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1 **Eastern Mediterranean hydroclimate over the late glacial and Holocene, reconstructed**
2 **from the sediments of Nar lake, central Turkey, using stable isotopes and carbonate**
3 **mineralogy**

4
5 Jonathan R. Dean^{a,b,c*}, Matthew D. Jones^{b,c}, Melanie J. Leng^{a,c}, Stephen R. Noble^a, Sarah E.
6 Metcalfe^{b,c}, Hilary J. Sloane^a, Diana Sahy^a, Warren J. Eastwood^d and C. Neil Roberts^e

7
8 *^aNERC Isotope Geosciences Facilities, British Geological Survey, Nottingham NG12 5GG*
9 *UK*

10 *^bSchool of Geography, University of Nottingham NG7 2RD UK*

11 *^cCentre for Environmental Geochemistry, University of Nottingham NG7 2RD UK*

12 *^dSchool of Geography, Earth and Environmental Sciences, University of Birmingham B15*
13 *2TT UK*

14 *^eSchool of Geography, Earth and Environmental Sciences, University of Plymouth PL4 8AA*
15 *UK*

16

17 *Corresponding author, jonathan.dean@bgs.ac.uk +44 (0)7944 747013

18 **Abstract**

19

20 There is a lack of high-resolution records of hydroclimate variability in the Eastern
21 Mediterranean from the late glacial and early Holocene. More knowledge of the speed of
22 climate shifts and the degree to which they were synchronous with changes in the North
23 Atlantic or elsewhere is required to understand better the controls on Eastern Mediterranean
24 climate. Using endogenic carbonate from a sediment sequence from Nar Gölü, a maar lake in
25 central Turkey, dated by varve counting and uranium-thorium methods, we present high-
26 resolution (~25 years) oxygen ($\delta^{18}\text{O}$) and carbon isotope records, supported by carbonate
27 mineralogy data, spanning the late glacial and Holocene. $\delta^{18}\text{O}_{\text{carbonate}}$ at Nar Gölü has been
28 shown previously to be a strong proxy for regional water balance. After a dry period (i.e.
29 evaporation far exceeding precipitation) in the Younger Dryas, the data show a transition into
30 the relatively wetter early Holocene. In the early Holocene there are two drier periods that
31 appear to peak at ~9.3ka and ~8.2ka, coincident with cooling ‘events’ seen in North Atlantic
32 records. After this, and as seen in other records from the Eastern Mediterranean, there is a
33 millennial-scale drying trend through the Mid Holocene Transition. The relatively dry late
34 Holocene is punctuated by centennial-scale drought intervals, at the times of 4.2ka ‘event’
35 and Late Bronze Age societal ‘collapse’. Overall, we show that central Turkey is drier when
36 the North Atlantic is cooler, throughout this record and at multiple timescales, thought to be
37 due to a weakening of the westerly storm track resulting from reduced cyclogenesis in the
38 North Atlantic. However, some features, such as the Mid Holocene Transition and the fact
39 the early Holocene dry episodes at Nar Gölü are of a longer duration than the more discrete
40 ‘events’ seen in North Atlantic records, imply there are additional controls on Eastern
41 Mediterranean hydroclimate.

42 **Highlights**

- 43 ➤ Sub-centennial resolution late glacial and Holocene isotope record from Turkey
- 44 ➤ Rapid transition from a dry late glacial into a wet early Holocene
- 45 ➤ Drier anomalies apparently at times of 9.3ka and 8.2ka events but last longer at Nar
- 46 ➤ Droughts at times of 4.2ka event and Late Bronze Age societal ‘collapse’
- 47 ➤ Strong teleconnection with North Atlantic, but additional other drivers

48

49 **Keywords**

50 Oxygen and carbon isotopes; Eastern Mediterranean; lake sediment; Holocene; late glacial;
51 Mid Holocene Transition; 9.3ka event; 8.2ka event; 4.2ka event; Late Bronze Age

52 1 Introduction

53

54 Water in the Eastern Mediterranean is a key and politically sensitive resource (Issar and
55 Adar, 2010) with rain-fed agriculture impossible across much of the region and regional
56 climate models suggesting conditions will become even drier through this century (Kitoh et
57 al., 2008). An improved understanding of hydroclimate over long timescales, >100 years, can
58 help identify the potential drivers of climate in the region under different boundary
59 conditions, assisting in the long-term sustainable management of water resources. This link
60 between people and their hydro-environment has been important for millennia, potentially
61 influencing the rise and fall of civilisations (e.g. Issar and Zohar, 2007; Rosen, 2007).

62 Current knowledge of regional palaeoclimatology suggests a dry, cool period in the
63 Eastern Mediterranean from ~12,900-11,700 years BP at the time of the Younger Dryas (Bar-
64 Matthews et al., 1999; Wick et al., 2003; Jones et al., 2007; Castañeda et al., 2010; Kotthoff
65 et al., 2011), followed by a wetter early Holocene marked by increased precipitation (Bar-
66 Matthews et al., 1999; Jones et al., 2007; Verheyden et al., 2008; Ocakoğlu et al., 2013).
67 There followed a significant shift in hydroclimate in the mid Holocene to a drier late
68 Holocene: the so-called Mid Holocene Transition (review of lake isotope data; Roberts et al.,
69 2008; 2011).

70 In the early Holocene in the North Atlantic region, two key centennial-scale cooling
71 episodes, at ~9,300 years BP and ~8,200 years BP, are well documented (e.g. von
72 Grafenstein et al., 1999; Rasmussen et al., 2006). These are expressed in many other northern
73 hemisphere regions as cool and/or dry periods, for example at the time of the 9.3ka 'event' in
74 China (Dykoski et al., 2005) and Oman (Fleitmann et al., 2003; 2007) and at the time of the
75 8.2ka 'event' in Turkey (Turner et al., 2008; Göktürk et al., 2011), Israel (Bar-Matthews et

76 al., 2003; Almogi-Labin et al., 2009), China (Dykoski et al., 2005) and Oman (Fleitmann et
77 al., 2003; 2007)

78 Late Holocene records (e.g. Jones et al., 2006) and present day climate (e.g. Cullen
79 and deMenocal, 2000; Harding et al., 2009) show clear links between the Eastern
80 Mediterranean and both the North Atlantic and the Indian Summer Monsoon. However, there
81 is a lack of records from the region with the required temporal resolution to allow for a
82 thorough investigation of centennial-scale climate change, and hence teleconnections to other
83 regions, beyond the late Holocene. To address this gap, we present a new, high-resolution
84 (~25 years) oxygen ($\delta^{18}\text{O}$) and carbon ($\delta^{13}\text{C}$) isotope record, with carbonate mineralogy data,
85 from Nar Gölü (Gölü = lake in Turkish) in central Turkey through the late glacial and
86 Holocene. This allows us to investigate the rapidity of climate shifts and centennial-scale
87 change throughout the whole Holocene and late glacial in a way that was not possible with
88 the previous, lower resolution records.

89

90 **2 Site description**

91

92 Nar Gölü (38°20'24''N, 34°27'23''E; 1363 m.a.s.l.; Figure 1) is a non-outlet, brackish maar
93 lake, 0.7 km² in area and >20 m deep, in the Cappadocia region of central Turkey (see Dean
94 et al., in press, for detailed catchment map). Modern $\delta^{18}\text{O}_{\text{lakewater}}$ values plot off the meteoric
95 water line (average for July surface samples from the centre of the lake 2001-2012 was –
96 1.3‰), suggesting high rates of evaporation (Jones et al., 2005; Dean et al., in press). The
97 crater geology is dominated by basalt and ignimbrite (Gevrek and Kazancı, 2000), reducing
98 the possibility for detrital carbonate contamination (cf. Leng et al., 2010). The climate of the
99 region is continental Mediterranean (Kutiel and Türkeş, 2005). Annual precipitation at Niğde,
100 45 km from Nar Gölü, averaged 339 mm between 1935 and 2010. July, August and

101 September receive only 6% of the total precipitation, while April and May are the wettest
102 months (27% of the total). The hottest months are July and August, when temperatures
103 average +23°C, while from December to February temperatures average +0.7°C
104 (meteorological data given in Dean et al., 2013).

105 Stable isotope (Jones et al., 2006; Dean et al., 2013), pollen (England et al., 2008) and
106 diatom (Woodbridge and Roberts, 2011) studies have previously been carried out on a 1,720
107 year core sequence (NAR01/02) from the lake. The distinctive carbonate-organic couplets in
108 the sediment have been shown to be annual (varves) and $\delta^{18}\text{O}_{\text{carbonate}}$ has been shown to
109 provide a means of reconstructing water balance (Jones et al., 2005; Dean et al., in press).

110

111 Figure 1

112

113 **3 Methods**

114

115 *3.1 Field work and chronology*

116

117 Three parallel cores from the deepest part of Nar Gölü were retrieved using a UWITEC
118 hammer-piston coring system, from the Laboratoire Environnement, Dynamiques et
119 Territoires de la Montagne (EDYTEM), Université Savoie Mont Blanc, in July 2010. The
120 three core sequences were matched visually at tie-points where turbidites or distinctive
121 sedimentological patterns could be clearly correlated, which led to the compilation of a 21.7
122 m master sequence (NAR10).

123 Where possible, chronologies for the sequence were established by varve counting.
124 Counts were made independently by two people and recounted until agreement (to within
125 five varve years) was reached. However, as the core sequence was not varved throughout,

126 additional age estimates were needed. Radiocarbon dating had previously been undertaken on
127 bulk organic and carbonate samples known to be ~500 years old (dated by varve counting),
128 but these gave apparent radiocarbon ages of 14,320 and 23,450 years BP respectively,
129 indicating a substantial old carbon reservoir linked to volcanic out-gassing (Jones, 2004).
130 Pollen and charcoal could not be extracted in sufficient quantities for radiocarbon dating of
131 these components, and there were no terrestrial macrofossils found in the cores.

132 Uranium-thorium (U-Th) dating was carried out on two aragonite and four calcite
133 dominated horizons (Dean, 2014). The use of the U-Th system in lacustrine environments is
134 predicated on the assumption that carbonates incorporate soluble U from the water column,
135 but little Th, as the latter is insoluble and found in lower quantities in lake water (Edwards et
136 al., 2003). However, Th can be incorporated into carbonates at the time of deposition from
137 detrital material (detrital Th) and from the water column (hydrogenous Th), meaning initial
138 [$^{230}\text{Th}/^{234}\text{U}$] is often not zero, and corrections to single-sample ages and multi-sample
139 isochrons are required for accurate age determination (e.g. Hasse-Schram et al., 2004).
140 Samples were processed using a total dissolution approach following Bischoff and Fitzpatrick
141 (1991) and Luo and Ku (1991). Our analytical protocol was based on Edwards et al. (1987)
142 with modifications as described in Douarin et al. (2013). This protocol was further
143 augmented here to ensure complete silicate component dissolution. After initial dissolution of
144 the carbonate fraction from each sample using HNO_3 and isolation from the detritus by
145 centrifugation, the remaining insoluble silicate component was dissolved in a mixture of
146 HNO_3 , HF and HClO_4 (all triple-distilled ultra-pure reagents) in an Evapoclean device to
147 ensure complete dissolution under clean lab conditions. The dissolved carbonate and detritus
148 fractions were then recombined and U and Th were separated and purified through anion
149 exchange chemistry. Samples were analysed on a Neptune Plus ICP-MS operating at c. 500-
150 600 V/ppm at an uptake rate of 50 $\mu\text{l}/\text{min}$ through an Aridus II desolvating nebuliser. Data

151 reduction was achieved using an in-house Excel spreadsheet and the Isoplot 3 add-in
152 (Ludwig, 2012) using the decay constants of Cheng et al. (2013). Isochrons were calculated
153 following Ludwig and Titterton (1994) and Ludwig (2012).

154

155 3.2 *Carbonate mineralogy*

156

157 Different mineralogies of calcium carbonate fractionate oxygen isotopes differently (Sharp,
158 2007) such that it is important to establish the mineralogy of the carbonates prior to
159 interpretation of carbonate isotope data (Leng and Marshall, 2004). Carbonate samples were
160 prepared in cavity mounts as described by Hardy and Tucker (1998). The scanning range
161 used on the X-ray Diffractometer (XRD; Siemens D500) was 5-65° 2θ and the scan rate was
162 2° 2θ per minute with a step size of 0.05. The TRACES program was used to identify which
163 minerals were present. Where two or more carbonate minerals were present, the proportions
164 of each were estimated by calculating the area under the XRD peaks and using
165 experimentally calibrated conversion curves (Hardy and Tucker, 1988). Scanning Electron
166 Microscopy (Philips XL30) and Energy-Dispersive X-ray Spectroscopy (EDS) (Oxford
167 Instruments INCA) were used to provide images of carbonate crystals and to calculate
168 elemental ratios.

169

170 3.3 *Oxygen and carbon isotope analysis of carbonates*

171

172 The XRD analyses showed that there are changes in carbonate mineralogy through the
173 NAR10 core sequence, with calcite, aragonite and dolomite present at different times. This
174 required samples for isotope analysis to be prepared in different ways.

175 Samples containing just calcite and/or aragonite were analysed for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$
176 using phosphoric acid, classic vacuum techniques, a standard +25.2°C reaction temperature
177 (McCrea, 1950; Craig, 1957) and a dual-inlet mass spectrometer. Data are given as ‰
178 deviations from VPDB and analytical reproducibility was 0.1‰ for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$. This
179 traditional reaction of carbonates at +25.2°C for 16 hours only partially reacts dolomite,
180 leading to an unpredictable kinetic fractionation. For samples composed of both dolomite and
181 aragonite/calcite, such as those in the Nar Gölü sequence, it is possible to extract CO_2 from
182 only the latter, without a significant reaction and accompanying fractionation of the dolomite,
183 by reducing the reaction time (Al-Aasm et al., 1990; Kyser et al., 2002; Baudrand et al.,
184 2012). We conducted experimental work to test whether a lower reaction temperature as well
185 as a shorter reaction time could be used to further limit the reaction of any dolomite within
186 the sample, and to test that this still led to the robust production of $\delta^{18}\text{O}$ data from calcite and
187 aragonite in these samples.

188 The experiments employed a calcite (KCM, marble) and dolomite standard (TDS,
189 hydrothermal dolomite) whose $\delta^{18}\text{O}$ values were -1.73‰ and -10.80‰ respectively. These
190 materials were chosen to mimic the calcite and dolomite at Nar Gölü. A range of mixtures of
191 the two standards were analysed and it was determined that for carbonate compositions with
192 $<\sim 20\%$ dolomite, reaction at +16°C for 1 hour was sufficient to release enough CO_2 from the
193 calcite for analysis (using triple the amount of carbonate sample (i.e. ~ 30 mg in a carbonate-
194 rich sample) as would normally be used). Detrimental fractionation artefacts were not
195 observed and potential contributions of CO_2 from the dolomite were insignificant (Figure
196 A.1) based on the known isotope compositions of the starting materials. For most calcite-
197 dolomite mixtures containing $>\sim 20\%$ dolomite, the offset from the accepted value of the
198 calcite component in the mixture was above expected analytical uncertainty and thus
199 interpreted to reflect a significant contribution from the dolomite (Figure A.1). Therefore,

200 samples containing >20% dolomite from the Nar Gölü core were not analysed for stable
201 isotopes, and samples containing <20% were analysed following the reduced time and
202 temperature reaction. The dolomite threshold adopted for Nar Gölü was very conservative to
203 ensure that the analytical data interpreted here were an accurate representation, with a high
204 degree of confidence, of the true calcite/aragonite isotope compositions. We acknowledge
205 that the analytical reproducibility of the data from these dolomite-bearing samples may be
206 greater than for those purely composed of calcite or aragonite, but given the large shifts seen
207 in the Nar Gölü record any increased analytical uncertainties are dwarfed by the observed
208 hydroclimate-induced changes to the isotopes. Our experiments are a promising first pass at
209 defining appropriate analytical thresholds for dolomite-bearing lake carbonates and there is
210 ample scope for further selective reaction method development to refine the analytical
211 approach for lake sediment samples with significantly >20% dolomite.

212

213 **4 Results**

214

215 *4.1 Lithology and chronology*

216

217 There are two main lithologies in the NAR10 sequence: laminated sediments (alternating
218 carbonate and organic layers, sometimes interrupted by turbidites) and non-laminated
219 sediments (Figure 2). It is clear from modern monitoring studies (e.g. Dean et al., in press)
220 and previous sediment core investigations (Jones et al., 2006; England et al., 2008;
221 Woodbridge and Roberts, 2011) that laminations of late Holocene age are annual in origin
222 (i.e. varved), and available evidence indicates that this is also true of most or all of the older
223 laminations. In one section of the NAR10 cores (798-1038 cm), the laminations are often
224 thicker (1-5 mm thick) suggesting the possibility of non-annual formation. However, because

225 U-Th dating is consistent with them being annual in origin (see below) we have assumed for
226 the chronology presented here that these are varves as well. From 1038 to 1141 cm the
227 laminations were often heavily deformed, making counting impossible. Mainly non-
228 laminated sections are found at core depth intervals 598-754 cm, 1965-2053 cm and from
229 2133 cm to the base of the core.

230

231 Figure 2

232

233 The chronology is summarised in Figure 2 and U-Th data are given in Table A.1 and
234 Figures A.2 and A.3.

235 Previous lake sediment studies (e.g. Hasse-Schramm et al., 2004) have shown that U-
236 Th analysis of carbonates that have had their age constrained by other means, for example
237 radiocarbon dating, provide critical constraints on the impact of potential hydrogenous Th.
238 When present and unaccounted for, the hydrogenous Th component will result in U-Th ages
239 that are older than the true age. Testing for the presence of hydrogenous Th was attempted by
240 analysing the U-Th isotope compositions of carbonate-rich layers at 0ka and 1ka, as
241 constrained by varve counting. The resulting U-Th data showed that the 0ka and 1ka
242 carbonate layers unfortunately have high detrital Th content in addition to insufficient
243 radiogenic ingrowth of ^{230}Th . No difference between the samples and typical continental
244 detritus compositions could be determined (Figure A.2). As a result, the magnitude of any
245 potential hydrogenous Th component at Nar Gölü remains indeterminate and therefore all of
246 the isochron ages presented here best represent maximum ages for the dated horizons. Figure
247 A.2 also shows that the turbidite samples from a layer ~6,500 years BP, considered here to be
248 the most representative samples of end member detritus available and composed of silt- to

249 clay-sized silicates and no carbonate, overlap well within uncertainty of the average
250 continental detritus composition.

251 U-Th analyses of carbonate-rich layers from various core depths were obtained to test
252 for locations most favourable for more detailed sampling. These layers proved to be calcite-
253 dominated, and although relatively carbonate-rich, they contained insufficiently high
254 carbonate/detritus ratios and U/Th ratios to result in isotope compositions departing
255 significantly from typical continental detritus compositions. Focus shifted to locating
256 aragonite-rich layers that might provide higher U/Th ratios and a better contrast with the
257 isotope composition of silicate detritus. Two horizons were thus identified, and the
258 stratigraphically lower horizon, from 1949 cm depth, sampled to provide 5 subsamples of
259 differing proportions of aragonite and detritus. U-Th analyses of these layers yielded a linear
260 array with an age of 11.82 ± 0.52 ka (Table A.1, Figure A.3). Scatter in the data, indicated by
261 a mean squared weighted deviation for the isochron regressed through the data points
262 (MSWD) = 10.5, could indicate heterogeneity in the detrital component U and Th isotope
263 composition. Following this analysis, the stratigraphically higher sample at 1021 cm was
264 analysed. As relatively high clay content was noted, additional physical separation steps were
265 taken to isolate the aragonite and coarse detritus components from the clay component, based
266 on in-house experience gained with other 'dirty' carbonates (Sahy et al., 2014). Subsamples
267 were sonicated in ultrapure water, and after a settling time of 1 hr the finest fraction
268 remaining in suspension was isolated and discarded. This physical separation step resulted in
269 a more favourable carbonate/detritus ratio composition, without risking disturbance of the U-
270 Th systematics (cf. Bischoff and Fitzpatrick, 1991). An isochron age for this sample is 4.41
271 $+0.16-0.17$ ka (MSWD=1.7).

272 To aid comparison with previously published sequences, all varve counts and U-Th
273 ages have been converted to years BP (i.e. before 1950) when plotted against age. It is

274 possible to use varve counting to provide a chronology for the sediments from the top of the
275 sequence (AD2010) through to 2,557 years BP (598 cm). There is a gap in the varved
276 sequence 598-754 cm but the $4.41 \pm 0.16-0.17$ ka U-Th age at 1021 cm ties the varved
277 section of core between 754-1038 cm to an absolute chronology. Therefore, we were able to
278 count up and down from this U-Th age to establish the chronology for this section. The
279 chronology for the section 598-754 cm could then be determined by linear interpolation,
280 assuming a fixed deposition rate for these largely homogenous sediments.

281 The 'floating' varved section of core 1161-1965 cm is tied by a U-Th age of 11.82
282 ± 0.52 ka at 1949 cm. Below 1965 cm, there is a non-varved section of sediment, before a
283 final floating varve sequence 2053-2133 cm. Again, we have assumed a linear deposition rate
284 for the non-varved section 1965-2053 cm. We were not able to obtain a U-Th age constraint
285 for the basal section of the core sequence, but the top of the basal varved section has been
286 assigned an age of 12,900 years BP based on correlation with NGRIP (Rasmussen et al.,
287 2006; Vinther et al., 2006). No data below 2133 cm are plotted against age.

288 The resulting age-depth model (Figure 2) indicates an overall sediment accumulation
289 rate of 1.5 mm/yr^{-1} between $\sim 12,000$ and $2,000$ years BP (~ 2000 -500 cm depth), with an
290 increase in mean sedimentation rate to $\sim 2.5 \text{ mm/yr}^{-1}$ in the last two millennia. Calculated
291 sedimentation rates are lower during non-laminated core sections between 598 and 754 cm
292 (dated 2,557-3,710 years BP) and 1965 to 2053 cm (dated 11,859-12,840 years BP), and are
293 higher during the interval of thick laminations, 798-1038 cm (dated to 3,987-4,383 years BP).
294 We estimated the sediments from 1141-1161 cm were lost during coring, but the sediments
295 from before 1141 cm and after 1161 cm are radically different, which suggests there may also
296 be a hiatus at this point in the NAR10 sequence. There are also deformed sediments 1038-
297 1141 cm. Because of these problems, we have not included data 1038-1141 cm in the age
298 model in this paper.

299 To check that the NAR01/02 and NAR10 sequences were overlapped at the correct
300 tie-points and recorded the same isotope signature, $\delta^{18}\text{O}_{\text{carbonate}}$ data from the NAR10 record
301 were compared to NAR01/02 data (Jones et al., 2006). The $\delta^{18}\text{O}_{\text{carbonate}}$ values from the
302 matched stratigraphic points are very similar (mostly within analytical error), suggesting the
303 $\delta^{18}\text{O}_{\text{carbonate}}$ values from the two core sequences are analogous, with age differences of <5
304 years at 1,400 years BP (Figure A.4). Counting of laminations from multiple replicate cores
305 shows no evidence that deposition of turbidite layers led to varve removal, and hence to any
306 under-estimate of true age.

307

308 4.2 Carbonate isotope and mineralogy data

309

310 Despite the changes in mineralogy, no correction has been made for the difference in the
311 mineral-water fractionation factors of calcite and aragonite here because the difference is
312 small ($\delta^{18}\text{O}$ of aragonite is $\sim 0.7\text{‰}$ more positive than $\delta^{18}\text{O}$ of calcite formed in the same
313 $\delta^{18}\text{O}_{\text{lakewater}}$ and temperatures; Grossman and Ku, 1986; Kim et al., 2007) compared to the
314 size of the shifts seen in this record. Additionally, calcite crystals from the NAR10 sequence
315 were analysed by EDS and the average Ca/Mg ratio was 18.2 mol%, so Nar Gölü calcite is of
316 a high-magnesium type (Gierlowski-Kordesch, 2010), and the offset in $\delta^{18}\text{O}$ between
317 aragonite and high-magnesium calcite formed under the same conditions is even smaller than
318 0.7‰ (Tarutani et al., 1969; Jimenez-Lopez et al., 2004).

319 The period from the base of the core sequence up to ~ 2057 cm (Figure 3) has variable
320 but generally lower $\delta^{18}\text{O}_{\text{carbonate}}$ (average -1.9‰) and $\delta^{13}\text{C}_{\text{carbonate}}$ (average $+13.7\text{‰}$) than the
321 period from ~ 2053 - 1965 cm (average -0.6‰ and $+18.8\text{‰}$ respectively). Calcite/aragonite
322 and varved sediments are found in the former whereas aragonite/dolomite and non-varved
323 sediments are found in the latter. There is then a rapid shift to lower $\delta^{18}\text{O}_{\text{carbonate}}$ and

324 $\delta^{13}\text{C}_{\text{carbonate}}$ values and varved, calcite and aragonite sediments in the early Holocene (1957-
325 1312 cm; average -2.9‰ and $+13.4\text{‰}$). $\delta^{18}\text{O}_{\text{carbonate}}$ values are fairly stable until increases to
326 peaks (-1.0‰ , -1.2‰ and 0.0‰) centred on ~ 1600 , 1520 and 1450 cm respectively, all
327 associated with shifts from calcite to aragonite. After ~ 1300 cm ($\delta^{18}\text{O}_{\text{carbonate}} -3.7\text{‰}$), there is
328 a sustained rise in $\delta^{18}\text{O}_{\text{carbonate}}$. The rise ends ~ 800 cm ($+1.8\text{‰}$) but high values are
329 maintained, albeit with centennial-scale periods of lower $\delta^{18}\text{O}_{\text{carbonate}}$, until ~ 350 cm.
330 $\delta^{13}\text{C}_{\text{carbonate}}$ values rise from $+12.5\text{‰}$ at ~ 1300 cm to $>+17\text{‰}$ ~ 800 cm. Dolomite is present
331 from ~ 490 - 1050 cm, with the periods between ~ 920 - 1000 , 600 - 680 and 540 - 580 cm having
332 $>20\%$ dolomite, precluding $\delta^{18}\text{O}_{\text{carbonate}}$ analysis for the reasons outlined in section 3.3. At
333 ~ 350 cm, there is a large transition to lower $\delta^{18}\text{O}_{\text{carbonate}}$ and $\delta^{13}\text{C}_{\text{carbonate}}$ and a shift from
334 aragonite to calcite.

335

336 Figure 3

337

338 **5 Discussion**

339

340 *5.1 Drivers of $\delta^{18}\text{O}_{\text{carbonate}}$ at Nar Gölü*

341

342 A comparison of meteorological records to $\delta^{18}\text{O}_{\text{lakewater}}$ and $\delta^{18}\text{O}_{\text{carbonate}}$ data since 1999 (Dean
343 et al., 2013; in press), and calibration with meteorological data (Jones et al., 2005), has shown
344 $\delta^{18}\text{O}_{\text{carbonate}}$ from Nar Gölü is a strong proxy for regional water balance (with lower
345 $\delta^{18}\text{O}_{\text{carbonate}}$ values when water balance was more positive, and vice versa). Several factors
346 support the contention that water balance was the driver of $\delta^{18}\text{O}_{\text{carbonate}}$ throughout the
347 NAR10 record.

348 Firstly, there is a strong co-variation between $\delta^{18}\text{O}_{\text{carbonate}}$ and $\delta^{13}\text{C}_{\text{carbonate}}$ ($r=0.84$,
349 $p<0.001$, $n=1502$, combining the NAR01/02 and NAR10 sequences; or $r=0.83$, $n=465$,
350 $p<0.001$ for just the NAR10 sequence) (Figures 3 and A.5). This can be taken to indicate
351 that the lake has been hydrologically closed throughout this period (Talbot, 1990; Li and Ku,
352 1997). Lakes with no surface outflow such as Nar Gölü tend to have $\delta^{18}\text{O}$ records driven by
353 changes in the evaporation:precipitation ratio (Leng and Marshall, 2004).

354 Secondly, varves are only preserved in lakes when water depth is sufficient to limit
355 turbidity caused by wind and when there is stratification leading to anoxic bottom waters and
356 consequent limited bioturbidity (Ojala et al., 2000; Ojala et al., 2012; Zolitschka et al., 2015).
357 A shift from varved to non-varved lake sediments is therefore likely to reflect a shift to lower
358 lake levels. The observation that varved sediments occur when $\delta^{18}\text{O}_{\text{carbonate}}$ is lowest, and non-
359 varved when $\delta^{18}\text{O}_{\text{carbonate}}$ is highest, supports the interpretation of $\delta^{18}\text{O}_{\text{carbonate}}$ as a water
360 balance proxy.

361 Thirdly, shifts from calcite to aragonite are believed to occur due to an increase in the
362 Mg/Ca ratio of lake water (Müller et al., 1972; Kelts and Hsu, 1978; Ito, 2001), which
363 favours the precipitation of aragonite over calcite (Berner, 1975; De Choudens-Sanchez and
364 Gonzalez, 2009). This shift has been observed in Nar Gölü over the last decade, as the lake
365 level has fallen and Mg/Ca ratios have increased (Dean et al., in press). Dolomite is found in
366 parts of the sequence, but unlike calcite and aragonite there is no evidence for dolomite
367 forming during our monitoring period (1997-present). Dolomite in lake sediments can
368 originate from the detrital inwash of old dolomite, from primary precipitation, or from
369 diagenetic precipitation in sediments (Armenteros, 2010; Leng et al., 2010). At Nar Gölü, the
370 former can be discounted as the crater geology is dominated by basalt and ignimbrite.
371 Primary dolomites are rare in lake sediments, however where they do occur they tend to have
372 rhombic crystals (Sabins, 1962) whilst those in Nar Gölü sediments are subhedral/anhedral

373 (Figure A.6). It is possible that dolomite formed authigenically within the sediments,
374 replacing calcite or aragonite during early diagenesis. An organogenic origin is plausible,
375 given the dolomite is calcium-rich (average Ca/Mg ratio of dolomite crystals is 2.3)
376 (Vasconcelos and McKenzie, 1997; Armenteros, 2010) and due to the porous nature of the
377 crystals (Figure A.6; Deng et al., 2010). Regardless of the actual mode of formation, it is
378 widely accepted that dolomite formation requires sufficient magnesium (Mazzullo, 2000), so
379 the appearance of dolomite in the sediments suggests magnesium was more highly
380 concentrated than at times when aragonite or calcite formed. Dolomite was also precipitated
381 at similar times in the late Holocene at another maar lake in the same region (Eski Acıgöl;
382 Roberts et al., 2001), implying a common origin linked to low lake levels and dry climatic
383 conditions. $\delta^{18}\text{O}_{\text{carbonate}}$ values are highest when dolomite is present, lower in aragonite zones
384 and lowest in calcite zones, again showing that $\delta^{18}\text{O}$ values follow evaporation trends.

385

386 5.2 *Hydroclimate reconstructions*

387

388 5.2.1 *The late glacial*

389

390 Full chronological control below 1965 cm is lacking. However, the low $\delta^{18}\text{O}_{\text{carbonate}}$ values
391 and varved sediments from 2053 cm to the bottom of the core sequence indicate a wetter
392 period probably at the time of the Bølling-Allerød and the higher $\delta^{18}\text{O}_{\text{carbonate}}$ values,
393 aragonite/dolomite and non-varved sediments 1965-2053 cm (Figure 3) indicate a dry period
394 at the time of the Younger Dryas. This would follow the pattern of late glacial hydroclimate
395 previously reconstructed in the region (e.g. Jones et al., 2007; Kotthoff et al., 2008; Wilson et
396 al., 2008).

397 The magnitude of change in $\delta^{18}\text{O}$ during the entire transition from the Younger Dryas-
398 aged dry period into the Holocene is 5.2‰ (Figure 4). Because of the chronological
399 uncertainty, it is not possible to calculate precisely how long the entire transition took, but
400 based on deposition rates of adjacent sections we can estimate it took <200 years. Over half
401 of the $\delta^{18}\text{O}_{\text{carbonate}}$ transition (2.9‰) occurs in just 9 varve years (Figure 4), although after this
402 shift there is a change back to higher $\delta^{18}\text{O}_{\text{carbonate}}$ values, in an excursion that lasts 27 varve
403 years, before a return to lower $\delta^{18}\text{O}_{\text{carbonate}}$. The nature of the transition recorded by
404 $\delta^{18}\text{O}_{\text{carbonate}}$ could indicate a non-linear response of $\delta^{18}\text{O}_{\text{carbonate}}$ to changing climate and/or that
405 the climate transition itself was non-linear.

406

407 Figure 4

408

409 5.2.2 *General Holocene trends*

410

411 The Nar Gölü $\delta^{18}\text{O}_{\text{carbonate}}$ record is similar to other lake $\delta^{18}\text{O}$ records from the Eastern
412 Mediterranean (Figure 5) (Roberts et al., 2008; 2011). Specifically, there are low $\delta^{18}\text{O}_{\text{carbonate}}$
413 values at Nar Gölü in the early Holocene and a clear and sustained transition to higher values
414 (albeit interrupted by centennial-scale fluctuations) that starts ~7,600 years BP and ends
415 ~4,000 years BP (Figure 5). This period of high $\delta^{18}\text{O}_{\text{carbonate}}$ at Nar Gölü lasts until ~1,500
416 years BP. These timings are similar to other records from the region (Roberts et al., 2011).
417 There are shifts from calcite prior to ~6,500 years BP to aragonite and dolomite for most of
418 the later Holocene, from varved to at times non-varved sediments and from low to high
419 $\delta^{13}\text{C}_{\text{carbonate}}$. This supports the interpretation of Holocene $\delta^{18}\text{O}_{\text{carbonate}}$ in Eastern
420 Mediterranean lake carbonates as responding to water balance (Jones and Roberts, 2008),
421 rather than to changes in the $\delta^{18}\text{O}$ of the source of precipitation (Litt et al., 2012) or in the

422 seasonality of precipitation (Stevens et al., 2001; 2006), since changes in carbonate
423 mineralogy, lithology and $\delta^{13}\text{C}_{\text{carbonate}}$ can be influenced by changes in water balance but not
424 directly by the other two factors (Leng and Marshall, 2004).

425

426 Figure 5

427

428 *5.2.3 Centennial-scale 'events' in the early Holocene*

429

430 Two main periods of centennial-scale climate change in the early Holocene have been
431 identified from North Atlantic region palaeoclimate records: the so-called 9.3ka and 8.2ka
432 'events' (Rasmussen et al., 2006). Climate changes at the time of the 8.2ka 'event' have been
433 identified in some Eastern Mediterranean records (Bar-Matthews et al., 2003; Landmann and
434 Kempe, 2005; Turner et al., 2008), however a lack of high-resolution records means
435 investigation of other centennial-scale changes in the early Holocene has been limited. The
436 uncertainties on the U-Th date at 1949 cm mean it is not possible to investigate fully whether
437 the early Holocene events occurred synchronously in Nar Gölü and NGRIP. However, it is
438 possible to count through the varved sediments from the start of the Holocene to establish
439 whether there were any changes in Nar Gölü that occurred the same amount of time from the
440 start of the Holocene in central Turkey as equivalent changes after the onset of the Holocene
441 in Greenland. We define the durations of the events simply by eye, following the logic of
442 Daley et al. (2011) that statistical approaches may not be suitable when comparing such
443 diverse data sets.

444 There is a shift to increasing dryness in Nar Gölü ~2,340 varve years after the start of
445 the Holocene, very similar to the number of years after the start of the Holocene the 9.3ka
446 cooling trend starts in NGRIP (Figure 6). However, whereas the cooling in NGRIP and other

447 records from the North Atlantic region such as Ammersee (von Grafenstein et al., 1999) lasts
448 ~100 years, in Nar Gölü the excursion lasts ~300 years (Figure 6). Relative dryness at this
449 time also lasts longer in other records more remote from the North Atlantic, for example
450 Dongge in China, where the excursion lasts ~200 years (Dykoski et al., 2005).

451 There is a peak in $\delta^{18}\text{O}_{\text{carbonate}}$ in Nar Gölü starting ~3,400 years after the start of the
452 Holocene, around the time of the 8.2ka 'event' in NGRIP. However, this appears to be the
453 peak of a longer isotope trend encompassing ~300 years (Figure 6). There is also a switch
454 from calcite to aragonite sediments for ~400 years at the time of the highest isotope values.
455 The 8.2ka 'event' is seen across the Northern Hemisphere (Alley et al., 1997; Alley and
456 Ágústsdóttir, 2005; Morrill and Jacobsen, 2005). In NGRIP it is defined as lasting 160 years
457 (Thomas et al., 2007), and in other isotope records from the North Atlantic region ~150±30
458 years (Daley et al., 2011). However, other than in a few records (e.g. Heshang Cave in China;
459 Liu et al., 2013), away from the North Atlantic region the effects are often spread over a
460 longer time period (Rohling and Pälike, 2005; Wiersma and Renssen, 2006; Thomas et al.,
461 2007), with more abrupt changes at 8,200 years BP superimposed on longer term
462 cooling/drying trends (Rohling and Pälike, 2005). Intervals where the climate became drier
463 compared to average early Holocene conditions are seen in records from across tropical
464 Africa ~8,500-7,800 years BP (Gasse, 2000), the Black Sea coast of Turkey ~8,400-7,800
465 years BP (Göktürk et al., 2011) and for several hundred years in Qunf Cave in Oman
466 (Fleitmann et al., 2003; 2007) (Figure 6). A dry event in between those centred on ~9,300 and
467 8,200 years BP is seen in both Nar Gölü and Qunf, as are shifts to lower $\delta^{18}\text{O}_{\text{carbonate}}$ either
468 side of this dry event (Figure 6).

469

470 Figure 6

471

472 5.2.4 Centennial-scale 'events' in the mid to late Holocene

473

474 Over 2,000 years of data are either missing from the core sequence or are not shown on
475 Figure 5 because of uncertainties in the chronology (section 4.1). Therefore, it is not possible
476 to investigate whether a drought previously identified in the region ~5,300-5,000 years BP
477 (Bar-Matthews and Ayalon, 2011; Kuzucuoğlu et al., 2011) occurred at Nar Gölü.

478 It is possible, however, to investigate changes from 4,400 years BP to the present day.
479 Firstly, there was >20% dolomite content ~4,300-4,150 years BP, indicating a period of very
480 negative water balance (i.e. dry conditions). Following this, once dolomite levels fell below
481 20%, the highest $\delta^{18}\text{O}_{\text{carbonate}}$ values in the whole sequence occurred at ~3,800 years BP.
482 Previous studies have identified substantial drying ~4,200-3,900 years BP in the Eastern
483 Mediterranean (Cullen et al., 2000; Eastwood et al., 2007; Ulgen et al., 2012). This dry
484 interval has been called the 4.2ka 'event' and coincides with the decline of the Akkadian
485 Empire in northern Mesopotamia (Weiss, 1993; Cullen et al., 2000) and the Old Kingdom of
486 Egypt (Stanley et al., 2003).

487 There is a further positive $\delta^{18}\text{O}_{\text{carbonate}}$ excursion in Nar Gölü ~3,400 years BP,
488 coincident with a period of dryness identified in Eski Acıgöl and Van (Roberts et al., 2011)
489 (Figure 5). There are two periods of high (>20%) dolomite ~3,150-2,550 and ~2,500-2,300
490 years BP. These again suggest drier conditions than the millennial-average, and the presence
491 of non-varved sediments for the first of these periods could be taken to indicate an even drier
492 climate than in the interval ~4,300-4,150 years BP. The period ~3,150-2,550 years BP is
493 synchronous within dating uncertainty with a drought seen across the Eastern Mediterranean
494 (Roberts et al., 2001; Verheyden et al., 2008; Langgut et al., 2013; Neugebauer et al., 2015),
495 at the time of the so-called Late Bronze Age Collapse/Crisis (Kaniewski et al., 2013) when
496 civilisations such as the Hittites in central Turkey went into decline (Weiss, 1982; Akurgal,

497 2001). The period ~2,500-2,300 years BP could be coincident with some peaks seen in the
498 Eski Acıgöl $\delta^{18}\text{O}_{\text{carbonate}}$ record, but more high-resolution records from the region are required
499 to help establish if there was a widespread excursion to increasing dryness at this time.

500

501 *5.2.5 The large shift in $\delta^{18}\text{O}_{\text{carbonate}}$ in the 6th century AD*

502

503 There is a large shift to more negative $\delta^{18}\text{O}_{\text{carbonate}}$ in the 6th century AD: ~1,450-1,400 years
504 BP (Figures 5 and A.4). This period of low $\delta^{18}\text{O}_{\text{carbonate}}$ lasted until ~550 years BP, interrupted
505 by a temporary rise to higher values ~1,090 years BP. Other than the multi-millennial scale
506 Mid Holocene Transition, it is the largest $\delta^{18}\text{O}$ shift seen in the record, even more pronounced
507 than the late glacial to Holocene transition. Around the Eastern Mediterranean, a shift to
508 wetter conditions is inferred at this time in Soreq Cave (Orland et al., 2009), Lake Tecer
509 (Kuzucuoğlu et al., 2011), the Eastern Mediterranean Sea (Schilman et al., 2001) and the
510 Dead Sea (Neumann et al., 2007). However, only at Nar Gölü is the shift of such a high
511 magnitude. As well as the shift to lower $\delta^{18}\text{O}_{\text{carbonate}}$, indicative of a shift to more positive
512 water balance, the shift from aragonite to calcite precipitation, and diatom assemblage data
513 (Woodbridge and Roberts, 2011), also indicate a shift to wetter conditions.

514

515 *5.3 Drivers of Eastern Mediterranean hydroclimate*

516

517 The rapidity of the late glacial to Holocene transition at Nar Gölü is comparable to
518 that seen in temperature records from the North Atlantic region (e.g. von Grafenstein et al.,
519 1999; Rasmussen et al., 2006). This suggests a strong teleconnection between the North
520 Atlantic and Eastern Mediterranean. A rapid transition into the Holocene is also seen in
521 Moomi Cave in Socotra, Yemen (Shakun et al., 2007), and in Hulu and Kulishu caves in

522 China (Wang et al., 2001; Ma et al., 2012; Orland et al., in press) (see Figure 1 for locations).
523 The transition is more gradual in Dongge; differences between the Chinese speleothem
524 records could be related to differences in the relative influences of the Indian Monsoon and
525 the East Asian Monsoon, and westerlies, at different sites (Huang et al., 2015).

526 On the whole, central Turkey was drier when the North Atlantic was cooler: at the
527 time of the Younger Dryas, at ~9,300 years BP, ~8,200 years BP, ~4,200 years BP and
528 ~3,100 years BP. Slowdowns of North Atlantic thermohaline circulation due to glacial
529 outburst floods have been suggested as the causes of the Younger Dryas (Teller, 2012), 9.3ka
530 (Fleitmann et al., 2008; Yu et al., 2010) and 8.2ka (Thomas et al., 2007; Hoogakker et al.,
531 2011; Hoffman et al., 2012) cooling episodes. Although there is no clear climate signal in the
532 NGRIP ice core ~4,200 and 3,100 years BP, increases in ice-rafted debris in the North
533 Atlantic known as Bond events 3 and 2 (Bond et al., 1997) occur around these times. A
534 significant amount of the precipitation that falls in central Turkey has North Atlantic origins
535 (Harding et al., 2009; Türkeş et al., 2009) so a reduction in cyclogenesis at these cooler times
536 is likely to have reduced the frequency, and potentially changed the path, of storm tracks
537 from the Atlantic. A resulting reduction in Mediterranean cyclogenesis is also likely.
538 Together, this would have led to less precipitation in the Eastern Mediterranean (Bartov et al.,
539 2003; Prasad et al., 2004; Rowe et al., 2012).

540 The influence of the North Atlantic through ocean and/or atmospheric circulation
541 could explain peaks in dryness at Nar Gölü at these times, but does not explain why the
542 excursions ~9,300 and 8,200 years BP last longer in Nar Gölü and other records outside of
543 the North Atlantic region (Rohling and Pälike, 2005) than the cooling ‘events’ in the North
544 Atlantic. Whilst changes in the North Atlantic are seen as a key driver of Eastern
545 Mediterranean hydroclimate in the present and past, it has been demonstrated that other
546 teleconnections are also important, such as Indian Summer Monsoon dynamics (Jones et al.,

547 2006; Ziv et al., 2006) and the North Sea-Caspian Pattern Index (Kutiel and Türkeş, 2005).
548 Jones et al. (2006) used the 1,720-year NAR01/02 record to show changes in the North
549 Atlantic have more of an influence on winter conditions and changes in the Indian Summer
550 Monsoon have more of an influence on summer conditions at Nar Gölü. Rohling and Pälike
551 (2005) suggest the sharp 8.2ka event signal is seen more in winter-based proxies with the
552 broader ~8,500-8,000 years BP climate deterioration more evident in summer-weighted
553 proxies. There seem to be drier summers at Nar Gölü over the past 1,720 years when the
554 monsoon is more intense (Jones et al., 2006), hypothesised to be related to a strengthening of
555 the descending branch of the Hadley cell and increased northerly winds over the Eastern
556 Mediterranean at these times (Raicich et al., 2003; Tyrllis et al., 2013). However, ~8,200
557 years BP, we see a different relationship, with seemingly concomitant shifts at Nar Gölü and
558 Qunf to drier conditions, related, at least in the latter case, to a less intense monsoon
559 (Fleitmann et al., 2003). Reduced solar output has been proposed as the cause of the broad
560 underlying climate deterioration ~8,500-8,000 years BP (Neff et al., 2001; Gupta et al., 2005;
561 Rohling and Pälike, 2005). This could explain a weakening of the monsoon, but it remains
562 unclear how reduced energy in the climate system would lead to increased evaporation in the
563 summer at Nar Gölü. Reduced summer precipitation would give a similar signal, to more
564 positive isotope values, but many authors suggest that summers had low precipitation
565 throughout the early Holocene, with significant rainfall only falling in the winter (see
566 discussion below). Whatever the cause, changes in the North Atlantic ~8,200 years BP
567 additionally led to drier winters (less precipitation) and even higher $\delta^{18}\text{O}_{\text{carbonate}}$ at Nar Gölü,
568 accounting for the maxima of the peak at Nar Gölü at this time.

569 There is a large, multi-millennial scale shift in Holocene hydroclimate seen in Nar
570 Gölü and other records from the Eastern Mediterranean (e.g. Roberts et al., 2001; Eastwood
571 et al., 2007), Asia (e.g. Fleitmann et al., 2003; Dykoski et al., 2005) and Africa (e.g. Adkins

572 et al., 2006; Renssen et al., 2006): the Mid Holocene Transition. This suggests that there are
573 additional drivers of Eastern Mediterranean hydroclimate, in this case at millennial-scales.
574 The Mid Holocene Transition has been linked to a decline in Northern Hemisphere summer
575 insolation from the early Holocene maximum (deMenocal et al., 2000; Braconnot et al.,
576 2007; Fleitmann et al., 2007; Renssen et al., 2007). Increased precipitation in Saharan Africa
577 in the early Holocene was caused by a northward movement of the Inter Tropical
578 Convergence Zone and monsoon rains related to this peak in insolation (Schneider et al.,
579 2014), but the direct influence of the African Monsoon is not generally considered to have
580 reached the Eastern Mediterranean (Arz et al., 2003; Brayshaw et al., 2011a). Summer
581 drought persisted for several millennia into the Holocene in the Eastern Mediterranean
582 (Turner et al., 2010; Peyron et al., 2011; Vanniere et al., 2011). Rather, the wet early
583 Holocene in the Eastern Mediterranean appears to have been the result of increased
584 precipitation in other seasons, especially the winter (Brayshaw et al., 2011b), made possible
585 because of the increased residual heat left in the North Atlantic and the Mediterranean Sea as
586 a result of the higher summer insolation (Tzedakis, 2007), increasing cyclogenesis. Through
587 the Holocene, a decrease in annual insolation led to a weakening and poleward shift of the
588 storm track (Black et al., 2011), and to drier conditions in the Eastern Mediterranean.

589

590 **6 Conclusion**

591

592 Using $\delta^{18}\text{O}_{\text{carbonate}}$ and carbonate mineralogy data, on a core sequence dated by U-Th
593 and varve counting, it has been possible to provide a highly-resolved hydroclimatic
594 reconstruction for the Eastern Mediterranean. We show relatively dry conditions at the time
595 of the Younger Dryas, a wet early Holocene and the Mid Holocene Transition to drier
596 conditions, which peaked ~4,200-1,500 years BP. There are centennial-scale periods where

597 climate became drier than the millennial average ~9,300, 8,200, 4,200 and 3,100 years BP.
598 Other studies have previously suggested a link between Eastern Mediterranean hydroclimate
599 and changes in the North Atlantic during the last glacial (e.g. Bartov et al., 2003) and late
600 Holocene (e.g. Jones et al., 2006), but it is only with the high-resolution record presented here
601 that we can demonstrate this was the case in the early Holocene as well. We show that dry
602 climatic anomalies during the late glacial and throughout the Holocene in central Turkey
603 appear to occur at the same time as cold anomalies in the North Atlantic. This suggests a
604 teleconnection between the two regions, via changes in cyclogenesis and in the frequency and
605 path of storm tracks from the Atlantic. However, the Mid Holocene Transition and the longer
606 duration of the 9.3ka and 8.2ka anomalies at Nar Gölü indicate there are additional controls
607 on Eastern Mediterranean hydroclimate.

608

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610

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628

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1038 **Figure captions** (*all figures black and white only*)

1039

1040 **Figure 1** Location of Nar Gölü in central Turkey, and the lake, ice and cave sites from which
1041 key isotope records referred to in this study have been produced.

1042

1043 **Figure 2** Age-depth plot for the NAR01/02 and NAR10 master sequences. The NAR01/02
1044 chronology was constructed using varve counts (Jones et al., 2005), and the NAR10
1045 chronology using a mixture of varve counts and two U-Th dates. We use linear interpolation
1046 for the non-varved sections 598-754 and 1965-2053 cm, and this is signified on the plot by
1047 the dashed lines. Depths for the NAR01/02 sequence are approximate, as they were taken by
1048 varve year, not against depth. The sections where there were gaps due to coring are shown by
1049 the white boxes.

1050

1051 **Figure 3** $\delta^{18}\text{O}_{\text{carbonate}}$ and $\delta^{13}\text{C}_{\text{carbonate}}$ data plotted against depth, with the locations of varved
1052 and non-varved sediments and carbonate mineralogy also shown. Relatively few samples are
1053 a mixture of aragonite and calcite, such that >50% calcite is defined as calcite and >50%
1054 aragonite as aragonite. Where dolomite is present, samples are shown as containing >20%
1055 dolomite (those samples not run for isotopes) and <20% dolomite (those samples run using
1056 the selective reaction method). Depths for the NAR01/02 sequence are approximate, as they
1057 were taken by varve year, not against depth.

1058

1059 **Figure 4** $\delta^{18}\text{O}_{\text{carbonate}}$ record for the late glacial to Holocene transition at Nar Gölü, with the
1060 varved section up to the second youngest sample analysed at 3-year resolution. Shifts
1061 between aragonite and calcite are also shown.

1062

1063 **Figure 5** $\delta^{18}\text{O}_{\text{carbonate}}$ records from Nar Gölü and other sites in the Eastern Mediterranean
1064 arranged in increasing distance from Nar Gölü, with more positive values indicating drier
1065 conditions: Eski Acıgöl (Roberts et al., 2001), Gölhisar Gölü (Eastwood et al., 2007), Soreq
1066 Cave (Bar-Matthews et al., 1997; 1999; Orland et al., 2009; Bar-Matthews and Ayalon,
1067 2011), Lake Van (core 90-4, 20-year interpolated values) (Wick et al., 2003) and Lake
1068 Zeribar (Stevens et al., 2001). All records show a transition in the mid Holocene to more
1069 positive values. The chronology at Nar Gölü further back in time than the point marked with
1070 the asterisk (*) is very tentative and based on comparison to NGRIP; we are careful not to
1071 over-interpret this period in the discussion. The sections where there were gaps due to coring
1072 or where isotope data are not shown due to the issues with the chronology (section 4.1) are
1073 shown by the white boxes.

1074

1075 **Figure 6** Nar Gölü $\delta^{18}\text{O}$ data for the late glacial and early Holocene, compared to records
1076 from further east in Asia, and the North Atlantic: Qunf in Oman (Fleitmann et al., 2003;
1077 2007), Kulishu (Ma et al., 2012), Dongge (Dykoski et al., 2005), Heshang (Hu et al., 2008;
1078 Liu et al., 2013) and Hulu (H82) (Wang et al., 2001) caves in China, Ammersee in Germany
1079 (reversed scale) (von Grafenstein et al., 1999) and NGRIP (reversed scale) (Vinther et al.,
1080 2006; Rasmussen et al., 2006). While there are rapid transitions into the Holocene in Nar
1081 Gölü, some records from China, as well as North Atlantic records, the difference is greater
1082 during the early Holocene ‘events’, where many records further away from the North Atlantic
1083 have less discrete and longer anomalies than the shorter 9.3 and 8.2 ka events in the North
1084 Atlantic region.



NGRIP

Ammersee

Nar Gölü

Van

Soreq

Qunf

Moomi

Kulishu

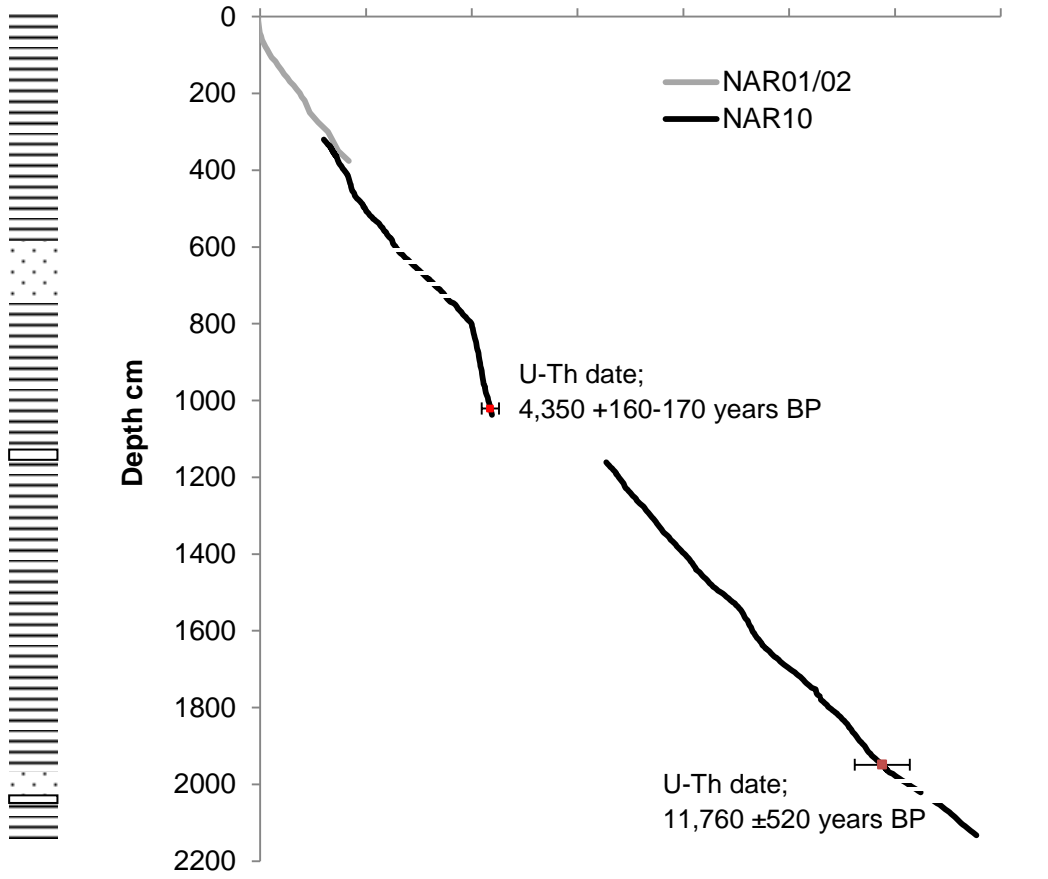
Heshang

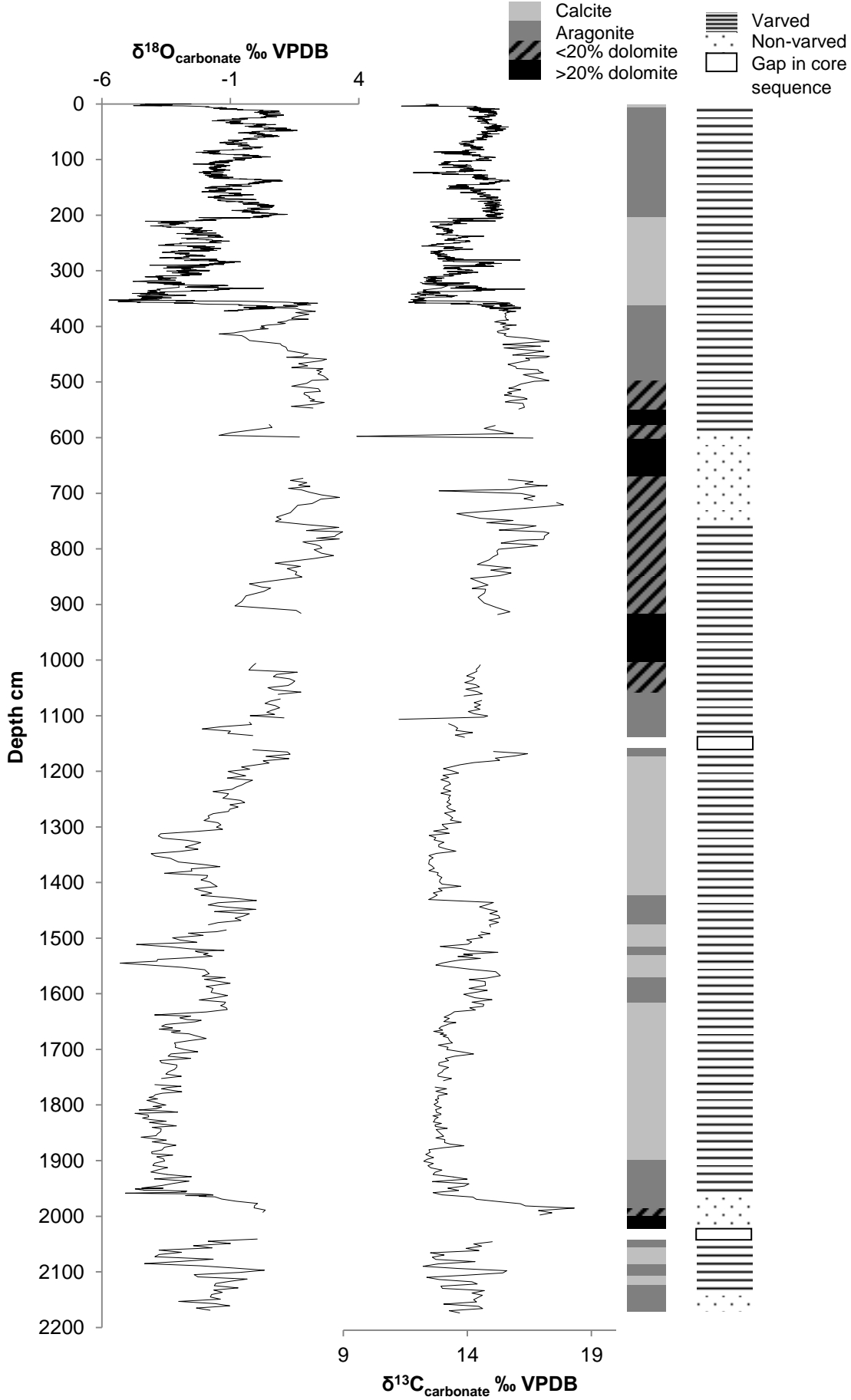
Hulu

Dongge

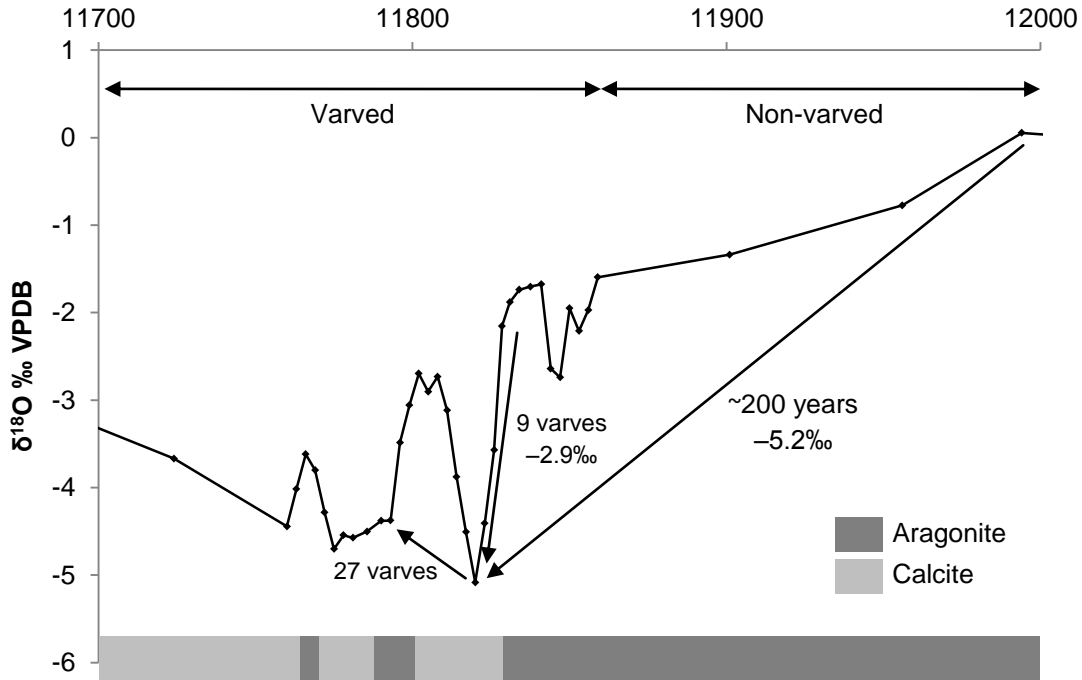
Varved
Non-varved
Gap in core sequence

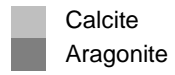
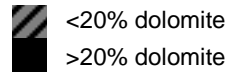
Years BP





Years BP





□ Gap in core sequence or data not plotted

Years BP

0 2000 4000 6000 8000 10000 12000

$\delta^{18}\text{O} \text{‰ Nar}$

$\delta^{18}\text{O} \text{‰ Golhisar}$

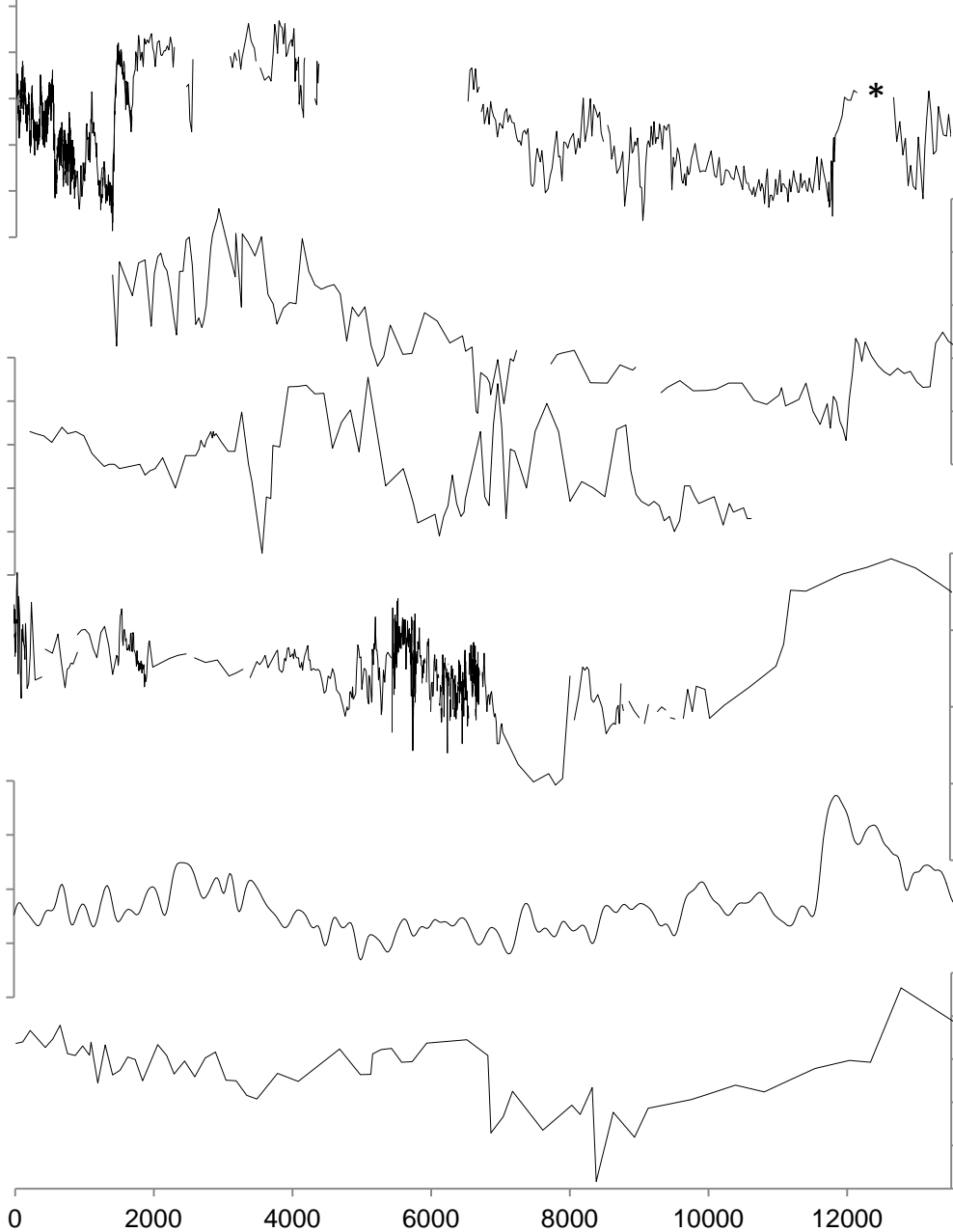
$\delta^{18}\text{O} \text{‰ Van}$

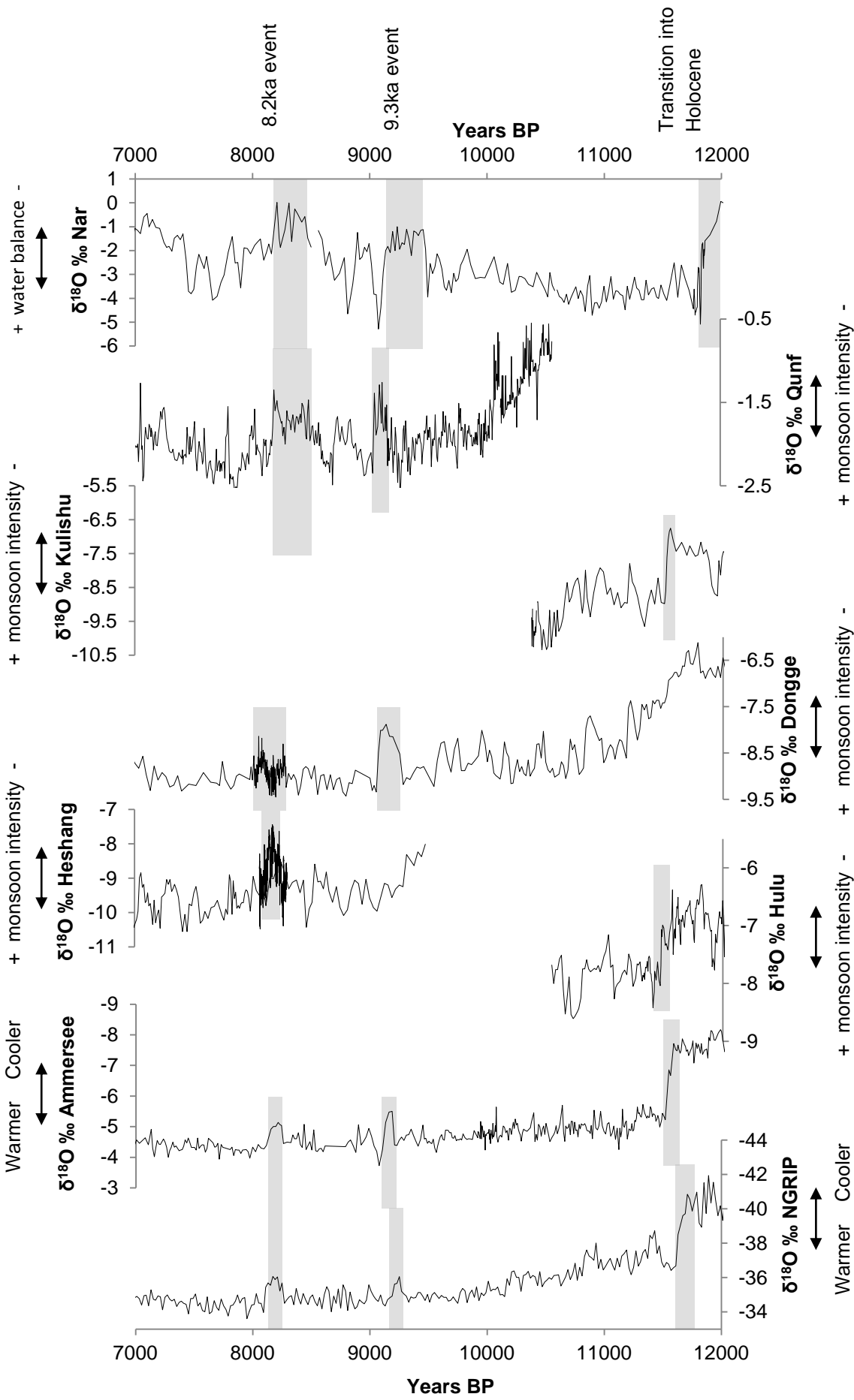
+ water balance
- water balance

$\delta^{18}\text{O} \text{‰ Eski}$

$\delta^{18}\text{O} \text{‰ Soreq}$

$\delta^{18}\text{O} \text{‰ Zeribar}$





1 ka (305 cm) A	0.6956	2.075	0.84	0.9802	0.20	0.8282	0.35	1.134	0.15
1 ka (305 cm) B	0.3764	1.114	0.84	0.9725	0.08	0.8140	0.35	1.121	0.14
1 ka (305 cm) C	0.4119	1.235	0.84	0.9852	0.09	0.8278	0.28	1.122	0.16
1 ka (305 cm) D	0.4003	1.237	0.83	1.015	0.08	0.8437	0.28	1.116	0.15
1 ka (305 cm) E	0.8744	2.748	0.85	1.033	0.12	0.8738	0.29	1.131	0.15
Turbidite A	2.066	10.92	0.85	1.737	0.60	1.473	0.66	0.9834	0.29
Turbidite B	2.393	11.37	0.84	1.562	0.66	1.308	0.71	0.9816	0.25
Turbidite C	2.443	10.78	0.89	1.450	0.27	1.296	0.37	0.9818	0.13
Turbidite D	1.407	7.134	0.85	1.666	0.56	1.424	0.62	0.9799	0.17

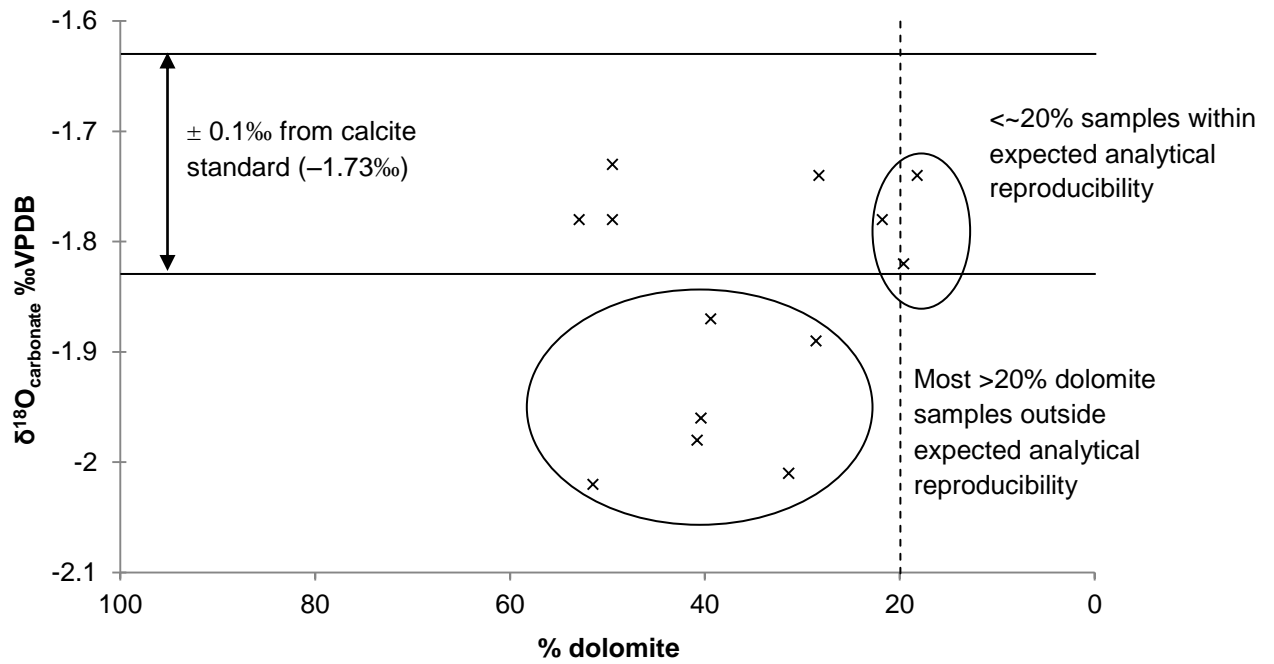


Figure A.1 $\delta^{18}\text{O}$ of samples reacted at 16°C for 1 hour containing mixtures of dolomite and calcite standards. Samples with $<20\%$ dolomite were all within expected analytical reproducibility.

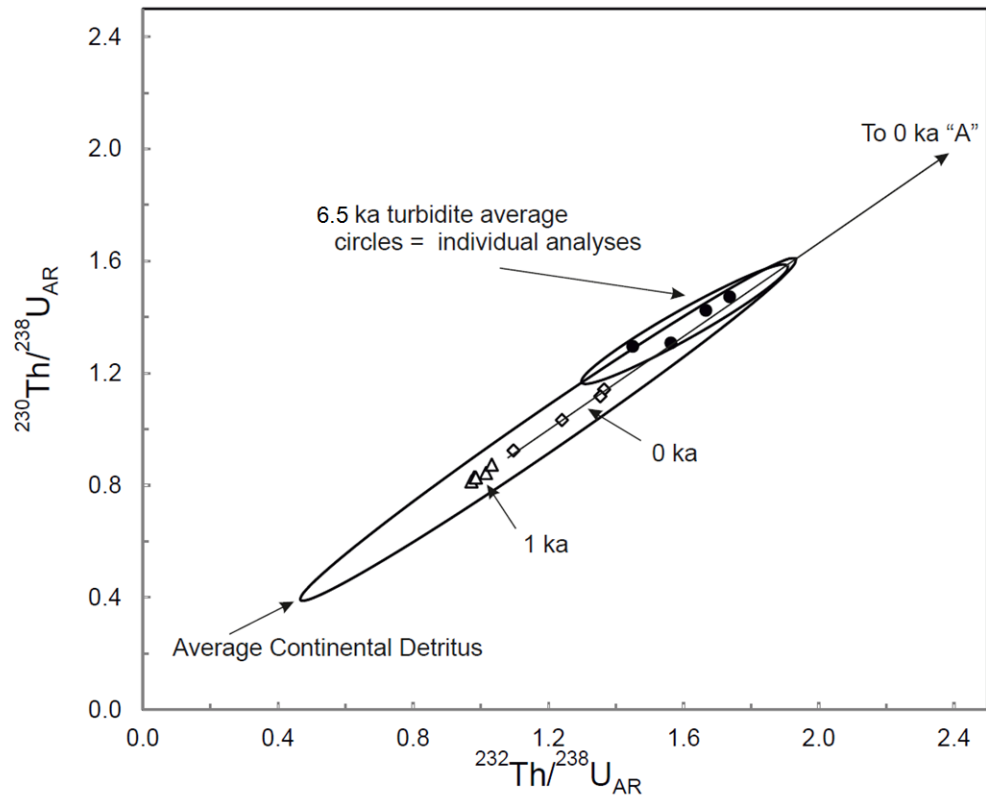


Figure A.2 $^{230}\text{Th}/^{238}\text{U}$ vs $^{232}\text{Th}/^{238}\text{U}$ plot of detritus and carbonates from the sediment core of known ages. Sub sample A from 0ka plots off the graph, indicated by the arrow marked “To 0ka A”, but still in a line with the other samples.

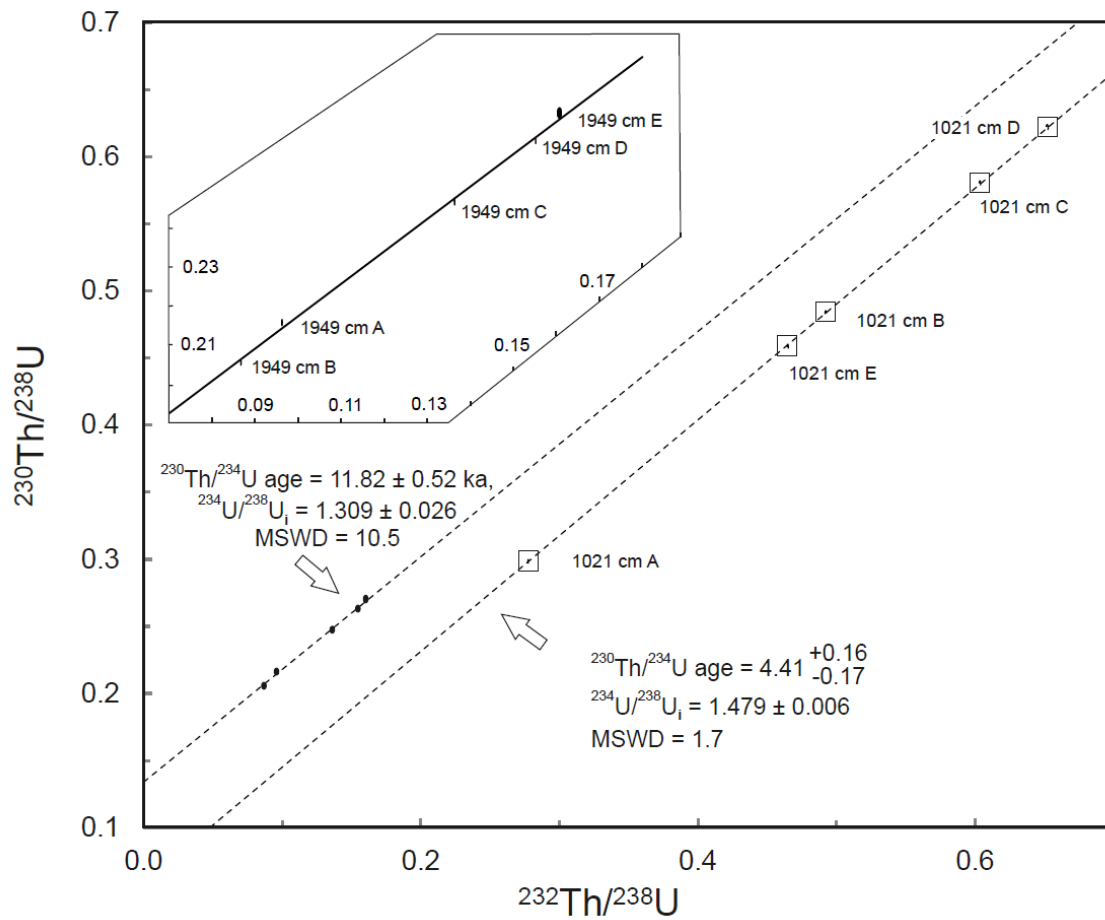


Figure A.3 2D projection of the 3D $^{230}\text{Th}/\text{U}$ isochron (Ludwig and Titterton, 1994), with 2 sigma uncertainty ellipses. The ellipses are small for the 1021 cm sample so are highlighted with boxes. A magnification of the isochron for 1949 cm is shown inset.

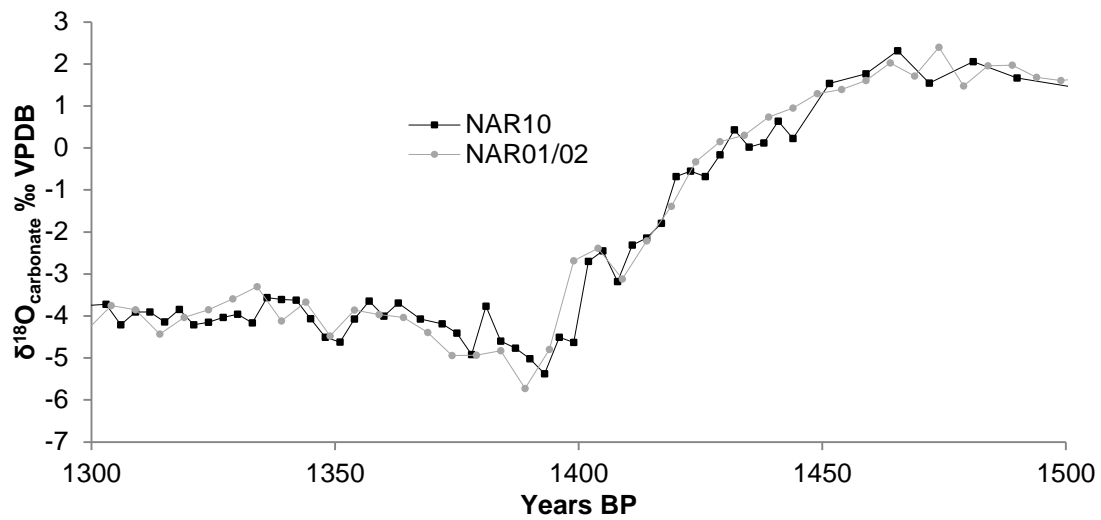


Figure A.4 Comparison of $\delta^{18}\text{O}_{\text{carbonate}}$ data from NAR10 and NAR01/02 cores through a major late Holocene climatic transition. There is very little apparent offset between the two records.

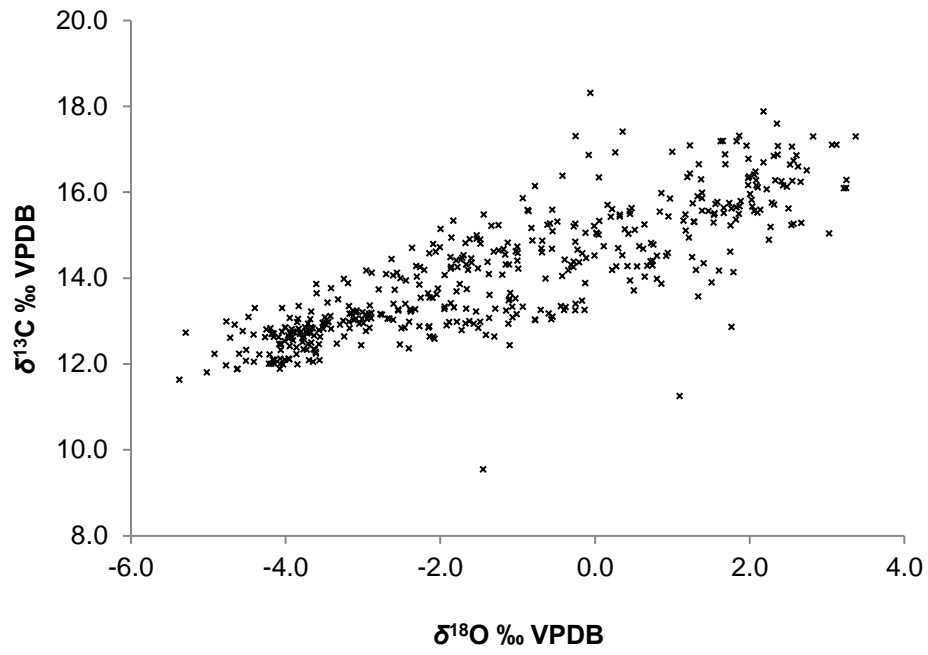


Figure A.5 $\delta^{13}\text{C}_{\text{carbonate}}$ data from the NAR10 core sequence plotted against $\delta^{18}\text{O}_{\text{carbonate}}$.

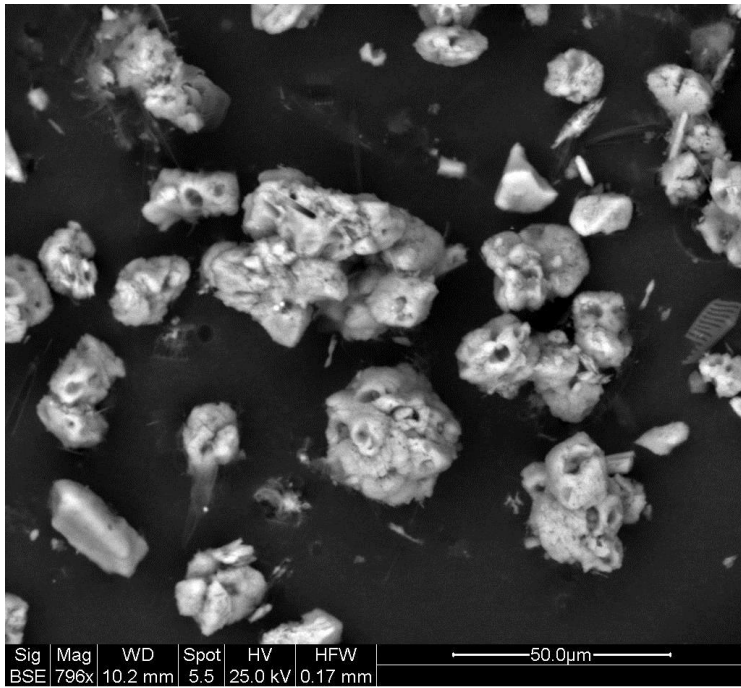


Figure A.6 Dolomite crystals viewed under SEM, showing non-rhombic shapes and microstructures, suggesting a diagenetic origin.