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

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Running head: Palaeolimnology of a Mexican Crater Lake

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

KEYWORDS: Trans-Mexican Volcanic Belt; Palaeolimnology; diatoms; ostracods; stable isotopes
Introduction and previous work

The highlands of the Trans-Mexican Volcanic Belt (TMVB), which cross Mexico at around 19°N, provide a range of opportunities for reconstructing past climates through the study of the sediments in its many lake basins, the availability of glacial deposits, tree ring records and a wealth of historical documents (primarily since the Spanish conquest in 1521). In spite of these many possibilities, understanding of climatic variability over the late Quaternary, particularly the late Pleistocene and early Holocene, is still rather limited (Caballero et al., 2010). A number of factors help to explain this, including poor dating control of many records, the effects of tectonic and volcanic activity, and especially human disturbance, which over at least the last 4000 years has had a profound impact on the natural environment to the extent that only records from the highest elevations may be unaffected (e.g. Lozano-Garcia and Vazquez-Selem, 2005). Lake sediment records have dominated studies of central Mexican palaeoclimatology, although recent speleothem records (Bernal et al., 2011) have also made a contribution. Palynology has played a central role in palaeolimnological investigations dating back to the 1950s (Sears and Clisby, 1995; Watts and Bradbury, 1982; Goman and Byrne, 1998; Lozano-Garcia et al., 2005). In this context, the dominance of pine-oak woodlands at higher altitudes in the TMVB has presented further challenges given the limited taxonomic resolution of these pollen types in standard palynology and the impact of long-range dispersal, especially of pine pollen (Correa-Metrio et al., 2012). The climatic interpretation of pollen diagrams from the TMVB has often hinged on decisions about the climatic conditions that changing proportions of pine and oak represent, or indeed whether the presence of large percentages of pine means that pine trees were actually present (c.f. Brown, 1985; Park et al., 2010). Pollen records are also highly susceptible to anthropogenic disturbance, although the distinctive presence of Zea mays pollen in records is a clear indicator of agricultural activity close to lakes. To help to resolve some of the uncertainties around
pollen-based reconstructions, the application of palaeolimnological methods, particularly diatom, geochemical, isotopic and mineralogical analyses, has become increasingly common across the TMVB. Whilst not immune to many of the complicating factors outlined above, this multi-proxy approach has helped to improve our understanding of both climatic and environmental change in this area. In this paper we present a well-dated palaeolimnological record from a crater lake, La Piscina de Yuriria, on the southern edge of the Valle de Santiago region, southern Guanajuato, which extends back some 30,000 years (all ages quoted in this paper that relate to the radiocarbon time-span are in calendar years unless otherwise stated). This is considered in the context of other data from adjacent crater lakes in the Valle de Santiago area and other lakes in the wider region to explore the timing and nature of climatic change and human impact.

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

**Study Region**

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

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

Unfortunately, the maar that contains La Piscina de Yuriria was not dated, although the adjacent shield volcano is believed to date to 6.9 Ma (Aranda-Gomez et al., 2013). It has been suggested that these maar lakes were once set within a large palaeolake, which extended from the modern Laguna de Yuriria, northwards around the modern town of Valle de Santiago. The maars have very small catchment areas and the lakes within them were supported by the waters of the underlying Salamanca aquifer. Unfortunately, this aquifer has been heavily exploited (there are more than 1600 active wells), which has resulted in a drawdown of about 2 m yr\(^{-1}\) over the past 25 years (Alcocer et al., 2000). As a result all the maars are now dry, even those that contained deep lakes in the early 1980s. Alcocer et al. (2000) report on the many undesirable consequences of this desiccation, including the loss of endemic fish species and lake margin wetland habitats, the economic effect of the loss of fisheries and alkali fly collection and adverse health
effects due to the mobilisation of alkaline dust. La Piscina de Yuriria contained a lake approximately 2 m deep in 1981 and was then observed to dry out through the 1980s, with the development of a salt crust across the basin floor. Freshwater springs around the margins of the lake within the crater also dried up. A shallow lake was re-established in the early 1990s, which was then made permanent in the early 2000s by the pumping of groundwater back into the crater. As this was one of the basins where adverse health effects were reported due to dust mobilisation, this re-wetting of the basin may have been a response.

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

The dramatic effects of water extraction on the maar lakes attracted work on what had been the deep lakes of Hoya la Alberca and the Hoya Rincon de Parangueo (Kienel et al., 2009; Park et al., 2010) partly in an effort to retrieve cores of laminated sediments before
these became lost through deflation or profoundly disturbed by desiccation and secondary precipitation of evaporite minerals. The Hoya Rincon de Parangueo record goes back to 9600 aBP. Prior to this, there had been work on the Hoya San Nicolas (Brown, 1985; Metcalfe et al., 1989), which had been cored in 1979 shortly after it dried out. The basal date on this core was 12,600-12,700 aBP. This maar was re-cored in 2001 with the results reported by Park et al. (2010) giving a record believed to extend back to 11,600 aBP (but not directly dated). As described further below, La Piscina de Yuriria was cored in 1981 and 1982 prior to its desiccation.

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

Study site
This study is based on lake sediment records from La Piscina de Yuriria (20°30'N; 101°08'W, 1740 m a.s.l.), which is one of the small (area = 0.75 km²), hydrologically-closed maar lakes in the Valle de Santiago region (Fig. 1). The basaltic basin experiences a subhumid, subtropical climate with annual precipitation of 700-800 mm and supports subtropical thorn bush scrub around the lake and sparse oak woodland above 2200 m a.s.l. (Metcalfe and Hales, 1994; Metcalfe et al., 1994). The lake seems to have been generally shallow in recent times (e.g. ~ 2 m deep in 1981 and 1982, <1 m in 1992, >1.8 m in 2004) and has dried up totally in some years (e.g.1989). The lake was saline as a result of evaporative enrichment, as well as highly eutrophic, with an alkalinity/Ca ratio >>1 and Na-CO₃-Cl-type composition indicating that evaporative evolution occurred along pathway IIIA of Eugster and Hardie (1978). The lake was fed by several circum-neutral to alkaline springs that were fresh to slightly brackish (Table S1) and also of Na-HCO₃ type. Evaporative enrichment is also reflected in the limited stable isotope data for the input water (-9.2 ‰) versus lake water (+0.7 ‰) (Table S1). Recent water-level changes have been the result of groundwater extraction for irrigation (e.g. Metcalfe and Hales, 1994) and later artificial recharge; variations during the late Quaternary, which form a major focus of the present study, have largely been driven by changes in effective moisture (precipitation minus evaporation, or P-E), as discussed below.

Materials and methods

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

Physical sedimentology

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

Bulk sediment geochemistry

Dried sediment samples from YC1 were digested using a combination of HNO₃, HF and H₂O₂ following Dean and Gorham (1976). The resulting residue was taken up in dilute HCl prior to analysis of major and minor metals using atomic absorption spectrophotometry (AAS). Samples from YC2 were prepared using the sequential digestion method of Engstrom and Wright (1984) and each separate fraction analyses for major and minor metals as for YC1. However, initial inspection of the results indicated that the fractionation had not worked well for these complex sediments. Consequently, the results for the separated fractions were summed and the data treated as ‘bulk’ analyses, as for YC1. Only selected metals are reported here, namely Fe, Mn, Al, and K, all of which are
abundant in catchment soils and sediments, and so are regarded as good tracers of inwash, as discussed in further detail below.

Stable isotopes in endogenic carbonate

Bulk, dried sediment samples from YC1 were sieved through an 80µm mesh to remove shell material, treated with Clorox to remove organic carbon and then dissolved in 100 % phosphoric acid. The evolved CO₂ was then analysed for oxygen and carbon isotopes using a VG ‘Micromass’ mass spectrometer at the Laboratoire d’Hydrologie et de Géochimie Isotopique, Université de Paris-Sud. Samples from YC2 were prepared in a similar way and analysed using a modified SIRA mass spectrometer at the University of Liverpool. All stable isotope results are reported in standard delta notation relative to the PDB standard.

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

Ostracods

Bulk sediment samples were disaggregated in ~5 % H₂O₂, sieved through a 63 µm sieve and the coarse fractions used for extraction of ostracod shells under a low-power binocular microscope. Because ostracod abundance varied dramatically, in some samples all of the ostracod valves were picked whereas in richer samples, only the first 350 valves were picked and overall abundance then estimated based on the weight of sediment examined. However, the unpicked fraction of rich samples was inspected for the presence of rare species not identified in the original count. Despite poor ostracod shell preservation in
many parts of the core, sufficiently well-preserved valves were found at a number of levels
to permit stable-isotope analyses, which were used to complement analyses on the
dendogenic carbonate.

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

Diatoms

Diatom samples were prepared for analysis at 10 cm intervals in YC1 and 10 – 20 cm
intervals (depending on sample availability) in YC2. In YC2, only material from 4 m to the
base was sampled in detail after preliminary analysis (samples prepared every 25 – 30
cm) had shown that the upper 4 m of the sequence seemed to match well with YC1. Poor
valve preservation between 7 m and 12 m meant that the majority of samples were taken
below and above this interval. Core samples were complemented by a surface sediment
sample and water samples from the lake margin. Samples were treated to remove
carbonates and organic matter following the method of Battarbee (1986). In some cases
strong acids (H$_2$SO$_4$ or HNO$_3$) were used to break up clumps of sediment. The final
suspension was mounted onto coverslips using Naphrax. The same method was used to
prepare surface sediment samples. Where possible, 400 valves were counted from each
level, although at levels with poor preservation this was sometimes reduced to 100 – 200 valves. Identifications were carried out using standard floras including Gasse (1986), Krammer and Lange Bertalot (1988, 1991a and b), Patrick and Reimer (1966, 1975), Schoeman and Archibald (1977). The separation of *Navicula* (*Craticula*) *elkab* from *Craticula halophila* proved to be an important part of the study and is discussed in Metcalfe (1990). The results from YC1 were published in Metcalfe and Hales (1994), and preliminary results from YC2 were described in Park (1999). Diatom results are presented here as percentages of the full counts. Species present at less than 2% and only occurring in one sample are not plotted here.

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

**Chronology**

Independent chronologies were established using AMS or radiometric radiocarbon dating on bulk organic carbon or carbonate (8 levels were dated from YC1 and 16 from YC2). Lack of appropriate material precluded the dating of terrestrial plant macrofossils from either core, but the absence of carbonate bedrock within the catchment means that hardwater error is unlikely to affect radiocarbon dates at this site. Radiocarbon measurements were performed at the Oxford University Radiocarbon Accelerator Unit, Simon Fraser University Radiocarbon facility, Laboratoire d’Hydrologie et de Géochimie
Isotopique, Université de Paris-Sud, the NERC Radiocarbon Facility, East Kilbride and the Arizona AMS Laboratory, University of Arizona. Dates were calibrated using IntCal13 (Reimer et al., 2013). For each core, outliers were identified and excluded from subsequent age modeling, which was undertaken using Clam 2.2 (Blaauw, 2010). Although some of the radiocarbon dates have been published previously, we report them again here for the sake of completeness and because we have revised the calibrations.

**Results**

**Chronology**

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

**Physical sedimentology**

Cores YC1 and YC2 core are both composed predominantly of brown to grey, diatomaceous gyttja. In the lower sections of YC2 (at ~26,300, 19,900 and 8300 aBP), centimetre-scale sand layers are present. Desiccation surfaces, represented by sub-vertical fractures infilled with darker material, occur at several depths in the interval between 16,800 and 14,300 and near the top of the sequence (~800 aBP). Nodular, siliceous concretions were observed between the base of the sequence and ~17,500 aBP. The interval 100 – 50 cm (representing ~1100 – 500 aBP) was not recovered in YC2.
Organic carbon content, estimated from loss-on-ignition (LOI) at 450˚C, varies between ~9 and 44 % (Fig. 3). From the beginning of the YC2 record until ~14,000 aBP, values average ~15%, increasing and becoming more variable after this time, with peaks up to 44 %. The LOI record in the upper part of YC2 is well replicated in YC1. Carbonate content averages around 18 % from the beginning of the record until 14,200 aBP, but with peaks up to 40 % in places. Between 14,000 and 7800 aBP, carbonate content falls to 14 % and becomes less variable. From 7800 aBP to the end of the YC2 record, there is a sharp increase to an average of 33 %, with transient peaks up to 45 %. The pattern of change in YC1 replicates that in the upper part of YC2, but values are systematically lower.

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

**Geochemistry and stable isotopes**

Concentrations of K, Fe, Mn and Al are variable throughout the cores. For much of the lower part of YC2 values are generally low, although there are some transient peaks. Values typically become higher and more variable after about 5100 aBP in YC2, a pattern that is generally replicated in YC1. There is strong and statistically-significant correlation amongst all of the metals discussed here in YC1 and in the section of YC2 covering the past ~5000 years. Correlations are weaker for the earlier part of YC2 (Table S2).
Stable isotope values are available for fine-grained carbonates and, for a few levels from YC2 only, for ostracod shells (Fig. 4). Mineralogical determinations showed that the calcite present within the core had low magnesium content (≤4 mole %); nevertheless, appropriate corrections were made to the $\delta^{18}$O values following Tarutani et al. (1969).

From the beginning of YC2 to 26,600 aBP, $\delta^{18}$O and $\delta^{13}$C values are typically low but variable (from about 0 ‰ to around -11 ‰ and +7.6 to -1 ‰, for $\delta^{18}$O and $\delta^{13}$C, respectively) and strongly covariant (Fig. 5). Between 25,000 and 14,200 aBP, the stable-isotope values are at relatively low resolution, and typically values are higher than in the preceding interval (up to about +3 ‰ and +8 ‰ for $\delta^{18}$O and $\delta^{13}$C, respectively) and covariant. Between 10,900 and 6600 aBP, there is a further rise in variability, but $\delta^{18}$O and $\delta^{13}$C covary as before. Between 6600 and 4500 aBP, there is a significant shift in the pattern of stable isotope variability, with $\delta^{18}$O values remaining relatively constant, but $\delta^{13}$C increasing up to +15.4 ‰. After 4500 aBP, $\delta^{18}$O values vary between about 0 and -2.7 ‰ and $\delta^{13}$C values are markedly reduced compared with the preceding interval. The overlapping sections of YC1 and YC2 show similar patterns of change, especially for $\delta^{13}$C.

The low-resolution ostracod values are typically $^{18}$O-enriched and $^{13}$C-depleted compared with those for fine-grained carbonates.

**Ostracods**

Ostracod abundance varies dramatically, between zero (i.e. barren levels) and ~35,000 valves per gram of sediment (Fig. 4). Significant zones of very high abundance are present from the beginning of the YC2 record to until 28,700 aBP and at 21,900 – 21,300 aBP, 17,800 – 15,900 aBP, 13,700 – 12,900 aBP, 4800 - 2700 aBP, 2000 – 1100 aBP.

Significant zones that are barren or contain very few ostracods occur at 28,500 – 27,800 aBP, 19,000 – 18,500 aBP, 15,400 – 14,400 aBP, 10,500 – 8300 aBP, 5300- 4900 aBP.

The zones of high abundance generally contain monospecific assemblages of
*Limnocythere sappaensis* (both males and females): zones of lower abundance typically support other species as well, including at least one species of candonid, *Heterocypris* sp., *Physocypria* sp., *Potamocypris* sp., *Strandesia* sp. and *Cypria* sp. Although *L. sappanesis* is regarded as conspecific with *Limnocythere inopinata* (e.g. Martens, 1994), we adopt the name *L. sappaensis* here since this is widely used for the species in North America.

**Diatoms**

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

Preliminary diatom analysis of samples from the top 4 m (past ~5000 years) of the YC2 core indicated the presence of similar assemblages to those reported in Metcalfe and Hales (1994), so efforts on this core were focused on the section below 4 m. Here, results are presented from both core sequences (Fig. 6). As with the ostracods, diatom preservation is highly variable through the sequences. In YC2 there was little or no valve preservation between 22,100 – 10,400 aBP cm. In YC1 there are gaps in the record due to poor preservation between 2000 – 1100 aBP. Both sequences are dominated by *N. (C) halophila* and *N. (C) elkab*, *A. costata* and *A. sphaerophora*, *C. muelleri* and a range of
Nitzschia species. These assemblages are all similar to those found in surface sediment samples taken from the lake at different times, reflecting high (if varying) pH and alkalinity, and highly evaporated conditions. A more distinctive aspect of the core records is the abundance, at times, of small Navicula species (e.g. N. fluens, N. minusculoides, N. muralis).

Discussion

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

Core YC1 covers the interval from about 4600 aBP to the coring date. The anomalous age at 44-46 cm probably represents inwash of older carbon from the catchment, substantiated by the magnetic susceptibility data. Core YC2 covers the interval from about 31,000 aBP to the coring date. The four dates that are omitted from the age model are all younger than expected, possibly the result of root penetration during times of low lake level. The general pattern of age-depth relationships for YC1 and the upper part of YC2 suggests good agreement between the two cores. Detailed comparisons based on the loss on ignition and magnetic susceptibility profiles (Fig. 2), which would be expected to agree for the two closely-located sequences, suggest a small (~200 years) age offset, with YC2 being consistently older. The fact that small differences exist is perhaps not surprising given that different equivalent depths and contrasting materials were dated in the two cores (Table 1). However, rather than correct either one of the cores we prefer to use the age models defined for each respective core and then refer to the resulting uncertainties as appropriate.
Loss on ignition (LOI) provides a good proxy for the organic carbon content of the sediments (Dean, 1974). The carbon/nitrogen (C/N) ratios (available for YC1 only: data not shown, but values vary between 11.6 and 49.4) indicate that the organic matter is of mixed aquatic and terrestrial sources, suggesting that LOI provides a record of aquatic productivity and terrestrial inwash at least for the past 4,600 years. The CaCO$_3$ content of the sediment is best explained by endogenic carbonate formation within the lake, since the catchment is devoid of carbonate rocks or sediments. In such settings, calcium carbonate precipitates from the water column when the lake becomes saturated with respect to carbon minerals as a result of evaporative enrichment of water or mediated by aquatic plants. However, enhanced aquatic productivity can also lead to carbonate dissolution (e.g. Cohen, 2003) and carbonate formation also depends on supply of ions from the catchment, meaning that the interpretation of sedimentary CaCO$_3$ records is not always straightforward.

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

The oxygen and carbon isotope composition of lacustrine carbonate is a function of the temperature and isotopic composition of the water in the case of oxygen, and the carbon-isotope composition of dissolved inorganic carbon (DIC) for carbon. In subtropical dryland lakes, such as La Piscina de Yuriria, the oxygen-isotope ratio of lake water is usually the
dominant control on the oxygen-isotope composition of carbonate, and this varies with the
degree of evaporative enrichment (Talbot, 1990). The carbon-isotope composition of DIC
is a complex function of carbon source (catchment- vs lake-derived) and in-lake
modification as a result of fractionation during DIC uptake by aquatic plants for
photosynthesis, and exchange with atmospheric CO$_2$ (Talbot, 1990) The same general
controls determine the isotopic values of biogenic carbonates, such as ostracod shells, but
there may be taxon-specific differences in the exact location and timing (especially
season) of carbonate formation compared with endogenic carbonate (e.g. Decrouy et al.,
2011). Moreover, biogenic carbonate may not be precipitated in isotopic equilibrium with
lake water or DIC: ostracod shells, for example, demonstrate offsets from oxygen-isotope
equilibrium and are typically $^{18}$O-enriched compared with endogenic carbonate
precipitated in equilibrium with lake water (von Grafenstein et al., 1999). Further
palaeolimnological inferences can be drawn for the strength of covariance amongst
carbonate oxygen and carbon isotope values in sediment sequences, strong covariance
typically being associated with hydrologically-closed systems (Talbot, 1990).

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

Habitat, pH, conductivity, ionic composition and nutrient levels are all major controls on the
presence and abundance of diatom species, with some species having well established
preferences in relation to some, or all, of these factors (e.g. Gasse, 1986; Kilham et al.,
In La Piscina de Yuriria, the dominant controls over the full record appear to be those associated with changes in evaporative concentration (pH, EC, ionic composition). NaCl and Na$_2$CO$_3$ waters have been shown to be particularly aggressive in relation to diatom dissolution (Barker et al., 1994), so it is likely that some species may be over-represented in the sediment record due to their robust form and heavy silicification. This seems particularly likely in the case of the cysts of *C. muelleri* and may also apply to the more robust forms of species such as *A. costata*, *Denticula elegans* and *Rhopalodia gibberula*. Although the impact of differential preservation needs to be borne in mind when interpreting the fossil record, saline-lake diatoms are well established as indirect tracers of climate change (Gasse et al., 1997).

From the base of YC2, which dates to a little before 30,000 aBP, until around 27,500 aBP, the $\delta^{18}$O$_{\text{carb}}$ record shows marked variability, with the lowest values equivalent to the minimum for the sequence as a whole, and the highest values close to the maximum (Fig. 4). There is strong covariance between $\delta^{18}$O$_{\text{carb}}$ and $\delta^{13}$C$_{\text{carb}}$ (Fig. 5) consistent with a hydrologically-closed lake undergoing temporal variations in the degree of evaporative enrichment (Talbot, 1990) in response to changes in effective moisture. The most negative $\delta^{18}$O$_{\text{carb}}$ values equate to un-evolved lake water that was probably fed by springs and rainfall with lower $\delta^{18}$O than at present ($\delta^{18}$O around -9 ‰ in summer 1992: Table S1), possibly coupled with cooler conditions. In contrast, the most positive $\delta^{18}$O$_{\text{carb}}$ values in this interval are best explained by evaporative enrichment under reduced effective moisture. The lowest $\delta^{13}$C$_{\text{carb}}$ can be explained by equilibration with atmospheric CO$_2$ whereas the more positive values require other, or additional, mechanisms to explain them. The uptake of $^{12}$C during aquatic photosynthesis by aquatic macrophytes or algae can lead to an increase in the $\delta^{13}$C$_{\text{DIC}}$ and hence of endogenic carbonates, but such an explanation appears inconsistent with low TOC values in this interval (Fig. 3). An alternative
explanation is the formation of co-genetic, $^{13}$C-enriched, CO$_2$ during methane formation. Despite the evidence pointing to low lake levels at this time, methane formation in shallow and eutrophic lakes has previously been reported in Mexico (Lake Pátzcuaro: Metcalfe et al., 2007) and elsewhere (e.g. Lamb et al., 2000; Gu et al., 2004). The high concentrations of L. sappaensis in this interval are consistent with the existence of a saline-alkaline lake (Forester, 1986): interestingly, the peaks in ostracod abundance coincide broadly with the peaks in $\delta^{18}$O$_{\text{carb}}$, suggesting that ostracod numbers increased with salinity, and hence evaporative enrichment. The occurrence of siliceous nodules within the interval, associated with the levels that have the highest $\delta^{18}$O$_{\text{carb}}$ values, is also consistent with the existence of highly saline and alkaline water. The sporadic isotope values from ostracod shells in this interval show $^{18}$O-enrichment compared with endogenic carbonates that is broadly consistent with the approximate +0.7 ‰ vital offset recorded for the genus *Limnocythere* (von Grafenstein et al., 1999). In contrast, the carbon isotope values in ostracod shells are similar to those in endogenic carbonate, suggesting that both sources of carbonate precipitated from DIC with a similar $\delta^{13}$C value once allowance has been made for differences in the timing and exact location of formation. This period covers diatom zones YC2-I and part of YC2-II (Fig. 6). The base of the core is dominated by *C. muelleri*, a diatom known to inhabit chloride-rich waters, but the other taxa here do not indicate hypersaline conditions, so it may be over-represented. The most negative $\delta^{18}$O$_{\text{carb}}$ value may be reflected in the increase in the freshwater *N. molestiformis* and *N. fluens* although the diatom assemblage overall continues to indicate shallow and alkaline conditions. The presence of *Nitzschia palea* and a form of *Nitzschia frustulum* (both obligate N heterotrophs) also indicates eutrophic conditions, which may help to explain methanogenesis in shallow water conditions (see above). The increasing abundance of *N. elkab*, and species of *Anomoeoneis* indicate more consistently high pH (> 8.5) and alkalinity, probably associated with shallowing of the lake. Low magnetic susceptibility and
low concentrations of ‘inwash’ indicator elements (Fe, Mn, K and Al) in this interval suggest that inwash of soil into the lake was limited despite the intervals of increased effective moisture (Fig. 3).

Between ~27,500 and 14,000 aBP, there was a shift to more positive $\delta^{18}O_{\text{carb}}$ and $\delta^{13}C_{\text{carb}}$ values, although with some stratigraphical variability and strong covariance between $\delta^{18}O_{\text{carb}}$ and $\delta^{13}C_{\text{carb}}$ (Fig. 4, Fig. 5). The elevated $\delta^{18}O_{\text{carb}}$ values suggest enhanced evaporative enrichment of lake water under conditions of low effective moisture: the presence of siliceous nodules at various points in this interval support the argument that the lake was shallow, saline and strongly evaporated. Short-term variations in carbonate content also support the occurrence of strong but variable evaporative enrichment (Fig. 3).

The presence of multiple surfaces that probably resulted from desiccation in the later part of this interval suggests that lake levels fell and the lake may have dried out totally on several occasions. Elevated $\delta^{13}C_{\text{carb}}$ values are too high to be explained solely by equilibration with atmospheric CO$_2$. Enhanced aquatic productivity, in which $^{12}C$-uptake by aquatic plants and algae causes DIC to be enriched in $^{13}C$, is incompatible with the low TOC content in this interval: the production of $^{13}C$-enriched co-genetic CO$_2$ in a stagnant, shallow, nutrient-rich lake could provide an alternative explanation as discussed above.

From 27,500 to 22,500 aBP, a period of evaporative enrichment and periodic desiccation is consistent with the diatom record for this interval (zones YC2-IIa and YC2-IIb), which ends around 22,500 aBP in a period when diatoms were sparse and poorly preserved (after which there is a break in diatom preservation, see above). There are two notable peaks in *N. minusculoides* reaching 71% of the count between 27,000 and 26,000 aBP and 44% at around 24,000 aBP. The earlier peak is associated with a layer of sand (or tephra, see below) and, moreover, there are no stable isotope data from this layer, so its significance remains unclear. The presence of some freshwater taxa (e.g. *N. fluens*,...
Caloneis bacillum) may reflect fluctuating conditions within a period of overall drying. There are several marked peaks in ostracod abundance during this interval, with assemblages strongly dominated by Limnocythere sappaensis, which supports the inference that the lake was generally saline and alkaline. The sporadic occurrence of other taxa suggests the periodic influx of fresher waters: rather than whole-lake freshening, these taxa could indicate surface or subsurface inflow of fresh water at various times. Low magnetic susceptibility values and low concentrations of ‘inwash’ elements suggest limited transfer of soil or sediment from the catchment during this interval; two sharp peaks in magnetic susceptibility around 26,000 and 23,000 aBP (Fig. 3) may represent tephra, rather than catchment inwash, although this remains to be confirmed.

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

After 11,000 aBP, there was a shift to more negative and also more variable $\delta^{18}O_{\text{carb}}$ and $\delta^{13}C_{\text{carb}}$ values, consistent with a general increase in effective moisture during the early Holocene (Fig. 4). The rise in magnetic susceptibility, coupled with minor increases in some of the ‘inwash’ elements, may also reflect enhanced inwash of catchment material during the early Holocene. Diatom preservation resumes (zone YC2-III), with assemblages dominated by Nitzschia palea and Chaetoceros muelleri (Fig. 6). Assuming that C. muelleri may be overrepresented, high percentages of N. palea and Nitzschia communis indicate lower pH and TDS (total dissolved solids) (Gasse, 1986) and eutrophic conditions.
This assemblage shows some similarities to that in zone YC2-Ia, but zone YC2-III may represent the period when the lake was freshest and deepest, although not deep enough to develop a truly planktonic flora. Today, assemblages with such high percentages of *N. palea* are found in shallow, freshwater lakes in Mexico with high levels of nutrient enrichment such as Lakes Zacapu and Cajititlan (Hill, 2006 and S. Metcalfe, unpublished data). Ostracod concentrations are low, but significantly the assemblages include a relatively high proportion of species other than *L. sappaensis*, consistent with fresher water than in much of the pre-Holocene.

Between about 8,000 and 4,500 aBP there was a marked change in the lake system. Increase in $\delta^{18}O_{\text{carb}}$ values suggests enhanced evaporative enrichment associated with decreased effective moisture (Fig. 4). Most notably, however, is the dramatic positive excursion in $\delta^{13}C_{\text{carb}}$ values, up to a maximum of about +16 ‰ and, associated with this, the breakdown in the positive covariance between $\delta^{18}O_{\text{carb}}$ and $\delta^{13}C_{\text{carb}}$ values that was apparent during earlier intervals (Fig. 5). The very high $\delta^{13}C_{\text{carb}}$ values are best explained by methanogenesis, as discussed for earlier intervals above. Interestingly, the single $\delta^{13}C_{\text{ostracod}}$ value from this interval does not track the $\delta^{13}C_{\text{carb}}$ values, but instead is much lower (Fig. 4). Although we cannot attach too much significance to a single value, this does suggest that the ostracods and the endogenic carbonate were formed in different micro-environments within the lake, or perhaps during different seasons, from DIC with contrasting $\delta^{13}C$ values. Although the $\delta^{18}O_{\text{ostracod}}$ values are more positive than the $\delta^{18}O_{\text{carb}}$ values, as would be expected, the difference is too large to be explained by vital offsets alone, possibly lending support to the view that the endogenic and ostracod carbonates were formed under contrasting conditions or at different times of the year. Ostracods occur sporadically in this interval, indicating that whatever conditions prevailed were not wholly unsuitable for ostracods to live. This period straddles the diatom record at the top of YC2.
(zone YC2-IV) and the bottom of YC1 (zone YC1-I). The diatom assemblage is notable for its dominance by *N. (C.) elkab* and *N. (C.) halophila* (both cores), with *Nitzschia frustulum* and *N. palea* (Fig. 6). The switch to an *N. elkab/N. halophila* flora is consistent with a return to more alkaline conditions, probably driven by increasing evaporation. *Navicula (C.) elkab* does seem to have a distinct ecology, with a preference for hyper-alkaline, Na$_2$CO$_3$ lakes, where Cl$^-$ is also important. *Nitzschia frustulum* tends to be more abundant with *N. (C.) elkab* than with *N. (C.) halophila* (e.g. YC2-4c and base of YC1-1) supporting the interpretation of high alkalinity (Gasse, 1986). Its presence with *N. palea* again seems to indicate high levels of nutrient enrichment. Overall, this assemblage is quite similar to that sampled from the modern lake in 1982, when it was around 2 metres deep. A peak in magnetic susceptibility between about 5,500 and 4,500 BP (Fig. 3) associated with a sharp reduction in $\delta^{18}$O$_{carb}$ (Fig. 4) points to a climatically-controlled inwash event, which may be reflected by rather poor diatom preservation in YC1.

After 4,500 aBP, $\delta^{18}$O$_{carb}$ remained high although with some short-lived negative excursions, whereas $\delta^{13}$C$_{carb}$ values are reduced dramatically (Fig. 4). For the interval of overlap between YC1 and YC2, there is good agreement between the isotope records, especially so for $\delta^{13}$C$_{carb}$, once allowance is made for the slight age difference between the two cores, as discussed earlier. There is a large increase in ostracod concentration within much of this interval: the dominance of assemblages by *L. sappaensis*, coupled with the large number of individuals, is consistent with the lake having been saline and alkaline for much of the time (Fig. 4). $\delta^{18}$O$_{ostracod}$ values are $^{18}$O-enriched compared to $\delta^{18}$O$_{carb}$ values, with the degree of enrichment consistent with vital offsets from oxygen-isotope equilibrium, as discussed above. The $\delta^{13}$C$_{ostracod}$ values agree well with those for $\delta^{13}$C$_{carb}$, suggesting that the two sources of carbonate formed under the same set of conditions, as was the case for the interval prior to the positive $\delta^{13}$C$_{carb}$ excursion. The diatom record for this
interval was published by Metcalfe and Hales (1994), although here we plot the data
against our new age model (Fig. 6). The species encountered confirm the presence of a
shallow alkaline lake throughout the late Holocene, although the balance between CO$_3^{2-}$
and Cl$^-$ seems to have varied. The assemblage in zone YC1-II, for example, may indicate
that Cl$^-$ replaced CO$_3^{2-}$ as the dominant anion. The assemblage is similar to Bradbury's
(1989) saline marsh group. Fresher conditions were then re-established, associated with
inwash from the catchment. This wetter/drier cycle is then repeated in zones YC1-III and
YC1-IV. Increasingly hostile conditions for diatom preservation are indicated through zone
YC1-V, with only a patchy diatom record between about 2000 and 1200 aBP. It seems
likely that the single count available through this period is significantly affected by
differential preservation. The diatom record resumes around 1000 aBP with a distinctive,
well-preserved sample dominated (55%) by *Navicula muralis* (zone YC1-VII). This diatom
is often found on mud flats and amongst aquatic vegetation (Hustedt, 1961-66). When
combined with very low magnetic susceptibility values, the diatom assemblages indicate
catchment stability. We note that this catchment stability occurs at a time in the late
Classic when many sites in the relatively dry parts of Central Mexico were abandoned
(Beekman, 2010; Park *et al.*, 2010). The most recent sediments preserve a flora quite
similar to that found in the various surface sediment samples, indicating a shallow alkaline
lake, but with increasing nutrient levels. The very highly evolved chemistry and rather
distinctive diatom flora of La Piscina de Yuriria is described in Davies *et al.* (2002) and
sampling in 2003 and 2004 confirmed its hypereutrophic status. Phases of inwash,
previously reported in YC1 and attributed to anthropogenic disturbance (Metcalfe *et al*.,
1994), are also seen at broadly the same times in YC2, i.e. around 3,600 and 1500 aBP
and over the past few centuries. An earlier phase at the base of YC1, which ended in that
core around 4700 aBP, appears in its entirety in YC2, starting around 5600 aBP. The
absence of *Z. mays* pollen from YC1 during this interval led Metcalfe *et al.* (1994) to
suggest that the inwash was climatically-mediated rather than the result of anthropogenic
disturbance. For the wider region however, Park et al. (2010) suggest that agricultural
activity began as early as around 5700 aBP (based on the presence of Zea pollen), and
expanded around 3000 aBP. Lozano et al. (2013) also report the occurrence of Z. mays
pollen from 3000 aBP in Lake Zirahuen, a lake previously thought to be un-affected by
human impact. The main climatic and environmental changes revealed in the records from
La Piscina de Yuriria are summarised in Table 2.

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

The number of palaeoclimatic records from the TMVB (and indeed the whole of Mexico),
extending back to 30,000 BP (Marine Isotope Stage 3) is small and limited to the Basin of
Mexico (e.g. Caballero and Ortega-Guerrero, 1998; Roy et al., 2009; Lozano-García et al.,
2015), Lake Cuitzeo (Israde et al., 2010), Lake Pátzcuaro (Watts and Bradbury, 1982;
Bradbury 2000) and Lake Zacapu (Correa-Metrio et al., 2012). Moreover, most of these
have limited dating control for the older sediments. Caballero et al. (2010) integrated
palaeolimnological records with evidence from glacial chronologies (Vazquez Selem and
Heine, 2004) to provide a palaeoclimatic scenario for the period from around 30,000 aBP
to the last glacial maximum (LGM). A more recent consideration of the record from Lake
Chalco in the Basin of Mexico has been published by Lozano-Garcia et al. (2015). With
the exception of Lake Pátzcuaro (one of the westernmost sites), the records suggest drying and cooling after about 30,000 aBP, with glacial advances restricted until around 22,000 aBP. The data from La Piscina de Yuriria suggest that there were rapid shifts between wet and dry conditions during this interval, with a general drying trend sometime between about 27,500 and 25,000 aBP. However, the relatively low resolution of the record after about 27,500 aBP means that a continuation of abrupt shifts in rainfall cannot be ruled out. Major glacial advances in the TMVB occurred around the LGM (22,000 – 18,000 aBP) with a suggested 6 – 8°C lowering in mean annual temperature. Reduced temperature, along with low effective moisture, could have contributed to the elevated δ^{18}O_{carb} values at this time. Although generally dry, there seem to have been short lived lake highstands in Cuitzeo and rising water levels in the Chalco Basin (southern Basin of Mexico) and the Lerma Basin around 18,000 and 19,000 aBP respectively (Caballero et al., 2002). Unfortunately the evidence from La Piscina de Yuriria is at too low resolution to be able to detect these highstands. The late glacial in the TMVB (18,000 – 15,000 aBP) is described as cold and dry, with minor glacial recession (Caballero et al., 2010): evidence from La Piscina de Yuriria indicates a continuation of dry conditions through the late glacial (=Younger Dryas) stadial. Significant warming and glacial retreat apparently started around 14,000 aBP.

Overall, the records of cool and dry conditions around the LGM and into the early Holocene from the central and eastern part of the TMVB are most easily explained by a reduction in summer season precipitation, which is driven today by the northward movement of the ITCZ and the onset of the NAM. Records from the Pátzcuaro Basin, to the west of this group, show high lake levels persisting through the LGM (Bradbury 2000; Metcalfe et al., 2007), while Correa-Metrio et al. (2012) suggest conditions moist enough for pine forest to dominate over grasslands in the Zacapu basin. In both cases an increase
in winter precipitation may provide an explanation although it is hard to reconcile this with the stable-isotope data from La Piscina de Yuriria, which suggest an overall decrease in effective moisture. Lachniet et al. (2013) have further suggested that the summer monsoon did not collapse during the last glacial, although once again the evidence from La Piscina de Yuriria does not appear to support this, nor does the latest interpretation of the Chalco record (Lozano-Garcia et al., 2015). The late glacial stadial was a time of dry conditions, with some suggestion that La Piscina de Yuriria may have dried out. A dry Younger Dryas stadial has also been reported from Zacapu (Correa-Metrio et al., 2012) and from sites in northern Mexico (e.g. Roy et al., 2013). Interestingly, the Younger Dryas is also reported as dry in the Juxtlahuaca speleothem record (Lachniet et al., 2013) where it is attributed to monsoon collapse. It appears that the weakening of the Atlantic Meridional circulation (AMOC), the subsequent southward displacement of the ITCZ and a weaker monsoon led to dry conditions over the northern hemisphere neotropics (Bush and Metcalfe, 2012). Only at the northern edge of the NAM region did the resumption of cold conditions allow more penetration of mid-latitude westerlies leading to wetter conditions (Metcalfe et al., 2015).

The classic pattern of climatic change in the NH tropics and subtropics is for wetter conditions in the early Holocene driven by poleward migration of the ITCZ and a stronger monsoon in response to insolation forcing. Whilst there is some support from this in Mexico, it seems that the establishment of the modern climatic regime was delayed by the presence of the residual Laurentide Ice Sheet and the influence of meltwater pulses entering the Gulf of Mexico (Metcalfe et al., 2015). The early Holocene interval in La Piscina de Yuriria indicates a change to wetter conditions overall, but with evidence for abrupt shifts between wet and dry conditions. A similar pattern is recorded by Park et al. (2010) for their other sites in the Valle de Santiago. The response of the wider NAM region...
to the insolation maximum in the early Holocene is complex, with the clearest response in
the south, where the direct influence of the ITCZ is strongest. Elsewhere, it seems that the
modern NAM regime may not have resumed until after 8000 aBP and there was a trade-off between increasing precipitation and increasing temperatures (Metcalfe et al., 2015).

La Piscina de Yuriria shows an overall trend of drying from the mid Holocene onwards.

This is consistent with the southward migration of the ITCZ and as the role of insolation
forcing became weaker, so the effect of other climate forcings such as ENSO, seems to
have become more important giving rise to increasingly complex patterns of change. This
drying was accompanied by increasing human impact, as shown by evidence for phases
of sediment inwash and by the presence of Z. mays pollen. Increasing human impact
during this interval is also evident from other sites in Mexico (Metcalfe et al., 1994).

Conclusions

In summary, evidence from La Piscina de Yuriria indicates that the climate of Central
Mexican highlands has changed dramatically over the past ~30,000 years. Between
30,000 and about 27,500 aBP it was highly variable with shifts, which may have been
abrupt, between dry and wet conditions. Over much of the glacial period, from ~27,500 to
about 14,000 aBP, climate became drier: there may have been abrupt shifts during this
interval, but the low resolution of our data means that any such shifts are not revealed.
The occurrence of strong millennial scale variability during MIS3, with a global signature,
has been widely noted (Clement and Peterson, 2008), apparently driven by changes in
AMOC. It is notable that there were three D/O warming events (2 – 4) between 30,000 and
22,000 aBP (Wolff et al., 2010) and two Heinrich events (H2 and H3). Modelling has
indicated differential sensitivity of AMOC under MIS3 and LGM conditions (e.g. Van
Meerbeeck et al. 2009), with the climate becoming less sensitive to AMOC changes as the
full glacial climate was established. The growth of the Laurentide Ice sheet from an
interstadial minimum around 35,000 aBP to its maximum by ca. 25,000 aBP (where it remained until around 15,000 aBP) (Dyke et al., 2002) also reflects the shift of the global climate system into full glacial mode, where other, less rapid forcings may have dominated (see Baker and Fritz, 2015). There is evidence of drought during the period that encompassed the late glacial stadial. During the Holocene, the climate initially became wetter, although the positive water balance was insufficient to lead to major changes in the lake’s chemistry and diatom flora. A fall in lake levels under drier climate in the mid Holocene was accompanied by a change in limnology that caused methane formation. Inwash of catchment soils and sediments was the result of a combination of natural climatic triggers and, for the later Holocene, anthropogenic disturbance.

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**Figures**

Fig. 1. Location of la Piscina de Yuriria and other sites referred to in the text.
Fig. 2. Age-depth relationships for (a) YC1 (b) YC2, based on radiocarbon dates. The datapoints are well described by 3rd order polynomial curves, \( \text{AGE} = -0.0001\text{Depth}^3 + 0.06645\text{Depth}^2 + 2.473\text{Depth} - 29.23 \) for YC1 and, \( \text{AGE} = -0.000005\text{Depth}^3 - 0.0006\text{Depth}^2 + 12.23\text{Depth} - 109.1 \) for YC2 where age is in calendar years BP and depth is in cm in both cases. Detailed synchronisation for YC1 and the upper part of YC2 based on (c) loss-on-ignition and (d) magnetic susceptibility.
Fig 3. Physical sedimentology and selected elemental geochemical variables for YC1 and YC2, plotted as a function of age in calendar years.

Fig. 4. Ostracod assemblages for YC2 and stable isotopes: endogenic carbonate for YC1 and YC2 and ostracod shells (triangles) for YC2. The ‘other taxa’ percentage curve includes all taxa except *Limnocythere sappaensis*.
Fig. 5. Covariance of $\delta^{18}O_{\text{carb}}$ and $\delta^{13}C_{\text{carb}}$ values for YC1 and YC2. For the interval YC2 7700-30,300 aBP only, there is significant covariance between $\delta^{18}O_{\text{carb}}$ and $\delta^{13}C_{\text{carb}}$ values ($R^2 = 0.68$).

Fig. 6. Diatom abundance (%) for (a) YC2 and (b) YC1. Only taxa present at > 2% and in more than one sample are plotted.
Fig. 7. Ostracod abundance and magnetic susceptibility for the last 10,000 years in core YC2.
# Tables

## Table 1. Radiocarbon dates from cores YC1 and YC2.

| Core  | Depth (cm) | Date type | Lab number | Radiocarbon | Cal age | Material | Notes |
|-------|------------|-----------|------------|-------------|---------|----------|-------|
|       |            |           |            | age (yr)    | range (yr) |          |       |
| YC1   | 44-46      | AMS       | 1x4A-118K1 | R/HR        | 80/100   | 673      | omitted from age model |
| YC1   | 76-78      | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | carbonates |
| YC1   | 122-123    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | carbonates |
| YC1   | 157-158    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | carbonates |
| YC1   | 159-160    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | carbonates |
| YC1   | 250-255    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | carbonates |
| YC1   | 251-252    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | carbonates |
| YC1   | 500-595    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | carbonates |
| YC2   | 126-127    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 216-217    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 321-322    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 450-440    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 522-332    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 652-402    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 682-522    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 702-682    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 802-702    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 692-682    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 652-402    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 682-522    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 702-682    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 802-702    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 652-402    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 682-522    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 702-682    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 802-702    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 652-402    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 682-522    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 702-682    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 802-702    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 652-402    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 682-522    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 702-682    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 802-702    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 652-402    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 682-522    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 702-682    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |
| YC2   | 802-702    | AMS       | 1x1A-118K1 | R/HR        | 80/100   | 673      | bulk organic carbon |

(95 % confidence intervals)
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-contorted 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 desiccation          | 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 |
### Supporting online information

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

| Year     | Water depth (m) | Type      | pH  | EC*  | Alkalinity (Total) | Cl  | SO₄ | K   | Na   | Ca   | Mg   | Fe  | δ¹⁸O H₂O | δ¹⁸O VSMOW | δ¹⁰N VSMOW |
|----------|-----------------|-----------|-----|------|--------------------|-----|-----|-----|------|------|------|-----|---------|-----------|------------|
| 1992 (April) | 2 (surface)     | Lake      | 11  | 20000 | 300.0             | 17.4 | 29.3 | 30.7 | 47.05 | 0.0  | 0.0  | ND* | ND      | ND        | ND         |
| 1992 (April) | 2 (at depth)    | Lake      | 11  | 20000 | 326.0             | 14.1  | 30.2 | 24.8 | 47.88 | 0.0  | 0.0  | ND  | ND      | ND        | ND         |
| 1992 (April) | 0.2 (margin)    | Lake      | 11  | 20000 | 324.0             | 16.3  | 29.5 | 32.1 | 41.33 | 0.0  | 0.0  | ND  | ND      | ND        | ND         |
| 1992 (April) |                | Spring    | 7.5 | 300   | 6.7               | 0.4  | 1.4  | 0.6  | 3.7   | 1.0  | 1.3  | ND  | ND      | ND        | ND         |
| 1992 (April) |                | Spring    | 7.0 | 100   | 5.0               | 1.1  | 3.6  | 1.0  | 9.8   | 1.1  | 1.6  | ND  | ND      | ND        | ND         |
| 1992 (May)  | 0.2 (margin)    | Lake      | 9.4 | 27500 | 222.0             | 20.5  | 22.1 | 54.2 | 7.0   | 0.0  | 0.0  | ND  | ND      | ND        | ND         |
| 1992 (May)  |                | Spring    | 7.7 | 700   | 8.0               | 4.2  | 0.3  | 0.6  | 8.0   | 0.0  | 1.6  | ND  | ND      | ND        | ND         |
| 1992 (August) | 0.2       | Lake      | 10.5 | 15000 | 248.2             | 117.4 | 50.0 | 22.8 | 298.5 | 0.3  | 0.1  | ND  | -0.2    | 0.7       | -35        |
| 1992 (August) |            | well     | 6.6 | 315   | 3.7               | 0.4  | 9.6  | 0.9  | 5.7   | 4.4  | 3.9  | ND  | -10.4   | -0.2      | -69        |
| 1992 (August) |            | Spring   | 8.0 | 452   | 7.7               | 0.1  | 7.6  | 2.0  | 8.1   | 3.3  | 3.7  | ND  | 11.6    | 9.1       | 69         |
| 1994 (March) | shore         | Lake      | 10.2 | 15000 | 322.3             | 21.3  | 19.2 | 1.0  | 95.0  | 5.9  | 5.9  | ND  | ND      | ND        | ND         |
| 2003 (March) | 0.7 (margin)   | Ion      | 7.5 | 1577  | 26.8              | 6.7  | 6.4  | 1.7  | 26.3  | 2.0  | 2.4  | ND  | ND      | ND        | ND         |
| 2003 (March) | 0.2 (margin)   | Lake      | 9.6 | 2910  | 16.1              | 16.7  | 0.8  | 1.5  | 163.4 | 2.6  | 0.0  | ND  | 71.5    | ND        | ND         |
| 2003 (March) |              | Groundw  | 0.0 | 1070  | ND                | ND   | ND  | ND  | ND   | ND   | ND   | ND  | ND      | ND        | ND         |

*EC* = Electrical Conductivity
*ND* = Not determined
*δ* = Per mil deviation from SMOW

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

### (a) Core YC1

| Variable | $R^2$ | $p$-value | $H_0$ (5%) |
|----------|-------|-----------|-------------|
| K        |       |           |             |
| Fe       | 0.65  | 8.03E-18  | rejected    |
| Mn       | 0.77  | 9.80E-06  | rejected    |
| Al       | 0.67  | 2.40E-17  | rejected    |

### (b) Core YC2 - all

| Variable | $R^2$ | $p$-value | $H_0$ (5%) |
|----------|-------|-----------|-------------|
| K        |       |           |             |
| Fe       | 0.05  | 0.00681   | rejected    |
| Mn       | 0.45  | 4.00E-17  | rejected    |
| Al       | 0.37  | 6.67E-07  | rejected    |

### (b) Core YC2 - post 5000 aBP

| Variable | $R^2$ | $p$-value | $H_0$ (5%) |
|----------|-------|-----------|-------------|
| K        |       |           |             |
| Fe       | 0.61  | 6.64E-08  | rejected    |
| Mn       | 0.75  | 7.15E-11  | rejected    |
| Al       | 0.0   | 1.17E-11  | rejected    |