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Investigating the potential of oxygen-isotope records from anthropogenic lakes as tracers of 20th century climate change

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Abstract

Historical climate change in southern England was investigated using ostracod oxygen-isotope ($\delta^{18}O$) records from two anthropogenic lakes in Hampshire, southern England. A strong relationship is observed between $\delta^{18}O_{ostracod}$, $\delta^{18}O_{precipitation}$ and $\delta^{18}O_{lake\_water}$ in the contemporary environment and therefore $\delta^{18}O_{ostracod}$ from the sedimentary record of these systems has the potential to reflect past climate variability. The possibility of these sites to act as archives of climate change through $\delta^{18}O_{ostracod}$ analysis is explored through the study of lake sediment cores that cover the period from the early 20th century onwards. Both lakes showed similar directionality of shifts in $\delta^{18}O_{ostracod}$ over this period, suggesting common driving mechanisms. Comparing $\delta^{18}O_{ostracod}$ timeseries to meteorological data is challenging in part because of the complexity with which climate parameters are recorded in the $\delta^{18}O_{lake\_water}$ and consequently within lacustrine carbonates. Our findings highlight the potential of sediments from anthropogenic lakes to act as archives of past climate and indicate they may be an important resource for generating climatic reconstructions across the medieval to instrumental period, which the sediments of many anthropogenic lakes cover. Such climate reconstructions would greatly improve our spatial and temporal understanding of climate variability where instrumental data are unavailable and other natural archives are scarce.

Keywords

anthropogenic lakes, climate change, historical, Late-Holocene, ostracods, oxygen isotopes

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Introduction

Investigating climate variability over the past millennium is critical for understanding the mechanisms and impacts of climate changes under scenarios of changing forcing factors (i.e. natural vs anthropogenic). However, obtaining paleoclimatic records for this period is challenging, not least because of the limited availability of archives that are able to produce high-resolution, chronologically well-constrained climate reconstructions that extend beyond the limit of instrumental datasets. The Medieval Climate Anomaly (MCA: ca. 950–1400 CE) and Little Ice Age (LIA: ca. 1200–1850 CE) are known natural climatic events during this time-frame that were driven by a range of decadal-scale variables. These include solar variability, volcanism and atmospheric controls and feedbacks, such as the North Atlantic Oscillation, when many controls on the Earth system (i.e. ice volume, orbital configuration, sea level, etc.) were much the same as the present day, including the influence of anthropogenic activity (Diaz et al., 2011; Goosse et al., 2012; Nesje and Dahl, 2003; Owens et al., 2017; Ruddiman, 2003; Trouet et al., 2009, 2012; Wanner et al., 2011). Despite the importance of these oscillations for understanding the scale and impact of such climate variability, there are limited high-resolution climate reconstructions that cover this time period, even in well studied areas such as northwest Europe. Study of the post-medieval to instrumental period is hindered by a lack of natural environmental archives and a greater reliance on documentary evidence, which is inconsistent in terms of its detail and continuity despite its contextual importance (e.g. Brönnimann et al., 2019; Burgdorf, 2022). Thus, there is a need for additional environmental archives across the last millennium that should ideally meet the following conditions: (1) have the potential for high-resolution (ideally decadal or better) proxy reconstruction in order to capture short-timescale climate variability; (2) contain a range of proxies, enabling human impacts and catchment processes to be distinguished from regional climate signals; and (3) a robust chronology.

Lake sediments have the potential to meet these requirements, but natural lakes covering the post-medieval onwards are quite rare in northwest Europe for a host of reasons. For example, many

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natural lakes that formed during the last glacial period have infilled and are no longer extant by the Late-Holocene. Equally it is likely that sedimentation that covers this period is strongly influenced by anthropogenic landscape management as much as by climate variability and untangling these impacts is challenging. Tree rings and speleothems have the potential to fill this temporal gap, at high resolution (potentially annually), but they too have limitations. For instance, to achieve interdecadal climate variability from tree rings requires tree-ring width data to be statistically treated and their chronologies sometimes produce unclear climate data (Dobrovolný et al., 2018; Freund et al., 2023). Furthermore, speleothems are restricted to regions of carbonate bedrock. Consequently, there are large spatial gaps in climate records for the past millennium, limiting our understanding of spatial variation in climate change and its impact over this period. The British Isles has an abundance of ornamental lakes often associated with period estates. These are anthropogenic lakes that became popular in garden design from the early 18th century in England with 450 on National Trust property in the UK as well as 1048 legacy Medieval fishing ponds listed by Historic England (2023). Similar anthropogenic lakes also exist across Europe (Bishop, 2021; Freshwater Habitats Trust, 2022). Although the latter may not all have extant water, the numbers show the huge opportunity to explore these features. Such lakes may be able to provide records of past climate across the historical period, helping to overcome the gaps in spatial coverage. This, alongside the variety of proxies applicable in lake settings and the likelihood of historical documentary evidence on such lakes recording periods of human disturbance, suggests anthropogenic lakes may be an important archive. However, thus far no studies have investigated the potential of the use of such sites as the basis for paleoclimate archives. The aim of this paper is to evaluate the potential of these anthropogenic lakes as archives of past climate through a study of two sites in Hampshire, southern England, both of which are located on chalk geology, namely Old Alresford Pond and The Grange Lake.

Oxygen isotopes in lake carbonates as a paleoclimate proxy

Oxygen isotopes in lake-sediment carbonates are a potential climate proxy in anthropogenic lakes. Their use as a climate proxy has been widely demonstrated in natural lakes in humid-temperate mid-latitude regions (Leng and Marshall, 2004). In such settings, the oxygen-isotope composition of endogenic or biogenic carbonate is a function of the oxygen-isotope composition of lake water ($\delta^{18}O_{lake\_water}$) from which the carbonate precipitates and the temperature at which the carbonate mineralises (Leng and Marshall, 2004). The $\delta^{18}O_{lake\_water}$, in turn, is controlled by the isotopic composition of inflows (i.e. rainfall, surface inflow and groundwater) together with any in-basin processes that lead to fractionation, particularly evaporation. In lakes with a high rate of recharge and a short water residence time, evaporitic modification may be minimal, meaning that the lake water isotope composition is close to that of precipitation ($\delta^{18}O_{precipitation}$). The isotope composition of precipitation is controlled by a range of factors including air temperature, and the source, transport pathway and rain-out history of rain-bearing air masses (Dansgaard, 1964; Rozanski et al., 1992, 1993). Providing the controls on the oxygen isotopic composition of lake sediment carbonates can be adequately constrained, such records therefore have the potential to provide us with information about climate. The shells of ostracods (small aquatic crustaceans that secrete shells of calcite) are good sources of carbonate for stable-isotope analysis. Some species have well-defined seasonality, so that the oxygen isotope signal can be tied to a specific time of year. Ostracods do not form shells that are in isotopic equilibrium with water, but the offsets are well understood and precisely quantified for some taxa (Holmes and Chivas, 2002; von Grafenstein et al., 1999). Finally, by using ostracod shells for isotope analyses ($\delta^{18}O_{ostracod}$), we can be sure that the calcite was produced within the lake and does not include any catchment-derived (or exogenic) material.

Here we investigate how well $\delta^{18}O_{ostracod}$ reflects prevailing climate, including seasonal variables in the present day as a baseline for our historical reconstruction by monitoring lake water temperature and $\delta^{18}O_{lake\_water}$ composition, supported by meteorological and rainfall-isotope data. Isotope analyses of modern ostracods from each lake were undertaken to understand how $\delta^{18}O_{lake\_water}$ and water temperature is encoded in the ostracod shell calcite. Isotope analyses of sub-fossil ostracods sampled from lake sediment cores from each lake were then undertaken to produce a historical climate record from this part of southern England. Although our primary focus is on oxygen-isotopes, we use the carbon-isotope ($\delta^{13}C$) records that are also produced as supporting information. The $\delta^{13}C$ of ostracod shells records the carbon-isotope composition of dissolved inorganic carbon (DIC), which in turn is controlled by the $\delta^{13}C$ of groundwater, together with in-lake modification. Stuart and Smedley (2009) report $\delta^{13}C$ value of −17.4 to −9.36 ‰ in groundwater of the Hampshire Chalk, with the more ^{13}C-deplete values relating to dominance of solid-derived carbon and the least deplete to carbon from bedrock source. Enhanced uptake of ^{12}C during aquatic photosynthesis as well as contributions of atmospheric CO_2, will also raise the $\delta^{13}C$ value of lake DIC (Kelts and Talbot, 1990). Clearly, the $\delta^{13}C$ of ostracods within the lakes can be influenced by both regional factors and lake-specific controls.

Our investigation focusses on approximately the last 100 years since this period overlaps in time with instrumental records. Our study also acts as a proof of concept for the use of such constructed lakes to record a climate signal over a longer period, potentially covering much of the past millennium at some sites.

Study sites

Old Alresford Pond (51°5′37.43″N, 1°9′30.92″W) and The Grange Lake (51°7′12.32″N, 1°11′46.66″W) are two constructed, historical, neighbouring lakes lying about 4 km apart in Hampshire, UK (Figure 1). The area is underlain by Cretaceous chalk bedrock and both lakes are fed by chalk streams that have been artificially dammed to generate lakes. The chalk streams are groundwater fed, with their flow greatest in winter and spring (Environment Agency, 2013). Both sites are shallow, with a maximum depth of <1 m. Old Alresford Pond is mostly infilled and extremely shallow (<0.5 m) for much of its area.

Old Alresford Pond

Constructed in ca. 1189 CE, Old Alresford Pond is a former medieval fishing pond that began use ca. 1208–1209 CE (Rollason, 2017; Turpin, 2008). The lake was created by constructing a 6 m high, 76 m long earth dam to produce a reservoir for mills along the River Itchen (Binnie, 1974; Environment Agency, 2013). Today the lake is 0.08 km^2 in area, smaller than its former extent of ca. 0.24 km^2 in the 13th century (Environment Agency, 2020; Roberts, 1986). The lake's water residence time is estimated at less than one day (Wang, 2013). Water sources include surface runoff from the catchment via the stream network, direct precipitation into the lake and groundwater. Boreholes to supply groundwater for the watercress beds upstream of Old Alresford Pond were introduced in the 20th century and contribute to water within the pond. Documentary evidence records dredging of the lake sometime between 1252 and 1254 CE, clearing built up sediment and then deepening one side of the lake in 1254 CE for easier eel fishing (Roberts, 1986).

Figure 1. Location maps. (a) Old Alresford Pond in relation to The Grange Lake with inset showing these in the context of Hampshire. (b) The Grange Lake with core location and flow direction (top) and sampling locations (bottom). (c) Old Alresford Pond sampling and coring locations. This study uses cores ALRE19 and GRA16. Imagery from Google Earth.

However, there is no record of more recent dredging. Peat from the nearby watercress farms was flushed into the lake at times during the 20th century (Tony Chambers and Rosemary Chambers, personal communication).

The Grange Lake

The Grange Lake is an ornamental lake on The Grange Estate, an addition typical of 18th century grounds design, and was built by damming the Candover Brook (also known as the Candover Stream) (Currie, 2001; Deveson, 2005). It is believed the lake was first ordered sometime between 1764 and 1772 CE by Robert Henley, Earl of Northington (Geddes, 1983). It is first seen on Milne's map of Hampshire in 1791 CE as a long, narrow lake and the southern part was later enlarged by Baron Ashburton, Alexander Baring, as shown on the ca. 1817 CE Ordnance Survey and the 1826 CE Greenwood's maps of the area (Norgate and Norgate, 2002a, 2002b, 2002c; Parks and Gardens, 2021). Today, the lake is divided by a waterfall and analyses refer to the 'Upper Lake', from where the core was recovered and some modern water and ostracod samples collected, and the 'Lower Lake' from where only modern water and ostracod samples were collected (Figure 1).

Methods

Modern limnology

Measurements of modern water oxygen and hydrogen isotope composition and temperature at each lake were undertaken to understand the relationship between the oxygen-isotope composition of rainfall and that of lake water and, consequently, how the lake water isotope signal is encoded into ostracod shells from the lake (Figure 1). Measurements and collections were made from May 2019 to November 2020. Samples at Old Alresford Pond were taken in May 2019, August 2019, November 2019, February and November 2020. Samples at The Grange Lake were taken in August 2019, November 2019, February 2020 and November 2020. Sampling had to be paused between February 2020 and November 2020 owing to the COVID-19 pandemic.

Hourly measurements of water temperature were made from August 2019 to November 2020, using TinyTag water temperature sensors installed in both lakes (Figure 1). There is an additional, short, temperature record from May to August 2019 at Old Alresford Pond that was discontinued because of device dysfunction.

Water samples were taken from various locations (Figure 1) at both lakes approximately every 3 months from May 2019 (Old Alresford Pond) or August 2019 (The Grange Lake) to November 2020 to help identify seasonal variations in water isotope composition. On each visit, 50 mL samples were collected in polyethylene bottles, which were sealed with electrical tape, and refrigerated below 5°C until analysis. In addition, an automated water sampler collected water samples approximately weekly from both lakes. These samples were 'capped' by a thin layer of paraffin oil that was placed in each bottle before sampling to minimise any evaporative effect on the water sample following IAEA/GNIP (2014) procedures for precipitation isotope sample collection. During each site visit these samples were transferred to 500 mL bottles, which were capped, sealed with electrical tape, and refrigerated below 5°C until analysis. The paraffin oil was removed before analysis by standing the water in a separation funnel for at least 2 h and allowing the oil to float. The separation funnel was covered to minimise evaporation. Once the oil had been displaced, a 50 mL water aliquot of water was collected in a centrifuge tube, sealed with electrical tape, and refrigerated below 5°C until analysis. Any samples that contained detrital material were additionally filtered through grade 1 filter paper and a grade 1 syringe filter. All water samples were analysed for oxygen and hydrogen isotopes at the British Geological Survey, Keyworth, using a CO_2 equilibrium method with an Isoprime 100 mass spectrometer plus Aquaprep device for oxygen isotope composition. Hydrogen isotopes were analysed on an online Cr reduction method with a EuroPyrOH-311-system coupled to a Micromass Isoprime mass spectrometer. Measurements were made against internal standards that are calibrated against the international standard VSMOW2 and presented in standard delta units. The typical uncertainties are ±0.05 ‰ $\delta^{18}O$ and ±1.0 ‰ for δD.

Ostracod samples were collected from the surface sediments and from a combination of submerged, floating and emergent macrophytes, depending on availability, at both lakes, on each visit using a 250 µm sieve and/or net. Samples were stored in sealed bottles prior to wet sieving through a 250 µm mesh with tap water in the laboratory and then oven drying at ca. 40°C–60°C in a drying cabinet. Specimens of *Candona neglecta* Sars, 1887 and *Candona candida* (Müller, 1776) were picked for oxygen and carbon isotope analysis. These species complete their lifecycles within a year and calcify their adult shells in the winter months, which reduces the influence of any summer evaporation of lake water on the recorded oxygen isotope signal (Decrouy, 2009; Dole-Olivier et al., 2000). The collected valves are likely to be from specimens that died at varying times before their collection and so the isotope signal will represent an average of the last few years. Carapaces in these samples were typically small and not consistently abundant across the samples and so valves were preferentially picked. The species were chosen because they are common in both modern lakes and their sediment records and because the offsets from oxygen-isotope equilibrium are well known (+2.2 ± 0.15 ‰ for members of the family Candondidae: Holmes and Chivas, 2002; von Grafenstein et al., 1999). Moreover, both species have similar, broad ecological tolerances and so can be used interchangeably for isotope analyses (Dole-Olivier et al., 2000; Külköylüoĝlu and Vinyard, 2000). The selected lake sites are so shallow that they are entirely littoral in habitat and therefore habitat variability is not a concern. For oxygen and carbon isotope analysis, samples consisted of multiple shells, representing a time average of the sample core slice, to a mass of between 34 and 100 µg of carbonate. Analyses were performed using an IsoPrime dual inlet mass spectrometer plus Multiprep device. Isotope values (^{13}C, ^{18}O) are reported as per mille (‰) deviations of the isotopic ratios ($^{13}C/^{12}C$, $^{18}O/^{16}O$) calculated to the VPDB scale using a within-run laboratory standard (KCM) calibrated against international standard NBS-19. The calcite-acid fractionation factor applied to the gas values is 1.00813. Due to the long run time a drift correction is applied across the run, calculated using the standards that bracket the samples. The Craig correction (Craig, 1957) is also applied to account for $\delta^{17}O$. The standard calcite values for KCM are for +2.00 ‰ for $\delta^{13}C$, and −1.73 ‰ $\delta^{18}O$ with an average analytical reproducibility of 0.05 ‰.

Air temperature data from 2019 to 2020 CE were obtained from the MetOffice Southampton National Oceanography Centre, Odiham and Otterbourne weather stations (Figure 1), the closest weather stations to the two lakes, to compare with water temperature measurements. Precipitation isotope data ($\delta^{18}O_{precipitation}$) for comparison with the lake water values ($\delta^{18}O_{lake\_water}$) were taken from Wallingford, Oxfordshire (Figure 1) the nearest GNIP (Global Network of Isotopes in Precipitation) station, which lies 69 km north of Alresford. A local meteoric water line (LMWL) was calculated for oxygen and hydrogen isotopes using the precipitation amount weighted least squares regression and given by $\delta D = 7.34$ and $\delta^{18}O_{precipitation} + 4.13$ (IAEA/WMO, 2023). There may be an offset in the isotope composition of precipitation at Wallingford compared to our study sites given the distance between them, but we would expect this to be smaller than the error. We also compare measured $\delta^{18}O_{lake\_water}$ with modelled $\delta^{18}O_{precipitation}$ at Old Alresford Pond using the Piso_AI software (Nelson et al., 2021). This allows a comparative estimate of $\delta^{18}O_{precipitaton}$ at the core location that also extends past the earliest GNIP measurement at Wallingford. The modelled output at both Old Alresford Pond and The Grange Lake were very similar and so we use the Old Alresford Pond output in our presented results.

Core recovery

Cores were collected from Old Alresford Pond and The Grange Lake (Figure 1). The lakes were cored with a UWITEC gravity corer for the uppermost sediments (0–20 cm), while the deeper sediments were collected using a 52 mm diameter Russian corer in 50 cm segments. Old Alresford Pond was originally cored in 2008 (ARL1) and then both lakes were cored in 2013 (ARL13 and GRA13), with an additional core collected from The Grange Lake in 2016 (GRA16: 51°7′12.32″N, 1°11′46.66″W) and Old Alresford Pond in 2017 (ARL17). These cores were collected as part of other studies focused on the lake's palaeoecology and the sediments were predominantly composed of marl (35–54 % of the inorganic fraction being calcium carbonate). In this study we use chronological markers and magnetic susceptibility tie points from the ARL13 and GRA13 cores (Supplemental Materials). A new sediment core was taken from Old Alresford Pond in 2019 (ALRE19: 51°5′37.43″N, 1°9′30.92″W) in three sections, using an adapted Livingstone corer to account for the shallow water depth (<1 m) and to collect the sediment-water interface, retrieving 260 cm of new sediment. The core was extruded into 1 cm slices and stored at <5°C. Here, we present data from core sediments between 0 and 93 cm, equivalent to ca. 1904–2019 CE.

Samples from ALRE19 and GRA16 were taken from both cores at 1 cm intervals, sieved at 250 µm with tap water and the coarse residue dried at a low temperature. Individual shells of *Candona neglecta* and *Candona candida* were picked from the dried >250 µm fraction and detrital material removed with a fine paint brush, under a low power biological microscope. Samples for analysis were composed of multiple (typically 5) shells to provide enough material for analysis and ensure that the results were representative of average lake conditions over the time-interval covered by a 1 cm increment of sediment. Analytical methods followed those for the modern ostracods.

Chronology

Chronologies for the cores were produced using a combination of short-lived radioisotopes (ALRE19) and spheroidal carbonaceous particle (SCP) counts (ARL13 and GRA13). Chronological determinations undertaken on the two cores used for isotope analysis in this study (ALRE19 and GRA16) were supplemented by tie points with parallel cores ARL13 and GRA13 using magnetic susceptibility data (Supplemental Figure 1). Tie points between ARL13 and ALRE19 occur at 44.5 cm (1958.1 ± 10 CE), 66.5 cm (1943.4 ± 13 CE) and 82.5 cm (1927 ± 15.5 CE) on the ARL13 age-depth scale reported using the intcal20_t age (years CE) and half the total years in the 95.4 % probability age range produced by the model. The age model for ALRE19 uses both independent ages from short-lived radioisotopes and SCP ages from ARL13. GRA16 relies on magnetic susceptibility tie points with GRA13 as there was not enough core material available to produce independent ages. There are 26 tie points between GRA13 and GRA16 detailed in the Supplemental Material.

Short-lived radioisotope analysis was applied to the ALRE19 core only as there was not enough sediment available from GRA16. Approximately 3 g of dried sediment were taken from ALRE19, crushed into a powder using a pestle and mortar and then analysed for ^{210}Pb, ^{226}Ra, ^{137}Cs and ^{241}Am on an ORTEC HPGe GWL series well-type coaxial low background intrinsic germanium detector by direct gamma assay at the Environmental Radiometric Facility, University College London. ^{210}Pb ages were then calculated using the constant rate supply (CRS) model (Appleby, 2001).

SCP ages were produced from the neighbouring ARLE13 and GRA13 cores. Samples were taken every 8 cm from Old Alresford Pond (ARLE13) and every 5 cm from The Grange Lake (GRA13) for SCP analyses (following Rose, 2008), to identify broad trends compared with the South and Central England calibration curve, following cumulative accumulation methods (Rose and Appleby, 2005). Subsequent samples were taken to clarify the depths of specific SCPs concentration peaks, which can be used to track industrial pollution over time, and hence these are used as a chronological marker for the last ca. 150+ years. SCP counts were expressed as cumulative percentage frequencies, which could then be related to ages determined using the South and Central England SCP age dataset, produced by Rose and Appleby (2005) that is based on ^{210}Pb chronologies from multiple lakes in the region. These data were used to produce age models for ARL13 and GRA13 using a Bayesian model in OxCal v4.4 (Bronk-Ramsey, 2009, 2021). All chronological data are in the Supplemental Material.

Magnetic susceptibility data for ALRE19 were produced by taking ca. 6 g of freeze-dried sediment and 1 cm intervals and placing in individual pots topped with cling film where sediment did not fill the pot. Low frequency magnetic susceptibility by sample mass (g) was recorded for each sample. ALRE13 was measured using mass specific magnetic susceptibility in HF mode. GRA16 data used volumetric susceptibility and GRA16 mass specific magnetic susceptibility in HF mode. These data were used for core correlation to produce single master cores for each site and date.

Age models for GRA13, GRA16, ARL13 and ALRE19 were generated using a Bayesian model in OxCal v4.4 (Bronk-Ramsey, 2009, 2021). Figures 2 and 3 in the Supplemental Figure document show the final age model for GRA16 and ALRE19 that are used in this study, and all data are available in the Supplemental Data File.

Results

Monthly recorded water temperature variations at both sites show expected seasonality and are similar at both sites. For example, maximum monthly temperatures were 18.68°C (2019) and 17.85°C (2020) at Old Alresford Pond, with similar values being recorded at The Grange Lake ‘Upper’ (18.79°C, 16.30°C in 2019–2020 CE respectively; Figure 2a–d). The monthly mean air temperatures recorded at Odiham, Otterbourne and Southampton NOC show the same seasonal trends as the recorded water temperatures (Figure 2e).

When comparing the $\delta^{18}O_{lake\_water}$ and δD_{lake_water} from both Old Alresford Pond and The Grange Lake across 2019–2020 CE, the values show a tight clustering of data points (Figure 3a). All the lake water isotope data lie close to, but slightly above, the LMWL for Wallingford (Figure 3a). For the 2019 data that overlap the monitoring period (Table 1), the $\delta^{18}O_{precipitation}$ values at Wallingford are within error of the $\delta^{18}O_{lake\_water}$ values at both Old Alresford Pond and The Grange Lake, despite the offset of the LMWL. The Piso_AI modelled data for both study sites and Wallingford demonstrate that the measured $\delta^{18}O_{lake\_water}$ and $\delta^{18}O_{precipitation}$ lie within those expected by the model for those locations (Figure 3c–e).

The modern $\delta^{18}O_{ostracod}$ and $\delta^{13}C_{ostracod}$ values from Old Alresford Pond and The Grange Lake are consistent with each other (Figure 3b). $\delta^{18}O_{ostracod}$ values at The Grange Lake span −1.17 ‰ (GRA-NOV19-LL-S) to −4.90 ‰ (GRA-FEB20-UL-M). These $\delta^{18}O_{ostracod}$ values vary from −3.89 ‰ (ALRE-NOV20-S) to −5.13 ‰ (ALRE-FEB20-S). There is no relationship between the month of collection and the $\delta^{18}O_{ostracod}$ value.

The ALRE19 core has a maximum $\delta^{18}O_{ostracod}$ value of −2.18 ‰ and a minimum of −4.71 ‰ compared to GRA16 that has a maximum $\delta^{18}O_{ostracod}$ value of −1.96 ‰ and a minimum of −4.29 ‰. The range of $\delta^{13}C_{ostracod}$ values in ALRE19 is from −9.94 ‰ to −5.90 ‰, and for GRA16 is −10.59 ‰ to −7.24 ‰. $\delta^{18}O_{ostracod}$ values are largely in the inverse direction of $\delta^{13}C_{ostracod}$ in both the ALRE19 and GRA16 records (Figure 4). A gradual negative shift in $\delta^{18}O_{ostracod}$ values from ca. 91 to 38 cm in ALRE19 followed by a positive shift from ca. 38 to ca. 33 cm, and then a decrease in values is recorded from ca. 38 cm to the top of the core (Figure 4a). Similar trends are observed in the GRA16 $\delta^{18}O_{ostracod}$ record with a negative trend observed from ca. 105 to 55 cm, followed by a positive shift from ca. 55 to 40 cm, which is followed by a decrease in values identified by just a few data points (Figure 4b). Overall, the trends across ALRE19 and GRA16 $\delta^{18}O_{ostracod}$ are similar.

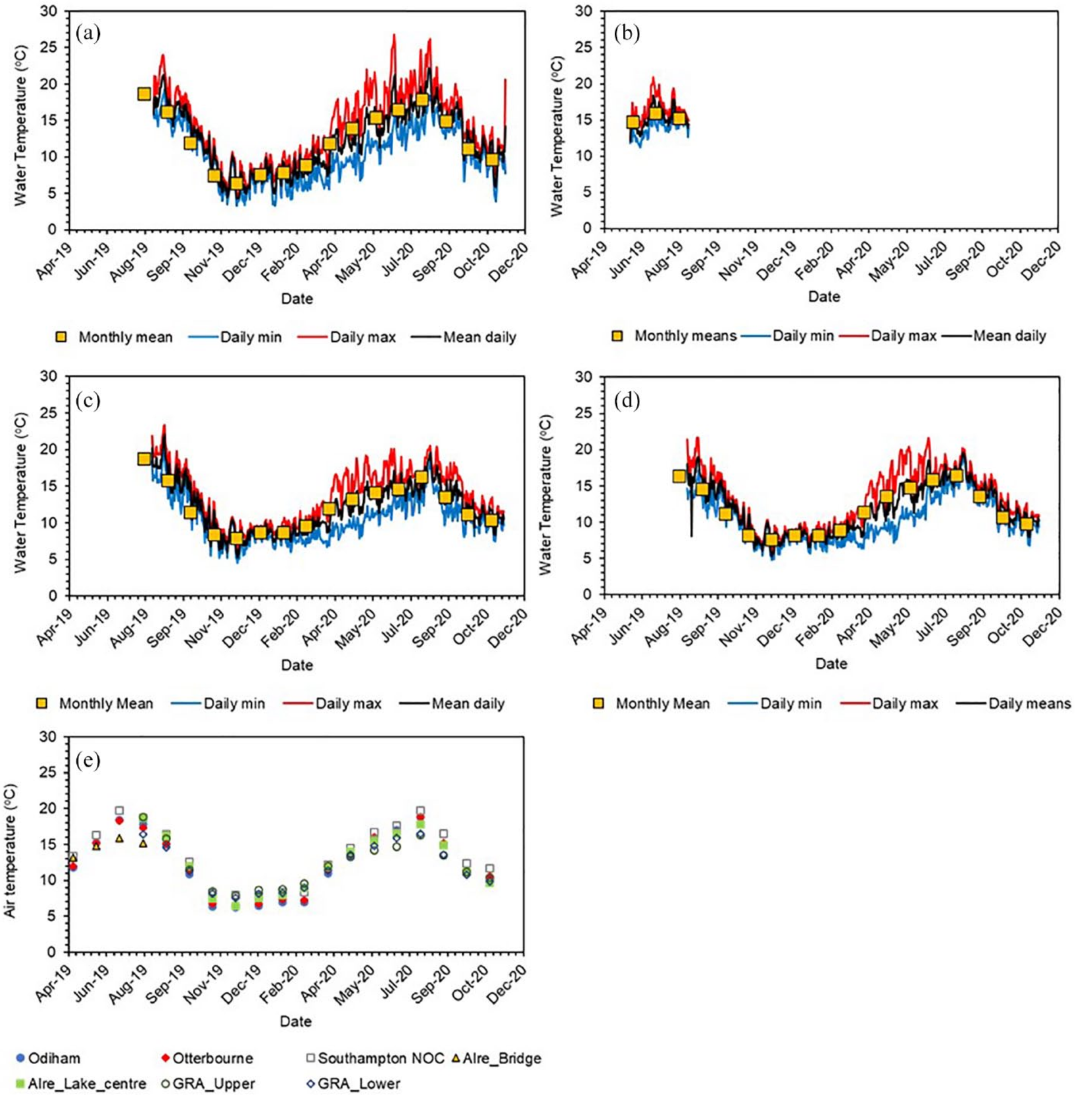

Figure 2. Water temperature at Old Alresford Pond lake centre (a) and under bridge (b) at outflow (Figure 1c). Water temperature at The Grange Lake Upper (c) and Lower Lake (d). a, c and d = August 2019 to November 2020; b = May 2019 to August 2019 (logger broke). Red = daily maximum temperature, blue = daily minimum temperature, black = daily mean temperature. Monthly mean temperatures (yellow) average the daily mean temperature over the period of one calendar month where values are available. (e) Monthly mean water temperature from Old Alresford Pond and The Grange Lake ('Upper' and 'Lower') compared to monthly mean air temperature from three MetOffice weather stations in the region (Otterbourne, Odiham, Southampton National Oceanography Centre (NOC)). Air temperature data provided by the National Meteorological Library and Archive – MetOffice UK. Site locations are shown in Figure 1.

Interpretation

Do modern lake waters at Old Alresford Pond and The Grange Lake record prevailing climatic conditions?

Oxygen-isotope values in ostracod-shell calcite are controlled by the temperature and isotopic composition of lake water at the time of ostracod shell secretion. To assess how modern $\delta^{18}O_{ostracod}$ values in the Alresford and Grange sequences record present-day climate conditions in the Hampshire lakes, we first assess how well lake temperature and $\delta^{18}O_{lake\_water}$ values record air temperature and $\delta^{18}O_{precipitation}$, respectively. For water temperature, Old Alresford Pond and The Grange Lake water temperature records show seasonal variability, indicating that the temperature of lake water responds quickly to air temperature (Figure 2a–d). The water temperatures show good agreement with air-temperature data from three local Met Office weather stations over the monitoring period (Figure 2e), indicating that we can use modern water temperature values as a good approximations of modern air temperature.

Secondly, we consider the relationship between $\delta^{18}O_{lake\_water}$ and $\delta^{18}O_{precipitation}$. The $\delta^{18}O_{lake\_water}$ values from the two lakes are similar and show minimal seasonal variation. The values lie parallel to the LMWL for Wallingford ca. 80 km away, but are slightly offset, suggesting that the data form a LMWL for each site (Figure 3a). To investigate this further, we generated a $\delta^{18}O_{precipitation}$ time series using the Piso_AI model (Nelson et al., 2021) (Figure 3c–e). Measured $\delta^{18}O_{precipitation}$ values at Wallingford lie within error of the modelled $\delta^{18}O_{precipitation}$ values (Figure 3e), indicating the Piso_AI model does produce a reasonable approximation of true $\delta^{18}O_{precipitation}$. We therefore use the model to estimate $\delta^{18}O_{precipitation}$ values at Old Alresford Pond and The Grange Lake and compare these to the measured $\delta^{18}O_{lake\_water}$ values. Then we test if these measured $\delta^{18}O_{lake\_water}$ values can be used as a proxy of $\delta^{18}O_{precipitation}$ values (3c–d). At both study sites, measured $\delta^{18}O_{lake\_water}$ values lie within error of the modelled $\delta^{18}O_{precipitation}$ values at each location, and we therefore conclude that $\delta^{18}O_{lake\_water}$ can be used as a proxy for $\delta^{18}O_{precipitation}$ (3c–d). The fact that the $\delta^{18}O_{lake\_water}$ values from both systems closely approximates the mean modelled $\delta^{18}O_{precipitation}$ values supports

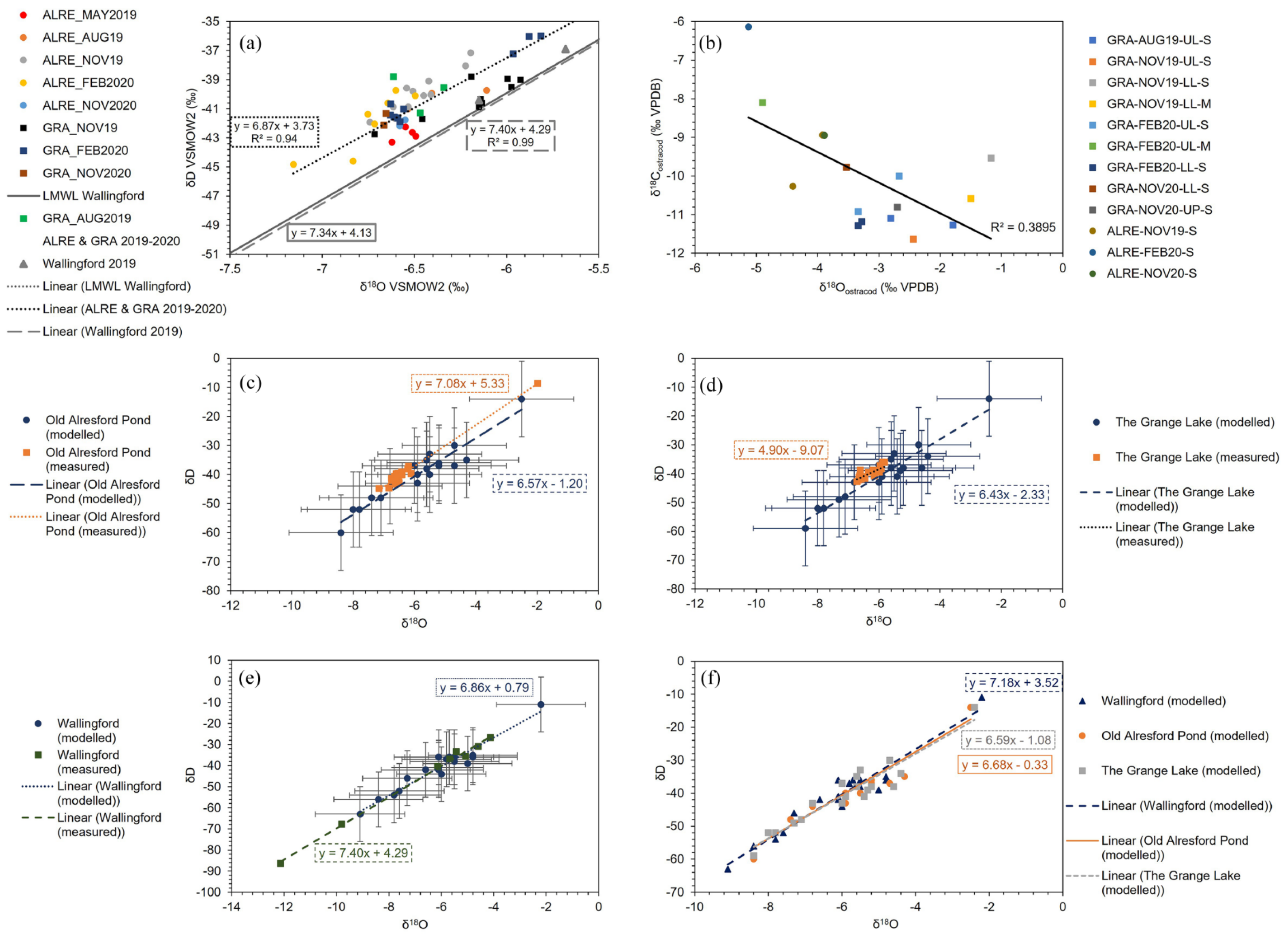

Figure 3. (a) $\delta^{18}O_{lake\_water}$ samples taken across Old Alresford Pond (circles) and The Grange Lake (squares – includes upper and lower lake measurements) across the monitoring period May 2019 to November 2020, with months depicted by colour and outliers removed. (b) $\delta^{18}O_{ostracod}$ and $\delta^{13}C_{ostracod}$ data from ostracod valves collected at Old Alresford Pond (circles) and The Grange Lake (squares) across the monitoring period. UL = Upper Lake, LL = Lower Lake, S = Sediment sample, M = Macrophyte sample. The linear trend line and R^2 use all data points. Outliers removed. (c) Compares Piso_AI modelled $\delta^{18}O_{precipitation}$ with measured $\delta^{18}O_{lake\_water}$ values from the Old Alresford Pond across the monitoring period May 2019 to November 2020. (d) compares Piso_AI modelled $\delta^{18}O_{precipitation}$ with measured $\delta^{18}O_{lake\_water}$ values from The Grange Lake across the monitoring period May 2019 to November 2020. The error on modelled values is ±1.7 ‰ $\delta^{18}O$ and ±13 ‰ $\delta D_{precipitation}$ as shown by error bars (Nelson et al., 2021). (e) Compares Piso_AI modelled $\delta^{18}O_{precipitation}$ from the Wallingford GNIP site across the monitoring period May 2019 to November 2020 with measured values from May to December 2019. (f) Compares Piso_AI modelled $\delta^{18}O_{precipitation}$ and $\delta D_{precipitation}$ values from Wallingford, Old Alresford Pond and The Grange Lake across the monitoring period May 2019 to November 2020.
Source: Piso_AI model from Nelson et al. (2021); Wallingford LMWL from IAEA/WMO (2023).

Table 1. Summary of $\delta^{18}O_{lake\_water}$ and δD_{lake_water} data from Old Alresford Pond and The Grange Lake May to Dec 2019 and of $\delta^{18}O_{precipitation}$ and $\delta D_{precipitation}$ values from Wallingford for the same period (IAEA/WMO, 2023).

Location	$\delta^{18}O_{lake\_water}$		δD_{lake_water}		$\delta^{18}O_{precipitation}$		$\delta D_{precipitation}$		Piso_AI $\delta^{18}O_{precipitation}$		Piso_AI $\delta D_{precipitation}$		$\delta^{18}O_{groundwater}$
	Mean	1 SD	Mean	1 SD	Mean	1 SD	Mean	1 SD	Mean	1 SD	Mean	1 SD	Mean
Old Alresford Pond	−6.39	0.9	−39.98	6.4					−5.9	1.4	−39.95	9.9	
The Grange Lake	−6.33	0.3	−40.13	1.9					−5.89	1.4	−40.21	9.8	
Wallingford GNIP					−6.63	2.8	−44.75	21	−6.1	1.5	−41.05	10.9	
Wessex Basin Chalk groundwaters													−6.6

SD: standard deviation.
Source: Piso_AI model data from Nelson et al. (2021) and Wessex Chalk groundwater data from Allen (2017).

the suggestion that lake water values reflect the weighted mean annual $\delta^{18}O_{precipitation}$ value.

What controls the $\delta^{18}O_{ostracod}$ of modern ostracod valves?

Having demonstrated that water temperature is a good approximation of air temperature (Figure 2) and that $\delta^{18}O_{lake\_water}$ is a good reflection of $\delta^{18}O_{precipitation}$ (Figure 3), we use the Kim and O'Neil (1997) equation to predict the oxygen-isotope value of carbonate precipitated in these lakes, referred to here as $\delta^{18}O_{synthetic}$, adding 2.2 ‰ for the offset from oxygen-isotope equilibrium for Candonidae (equation (1)). This equation is often used in ostracod studies and has good reproducibility in experimental conditions (Keatings et al., 2002). We use the mean annual $\delta^{18}O_{lake\_water}$ values for each site as a representation of mean annual $\delta^{18}O_{precipitation}$ and mean measured winter temperature (DJF) values since both *C. candida* and *C. neglecta* calcify in the cooler, winter months (Decrouy, 2009; Dole-Olivier et al., 2000). We then compare the $\delta^{18}O_{synthetic}$ value to measured $\delta^{18}O_{ostracod}$ values over the monitoring period (Figure 3b and Table 2), with the caveat that the latter may represent the last few years because the ostracods were not living at the time of collection.

$$\delta^{18}O_{carbonate(Kim\,\&\,O'Neil(1997))} = \delta^{18}O_{lake\_water} + (28.625 - (SQRT((16.5604 + (0.32 * DJF\ mean\ water\ temperature)))/0.16))$$

$$\delta^{18}O_{synthetic} = \delta^{18}O_{carbonate(Kim\,\&\,O'Neil(1997))} + 2.2 \tag{1}$$

At Old Alresford Pond mean $\delta^{18}O_{ostracod}$ was -4.07 ± 0.29 ‰, which falls outside the range of the $\delta^{18}O_{synthetic}$ value using DJF temperatures at −2.73 ‰. The difference between $\delta^{18}O_{synthetic}$ and $\delta^{18}O_{ostracod}$ is closer when using SON temperatures (−3.77 ‰). Whilst we know that the two *Candona* species that we analysed typically calcify in the winter, the individuals collected could represent several years, as they were not necessarily living at the time of collection, whereas our measured lake temperatures are from just 1 year. It is also possible that some individuals may have calcified slightly earlier in the year than others. Moreover, there is additional uncertainty around how representative the measured modern $\delta^{18}O_{ostracod}$ values at Old Alresford Pond are, because only three samples were analysed from this site. Furthermore, the measured lake temperatures across SON and DJF have quite large standard deviations (±3.94°C and ±1.41°C, respectively), so that average temperatures from 2019 to 2020 CE are not a perfect representation of the conditions in which specimens analysed had calcified. These reasons could explain the larger offset between $\delta^{18}O_{synthetic}$ and $\delta^{18}O_{ostracod}$ in DJF at Old Alresford Pond as this difference is reduced when considering SON temperatures, also representing the cooler part of the year. The 'Upper Lake' at The Grange Lake has a $\delta^{18}O_{synthetic}$ value that lies within the range of $\delta^{18}O_{ostracod}$ value, indicating that $\delta^{18}O_{ostracod}$ are predominately a reflection of temperature and $\delta^{18}O_{precipitation}$. The proximity of the lakes to one another means we expect similar $\delta^{18}O_{ostracod}$ trends if these values are being driven by climate variables. The range of $\delta^{18}O_{ostracod}$ values from The Grange Lake lie largely within the range of those from Old Alresford Pond (Figure 3b). These similarities between the two lakes further support our conclusion that in the modern day $\delta^{18}O_{ostracod}$ values at these lakes are primarily a reflection of the temperature of the coldest part of the year, albeit with greater uncertainty from the Old Alresford Pond data presented, and mean $\delta^{18}O_{precipitation}$.

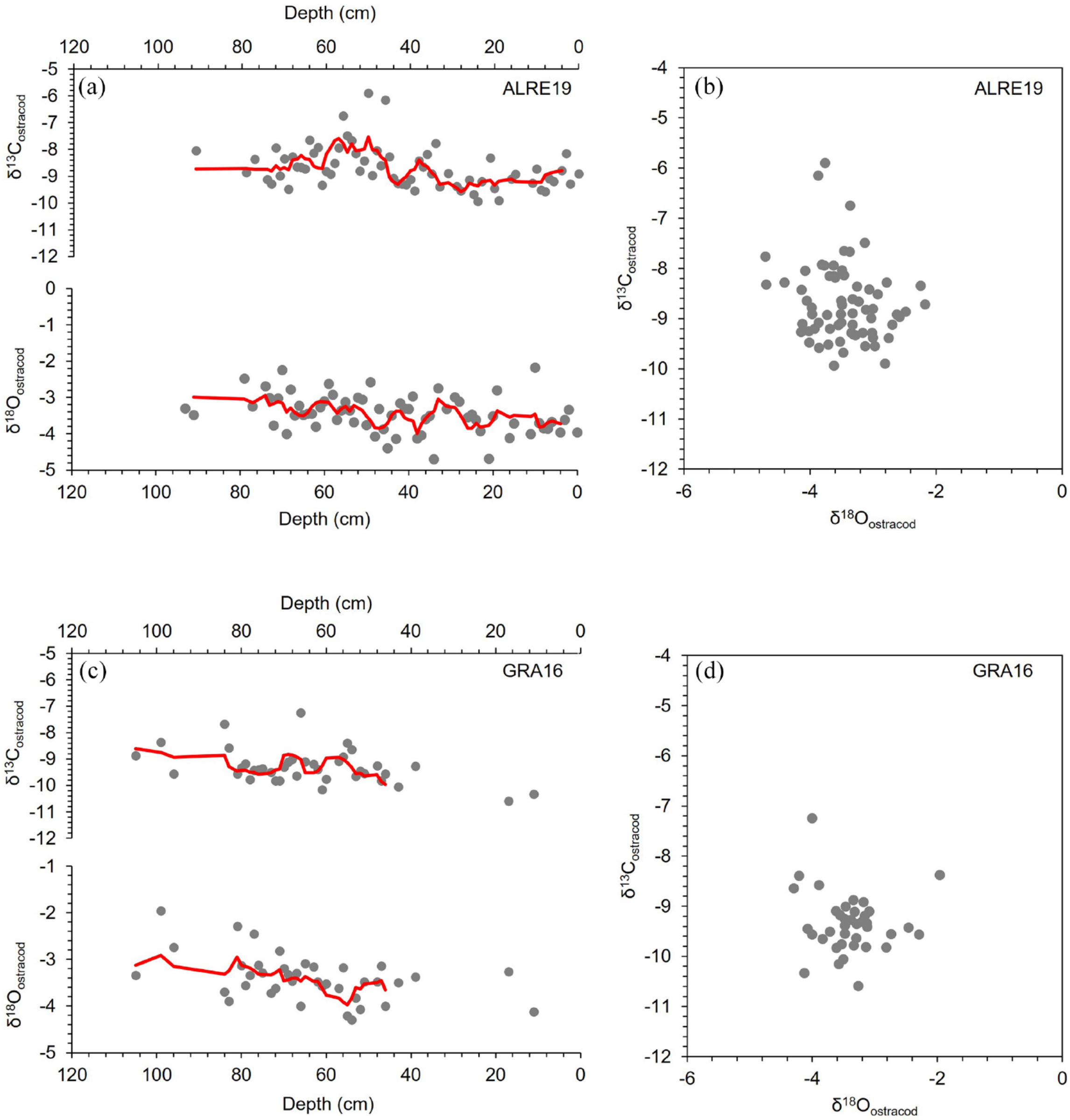

Figure 4. Downcore oxygen and carbon isotope data from (a) ALRE19 with (c) cross-plot of ALRE19 isotope data and (b) GRA16 core with (d) cross-plot of GRA16 isotope data. Data correspond to ca. 1904–2019 CE. Grey circles = individual data points with outliers removed. Red line = five-point moving average.

Does sub-fossil $\delta^{18}O_{ostracod}$ from Old Alresford Pond and The Grange Lake reflect 20th century climate variability?

If the modern temperature and isotope chemistry of lake waters in the Old Alresford Pond and The Grange Lake systems reflect prevailing climatic conditions, and both factors control $\delta^{18}O_{ostracod}$, then downcore changes in $\delta^{18}O_{ostracod}$ should record past climate variability. To test this, we compare the measured $\delta^{18}O_{ostracod}$ values derived from downcore analysis of these core sequences with a 'synthetic' $\delta^{18}O_{ostracod}$ record calculated using the fractionation equation of Kim and O'Neil (1997) to combine the annual DJF temperature record from the Southampton MetOffice station and annual $\delta^{18}O_{precipitation}$ values simulated using Piso_AI. A standard value of 2.2 ‰ is added to these values to allow for the vital offsets from oxygen-isotope equilibrium in Candonidae. We refer to this downcore synthetic record here as $\delta^{18}O_{synthetic}$. We present the $\delta^{18}O_{synthetic}$ record for Old Alresford Pond (Figure 6c) as this site has the most $\delta^{18}O_{ostracod}$ datapoints to compare to, and we have already demonstrated the similarity between the $\delta^{18}O_{ostracod}$ at the two lakes (Figure 5). We assume that $\delta^{18}O_{precipitation}$ was a close approximation of $\delta^{18}O_{lake\_water}$ for the 20th century, as it is at present, and that neither of the lakes has been significantly impacted by evaporative enrichment, for example as a result of local hydrological changes, in the past. Several lines of evidence suggest that this assumption is reasonable. First, the $\delta^{18}O_{ostracod}$ records for the two lakes broadly agree for the period of temporal overlap. Had local hydrological effects, such as anthropogenic catchment modifications, led to significant evaporative enrichment, this would be most likely to have affected one or the other of the two catchments and not both. Climatic influences, in particular a reduction of effective moisture, could have led to significant evaporative enrichment, but there is no evidence for this in the meteorological data. Secondly, there is no covariance between $\delta^{18}O_{ostracod}$ and $\delta^{13}C_{ostracod}$ in the down-core data (Figure 4c and d). Although not an unequivocal indicator of increased residence time and evaporative enrichment in mid-latitude lakes (Drummond et al., 1995; Talbot, 1990), the lack of covariance does lend some support to our assumption that such enrichment has not been a significant component of the lakes' hydrological balance during the 20th century.

Table 2. Comparing $\delta^{18}O_{synthetic}$ to $\delta^{18}O_{ostracod}$ at Old Alresford Pond and The Grange Lake in the 2019–2020 monitoring period.

Site	Season	Mean annual $\delta^{18}O_{lake\_water}$	$\delta^{18}O_{lake\_water}$ SD	Number of mean annual $\delta^{18}O_{lake\_water}$ samples	Mean water temperature (°C)	Water temperature SD	Number of water temper-ature samples	Mean $\delta^{18}O_{synthetic}$	Mean $\delta^{18}O_{ostracod}$	Number of $\delta^{18}O_{ostracod}$ samples	$\delta^{18}O_{ostracod}$ SD	$\delta^{18}O_{ostracod}$ range	Difference between $\delta^{18}O_{synthetic}$ and $\delta^{18}O_{ostracod}$
Old Alresford Pond	DJF	−6.39	0.89	28	7.29	1.41	90	−2.73	−4.07	3	0.29	−3.78 to −4.36	1.34
	SON				11.89	3.94		−3.77					0.3
The Grange Lake (Upper)	DJF	−6.28	0.31	19	8.44	0.95	90	−2.88	−3.14	6	0.83	−2.31 to −3.97	0.26

All measured values from The Grange Lake come from the 'Upper', lake to align with core location. Outliers have been removed. All data available in Supplemental Material.
DJF: December, January, February; SON: September, October, November.

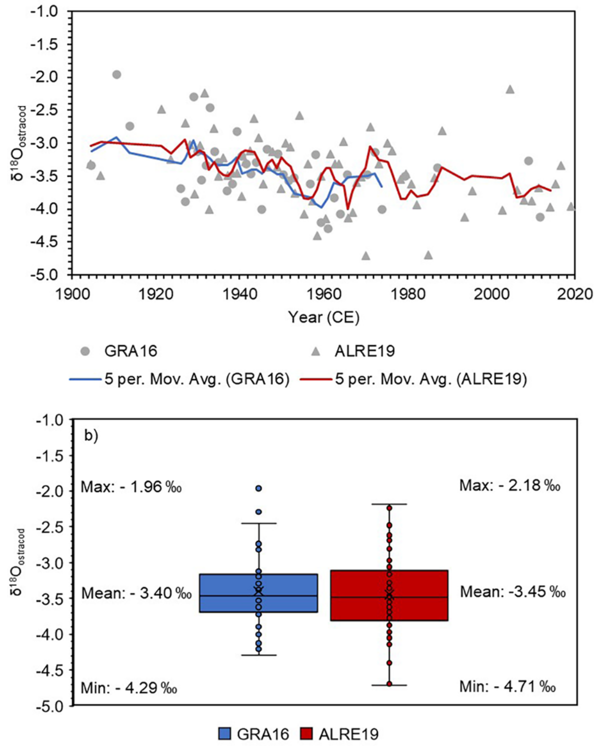

Figure 5. (a) $\delta^{18}O_{ostracod}$ data from ALRE19 (grey triangles) and GRA16 (grey circles) with a five-point moving average (ALRE19 = red and GRA16 = blue); (b) boxplot summarising the $\delta^{18}O_{ostracod}$ data from ALRE19 (red) and GRA16 (blue) from 1904 CE. Outliers have been removed (all data are presented in Supplemental Material).

Comparison of the structure of the $\delta^{18}O_{ostracod}$ and $\delta^{18}O_{synthetic}$ records for the 20th century shows reasonable agreement given the chronological uncertainties of $\delta^{18}O_{ostracod}$ (Figure 6c–d). The key characteristic of the $\delta^{18}O_{ostracod}$ record is the positive shift from the 1960s, peaking ca. 1971 CE (Figure 6d). This is followed by a fluctuation of values around a stable mean, which is lower than the 1960s–70s peak. We also observe a peak in the $\delta^{18}O_{synthetic}$ record ca. 1960s, but the stabilising of values post-$\delta^{18}O_{synthetic}$ are not necessarily more negative. The long-term negative shift in $\delta^{18}O_{ostracod}$ values from 1904 to 1960 CE is interrupted by a positive shift in the 1940s CE, which is also evident in the $\delta^{18}O_{synthetic}$ record. Whilst the absolute values of the two records are offset on average by ca. 0.6 ‰, the range between maximum and minimum of the moving average values is ca. 1 ‰. The overall long term negative trend in the $\delta^{18}O_{ostracod}$ is not as clearly captured in the $\delta^{18}O_{synthetic}$ record (Figure 6c and d). This is unlikely to be a result of basin-specific effect because this difference is observed in both the Old Alresford Pond and The Grange Lake records, as well as both $\delta^{18}O_{ostracod}$ and $\delta^{13}C_{ostracod}$ datasets (Figure 4).

The consistency between these two records is important for the 'ground-truthing' of the data. For example, modern monitoring indicates that $\delta^{18}O_{ostracod}$ at Old Alresford Pond is controlled by $\delta^{18}O_{lake\_water}$ (which in this instance is effectively $\delta^{18}O_{precipitation}$) and autumn/winter temperatures. The reasonable agreement between the $\delta^{18}O_{ostracod}$ and $\delta^{18}O_{synthetic}$, which is generated using winter temperatures and $\delta^{18}O_{precipitation}$, indicates that this relationship holds true for the 20th century at least. Equally, the good agreement between the modelled $\delta^{18}O_{precipitation}$ data, which is incorporated in the $\delta^{18}O_{synthetic}$ values, and the empirical data provides added confidence that the modelled outputs are robust. For instance, the positive shift in $\delta^{18}O_{precipitation}$ in the 1960s–1970s corresponds with positive shifts observed in both $\delta^{18}O_{ostracod}$ and $\delta^{18}O_{synthetic}$. The uncertainty surrounding the ALRE19 age model

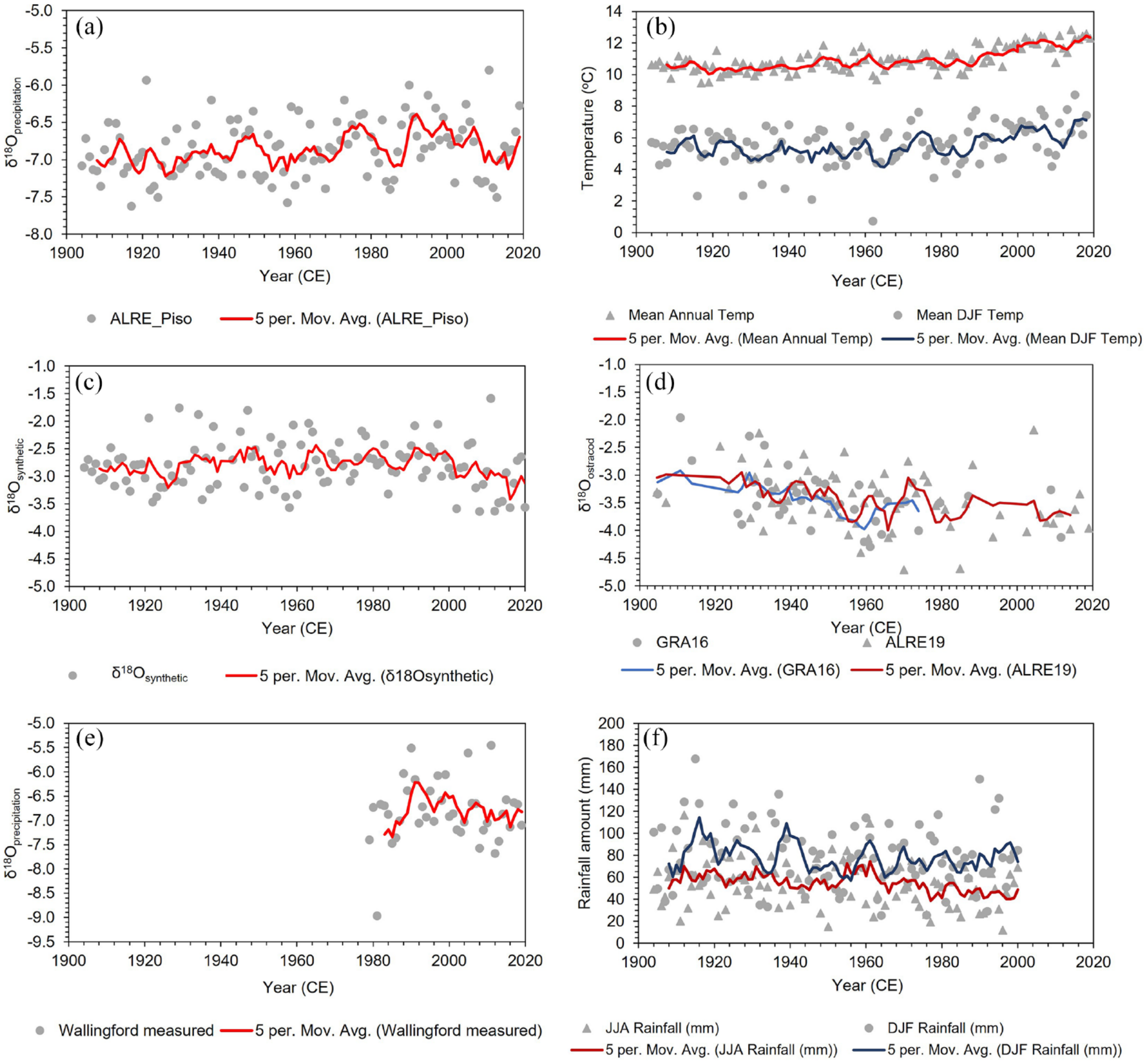

Figure 6. (a) Modelled $\delta^{18}O_{precipitation}$ at Old Alresford Pond from Piso_AI model (Nelson et al., 2021) 1904 to 2020 CE where grey circles = all data points and red line = five-point moving average. (b) Mean annual temperature in Southampton (grey triangles) with five-point moving average (red line) and mean winter (December, January, February) temperature in Southampton (grey circles) with five-point moving average 1904–2019 CE from MetOffice. (c) $\delta^{18}O_{synthetic}$ at Old Alresford Pond (grey circles) with red line = five-point moving average. (d) Measured $\delta^{18}O_{ostracod}$ values from ALRE19 (grey triangles) with five-point moving average (red line) and GRA16 (grey circles) with five-point moving average (blue line). (e) Measured $\delta^{18}O_{precipitation}$ at Wallingford GNIP site 1979–2019 CE and five-point moving average (IAEA/WMO, 2023). (f) Summer (June, July, August) rainfall amount (grey triangles) with five-point moving average (red line) and winter (December, January, February) rainfall amount (grey circles) with five-point moving average (blue line) from Southampton MetOffice, 1904–2000 CE. All data in Supplemental Material. MetOffice data made available from the National Meteorological Library and Archive.

(with age uncertainties as much as ±23.5 years at some levels) make the statistical testing of this similarity challenging. However, to gain some insight into the similarity, we split the record based on the sampling density in the ALRE19 record; from ca. 1904 to 1940 CE the Pearson correlation coefficient (*r*) value between $\delta^{18}O_{ostracod}$ and $\delta^{18}O_{synthetic}$ is 0.04, ca. 1941 to 1970 CE equal to 0.23 and from ca. 1971 to 2019 CE at 0.27. This demonstrates some statistical similarity throughout the record and that there is some similarity in the direction of correlation between the $\delta^{18}O_{ostracod}$ and $\delta^{18}O_{synthetic}$ prior to 1980 CE. The difference post-1980 CE may be a factor of the distribution of $\delta^{18}O_{ostracod}$ data points, particularly how few $\delta^{18}O_{ostracod}$ data points are in the ca. 1990s, as the moving averages in Figure 6c and 6d indicate that a slight negative trend is present in both $\delta^{18}O_{ostracod}$ and $\delta^{18}O_{synthetic}$.

Whilst a reasonable agreement exists between the $\delta^{18}O_{ostracod}$ and $\delta^{18}O_{synthetic}$ records there is less of a clear relationship between $\delta^{18}O_{ostracod}$ and any single climatic parameter. For example, both mean annual and winter temperatures have increased across the 20th century in the Southampton dataset (Figure 6b), but no comparable, simple trend exists within the isotopic dataset. It is likely that this discrepancy reflects the fact that both the $\delta^{18}O_{ostracod}$ and $\delta^{18}O_{synthetic}$ records reflect an amalgamation of multiple climate factors not just autumn/winter temperatures but also the different climate factors that control $\delta^{18}O_{precipitation}$ (i.e. air temperature, the amount/seasonality rainfall, air mass trajectory, etc.; see Rozanski et al., 1992, 1993). Only the trend in summer rainfall amount shows any comparability with the trend seen in $\delta^{18}O_{ostracod}$ with a value of 0.15 between ca. 1904 and 2000 CE (Figure 6f). Summer rainfall variability has the potential to produce the negative/positive trends seen in the $\delta^{18}O_{ostracod}$ record, that is, as summer rainfall typically has the most positive $\delta^{18}O_{precipitation}$ values of any rainfall occurring throughout the year, its increase/decrease will potentially result in more positive/negative $\delta^{18}O_{ostracod}$ values respectively by its contribution to average $\delta^{18}O_{lake\_water}$ values.

The magnitude of changes seen in the $\delta^{18}O_{ostracod}$ record are, however, unlikely to be explained by the changes in summer rainfall based on the Southampton weather station data.

Discussion

Can we identify individual climate drivers from the $\delta^{18}O_{ostracod}$ record?

In many Pleistocene studies in Northwestern Europe the variability in the $\delta^{18}O$ of lake carbonates is frequently interpreted as reflecting temperature changes (Blockley et al., 2018; Candy et al., 2016). In the present study, it is unlikely that temperature is the only driving factor, however, because the magnitude of the temperature changes that are observed in some of these Late Pleistocene transitions, such as the Younger Dryas, is so large that it is often interpreted as the primary control and is assumed to overwhelm the influence of other environmental factors. The present study has demonstrated, however, that during the Late-Holocene, and in particular the 20th century when the magnitude of temperature change is much smaller (Figure 6b), several climatic variables interact to control the $\delta^{18}O$ of lake carbonates, in this case ostracods. Furthermore, these interactions may not be straightforward, as measured temperature and rainfall amount do not necessarily follow the same trends (Figure 6b and f). In this example there is a weak correlation between winter temperature (the season when the candonids used for this study calcify) and summer rainfall (thought to have a greater influence on $\delta^{18}O_{precipitation}$ and therefore $\delta^{18}O_{lake\_water}$). The Pearson correlation coefficient (r) from 1904 to 2000 CE for these variables is −0.02. Whilst it is possible to use anthropogenic lakes, such as Old Alresford Pond, as an archive for isotopic changes across the last 1000 years, interpreting this proxy record based on a single climate variable is problematic. This study has shown that such lakes do have a key role, however, in validating the output of isotope enabled climate models. These archives allow high-resolution isotopic records to be generated that can be directly compared with modelled $\delta^{18}O_{precipitation}$ values. The coupling of model and empirical isotope data, through studies such as these, therefore, has the potential to validate and improve modelling studies in the future.

Can anthropogenic lakes be used as climate archives?

The British Isles hosts many anthropogenic lakes including ornamental lakes and medieval fishing or stew ponds. These are typically overlooked as archives of past climate change, largely because of the influence of human activity on the record and the challenges faced in creating robust chronologies over recent centuries. However, such sites are likely to have written histories that may help overcome some of these challenges, as well as extant water bodies that allow us to disentangle climate and catchment variability captured in our chosen proxy, as in the present study. This study has shown that there is great potential in using anthropogenic lakes, in this instance those that precipitate carbonate, as archives of past climatic and environmental change over the historical period. Consequently, using anthropogenic lakes in this way has the potential to make a significant contribution to our knowledge and understanding of the regional impact of known climate changes in this period (e.g. MCA and LIA). Considering the known variability in the expression of these events across NW Europe, an ability to produce proxy records that may be able to quantify climate change variability at a more localised level, will be of significant interest to the scientific community. From the perspective of the British Isles, decades of Holocene paleoclimatic research have demonstrated that there is often regional variability and asynchronicity in response to climate change (e.g. Charman, 2010). As we continue to understand what modern climate change may look like in the British Isles it becomes increasingly important to develop our understanding of the magnitude and variability of past climate change, especially in the Late-Holocene where external climate driving mechanisms have changed little. Therefore, improving the spatial and temporal resolution of such records will provide a useful reference point. Future work should investigate additional anthropogenic lakes across the British Isles to assess their potential for paleoclimatic reconstruction. Oxygen isotopes in particular have the advantage of yielding quantified climate reconstructions and are therefore a valuable proxy to employ in such investigations. A network of sites across the region would provide the opportunity to compare sites over a wider spatial scale and identify possible trends, synchronicities and complexities in the expression of climate variability over this time. This may provide important insights into the expression and impact of known events, such as the LIA, on the region. Furthermore, based on the success of this study in comparing to modelled values, it may be possible to further explore the use of a proxy-model comparison, employing isotope-enabled climate models to help identify possible driving mechanisms behind observed climate trends recorded in proxy oxygen-isotope datasets.

Conclusions

This work has brought together modern limnological monitoring alongside palaeoclimatological methods to investigate the potential of constructed lakes as archives of 20th century climate. We demonstrated for our Hampshire sites that $\delta^{18}O_{lake\_water}$ is representative of local $\delta^{18}O_{precipitation}$, and this is recorded in $\delta^{18}O_{ostracod}$. Furthermore, we have shown that water temperature is in equilibrium with air temperature in the present day, and therefore assume that this is true in the past. We observe similarities in the direction of $\delta^{18}O_{ostracod}$ trends in both Old Alresford Pond and The Grange Lake over the 20th century, indicating they are most likely driven by a common climate driver rather than basin-specific effects. We explore this further by comparing $\delta^{18}O_{ostracod}$ trends to recorded changes in temperature and rainfall amount, two climatic factors that may drive changes in $\delta^{18}O_{ostracod}$. These comparisons demonstrated that there is no simple relationship between measurable climate variables (temperature and rainfall amount) and $\delta^{18}O_{ostracod}$ and so we argue that a complex interaction of climatic variables drive the observed $\delta^{18}O_{ostracod}$ shifts in the 20th century. We demonstrate a good relationship between measured and modelled isotopic values in this study and discuss the potential role anthropogenic lakes containing paleoclimatic proxy records may have in 'ground-truthing' modelled oxygen-isotope outputs. Future work that employs a more extensive data-model comparison approach using isotope-enabled climate models in lacustrine environments may allow for a greater variety of climate driving mechanisms on $\delta^{18}O_{ostracod}$ to be assessed, given that there is no simple driver in our 20th century record. Overall, we have shown the sediments of anthropogenic lakes are valuable archives of climatic variability over the last millennium.

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Supplemental material

Supplemental material for this article is available online.

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