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1 **Seasonality of Holocene hydroclimate in the Eastern Mediterranean reconstructed**
2 **using the oxygen isotope composition of carbonates and diatoms from Lake Nar,**
3 **central Turkey**

4

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19

20

21 **Abstract**

22

23 A positive shift in the oxygen isotope composition ($\delta^{18}\text{O}$) of lake carbonates in the
24 Eastern Mediterranean from the early to late Holocene is usually interpreted as a change
25 to drier (reduced P/E) conditions. However, it has also been suggested that changes in
26 the seasonality of precipitation could explain these trends. Here, Holocene records of
27 $\delta^{18}\text{O}$ from both carbonates and diatom silica, from Lake Nar in central Turkey, provide
28 insights into palaeoseasonality. We show how $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ (the difference between
29 spring and summer reconstructed $\delta^{18}\text{O}_{\text{lakewater}}$) was minimal in the early Holocene and
30 for most of the last millennium, but was greater at other times. For example, between
31 ~4,100-1,600 years BP we suggest that increased $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ could have been the
32 result of relatively more spring/summer evaporation, amplified by a decline in lake
33 level. In terms of change in annual mean $\delta^{18}\text{O}$, isotope mass balance modelling shows
34 that this can be influenced by changes in seasonal P/E as well as inter-annual P/E, but
35 lake level falls inferred from other proxies confirm there was a mid Holocene transition
36 to drier climatic conditions in central Turkey.

37 **Keywords**

38 Oxygen isotopes; Eastern Mediterranean; lake sediment; Mid Holocene Transition;
39 palaeoseasonality; Turkey

40 1 Introduction

41

42 Understanding the detail of hydrological variability over multiple timescales is
43 important in regions such as the Eastern Mediterranean where water stress is increasing
44 (Issar and Adar, 2010) and where management of water supplies under a changing
45 climate is essential (e.g. Kelley et al., 2015). Water availability issues have potentially
46 been critical for societies in the region for millennia (e.g. Weiss et al., 1993) and an
47 understanding of both changes in mean state and seasonality are required (Rohling,
48 2016). Many studies from the region have shown a shift in the mid Holocene to higher
49 oxygen isotope ratios of lake carbonates ($\delta^{18}\text{O}_{\text{carbonate}}$) (Roberts et al., 2008). These are
50 usually interpreted as responding to changes in the balance between precipitation and
51 evaporation (P/E) (Jones and Roberts, 2008), thus showing a mid Holocene transition
52 from a wetter early Holocene, with relatively more precipitation, to a drier late
53 Holocene, where evaporation losses were relatively increased. However, the extent to
54 which there were shifts in the seasonality of precipitation in the Holocene, and the
55 degree to which these would have affected $\delta^{18}\text{O}_{\text{carbonate}}$, remains an unresolved issue in
56 Eastern Mediterranean Holocene palaeoclimatology. Stevens et al. (2001, 2006)
57 suggested that a change from winter- to spring-dominated precipitation was potentially
58 a driver of the increasing $\delta^{18}\text{O}_{\text{carbonate}}$ trend in the mid Holocene, based on analysis of
59 the sediments of Lakes Zeribar and Mirabad in Iran. Other authors, using pollen and

60 microcharcoal records, have also argued that there were shifts in the seasonality of
61 precipitation in the region through the Holocene (e.g. Djamali et al., 2010; Turner et al.,
62 2010; Peyron et al., 2011).

63

64 Seasonality change analysis requires proxies that are sensitive to different seasons. Dean
65 et al. (2013) showed that comparing $\delta^{18}\text{O}$ from endogenic carbonates and diatoms at
66 Nar Gölü (Gölü = lake in Turkish) in central Anatolia can provide insights into
67 seasonality as they formed/grew at different times of the year. Such records, combining
68 $\delta^{18}\text{O}$ from diatoms and carbonates in the same core, remain rare. Here, we present a
69 $\delta^{18}\text{O}_{\text{carbonate}}$ vs. $\delta^{18}\text{O}_{\text{diatom}}$ record from Nar Gölü for the entire Holocene, developing a
70 rigorous methodology for diatom isotope data correction, coupled with an isotope mass
71 balance model, to investigate how and why intra-annual variability (seasonality) of
72 $\delta^{18}\text{O}_{\text{lakewater}}$ changed over time.

73

74 **2 Site description and core material**

75

76 Nar Gölü (38°20'24''N, 34°27'23''E; 1363 m.a.s.l.; Figure 1) is a maar lake, ~0.6 km²
77 in area and >20 m deep, located in the Cappadocia region of central Turkey. The
78 climate of the region is continental Mediterranean (Kutiel and Türkeş, 2005), with
79 precipitation at a nearby meteorological station in Niğde, 45 km from Nar Gölü,

80 averaging 339 mm per year and peaking in April and May. The crater geology is
81 dominated by basalt and ignimbrite (Gevrek and Kazancı, 2000). The limnology and
82 contemporary sedimentation patterns are described in detail in Dean et al. (2015a), but
83 in summary endogenic carbonate precipitation in the lake surface waters is weighted
84 towards the early summer (end of June/beginning of July), whereas diatom production
85 is weighted towards the spring (end of March/beginning of April). There was ~1.6‰
86 intra-annual variability in $\delta^{18}\text{O}_{\text{lakewater}}$ through our June 2011 to July 2012 monitoring
87 period (the period for which we have samples through all seasons), ~0.5‰ of which
88 occurred between the estimated time of peak diatom growth in spring 2012 and
89 carbonate formation in the early summer 2012 (Figure 2). We believe the timing of
90 diatom growth and carbonate precipitation is likely to have stayed roughly the same
91 through the Holocene. As we show in section 4, $\delta^{18}\text{O}_{\text{lakewater}}$ reconstructed for the time
92 of diatom growth is almost always lower than $\delta^{18}\text{O}_{\text{lakewater}}$ reconstructed for the time of
93 carbonate precipitation, and this would not be the case if diatom growth was weighted
94 to the summer or early autumn (Figure 2). Indeed, previous work showed there were
95 three planktonic/facultative planktonic ‘bloom’ taxa common in the Nar Gölü diatom
96 record over the last 1,700 years that are likely to have been spring blooming: *Synedra*
97 *acus*, *Nitzschia palaeacea* and *Cyclotella meneghiniana* (Woodbridge and Roberts
98 2011). These taxa were also the dominant ‘bloom’ diatoms in the early Holocene
99 (11,700-6,500 years BP) and it is reasonable to assume that their seasonal ecology was

100 the same at that time as during the late Holocene. The only additional early Holocene
101 bloom diatom is *Aulacoseira ambigua*, but this is only important in two samples
102 (11,657 and 11,403 years BP). In the section from 4,400-3,900 years BP it is possible
103 that *Nitzschia palaea* was a bloom taxon and it is likely to have been spring blooming
104 like *N. palaeacea*. The majority of carbonate is always likely to have precipitated in the
105 early summer in response to increasing evaporation (Dean et al., 2015a).

106

107 Figure 1

108

109 Figure 2

110

111 There have been a number of previous palaeolimnological investigations of the Nar
112 Gölü sediments (e.g. Jones et al., 2006; England et al., 2008; Woodbridge and Roberts,
113 2010). Here we combine data from the original core sequence taken in 2001/2
114 (NAR01/02) with new data from a longer core sequence taken in 2010 (Roberts et al.,
115 2016). The chronology of the NAR10 core was constructed by combining varve
116 counting and U-Th dates (Dean et al., 2015b).

117

118 **3 Methods**

119

120 *3.1 Isotope sample preparation and mass spectrometry*

121

122 $\delta^{18}\text{O}_{\text{carbonate}}$ data were produced using classic vacuum techniques and an Optima dual-
123 inlet mass spectrometer, as described in detail in Dean et al. (2015b). Specifically, the
124 carbonate analysed for isotopes from the Nar Gölü record was calcite and aragonite, as
125 detailed in Dean et al. (2015b). Data are given as ‰ deviations from VPDB and
126 analytical reproducibility was 0.1‰ for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$.

127

128 Samples for $\delta^{18}\text{O}_{\text{diatom}}$ analysis need to be as free as possible of non-diatom material
129 since the analytical methods used will liberate oxygen from these other components of
130 the sediment, such as carbonate and detrital silicates. Samples were therefore processed
131 using techniques similar to those of Morley et al. (2004), with the use of hydrogen
132 peroxide, nitric acid (to help remove organics; Tyler et al., 2007), hydrochloric acid,
133 differential settling, sieving at 10 μm and heavy liquid separation stages. $\delta^{18}\text{O}_{\text{diatom}}$
134 analysis was carried out on cleaned diatom samples using the stepwise fluorination
135 technique and a Thermo Finnigan MAT 253 at the NERC Isotope Geosciences
136 Facilities. The method is described in Leng and Sloane (2008) and has been verified
137 through an inter-laboratory comparison exercise (Chapligin et al., 2011). The data are
138 presented as ‰ deviations from VSMOW and analytical reproducibility was 0.3‰.

139

140 Diatom isotope samples prepared from ~8,800-7,900 and ~4,000-2,350 years BP had
 141 insufficient diatom silica for analysis, although there were still diatoms growing in the
 142 lake at this time (Roberts et al., 2016).

143

144 3.2 Correction of diatom isotope data

145

146 The samples from Nar Gölü still contained residual detrital silicates after the preparation
 147 described above due to a lack of density contrast between the detrital silicates and the
 148 diatoms, which reduced the efficacy of heavy liquid separation (Dean et al., 2013). A
 149 correction was, therefore, applied to account for the impact of detrital silicates on $\delta^{18}\text{O}$
 150 (Mackay et al. 2011):

151

$$152 \delta^{18}\text{O}_{\text{corrected-diatom}} = (\delta^{18}\text{O}_{\text{diatom}} - \delta^{18}\text{O}_{\text{contamination}} \times [\%_{\text{contamination}} / 100]) / (\%_{\text{diatom}} / 100) \quad (1)$$

153

154 where $\delta^{18}\text{O}_{\text{diatom}}$ is the original isotope value of the prepared diatom sample, $\%_{\text{contamination}}$
 155 and $\%_{\text{diatom}}$ are calculated using Eq. 2 (details below) and $\delta^{18}\text{O}_{\text{contamination}}$ is the isotope
 156 value of contamination.

157

158 A number of modifications were made to the methodology for the contamination
 159 correction of $\delta^{18}\text{O}_{\text{diatom}}$ samples that was previously used for Nar Gölü sediments (Dean

160 et al., 2013) to make it more robust. For element concentration data, here we use an
161 XRF (Panalytical epsilon 3 XL) rather than an Energy-Dispersive X-ray Spectroscopy
162 (EDS) probe, allowing for more precise measurements of aluminium concentrations (a
163 good marker for the amount of detrital silicates present (Mackay et al., 2011)), with an
164 analytical reproducibility of 0.03%. The XRF was set up to quantify the proportions of
165 Na, Mg, Al, Si, P, S, K, Ca, Ti, Mn and Fe using the Panalytical Omnic program.
166 Instead of calculating the $\delta^{18}\text{O}$ of contamination through the intercept of the $\delta^{18}\text{O}_{\text{diatom}}$
167 vs. contamination plot, nine turbidites from along the NAR10 core were prepared and
168 run in the same way as the diatom isotope samples. They had a mean $\delta^{18}\text{O}$ value of
169 16.0‰ ($\pm 1.0\%$), which is within uncertainty of the value of 16.5‰ estimated in Dean et
170 al. (2013) from NAR01/02. It is likely that % contamination was overestimated in Dean
171 et al. (2013) because some minerogenic contamination will be removed by the first
172 fluorination stage before $\delta^{18}\text{O}$ is measured (Swann and Leng, 2009) and diatom
173 frustules can incorporate aluminium, so $\text{Al}_2\text{O}_3\%$ in the samples does not only reflect
174 minerogenic contamination (Beck et al., 2002; Koning et al., 2007; Swann, 2010; Ren et
175 al., 2013). To investigate the latter effect, Scanning Electron Microscopy (SEM) was
176 used to identify individual clean diatoms (i.e. with no detrital silicates visible at all) and
177 the Al_2O_3 wt% of the individual diatoms was measured by EDS, averaging $1.0\% \pm 0.4$
178 (1σ) for the individual diatoms measured across 16 samples. This suggests that there is a
179 significant amount of diatom-bound aluminium, so a correction factor was applied to

180 account for this. Based on the average Al_2O_3 value of the turbidite layers throughout the
181 core sequence that were prepared and run as $\delta^{18}\text{O}_{\text{diatom}}$ samples, 14.56% Al_2O_3
182 represents 100% contamination (i.e. all detrital silicates, no diatoms). 1‰ Al_2O_3
183 represents 0% contamination. Thus, there is an equation, derived from Figure SI-1, that
184 can be used to calculate the new %_{contamination} values for our samples:

185

$$186 \quad \%_{\text{contamination}} = (7.3746 \times \text{sample}_{\text{Al}}) - 7.3746 \quad (2)$$

187

188 where $\text{sample}_{\text{Al}}$ is the measured Al_2O_3 concentration in each sample analysed for
189 $\delta^{18}\text{O}_{\text{diatom}}$. Eq. 2 was used to calculate the %_{contamination} values for Eq. 1. This modified
190 methodology was used on the new samples from NAR10, as well as to recalculate the
191 corrections to the NAR01/02 data presented in Dean et al. (2013). Henceforth, $\delta^{18}\text{O}_{\text{diatom}}$
192 refers to the corrected $\delta^{18}\text{O}_{\text{diatom}}$ data.

193

194 Uncertainties from individual components of the correction are outlined in Table 1 and
195 were combined to calculate the overall uncertainty associated with the correction.

196 Uncertainties are reduced compared to those reported in Dean et al. (2013) because of
197 the improved methodology. Figure SI-2 shows the original corrected NAR01/02 data
198 published in Dean et al. (2013) compared to re-calculated values used in this paper.

199 Although the actual values are slightly different and not all of the samples from Dean et

200 al. (2013) had sufficient material remaining for re-analysis by XRF (so data are now
 201 excluded), the general trends are very similar, with periods of lower $\delta^{18}\text{O}$ particularly at
 202 1,450, 1,250 and 120 years BP. The overall similarities in trends mean that the
 203 interpretations of Dean et al. (2013) are still valid, although for consistency in this paper
 204 we present the re-analysed NAR01/02 data along with the NAR10 data.

205

206 3.3 *Calculating $\delta^{18}\text{O}_{\text{lakewater}}$*

207

208 To allow for direct comparison of the $\delta^{18}\text{O}$ data from carbonates and diatoms, we
 209 estimate $\delta^{18}\text{O}_{\text{lakewater}}$ at the time of carbonate precipitation and diatom growth using the
 210 calcite (Kim and O'Neil, 1997), aragonite (Grossman and Ku, 1986) and diatom
 211 (Crespin et al., 2010) palaeotemperature equations respectively:

212

$$213 \delta^{18}\text{O}_{\text{lakewater}} = \delta^{18}\text{O}_{\text{calcite}} - (4.58 \pm [4.58^2 - 4 \times 0.08 \times (13.8 - T)]^{1/2}) / 2 \times 0.08 \quad (3)$$

214

$$215 \delta^{18}\text{O}_{\text{lakewater}} = \delta^{18}\text{O}_{\text{aragonite}} - (T - 19.7) / -4.34 \quad (4)$$

216

$$217 \delta^{18}\text{O}_{\text{lakewater}} = \delta^{18}\text{O}_{\text{diatom}} - (T - 245) / -6.25 \quad (5)$$

218

219 where $\delta^{18}\text{O}_{\text{lakewater}}$ and $\delta^{18}\text{O}_{\text{diatom}}$ are expressed on the VSMOW scale, $\delta^{18}\text{O}_{\text{calcite}}$ and
220 $\delta^{18}\text{O}_{\text{aragonite}}$ against VPDB and T in °C. We use a temperature range of +15 to +20°C for
221 the time of carbonate precipitation and +5 to +10°C for the time of diatom growth,
222 justified by our measurements of seasonal lake waters from 2011-2013 (Figure 2 and
223 Eastwood et al., unpublished data). The temperature range for the time of diatom growth
224 has been reduced from that used in Dean et al. (2013), where we estimated +5 to +15°C,
225 because of our increased knowledge of intra-annual epilimnion temperature variability
226 with the additional years of temperature logging data from Nar Gölü. While we
227 recognise that there will have been changes in temperature during the Holocene, these
228 changes are likely to have been only a few degrees centigrade (see references in section
229 5.1), smaller than the ranges of 5°C given for the times of diatom growth and carbonate
230 precipitation.

231

232 3.4 *Lake isotope mass balance models*

233

234 To examine further the changes in hydroclimate seasonality and how this would be
235 recorded in the seasonality of the lake $\delta^{18}\text{O}$ system, we use an isotope mass balance
236 model, employing the equations outlined in Jones and Imbers (2010) and Jones et al.
237 (2016), and fully explained in the Supplementary Information. The equations are based
238 on monthly time steps to allow investigations of changing intra-annual $\delta^{18}\text{O}_{\text{lakewater}}$

239 variability under different climatic states that have been identified from the isotope data:
240 for the present day (Modern), the Mid Holocene (here meaning from approximately
241 6,000 to 1,600 years BP) and the Early Holocene.

242

243 For the present day, average monthly values of temperature (average [Tav], minimum
244 [Tmin] and maximum [Tmax]), total precipitation (P) and snowfall between 2005 and
245 2011 (only until 2010 for snowfall) from the meteorological station at Niğde were used
246 to drive a model of modern conditions in a lake with the same volume ($\sim 750,000 \text{ m}^3$)
247 and lake area ($556,500 \text{ m}^2$) as Nar Gölü (Table 2 and Supplementary Information).

248

249 In this modern lake setting, annual average $\delta^{18}\text{O}_{\text{lakewater}}$ in the model is 0.59‰ with a
250 range (intra-annual $\delta^{18}\text{O}_{\text{lakewater}}$ variability) of 1.06 (Table 2). This compares to
251 measured summer values at Nar Gölü of between -1.9 and -0.2 ‰ for the same period
252 (2005-2011), and an intra-annual range of ~ 1.6 ‰ (Dean et al., 2015a). The difference
253 between the measured data and the model are due to a number of factors. Firstly, the
254 model is for a lake in Niğde, the location of the nearest meteorological station, not for
255 Nar Gölü. This will affect the precipitation and evaporation components of the model,
256 and therefore the parameterisation of surface and groundwater inflow and outflow,
257 which have narrow windows for a given lake in a given location (Jones et al., 2016).
258 Nar Gölü is also stratified, adding a level of complexity to the isotope hydrology not

259 included in the model. However, the model in the Modern scenario has mean and intra-
260 annual $\delta^{18}\text{O}$ values in the same order as Nar Gölü, and is used here not to recreate
261 conditions at Nar Gölü precisely, but to inform our discussion of why $\delta^{18}\text{O}$ may change
262 in time. As such, the model is deliberately simple, and appropriate. Inputs to the model
263 for the palaeoclimate scenarios are based on our best understanding of regional
264 temperature and precipitation changes from the literature (see discussions below).

265

266 **4 Results**

267

268 Figure 3 shows $\delta^{18}\text{O}_{\text{carbonate}}$ and $\delta^{18}\text{O}_{\text{diatom}}$ plotted against depth. There are gaps in both
269 the $\delta^{18}\text{O}_{\text{carbonate}}$ record, where interpretation of $\delta^{18}\text{O}_{\text{carbonate}}$ values is complicated by
270 dolomite precipitation (Dean et al., 2015b), and the $\delta^{18}\text{O}_{\text{diatom}}$ record, because there was
271 not enough diatom silica for isotope analysis and/or samples were too contaminated
272 (with detrital silicates and at times additionally with dolomite), even after cleaning, to
273 run. Because of issues with the chronology discussed elsewhere (Dean et al., 2015b;
274 Roberts et al., 2016), the data between 1034-1161 cm are not plotted on Figure 4.

275

276 Figure 3

277

278 Figure 4

279

280 The overall trends in $\delta^{18}\text{O}_{\text{carbonate}}$ and $\delta^{18}\text{O}_{\text{diatom}}$ are similar. Both have lower values
281 towards the bottom of the core in the period likely to be at the time of the Bølling-
282 Allerød, higher values at the time of the Younger Dryas, and lower values in the early
283 Holocene (Figure 4). Both $\delta^{18}\text{O}_{\text{diatom}}$ and $\delta^{18}\text{O}_{\text{carbonate}}$ increase at ~7,500 years BP to
284 higher values (by 4‰ VSMOW for $\delta^{18}\text{O}_{\text{diatom}}$ and ~5‰ VPDB for $\delta^{18}\text{O}_{\text{carbonate}}$).
285 However, a major difference is that while there is another increase in $\delta^{18}\text{O}_{\text{carbonate}}$ (>2‰
286 VPDB) ~4,100 years BP, ending with peak Holocene values that are maintained until
287 ~1,600 years BP, there is no corresponding second increase in $\delta^{18}\text{O}_{\text{diatom}}$ values. Where
288 data are available, $\delta^{18}\text{O}_{\text{diatom}}$ values are relatively stable, at c.+37‰ VSMOW for the
289 period ~7,000 to 1,600 years BP after rising from early Holocene values of c.+33‰.
290 Both $\delta^{18}\text{O}_{\text{carbonate}}$ and $\delta^{18}\text{O}_{\text{diatom}}$ decline dramatically at ~1,600 years BP for ~400 years,
291 before returning to higher values for most of the last 1,000 years.

292

293 Figure 4 also shows $\delta^{18}\text{O}_{\text{lakewater}}$ estimated for the times of diatom growth and carbonate
294 precipitation. Because late glacial temperatures are not well known, we only use the
295 palaeotemperature equations to reconstruct $\delta^{18}\text{O}_{\text{lakewater}}$ for the Holocene, during which
296 annual average temperatures probably only changed by a few degrees in the region (e.g.
297 Emeis et al., 2000). The shaded areas on Figure 4C combine maximum and minimum
298 $\delta^{18}\text{O}_{\text{lakewater}}$ values possible for the temperature ranges noted above, plus the

299 uncertainties associated with the $\delta^{18}\text{O}_{\text{diatom}}$ contamination correction. $\delta^{18}\text{O}_{\text{lakewater}}$ at the
300 time of diatom growth increased from c. -5‰ in the early Holocene to c. -1‰ in the mid
301 Holocene, before falling to c. -15‰ \sim 1,600-1,200 years BP and then returning to higher
302 values (c. -2 to -3‰) for the last 1,000 years. $\delta^{18}\text{O}_{\text{lakewater}}$ at the time of carbonate
303 precipitation increased from c. -3‰ in the early Holocene to c. $+1\text{‰}$ \sim 6,600 years BP
304 and to c. $+3\text{‰}$ by \sim 4,000 years BP, before falling to c. -4‰ \sim 1,600-1,200 years BP and
305 then increasing to c. -1‰ for the last 1,000 years.

306

307 $\Delta\delta^{18}\text{O}_{\text{lakewater}}$, the difference between $\delta^{18}\text{O}_{\text{lakewater}}$ at the time of carbonate precipitation
308 compared to the time of diatom growth, was only $\sim 1\text{‰}$ in the early Holocene. It then
309 increased to $\sim 4\text{‰}$ for much of the time from \sim 4,100 to 1,600 years BP, as $\delta^{18}\text{O}_{\text{lakewater}}$ at
310 the time of carbonate precipitation increased 4,100 years BP, but $\delta^{18}\text{O}_{\text{lakewater}}$ at the time
311 of diatom growth did not (Figure 4C). Then, \sim 1,600-1,200 years BP, because the fall in
312 $\delta^{18}\text{O}_{\text{diatom}}$ is much greater than the fall in $\delta^{18}\text{O}_{\text{carbonate}}$, $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ values are $>10\text{‰}$.
313 For the last 1,000 years, $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ declined to levels more similar to the early
314 Holocene. Limited variability in recent times is also shown in our monitoring data, with
315 only a 0.5‰ difference in our lakewater samples between April and July in 2012
316 (Figure 2) and a 0.7‰ difference seen between April and August 2002 (Jones et al.,
317 2005).

318

319 5 Discussion

320

321 From the isotope data, there appear to be three key lake states: 1. limited difference
322 between $\delta^{18}\text{O}_{\text{lakewater}}$ at the times of diatom growth and carbonate precipitation, i.e.
323 $\Delta\delta^{18}\text{O}_{\text{lakewater}} \sim 1\text{‰}$ (during the early Holocene and last 1,000 years); 2. intermediate
324 $\Delta\delta^{18}\text{O}_{\text{lakewater}}$, at $\sim 4\text{‰}$ (mid Holocene and up to $\sim 1,600$ years BP), and 3. maximum
325 $\Delta\delta^{18}\text{O}_{\text{lakewater}}$, at $\sim 10\text{‰}$ ($\sim 1,600$ - $1,200$ years BP). We discuss these in turn. The
326 differences in resolution between the carbonate and diatom isotope data means that we
327 limit ourselves to comparing the long-term general trends in the data through the early
328 and mid Holocene.

329

330 5.1 *The early Holocene (11,700 to 6,500 years BP)*

331

332 $\delta^{18}\text{O}_{\text{diatom}}$ and $\delta^{18}\text{O}_{\text{carbonate}}$ values for the early Holocene are both low relative to the mid
333 and late Holocene (Figure 4), which could indicate higher annual average P/E (i.e.
334 effectively wetter conditions) in the early Holocene, as has been suggested by other
335 studies (summarised in Roberts et al., 2008). Specifically, pollen data (Djamali et al.,
336 2010; Kotthoff et al., 2008; Peyron et al., 2011; Peyron et al., 2017), microcharcoal data
337 (Wick et al., 2003; Turner et al., 2008; Vannièrè et al., 2011), climate modelling results
338 (Brayshaw et al., 2010) and $\delta^{18}\text{O}$ data of freshwater mollusc shells from Çatalhöyük

339 ~160 km SW of Nar (Bar-Yosef Mayer et al., 2012; Lewis et al., 2017) have suggested
340 that the early Holocene in the Eastern Mediterranean region had wetter winters than
341 present, but with many of the studies suggesting drier springs and/or summers. Annual
342 average temperatures were several degrees cooler in the early Holocene compared to the
343 late Holocene, as reconstructed by alkenone-derived sea surface temperatures (Emeis et
344 al., 2000; Triantaphyllou et al., 2009) and speleothem fluid inclusions (McGarry et al.,
345 2004). However, the prominence of *Pistacia* in the pollen record from Nar Gölü
346 (Roberts et al., 2016) and from nearby Eski Acıgöl (Roberts et al., 2001; Woldring and
347 Bottema, 2003), between 11,000 and 8,000 years BP, suggests winters were milder than
348 today (Rossignol-Strick, 1999). Therefore, the inferred drops in annual temperature may
349 have been concentrated in the summer. There is, however, a gap in the $\delta^{18}\text{O}_{\text{diatom}}$ record
350 between 8,800 and 7,900 years BP due to there being too little diatom silica for diatom
351 isotope measurements to be made. Intriguingly, this period coincides with a phase of
352 marked spring floods on the Çarşamba river in Anatolia (Boyer et al., 2006), which
353 would have been caused by enhanced spring snowmelt in its upper watershed in the
354 Taurus mountains. Despite the fact that spring and summer precipitation may have been
355 lower in the early Holocene than the present day, $\delta^{18}\text{O}_{\text{carbonate}}$ is still lower in the early
356 Holocene and there is limited $\Delta\delta^{18}\text{O}_{\text{lakewater}}$. Presumably, the lower $\delta^{18}\text{O}_{\text{carbonate}}$ and
357 limited $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ is due to relatively less summer evaporation of the lake waters
358 compared to the mid and late Holocene, which is to be expected if there were lower

359 temperatures in the early Holocene spring/summer, as well as increased winter
360 precipitation. Our mass balance modelling allows us to refine our basic interpretation of
361 hydroclimate in the early Holocene.

362

363 In our early Holocene model, we have reduced the annual average temperature by 1°C,
364 as estimated from the studies cited above and as used in Jones et al. (2007); details in SI
365 Tables. Annual precipitation values are kept the same as the present day, but the
366 seasonal distribution has been shifted to more winter-dominated with no snow, as is
367 indicated by the literature discussed above. Under this scenario, average annual lake
368 water values are lower than the present day model (-2.81‰), and could be even more so
369 if annual-averaged precipitation was increased under the same P/E seasonality regime,
370 as seems possible (Roberts et al., 2008). This demonstrates that the seasonality of P/E,
371 in addition to the average annual conditions, is important in controlling inter-annual
372 changes in $\delta^{18}\text{O}_{\text{lakewater}}$.

373

374 To investigate further the relative contributions of precipitation and temperature (linked
375 closely to evaporation in this model), an early Holocene scenario, using modern day
376 temperatures (as well as modern day annual-average precipitation levels again) and
377 changing only the seasonal distribution of precipitation, was also undertaken. Here
378 $\delta^{18}\text{O}_{\text{lakewater}}$ was still lower than the present day scenario (-0.57‰) and the average of

379 monthly P/E increases (Table 2). This result drives a difference in this model because
380 groundwater inflow and outflow are dependent on P/E, with additional groundwater
381 outflow required in the early Holocene compared to present day to balance the lake
382 system, and suggesting higher lake levels under early Holocene conditions. This
383 indicates that changing the seasonal distribution of P/E, irrespective of annual average
384 conditions, can lead to changes in both lake hydrology and lake isotope composition. It
385 highlights the need to be careful when suggesting that the early Holocene was ‘wetter’
386 than the mid and late Holocene based solely on evidence from lake sediment isotopes,
387 as now it is clear that changes in the seasonality of P/E have an impact on $\delta^{18}\text{O}$, in part
388 due to changes in seasonal water balance as well as due to changes in $\delta^{18}\text{O}$ of
389 precipitation (Table 2), as suggested by Stevens et al. (2001, 2006) for Lakes Zeribar
390 and Mirabad.

391

392 5.2 *The mid Holocene (~6,500 to ~1,600 years BP)*

393

394 At Nar Gölü, a number of proxies respond to changes in lake level, usually driven by
395 changes in P/E, such as lithology (varved vs. non-varved), carbonate mineralogy (calcite
396 vs. aragonite and dolomite) (Dean et al., 2015b), the Sr-Ca elemental ratio and certain
397 diatom species (Roberts et al., 2016). These multiple proxies indicate that annual
398 average P/E was probably lower after ~6,500 years BP compared to the early Holocene.

399 We know at Nar Gölü that lake level falls lead to more positive $\delta^{18}\text{O}_{\text{carbonate}}$ (Dean et al.,
400 2015a) and therefore a significant part of the $\delta^{18}\text{O}$ trend in carbonates and diatoms to
401 higher values in the mid and late Holocene, compared to the early Holocene, is likely
402 related to a shift to drier conditions. Other influences on $\delta^{18}\text{O}$, such as changes in the
403 isotopic composition of the source of precipitation, amount effect or temperature, could
404 not have accounted for the large size of the shift in both $\delta^{18}\text{O}_{\text{carbonate}}$ and $\delta^{18}\text{O}_{\text{diatom}}$ from
405 the early to mid and late Holocene.

406

407 $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ does not initially increase in the mid Holocene because both $\delta^{18}\text{O}_{\text{carbonate}}$
408 and $\delta^{18}\text{O}_{\text{diatom}}$ increase, but in the period ~4,100 to ~1,600 years BP $\delta^{18}\text{O}_{\text{lakewater}}$ at the
409 time of diatom growth is up to ~4‰ lower than at the time of carbonate precipitation
410 (Figure 4). Annual average precipitation must have been lower for most of the mid and
411 late Holocene compared to the early Holocene (Jones et al., 2007). It is possible that a
412 significant share of this precipitation decline occurred ~7,500 years BP, while at ~4,100
413 years BP there was a rise in summer evaporation but winter/spring precipitation levels
414 did not change substantially. If that was the case, that would explain why both $\delta^{18}\text{O}_{\text{diatom}}$
415 (responding more to winter/spring precipitation) and $\delta^{18}\text{O}_{\text{carbonate}}$ (responding more to
416 summer evaporation) increased ~7,500 years BP but only $\delta^{18}\text{O}_{\text{carbonate}}$ increased at
417 ~4,100 years BP (thus leading to increased $\Delta\delta^{18}\text{O}_{\text{lakewater}}$). However, lake level change
418 could account for some of this increased $\Delta\delta^{18}\text{O}_{\text{lakewater}}$. $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ will be more

419 sensitive to inputs and outputs when the lake level and volume were lower, with less of
420 a buffering effect than when the lake level is higher: this is a well-known phenomenon
421 in limnology (e.g. Leng and Anderson, 2003; Steinman et al., 2010).

422

423 To test this with the lake isotope mass balance model, two model conditions are set for
424 this period. In both, precipitation is reduced compared to the present day as multi-proxy
425 evidence from Nar Gölü (Dean et al., 2015b; Roberts et al., 2016) and elsewhere in the
426 region (Roberts et al., 2008) points to lower lake levels at this time. In the first Mid
427 Holocene scenario (MHi), temperatures are held the same as the present day, resulting
428 in an average $\delta^{18}\text{O}_{\text{lakewater}}$ value of +1.06‰, which is higher than the early Holocene
429 scenarios and thus supports our contention that some of the increase in $\delta^{18}\text{O}$ could be
430 due to reduced annual precipitation. However, the range in the model is only 1.10‰
431 (Table 2), which is similar to the early Holocene model, despite the higher $\Delta\delta^{18}\text{O}_{\text{lakewater}}$
432 seen in the data in the mid Holocene compared to the early Holocene. In the second Mid
433 Holocene scenario (MHii), summer temperatures are raised to increase summer
434 evaporation such that P/E seasonality is increased relative to MHi. Average $\delta^{18}\text{O}_{\text{lakewater}}$
435 values become even more positive (+2.00‰) and the range increases (1.22‰; Table 2).
436 Further, a shift from a steady state lake with the same volume as the present day
437 scenario, in MHii conditions, to one with a 20% smaller volume, increases the intra-

438 annual $\delta^{18}\text{O}_{\text{lakewater}}$ range to 1.52‰, showing how a change to lower lake levels could
439 account for some of the increase in $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ at this time (as discussed above).

440

441 To ensure steady state lakes under the mid Holocene climatic scenarios, the
442 groundwater outflow constant has to be reduced (see Supplementary Information for
443 model details). In the model, this is partly a function of P/E as more water entering the
444 lake will push more of it out, however here it needs to be further reduced relative to
445 present day to ensure a steady state lake, i.e. one where volume is not always increasing
446 or decreasing at an annual time step. This suggests there are further controls on
447 groundwater outflow that are not described by our simple model, possibly linked to lake
448 volume and depth, with the lower lake levels of the mid Holocene also potentially
449 contributing to reduced groundwater outflow at these times.

450

451 5.2.3 *Late Holocene (last 1,600 years)*

452

453 Around 1,600-1,200 years BP, $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ was at times >10‰. Dean et al. (2013)
454 hypothesised that this was due to a seasonal freshwater lid of low $\delta^{18}\text{O}$ snowmelt
455 occurring at this time, in which the diatoms lived. To further investigate the sensitivity
456 of the Nar Gölü system to snow volume, the modern lake isotope mass balance model
457 was altered to have no snow, or double the amount of snow, keeping all other variables

458 the same. This produced more positive or more negative annual average $\delta^{18}\text{O}_{\text{lakewater}}$
459 values respectively, as would be expected by putting less or more negative $\delta^{18}\text{O}$ water
460 into the system (Table 2). There is no impact on the range if these changes are made
461 into a well-mixed lake system as in the model, further suggesting that density
462 differences and stratification are probably important in explaining the $\Delta\delta^{18}\text{O}_{\text{lakewater}}$
463 variability reconstructed down-core at Nar Gölü as proposed by Dean et al. (2013) for
464 this unusual period during the late Holocene.

465

466 **6 Conclusions**

467

468 The combination of two $\delta^{18}\text{O}$ records, from diatoms and endogenic carbonate that
469 formed in Nar Gölü in central Turkey at different times of the year, helps to inform
470 discussion of palaeoseasonality. Our record indicates that there are three lake states
471 through the Holocene: the early Holocene and the last 1,000 years when there is limited
472 $\Delta\delta^{18}\text{O}_{\text{lakewater}}$, the mid Holocene and up to ~1,600 years BP when $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ was at
473 times ~4‰ and a short period ~1,600-1,200 years BP when $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ was ~10‰.
474 Modelling results indicate that the increase in $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ from the early to the mid
475 Holocene could be related to changes in P/E seasonality, but a shift to lower lake levels
476 (and volumes) would have amplified the impact of any changes in P/E. Therefore,
477 while we have shown that using $\Delta\delta^{18}\text{O}_{\text{lakewater}}$ to compare lake conditions at different

478 times of the year can provide insights into seasonality, it is not a simple proxy for intra-
479 annual P/E variability. In terms of inter-annual $\delta^{18}\text{O}$ change, we suggest that lower
480 $\delta^{18}\text{O}_{\text{carbonate}}$ and $\delta^{18}\text{O}_{\text{diatom}}$ values in the early Holocene compared to the present day
481 could partly be the result of changes in the seasonality of P/E. However, the multi-proxy
482 evidence available from Nar Gölü clearly points to a mid Holocene transition to lower
483 lake levels driven by annual-mean shifts to reduced P/E.

484

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498

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695

696 **Table 1** Sources of uncertainty associated with the correction of $\delta^{18}\text{O}_{\text{diatom}}$ data used in
 697 this paper.

Source of uncertainty	Magnitude of uncertainty
Diatom isotope measurement analytical reproducibility (1σ)	0.3‰
Al_2O_3 measurement analytical reproducibility (1σ)	0.03%
Variance in Al_2O_3 composition of turbidites (from \bar{x} of 14.56%) (1σ)	1.6%
Variance in $\delta^{18}\text{O}$ value of turbidites from \bar{x} of 16.0‰ (1σ)	1.0‰

698

699

700 **Table 2** Lake isotope mass balance model summary

701

	δl (‰)		Tav (°C)	P (mm)	δp (‰)	Qi average	Qo average	Volume	P/E
	Mean	Range	Mean	Total	Weighted Mean	(m ³ /month)	(m ³ /month)	(m ³)	Annual average
Modern	0.59	1.06	11.7	356.2	-9.4	76328	39812	7500000	0.422
with no snow	0.71	1.05	11.7	356.2	-8.5	76328	39812	7500000	0.422
with double snow	0.50	1.06	11.7	356.2	-10.4	76328	39812	7500000	0.422
Mid Holocene i	1.06	1.10	11.7	295.2	-8.8	71991	32851	7500000	0.398
Mid Holocene ii	2.00	1.22	12.6	295.2	-8.8	70781	25635	7500000	0.391
	2.00	1.52	12.6	295.2	-8.8	70781	25635	6000000	0.391
Early Holocene	-2.81	1.19	10.7	356.2	-8.9	116438	86320	7500000	0.643
	-2.81	0.99	10.7	356.2	-8.9	116438	86320	9000000	0.643
with modern temperatures	-0.57	1.21	11.7	356.2	-8.9	90813	54422	7500000	0.502

702

703

704 **Figure captions**

705

706 **Figure 1** Location of Nar Gölü in Turkey and lakes Zeribar and Mirabad in Iran.

707

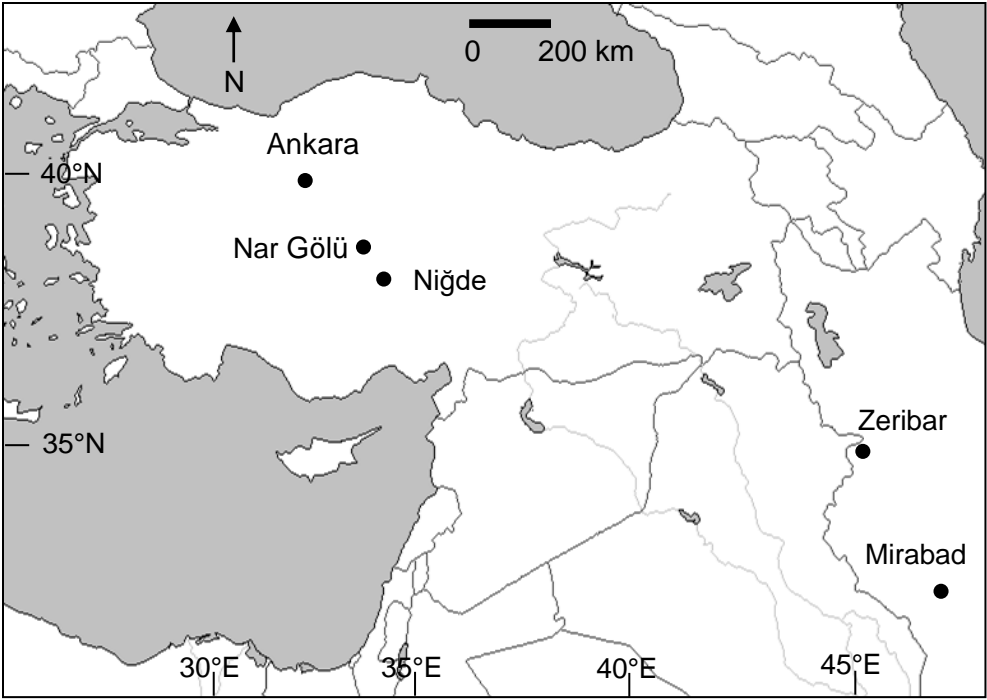
708 **Figure 2** Seasonal data from 2011-2012, showing increase in lake water $\delta^{18}\text{O}$ (A) and
709 temperature (B) between the estimated times of year of diatom growth (i) and carbonate
710 formation (ii).

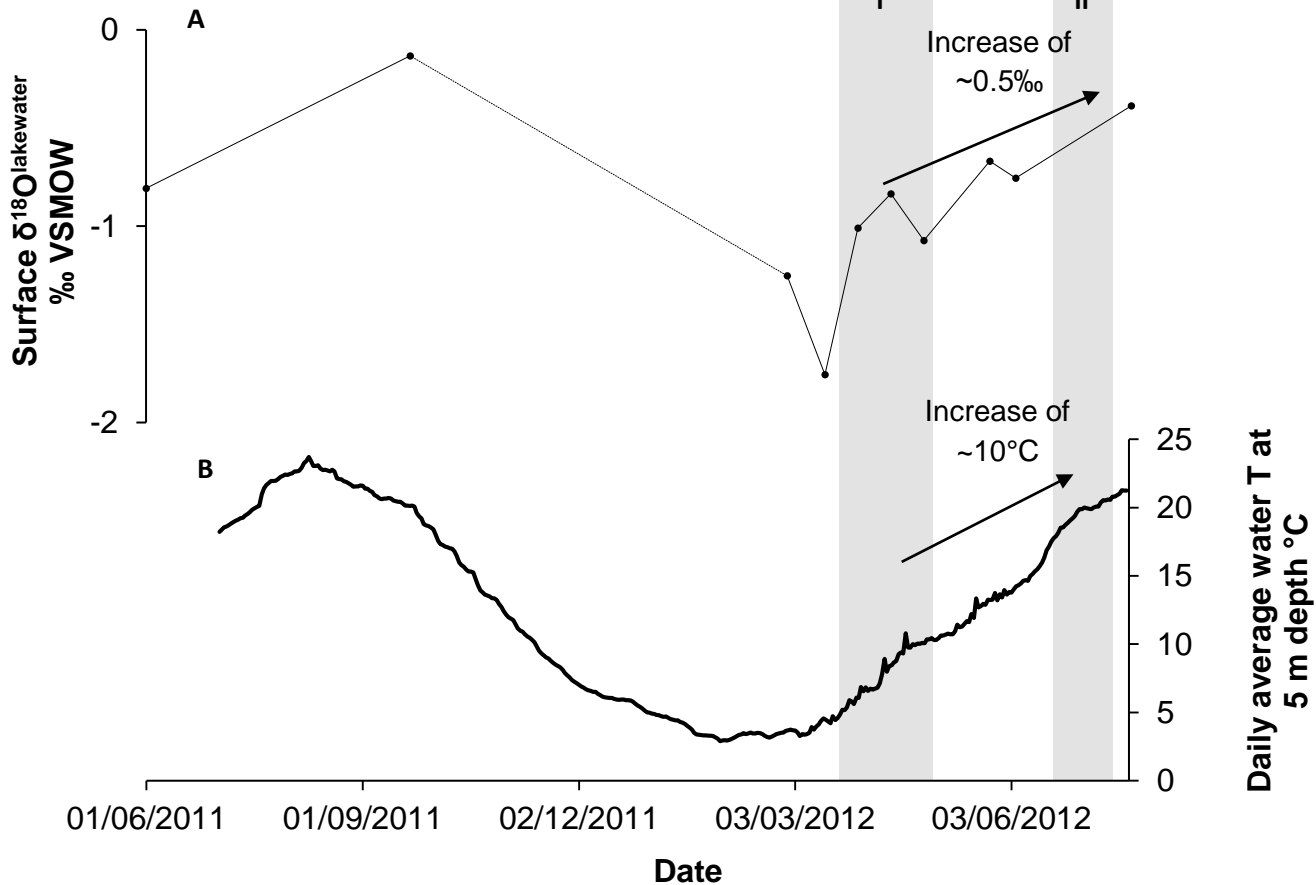
711

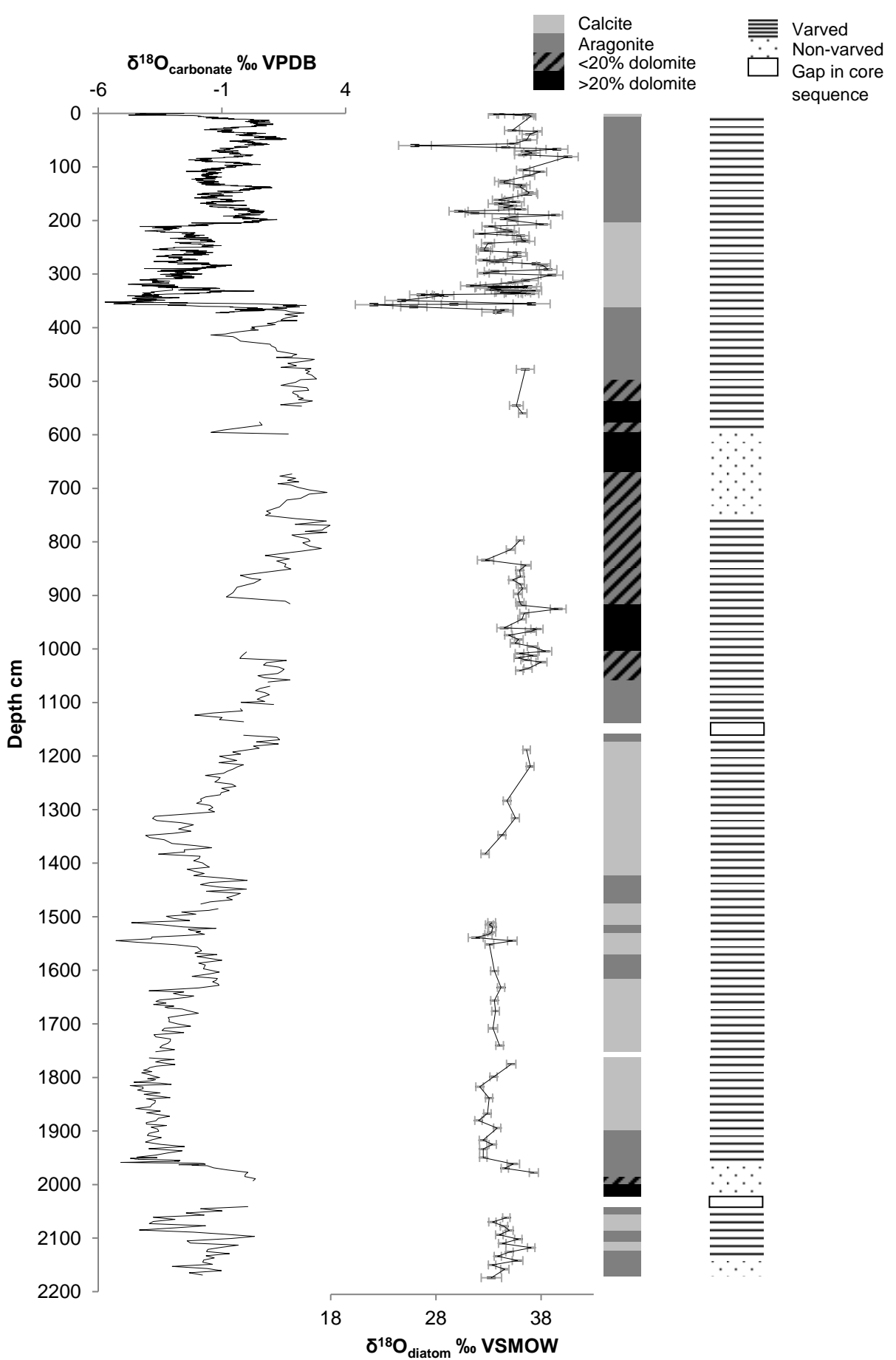
712 **Figure 3** $\delta^{18}\text{O}_{\text{diatom}}$ and $\delta^{18}\text{O}_{\text{carbonate}}$ data plotted against depth, with the error bars on $\delta^{18}\text{O}_{\text{diatom}}$
713 representing the combined uncertainties from Table 1. There are no carbonate isotope data in
714 sections where there were gaps due to coring (shown by white boxes on the lithology plot) or
715 where there were high levels (>20%) of dolomite (explained in detail in Dean et al., 2015b).
716 Gaps in the diatom isotope data are due to gaps in coring or insufficient amounts of diatom
717 silica.

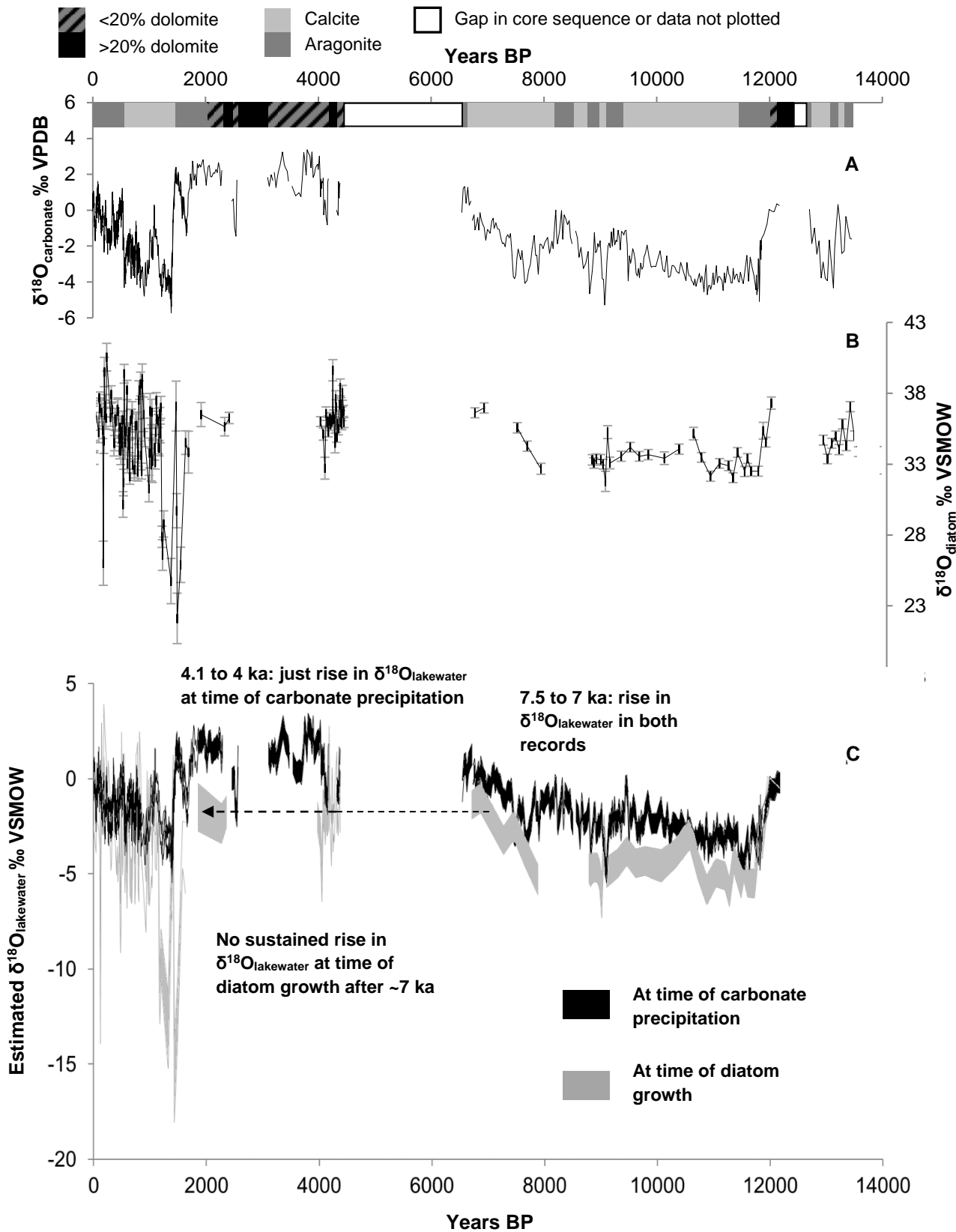
718

719 **Figure 4** (A) $\delta^{18}\text{O}_{\text{carbonate}}$ (with carbonate mineralogy data) and (B) $\delta^{18}\text{O}_{\text{diatom}}$, with (C) data
720 converted to $\delta^{18}\text{O}_{\text{lakewater}}$ assuming a temperature range of +15 to +20°C for the time of
721 carbonate precipitation and +5 to +10°C for the time of diatom growth. Some isotope data
722 plotted against depth are not shown against age due to issues with the chronology (discussed
723 in detail in Dean et al., 2015b).









Supplementary Information for Seasonality of Holocene hydroclimate in the Eastern Mediterranean reconstructed using the oxygen isotope composition of carbonates and diatoms from Lake Nar, central Turkey

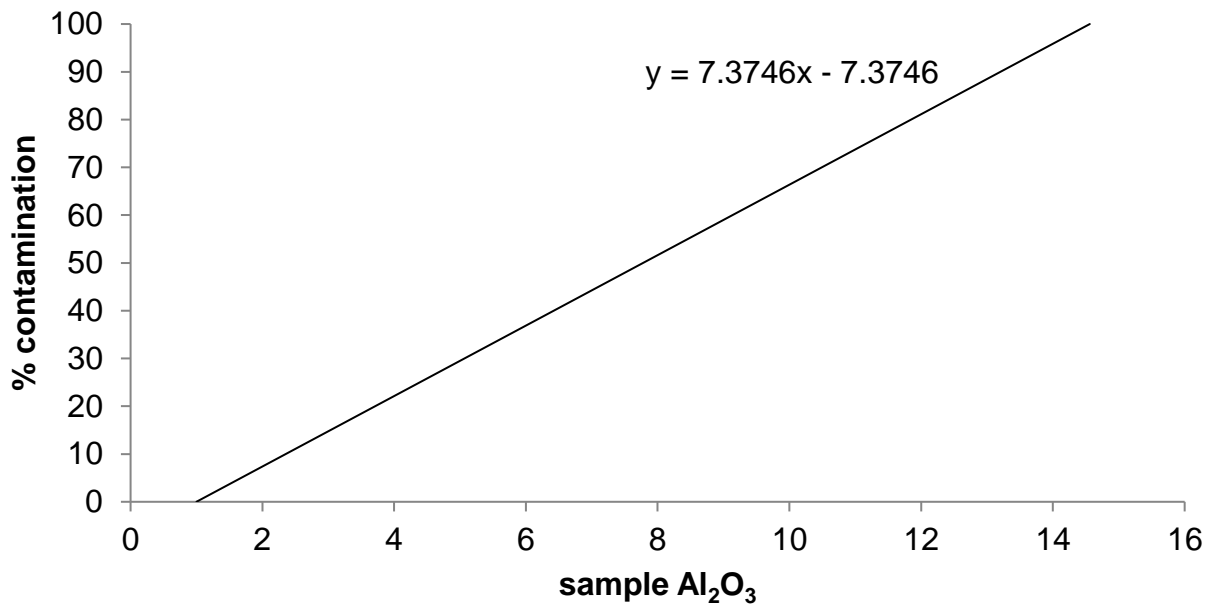


Figure SI-1 Regression line showing equation used to derive Eq. 2: a mixing line between the point when Al₂O₃ is 14.56% indicating 100% contamination and when Al₂O₃ is 1% indicating 0% contamination (i.e. 100% diatom).

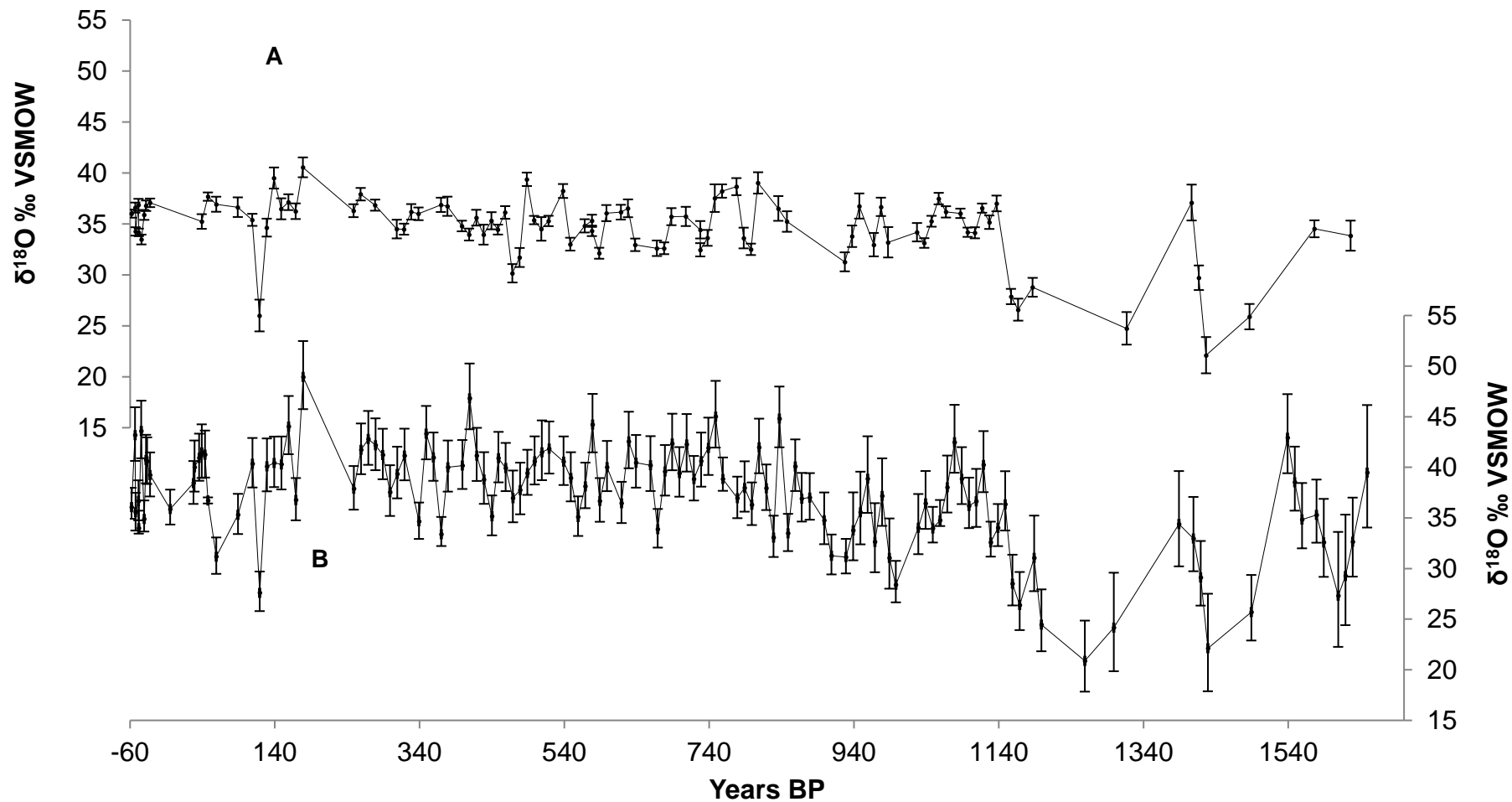


Figure SI-2 The difference between NAR01/02 diatom isotope trends in this paper (A) and as published in Dean et al. (2013) (B). Not all samples originally run and corrected in B could be included in A because many did not have sufficient material left to allow for XRF analysis. Error bars show the combined uncertainties from the factors given in Table 1.

Isotope Mass Balance Models

Theoretical model

The following is edited from Jones et al. (2016) and Jones and Imbers (2010) for the model lake used in this study.

The water mass and isotopic mass balance of a well-mixed lake is, respectively:

$$\frac{dV}{dt} = P + Qi - E - Qo \quad (1)$$

$$\frac{d}{dt}(V\delta_L) = P\delta_P + Qi\delta_P - E\delta_E - Qo\delta_L \quad (2)$$

where V is the lake volume, t , time, P , precipitation on lake surface per unit time, E is evaporation from lake surface per unit time and Q_o and Q_i are obtained as $Q_x = S_x + G_x$, where S_o and G_o and S_i and G_i are the surface and groundwater outflows and inflows respectively, and are measured in the same units as P and E . δ_P , δ_E and δ_L are the isotope values, either $\delta^{18}\text{O}$ or δD , of the precipitation, evaporation and lake waters respectively.

δ_E is difficult to measure and is therefore usually calculated (e.g. Steinman et al., 2010) using equations based on the evaporation model of Craig and Gordon (1965) such that

$$\delta_E = \frac{\alpha^*\delta_L - h\delta_A - \epsilon}{1 - h + 0.001\epsilon_k} \quad (3)$$

where α^* is the equilibrium isotopic fractionation factor dependent on the temperature at the evaporating surface and

$$\frac{1}{\alpha^*} = \exp(1137T_L^{-2} - 0.4256T_L^{-1} - 2.0667 \times 10^{-3}) \quad (4)$$

for oxygen and

$$\frac{1}{\alpha^*} = \exp(24844T_L^{-2} - 76.248T_L^{-1} - 52.61 \times 10^{-3}) \quad (5)$$

for hydrogen. T_L is the temperature of the lake surface water in degrees Kelvin (Majoube 1971). h is the relative humidity normalised to the saturation vapour pressure at the temperature of the air

water interface and ϵ_k is the kinetic fraction factor; for $\delta^{18}\text{O}$ ϵ_k has been shown to approximate $14.2(1-h)$ and $12.5(1-h)$ for $\delta^2\text{H}$ (Gonfiantini, 1986). δ_A is the isotopic value of the air vapour over the lake and $\epsilon = \epsilon^* + \epsilon_k$ where $\epsilon^* = 1000(1-\alpha^*)$.

In the model we use an equation derived from those above to calculate the isotopic value of lake waters (δ_L) at a given time, $t+\Delta t$, based on the value of δ_L at time t , and the inputs and outputs from the lake between t and $t + \Delta t$.

The left-hand side of Eq. 2 is expanded and Eq.1 substituted into it:

$$\frac{d}{dt}(V\delta_L) = V \frac{d\delta_L}{dt} + \delta_L \frac{dV}{dt} = \delta_L(P + Qi - E - Qo) + V \frac{d\delta_L}{dt} \quad (6)$$

and then re-written, such that δ_L dependences are explicit.

δ_E is expressed as a function of δ_L such that

$$\delta_E = A\delta_L + C \quad (7)$$

where, for Equation 3

$$A = \frac{\alpha^*}{1-h+0.001\epsilon_k} \text{ and } C = -\frac{h\delta_A+\epsilon}{1-h+0.001\epsilon_k}$$

Taking Eq. (2) and (6) and replacing δ_E using Eq. (7):

$$V \frac{d\delta_L}{dt} + \delta_L(P + Qi - E - Qo) = \delta_P(P + Qi) - E(A\delta_L + C) - Qo\delta_L \quad (8)$$

Rearranging all terms in Eq.(8) then leads to:

$$V \frac{d\delta_L}{dt} = \delta_P(P + Qi) - EC - \delta_L(P + Qi - E(1 - A)) \quad (9)$$

We define λ and β as: $\lambda = (P+Qi) \delta_P - EC$ and $\beta = P+Qi - E(1-A)$ such that equation (9) can be rewritten as:

$$V \frac{d\delta_L}{dt} = \lambda - \beta \delta_L \quad (10)$$

We assume that dV/dt can be adequately approximated as equal to the change of volume over 1 month and all other variables are also put into the model as rates per month.

Integrating equation (10) obtains an expression for the evolution of δ_L with time. At this stage we introduce a first approximation by assuming a constant value for V for each month; consistent with constant values of P and Q_i etc. over each month. The following parameterisation for V is used:

$$\bar{V} = \frac{V_{30th} + V_0}{2} \quad (11)$$

where V_{30th} is the total volume on the last day of each month, and V_0 is the initial volume on the first day of the month.

Integration of Eq. (10) after considering the approximation in equation (11) results in:

$$\ln \left(\frac{\lambda - \beta \delta_{L0}}{\lambda - \beta \delta_L} \right) = \frac{\beta}{\bar{V}} \Delta t \quad (12)$$

Where δ_{L0} is the initial isotopic composition (i.e. at the beginning of each month) and $\Delta t=1$ for each monthly step of our model. Finally exponentials of both sides of Eq. (12) give an expression for δ_L :

$$\delta_L = \frac{1}{\beta} (\lambda - (\lambda - \beta \delta_{L0}) \exp(-\frac{\beta}{\bar{V}})) \quad (13)$$

Values for this model

T_L: temperature of the lake surface water

From monitoring data of Lake Nar (Jones et al., 2005, Dean et al., 2015) and other studies (Jones et al., 2016) lake surface temperatures in the model are taken as the average of mean and maximum air temperatures.

h: normalised relative humidity

Relative humidity values were calculated based on present day relationships with temperature (c.f. Jones et al., 2005) such that these values could change in time in palaeo scenarios.

These values were normalised to the conditions at the lake surface using the saturation vapour pressure of the air and surface water as defined in Steinman et al. (2010).

E: Evaporation

Evaporation is calculated based on the equation of Linacre (1992) that has been shown previously (Jones et al., 2005; Jones et al., 2007) to be a reasonable measure of evaporation and is especially useful for palaeo-contexts where instrumental measurements are non-existent.

$$E(\text{mm/day}) = [0.015 + 4 \times 10^{-4} T_a + 10^{-6} z] \times [480 (T_a + 0.006z) / (84 - A) - 40 + 2.3 u (T_a - T_d)] \quad (14)$$

where T_a is air temperature ($^{\circ}\text{C}$), z = altitude (m), A = latitude, T_d = dew point temperature = $0.52 T_{a \text{ min}} + 0.60 T_{a \text{ max}} - 0.009 (T_{a \text{ max}})^2 - 2$ $^{\circ}\text{C}$.

δ_P : isotopic composition of precipitation

Values for the isotopic composition of rainfall at Nar came from the Online Isotopes in Precipitation Calculator (Bowen et al., 2005; Bowen, 2016).

Isotopic values of snow were based on sampling of snowfall from the catchment (Dean et al., 2013) and were fixed at -15‰ (i.e. more negative than rainfall).

Monthly values are kept as modern throughout, although the weighted annual mean values change as the amount of precipitation in a given month changes in each scenario (Table 2).

Qi: surface and groundwater inflow

The model lake has no surface inflow; this is similar to Lake Nar where there are no permanent stream inflows to the lake.

Monitoring of springs within the Nar catchment (Jones et al., 2005) has shown these to be meteoric water, such that the isotopic composition of inflowing waters to the model lake are considered to be the same as rainfall.

Values of Q_i and Q_o are optimised in the model to allow a stable lake with mean isotope values, and intra-annual range, similar to that of Lake Nar. In this model Q_i is a function of P:E.

Q_o: surface and groundwater outflow

There is no surface run off from the model lake, or from Lake Nar.

The amount of groundwater outflow is optimised for the model as described above and in the model lake is dependent on P:E, as the amount of groundwater inflow will change the flow of water through the lake, and a constant for when Q_i is potentially 0 such that the lake is balanced.

Table SI-1: precipitation values (mm) used in models

Month	Modern			Mid Holocene Rainfall	Early Holocene Rainfall
	Snow	Rainfall	Total		
Jan	17.0	16.2	33.2	40.0	51.0
Feb	15.1	21.7	36.7	36.7	46.0
Mar	7.3	31.1	38.4	30.0	40.0
Apr	2.8	44.5	47.2	25.0	30.0
May		38.8	38.8	20.0	20.0
Jun		21.4	21.4	15.0	10.0
Jul		7.7	7.7	7.0	7.7
Aug		7.3	7.3	7.3	7.3
Sep		17.2	17.2	17.2	17.2
Oct		31.6	31.6	25.0	31.0
Nov	6.5	35.3	41.8	32.0	45.0
Dec	13.4	21.4	34.8	40.0	51.0

Table SI-2: temperatures for Modern and Mid Holocene I scenarios (°C)

Month	Average (T _{av})	Minimum (T _{min})	Maximum (T _{max})
Jan	0.16	-4.05	5.47
Feb	1.46	-3.08	6.88
Mar	5.92	0.70	11.87
Apr	10.57	4.82	16.54
May	15.88	9.04	22.27
Jun	20.28	12.88	26.62
Jul	23.69	15.83	30.31
Aug	23.41	15.69	30.41
Sep	18.33	11.13	25.85
Oct	12.57	6.72	19.74
Nov	6.37	1.25	13.27
Dec	2.34	-1.93	7.96

Table SI-3: temperatures for Mid Holocene ii scenario (°C)

Month	Average (Tav)	Minimum (Tmin)	Maximum (Tmax)
Jan	0.16	-4.05	5.47
Feb	1.46	-3.08	6.88
Mar	5.92	0.70	11.87
Apr	10.57	4.82	16.54
May	17.00	10.00	23.00
Jun	21.50	14.00	28.50
Jul	25.00	17.00	31.50
Aug	25.50	16.50	31.00
Sep	21.50	12.00	25.85
Oct	15.00	7.00	19.74
Nov	6.37	1.25	13.27
Dec	2.34	-1.93	7.96

Table SI-4: temperatures for Early Holocene ii scenario (°C)

Month	Average (Tav)	Minimum (Tmin)	Maximum (Tmax)
Jan	0.16	0.00	5.47
Feb	1.46	0.50	6.88
Mar	5.92	0.70	11.87
Apr	10.57	4.82	16.54
May	15.00	9.04	21.00
Jun	18.00	10.00	25.00
Jul	20.00	13.00	28.00
Aug	20.00	13.00	28.00
Sep	17.00	10.00	25.00
Oct	12.57	5.00	18.00
Nov	6.37	3.00	12.00
Dec	2.34	0.00	7.00

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