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1 **Climatic variability over the last 30,000 years recorded in La Piscina de Yuriria, a**  
2 **Central Mexican Crater lake (DOI: 10.1002/jqs.2846)**  
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4

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19

20 Running head: Palaeolimnology of a Mexican Crater Lake  
21

22 **ABSTRACT:** The Trans-Mexican Volcanic Belt provides an excellent setting for  
23 reconstruction of late Quaternary climate from different natural archives. Moreover human  
24 impact on the landscape since the mid Holocene provides a good opportunity to  
25 investigate the complex interplay of natural and anthropogenic forcing of landscape  
26 change. However despite the wealth of records, understanding of the environmental  
27 history of the region and its wider significance for climate change across the northern  
28 neotropics remains incomplete. We present a radiocarbon-dated, multiple-proxy  
29 (sedimentology, sedimentary geochemistry, ostracods, diatoms, stable isotopes) record of  
30 climatic and environmental change based on the lacustrine sediments from La Piscina de  
31 Yuriria, a hydrologically-closed volcanic crater in the northern TMVB. Much of the last  
32 glacial interval was characterised by low effective moisture associated with a weakened  
33 North American Monsoon (NAM) although the interval from 30,000 to 27,500 aBP  
34 experienced abrupt changes in rainfall. The period corresponding to the late glacial stadial  
35 was also dry and the lake may have dried out at this time. There was a change to wetter  
36 but variable conditions during the early Holocene as the NAM strengthened. Progressive  
37 drying during the later Holocene was accompanied by phases of catchment disturbance,  
38 which were partly the result of human impact.  
39  
40  
41

42 **KEYWORDS:** Trans-Mexican Volcanic Belt; Palaeolimnology; diatoms; ostracods; stable  
43 isotopes  
44  
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## 49 **Introduction and previous work**

50 The highlands of the Trans-Mexican Volcanic Belt (TMVB), which cross Mexico at around  
51 19°N, provide a range of opportunities for reconstructing past climates through the study of  
52 the sediments in its many lake basins, the availability of glacial deposits, tree ring records  
53 and a wealth of historical documents (primarily since the Spanish conquest in 1521). In  
54 spite of these many possibilities, understanding of climatic variability over the late  
55 Quaternary, particularly the late Pleistocene and early Holocene, is still rather limited  
56 (Caballero *et al.*, 2010). A number of factors help to explain this, including poor dating  
57 control of many records, the effects of tectonic and volcanic activity, and especially human  
58 disturbance, which over at least the last 4000 years has had a profound impact on the  
59 natural environment to the extent that only records from the highest elevations may be  
60 unaffected (e.g. Lozano-Garcia and Vazquez-Selem, 2005). Lake sediment records have  
61 dominated studies of central Mexican palaeoclimatology, although recent speleothem  
62 records (Bernal *et al.*, 2011) have also made a contribution. Palynology has played a  
63 central role in palaeolimnological investigations dating back to the 1950s (Sears and  
64 Clisby, 1995; Watts and Bradbury, 1982; Goman and Byrne, 1998; Lozano-Garcia *et al.*,  
65 2005). In this context, the dominance of pine-oak woodlands at higher altitudes in the  
66 TMVB has presented further challenges given the limited taxonomic resolution of these  
67 pollen types in standard palynology and the impact of long-range dispersal, especially of  
68 pine pollen (Correa-Metrio *et al.*, 2012). The climatic interpretation of pollen diagrams  
69 from the TMVB has often hinged on decisions about the climatic conditions that changing  
70 proportions of pine and oak represent, or indeed whether the presence of large  
71 percentages of pine means that pine trees were actually present (c.f. Brown, 1985; Park *et*  
72 *al.*, 2010). Pollen records are also highly susceptible to anthropogenic disturbance,  
73 although the distinctive presence of *Zea mays* pollen in records is a clear indicator of  
74 agricultural activity close to lakes. To help to resolve some of the uncertainties around

75 pollen-based reconstructions, the application of palaeolimnological methods, particularly  
76 diatom, geochemical, isotopic and mineralogical analyses, has become increasingly  
77 common across the TMVB. Whilst not immune to many of the complicating factors  
78 outlined above, this multi-proxy approach has helped to improve our understanding of both  
79 climatic and environmental change in this area. In this paper we present a well-dated  
80 palaeolimnological record from a crater lake, La Piscina de Yuriria, on the southern edge  
81 of the Valle de Santiago region, southern Guanajuato, which extends back some 30,000  
82 years (all ages quoted in this paper that relate to the radiocarbon time-span are in  
83 calendar years unless otherwise stated). This is considered in the context of other data  
84 from adjacent crater lakes in the Valle de Santiago area and other lakes in the wider  
85 region to explore the timing and nature of climatic change and human impact.

86  
87 The present-day climate of the region, and climate forcing mechanisms over the late  
88 Quaternary, are complex. Central Mexico falls under the influence of the North American  
89 Monsoon (NAM). Variations in rainfall are forced by northern hemisphere summer  
90 insolation over the longer term (millennial timescales) linked to precession, which has  
91 caused changes in the position of the inter-tropical convergence zone (ITCZ) (Metcalf *et*  
92 *al.*, 2015). On shorter (centennial and shorter timescales), changes in the Pacific and  
93 Atlantic Oceans, both of which are important moisture sources, have influenced rainfall  
94 over tropical North America and Central America. During the last glacial and first half of the  
95 Holocene, the presence of the Laurentide Ice Sheet influenced the climate of the NAM  
96 region. The ice sheet had direct effects, through its impact on the position of the jet stream  
97 and the mid-latitude westerly winds, and indirectly via meltwater influx into the Gulf of  
98 Mexico, a significant moisture source (Aharon, 2003). As the influence of orbitally-forced  
99 summer insolation waned over the course of the Holocene, the effect of other factors such  
100 as sea-surface temperature variations in the Pacific and Atlantic, linked to various inter-

101 annual modes of ocean-atmosphere re-organisation, has become increasingly important  
102 (Metcalf *et al.*, 2015).

103

#### 104 **Study Region**

105 La Piscina de Yuriria lies within the Valle de Santiago, which is located at the northern  
106 edge of the Michoacán-Guanajuato Volcanic field (Fig. 1), and is distinctive because of the  
107 presence of at least seventeen maar type volcanoes, of which a number contained lakes  
108 (Aranda Gomez *et al.*, 2013). It lies on the margin of the volcanic uplands to the south and  
109 the lowlands of the Rio Lerma, in an area known as the Bajío, which became a very  
110 important agricultural region during the Spanish Colonial period (Butzer and Butzer, 1993).

111

112 Seven maar lakes have been identified around the town of Valle de Santiago, of which  
113 four contained water in 1900 (Ordonez, 1900). K-Ar dating of some of the maars puts their  
114 formation to between 1.2 Ma (Hoya San Nicolas) and 0.07 Ma (Hoya La Alberca).

115 Unfortunately, the maar that contains La Piscina de Yuriria was not dated, although the  
116 adjacent shield volcano is believed to date to 6.9 Ma (Aranda-Gomez *et al.*, 2013). It has  
117 been suggested that these maar lakes were once set within a large palaeolake, which  
118 extended from the modern Laguna de Yuriria, northwards around the modern town of  
119 Valle de Santiago. The maars have very small catchment areas and the lakes within them  
120 were supported by the waters of the underlying Salamanca aquifer. Unfortunately, this  
121 aquifer has been heavily exploited (there are more than 1600 active wells), which has  
122 resulted in a drawdown of about 2 m yr<sup>-1</sup> over the past 25 years (Alcocer *et al.*, 2000). As  
123 a result all the maars are now dry, even those that contained deep lakes in the early  
124 1980s. Alcocer *et al.* (2000) report on the many undesirable consequences of this  
125 desiccation, including the loss of endemic fish species and lake margin wetland habitats,  
126 the economic effect of the loss of fisheries and alkali fly collection and adverse health

127 effects due to the mobilisation of alkaline dust. La Piscina de Yuriria contained a lake  
128 approximately 2 m deep in 1981 and was then observed to dry out through the 1980s, with  
129 the development of a salt crust across the basin floor. Freshwater springs around the  
130 margins of the lake within the crater also dried up. A shallow lake was re-established in  
131 the early 1990s, which was then made permanent in the early 2000s by the pumping of  
132 groundwater back into the crater. As this was one of the basins where adverse health  
133 effects were reported due to dust mobilisation, this re-wetting of the basin may have been  
134 a response.

135

136 The Valle de Santiago area lies towards the northern margin of what was the  
137 MesoAmerican cultural area in the pre-Hispanic period. There is limited evidence for  
138 settlements during the Preclassic period associated with the Chupicuaro culture (ca. 800 –  
139 0 BC). Population expanded during the Classic period (ca. AD 300 – 900) when urban  
140 centres developed across the Bajío. In the late Postclassic (after ca. AD 1300) the area  
141 lay near the frontier between the settled Purépecha (Tarascans) to the south and the  
142 nomadic Chichimec to the north (Gorenstein and Pollard, 1983). A church (convento) was  
143 built at Yuririapundaro (the Purépecha name for Yuriria) in 1550. The translation of  
144 Yuririapundaro is ‘Lake of Blood’, referring to the distinct red colour of the water in the  
145 crater (de Escobar 1729 in Gomez de Orozco, 1972). A similar red colour, probably the  
146 result of blooms of sulphur bacteria, was observed at the lake in the 1980s. According to  
147 Park *et al.* (2010) the Spanish settled in Valle de Santiago in the early 17<sup>th</sup> century  
148 initiating a period of intensive agricultural exploitation that has lasted until the present day.

149

150 The dramatic effects of water extraction on the maar lakes attracted work on what had  
151 been the deep lakes of Hoya la Alberca and the Hoya Rincon de Parangueo (Kienel *et al.*,  
152 2009; Park *et al.*, 2010) partly in an effort to retrieve cores of laminated sediments before

153 these became lost through deflation or profoundly disturbed by desiccation and secondary  
154 precipitation of evaporite minerals. The Hoya Rincon de Paranguero record goes back to  
155 9600 aBP. Prior to this, there had been work on the Hoya San Nicolas (Brown, 1985;  
156 Metcalfe *et al.*, 1989), which had been cored in 1979 shortly after it dried out. The basal  
157 date on this core was 12,600-12,700 aBP. This maar was re-cored in 2001 with the results  
158 reported by Park *et al.* (2010) giving a record believed to extend back to 11,600 aBP (but  
159 not directly dated). As described further below, La Piscina de Yuriria was cored in 1981  
160 and 1982 prior to its desiccation.

161

162 The interpretation of sequences from the Valle de Santiago maar lakes has been subject  
163 to the common uncertainties that affect records from the TMVB region. The interpretation  
164 of pine pollen (or the ratio of pine to pine + oak) has been particularly significant here, with  
165 Park *et al.* (2010) rejecting Brown's earlier interpretation of wetter conditions between ca.  
166 5700 aBP and 3800 aBP in the Hoya San Nicolas. The shallower lakes also seem to have  
167 dried up quite regularly through the Holocene, possibly three times in the case of Hoya  
168 San Nicolas (Park *et al.*, 2010). The basic framework of change based on these earlier  
169 studies seems to be as follows: a cool and relatively moist late glacial (prior to ~12,700  
170 aBP); a variable late glacial to Holocene transition; a dry (possibly very dry) early  
171 Holocene; the rapid establishment of wetter conditions around 8400 aBP lasting until 5700  
172 aBP; dry 5700 – 3800 aBP, then wetter again, but not as wet as the period between  
173 around 8400 and 6000 aBP. The late Holocene has been profoundly influenced by human  
174 activity with evidence of maize cultivation and enhanced erosion (especially 2200 to 1300  
175 aBP). Both Metcalfe *et al.* (1989) and Park *et al.* (2010) report a cessation of human  
176 disturbance around 1000 aBP followed by renewed activity after 400 aBP probably  
177 associated with Spanish settlement. Our new data from La Piscina de Yuriria allow us

178 both to extend this record back into the last glacial and to test the framework outlined  
179 above.

180

## 181 **Study site**

182 This study is based on lake sediment records from La Piscina de Yuriria (20°30'N;  
183 101°08'W, 1740 m a.s.l), which is one of the small (area = 0.75 km<sup>2</sup>), hydrologically-closed  
184 maar lakes in the Valle de Santiago region (Fig. 1). The basaltic basin experiences a  
185 subhumid, subtropical climate with annual precipitation of 700-800 mm and supports  
186 subtropical thorn bush scrub around the lake and sparse oak woodland above 2200 m  
187 a.s.l. (Metcalf and Hales, 1994; Metcalfe *et al.*, 1994). The lake seems to have been  
188 generally shallow in recent times (e.g. ~ 2 m deep in 1981 and 1982, <1 m in 1992, >1.8  
189 m in 2004) and has dried up totally in some years (e.g.1989). The lake was saline as a  
190 result of evaporative enrichment, as well as highly eutrophic, with an alkalinity/Ca ratio >>1  
191 and Na-CO<sub>3</sub>-Cl-type composition indicating that evaporative evolution occurred along  
192 pathway IIIA of Eugster and Hardie (1978). The lake was fed by several circum-neutral to  
193 alkaline springs that were fresh to slightly brackish (Table S1) and also of Na-HCO<sub>3</sub> type.  
194 Evaporative enrichment is also reflected in the limited stable isotope data for the input  
195 water (-9.2 ‰) versus lake water (+0.7 ‰) (Table S1). Recent water-level changes have  
196 been the result of groundwater extraction for irrigation (e.g. Metcalfe and Hales, 1994) and  
197 later artificial recharge; variations during the late Quaternary, which form a major focus of  
198 the present study, have largely been driven by changes in effective moisture (precipitation  
199 minus evaporation, or P-E), as discussed below.

200

## 201 **Materials and methods**

### 202 *Field collection*



203 Two lake sediment cores, namely cores YC1 (length 4 m) and YC2 (14.3 m), were  
204 recovered from La Piscina de Yuriria from under 2 m of water in 1981 and 1982,  
205 respectively, by members of the Tropical Palaeoenvironments Research Group, at the  
206 time based at the University of Oxford, UK. Cores were initially wrapped in clingfilm and  
207 aluminium foil and stored in plastic tubes at 4 °C prior to sectioning into 1 cm slices,  
208 typically at 5 cm intervals, for analysis. Some of the data for core YC1 have been  
209 published previously (Metcalf and Hales, 1994; Metcalfe *et al.*, 1994) but are summarised  
210 here alongside previously unreported data for that core and for YC2.

211

### 212 *Physical sedimentology*

213 Sediment samples from YC1 and YC2 were analysed for organic carbon content using  
214 loss on ignition at 450 °C, and for carbonate content by calcimetry. Low-frequency  
215 magnetic susceptibility was measured using a Bartington MS1 magnetic susceptibility  
216 meter.

217

### 218 *Bulk sediment geochemistry*

219 Dried sediment samples from YC1 were digested using a combination of HNO<sub>3</sub>, HF and  
220 H<sub>2</sub>O<sub>2</sub> following Dean and Gorham (1976). The resulting residue was taken up in dilute HCl  
221 prior to analysis of major and minor metals using atomic absorption spectrophotometry  
222 (AAS). Samples from YC2 were prepared using the sequential digestion method of  
223 Engstrom and Wright (1984) and each separate fraction analyses for major and minor  
224 metals as for YC1. However, initial inspection of the results indicated that the fractionation  
225 had not worked well for these complex sediments. Consequently, the results for the  
226 separated fractions were summed and the data treated as 'bulk' analyses, as for YC1.  
227 Only selected metals are reported here, namely Fe, Mn, Al, and K, all of which are

228 abundant in catchment soils and sediments, and so are regarded as good tracers of  
229 inwash, as discussed in further detail below.

230

### 231 *Stable isotopes in endogenic carbonate*

232 Bulk, dried sediment samples from YC1 were sieved through an 80µm mesh to remove  
233 shell material, treated with Clorox to remove organic carbon and then dissolved in 100 %  
234 phosphoric acid. The evolved CO<sub>2</sub> was then analysed for oxygen and carbon isotopes  
235 using a VG 'Micromass' mass spectrometer at the Laboratoire d'Hydrologie et de  
236 Géochimie Isotopique, Université de Paris-Sud. Samples from YC2 were prepared in a  
237 similar way and analysed using a modified SIRA mass spectrometer at the University of  
238 Liverpool. All stable isotope results are reported in standard delta notation relative to the  
239 PDB standard.

240

241 As part of the evaluation of the stable isotope results, the mineralogy of the carbonate  
242 fractions was assessed by X-ray diffraction using Phillips PW1320/10 and PW1050 X-ray  
243 diffractometers. Specifically, the Mg content of calcite was assessed using the method of  
244 Goldsmith *et al.* (1961)

245

### 246 *Ostracods*

247 Bulk sediment samples were disaggregated in ~5 % H<sub>2</sub>O<sub>2</sub>, sieved through a 63 µm sieve  
248 and the coarse fractions used for extraction of ostracod shells under a low-power binocular  
249 microscope. Because ostracod abundance varied dramatically, in some samples all of the  
250 ostracod valves were picked whereas in richer samples, only the first 350 valves were  
251 picked and overall abundance then estimated based on the weight of sediment examined.  
252 However, the unpicked fraction of rich samples was inspected for the presence of rare  
253 species not identified in the original count. Despite poor ostracod shell preservation in

254 many parts of the core, sufficiently well-preserved valves were found at a number of levels  
255 to permit stable-isotope analyses, which were used to complement analyses on the  
256 endogenic carbonate.

257

258 Specimens of ostracods from the picked fraction of YC2 were selected and studied  
259 carefully for signs of damage, dissolution or replacement. Suitable individuals were then  
260 cleaned carefully using a 000 size clean nylon paintbrush and ultra pure deionised water  
261 under a binocular microscope to remove surface contaminants. Stable-isotope analyses of  
262 multiple shells of *Limnocythere sappaensis* were analysed for oxygen and carbon isotopes  
263 using a VG Isocarb coupled to a VG Optima mass spectrometer at the NERC Isotope  
264 Geosciences Laboratory, Keyworth (UK). Two samples, composed of up to 8 individual  
265 valves, were analyzed at most core levels selected and a weighted mean value for  $\delta^{18}\text{O}$   
266 and  $\delta^{13}\text{C}$  calculated for plotting purposes.

267

#### 268 *Diatoms*

269 Diatom samples were prepared for analysis at 10 cm intervals in YC1 and 10 – 20 cm  
270 intervals (depending on sample availability) in YC2. In YC2, only material from 4 m to the  
271 base was sampled in detail after preliminary analysis (samples prepared every 25 – 30  
272 cm) had shown that the upper 4 m of the sequence seemed to match well with YC1. Poor  
273 valve preservation between 7 m and 12 m meant that the majority of samples were taken  
274 below and above this interval. Core samples were complemented by a surface sediment  
275 sample and water samples from the lake margin. Samples were treated to remove  
276 carbonates and organic matter following the method of Battarbee (1986). In some cases  
277 strong acids ( $\text{H}_2\text{SO}_4$  or  $\text{HNO}_3$ ) were used to break up clumps of sediment. The final  
278 suspension was mounted onto coverslips using Naphrax. The same method was used to  
279 prepare surface sediment samples. Where possible, 400 valves were counted from each

280 level, although at levels with poor preservation this was sometimes reduced to 100 – 200  
281 valves. Identifications were carried out using standard floras including Gasse (1986),  
282 Krammer and Lange Bertalot (1988, 1991a and b), Patrick and Reimer (1966, 1975),  
283 Schoeman and Archibald (1977). The separation of *Navicula (Craticula) elkab* from  
284 *Craticula halophila* proved to be an important part of the study and is discussed in  
285 Metcalfe (1990). The results from YC1 were published in Metcalfe and Hales (1994), and  
286 preliminary results from YC2 were described in Park (1999). Diatom results are presented  
287 here as percentages of the full counts. Species present at less than 2% and only  
288 occurring in one sample are not plotted here.

289

290 Initial zonation was carried out using CONISS, the stratigraphically constrained cluster-  
291 analysis program within Tilia (Grimm, 1987). The counts from the core samples were also  
292 analysed using CANOCO and TWINSPAN. TWINSPAN (Hill, 1979; Jongman *et al.*, 1992)  
293 provided a clustering that was not constrained stratigraphically. CANOCO (ter Braak,  
294 1988) was used to explore further the variation between samples and species. The latter  
295 analyses were used to determine whether the lake diatom assemblages showed repeat  
296 occurrences through time and to explore that trajectory of change.

297

### 298 *Chronology*

299 Independent chronologies were established using AMS or radiometric radiocarbon dating  
300 on bulk organic carbon or carbonate (8 levels were dated from YC1 and 16 from YC2).  
301 Lack of appropriate material precluded the dating of terrestrial plant macrofossils from  
302 either core, but the absence of carbonate bedrock within the catchment means that  
303 hardwater error is unlikely to affect radiocarbon dates at this site. Radiocarbon  
304 measurements were performed at the Oxford University Radiocarbon Accelerator Unit,  
305 Simon Fraser University Radiocarbon facility, Laboratoire d'Hydrologie et de Géochimie

306 Isotopique, Université de Paris-Sud, the NERC Radiocarbon Facility, East Kilbride and the  
307 Arizona AMS Laboratory, University of Arizona. Dates were calibrated using IntCal13  
308 (Reimer *et al.*, 2013). For each core, outliers were identified and excluded from  
309 subsequent age modeling, which was undertaken using Clam 2.2 (Blaauw, 2010).  
310 Although some of the radiocarbon dates have been published previously, we report them  
311 again here for the sake of completeness and because we have revised the calibrations.

312

## 313 **Results**

### 314 *Chronology*

315 The age model for YC1 is based on 7 of the 8 radiocarbon dates for that core (Table 1,  
316 Fig. 2) and the age-depth pairs well described by a 3<sup>rd</sup>-order polynomial curve. The age  
317 model for YC2 is based on 12 of the 16 radiocarbon dates (Table 1, Fig. 2) and the age-  
318 depth pairs well described by a 3<sup>rd</sup>-order polynomial curve. The age models were used to  
319 assign ages to each depth within the respective cores. An assessment of the comparability  
320 of the independent age models for YC1 and YC2 was also undertaken by comparing  
321 variations in their physical sedimentology, which would be expected to change  
322 synchronously in the two cores: this assessment is discussed below.

323

### 324 *Physical sedimentology*

325 Cores YC1 and YC2 core are both composed predominantly of brown to grey,  
326 diatomaceous gyttja. In the lower sections of YC2 (at ~26,300, 19,900 and 8300 aBP),  
327 centimetre-scale sand layers are present. Desiccation surfaces, represented by sub-  
328 vertical fractures infilled with darker material, occur at several depths in the interval  
329 between 16,800 and 14,300 and near the top of the sequence (~800 aBP). Nodular,  
330 siliceous concretions were observed between the base of the sequence and ~17,500 aBP.  
331 The interval 100 – 50 cm (representing ~1100 – 500 aBP) was not recovered in YC2.

332

333 Organic carbon content, estimated from loss-on-ignition (LOI) at 450°C, varies between ~9  
334 and 44 % (Fig. 3). From the beginning of the YC2 record until ~14,000 aBP, values  
335 average ~15%, increasing and becoming more variable after this time, with peaks up to 44  
336 %. The LOI record in the upper part of YC2 is well replicated in YC1. Carbonate content  
337 averages around 18 % from the beginning of the record until 14,200 aBP, but with peaks  
338 up to 40 % in places. Between 14,000 and 7800 aBP, carbonate content falls to 14 % and  
339 becomes less variable. From 7800 aBP to the end of the YC2 record, there is a sharp  
340 increase to an average of 33 %, with transient peaks up to 45 %. The pattern of change in  
341 YC1 replicates that in the upper part of YC2, but values are systematically lower.

342

343 Magnetic susceptibility ( $\chi$ ) values (Fig. 3) are typically lower than  $10 \times 10^{-7} \text{m}^3 \text{kg}^{-1}$  but with a  
344 number of sharp peaks at several levels and a more gradual increase and then decline  
345 between ~10,100 and 6900 aBP. Values rise sharply at 3600 aBP, after which they remain  
346 high and show increased variability. The pattern of change in YC1 replicates that in the  
347 upper part of YC2.

348

#### 349 *Geochemistry and stable isotopes*

350 Concentrations of K, Fe, Mn and Al are variable throughout the cores. For much of the  
351 lower part of YC2 values are generally low, although there are some transient peaks.

352 Values typically become higher and more variable after about 5100 aBP in YC2, a pattern  
353 that is generally replicated in YC1. There is strong and statistically-significant correlation  
354 amongst all of the metals discussed here in YC1 and in the section of YC2 covering the  
355 past ~5000 years. Correlations are weaker for the earlier part of YC2 (Table S2).

356

357 Stable isotope values are available for fine-grained carbonates and, for a few levels from  
358 YC2 only, for ostracod shells (Fig. 4). Mineralogical determinations showed that the calcite  
359 present within the core had low magnesium content ( $\leq 4$  mole %): nevertheless,  
360 appropriate corrections were made to the  $\delta^{18}\text{O}$  values following Tarutani *et al.* (1969).  
361 From the beginning of YC2 to 26,600 aBP,  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values are typically low but  
362 variable (from about 0 ‰ to around -11 ‰ and +7.6 to -1 ‰, for  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$ ,  
363 respectively) and strongly covariant (Fig. 5). Between 25,000 and 14,200 aBP, the stable-  
364 isotope values are at relatively low resolution, and typically values are higher than in the  
365 preceding interval (up to about +3 ‰ and +8 ‰ for  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$ , respectively) and  
366 covariant. Between 10,900 and 6600 aBP, there is a further rise in variability, but  $\delta^{18}\text{O}$  and  
367  $\delta^{13}\text{C}$  covary as before. Between 6600 and 4500 aBP, there is a significant shift in the  
368 pattern of stable isotope variability, with  $\delta^{18}\text{O}$  values remaining relatively constant, but  $\delta^{13}\text{C}$   
369 increasing up to +15.4 ‰. After 4500 aBP,  $\delta^{18}\text{O}$  values vary between about 0 and -2.7 ‰  
370 and  $\delta^{13}\text{C}$  values are markedly reduced compared with the preceding interval. The  
371 overlapping sections of YC1 and YC2 show similar patterns of change, especially for  $\delta^{13}\text{C}$ .  
372 The low-resolution ostracod values are typically  $^{18}\text{O}$ -enriched and  $^{13}\text{C}$ -depleted compared  
373 with those for fine-grained carbonates.

374

### 375 *Ostracods*

376 Ostracod abundance varies dramatically, between zero (i.e. barren levels) and ~35,000  
377 valves per gram of sediment (Fig. 4). Significant zones of very high abundance are  
378 present from the beginning of the YC2 record to until 28,700 aBP and at 21,900 – 21,300  
379 aBP, 17,800 – 15,900 aBP, 13,700 – 12,900 aBP, 4800 - 2700 aBP, 2000 – 1100 aBP.  
380 Significant zones that are barren or contain very few ostracods occur at 28,500 – 27,800  
381 aBP, 19,000 – 18,500 aBP, 15,400 – 14,400 aBP, 10,500 – 8300 aBP, 5300- 4900 aBP.  
382 The zones of high abundance generally contain monospecific assemblages of

383 *Limnocythere sappaensis* (both males and females): zones of lower abundance typically  
384 support other species as well, including at least one species of candonid, *Heterocypris* sp.,  
385 *Physocypris* sp., *Potamocypris* sp., *Strandesia* sp. and *Cyprina* sp. Although *L. sappaensis*  
386 is regarded as conspecific with *Limnocythere inopinata* (e.g. Martens, 1994), we adopt the  
387 name *L. sappaensis* here since this is widely used for the species in North America.

388

### 389 *Diatoms*

390 The diatom assemblages from surface sediment samples collected in 2 m of water in 1982  
391 were dominated by *Navicula (Craticula) elkab* (22%), *N. (C.) halophila* (13%) and a variety  
392 of *Nitzschia* species. A sample taken in 1997 in 0.3 m of water, just as a lake was  
393 becoming re-established, was dominated by *Anomoeoneis sphaerophora* (48%) and  
394 *Navicula (Craticula) elkab* (16%). *Anomoeoneis costata* and *Chaetoceros mulleri* spores  
395 made up 8% of the count each (Davies, 2000). A sample taken in 2004, at the edge of the  
396 lake, was dominated by *Navicula (Craticula) halophila* (56%), a range of *Nitzschia* species,  
397 *Chaetoceros muelleri* (some as resting spores) and *Anomoeoneis sphaerophora* f.  
398 *costata* (Hill, 2006), although these each formed less than 6% of the assemblage. In 2004,  
399 the lake was hypereutrophic, with TP = 584  $\mu\text{g l}^{-1}$  and chlorophyll-a 127  $\mu\text{g l}^{-1}$  (Hill, 2006).

400

401 Preliminary diatom analysis of samples from the top 4 m (past ~5000 years) of the YC2  
402 core indicated the presence of similar assemblages to those reported in Metcalfe and  
403 Hales (1994), so efforts on this core were focused on the section below 4 m. Here, results  
404 are presented from both core sequences (Fig. 6). As with the ostracods, diatom  
405 preservation is highly variable through the sequences. In YC2 there was little or no valve  
406 preservation between 22,100 – 10,400 aBP cm. In YC1 there are gaps in the record due  
407 to poor preservation between 2000 – 1100 aBP. Both sequences are dominated by *N. (C)*  
408 *halophila* and *N. (C) elkab*, *A. costata* and *A. sphaerophora*, *C. muelleri* and a range of



409 *Nitzschia* species. These assemblages are all similar to those found in surface sediment  
410 samples taken from the lake at different times, reflecting high (if varying) pH and alkalinity,  
411 and highly evaporated conditions. A more distinctive aspect of the core records is the  
412 abundance, at times, of small *Navicula* species (e.g. *N. fluens*, *N. minusculoides*, *N.*  
413 *muralis*).

414

## 415 **Discussion**

416 We discuss the chronology for the lake-sediment sequences and then the interpretation of  
417 each of the palaeolimnological variables, before proceeding to reconstruct the  
418 palaeolimnological history of La Piscina de Yuria for the late Quaternary.

419

420 Core YC1 covers the interval from about 4600 aBP to the coring date. The anomalous age  
421 at 44-46 cm probably represents inwash of older carbon from the catchment, substantiated  
422 by the magnetic susceptibility data. Core YC2 covers the interval from about 31,000 aBP  
423 to the coring date. The four dates that are omitted from the age model are all younger than  
424 expected, possibly the result of root penetration during times of low lake level. The general  
425 pattern of age-depth relationships for YC1 and the upper part of YC2 suggests good  
426 agreement between the two cores. Detailed comparisons based on the loss on ignition  
427 and magnetic susceptibility profiles (Fig. 2), which would be expected to agree for the two  
428 closely-located sequences, suggest a small (~200 years) age offset, with YC2 being  
429 consistently older. The fact that small differences exist is perhaps not surprising given that  
430 different equivalent depths and contrasting materials were dated in the two cores (Table  
431 1). However, rather than correct either one of the cores we prefer to use the age models  
432 defined for each respective core and then refer to the resulting uncertainties as  
433 appropriate.

434

435 Loss on ignition (LOI) provides a good proxy for the organic carbon content of the  
436 sediments (Dean, 1974). The carbon/nitrogen (C/N) ratios (available for YC1 only: data not  
437 shown, but values vary between 11.6 and 49.4) indicate that the organic matter is of mixed  
438 aquatic and terrestrial sources, suggesting that LOI provides a record of aquatic  
439 productivity and terrestrial inwash at least for the past 4,600 years. The CaCO<sub>3</sub> content of  
440 the sediment is best explained by endogenic carbonate formation within the lake, since the  
441 catchment is devoid of carbonate rocks or sediments. In such settings, calcium carbonate  
442 precipitates from the water column when the lake becomes saturated with respect to  
443 carbon minerals as a result of evaporative enrichment of water or mediated by aquatic  
444 plants. However, enhanced aquatic productivity can also lead to carbonate dissolution  
445 (e.g. Cohen, 2003) and carbonate formation also depends on supply of ions from the  
446 catchment, meaning that the interpretation of sedimentary CaCO<sub>3</sub> records is not always  
447 straightforward.

448

449 The magnetic susceptibility of the sediments is strongly linked to inwash because the  
450 volcanic soils are rich in magnetic minerals. Increases in  $\chi$  in YC1 have therefore been  
451 interpreted as inwash events associated with either natural or anthropogenic catchment  
452 disturbance (Metcalf and Hales, 1994). The elements Fe, Mn, Al and K are all associated  
453 with weathered volcanic soils, and their concentrations in the lake sediments are  
454 controlled by catchment inwash: Fe and Mn may also have been mediated by redox  
455 conditions within the lake although we do not have direct evidence for this.

456

457 The oxygen and carbon isotope composition of lacustrine carbonate is a function of the  
458 temperature and isotopic composition of the water in the case of oxygen, and the carbon-  
459 isotope composition of dissolved inorganic carbon (DIC) for carbon. In subtropical dryland  
460 lakes, such as La Piscina de Yuriria, the oxygen-isotope ratio of lake water is usually the

461 dominant control on the oxygen-isotope composition of carbonate, and this varies with the  
462 degree of evaporative enrichment (Talbot, 1990). The carbon-isotope composition of DIC  
463 is a complex function of carbon source (catchment- vs lake-derived) and in-lake  
464 modification as a result of fractionation during DIC uptake by aquatic plants for  
465 photosynthesis, and exchange with atmospheric CO<sub>2</sub> (Talbot, 1990) The same general  
466 controls determine the isotopic values of biogenic carbonates, such as ostracod shells, but  
467 there may be taxon-specific differences in the exact location and timing (especially  
468 season) of carbonate formation compared with endogenic carbonate (e.g. Decrouy *et al.*,  
469 2011). Moreover, biogenic carbonate may not be precipitated in isotopic equilibrium with  
470 lake water or DIC: ostracod shells, for example, demonstrate offsets from oxygen-isotope  
471 equilibrium and are typically <sup>18</sup>O-enriched compared with endogenic carbonate  
472 precipitated in equilibrium with lake water (von Grafenstein *et al.*, 1999). Further  
473 palaeolimnological inferences can be drawn for the strength of covariance amongst  
474 carbonate oxygen and carbon isotope values in sediment sequences, strong covariance  
475 typically being associated with hydrologically-closed systems (Talbot, 1990).

476  
477 Although the occurrence and abundance of different lacustrine ostracod taxa are  
478 determined by a range of factors, in La Piscina de Yuriria, salinity, hydrochemistry and the  
479 extent to which the lake is seasonally permanent are likely to be the dominant controls. In  
480 saline lakes, there is moreover often a relationship between species diversity and ostracod  
481 abundance: highly saline lakes are typically dominated by a single species that is present  
482 in very high abundance (De Deckker and Forester, 1988).

483  
484 Habitat, pH, conductivity, ionic composition and nutrient levels are all major controls on the  
485 presence and abundance of diatom species, with some species having well established  
486 preferences in relation to some, or all, of these factors (e.g. Gasse, 1986; Kilham *et al.*,

487 1986). In La Piscina de Yuriria, the dominant controls over the full record appear to be  
488 those associated with changes in evaporative concentration (pH, EC, ionic composition).  
489 NaCl and Na<sub>2</sub>CO<sub>3</sub> waters have been shown to be particularly aggressive in relation to  
490 diatom dissolution (Barker *et al.*, 1994), so it is likely that some species may be over-  
491 represented in the sediment record due to their robust form and heavy silicification. This  
492 seems particularly likely in the case of the cysts of *C. muelleri* and may also apply to the  
493 more robust forms of species such as *A. costata*, *Denticula elegans* and *Rhopalodia*  
494 *gibberula*. Although the impact of differential preservation needs to be borne in mind when  
495 interpreting the fossil record, saline-lake diatoms are well established as indirect tracers of  
496 climate change (Gasse *et al.*, 1997).

497

498 From the base of YC2, which dates to a little before 30,000 aBP, until around 27,500 aBP,  
499 the  $\delta^{18}\text{O}_{\text{carb}}$  record shows marked variability, with the lowest values equivalent to the  
500 minimum for the sequence as a whole, and the highest values close to the maximum (Fig.  
501 4). There is strong covariance between  $\delta^{18}\text{O}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{carb}}$  (Fig. 5) consistent with a  
502 hydrologically-closed lake undergoing temporal variations in the degree of evaporative  
503 enrichment (Talbot, 1990) in response to changes in effective moisture. The most negative  
504  $\delta^{18}\text{O}_{\text{carb}}$  values equate to un-evolved lake water that was probably fed by springs and  
505 rainfall with lower  $\delta^{18}\text{O}$  than at present ( $\delta^{18}\text{O}$  around -9 ‰ in summer 1992: Table S1),  
506 possibly coupled with cooler conditions. In contrast, the most positive  $\delta^{18}\text{O}_{\text{carb}}$  values in this  
507 interval are best explained by evaporative enrichment under reduced effective moisture.  
508 The lowest  $\delta^{13}\text{C}_{\text{carb}}$  can be explained by equilibration with atmospheric CO<sub>2</sub> whereas the  
509 more positive values require other, or additional, mechanisms to explain them. The uptake  
510 of <sup>12</sup>C during aquatic photosynthesis by aquatic macrophytes or algae can lead to an  
511 increase in the  $\delta^{13}\text{C}_{\text{DIC}}$  and hence of endogenic carbonates, but such an explanation  
512 appears inconsistent with low TOC values in this interval (Fig. 3). An alternative

513 explanation is the formation of co-genetic,  $^{13}\text{C}$ -enriched,  $\text{CO}_2$  during methane formation.  
514 Despite the evidence pointing to low lake levels at this time, methane formation in shallow  
515 and eutrophic lakes has previously been reported in Mexico (Lake Pátzcuaro: Metcalfe *et*  
516 *al.*, 2007) and elsewhere (e.g. Lamb *et al.*, 2000; Gu *et al.*, 2004). The high concentrations  
517 of *L. sappaensis* in this interval are consistent with the existence of a saline-alkaline lake  
518 (Forester, 1986): interestingly, the peaks in ostracod abundance coincide broadly with the  
519 peaks in  $\delta^{18}\text{O}_{\text{carb}}$ , suggesting that ostracod numbers increased with salinity, and hence  
520 evaporative enrichment. The occurrence of siliceous nodules within the interval,  
521 associated with the levels that have the highest  $\delta^{18}\text{O}_{\text{carb}}$  values, is also consistent with the  
522 existence of highly saline and alkaline water. The sporadic isotope values from ostracod  
523 shells in this interval show  $^{18}\text{O}$ -enrichment compared with endogenic carbonates that is  
524 broadly consistent with the approximate +0.7 ‰ vital offset recorded for the genus  
525 *Limnocythere* (von Grafenstein *et al.*, 1999). In contrast, the carbon isotope values in  
526 ostracod shells are similar to those in endogenic carbonate, suggesting that both sources  
527 of carbonate precipitated from DIC with a similar  $\delta^{13}\text{C}$  value once allowance has been  
528 made for differences in the timing and exact location of formation. This period covers  
529 diatom zones YC2-I and part of YC2-II (Fig. 6). The base of the core is dominated by *C.*  
530 *muelleri*, a diatom known to inhabit chloride-rich waters, but the other taxa here do not  
531 indicate hypersaline conditions, so it may be over-represented. The most negative  $\delta^{18}\text{O}_{\text{carb}}$   
532 value may be reflected in the increase in the freshwater *N. molestiformis* and *N. fluens*  
533 although the diatom assemblage overall continues to indicate shallow and alkaline  
534 conditions. The presence of *Nitzschia palea* and a form of *Nitzschia frustulum* (both  
535 obligate N heterotrophs) also indicates eutrophic conditions, which may help to explain  
536 methanogenesis in shallow water conditions (see above). The increasing abundance of *N.*  
537 *elkab*, and species of *Anomoeoneis* indicate more consistently high pH (> 8.5) and  
538 alkalinity, probably associated with shallowing of the lake. Low magnetic susceptibility and

539 low concentrations of 'inwash' indicator elements (Fe, Mn, K and Al) in this interval  
540 suggest that inwash of soil into the lake was limited despite the intervals of increased  
541 effective moisture (Fig. 3).

542

543 Between ~27,500 and 14,000 aBP, there was a shift to more positive  $\delta^{18}\text{O}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{carb}}$   
544 values, although with some stratigraphical variability and strong covariance between  
545  $\delta^{18}\text{O}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{carb}}$  (Fig.4, Fig. 5). The elevated  $\delta^{18}\text{O}_{\text{carb}}$  values suggest enhanced  
546 evaporative enrichment of lake water under conditions of low effective moisture: the  
547 presence of siliceous nodules at various points in this interval support the argument that  
548 the lake was shallow, saline and strongly evaporated. Short-term variations in carbonate  
549 content also support the occurrence of strong but variable evaporative enrichment (Fig. 3).  
550 The presence of multiple surfaces that probably resulted from desiccation in the later part  
551 of this interval suggests that lake levels fell and the lake may have dried out totally on  
552 several occasions. Elevated  $\delta^{13}\text{C}_{\text{carb}}$  values are too high to be explained solely by  
553 equilibration with atmospheric  $\text{CO}_2$ . Enhanced aquatic productivity, in which  $^{12}\text{C}$ -uptake by  
554 aquatic plants and algae causes DIC to be enriched in  $^{13}\text{C}$ , is incompatible with the low  
555 TOC content in this interval: the production of  $^{13}\text{C}$ -enriched co-genetic  $\text{CO}_2$  in a stagnant,  
556 shallow, nutrient-rich lake could provide an alternative explanation as discussed above.

557 From 27,500 to 22,500 aBP, a period of evaporative enrichment and periodic desiccation  
558 is consistent with the diatom record for this interval (zones YC2-IIa and YC2-IIb), which  
559 ends around 22,500 aBP in a period when diatoms were sparse and poorly preserved  
560 (after which there is a break in diatom preservation, see above). There are two notable  
561 peaks in *N. minusculoides* reaching 71% of the count between 27,000 and 26,000 aBP  
562 and 44% at around 24,000 aBP. The earlier peak is associated with a layer of sand (or  
563 tephra, see below) and, moreover, there are no stable isotope data from this layer, so its  
564 significance remains unclear. The presence of some freshwater taxa (e.g. *N. fluens*,

565 *Caloneis bacillum*) may reflect fluctuating conditions within a period of overall drying.  
566 There are several marked peaks in ostracod abundance during this interval, with  
567 assemblages strongly dominated by *Limnocythere sappausensis*, which supports the  
568 inference that the lake was generally saline and alkaline. The sporadic occurrence of other  
569 taxa suggests the periodic influx of fresher waters: rather than whole-lake freshening,  
570 these taxa could indicate surface or subsurface inflow of fresh water at various times. Low  
571 magnetic susceptibility values and low concentrations of 'inwash' elements suggest limited  
572 transfer of soil or sediment from the catchment during this interval; two sharp peaks in  
573 magnetic susceptibility around 26,000 and 23,000 aBP (Fig. 3) may represent tephra,  
574 rather than catchment inwash, although this remains to be confirmed.

575

576 There is a gap in the stable isotope record between 14,000 and 11,000 aBP, but evidence  
577 for dry conditions during this interval comes from the presence of a desiccation surface  
578 and the occurrence of a possible palaeosol. It is also supported by the lack of diatom  
579 preservation. Sporadic ostracod occurrence also suggests that the lake may have been  
580 ephemeral during this time. This interval covers the northern hemisphere late glacial  
581 stadial event and confirms that this was a time of low effective moisture in central Mexico.

582

583 After 11,000 aBP, there was a shift to more negative and also more variable  $\delta^{18}\text{O}_{\text{carb}}$  and  
584  $\delta^{13}\text{C}_{\text{carb}}$  values, consistent with a general increase in effective moisture during the early  
585 Holocene (Fig. 4). The rise in magnetic susceptibility, coupled with minor increases in  
586 some of the 'inwash' elements, may also reflect enhanced inwash of catchment material  
587 during the early Holocene. Diatom preservation resumes (zone YC2-III), with assemblages  
588 dominated by *Nitzschia palea* and *Chaetoceros muelleri* (Fig. 6). Assuming that *C.*  
589 *muelleri* may be overrepresented, high percentages of *N. palea* and *Nitzschia communis*  
590 indicate lower pH and TDS (total dissolved solids) (Gasse, 1986) and eutrophic conditions.

591 This assemblage shows some similarities to that in zone YC2-Ia, but zone YC2-III may  
592 represent the period when the lake was freshest and deepest, although not deep enough  
593 to develop a truly planktonic flora. Today, assemblages with such high percentages of *N.*  
594 *palea* are found in shallow, freshwater lakes in Mexico with high levels of nutrient  
595 enrichment such as Lakes Zacapu and Cajititlan (Hill, 2006 and S. Metcalfe, unpublished  
596 data). Ostracod concentrations are low, but significantly the assemblages include a  
597 relatively high proportion of species other than *L. sappaensis*, consistent with fresher  
598 water than in much of the pre-Holocene.

599

600 Between about 8,000 and 4,500 aBP there was a marked change in the lake system.  
601 Increase in  $\delta^{18}\text{O}_{\text{carb}}$  values suggests enhanced evaporative enrichment associated with  
602 decreased effective moisture (Fig. 4). Most notably, however, is the dramatic positive  
603 excursion in  $\delta^{13}\text{C}_{\text{carb}}$  values, up to a maximum of about +16 ‰ and, associated with this,  
604 the breakdown in the positive covariance between  $\delta^{18}\text{O}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{carb}}$  values that was  
605 apparent during earlier intervals (Fig. 5). The very high  $\delta^{13}\text{C}_{\text{carb}}$  values are best explained  
606 by methanogenesis, as discussed for earlier intervals above. Interestingly, the single  
607  $\delta^{13}\text{C}_{\text{ostracod}}$  value from this interval does not track the  $\delta^{13}\text{C}_{\text{carb}}$  values, but instead is much  
608 lower (Fig. 4). Although we cannot attach too much significance to a single value, this  
609 does suggest that the ostracods and the endogenic carbonate were formed in different  
610 micro-environments within the lake, or perhaps during different seasons, from DIC with  
611 contrasting  $\delta^{13}\text{C}$  values. Although the  $\delta^{18}\text{O}_{\text{ostracod}}$  values are more positive than the  $\delta^{18}\text{O}_{\text{carb}}$   
612 values, as would be expected, the difference is too large to be explained by vital offsets  
613 alone, possibly lending support to the view that the endogenic and ostracod carbonates  
614 were formed under contrasting conditions or at different times of the year. Ostracods occur  
615 sporadically in this interval, indicating that whatever conditions prevailed were not wholly  
616 unsuitable for ostracods to live. This period straddles the diatom record at the top of YC2



617 (zone YC2-IV) and the bottom of YC1 (zone YC1-I). The diatom assemblage is notable for  
618 its dominance by *N. (C.) elkab* and *N. (C.) halophila* (both cores), with *Nitzschia frustulum*  
619 and *N. palea* (Fig. 6). The switch to an *N. elkab/N. halophila* flora is consistent with a  
620 return to more alkaline conditions, probably driven by increasing evaporation. *Navicula*  
621 (*C.*) *elkab* does seem to have a distinct ecology, with a preference for hyper-alkaline,  
622 Na<sub>2</sub>CO<sub>3</sub> lakes, where Cl<sup>-</sup> is also important. *Nitzschia frustulum* tends to be more abundant  
623 with *N. (C.) elkab* than with *N. (C.) halophila* (e.g. YC2-4c and base of YC1-1) supporting  
624 the interpretation of high alkalinity (Gasse, 1986). Its presence with *N. palea* again seems  
625 to indicate high levels of nutrient enrichment. Overall, this assemblage is quite similar to  
626 that sampled from the modern lake in 1982, when it was around 2 metres deep. A peak  
627 in magnetic susceptibility between about 5,500 and 4,500 BP (Fig. 3) associated with a  
628 sharp reduction in  $\delta^{18}\text{O}_{\text{carb}}$  (Fig. 4) points to a climatically-controlled inwash event, which  
629 may be reflected by rather poor diatom preservation in YC1.

630

631 After 4,500 aBP,  $\delta^{18}\text{O}_{\text{carb}}$  remained high although with some short-lived negative  
632 excursions, whereas  $\delta^{13}\text{C}_{\text{carb}}$  values are reduced dramatically (Fig. 4). For the interval of  
633 overlap between YC1 and YC2, there is good agreement between the isotope records,  
634 especially so for  $\delta^{13}\text{C}_{\text{carb}}$ , once allowance is made for the slight age difference between the  
635 two cores, as discussed earlier. There is a large increase in ostracod concentration within  
636 much of this interval: the dominance of assemblages by *L. sappaensis*, coupled with the  
637 large number of individuals, is consistent with the lake having been saline and alkaline for  
638 much of the time (Fig. 4).  $\delta^{18}\text{O}_{\text{ostracod}}$  values are <sup>18</sup>O-enriched compared to  $\delta^{18}\text{O}_{\text{carb}}$  values,  
639 with the degree of enrichment consistent with vital offsets from oxygen-isotope equilibrium,  
640 as discussed above. The  $\delta^{13}\text{C}_{\text{ostracod}}$  values agree well with those for  $\delta^{13}\text{C}_{\text{carb}}$ , suggesting  
641 that the two sources of carbonate formed under the same set of conditions, as was the  
642 case for the interval prior to the positive  $\delta^{13}\text{C}_{\text{carb}}$  excursion. The diatom record for this

643 interval was published by Metcalfe and Hales (1994), although here we plot the data  
644 against our new age model (Fig. 6). The species encountered confirm the presence of a  
645 shallow alkaline lake throughout the late Holocene, although the balance between  $\text{CO}_3^{2-}$   
646 and  $\text{Cl}^-$  seems to have varied. The assemblage in zone YC1-II, for example, may indicate  
647 that  $\text{Cl}^-$  replaced  $\text{CO}_3^{2-}$  as the dominant anion. The assemblage is similar to Bradbury's  
648 (1989) saline marsh group. Fresher conditions were then re-established, associated with  
649 inwash from the catchment. This wetter/drier cycle is then repeated in zones YC1-III and  
650 YC1-IV. Increasingly hostile conditions for diatom preservation are indicated through zone  
651 YC1-V, with only a patchy diatom record between about 2000 and 1200 aBP. It seems  
652 likely that the single count available through this period is significantly affected by  
653 differential preservation. The diatom record resumes around 1000 aBP with a distinctive,  
654 well-preserved sample dominated (55%) by *Navicula muralis* (zone YC1-VII). This diatom  
655 is often found on mud flats and amongst aquatic vegetation (Hustedt, 1961-66). When  
656 combined with very low magnetic susceptibility values, the diatom assemblages indicate  
657 catchment stability. We note that this catchment stability occurs at a time in the late  
658 Classic when many sites in the relatively dry parts of Central Mexico were abandoned  
659 (Beekman, 2010; Park *et al.*, 2010). The most recent sediments preserve a flora quite  
660 similar to that found in the various surface sediment samples, indicating a shallow alkaline  
661 lake, but with increasing nutrient levels. The very highly evolved chemistry and rather  
662 distinctive diatom flora of La Piscina de Yuriria is described in Davies *et al.* (2002) and  
663 sampling in 2003 and 2004 confirmed its hypereutrophic status. Phases of inwash,  
664 previously reported in YC1 and attributed to anthropogenic disturbance (Metcalfe *et al.*,  
665 1994), are also seen at broadly the same times in YC2, i.e. around 3,600 and 1500 aBP  
666 and over the past few centuries. An earlier phase at the base of YC1, which ended in that  
667 core around 4700 aBP, appears in its entirety in YC2, starting around 5600 aBP. The  
668 absence of *Z. mays* pollen from YC1 during this interval led Metcalfe *et al.* (1994) to

669 suggest that the inwash was climatically-mediated rather than the result of anthropogenic  
670 disturbance. For the wider region however, Park *et al.* (2010) suggest that agricultural  
671 activity began as early as around 5700 aBP (based on the presence of *Zea* pollen), and  
672 expanded around 3000 aBP. Lozano *et al.* (2013) also report the occurrence of *Z. mays*  
673 pollen from 3000 aBP in Lake Zirahuen, a lake previously thought to be un-affected by  
674 human impact. The main climatic and environmental changes revealed in the records from  
675 La Piscina de Yuriria are summarised in Table 2.

676

677 There is a reasonable correspondence between phases of inwash, as indicated by  
678 magnetic susceptibility and 'inwash' elements, and intervals of low ostracod abundance for  
679 the whole of the Holocene (Fig. 7). This suggests that increased turbidity in the lake, which  
680 would likely have arisen during phases of increased inwash, was unfavourable for  
681 ostracod survival as has been noted previously (e.g. Bridgwater *et al.*, 1999). Therefore  
682 water turbidity is an additional control to hydrochemistry on ostracod assemblages in La  
683 Piscina de Yuriria.

684

685 The number of palaeoclimatic records from the TMVB (and indeed the whole of Mexico),  
686 extending back to 30,000 BP (Marine Isotope Stage 3) is small and limited to the Basin of  
687 Mexico (e.g. Caballero and Ortega-Guerrero, 1998; Roy *et al.*, 2009; Lozano-García *et al.*,  
688 2015), Lake Cuitzeo (Israde *et al.*, 2010), Lake Pátzcuaro (Watts and Bradbury, 1982;  
689 Bradbury 2000) and Lake Zacapu (Correa-Metrio *et al.*, 2012). Moreover, most of these  
690 have limited dating control for the older sediments. Caballero *et al.* (2010) integrated  
691 palaeolimnological records with evidence from glacial chronologies (Vazquez Selem and  
692 Heine, 2004) to provide a palaeoclimatic scenario for the period from around 30,000 aBP  
693 to the last glacial maximum (LGM). A more recent consideration of the record from Lake  
694 Chalco in the Basin of Mexico has been published by Lozano-Garcia *et al.* (2015). With

695 the exception of Lake Pátzcuaro (one of the westernmost sites), the records suggest  
696 drying and cooling after about 30,000 aBP, with glacial advances restricted until around  
697 22,000 aBP. The data from La Piscina de Yuriria suggest that there were rapid shifts  
698 between wet and dry conditions during this interval, with a general drying trend sometime  
699 between about 27,500 and 25,000 aBP. However, the relatively low resolution of the  
700 record after about 27,500 aBP means that a continuation of abrupt shifts in rainfall cannot  
701 be ruled out. Major glacial advances in the TMVB occurred around the LGM (22,000 –  
702 18,000 aBP) with a suggested 6 – 8°C lowering in mean annual temperature. Reduced  
703 temperature, along with low effective moisture, could have contributed to the elevated  
704  $\delta^{18}\text{O}_{\text{carb}}$  values at this time. Although generally dry, there seem to have been short lived  
705 lake highstands in Cuitzeo and rising water levels in the Chalco Basin (southern Basin of  
706 Mexico) and the Lerma Basin around 18,000 and 19,000 aBP respectively (Caballero *et*  
707 *al.*, 2002). Unfortunately the evidence from La Piscina de Yuriria is at too low resolution to  
708 be able to detect these highstands. The late glacial in the TMVB (18,000 – 15,000 aBP) is  
709 described as cold and dry, with minor glacial recession (Caballero *et al.*, 2010): evidence  
710 from La Piscina de Yuriria indicates a continuation of dry conditions through the late glacial  
711 (=Younger Dryas) stadial. Significant warming and glacial retreat apparently started  
712 around 14,000 aBP.

713

714 Overall, the records of cool and dry conditions around the LGM and into the early  
715 Holocene from the central and eastern part of the TMVB are most easily explained by a  
716 reduction in summer season precipitation, which is driven today by the northward  
717 movement of the ITCZ and the onset of the NAM. Records from the Pátzcuaro Basin, to  
718 the west of this group, show high lake levels persisting through the LGM (Bradbury 2000;  
719 Metcalfe *et al.*, 2007), while Correa-Metrio *et al.* (2012) suggest conditions moist enough  
720 for pine forest to dominate over grasslands in the Zacapu basin. In both cases an increase

721 in winter precipitation may provide an explanation although it is hard to reconcile this with  
722 the stable-isotope data from La Piscina de Yuriria, which suggest an overall decrease in  
723 effective moisture. Lachniet *et al.* (2013) have further suggested that the summer  
724 monsoon did not collapse during the last glacial, although once again the evidence from  
725 La Piscina de Yuriria does not appear to support this, nor does the latest interpretation of  
726 the Chalco record (Lozano-Garcia *et al.*, 2015). The late glacial stadial was a time of dry  
727 conditions, with some suggestion that La Piscina de Yuriria may have dried out. A dry  
728 Younger Dryas stadial has also been reported from Zacapu (Correa-Metrio *et al.*, 2012)  
729 and from sites in northern Mexico (e.g. Roy *et al.*, 2013). Interestingly, the Younger Dryas  
730 is also reported as dry in the Juxtlahuaca speleothem record (Lachniet *et al.*, 2013) where  
731 it is attributed to monsoon collapse. It appears that the weakening of the Atlantic  
732 Meridional circulation (AMOC), the subsequent southward displacement of the ITCZ and a  
733 weaker monsoon led to dry conditions over the northern hemisphere neotropics (Bush and  
734 Metcalfe, 2012). Only at the northern edge of the NAM region did the resumption of cold  
735 conditions allow more penetration of mid-latitude westerlies leading to wetter conditions  
736 (Metcalfe *et al.*, 2015).

737

738 The classic pattern of climatic change in the NH tropics and subtropics is for wetter  
739 conditions in the early Holocene driven by poleward migration of the ITCZ and a stronger  
740 monsoon in response to insolation forcing. Whilst there is some support from this in  
741 Mexico, it seems that the establishment of the modern climatic regime was delayed by the  
742 presence of the residual Laurentide Ice Sheet and the influence of meltwater pulses  
743 entering the Gulf of Mexico (Metcalfe *et al.*, 2015). The early Holocene interval in La  
744 Piscina de Yuriria indicates a change to wetter conditions overall, but with evidence for  
745 abrupt shifts between wet and dry conditions. A similar pattern is recorded by Park *et al.*  
746 (2010) for their other sites in the Valle de Santiago. The response of the wider NAM region

747 to the insolation maximum in the early Holocene is complex, with the clearest response in  
748 the south, where the direct influence of the ITCZ is strongest. Elsewhere, it seems that the  
749 modern NAM regime may not have resumed until after 8000 aBP and there was a trade-  
750 off between increasing precipitation and increasing temperatures (Metcalf *et al.*, 2015).  
751 La Piscina de Yuriria shows an overall trend of drying from the mid Holocene onwards.  
752 This is consistent with the southward migration of the ITCZ and as the role of insolation  
753 forcing became weaker, so the effect of other climate forcings such as ENSO, seems to  
754 have become more important giving rise to increasingly complex patterns of change. This  
755 drying was accompanied by increasing human impact, as shown by evidence for phases  
756 of sediment inwash and by the presence of *Z. mays* pollen. Increasing human impact  
757 during this interval is also evident from other sites in Mexico (Metcalf *et al.*, 1994).

758

## 759 **Conclusions**

760 In summary, evidence from La Piscina de Yuriria indicates that the climate of Central  
761 Mexican highlands has changed dramatically over the past ~30,000 years. Between  
762 30,000 and about 27,500 aBP it was highly variable with shifts, which may have been  
763 abrupt, between dry and wet conditions. Over much of the glacial period, from ~27,500 to  
764 about 14,000 aBP, climate became drier: there may have been abrupt shifts during this  
765 interval, but the low resolution of our data means that any such shifts are not revealed.  
766 The occurrence of strong millennial scale variability during MIS3, with a global signature,  
767 has been widely noted (Clement and Peterson, 2008), apparently driven by changes in  
768 AMOC. It is notable that there were three D/O warming events (2 – 4) between 30,000 and  
769 22,000 aBP (Wolff *et al.*, 2010) and two Heinrich events (H2 and H3). Modelling has  
770 indicated differential sensitivity of AMOC under MIS3 and LGM conditions (e.g. Van  
771 Meerbeek *et al.* 2009), with the climate becoming less sensitive to AMOC changes as the  
772 full glacial climate was established. The growth of the Laurentide Ice sheet from an

773 interstadial minimum around 35,000 aBP to its maximum by ca. 25,000 aBP (where it  
774 remained until around 15,000 aBP) (Dyke et al., 2002) also reflects the shift of the global  
775 climate system into full glacial mode, where other, less rapid forcings may have dominated  
776 (see Baker and Fritz, 2015). There is evidence of drought during the period that  
777 encompassed the late glacial stadial. During the Holocene, the climate initially became  
778 wetter, although the positive water balance was insufficient to lead to major changes in the  
779 lake's chemistry and diatom flora. A fall in lake levels under drier climate in the mid  
780 Holocene was accompanied by a change in limnology that caused methane formation.  
781 Inwash of catchment soils and sediments was the result of a combination of natural  
782 climatic triggers and, for the later Holocene, anthropogenic disturbance.

783

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798

799

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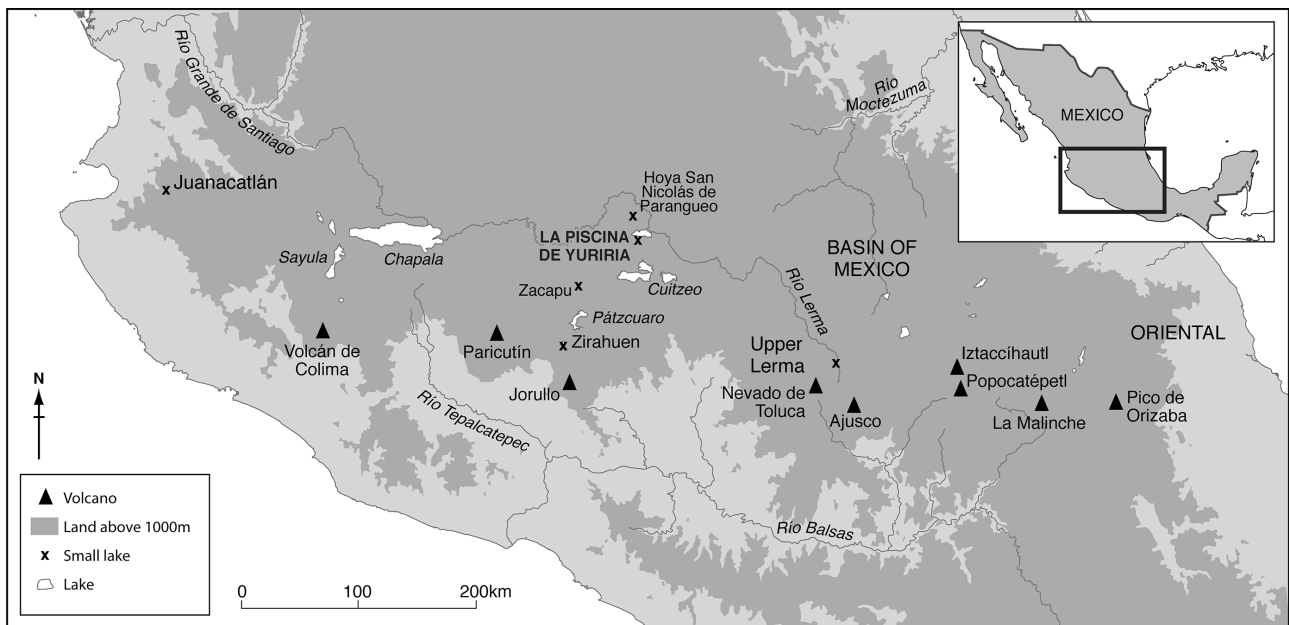
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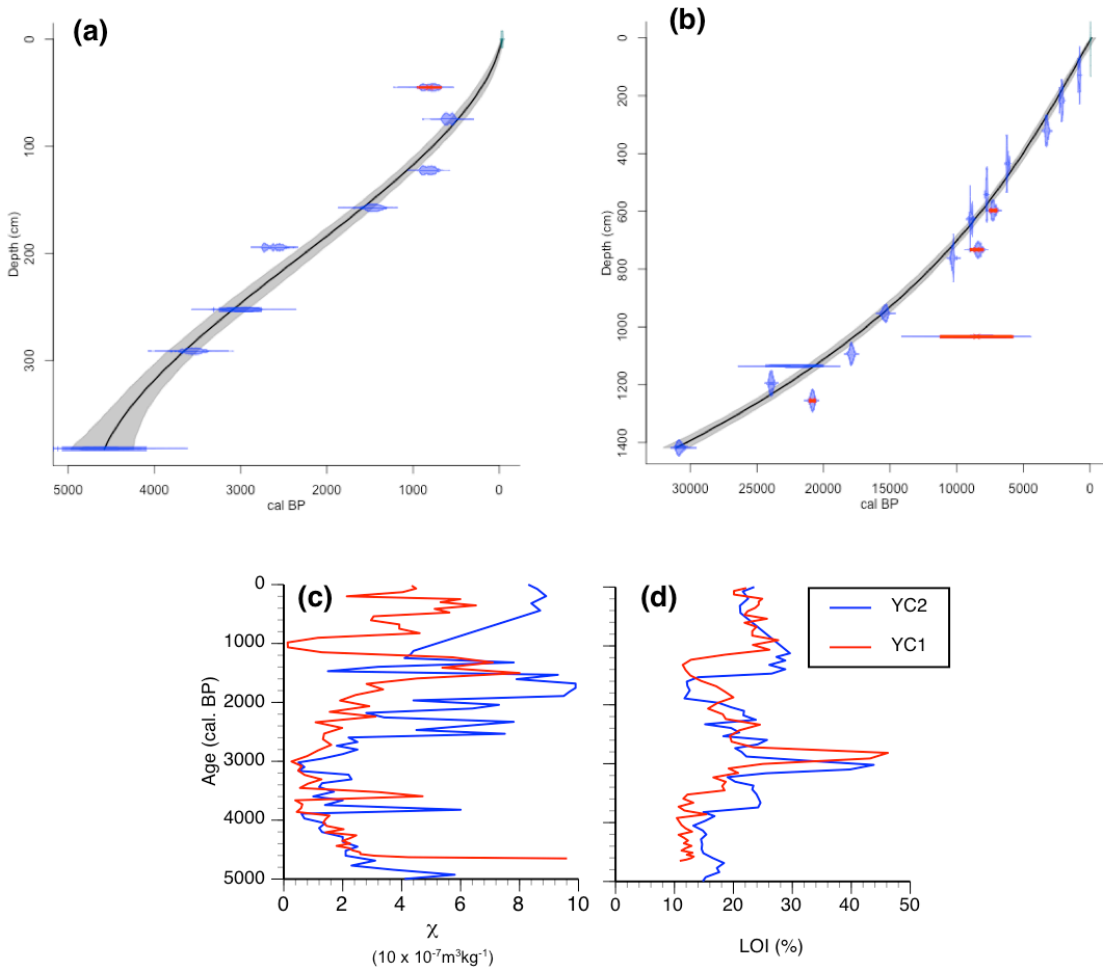
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1112 **Figures**



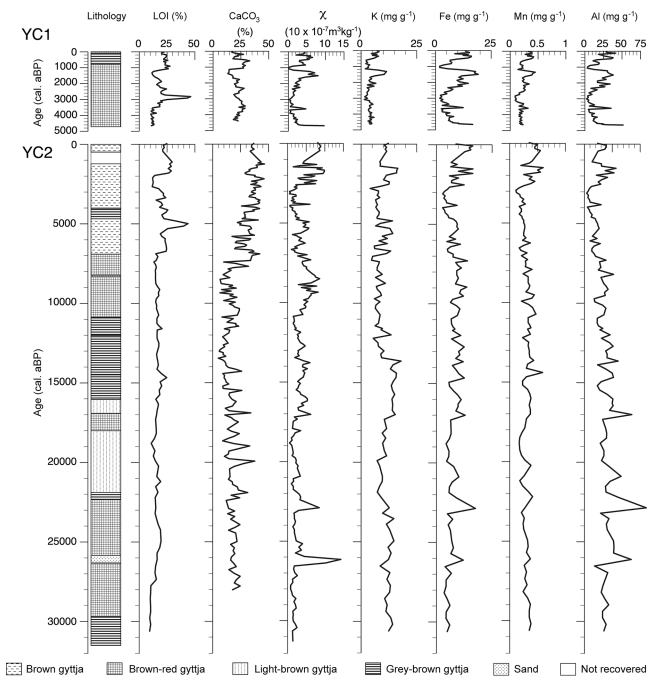
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1114 Fig. 1. Location of la Piscina de Yuriria and other sites referred to in the text.



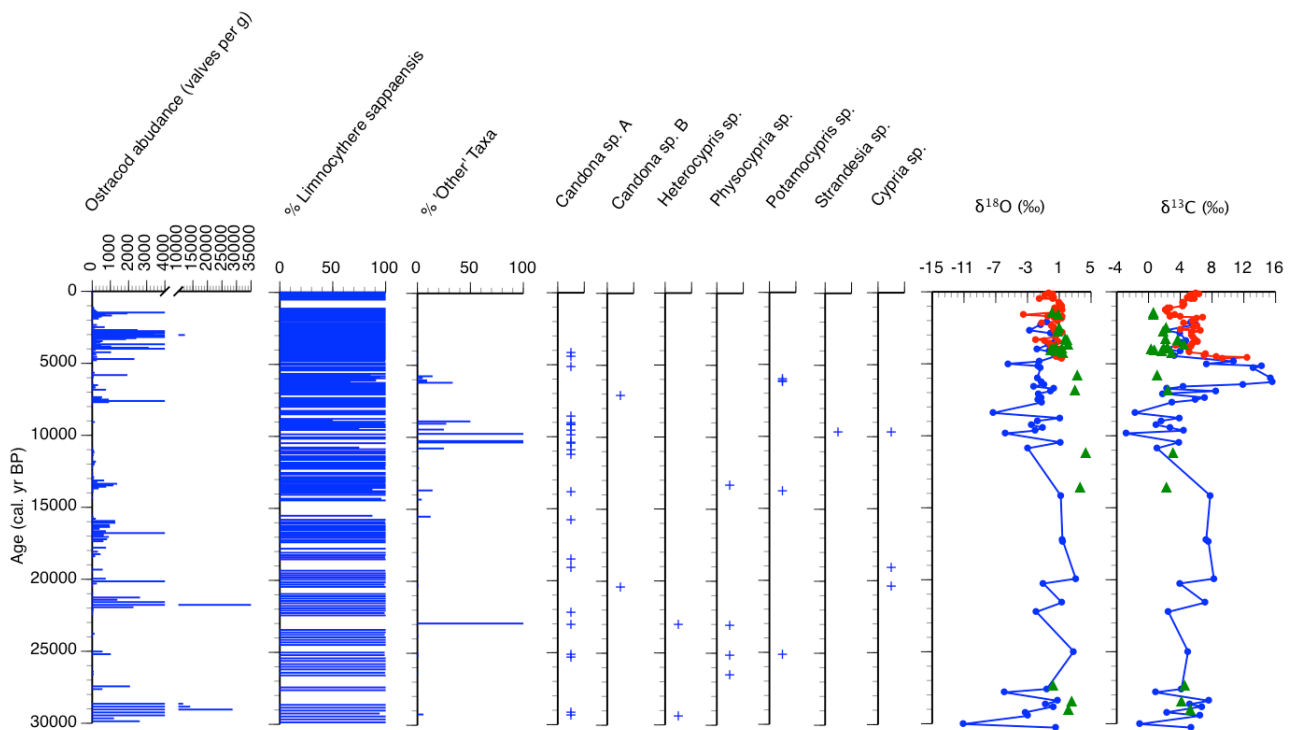
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1116 Fig. 2. Age-depth relationships for (a) YC1 (b) YC2, based on radiocarbon dates. The  
 1117 datapoints are well described by 3rd order polynomial curves,  $AGE = -0.0001Depth^3 +$   
 1118  $0.06645Depth^2 + 2.473Depth - 29.23$  for YC1 and,  $AGE = -0.000005Depth^3 -$   
 1119  $0.0006Depth^2 + 12.23Depth - 109.1$  for YC2 where age is in calendar years BP and depth  
 1120 is in cm in both cases. Detailed synchronisation for YC1 and the upper part of YC2 based  
 1121 on (c) loss-on-ignition and (d) magnetic susceptibility.



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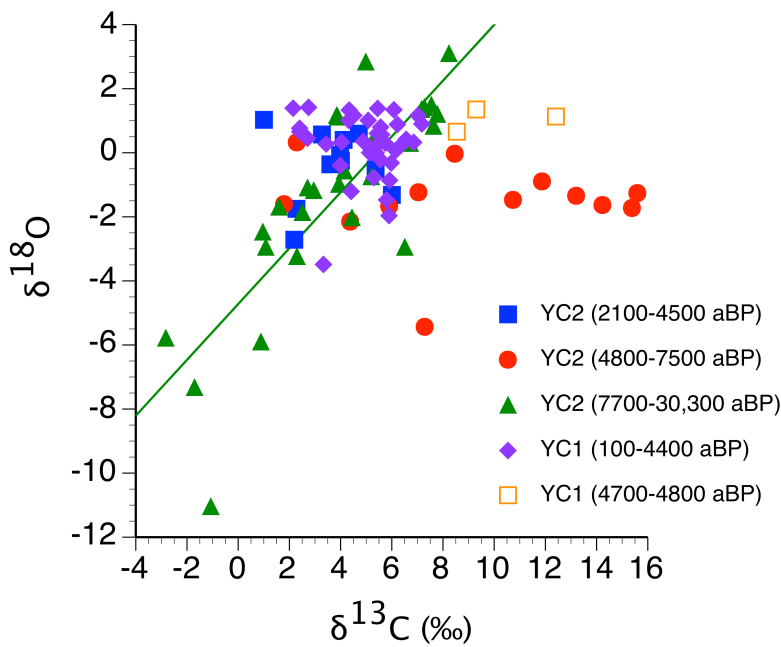
1123 Fig 3. Physical sedimentology and selected elemental geochemical variables for YC1 and  
 1124 YC2, plotted as a function of age in calendar years.



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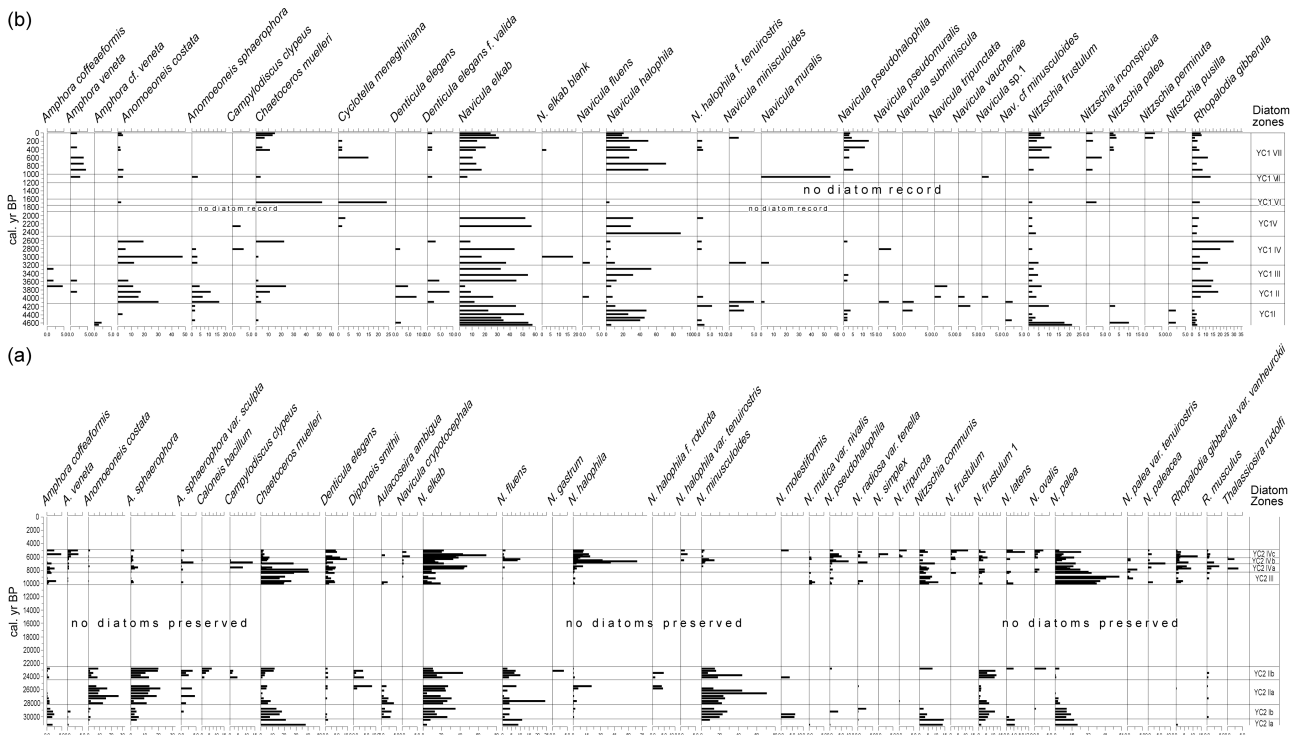
1126 Fig. 4. Ostracod assemblages for YC2 and stable isotopes: endogenic carbonate for YC1  
 1127 and YC2 and ostracod shells (triangles) for YC2). The 'other taxa' percentage curve  
 1128 includes all taxa except *Limnocythere sappaensis*.

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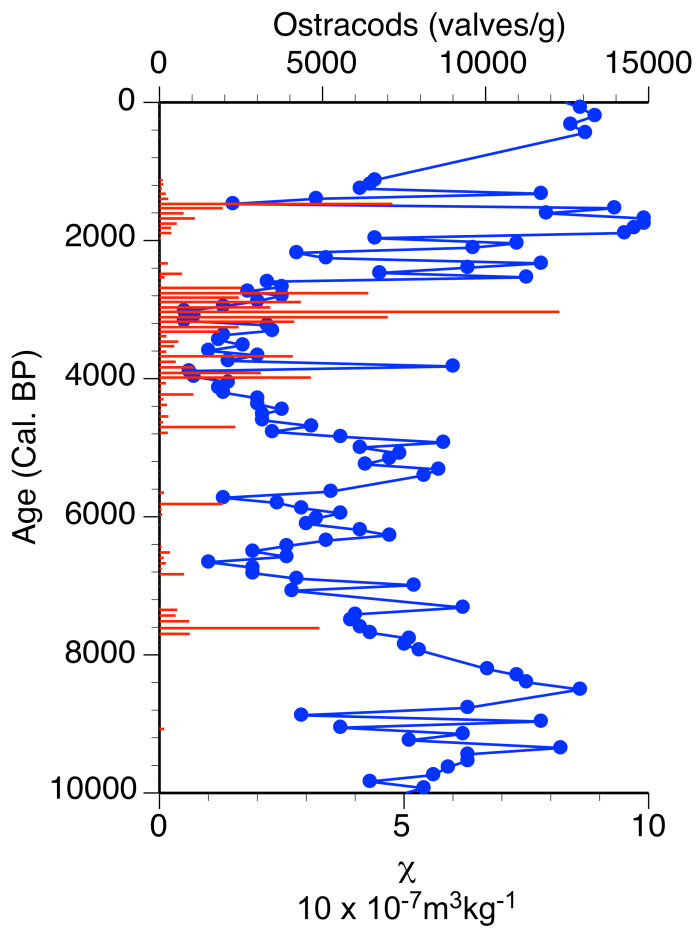
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1131 Fig. 5. Covariance of  $\delta^{18}\text{O}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{carb}}$  values for YC1 and YC2. For the interval YC2  
 1132 7700-30,300 aBP only, there is significant covariance between  $\delta^{18}\text{O}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{carb}}$  values  
 1133 ( $R^2 = 0.68$ ).



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1135 Fig. 6. Diatom abundance (%) for (a) YC2 and (b) YC1. Only taxa present at > 2% and in  
 1136 more than one sample are plotted.



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1138 Fig. 7. Ostracod abundance and magnetic susceptibility for the last 10,000 years in core

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1150 **Tables**1151 **Table 1. Radiocarbon dates from cores YC1 and YC2.****Table 1. Radiocarbon dates from cores YC1 and YC2**

Core	Depth (cm)	Date Type	Lab number	Radiocarbon age (yr)	Error (yr)	Cal age range (yr) (95 % confidence intervals)		Material	Notes
YC1	44-46	AMS	OxA-1963	875	90	670	951	carbonate	omitted from age model
YC1	74-75	AMS	OxA-1964	570	80	500	673	carbonate	
YC1	122-123	AMS	OxA-1965	900	60	705	927	carbonate	
YC1	157-158	AMS	OxA-1966	1570	80	1307	1680	carbonate	
YC1	194-195	AMS	OxA-1967	2540	60	2383	2759	carbonate	
YC1	250-255	AMS	RIDDL-62	2840	120	2751	3323	carbonate	
YC1	291-292	AMS	OxA-1968	3320	90	3372	3824	carbonate	
YC1	380-385	AMS	RIDDL-63	4100	200	4084	5278	carbonate	
YC2	129-130	AMS	AA13908	860	50	689	908	bulk organic carbon	
YC2	216-217	AMS	AA13907	2125	50	1955	2306	bulk organic carbon	
YC2	321-322	AMS	AA13906	3040	70	3008	3391	bulk organic carbon	
YC2	430-440	Radiometric	SRR-5189	5410	50	6018	6300	bulk organic carbon	
YC2	537-547	Radiometric	SRR-5190	6925	45	7670	7911	bulk organic carbon	
YC2	592-602	Radiometric	2815-Ors	6340	140	6939	7552	bulk organic carbon	omitted from age model
YC2	622-632	Radiometric	SRR-5191	8055	50	8730	9120	bulk organic carbon	
YC2	728-738	Radiometric	2844-Ors	7570	188	8001	8967	bulk organic carbon	omitted from age model
YC2	761-762	AMS	AA13905	9105	80	9969	10511	bulk organic carbon	
YC2	953-954	AMS	AA13904	12865	105	15085	15728	bulk organic carbon	
YC2	1029-1039	Radiometric	2812-Ors	7100	1200	5752	11244	carbonate	omitted from age model
YC2	1093-1094	AMS	AA13903	14705	105	17617	18165	bulk organic carbon	
YC2	1130-1140	Radiometric	2869-Ors	18180	980	19938	24375	bulk organic carbon	
YC2	1190-1200	Radiometric	SRR-5192	19880	75	23676	24160	bulk organic carbon	
YC2	1250-1260	Radiometric	SRR-5193	17275	90	20577	21103	bulk organic carbon	omitted from age model
YC2	1413-1423	AMS	AA16907	26590	280	30308	31189	bulk organic carbon	

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1166 Table 2. Summary of climatic and environmental changes at La Piscina de Yuriria over the  
 1167 past 30,000 years

**Table 2. Summary of climatic and environmental changes at La Piscina de Yuriria over the past 30,000 years**

Age Range (cal. aBP)	Climatic and environmental conditions	Key evidence
0-4,500	Lower effective moisture under drier climate regime but with short-lived wetter intervals	Oxygen isotopes
	Saline -alkaline lake	Ostracods, diatoms
	Transient anthropogenically-induced inwash events associated with catchment instability	Elemental geochemistry and magnetic susceptibility
4,500-8,000	Reduced effective moisture leading to enhanced evaporative enrichment under drier climate	Oxygen isotopes
	Saline - alkaline lake with nutrient enrichment and short-lived fresher interval	Ostracods, diatoms
	Intense methane formation	Carbon isotopes
	Climatically-controlled inwash event during wet intervals	Elemental geochemistry and magnetic susceptibility
8,000-11,000	Higher effective moisture, but with evidence of variability under wetter but variable climatic regime	Oxygen isotopes
	Fresher and deeper lake, but eutrophic	Ostracods, diatoms
11,000-14,000	Dry climate, lake dessication	Stratigraphy (presence of desiccation surface)
14,000-25,500	Reduced effective moisture leading to enhanced evaporative enrichment under drier climate	Oxygen isotopes
	Periodic lake desiccation	Stratigraphy (presence of desiccation surface)
	Saline - alkaline lake with short-lived fresher intervals	Ostracods, diatoms
	Shallow eutropic lake with methane formation	Diatoms, carbon isotopes
	Stable catchment with limited inwash	Elemental geochemistry and magnetic susceptibility
27,500-30,000	Rapid shifts between low and high effective moisture under variable dry to wet climatic regime, possibly accompanied by cooler conditions	Oxygen isotopes
	Saline - alkaline lake	Ostracods, diatoms
	Shallow eutropic lake with methane formation	Diatoms, carbon isotopes
	Stable catchment with limited inwash	Elemental geochemistry and magnetic susceptibility

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1171 **Supporting online information**

1172 **Table S1. Water Chemistry for La Piscina de Yuriria. Date from Davies (1995) and various**  
 1173 **unpublished sources.**

**Table S1. Water Chemistry for La Piscina de Yuriria. Date from Davies (1995) and various unpublished sources.**

Year	Water depth (m)	Type	pH	EC <sup>a</sup> (µScm <sup>-1</sup> )	Alkalinity (Total) (meqL <sup>-1</sup> )	Cl (meqL <sup>-1</sup> )	SO <sub>4</sub> (meqL <sup>-1</sup> )	K (meqL <sup>-1</sup> )	Na (meqL <sup>-1</sup> )	Ca (meqL <sup>-1</sup> )	Mg (meqL <sup>-1</sup> )	TP (µgL <sup>-1</sup> )	δ <sup>13</sup> C <sub>DIC</sub> ‰VPDB	δ <sup>18</sup> O ‰VSMOW	δD ‰VSMOW
1982 (April)	2 (surface)	Lake	11.0	26000	300.0	174.3	29.3	30.7	478.5	0.0	0.0	ND <sup>#</sup>	ND	ND	ND
1982 (April)	2 (at depth)	Lake	NA	23000	326.0	141.5	30.2	24.8	478.5	0.0	0.0	ND	ND	ND	ND
1982 (April)	0.2 (margin)	Lake	11.0	29000	294.0	116.3	29.5	32.1	413.3	0.0	0.0	ND	ND	ND	ND
1982 (April)		Spring	7.5	500	6.2	0.4	1.4	0.6	3.7	1.0	1.3	ND	ND	ND	ND
1982 (April)		Spring	7.0	1100	5.0	1.1	3.6	1.0	9.6	1.1	1.6	ND	ND	ND	ND
1982 (May)	0.2 (margin)	Lake	9.6	27500	333.0	203.0	33.5	32.1	543.7	0.0	0.0	ND	ND	ND	ND
1982 (May)		Spring	7.7	700	8.0	4.2	0.3	0.6	8.9	0.9	1.6	ND	ND	ND	ND
1992 (August)	0.2	Lake	10.5	15620	248.2	117.4	50.0	22.8	298.5	0.3	0.1	ND	-0.2	0.7	-25
1992 (August)		well	6.8	315	3.7	0.4	9.6	0.9	5.7	4.4	3.9	ND	-10.4	-9.2	-69
1992 (August)		spring	8.0	452	7.7	0.4	7.5	2.0	8.4	3.3	3.7	ND	-11.6	-9.1	-69
1997 (March)	shore	Lake	10.2	8130	68.3	27.5	9.2	7.0	95.8	5.9	1.9	ND	ND	ND	ND
2003 (March)	0.2 (margin)	Lake	9.5	1907	20.8	6.2	0.4	1.7	26.3	2.9	2.4	562.0	ND	ND	ND
2004 (July)	0.2 (margin)	Lake	9.6	2910	16.4	16.7	0.8	1.5	163.4*	2.6	0.0	744.5	ND	ND	ND
2003 (March)		Groundwater	8.0	1076	ND	ND	ND	ND	ND	ND	ND	ND	ND	ND	ND

<sup>a</sup>Electrical conductivity  
<sup>#</sup>ND = not determined  
 \* possible contamination

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1190 Table S2. Correlation matrices ( $R^2$  values) for correlation amongst selected geochemical  
 1191 variables from (a) YC1 (b) YC2, whole core (c) YC2 post 5000 aBP. In each table the null  
 1192 hypothesis ( $H_0$ ) is that there is no statistically-significant relationship between the two  
 1193 variables

**Table S2. Correlation amongst selected geochemical variables**

(a) Core YC1

		K	Fe	Mn	Al
<b>K</b>	<b>R<sup>2</sup></b> <i>p-value</i> <i>H0 (5%)</i>	1.			
<b>Fe</b>	<b>R<sup>2</sup></b> <i>p-value</i> <i>H0 (5%)</i>	<b>0.65</b> 1.68E-16 <i>rejected</i>	1.		
<b>Mn</b>	<b>R<sup>2</sup></b> <i>p-value</i> <i>H0 (5%)</i>	<b>0.68</b> 8.35E-18 <i>rejected</i>	<b>0.77</b> 0.00E+00 <i>rejected</i>	1.	
<b>Al</b>	<b>R<sup>2</sup></b> <i>p-value</i> <i>H0 (5%)</i>	<b>0.52</b> 4.76E-12 <i>rejected</i>	<b>0.901</b> 0.00E+00 <i>rejected</i>	<b>0.67</b> 2.46E-17 <i>rejected</i>	1.

(b) Core YC2 - all

		K	Mn	Fe	Al
<b>K</b>	<b>R<sup>2</sup></b> <i>p-value</i> <i>H0 (5%)</i>	1.			
<b>Fe</b>	<b>R<sup>2</sup></b> <i>p-value</i> <i>H0 (5%)</i>	<b>0.15</b> 0.00001 <i>rejected</i>	1.		
<b>Mn</b>	<b>R<sup>2</sup></b> <i>p-value</i> <i>H0 (5%)</i>	<b>0.06</b> 0.00581 <i>rejected</i>	<b>0.45</b> 4.00E-17 <i>rejected</i>	1.	
<b>Al</b>	<b>R<sup>2</sup></b> <i>p-value</i> <i>H0 (5%)</i>	<b>0.22</b> 4.32E-08 <i>rejected</i>	<b>0.19</b> 6.67E-07 <i>rejected</i>	<b>0.37</b> 0. <i>rejected</i>	1.

(b) Core YC2 - post 5000 aBP

		K	Mn	Fe	Al
<b>K</b>	<b>R<sup>2</sup></b> <i>p-value</i> <i>H0 (5%)</i>	1.			
<b>Fe</b>	<b>R<sup>2</sup></b> <i>p-value</i> <i>H0 (5%)</i>	<b>0.61</b> 9.64E-08 <i>rejected</i>	1.		
<b>Mn</b>	<b>R<sup>2</sup></b> <i>p-value</i> <i>H0 (5%)</i>	<b>0.49</b> 5.96E-06 <i>rejected</i>	<b>0.75</b> 7.15E-11 <i>rejected</i>	1.	
<b>Al</b>	<b>R<sup>2</sup></b> <i>p-value</i> <i>H0 (5%)</i>	<b>0.59</b> 2.00E-07 <i>rejected</i>	<b>0.71</b> 9.10E-10 <i>rejected</i>	<b>0.8</b> 1.67E-12 <i>rejected</i>	1.

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