\text{-}2\text{ mm yr}^{-1}$ tectonic subsidence, the fans being
217 in the hangingwall of the active Schinos Fault (Jackson et al., 1982; Collier et al., 1998).

218 This study concentrates on the early Holocene part of Fan F of Leeder et al. (1998,
 219 2002) where the sea cliff and a roadside quarry allow unusually good access to the lower fan
 220 stratigraphy. Fan F has its apex at about 200 m elevation and a basal perimeter of approximately
 221 1.3 km. A single gorge, cut through the limestone range front, fed the fan from a catchment
 222 approximately 1 km long and up to 0.6 km wide (700 m maximum elevation). The fan surface
 223 is densely wooded except parts of the western lower slopes, which have been cleared for house
 224 building. We estimate Fan F to have a present day volume of $4.76 \times 10^7 \text{ m}^3$, and using a
 225 sediment bulk density of 1520 kg m^{-3} , a sedimentary mass of $\sim 7.24 \times 10^{10} \text{ kg}$. Fan F currently
 226 shows no evidence of active channel sedimentation; this reflects exhaustion of readily erodible
 227 ophiolite, combined with exposure of underlying limestone bedrock in the upper catchment,
 228 inducing subsurface (karst) drainage. There is some evidence (discussed below) that the eastern
 229 part of Fan F may have been largely inactive since the mid Holocene.



230
 231 Fig. 3. Panoramic view and interpretive sketch over Schinos Bay looking south from 50 m
 232 above sea level into the footwall scarps and bajada (skyline is c. 850 m above m.s.l.) along
 233 the overlap zone between the Pisias (upper scarp) and Schinos (lower scarp) active normal
 234 faults (for general location see Fig. 1). The large sea-cliffed ('toecut') coalesced alluvial fan
 235 is Fan 'F' (left side) and 'G' (right side) of Leeder et al. (2002), fed by currently inactive
 236 catchments that drain Mesozoic limestone and serpentinite hinterlands.

237 Fan F is located in the hangingwall of both the Schinos and Pisia Faults (Jackson et al.,
238 1982; Collier et al., 1998), and the footwall of the offshore West Alkyonides Fault. However,
239 late Quaternary displacements were dominated by subsidence in the hanging walls of the
240 onshore faults (Leeder et al., 2002; Mechernich et al., 2018). The effects of individual
241 earthquakes on fan morphology are well constrained by research on Fan D, 2 km to the E of
242 our study site. In 1981 a series of three earthquakes (February 24, 1981, 6.7 Ms; February 25,
243 1981, 6.4 Ms; March 4, 1981, 6.4 Ms) struck the Alkyonides Basin (Jackson et al., 1982). On
244 Fan D, surface displacements of between 0.4-1.3 m on the Schinos Fault were recorded,
245 probably formed by one or both of the first two 1981 events (Collier et al., 1998). A maximum
246 recurrence interval of 330 years for such surface breaks has been calculated based on dated
247 historical events (Collier et al., 1998). The presence of a 5 m high scarp on Fan G (~2.5 km
248 SW of our study site) led Collier et al. (1998) to conclude that rates of displacement may be
249 comparable along much of the length of the Schinos Fault, since at least the mid-Holocene.
250 These well-characterised alluvial sediment-hosted fault scarps have surprisingly little effect on
251 overall fan morphology (particularly in the lower fans): in active channel areas they are rapidly
252 degraded/overridden by flood sedimentation events. In short, there is no evidence that
253 individual fault surface scarps have much discernible influence on wider fan sedimentation
254 patterns downslope.

255

256 **4. Materials and methods**

257

258 *4.1. Limnon Cave*

259 Stalagmite KTR-2, was collected about 600 m from the natural entrance (Fig. 2) where
260 epikarst thickness is about 200 m. Cave water samples (Table S1) were collected at the time of
261 sampling and in addition two drip sites (1 and 2) were subsequently monitored in December
262 2006 and January 2007, February, March and April 2009, and February and March 2010. It
263 was possible to sample two modern calcite-water pairs, a small stalagmite (KTR-1) thought to
264 be active at 400 m from the natural entrance (Fig. 2) growing beneath a dripping stalactite (drip
265 1); a nearby straw stalactite (drip 2). The tip of a stalactite drape at 550 m and calcite
266 precipitating on the steel walkway at 500 m was also sampled (Table 1). In addition, water
267 samples from three springs were collected, one 200 m from the natural cave entrance, a second
268 2 km N of the cave (972 m elevation) and a third at Kalavryta, 17 km NNW of the cave (795
269 m elevation).

270 U-Series dating was carried out at the NERC Isotope Geosciences Laboratory,
 271 Keyworth, UK (full method in Supplementary Information) using 200-250 mg samples from
 272 drilled locations (Fig. 2). Uranium and thorium isotope data were obtained on a Thermo
 273 Neptune Plus MC-ICP-MS using an Aridus II desolvating nebulizer and standard-sample
 274 bracketing and instrument procedures modified from Andersen et al. (2008) and Hiess et al.
 275 (2012). Hydride and tailing corrections were on the order of 2 ppm of the adjacent peaks. Total
 276 ^{238}U and ^{232}Th blanks were <10 pg and <4 pg and were negligible relative to the sample U and
 277 Th. Standard accuracy (within 0.1%) and reproducibility (within 0.2%) of $^{234}\text{U}/^{238}\text{U}$ was
 278 monitored by replicate analyses of Harwell uraninite HU-1. Replicate measurements of the
 279 reference solution showed $^{229}\text{Th}/^{230}\text{Th}$ accuracy and reproducibility to be $\pm 0.2\text{-}0.3\%$ for ^{230}Th
 280 ion beams > 5000 cps. Data reduction incorporated the revised average $^{235}\text{U}/^{238}\text{U}$ ratio of
 281 137.818 (Hiess et al., 2012) and U-Th ages were calculated using the decay constants of Cheng
 282 et al. (2013). Holocene U/Th ages have errors $< \pm 82$ years (Table 2) and corrected ages are BP
 283 (before 1950 AD).

284
285

286 Table 1. Stable isotope composition of active calcites in Limnon Cave forming within 150 m
 287 of KTR-2.

	$\delta^{18}\text{O}$ (‰ VPDB)	$\delta^{13}\text{C}$ (‰ VPDB)	T °C (Kim and O’Neil 1997)	T °C (Tremaine et al., 2011)
Active straw stalactite	-6.3	-8.7	9.4	13.1
Top of active stalagmite KTR-1	-6.2	-8.5	8.9	12.6
Active stalactite drape	-6.2	-7.9	8.9	12.6
Calcite deposit on metal walkway	-6.8	-11.6	11.6	15.7
Mean	-6.2	-8.3		

288
 289 The temperatures are calculated from the calcite $\delta^{18}\text{O}$ values using the equilibrium equation of
 290 Tremaine et al. (2011) and, for comparison, that of Kim and O’Neil (1997). A mean cave water
 291 $\delta^{18}\text{O}$ of -7.5‰ VSMOW was used for all temperature calculations (see text). Values in bold
 292 calculated from Tremaine et al. (2011) overlap with measured winter cave temperatures.
 293 Walkway deposit value is probably disequilibrium (see Supplementary information) and is
 294 excluded from means.

295

296 Petrography on KTR-2 was done using multiple overlapping thin-sections and samples
297 for stable isotope analysis were drilled at 1 mm spacing through the axial part of the stalagmite
298 (127 samples; hereafter low-resolution). In addition, a 33 mm section between 33 and 66 mm,
299 was micro-milled at high-resolution in an attempt to capture details of any 8.2 ka signal.
300 Samples were drilled in trenches ~250 μm wide and ~100 μm deep, normal to the growth axis.
301 Each sample 'sweep' abutted the preceding one such that sample trenches were quasi
302 continuous along the growth axis. This sampling achieved decadal resolution based on the age
303 model. Isotopic analyses (University of East Anglia Stable Isotope Laboratory) were made on
304 75 ± 5 μg samples, run alongside 75 ± 5 μg internal standards of UEACMST (University of East
305 Anglia Carrara Marble Standard; $\delta^{18}\text{O}$ -2.05 ‰ VPDB; $\delta^{13}\text{C}$ 1.99 ‰ VPDB), reacted with 105%
306 ($\rho = 1.92$ gml⁻³) phosphoric acid (H_3PO_4) at 90°C in an on-line common acid bath. The
307 evolved CO_2 was purified and analysed for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ using a Europa SIRA II dual inlet
308 isotope ratio mass spectrometer. The data are calibrated to international reference scales
309 (VPDB and VSMOW) using IAEA Certified Reference Material NBS-19 ($\delta^{18}\text{O}$ -2.20
310 ‰ VPDB; $\delta^{13}\text{C}$ 1.95 ‰ VPDB). Repeat analysis of both international and internal reference
311 materials gave 1σ errors of less than $\pm 0.1\%$ for both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$.

312 A tablet immediately adjacent to the micromilled section was used for laser ablation
313 ICPMS trace element transect, using a spot size of 30 μm and increment between spots of 200
314 μm (method in Royle et al., 2015). Sr and Mg data were highly reproducible with RSDs of
315 2.8% and 3.8% respectively. Exact matching of micro-milled samples and LA-ICPMS spots
316 was not possible due to the poor the optics of the laser microscope and the destructive style of
317 drilling, but sample widths of both techniques were close enough to ensure that decadal-scale
318 sampling coherence was achieved.

319

320 *4.2. Alluvial fan palaeosols*

321 The alluvial fan studies focus on the palaeosols and associated sediments in the basal
322 10 m of the Schinos bajada alluvial fan F of Leeder et al. (1998, 2002); field colours were
323 recorded with reference to Munsell colour chips. Samples for radiocarbon analysis were
324 taken from the upper few centimetres of the palaeosols, excavated >5 cm behind pre-cleaned
325 vertical surface exposures, taking great care to avoid modern root material. Bulk sediment
326 was processed by Beta Analytic and the AMS dated material is the organic fraction remaining
327 after sieving the sediment to <180 μm to remove any roots or macrofossils and then acid
328 washed to remove carbonate. The organic component in these oxic sediments is assumed to

329 be finely-disseminated inert micro-charcoal, accumulated as a concentrate from wildfires.
330 Radiocarbon dates have errors $< \pm 40$ years and were converted to calibrated age ranges BP
331 (before 1950 AD) using INTCAL 13 (Reimer et al. 2013; Table 2).

332

333 **5. Results**

334

335 *5.1. Limnon cave stalagmite*

336 KTR-2 is a twin stalagmite; we sampled the left hand stalagmite shown in Figure 2 and
337 nine U-series dates were used to constrain ages (Table 2). The two lower dates bracket a
338 detritus-rich horizon ~ 2 mm in width, which lies at or just below the Pleistocene-Holocene
339 boundary (top at 127 mm). The data show evidence of slight detrital contamination ($^{230}\text{Th}/^{232}\text{Th}$
340 = 99-310) and ages have been corrected assuming a contaminant of bulk Earth composition
341 with a Th/U weight ratio of 3.8 (Taylor and McLennan 1995) and ^{238}U , ^{234}U and ^{230}Th in secular
342 equilibrium. Age corrections are generally ≤ 100 years, although ~ 160 years for KTR2-2 and
343 KTR2-3. The dates are in stratigraphic order except KTR2-5 and KTR2-6 with an age
344 differential of 180 years, and stratigraphically inverted beyond 2σ errors: there are no obvious
345 geochemical grounds to prefer one date over the other (Table 2). KTR2-5 (8641 ± 43 years BP)
346 has a lower $^{234}\text{U}/^{238}\text{U}$ ratio than all other samples, possibly indicating uranium isotope mobility,
347 but KTR2-6 (8821 ± 50 years BP) may have experienced detrital contamination by sediment
348 with lower Th/U ratio than bulk Earth leading to higher ^{230}Th and increased age. The
349 speleothem age modelling program StalAge (Scholz and Hoffmann, 2011), rejected sample
350 KTR2-5; however, this solution requires an implausible 10.2 mm of stalagmite extension in 3
351 years between 61.82 to 72.0 mm. On this basis we think it likely that KTR2-5 is the more
352 reliable age and it was incorporated into the StalAge model used for subsequent data
353 interpretation (Fig. 4).

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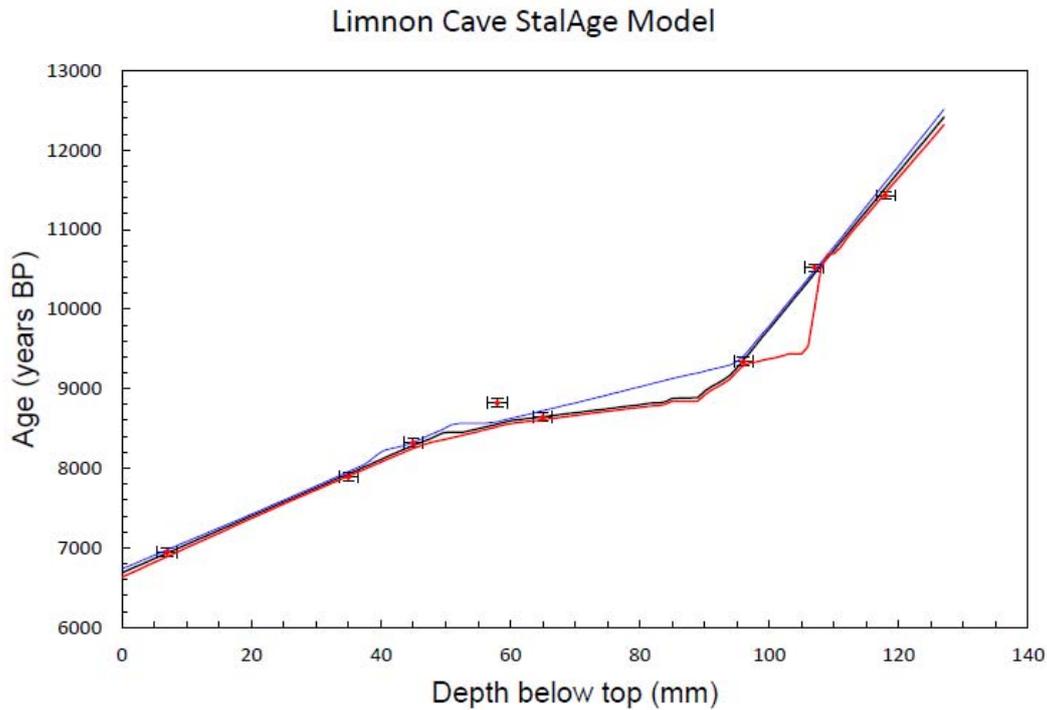
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Table 2. U-series data for stalagmite KTR2.

Sample Number	Depth from top (mm)	²³⁸ U (ppm)	²³² Th (ppb)	(²³⁴ U/ ²³⁸ U)	(²³⁰ Th/ ²³⁸ U)	(²³⁰ Th/ ²³² Th)	(²³⁴ U/ ²³⁸ U) ₀	Age Uncorrected (years BP)	Age Corrected (years BP)
KTR2 9	7	0.1778	0.1271	1.0209(12)	0.0636(32)	271.6	1.0213(12)	7033 ± 34	6947 ± 37
KTR2 8	35	0.2841	0.1985	1.0094(11)	0.0710(29)	309.9	1.0096(11)	7973 ± 32	7888 ± 35
KTR2 7	45	0.2290	0.2618	1.0065(11)	0.0745(35)	199.1	1.0066(11)	8419 ± 34	8321 ± 42
KTR2 6	58	0.1681	0.2714	1.0093(12)	0.0790(43)	149.7	1.0096(12)	8934 ± 38	8821 ± 50
KTR2 5	65	0.2173	0.2858	1.0029(12)	0.0770(36)	178.8	1.0030(11)	8746 ± 33	8641 ± 43
KTR2 4	96	0.1796	0.2306	1.0063(12)	0.0832(42)	197.9	1.0064(12)	9442 ± 43	9339 ± 50
KTR2 3	107	0.2575	0.8751	1.0082(12)	0.0933(69)	84.3	1.0084(12)	10,688 ± 41	10,522 ± 82
KTR2 2	118	0.2259	0.7026	1.0105(15)	0.1011(64)	99.7	1.0088(12)	11,585 ± 40	11,428 ± 76
KTR2 1	137	0.2147	5.4331	1.0308(36)	0.2232(42)	27.5	1.0332(38)	27,312 ± 930	26,514 ± 529

362

363 Sample weights ~150 mg. Note age inversion in samples KTR2-5 and KTR2-6 (shaded), the
364 latter was omitted from the age model (see text). All errors are 2σ. Isotope ratios in brackets
365 denote activity ratios and were calculated using the decay constants of Cheng et al. (2013).
366 Numbers in parenthesis are ratio errors for the last reported digits. Ages BP refer to 2016, the
367 date of analysis. Ages were corrected assuming a contaminant of bulk earth composition with
368 a Th/U weight ratio = 3.8 (Taylor and McLennan, 1995), assumed error of 50% and ²³⁸U, ²³⁴U
369 and ²³⁰Th in secular equilibrium.



370

371 Fig. 4. U-Th age-depth model derived by StalAge for the Holocene section of the KTR-2 (data
 372 in Table 2) excluding KTR2-6 (see text). Upper (blue) and lower (red) lines represent 2 s.d.
 373 errors. There is only one likely minor hiatus (at 96 mm), indicating largely continuous
 374 extension.

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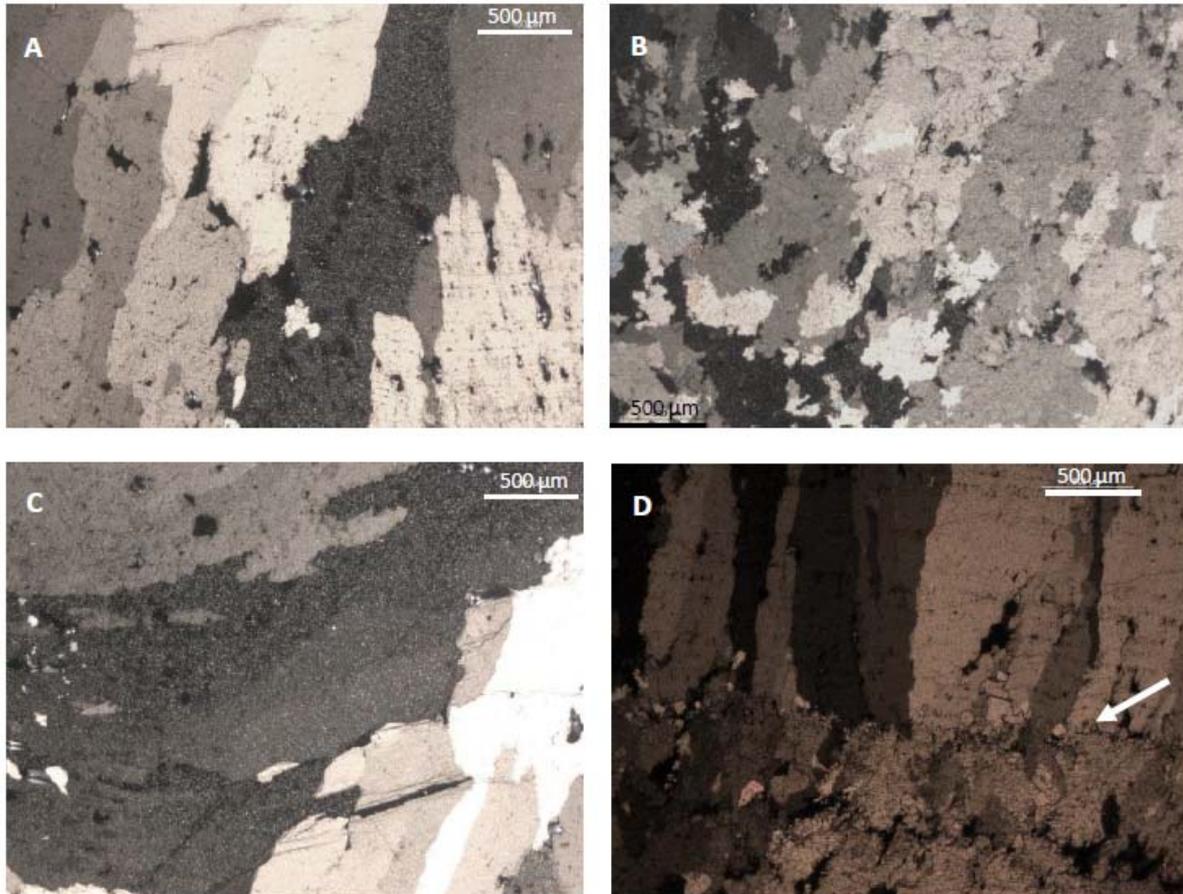
377 An age model that includes KTR2-5 but excludes KTR2-6 modifies ages by <50 years
 378 relative to the opposite selection, except between 8.5 ka and 9.0 ka where ages differ by up to
 379 270 years at 8.9 ka (Fig. S1). However, the magnitude of the variation between the two model
 380 chronologies is insufficient to have a significant impact on the palaeoclimatic reconstructions
 381 discussed in this paper. The dates shows that speleothem growth above the basal detritus-rich
 382 horizon began ~12.4 ka and continued until ~6.7 ka. There is only one likely minor hiatus (at
 383 96 mm), indicating largely continuous extension. Stalagmite extension rates began ~1.1 cm
 384 ka⁻¹ between 12.4 ka and 8.9 ka increasing to ~8.4 cm ka⁻¹ between 8.9 ka and ~8.5 ka before
 385 falling to ~2.9 cm ka⁻¹ from ~8.5 ka to 6.7 ka (Fig. 4).

386

387 Petrography shows that KTR-2 is wholly calcitic mostly of columnar open (Co) fabric
 388 (Frisia 2015) with patchy horizons of columnar microcrystalline (Cm) fabrics (Frisia 2015)
 389 seen below 60 mm, particularly at 115 and 96 mm (Figs 5a, b). Cm fabrics are in places
 accompanied by irregular calcite crystals that grew laterally toward the speleothem flank (Fig.

390 5c); a prominent black horizon in hand specimen at 32 mm also contains lateral crystal growth
391 fabrics. The horizon at 96 mm contains the only evidence of clay-rich detritus as a layer that
392 truncates lateral crystal growth (Fig. 5d). The top 2 mm of the stalagmite shows evidence of
393 post-growth corrosion.

394



395

396 Fig. 5. Thin section photomicrographs of KTR-2 fabrics. a) typical columnar open (Co) calcite,
397 b) columnar microcrystalline (Cm) calcite; c) horizontal growth of Co calcite from left flank at
398 31 mm (~7.8 ka); d) abrupt transition (arrow) between Cm and Co calcites defined by a detritus-
399 rich layer at 96 mm (~9.4 ka).

400

401

402 The most instructive isotopic and trace element data are shown in Figures 6-8, with
403 additional data and figures available in the Supplementary Information. The $\delta^{18}\text{O}$ record (Fig.
404 6a) begins with values around -6.5‰ with low variability until ~ 8.8 ka, after which $\delta^{18}\text{O}$
405 generally increases to around -6.2‰ . Superimposed on this are $\sim 1\text{‰}$ negative excursions at 8.5
406 and 8.0 ka, interspersed with a positive excursion centred on 8.2 ka and most clearly seen in

407 the micro-milled data (Fig. 7a). Another marked positive shift occurs at 7.0 ka (~1.4‰) prior
408 to termination of growth. $\delta^{13}\text{C}$ fluctuates around -5.5‰ before 10.3 ka and mostly between -
409 6.0 to -8.0‰ thereafter (Fig. 6b). However, variability in $\delta^{13}\text{C}$ is high throughout, and a period
410 of less negative values between -5.5 and -6.0‰, is evident between ~8.8 and 8.1 ka (Figs 6b
411 and 7b): even in this period there is a negative excursion at ~8.5 ka. Excursions to less negative
412 $\delta^{13}\text{C}$ are clear in the micro-milled record at 8.3 ka and between 8.2 and 8.1 ka (Fig. 7b), the
413 latter coincident with the highest $\delta^{18}\text{O}$ value. A major positive excursion at the end of the record
414 matches that in $\delta^{18}\text{O}$.

415 There is no obvious relationship between variation in high-resolution $\delta^{13}\text{C}$ and any trace
416 element (Figs S3-S5). However, the smoothed high-resolution $\delta^{18}\text{O}$ (Fig. S2) shows some
417 similarity (within dating error) to smoothed trends in molar Mg/Sr, where Sr content is used as
418 a surrogate for Ca variation (Fig. 8); this ratio is inferred to record epikarst processes including
419 source effects (Roberts et al., 1999; Fairchild et al., 2000), residence times and degassing and/or
420 prior precipitation (see Brasier et al., 2010). There is also weak relationship between Mg/Ca
421 (and Sr/Ca) vs $\delta^{18}\text{O}$ profile shape, particularly the first 200 years of the record (Fig. 8) and
422 again from ~8.3 ka to the end of the record at 7.9 ka. Mg/Ca generally increases from ~8.4 ka
423 to ~8.1 ka but with a marked reversal near the 8.2 ka peak in $\delta^{18}\text{O}$ (-6.2 ‰): Mg/Sr (and Sr/Ca)
424 show a substantial increase at this point (Fig. 8).

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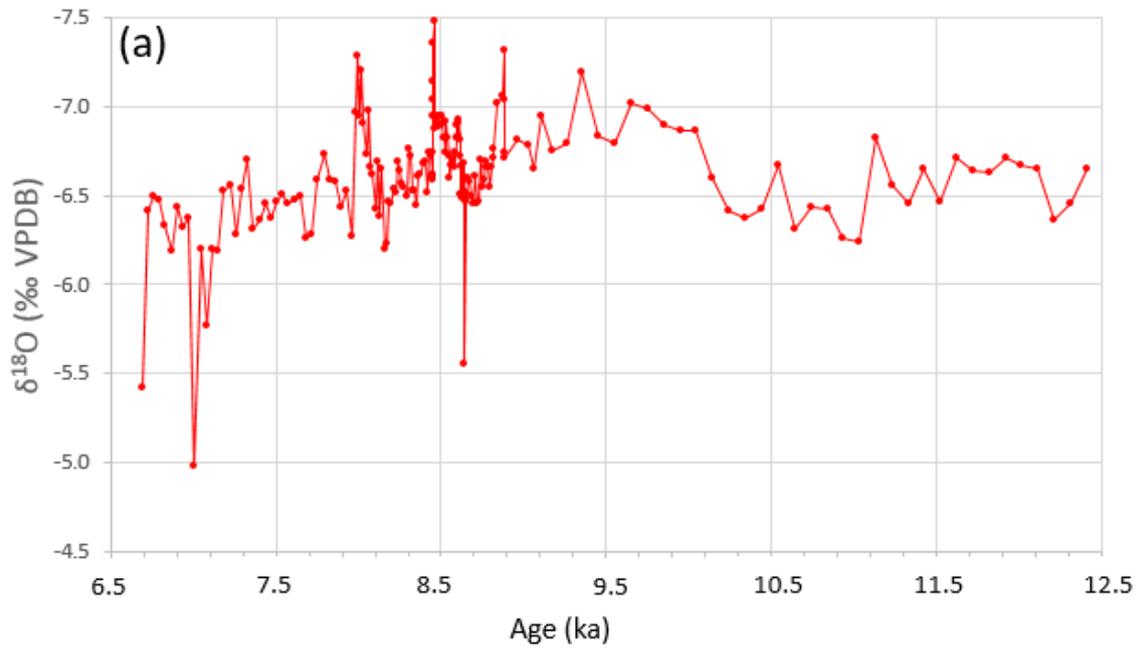
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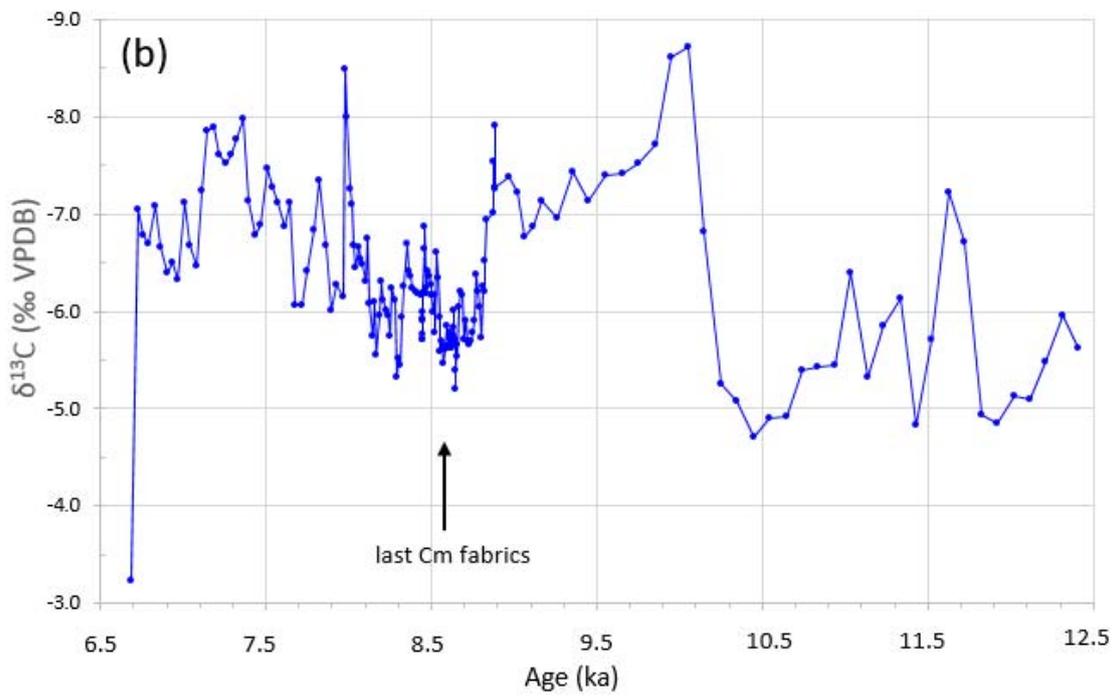
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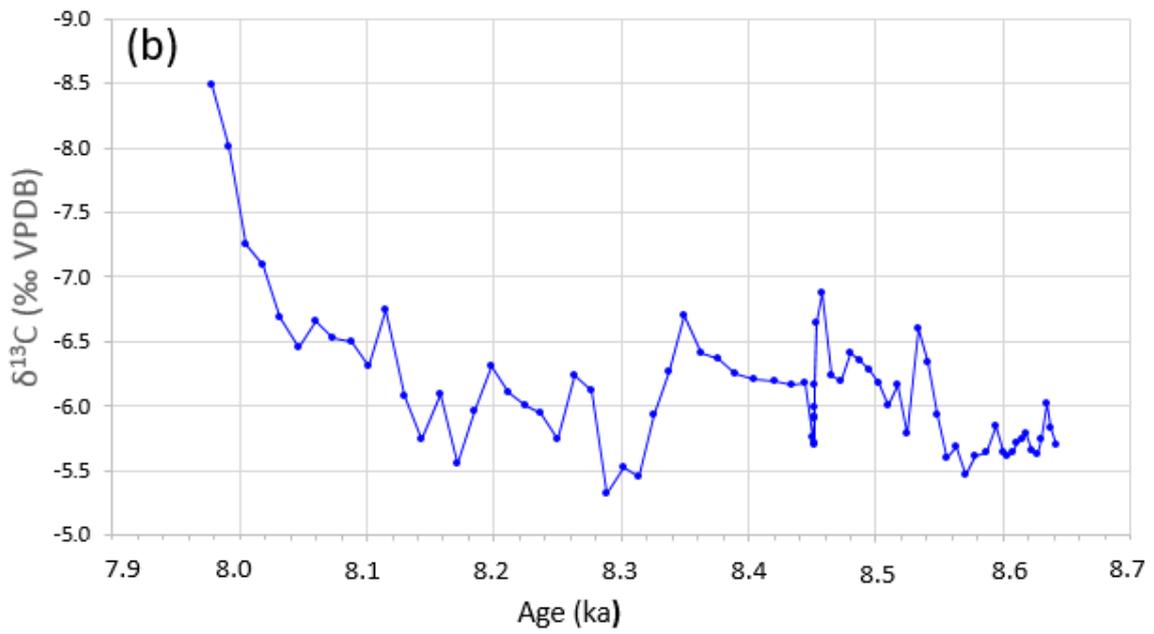
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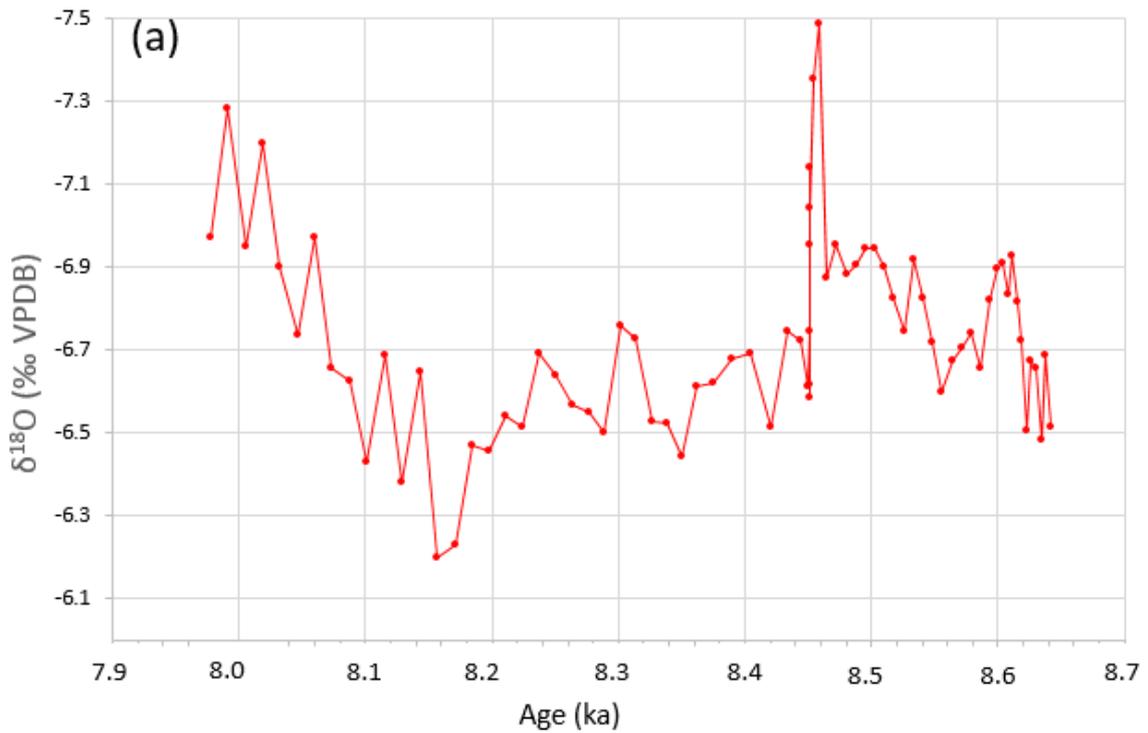
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434 Fig 6. Axial low resolution $\delta^{18}\text{O}$ (panel a, red) and $\delta^{13}\text{C}$ (panel b, blue) data plotted on the
 435 StalAge timescale (Fig. 4).

436



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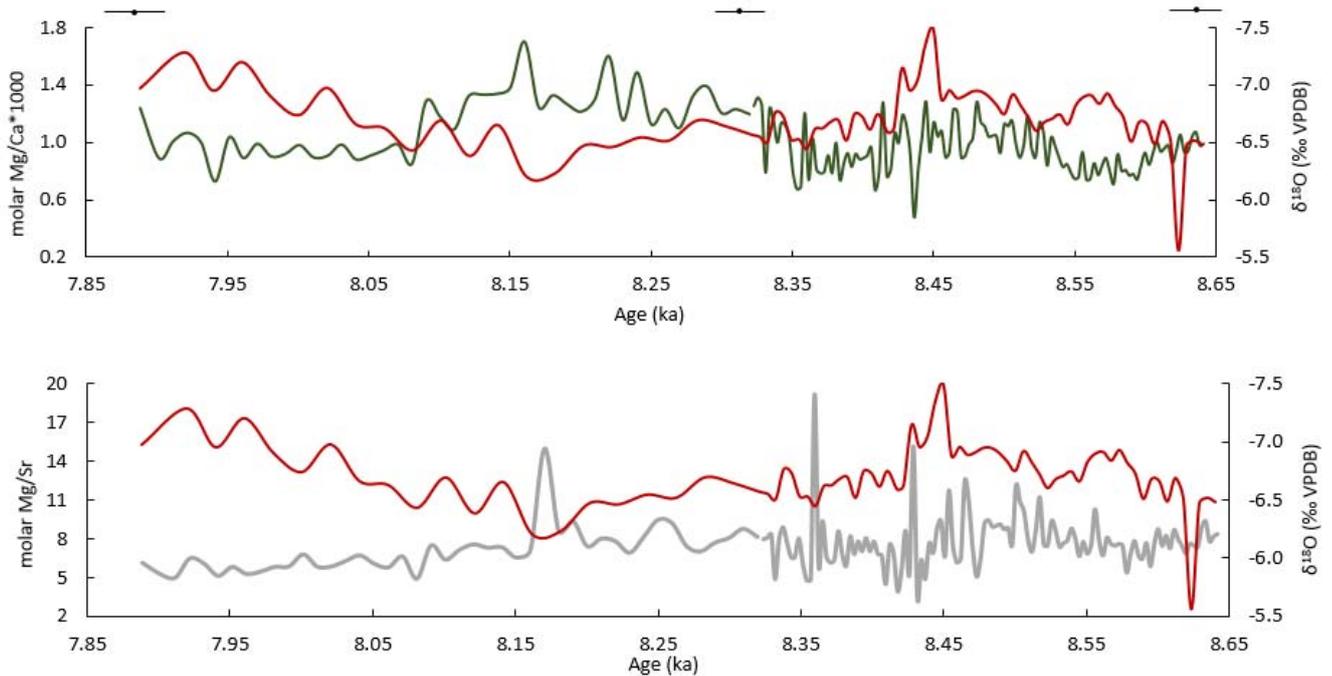
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440 Fig 7. Axial high resolution $\delta^{18}\text{O}$ (panel a, red) and $\delta^{13}\text{C}$ (panel b, blue) micro-milled data
 441 plotted on the StalAge timescale between 8.6 and 7.9 ka.

442

443



444
 445 Fig. 8. Variation in high-resolution $\delta^{18}\text{O}$ (red) and Mg/Ca (green; upper panel) and Mg/Sr
 446 (grey; lower panel) showing some similarity in profile shape. The age model is constrained by
 447 three U/Th ages shown as black dots with error bars above the plot.

448

449 5.2. Alluvial fan palaeosols

450 The sea-cliffed fans have vertical faces up to 9 m high that expose crudely-stratified,
 451 dm-m thick, lenticular-bedded, open framework gravels that are grain-supported, comprising
 452 subangular to subrounded serpentinite/limestone clasts (Fig. 9a and b). The lenticular-bedded
 453 gravel units are in places accentuated by intercalated pale-coloured (2.5Y 8/2) fine silt of
 454 serpentinite composition (checked by XRD), probably reworked by water flow downslope
 455 since the silt has intercalated sand and granule stringers and is often cut out laterally by dm-
 456 scale erosional scours hosting coarser sediment (Fig. 9b). Intercalated palaeosols (Fig. 9a and
 457 b) range from centimetric- to decimetric-thick, brownish to red (see Munsell colours below)
 458 iron-rich horizons that partition the alluvial sediments into successive units (Fig. 10). The
 459 INTCAL 13 calibrated ages of the palaeosols are shown in Table 3 and on Figures 9-11.
 460 Palaeosol dates are mean residence time (MRT) ages, the average age of the organic carbon
 461 component in the sample. MRT ages are typically older than the age of the latest soil
 462 development and can suffer from reworking of older material into the soil (Collier et al., 1998).
 463 We thus consider the palaeosol ages as maxima and accept that the real age of the soils could

464 be younger. Likewise, MRT ages on silt layers will suffer incorporation of older material into
465 the sediment during deposition and are thus likely to be older than the depositional age. This
466 said, the overall stratigraphic consistency of the ages suggest they are representative.

467 The oldest lower fan sediments crop out at modern sea-level in a 4.4 m high sea cliff
468 (38°03' 10.1" N 23°03' 08.6"; Figs 9 and 10). The oldest sediments here are streamflow
469 alluvium with a thin, irregular, mottled pale brown - brown (2.5Y 8/2 and 7.5YR 5/4) coloured
470 palaeosol, dated to 9.3 - 9.5 ka (sample JEA140916-3; Table 3). The 2.95 m thick gravel unit
471 overlying this basal palaeosol is particularly rich in limestone clasts, its upper part punctuated
472 by an erosion surface of prominent clasts with low matrix content. Above this is a prominent,
473 laterally continuous palaeosol, 30 cm thick, that weathers light red (10R 6/8) in its upper part.
474 The sample from this palaeosol gave an age of 9.5-9.7 ka (sample JEA140916-5; Figs 9-10).
475 The age is thus apparently the same or older than the underlying palaeosol, the degree of
476 inversion depending on the calibration age range chosen, and may be indicative of inaccuracies
477 with MRT ages (see above).

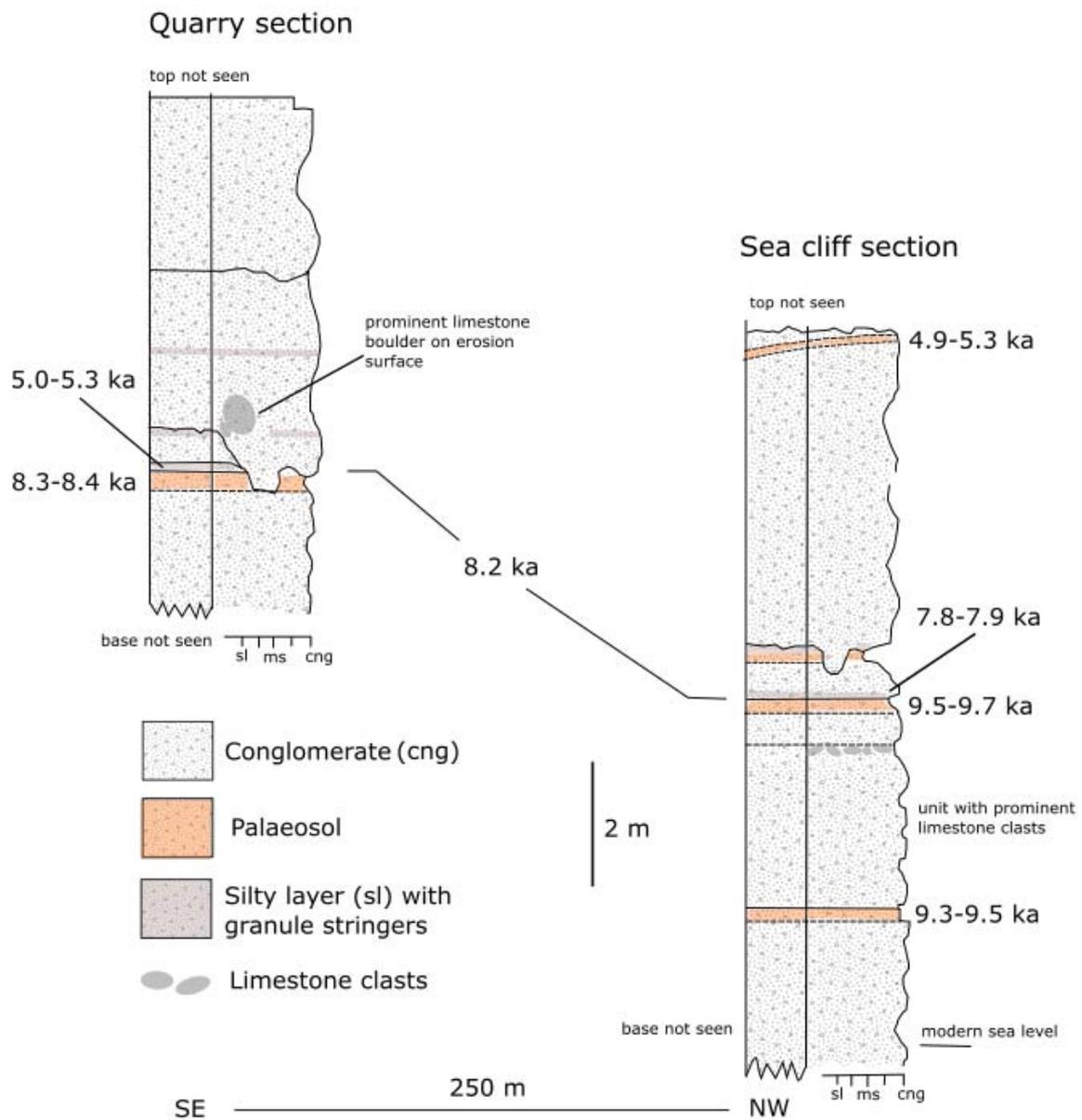


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479

480 Fig. 9. Palaeosols in the modern seacliff. a) Upper part of a mature 30 cm thick light red
481 palaeosol (~9.5-9.7 ka) sharply overlain by a silty layer (~7.8-7.9 ka). The overlying 60 cm of
482 streamflow gravels are topped by a thin, discontinuous palaeosol overlain by another silty layer.
483 b) Showing lateral continuity and sharp upper surface of the 30 cm thick palaeosol, also
484 erosional gutters cutting through the upper silt layer and discontinuous palaeosol.



485

486 Fig. 10. Stratigraphic logs for Fan F (location on Fig. 3) showing stream flow gravel units, silty
 487 layers, erosion surfaces and palaeosol horizons with calibrated radiocarbon date ranges (Table
 488 3). Time marker for 8.2 ka shows temporal correlation. Between ~9.5 ka and 8.0 ka, palaeosol
 489 (orange shade) development was ongoing, possibly intensifying from NW to SE with time.
 490 Palaeosols are probably compound with numerous non-deposition time breaks. The streamflow
 491 gravel lobe above the 5.0 ka palaeosol in the sea cliff section is probably thickening to the SE
 492 (apparent dip on palaeosol ~9° E but its exact relationship with the post 5.0 ka lobe in the quarry
 493 section is not exposed.

494

495

496 Table 3. AMS ¹⁴C dates of palaeosol and associated fine-grained matrix sediments from
 497 Schinos Fan F.

	Sample No	Lab. No	Conventional ¹⁴ C age (1σ) (radiocarbon years)	IntCal13 2σ calibration (cal. years BP)
Schinos fan F, roadside quarry				
Light-grey fine silt unit 10 cm above overlying palaeosol (sample MRL109914-5).	MRL090911-4	Beta-309118	4460±30	4970-5280
Upper 10 cm of prominent 30 cm thick light red palaeosol, 1.5 m above base of quarry floor.	MRL109914-5	Beta-150165	7620±40	8360-8440
Schinos fan F sea cliff section				
Impersistent light reddish brown palaeosol, 5.7 m above palaeosol sample JEA140916-5.	JEA140916-6	Beta-448200	4440±30	4890-5275
Light-grey fine silt unit, 10 cm above palaeosol sample JEA140916-5.	JEA140916-4	Beta-448198	7030±30	7795-7935
Prominent light reddish brown palaeosol horizon, 20 cm thick, 5.05 m above modern sea-level	JEA140916-5	Beta-448199	8660±30	9545-9680
Light brown palaeosol horizon, 20 cm thick, 2.0 m above modern sea-level	JEA140916-3	Beta-448197	8350±40	9280-9470

498 All dates by *Beta Analytic Inc.* calibrated using IntCal13. Age for sample MRL109914-5
 499 originally published in Leeder et al. (2002).

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502

503 This prominent palaeosol is sharply overlain by a silty lens up to 20 cm thick which
504 returned an age of ~7.8 to 7.9 ka (sample JEA140916-4; Table 3 and Figs 9 and 10). Both the
505 silts and the palaeosol are locally cut-out by erosional scour surfaces. The silts are in turn
506 overlain by 60 cm of pebbly gravels topped by a weakly developed, 5 cm thick palaeosol (not
507 dated) overlain sharply by another 10 cm thick silt layer. This silty layer is also cut out locally
508 by gutter-like erosional scours (Fig. 9b). The upper cliff section comprises c. 5 m of streamflow
509 gravels (Fig. 10) with a thin, discontinuous, yellowish red (5YR 5/6) palaeosol at the top which
510 returned an age of 4.9 -5.3 ka (sample JEA140916-6; Table 3). This palaeosol underlies a gravel
511 lens that thickens eastward.

512 An east-west quarry section at about 7 m elevation (38°03' 05.7" N 23°03' 15.8"; Figs
513 10 and 11), 250 m SE of the seacliff section features a striking 0.3 m thick laterally-continuous,
514 red (10R 5/8) ferralitic palaeosol which caps a >2 m thickness (base not seen) of streamflow
515 alluvium (Figs 10 and 11a). It is itself overlain by c. 6 m of serpentinite-rich alluvial gravels.
516 The palaeosol here comprises light-red (10R 6/8) surface coats to serpentinite clasts and clay-
517 matrix in its upper 10 cm, with rubification decreasing downwards to more red-brown (10R
518 4/6 to 2.5YR 4/6) hues. The topmost few cm were sampled (MRL109914-5; Table 3 and Fig.
519 11b) returning an age of 8.3 to 8.4 ka. Overlying streamflow gravels have a sharp and
520 sometimes erosive contact with the palaeosol (Fig. 11b). Sample MRL090911-4A (Table 3)
521 was from a pale coloured (2.5Y 8/2), poorly-sorted sandy-granuley coarse silt in an irregular
522 lens-shaped unit, in the base of the post-palaeosol alluvium, 10 cm above the top of the 8.3 to
523 8.4 ka palaeosol (Fig.11b). This sample returned an age of 5.0-5.3 ka (Table 3). The silt layer
524 is overlain by 60 cm of gravel, followed by a second 10 cm thick silty layer, locally cut by
525 erosional scours that in places also cut through the underlying gravels and the palaeosol.

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535 Fig. 11. a) Panorama of the c. 200 m lateral extent of prominent light red palaeosol in roadside
536 quarry to the SE of the cliffed lateral margin to Fan F (Fig. 3). b) Close-up of the main palaeosol
537 with the shallow-scoured base to the angular fan gravels that contain light-grey fine silt-with-
538 granules units 10 cm or so above the underlying palaeosol. The palaeosol gave an 8.3 to 8.4 ka
539 age, while the base of the light grey gravel unit above the palaeosol gave an age of 4.9 to 5.3
540 ka. The junction between soil and gravel represents a depositional hiatus of some 3000 years.

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6. Interpretation of results

6.1. Stable isotopes background

Isotopic compositions of IAEA Patras precipitation augmented with OIPC v3.1 interpolated precipitation values for the Limnon Cave region (www.waterisotopes.org; Bowen and Wilkinson 2002; Bowen and Revenaugh 2003) mostly plot on, or just above, the GMWL (Fig. S6) while modern cave drip and pool waters, and nearby spring waters (Table S1) plot midway between the GMWL and the EMMWL, with similar gradient. In common with most Mediterranean localities, winter precipitation in the Peloponnese is ~6 ‰ more negative in $\delta^{18}\text{O}$ than summer precipitation (Fig. S7). Weak anti-correlation between modern air temperature and rainfall $\delta^{18}\text{O}$, but strong correlation ($r^2 = 0.80$) between mean monthly rainfall amount and mean rainfall $\delta^{18}\text{O}$ (Fig. S7) suggests that the amount effect is largely responsible for the negative winter precipitation $\delta^{18}\text{O}$, as noted in other Mediterranean region palaeoclimate records (Bar-Matthews et al. 2003, Drysdale, 2009, Finné et al., 2014). The modern cave water $\delta^{18}\text{O}$ overlap the four most negative (December-March) OIPC, which overall suggests a mixed Atlantic and Mediterranean moisture source, with recharge predominantly during autumn and winter. The two spring waters sampled >2 km from the cave have more negative $\delta^{18}\text{O}$ than the cave waters, caused by orographic effects of recharge at higher altitude; they do however, help define the local meteoric line and its identical slope to the EMMWL (Fig. S6).

Limnon cave water samples have a mean $\delta^{18}\text{O}$ of $-7.50 \pm 0.12\text{‰}$ VSMOW. Excluding the slightly enriched drip 1 sample of January 2007 (-7.1‰), the average is $-7.52 \pm 0.06\text{‰}$ VSMOW ($n=17$). Considerable isotopic homogeneity is therefore evident, suggesting effective groundwater mixing in the epikarst.

The extent to which oxygen isotopic equilibrium is maintained during precipitation of speleothem calcite from parent seepage water has typically been evaluated using the equilibrium fractionation equation of Kim and O'Neil (1997) derived from laboratory precipitation experiments. The calculated temperatures can then be compared with measured cave temperatures. However, extensive investigation of empirical speleothem and cave water oxygen isotope data (Tremaine et al., 2011) suggests that natural carbonate-water isotopic systems may not be well-reproduced by laboratory experiments. The best-fit "cave calcite" line through a plot of the available global speleothem-water $\delta^{18}\text{O}$ data is described by the equation:

578 $1000\ln\alpha = 16.1 (10^3T^{-1}) -24.6$ (1)

579 suggesting that water-calcite equilibrium fractionation factors are likely higher in natural
580 cave systems than in laboratory experiments.

581 Air temperatures at the sample points varied between 12.5- 13.3 °C (RH 89-90%) in
582 winter and 14.2 °C to 14.5 °C (RH of 96%) in summer (Iliopouou-Georgudaki and
583 Economidou, 1991), and a spot reading during sampling in November 2006 was 14.5 °C.
584 Winter pool water temperatures in this part of the cave are between 12-13 °C (Iliopouou-
585 Georgudaki and Economidou, 1991). The mean annual temperature at Tripoli, 50 km SE of
586 the cave is 14.1 °C (1961-1990; altitude 650 masl) and mean annual temperature at Kalavryta
587 (10 kms NNW, 731 masl) is 13.6°C (Pope et al., 2017), the equivalent temperature at the cave
588 site (850 masl) being ~12.8°C. Cave temperatures are thus within ±1.8 °C of the local annual
589 average temperature. Under wet winter – dry summer Mediterranean conditions calcite
590 precipitation may occur predominantly in winter and therefore calculated cave temperatures
591 of around 12°C are expected. Calculated temperatures using the Tremaine et al. (2011)
592 equation are shown in (Table 1) which also shows the Kim and O’Neil (1997) temperatures
593 for comparison. The modern calcite calculated temperatures fall within the expected range
594 (12.6°C and 13.1°C) excepting the deposit on the metal walkway which appears anomalous
595 (see Supplementary Information). These data show that most modern Limnon Cave
596 speleothem calcite is forming in near-equilibrium with its winter drip waters. We assume
597 these conditions largely held during the early Holocene, and in support of this the
598 petrographic fabrics in KTR-2 also suggest low degassing efficiency (see below).

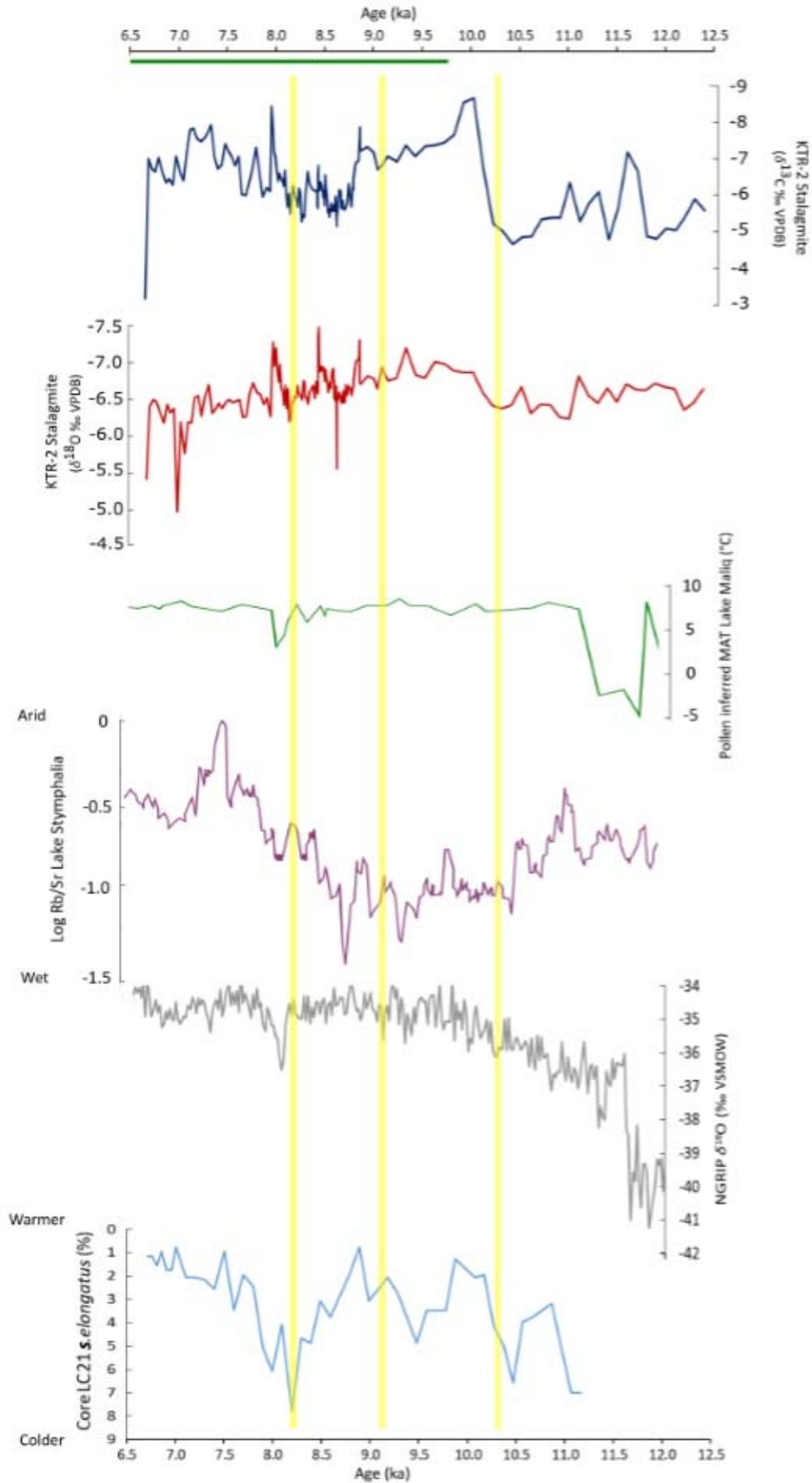
599 *6.2. KTR 2 record*

600 The columnar open (Co) fabrics seen in most of KTR-2 typically form under constant
601 and relatively high drip rate (0.1- 0.3 ml/min; Frisia et al., 2000; Boch et al. 2011) and in a
602 thicker water film than columnar compact calcites. Under these conditions, degassing is less
603 efficient (Kendall and Broughton 1978; Boch et al. 2011) which discourages complete
604 coalescence of crystallites. Co calcites typically form in dripwater with Mg/Ca ratios <0.3
605 and pH from 7.4 to 8.0 (Boch et al., 2011), the resulting high HCO₃/CO₃ ratios promoting
606 vertical linear extension. Columnar microcrystalline (Cm) fabrics form under more variable
607 drip rates (30 ml to <0.1 ml/min; Frisia and Borsato 2010) but most importantly with clear
608 input of impurities and organic colloids when compared to Co conditions (Frisia 2015); the
609 highly irregular crystal boundaries, typical of Cm fabrics, form where foreign particles induce

610 crystal defects (Frisia et al. 2000). In Alpine settings, typically with mixed conifer and
611 deciduous forest cover, combination of low dripwater supersaturation, low degassing and
612 increased flushing of colloidal particles appears to occur in autumn (Frisia et al., 2005) and
613 suggests that C_m is indicative of seasonal temperature and rainfall (increase in autumn)
614 contrast. Seasonal change in cave ventilation may also be indicated with less efficient
615 exchange between cave and atmospheric air occurring when inflow of soil-derived colloidal
616 particles is greater (Frisia 2015).

617 The association of C_m fabrics with irregular lateral crystal growth toward the
618 stalagmite flank in KTR-2, may indicate growth in very thin water films (and thus low drip
619 rates at these times), precluding substantial vertical extension.

620 Much of the Holocene growth of KTR-2 and its subsequent cessation is coincident
621 with the timing of S1 in the Eastern Mediterranean (see above) with conditions at this time
622 broadly wetter and warmer than the rest of the Holocene. Wetter conditions than present are
623 borne out by KTR-2 $\delta^{18}O$, which are typically up to 0.5‰ more negative than modern
624 speleothem calcite values. KTR-2 $\delta^{13}C$ values are nearly all less negative (typically by 1.5 –
625 2.0‰) than the mean modern speleothem calcite value of -8.3‰ (Table 1), and this is
626 particularly marked in the earliest part of the record until 10.3 ka (Fig. 12). These ‘high’ $\delta^{13}C$
627 values suggest less input of isotopically negative soil-carbon relative to today, particularly in
628 the period before 10.3 ka. Cool conditions evident in the Adriatic from 11.0 ka to 10.0 ka
629 (Rohling et al. 1997), and as late as ~9.6 ka in the Northern Aegean (Gogou et al. 2007; Fig.
630 12) may have limited soil development, particularly if accompanied by summer aridity (see
631 e.g. Heymann et al., 2013). Petrographic fabrics between 11.2 ka and 9.4 ka alternated
632 between C_o and C_m , suggesting short periods of constant drip water supply (C_o) giving way
633 to periods of more variable drip rate (C_m). KTR-2 Holocene extensions rates were mostly at
634 their lowest during this period, ~1.1 cm ka⁻¹, until 8.9 ka (Fig. 4) . There is no clear evidence
635 of the PBO climatic anomaly (cold and dry) between 11.4-11.2 ka excepting the possibility
636 that it could have contributed to a ‘high’ in $\delta^{13}C$ (Fig. 12) at this time.



638 Fig. 12. Low resolution KTR-2 stable isotope data ($\delta^{13}\text{C}$ blue, $\delta^{18}\text{O}$ red) related to regional
639 and global palaeoclimate events. Horizontal green bar indicates duration of S1 in the Aegean.
640 Vertical yellow bars help correlate events, bar width representing minimum error envelope of
641 ± 60 years (from U-series dates). Onset of a warm and wet (climate optimum) conditions in
642 KTR-2 ~ 10.3 ka (vertical yellow bar), marked by rapidly declining $\delta^{13}\text{C}$, coincident with a
643 cold phase (%cold water cyanobacterium *S. elongatus*; Marino et al. 2009; light blue record)
644 in otherwise warming SST trend in Aegean. Wetter conditions in Lake Stymphalia (Heymann
645 et al., 2013; purple dataset) also start at this time. ~ 9.3 ka (vertical yellow bar), and between
646 8.8-8.2 ka $\delta^{13}\text{C}$ indicates periods of dryness and cooler temperatures. Both KTR-2 isotopes
647 are relatively high between 8.1 and 8.2 ka (see Figs 7a and 13 for high resolution data),
648 coincident with peak abundance (cool) of *S. elongatus* (Marino et al. 2009) and within error
649 of the 8.2 ka cold event in NGRIP (grey record) (Andersen et al. 2004) and Lake Maliq
650 (green record, Bordon et al. 2009). Warmer and wetter conditions in KTR-2 re-established
651 after 8.2 ka.

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655 At ~ 10.3 ka there is a sharp 3‰ decrease in $\delta^{13}\text{C}$ (to -7.8 ‰; Fig. 12), heralding a
656 period of ~ 1000 years when $\delta^{13}\text{C}$ remained below -7.0 ‰ and indicating increased input of
657 isotopically negative soil-carbon, not dissimilar to present day conditions. During this same
658 1000 years $\delta^{18}\text{O}$ is < -6.7 ‰, the most sustained period of isotopically light compositions in
659 the Holocene record and indicative of increased winter rainfall. These ‘warm and wet’
660 indicators coincide with the start of S1 (Grant et al., 2016) and the onset of a Holocene
661 Climate Optimum. In this interval Cm fabrics underlie a detritus-rich layer at 96 mm, ~ 9.4 ka
662 (Fig. 5d) suggesting at least one period (~ 200 years based on layer width) when infiltration
663 was capable of transporting soil-derived colloids and particles through conduits. The $\delta^{13}\text{C}$
664 record suggests optimum conditions were interrupted briefly between 9.2 and 9.1 ka, and
665 decisively at ~ 8.8 ka in KTR-2 (Fig. 12) when values increased markedly, both perturbations
666 indicating drier conditions.

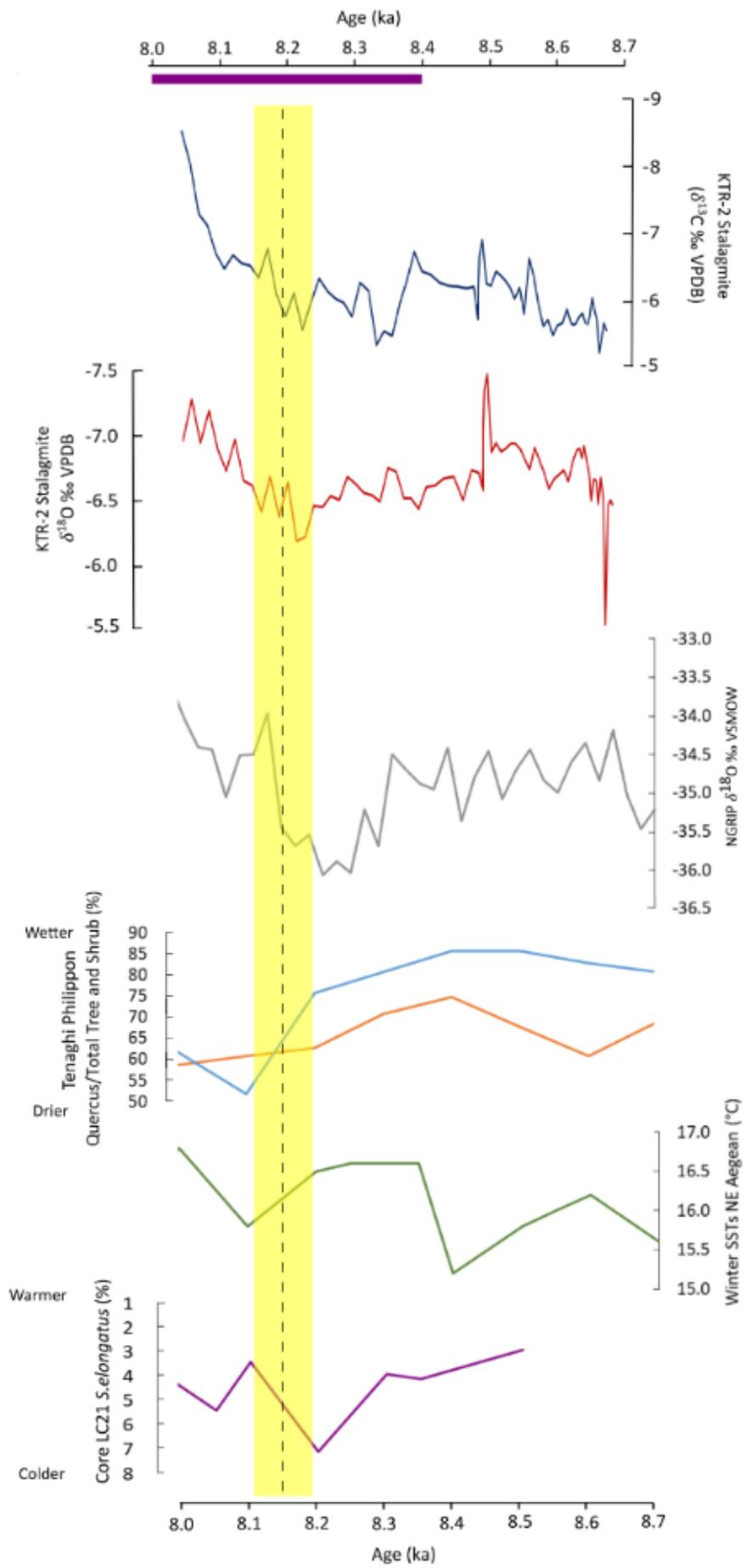
667

668 The early KTR-2 ‘optimum’ is broadly coincident with a number of regional
669 terrestrial and marine palaeoclimate indicators. The largest and most rapid increase in early
670 Holocene Aegean sea surface temperatures occurred between 10.0 and 9.0 ka (Triantaphyllou
671 et al., 2016), combined with pulsed input of terrestrial organics (Gogou et al. 2007) and
672 lowering of surface salinity, caused by increased fluvial discharge (Kotthoff et al. 2008b).
673 Onset of Aegean and Ionian sapropel formation occurred ~ 9.8 ka (Gogou et al., 2007;
674 Kotthoff et al., 2008b; Geraga et al., 2008) following the period of most negative $\delta^{13}\text{C}$ values

675 in KTR-2. At Tenaghi Philippon and Nisi Fen (N. Greece) terrestrial pollen data indicates
676 increased winter precipitation and stable winter temperatures between 10.4 ka and 9.5 ka
677 (Kottholf et al 2008a), the younger age within error of the ~9.4 ka detritus-rich layer in KTR-
678 2. The ~0.5‰ increase in $\delta^{13}\text{C}$ between 9.2 and 9.1 ka in KTR-2 is within error of the 9.2 ka
679 climatic anomaly (Fleitmann et al., 2008) the shift to less negative values consistent with
680 drier conditions.

681 Between ~8.8 and 8.2 ka, $\delta^{18}\text{O}$ values are typically around -6.6 to -6.5 ‰ (Fig. 12)
682 while $\delta^{13}\text{C}$ increase to ~ -6.6 ‰, values broadly similar to those before 10.3 ka, suggesting a
683 return to decreased winter rainfall and re-established dryness. These timings correspond to
684 regional climatic deterioration (aridity) that began around 8.8 ka (Rohling and Pälike 2005;
685 Marino et al. 2009) culminating in the northern hemisphere '8.2 ka event' of cooler and drier
686 conditions centred between 8.2 and 8.1 ka (Alley et al. 1997). However, a ~200 year negative
687 excursion in both isotopes ~8.5 ka in KTR-2 is a clear exception in this trend, the possible
688 significance of which is discussed later.

689 High-resolution $\delta^{18}\text{O}$ (micro-milled profile) between 8.6 ka to 8.4 ka decrease to a
690 minimum of -7.5 ‰ between 8.5 and 8.4 ka (Figs 7a and 13) accompanied by negative $\delta^{13}\text{C}$
691 and peaks in Sr, Ba, Na and P content (Fig. S4): stalagmite extension rates were also at their
692 highest, ~ 8.4 cm ka^{-1} sometime between 8.9 ka and 8.5 ka (Fig. 4). The combined
693 information suggest significant rainfall infiltration ($\delta^{18}\text{O}$) that mobilised soil-based lithogenic
694 colloids and soil organic matter (trace element and $\delta^{13}\text{C}$ response). However, from 8.4 ka,
695 $\delta^{18}\text{O}$ progressively increases to a maximum of -6.2 ‰ at ~8.2 ka (Figs 7a and 12), while $\delta^{13}\text{C}$
696 shows two more low negative (> -6.0 ‰) excursions ~8.3 ka and ~8.2 ka. These isotopic
697 trends are accompanied by increasing Mg/Ca (but with a reversal that matches the timing of
698 the ~8.2 ka low negative $\delta^{13}\text{C}$) and a peak in Mg/Sr ratio ~ 8.2 ka (Fig. 8). The combined
699 data are indicative of increasing dryness ($\delta^{18}\text{O}$) and increasing water residence time in the
700 epikarst ($\delta^{13}\text{C}$, Mg and to a lesser extent Sr), possibly accompanied by prior calcite
701 precipitation (PCP). The overall $\delta^{13}\text{C}$ response between 8.6 and 8.2 ka is clearly not one of
702 progressive change; instead, it shows marked 1‰ fluctuations around a value of -6.0 ‰,
703 changing to lighter compositions after 8.1 ka (Fig. 12). The $\delta^{13}\text{C}$ values while thus



705 Fig. 13. High-resolution stable isotope data (micro-milled samples) for the period around 8.2
706 ka in context with global and regional observations. Dryness indicators in KTR-2 (mainly in
707 $\delta^{13}\text{C}$ blue but also in $\delta^{18}\text{O}$ in red) combine between 8.2 and 8.1 ka indicated by the vertical
708 yellow bar. The bar width is a minimum error envelope (from U-series dates) of ± 60 years.
709 This dry period is within error of the latter part of the NGRIP 8.2 kyr cold event (dataset from
710 Andersen et al. 2004). Regional dryness at this time is indicated by rapid declines in Tenaghi
711 Philippon total tree and shrub pollen percentage from 8.4 ka (blue dataset; Peyron et al.,
712 2011) as are the proportion of evergreen oaks (orange dataset). Cooling NE Aegean (Marino
713 et al. 2009; green dataset) winter SSTs between 8.2 and 8.0 ka are indicated by higher
714 percentages of *S. elongatus* a cold water cyanobacterium (Rohling et al. 2002). The KTR-2
715 dry (and by inference cool) phase also corresponds broadly to the timing of S1 disruption in
716 the coastal Aegean Sea (purple bar; Kottholf et al. 2008b).

717

718

719 relatively high and broadly consistent with episodic dryness, are mostly not as high as values
720 attained in the period before 10.3 ka. This suggests aridity was not as marked as in the early
721 Holocene and is supported by development of more open columnar calcite (after 9.2 ka)
722 indicating more consistent recharge and steady infiltration. KTR-2 extension rates had slowed
723 $\sim 2.9 \text{ cm ka}^{-1}$ after 8.5 ka (Fig. 4). Cm fabrics are not present after 8.6 ka (Fig. 7a) suggesting
724 drip rate was reasonably constant from this time onward. These observations may indicate
725 that above-cave vegetation was sustained by episodic convective summer precipitation,
726 coinciding with the growing season and expressed by the short-lived negative $\delta^{13}\text{C}$
727 excursions (discussed further below).

728

729 The developing dryness recorded in KTR-2 between ~ 8.5 and 8.2 ka is consistent
730 with growing evidence that the '8.2 ka event' is superimposed on a climatic deterioration
731 trend between 8.8 ka to 7.8 ka (Rohling and Pälike 2005). The KTR-2 dryness is also
732 consistent with lower resolution chronologies for the onset of water level reduction in nearby
733 Lake Stymphalia (Fig. 1; Heymann et al. 2013) and in the southern Balkans and Macedonia
734 (lakes Maliq and Dojran; Bordon et al. 2009; Francke et al. 2013). The combined 'peak
735 dryness' indicators in KTR-2, ~ 8.2 ka, are all within error of the timing of minimum tree
736 pollen percentages at Tenaghi Philippon in N Greece (Peyron et al. 2011) and within error of
737 lake level low-stand ~ 8.2 ka at Stymphalia (Fig. 1; Heymann et al. 2013). In marine records
738 an increase in Ionian Sea surface salinity occurs around ~ 8.0 ka (Emeis et al. 2000) and
739 disruption of Aegean S1 closest to the Greek coastline occurs between 8.4 and 8.0 ka
740 (Kottholf et al. 2008b).

741 After 8.1 ka both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ become progressively more negative (Fig. 13) and
742 Mg/Ca and Mg/Sr ratios decrease (Fig. 8), all consistent with renewed increase in
743 precipitation for 200 years as ‘optimum conditions’ re-established until ~ 8.0 ka when
744 increasing $\delta^{18}\text{O}$ marks a phase of aridity preceding cessation of speleothem growth.
745 Petrography supports this final phase of aridity, a more compact columnar fabric with
746 increasing crystal coalescence indicating a slowing drip rate and more effective degassing
747 (Kendall and Broughton 1978). However, this aridity is not recorded consistently in $\delta^{13}\text{C}$
748 values, which fluctuate between -6 and -8‰ to the end of the record (excepting the terminal
749 value). This noted, a dark layer visible in hand specimen at ~ 7.8 ka coincides with lateral
750 crystal growth fabrics and a peak in $\delta^{13}\text{C}$. The crystal fabrics may indicate growth in a thin
751 water film, but not in this case accompanied by either a marked hiatus or Cm fabrics; drip
752 rate may have been slow (consistent with high $\delta^{13}\text{C}$) but reasonably constant. The $\delta^{13}\text{C}$
753 response suggests there was mostly enough rainfall to support some vegetation during this
754 period, perhaps because of effective summer rainfall, or if temperatures were cool (solar
755 insolation being in decline at this time), because of reduced effective evapotranspiration. By
756 6.7 ka both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ show large positive excursions indicating more intense aridity and
757 the following cessation of stalagmite growth probably indicates complete dryness in the
758 epikarst.

759 Developing aridity ~ 8.0 ka agrees with the timing of the end of Climate Optimum
760 conditions in Tenaghi Phillipon (Peyron et al., 2011). The increased aridity, marked by
761 increased $\delta^{13}\text{C}$ at 7.2 ka, corresponds with the end of S1 deposition (Fig. 13) at ~ 7.0 to 7.1 ka
762 in the N Aegean, Ionian and Adriatic Seas (Kottholf et al. 2008b; Emeis et al. 2000, Geraga
763 et al. 2008; Rohling et al. 1997). Drying at this time is also manifest in reduced precipitation
764 at the Alkyas Lagoon, Zakynthos (Avramidis et al. 2013) and in Lake Accesa, Italy (Peyron
765 et al. 2011). This timing also broadly matches the end of the S1 event as recorded in
766 speleothems from Corchia Cave, Italy (Zanchetta et al. 2007), and Soreq Cave in Israel (Bar
767 Matthews et al. 1999).

768 769 *6.3. Alluvial fan record*

770 The gravel-dominated parts of the sedimentary sequence formed under ‘normal’ local
771 postglacial Mediterranean climate (high winter runoff/summer drought) probably with
772 extensive deciduous oak woodland as indicated by palynological data from the offshore
773 Corinthian Gulf (Collier et al. 2000).

774 The Fe-rich nature of the palaeosols (Fig. 11) reflects the ultrabasic substrate upon
775 which they have developed, rich in Fe-bearing olivine and orthopyroxene. These provide high
776 pH soil microenvironments conducive to oxidative weathering, clay diagenesis and Fe³⁺
777 accumulation, probably from ferrihydrite precursors (see Schwertmann et al. 2004). The
778 similarity of silt layer spacings and erosion features above the most prominent 30 cm thick
779 palaeosol at both locations allow correlation between the sea cliff and quarry sections (Fig. 10).
780 If correctly correlated, the different ages determined for the same palaeosol require
781 explanation. We suggest the palaeosol developed slowly, episodically and possibly
782 diachronously over a long period of time (the ages allow ~9.5 to 8.3 ka) with millimetric
783 differences in sample depth below top, having a profound effect on age. This suggests that the
784 younger age in the quarry section (8.3-8.4 ka) marks the final phase, and culmination of
785 pedogenesis.

786 The basal palaeosol in the sea cliff section, with an age ~9.3 ka, is separated from the
787 overlying palaeosol by 'normal' alluviation, and the bracketing ages (despite possible inversion
788 of <400 years) suggest almost instantaneous flood deposition of this unit. Overall the ages are
789 too young to register the PBO but they broadly confirm that periodic pedogenesis was ongoing
790 by ~9.5 ka following ~700 years of wet conditions as recorded in KTR2 (Fig. 12). The 9.3 ka
791 palaeosol age corresponds with the brief drying episode between 9.2 and 9.1 ka in the KTR-2
792 record (see above), within error of the ~9.2 ka climatic anomaly (Fleitmann et al., 2008).

793 The KTR-2 record suggests that drying had re-established by 8.8 ka, leaving only a few
794 hundred years for 'normal' alluviation following the 9.2 ka climatic anomaly. The 30 cm thick,
795 light red palaeosol in the quarry section (Fig. 11) fixes culmination of this (probably prolonged)
796 pedogenesis at 8.3 to 8.4 ka, within error of the 8.2 ka climatic event. The KTR-2 low negative
797 $\delta^{13}\text{C}$ values indicate cool and dry conditions at this time. Palaeosol development records
798 abandonment of this part of the active fan with resulting pedogenesis.

799 Fine-grained alluvial fan sedimentation had resumed ~7.8-7.9 ka in the seacliff section
800 (Figs 9 and 10), coincident with indications of wetter climate (from $\delta^{13}\text{C}$ in KTR-2) particularly
801 ~8.1-8.0 ka, but also episodically between 7.8 and 7.2 ka, the latter overlapping wetter climate
802 indications ~7.5 ka in nearby lake Stymphalia (Heymann et al., 2103). Resumption of
803 alluviation in the quarry section occurred later, the 8.3 to 8.4 ka age palaeosol not overwhelmed
804 by 'normal' streamflow alluvium for a further ~3000 years, when aggradation allowed spillover

805 onto the reactivated fan surface. A total of around 6 m accumulated over an unconstrained time
806 to the present-day inactive fan surface (Fig. 10).

807 Lower down the fan, the youngest immature upper palaeosol seen in the sea cliff
808 developed around 5.0 ka, in response to a channel/lobe switch that subsequently reversed and
809 deposited the youngest prism of sediment on the easternmost flanks of the fan. This last
810 depositional event was foreclosed as marine erosion began the slow retreat of the fan's eastern
811 coastal cliff line promoting channel incision and abandonment of many of the lower fan lobes.
812 The palaeosol age of 5.0 ka is too young for comparison with the KTR-2 record but consistent
813 with aridity indicators ~5.0 ka in the Lake Stymphalia record (Heymann et al., 2013). In central
814 Italy and the Levant, Zanchetta et al. (2014) detect a speleothem isotopic excursion argued to
815 reflect relatively drier winters including a short sub-centennial period around 5.2 ka. In
816 Lebanon, Cheng et al. (2015) detect strong Bond event aridity at 5.1 ka.

817

818 **7. 8.2 ka regional climate effects**

819 While the KTR-2 record contains evidence of cool conditions and developing if episodic
820 dryness between 8.8 and 8.2 ka, wetter conditions between 8.5 and 8.4 ka are evident in both
821 the isotopic and elemental proxies. Indications of more humid summers between 8.7 and 8.5
822 ka in nearby Lake Stymphalia (Heymann et al., 2013) might implicate a change in the timing
823 of recharge, from winter to summer, around 8.5 ka. The combined information indicates that
824 developing dryness approaching 8.2 ka was not as marked as it had been in the early Holocene,
825 an interpretation corroborated by the moderate KTR-2 extension rates ($\sim 2.9 \text{ cm ka}^{-1}$; Fig. 4)
826 approaching 8.2 ka, relative to the slower early Holocene rates ($\sim 1.1 \text{ cm ka}^{-1}$; Fig. 4). Moreover,
827 between 8.5 and 8.4 ka, decrease in $\delta^{13}\text{C}$ could be indicating that above-cave vegetation growth
828 was reinvigorated, perhaps by episodic convective summer precipitation coinciding with the
829 growing season.

830 The stalagmite data is thus consistent with Aegean hinterland vegetation records that
831 indicate a fundamental change in hydrological conditions from wet to drier winters.
832 Specifically, deciduous tree pollen percentage in Aegean marine cores declines from 8.4 ka
833 with sharp reduction at 8.2 ka, especially noticeable in the reduced proportion of evergreen
834 oaks (Fig. 13) that are sensitive to winter drought (Kottoff et al. 2008a, 2008b; Pross et al.
835 2009). Analogy with present day Mediterranean climate dynamics suggests this was caused by
836 blocking of Atlantic fronts that intrude westward and trigger internal winter Mediterranean

837 cyclogenesis (Meteorological Office 1962; Trigo et al. 2000, 2002). However, microfaunal and
838 palynological data from both terrestrial northern Greece (Peyron et al. 2011) and the SE Aegean
839 (Triantaphyllou et al. 2009) also indicate a parallel increase in summer precipitation. This may
840 implicate an 8.2 ka-driven (cold N Atlantic) intensification of the Siberian high pressure,
841 blocking Atlantic fronts, a weakened summer monsoon (Wang et al., 2005; Cheng et al., 2009)
842 and reduced subsidence over the eastern Mediterranean promoting vigorous summer
843 cyclogenesis (cf. Trigo et al., 2002). Whatever the precise mechanism the precipitation regime
844 of the eastern Mediterranean changed its pattern from winter frontal to summer convective over
845 less than 1000 years.

846 If palaeosol development on Fan F was randomly distributed in time and space, then it is
847 simply coincidence that our palaeosol MRT ages align broadly with two episodes of known
848 climatic aridity at 9.2 ka and 8.2 ka. Alternatively, and perhaps more likely, the development
849 of palaeosols was driven by climatic events. If this is accepted, there are two scenarios that
850 may account for palaeosol development:

- 851 1. The whole of the eastern lower fan essentially dried up, the cessation of sedimentation
852 allowing palaeosol development.
- 853 2. A change in precipitation regime from winter frontal to summer convective promoted
854 ‘flashier’ summer rainfall regimes, perhaps with exceptional floods that caused rapid
855 fan-channel incision. The incision isolated large areas of surface fan and the onset of
856 prolonged pedogenesis there. Reduced rains from diminished winter cyclogenesis may
857 have promoted winter drought, slowing cool season pedogenesis and reducing
858 evergreen *Quercus* canopies further, enhancing summer runoff and sediment erosion.

859 Given that the speleothem record suggests episodic dryness, and a number of proxies
860 suggest episodic convective summer rainfall in the period between 8.8 and 8.1 ka, we prefer
861 scenario 2 as a likely driver for palaeosol development at least for the 8.2 ka event. Allocyclic
862 channel incision on the lower fan is likely to have been caused mainly by flash flooding before
863 the sea level highstand (~7 ka), whereas after the highstand, sea cliffing of the fan toe would
864 have also contributed to incision. The main weakness of this interpretation is that so far,
865 exposures in the lower fan have not been extensive enough to reveal the incised channel
866 network required in this scenario.

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868

869 **8. Conclusions**

870

871 We have outlined a case study from southern Greece, that for the first time, links regional
872 Holocene palaeoclimate change from a montane speleothem record with the sedimentary
873 response of a small, range front, semi-arid alluvial fan using dated sedimentary records.

874

- 875 1. The stalagmite palaeoclimate record from stable isotope and trace element data is
876 consistent with other regional proxies, and records the presence of two (9.2 ka and 8.2
877 ka) of the three early Holocene cool and dry climatic anomalies.
- 878 2. While developing, if episodic, dryness is clearly evident in the stalagmite record
879 between 8.8 and 8.2 ka, the markedly cool and dry conditions predicted for the N
880 Hemisphere at 8.2 ka, are not strongly developed, in common with a number of other
881 stalagmite climate records from the wider Mediterranean regions.
- 882 3. Palaeosols on alluvial Fan F have calibrated radiocarbon ages that align broadly with
883 two episodes of documented climatic aridity at 9.2 ka and 8.2 ka. We interpret palaeosol
884 development as a non-random, allocyclic response, driven by climatic events that are
885 recorded in the stalagmite climatic proxies.
- 886 4. For the 8.2 ka event, temporary development of ‘flashier’ summer rainfall regime
887 causing fan-channel incision is a plausible mechanism for allocyclic control on
888 palaeosol development. Our attribution of the causes of incision remains speculative in
889 the absence of better exposures; however, our approach outlines how radiocarbon
890 chronology for alluvial fan palaeosols can be used for centennial-timeframe
891 interpretation of alluvial fan response to climatic drivers.

892

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