Quantitative reconstruction of early Holocene and last glacial climate on the Balkan Peninsula using coupled hydrological and isotope mass balance modelling

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Abstract

We investigate the modern hydrology of Lake Ohrid (Macedonia/Albania) using a combined hydrological and isotope-based modelling approach and present a new evaluation of contemporary water balance and palaeoclimate estimates. The combined model is able to estimate hydrological components that cannot be directly measured, and indicates that sublacustrine spring inflow is in the order of 50% higher than previous estimates and groundwater outflow comprises approximately a third of overall water outflow. In combination with sediment core oxygen isotope data, we used the combined model to quantitatively reconstruct past climate, in particular precipitation, during the early Holocene and last glacial period. Calculated precipitation in the early Holocene was higher than the value for present day and was approximately 44% lower than present during the last glacial, assuming the majority of precipitation fell as snow. The estimated amount of precipitation in the last glacial would have been high enough to provide refugial conditions at Lake Ohrid and to support the continuous existence of arboreal vegetation in the catchment. The improved understanding of the modern isotope hydrology of Lake Ohrid is fundamental for explaining the systematics of

past isotope variation and providing context for extended sediment records from the lake, which will provide longer-term palaeoclimate reconstructions covering multiple glacial-interglacial cycles.

1. Introduction

Lake Ohrid, located on the Balkan Peninsula (Figure 1), is thought to be Europe's oldest freshwater lake with continuous lacustrine sedimentation for at least the past 1.2 Ma (Wagner et al., 2017). The sensitivity of Lake Ohrid and its catchment to hydroclimate variability over the last glacial-interglacial cycle has been documented in several studies using the stable isotope composition of carbonates (Leng et al., 2010; Lacey et al., 2015), geochemical and sediment proxies (Vogel et al., 2010; Wagner et al., 2010), and terrestrial vegetation composition from pollen (Wagner et al., 2009; Panagiotopoulos et al., 2013, 2014). In 2013, an International Continental scientific Drilling Program (ICDP) deep drilling campaign, the Scientific Collaboration on Past Speciation Conditions in Lake Ohrid (SCOPSCO) project, recovered a 584-m composite sediment core sequence spanning the entire history of the lake (Wagner et al., 2014, 2017). Analytical work to date encompasses the upper section of this record (ca. <640 ka) and reveals shifts to more positive values in the reconstructed oxygen isotope composition of lake water ($\delta^{18}O_L$) between glacial and interglacial phases, in particular during the transition between the last glacial and early Holocene (Lacey et al., 2016). This transition from lower glacial to higher interglacial $\delta^{18}O_L$ is opposite to the trend observed in other lake and speleothem records from the region (e.g. Roberts et al., 2008; Masi et al., 2018). In addition, the pollen record from SCOPSCO cores indicates a good correspondence between changes in the vegetation assemblage and glacial-interglacial cycles, and suggests that moisture availability was an important forcing mechanism in controlling the presence and abundance of arboreal vegetation in the catchment (Sadori et al., 2016).

Therefore, to explain the observed changes in $\delta^{18}O_L$ at Lake Ohrid between glacial and interglacial periods, and to estimate past changes in moisture availability, it is necessary to evaluate the drivers of $\delta^{18}O$. In the Mediterranean region, water balance is typically considered to be the primary driver of lake stable isotope hydrology (Roberts et al., 2008, 2010), however it is crucial to understand the modern hydrology of individual lake systems to act as a basis for calibrating proxy-based reconstructions of past climate and environmental change. Palaeoclimate investigations based on stable isotope data from Lake Ohrid have so far only utilised simple linear regression models to understand the hydrological balance of the lake. The current understanding of water balance is derived from estimates that have been modelled using a hydrological mass balance approach (Watzin et al., 2002; Matzinger et al., 2006b), which assume negligible groundwater outflow from the lake.

In this study, we use existing monitoring datasets to constrain groundwater flows and calculate a new contemporary water balance for Lake Ohrid using coupled hydrological and stable isotope mass balance modelling. We then use this model to provide a quantitative reconstruction of early Holocene and last glacial climate, in particular past changes in precipitation, in order to better explain glacial-interglacial shifts in $\delta^{18}O_L$ observed in the proxy record (Lacey et al., 2016). The reconstructed changes in climate are also used to test the hypothesis of Sadori et al. (2016) that the Lake Ohrid catchment had sufficient moisture levels during glacial phases to act as a refugium for arboreal vegetation. The new water balance model is critical to providing an improved, quantitative understanding of the modern isotope hydrology of Lake Ohrid, which will be helpful for discerning the systematics of hydroclimate in longer-term reconstructions from Lake Ohrid that cover multiple glacial-interglacial cycles.

2. Study site

Lake Ohrid (693 m a.s.l.; 40°54’-41°10’N, 20°38’-20°48’E; Figure 1) is situated on the border between the Former Yugoslav Republic of Macedonia and Albania in a Pliocene-formed tectonic graben, bounded to the east by the Galičica (2262 m a.s.l.) and Mali Thate (2287 m a.s.l.) mountains and to the west by the Mokra Mountain chain (1500 m a.s.l.). The lake is approximately 30 km long by 15 km wide and covers an area of 358 km^2. The lake basin has a tub-shaped morphology with a water volume of 50.7 km^3, and a maximum and average water depth of 293 m and 150 m, respectively. Lake Ohrid is fed by a direct catchment area of around 1000 km^2, however an underground connection via karst channels to neighbouring Lake Prespa (849 m a.s.l.; 10 km east of Ohrid) expands the watershed to 2600 km^2 (Amataj et al., 2007). Remaining water input is derived from direct precipitation and river inflow, and output is dominated by evaporation and river outflow. Groundwater outflow is currently assumed to be negligible and has not been observed to date (Wagner et al., 2008).

The climate of Lake Ohrid and its watershed is strongly dependent on both Mediterranean and continental influences, owing to the lake’s location in a deep valley surrounded by high mountains and its proximity to the Adriatic Sea, and is also modified by the thermal capacity of the lake itself (Watzin et al., 2002). Average monthly air temperatures range between 2°C in winter and 21°C in summer, with absolute minimum and maximum values of approximately −15°C and 40°C in winter and summer, respectively (Figure 2). The annual distribution of precipitation belongs to the Mediterranean pluviometric regime and varies considerably depending on geographical position in the catchment (Watzin et al., 2002). Rainfall stations around the shoreline of Lake Ohrid receive an average precipitation of 773 mm/year (Figure 2), however this increases up to 1445 mm/year at higher altitudes in the catchment (Wagner et al., 2008). Prevailing wind directions are governed by the basin morphology, with northerly winds prevailing in winter and southerly winds in summer (Stankovic, 1960; Watzin et al., 2002).

3. Methodology

3.1 Hydrological mass balance

The annual water mass balance of a well-mixed lake can be described (e.g. Gibson et al., 2002; Steinman et al., 2010) as the change in lake volume (V) per unit time (T), which is a function of the sum of water inputs (I) and outputs (Q), and may be written as

$$\frac{dV}{dT} = \sum I - \sum Q \qquad (1)$$

Water inputs to a lake comprise direct precipitation on the lake surface (P_L), surface runoff (S_i) and groundwater inflow (G_i). Water outputs include evaporation (E), surface outflow (S_q) and groundwater discharge (G_q), such that

$$\frac{dV}{dT} = P_L + S_i + G_i - E - S_q - G_q \qquad (2)$$

For Lake Ohrid, inputs for equation (2) are derived from previous investigations of lake and catchment hydrology as described below.

Precipitation (P_L)

The average annual precipitation recorded at meteorological stations situated throughout the Lake Ohrid catchment varies between 703 and 1445 mm/year, however for stations located close to the lake the yearly average is 773 mm/year (Figure 2; Watzin et al., 2002). Given that Lake Ohrid has a surface area of 358 km^2, the total amount of precipitation falling over the entire surface area of the lake is calculated to be 8.8 m^3/s.

Surface (S_i) and groundwater (G_i) inflow

The primary surface inflow to Lake Ohrid is the River Sateska, which has a total discharge of 7.2 m^3/s (Figure 2; Watzin et al., 2002). However, as the river was previously a direct tributary of the main outflow from Ohrid and diverted into the lake in 1962, we do not incorporate the Sateska inflow in our calculations here as our focus is on the long-term palaeo record (the impact on the isotope composition of lake water, δ_L, is discussed later). Other tributaries, for example the Pogradec, Koselka, and Verdova rivers, and catchment runoff have lower discharge rates totalling 7.2 m^3/s (Watzin et al., 2002).

Groundwater inflow to Lake Ohrid occurs through a network of surface and sub-lacustrine springs. The surface springs consist of three main complexes to the south and north-east of the lake, of which the largest is a collection of 15 springs located at the southern site of St Naum with an average discharge of 7.5 m^3/s (Figure 2; Popovska and Bonacci, 2007). To the west of St Naum, near the village of Tushemisht, a second zone comprising 80 springs has an annual discharge of 2.5 m^3/s and the Biljana springs to the north-east of Lake Ohrid have a discharge of 0.3 m^3/s (Watzin et al., 2002).

Artificial and environmental tracer experiments have shown that the water in surface springs is not solely derived from atmospheric precipitation in the catchment, as a proportion is transferred from nearby Lake Prespa through underground karst channels (Amataj et al., 2007; Eftimi et al., 2007). Lake Prespa has a higher surface area to volume ratio in comparison to Lake Ohrid and its waters have a more positive average isotope composition (Leng et al., 2010), which imparts a characteristic shift when combined with meteoric water in the underground karst system. Two-component mixing analysis conducted using stable isotope and Cl- data suggests that the ratio of water originating from Lake Prespa, compared to meteoric precipitation, is around 53% at the Tushemisht springs and 42% at St Naum (Table 1; Anovski et al., 1991; Eftimi and Zoto, 1997; Anovski, 2001; Eftimi et al., 2001; Matzinger et al., 2006a). The Biljana spring waters are derived solely from meteoric precipitation and not influenced by Lake Prespa (Eftimi et al., 2007).

The surface springs around Lake Ohrid receive approximately 4.5 m^3/s of water from Lake Prespa (Table 1), however total water outflow from Lake Prespa draining into the underground karst system is estimated to total 7.7 m^3/s (Anovski, 2001). The remaining 3.2 m^3/s of outflow from Lake Prespa is most likely transferred to Lake Ohrid through the sublacustrine network of springs along the lake's eastern margin (Matzinger et al., 2006a). A precise value for the total inflow derived from the sublacustrine springs is currently unknown. Matzinger et al. (2006b) assume a value of 9.9 m^3/s for total sublacustrine spring inflow, thereby implying a meteoric component of 6.7 m^3/s when the contribution from Lake Prespa is considered, however this value was determined by closing the balance rather than being a direct measure of flowrate. The meteoric component of sublacustrine spring inflow is therefore unknown and termed G_iX here.

Evaporation (E)

Although a direct measurement of evaporation is unavailable for Lake Ohrid, the rate can be estimated using the Linacre (1992) simplification of the Penman (1948) formula for open water evaporation:

$$E = [0.015 + 4 \times 10^{-4}\, T_a + 10^{-6}\, z] \times [480(\,T_a + 0.006\, z)/(84 - A) - 40 + 2.3\, u\,(\,T_a - T_d)]\ \text{(mm/day)} \qquad (3)$$

where T_a is the average air temperature (°C), z is the altitude, A is latitude, u is wind speed (m/s), and T_d is the dew point temperature (T_d = 0.52 T_{min} + 0.60 T_{max} – 0.009 T_{max}^2 – 2.0 °C). Based on climatological measurements between 1961 and 1990 at Pogradec (Figure 2), the average air temperature at Lake Ohrid is 11.7°C, average maximum temperature is 26.2°C, average minimum temperature is −0.8°C, and average wind speed is 2.3 m/s (Watzin et al., 2002). Using Equation 3 (Linacre, 1992), evaporation from Lake Ohrid is estimated to be 13.7

m^3/s, which is similar to a previous estimate calculated using Penman (1948) of 13.0 m^3/s (Watzin et al., 2002; Matzinger et al., 2006b).

Surface (S_q) and groundwater (G_q) outflow

The only surface outflow from Lake Ohrid is the river Crn Drim at the northern margin of the lake, which has a measured average discharge rate of 22 m^3/s (Watzin et al., 2002). When the diversion of the River Sateska is considered, and assuming that any increased outflow is directly proportional to increased inflow, the pre-1962 rate is taken to be 14.8 m^3/s.

Groundwater outflow from Lake Ohrid has not been observed to date (Wagner et al., 2008) and is not considered by previous water balance models (Watzin et al., 2002; Matzinger et al., 2006b). However, given that Triassic limestone crops out along the western margin of Lake Ohrid and the basin is characterised by active faulting (Reicherter et al., 2011; Lindhorst et al., 2015), the potential for a component of groundwater outflow should not be excluded.

Hydrological mass balance

A revised water balance for Lake Ohrid that includes estimates for groundwater fluxes into and out of the lake, based on data outlined above, is shown in Table 2. The unquantified component of groundwater input through the sublacustrine spring network sourced from meteoric precipitation is substituted as G_iX. If a steady state is assumed for Lake Ohrid, such that no change in lake volume is observed over a given period (dV/dT = 0), then the sum of water inputs is equal to the sum of water outputs and Equation (2) can be rewritten for Lake Ohrid:

$$P_L + S_i + G_iP + G_iS + G_iX = E + S_q + G_q \tag{4}$$

where G_i comprises the output from Lake Prespa (G_iP), and the measured surface spring (G_iS) and unknown sublacustrine spring (G_iX) components of groundwater inflow derived from meteoric precipitation (i.e. $G_i = G_iP + G_iS + G_iX$).

Using the revised water balance (Table 2) and Equation (4), the hydrological mass balance for Lake Ohrid may be written as

$$29.5 + G_i X = 28.5 + G_q \quad (5)$$

which can be simplified to

$$G_q - G_i X = 1.0 \text{ (m}^3\text{/s)} \quad (6)$$

Although the parameters G_iX and G_q cannot be directly measured, it is possible to calculate their values through isotope mass balance.

3.2 Isotope mass balance

The isotope mass balance of a lake is defined (e.g. Steinman et al., 2010; Gibson et al., 2016; Jones et al., 2016) as the sum of the products of water flux (P_L, S_i, G_i, E, S_q, G_q) and the isotope composition of the respective inflows (δ_{PL}, δ_{Si}, δ_{Gi}) and outflows (δ_E, δ_{Sq}, δ_{Gq}), which can be expressed as

$$\frac{dV\delta_L}{dT} = P_L\delta_{PL} + S_i\delta_{Si} + G_i\delta_{Gi} - E\delta_E - S_q\delta_{Sq} - G_q\delta_{Gq} \quad (7)$$

Isotope composition of inflows (δ_{PL}, δ_{Si}, δ_{Gi})

As part of an IAEA Regional Project the isotope composition of precipitation falling directly on the lake's surface (δ_{PL}) was measured at the St Naum spring complex, which determined that mean annual weighted $\delta^{18}O$ = –8.4 ‰ and δD = –52.9 ‰ (Figure 3; Anovski, 2001).

We take $\delta^{18}O$ = –10.1±0.5 ‰ and δD = –67.4±3.1 ‰, average spring water values from data collected periodically over a 30-year period (Figure 3; Anovski et al., 1980, 1991, 2001; Eftimi and Zoto, 1997; Leng et al., 2010), to represent the isotope composition of surface and groundwater inflows fed directly by atmospheric precipitation (δ_{IN}), such that $\delta_{IN} = \delta_{Si} = \delta_{Gi}$.

These values are more negative than for δ_{PL} as infiltration will be principally derived from precipitation at higher altitudes in the Ohrid catchment (Anovski, 2001), which rises to approximately 1600 m above lake level in the Galičica mountain range separating Lake Ohrid and Lake Prespa (Francke et al., 2016). In addition, a large proportion of the precipitation across the catchment likely falls as snow. The pattern of annual discharge of the River Sateska is at a maximum in early spring following snowmelt (Figure 2; Matzinger et al., 2006b). Snow is typically characterised as having a lower isotope composition than the equivalent rainfall as it reflects fractionation at lower temperatures at within-cloud conditions (Gat, 1996; Darling et al., 2006; Dean et al., 2013).

The Prespa-fed component of surface and sublacustrine springs is assumed to be homogenous with Prespa lakewater (δ_{LP}), which has been measured over a 30-year period to have average $\delta^{18}O = -1.5 \pm 0.6$ ‰ and $\delta D = -20.5 \pm 3.6$ ‰ (Figure 3; Leng et al., 2010).

Isotope composition of outflows (δ_E, δ_{Sq}, δ_{Gq}*)*

The isotope composition of evaporation (δ_E) is difficult to measure directly, and so is typically calculated using the Craig and Gordon (1965) evaporation model (e.g. Steinman et al., 2010):

$$\delta_E = \frac{(\alpha^* \times \delta_L) - (h \times \delta_A) - \varepsilon}{1 - h + (0.001 \times \varepsilon_K)} \quad (8)$$

where α^* is the reciprocal of the equilibrium isotope fractionation factor (α) calculated for $\delta^{18}O$ (eq. 9) and δD (eq. 10) using the equations of Horita and Wesolowski (1994), and T_W is the temperature of lake surface water (in degrees K) assumed to be 287.2 K (Stankovic, 1960).

$$\ln \alpha = 0.35041\left(\frac{10^6}{T_W{}^3}\right) - 1.6664\left(\frac{10^3}{T_W{}^2}\right) + 6.7123\left(\frac{1}{T_W}\right) - 7.685 \times 10^{-3} \quad (9)$$

$$\ln \alpha = 1.1588\left(\frac{T_W{}^3}{10^9}\right) - 1.6201\left(\frac{T_W{}^2}{10^6}\right) + 0.79484\left(\frac{T_W}{10^3}\right) + 2.9992\left(\frac{10^6}{T_W{}^3}\right) - 161.04 \times 10^{-3} \quad (10)$$

The normalised relative humidity (h; eq. 11) is the quotient of the saturation vapour pressure of the overlying air ($e_{s\text{-}a}$) and the saturation vapour pressure at the surface water temperature ($e_{s\text{-}w}$) (eq. 12; Steinman et al., 2010), which relates measured relative humidity (RH = 72.0%) to average annual temperature (T) of air (11.7°C) or lake water (14.0°C).

$$h = RH \times \frac{e_{s-a}}{e_{s-w}} \tag{11}$$

$$e_{s-a\ \&\ s-w} = 6.108 \times e^{\frac{17.27 \times T}{T+237.7}} \tag{12}$$

The isotope composition of atmospheric moisture (δ_A) is assumed to be in equilibrium with precipitation (eq. 13). The equilibrium isotopic separation factor (ε^*; eq. 14) is the difference between the isotope composition of precipitation and atmospheric moisture (Gibson et al., 2002), which is known to be a function of temperature (eq. 9 and 10; Gonfiantini, 1986).

$$\delta_A = \delta_P - \varepsilon^* \tag{13}$$

$$\varepsilon^* = 1000 \times (1 - \alpha^*) \tag{14}$$

In addition to ε^*, the total isotope separation factor (ε; eq. 15) also comprises a kinetic component (ε_K; Gibson et al., 2002), which is constrained for both oxygen and hydrogen (eq. 16 and 17; Gonfiantini, 1986).

$$\varepsilon = \varepsilon^* + \varepsilon_K \tag{15}$$

$$\varepsilon_K = 14.2 \times (1 - h) \text{ for } \delta^{18}\text{O} \tag{16}$$

$$\varepsilon_K = 12.5 \times (1 - h) \text{ for } \delta\text{D} \tag{17}$$

In larger lakes, such as Ohrid, evaporation can have a significant influence on the overlying atmosphere producing a moisture feedback, and it is therefore important to consider the effects on kinetic fractionation (eq. 18). As lakewater evaporates, the fraction (*f*) of evaporate incorporated in the overlying atmosphere modifies δ_A by the addition of δ_E to form δ'_A (Gibson et al., 2016).

$$\delta'_A = (1-f)\delta_A + f\delta_E \tag{18}$$

In a feedback system, the modified isotope composition of evaporation (δ'_E) is therefore defined as:

$$\delta'_E = \frac{(\alpha^* \times \delta_L) - (h \times \delta'_A) - \varepsilon}{1 - h + (0.001 \times \varepsilon_K)} \tag{19}$$

In addition to evaporation, outflow through the river Crn Drim (δ_{Sq}) and any groundwater flux (δ_{Gq}) is assumed to have the same isotope composition as average δ_L, where $\delta^{18}O$ = –3.5±0.3 ‰ and δD = –31.7±1.6 ‰ (Figure 3; Anovski et al., 1980, 1991; Eftimi and Zoto, 1997; Matzinger et al., 2006b; Leng et al., 2010).

Isotope mass balance

The revised water balance (Table 2) allows Equation (7) to be re-expressed for Lake Ohrid:

$$\frac{dV\delta_L}{dT} = P_L\delta_{PL} + S_i\delta_{IN} + G_iP\delta_{LP} + G_iS\delta_{IN} + G_iX\delta_{IN} - E\delta_{E'} - S_q\delta_L - G_q\delta_L \tag{20}$$

Assuming lake volume is constant through time, such that $dV\delta_L/dT = 0$, Equation (20) can be re-expressed and simplified to

$$P_L\delta_{PL} + (S_i + G_iS + G_iX)\delta_{IN} + G_iP\delta_{LP} = E\delta_{E'} + \left(S_q + G_q\right)\delta_L \tag{21}$$

Given the above, by then iteratively solving equations 18 and 19, for both $\delta^{18}O$ and δD with varying *f*, and by simultaneously evaluating equations 6 and 21 until G_iX and G_q converge, for both $\delta^{18}O$ and δD scenarios, a balanced model, both hydrologically and isotopically, for Lake Ohrid can be obtained.

4. Results and discussion

4.1. Isotope mass balance

The iterative calculation of G_iX and G_q suggests flow rates of 15.3 and 16.3 m^3/s, respectively, providing a new estimate of water balance for Lake Ohrid (Table 3). Sublacustrine spring inflow of 15.3 m^3/s is approximately 50% higher than in existing hydrological models for the lake (e.g. Matzinger et al., 2006b), and groundwater outflow, previously assumed to be negligible, of 16.3 m^3/s comprises roughly a third of total water output from Lake Ohrid. For conservation of isotope mass balance the fraction of evaporate incorporated into the overlying atmosphere is approximately $f = 33\%$ (Figure 4), which is of a consistent order of magnitude with other larger lakes such as Lake Superior (40%), Lake Michigan (33%), and Lake Ontario (27%) (Jasechko et al., 2014). The new water balance gives total water output (evaporation, surface and groundwater outflow; Table 3) from Lake Ohrid to be 44.8 m^3/s, which, combined with the lake's volume (50.7 km^3), suggests a calculated water residence time for the lake of approximately 36 years. As the entire water column experiences complete overturn once every 7 years and the upper 200-m on an annual basis (Matzinger et al., 2006b, 2007), the lake water mixes completely several times within the calculated residence time, which may be lower than actual residence time by up to a factor of 4 (Ambrosetti et al., 2003; Wagner et al., 2017). Further, the new calculated value for total water input is < 3% of the overall lakewater volume, and given the lake is well-mixed within its water residence time, any seasonal and inter-annual variations in δ_L will likely be buffered by the large volume and long residence time. This is highlighted by the contemporary monitoring data (Figure 3), which show that δ_L has remained very consistent over the past 30 years and that the lake is an isotopically well-mixed system ($\delta^{18}O = -3.5 \pm 0.3$ ‰; Leng et al., 2013 and references therein).

4.2 Estimating past hydrological balance

Isotope-based reconstructions of past climate require a good understanding of the contemporary hydrological system, and by using the established stable isotope mass balance model for the modern environment (as presented above) we can use isotope measurements from core sequences to give quantitative estimates of past changes in the hydrological balance at Lake Ohrid. Over the past 640 ka, one of the largest changes in reconstructed $\delta^{18}O_L$ is between the last glacial and the Holocene (Lacey et al., 2016). Average $\delta^{18}O_L$ during the last glacial is roughly 3 ‰ more negative than during the Holocene (Lacey et al., 2016), which is the same magnitude of change as indicated for neighbouring Lake Prespa (Leng et al., 2013). This substantial shift in $\delta^{18}O_L$ could be related to changes in moisture availability, which is also suggested to be a primary driver of changes in catchment vegetation (Lézine et al., 2010; Panagiotopoulos et al., 2014; Sadori et al., 2016). Moisture availability is important for sustaining tree populations and it has been suggested that the Lake Ohrid catchment received enough moisture to enable the survival of arboreal vegetation, even during glacial periods (Sadori et al., 2016). However, glacial phases are typically characterised by more positive $\delta^{18}O$ values in central and eastern Mediterranean lake sequences (Roberts et al., 2008; Giaccio et al., 2015) and in speleothem records (Regattieri et al., 2018).

To better qualify the extent of water availability, and the precipitation changes that control it, across this time frame, we reconstruct here the change in precipitation during the last glacial and the early Holocene using the stable isotope mass balance model for Lake Ohrid and compare the output to other regional records and climate models.

Precipitation (P_L), surface inflows (S_i) and groundwater inflow (G_i)

Values for P_L at Lake Ohrid during the early Holocene and last glacial are unknown. The inflows S_i, G_iS, and G_iX are all a component of catchment-derived meteoric precipitation, therefore the values can be represented by a single inflow term, I_i, where:

$$I_i = S_i + G_iS + G_iX \tag{22}$$

If precipitation over the catchment increases or decreases, P_L will change together with concomitant change in the components of I_i. For estimating past hydrological balance, we assume that variations in P_L and I_i during the early Holocene and last glacial are consistent with their present-day ratio. As P_L is equivalent to 8.8 m^3/s and I_i to 27.8 m^3/s (Table 3; Equation 22), then:

$$I_i = 3.2 \times P_L \tag{23}$$

We take the present outflow from Lake Prespa (G_iP) to be constant for the early Holocene and last glacial at 7.7 m^3/s.

Evaporation (E), surface outflow (S_q), and groundwater outflow (G_q)

To estimate past rates of evaporation, a pollen record from nearby Lake Maliq can be used to evaluate local temperature change. The average temperature difference between the present and early Holocene is reconstructed to be −1°C and for the last glacial −7°C (Bordon et al., 2009), which is consistent with the reconstructed pattern of regional temperature change (Davis et al., 2003), also from pollen data. Although there is no way to calculate palaeo-wind speeds, Jones et al. (2007) suggest Late Glacial average wind velocities may be double that of contemporary measured values in the eastern Mediterranean. At Lake Ohrid, maxima in Cr/Ti and Zr/Ti infer stronger wind activity during glacial periods (Vogel et al., 2010), and so a value of 4.6 m/s is used for last glacial wind speed. If it is assumed that maximum and minimum temperatures are similarly reduced as for average temperature, evaporation decreases to 12.3 m^3/s during the early Holocene and to 6.3 m^3/s in the last glacial.

Surface and groundwater outflow for the early Holocene and last glacial are unknown, but as both are a function of lakewater export (Q_q), the parameters S_q and G_q can be combined:

$$Q_q = S_q + G_q \tag{24}$$

Isotope composition of inflows (δ_{PL}, δ_{IN}, δ_{LP})

In the Mediterranean region, contemporary rainfall isotope data show a positive correlation between temperature and the isotope composition of precipitation (δ_P) of around +0.3 ‰/°C, which compares well with simulated palaeo relationships at the Last Glacial Maximum (LGM; Bard et al., 2002; Zanchetta et al., 2007). There is also a correspondence between the amount of precipitation and δ_P, where modern $\delta^{18}O_P$ decreases by −1.6 ‰ for every 100 mm increase in monthly precipitation (Bard et al., 2002), however as changes in P_L are unknown the amount effect is discussed later. Given the temperature reconstruction from nearby Lake Maliq (Bordon et al., 2009), this implies that δ_P would have been −2.1 ‰ lower in the last glacial (−7 °C) when compared to the late Holocene. When considering glacial-interglacial shifts in δ_P, changes at the source of δ_P must also be taken in to account. Glacial seawater was roughly 1 ‰ higher on average during the LGM due to the expansion of global ice volume (Schrag et al., 2002), and local evaporative enrichment in the Mediterranean resulted in a change of +1.2 ‰ (Paul et al., 2001). In the Ionian Sea, west of Lake Ohrid, the glacial-interglacial change in $\delta^{18}O$ is estimated to be nearer to +1.3 ‰ (Emeis et al., 2000). This suggests that the combined effect of temperature and source $\delta^{18}O$ changes between the last glacial and late Holocene would therefore be approximately −0.8 ‰. Assuming that $\delta^{18}O$ of Mediterranean seawater in the Holocene had a relatively similar isotope composition to today, as observed for the Ionian Sea (Emeis et al., 2000), we take early Holocene (−1 °C) δ_P to be −0.3 ‰ compared to late Holocene values.

As temperatures in the early Holocene were similar to those at present, we assume a comparable precipitation regime (rainfall vs. snowfall) and take the variation in δ_P of −0.3 ‰ to also apply for δ_{IN}. However, in the last glacial, much of the precipitation at higher altitudes across the

Ohrid-Prespa catchment may have fallen as snow, as indicated by climate model simulations for the region at the Last Glacial Maximum (Robinson et al., 2006). The snow may have also been incorporated into ice sheets during phases of glacial expansion (Ribolini et al., 2011). Snowfall reflects equilibrium conditions at the point of in-cloud formation and so comprises significantly lower $\delta^{18}O$ (Darling et al., 2006), which is highlighted by Dean et al. (2013) who report snowfall $\delta^{18}O$ of around −16 ‰ in the catchment of Lake Nar in central Turkey, compared to typical average rainfall values of around −10.6 ‰ (Jones et al., 2005). As much of the present stream and spring inflow to Lake Ohrid is fed by higher altitude precipitation over the catchment and spring snowmelt (Matzinger et al., 2006b), we approximate δ_{IN} during the last glacial to $\delta^{18}O = -16$ ‰.

The transfer of water from Lake Prespa to Lake Ohrid during the early Holocene and last glacial is assumed constant, although δ_{LP} will vary between the two intervals. Measured $\delta^{18}O$ for endogenic calcite (Holocene) and authigenic siderite (last glacial) is available from cores recovered from Lake Prespa (core Co1215; Figure 1), where $\delta^{18}O_{calcite}$ in the early Holocene is −2.8 ‰ and $\delta^{18}O_{siderite}$ in the last glacial is −1.4 ‰ (Figure 5; Leng et al., 2010, 2013). Assuming a temperature of 19°C for summer lakewater at Prespa (time of endogenic calcite precipitation) in the early Holocene and 4.7°C (air temperature) for glacial bottom water (environment of authigenic siderite precipitation), δ_{LP} is calculated to be −2.1 ‰ for the early Holocene (using Hays and Grossman, 1991) and −5.8 ‰ during the last glacial (using Zhang et al., 2001).

δ^2H is estimated for δ_{PL}, δ_{IN}, and δ_{LP} using the modern local evaporation line ($\delta^2H = 5.4\delta^{18}O - 12.8$), defined by water measurements collated over a ca. 30-year period (Anovski et al., 1980, 1991; Eftimi and Zoto, 1997; Anovski, 2001; Matzinger et al., 2006b; Jordanoska et al., 2010; Leng et al., 2010, 2013).

Isotope composition of outflows (δ_E, δ_{Sq}, δ_{Gq}*)*

The isotope composition of evaporation is calculated iteratively using equations 18 and 19, and a variable *f*. This is achieved by simultaneously evaluating hydrological and isotope mass balance equations 25 and 26 to solve for P_L and Q_q (Figure 4), which are balanced for both $\delta^{18}O$ and δD as in the present-day mass balance model.

$$4.2P_L + G_iP = E + Q_q \tag{25}$$

$$P_L\delta_{PL} + 3.2P_L\delta_{IN} + G_iP\delta_{LP} = E\delta_{E'} + Q_q\delta_L \tag{26}$$

Equations 25 and 26 are derived by combining equations 4 and 21 with equations 23 and 24, respectively. To calculate δ_E (for use in Equation 18), we take the same temperature change as for calculating E and assume a relative humidity of 73% for the early Holocene (based on the present relationship between RH and temperature) and 50% for the last glacial. Over interglacial-glacial timescales, relative humidity is suggested to reduce with decreasing temperatures as less moisture is available due to lower evaporation rates (Lemcke and Sturm, 1997; Jones et al., 2007), and lower RH during the last glacial is confirmed for the Balkan region by a pollen-based humidity-index from the Aegean Sea (Kouli et al., 2012).

To determine the past isotope composition of lakewater outflow, assumed to be equivalent to δ_L during respective time periods, we use measured $\delta^{18}O_{calcite}$ of −6.0 ‰ for the early Holocene (average for 8.5-9 ka from core Co1262; Lacey et al., 2015) and measured $\delta^{18}O_{siderite}$ of −4.0 ‰ for the last glacial (average for 16-42 ka from core 5045-1, Figure 5; Lacey et al., 2016). As for the Lake Prespa calculations, we assume a temperature of 19°C for summer lake water in the early Holocene and 4.7°C for glacial bottom water. Conversion to $\delta^{18}O_L$ gives −5.3 ‰ during the early Holocene and −8.1 ‰ during the last glacial (calculated using Hays and Grossman, 1991; Zhang et al., 2001).

Model output and sensitivity tests

The calculated hydrological balance for Lake Ohrid during the early Holocene and last glacial period is given in Table 4. The iterative calculation of P_L suggests that precipitation was around 26% higher in the early Holocene (11.1 m^3/s or 978 mm/year) and 44% lower in the last glacial (4.9 m^3/s or 432 mm/year), in comparison to the late Holocene (Table 4).

The hydrological balance model output is dependent on estimates of past temperature, evaporation, wind speeds, and δ_P (including the isotope composition and seasonality of precipitation). The possible variability in these parameters, and any influence this may have on the calculation of past hydrological balance, can be assessed using sensitivity tests. Palaeotemperatures are estimated from a reconstruction based on a nearby pollen sequence (Bordon et al., 2009), and influence the calculation of evaporative flux from the lake and the amount of direct precipitation (P_L) through the iterative calculation of $\delta'E$. The calculation of past evaporative flux uses estimates for temperature and wind speed, and sensitivity analysis suggests that wind speed has less influence on evaporation compared to changes in temperature (Figure 6). Wind speeds would have to increase to ~22 m/s (assuming mean air temperature = 4.7 °C), or average temperature would have to be similar to the early Holocene (10.4 °C; assuming wind speed = 4.6 m/s), before evaporation during the last glacial was equivalent to the modern evaporative flux. Changing the estimate of past temperature by ±2°C for the iterative calculation of $\delta'E$ and P_L (i.e. twice the reconstructed change between the early and late Holocene; Bordon et al., 2009) suggests that temperature changes do not overly effect the resulting value for P_L, where a +1°C change leads to +1 m^3/s in P_L in the early Holocene (Table 5). This also assumes a simultaneous change in evaporation, with other parameters held constant, as varying air temperature influences the calculation of evaporative flux. In the late Glacial, temperature changes have less effect on P_L than in the early Holocene, as a +1°C change leads to around +0.7 m^3/s in P_L (Table 5).

In the last glacial, δ_{IN} is approximated to $\delta^{18}O = -16$ ‰ as high-altitude precipitation in the catchment and snowmelt is a major component of stream and spring inflow to Lake Ohrid, so a greater component of annual precipitation would likely comprise snowfall during colder glacial phases. Sensitivity analysis shows that changes in δ_{IN} only have a minor effect on the calculated value for P_L, where varying $\delta^{18}O_{IN}$ between −16 ‰ and −13 ‰ produces up to a 2.9 m^3/s change in P_L (Figure 6). It is only when δ_{IN} approaches δ_{PL} that larger variations in P_L are predicted, however δ_{IN} will always be lower than δ_{PL} due to an altitude effect and the elevation difference between Lake Ohrid and its catchment. Therefore, reduced P_L during the last glacial is possible even when seasonality changes in precipitation are taken into account.

The sensitivity analysis of changing P_L with respect to δ_{PL} also suggests that the amount effect, where variable P_L forces changes in δ_{PL}, would only drive small changes in calculated values for P_L. The relationship between modern monthly precipitation and $\delta^{18}O$ is around −1.6 ‰ per 100 mm change (based on Global Network for Isotopes in Precipitation data from Pisa, Genoa, and Palermo), although this may have been up to −3.9 ‰ per 100 mm at the LGM (Bard et al., 2002). The estimated change in P_L in the early Holocene of +205 mm/year may therefore equate to an amount effect of between −0.3 ‰ and −3.3 ‰, depending on the seasonality of additional precipitation. If these values are taken into account and lower δ_{PL} and δ_{IN} are incorporated into the model for the early Holocene the estimated value for P_L is reduced, which is unlikely given regional precipitation reconstructions for this time (e.g. Brayshaw et al., 2011; Peyron et al., 2017). Similarly, during the last glacial, the estimated change of −341 mm/year may equate to an amount effect of between +0.5 ‰ and +13.3 ‰, depending on the seasonality of precipitation and whether the contemporary or LGM relationship is considered. Sensitivity analysis for the last glacial suggest that varying δ_{PL} results only in a minor change in P_L, and only if the seasonality of precipitation was such that δ_{IN} approached δ_{PL} (i.e. restricted snowfall) would changes in δ_{PL} be overly influenced by the amount effect (Figure 6).

4.3 Past hydrological balance

Greater precipitation in the early Holocene is consistent with the shift to lower $\delta^{18}O$ observed in lake carbonate and speleothem records from across the Balkan Peninsula (Constantin et al., 2007; Francke et al., 2013; Leng et al., 2013; Drăguşin et al., 2014), and a regional shift to lower $\delta^{18}O$ across other Mediterranean lake records (Lamb et al., 1989; Frogley et al., 2001; Zanchetta et al., 2007; Roberts et al., 2011; Dean et al., 2015). This is further supported by lake level reconstructions from Italy and Greece that indicate deeper water conditions during the early Holocene (Digerfeldt et al., 2000; Magny et al., 2007, 2011; Joannin et al., 2012), and increased river discharge into the Gulf of Salerno (Naimo et al., 2005). Pollen-inferred reconstructions of precipitation from marine and terrestrial records show a wetter regional climate regime across the central and eastern Mediterranean during the early Holocene (Peyron et al., 2017), where rainfall is estimated to have been roughly 20% higher than present in central Anatolia and the southern Levant based on other isotope records (Bar-Matthews et al., 2003; Jones et al., 2007). Global and regional climate model simulations also suggest that the southern Balkan Peninsula experienced one of the largest increases in rainfall during the early Holocene set against stronger precipitation across the Mediterranean region as a whole compared to the present (Brayshaw et al., 2011).

The substantial decrease in precipitation calculated for the last glacial period is broadly consistent with pollen-based rainfall estimates for the Late Glacial and Younger Dryas from nearby Lake Maliq (~ 300 mm/year; Bordon et al., 2009), and a 50% reduction in winter precipitation between the Late Glacial and early Holocene over the borderlands of the Aegean Sea (Kotthoff et al., 2008). Model simulations of past climates for the last glacial, typically focussed on the Last Glacial Maximum (ca. 21 ka), indicate reduced precipitation relative to present day, but also suggest that evaporation still likely exceeded precipitation at this time (Robinson et al., 2006), which may be due a southward shift in Mediterranean storm tracks

(Goldsmith et al., 2017). The pollen record from Lake Ohrid suggests that glacial periods were typically characterised by cold and dry conditions, as shown by the dominance of non-arboreal pollen indicative of an open environment, which was dominated by steppes and steppe forests during the last two glacial periods (Sadori et al., 2016). However, even during glacial periods, environmental conditions at Lake Ohrid did not appear to cross ecological tolerance thresholds as most arboreal taxa have a continuous presence in the record over the past ca. 500 ka. This suggests that the lake's catchment may have acted as a refugium area for tree populations (Sadori et al., 2016), similar to Lake Ioannina in western Greece (Tzedakis et al., 2002), but in contrast to other eastern Mediterranean sites where arboreal taxa often disappear during glacials due to a more continental climate and lower moisture availability (Okuda et al., 2001; Tzedakis et al., 2004). At Lake Ohrid, the calculated annual precipitation of around 432 mm/year (or 4.9 m^3/s) during the last glacial (Table 4) is above the threshold of approximately 300 mm for the survival of temperate tree populations (e.g. Zohary, 1973). Rainfall is also observed to be greater across the catchment compared to directly over the lake, as average rainfall across the watershed is 907 mm, whereas direct precipitation on the lake is lower at 773 mm (Watzin et al., 2002; Popovska and Bonacci, 2007). This suggests that the calculated value for direct precipitation of 432 mm during the last glacial will be lower than for the catchment as a whole. In addition, the fraction of evaporate added to overlying atmospheric vapour is calculated to be only slightly higher than present for the early Holocene (0.36), but is estimated to be much higher for the last glacial (0.73), suggesting that the lake would have provided additional moisture to its surroundings during dry phases. Therefore, the estimate for last glacial precipitation supports the suggestion of Sadori et al. (2016) that refugial conditions most likely occurred in the Lake Ohrid catchment area during glacial periods.

5. Conclusions

This work provides an improved, quantitative understanding of the modern isotope hydrology of Lake Ohrid by re-evaluating groundwater fluxes, which is helpful for explaining the systematics of past climate variations recorded in proxy records from the lake. By incorporating contemporary isotope data into hydrological and isotope mass balance models, we have been able to provide a more robust estimate for the water balance of Lake Ohrid. The new model incorporates underground inflow and outflow components that cannot be directly measured. Groundwater inflow through sublacustrine springs derived from meteoric precipitation is calculated to be 15.3 m^3/s, which is around 50% more than predicted in previous water balance models. Groundwater outflow, previously assumed to be negligible, is estimated to be 16.3 m^3/s and comprise roughly a third of outflow from Lake Ohrid. The new estimate of groundwater outflow decreases the importance of evaporation at only a third of total water output. Therefore, overall changes in the amount of precipitation, and associated changes in throughflow, may have greater influence on $\delta^{18}O$ rather than isotope variations being intrinsically linked to changes in the precipitation to evaporation ratio.

Estimated values for hydrological balance in the early Holocene suggest that precipitation at Lake Ohrid was up to 26% higher than the value for present day, which is consistent with local and regional palaeoclimate records and climate model simulations. Precipitation during the last glacial is calculated to have been around 44% lower than present. The model also suggests that during recent glacial phases the reconstructed shift to low $\delta^{18}O_L$ from sediment core data can be accounted for, even when precipitation is greatly reduced. This assumes that the majority of precipitation fell in winter as snow within the Lake Ohrid catchment, similar to climate model predictions for the Last Glacial Maximum. The amount of precipitation during the last glacial was above the critical threshold to support the continuous presence of arboreal vegetation within the catchment, suggesting that refugial conditions existed even through glacial phases.

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Figure 1 (A) Location of Lake Ohrid and Lake Prespa on the Balkan Peninsula (black rectangle), and (B) map of the Ohrid and Prespa basins, showing bathymetry of Lake Ohrid (Lindhorst et al., 2015). The location of coring sites mentioned in the text (5045-1, Co1262, Co1215; Leng et al., 2013; Lacey et al., 2015, 2016; Wagner et al., 2017) are shown by white squares.

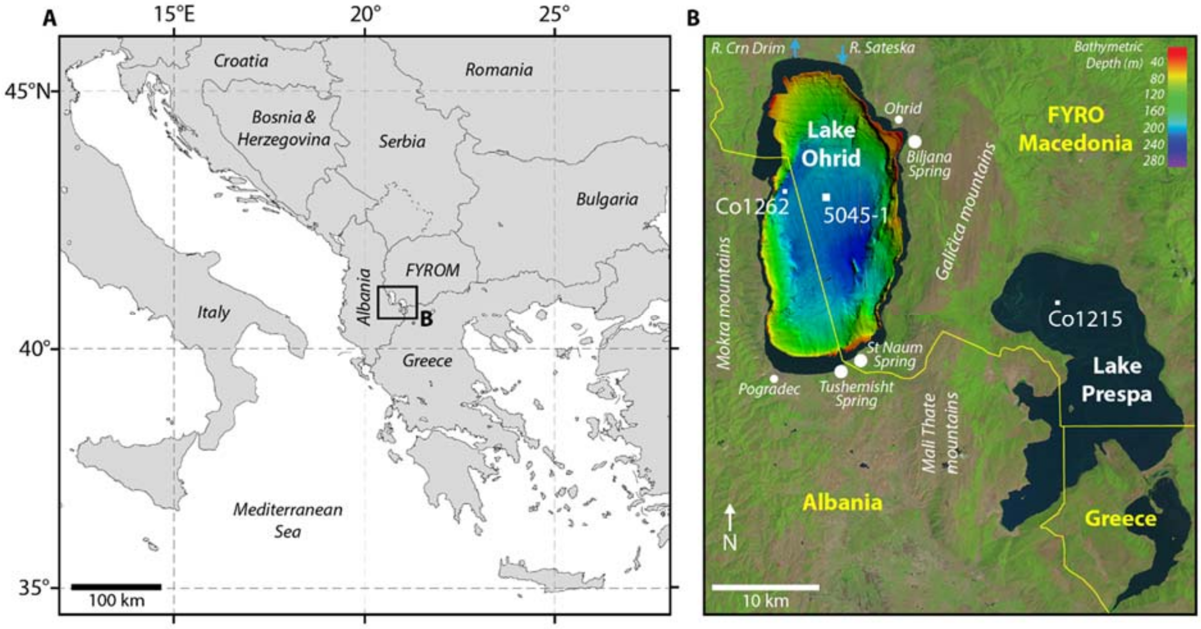

Figure 2 Climatological indices of air temperature (T) and precipitation (P) from the meteorological station at Pogradec, Albania (1961-1990; Watzin et al., 2002), relative humidity (RH) from the station at Bitola (1972-1977; Outcalt and Allen, 1982), and the seasonal discharge of the River Sateska (1996-2000; Matzinger et al., 2006b).

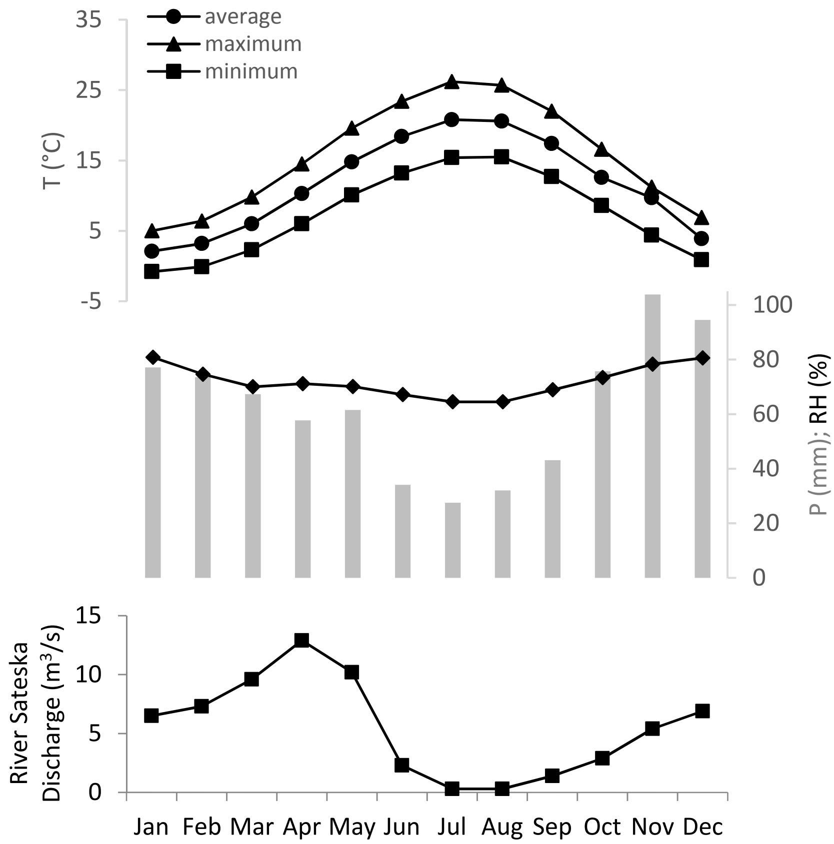

Figure 3 Modern isotope composition ($\delta^{18}O$ and δD) of water from lakes Ohrid and Prespa, springs, and local direct/catchment rainfall (Anovski et al., 1980; Anovski et al., 1991, 2001; Eftimi and Zoto, 1997; Matzinger, 2006b; Jordanoska et al., 2010; Leng et al., 2010, 2013). The global meteoric water line (GMWL; Craig, 1961), local meteoric water line (LMWL; Anovski et al., 1991, Eftimi and Zoto, 1997), and calculated local evaporation line (LEL) are shown.

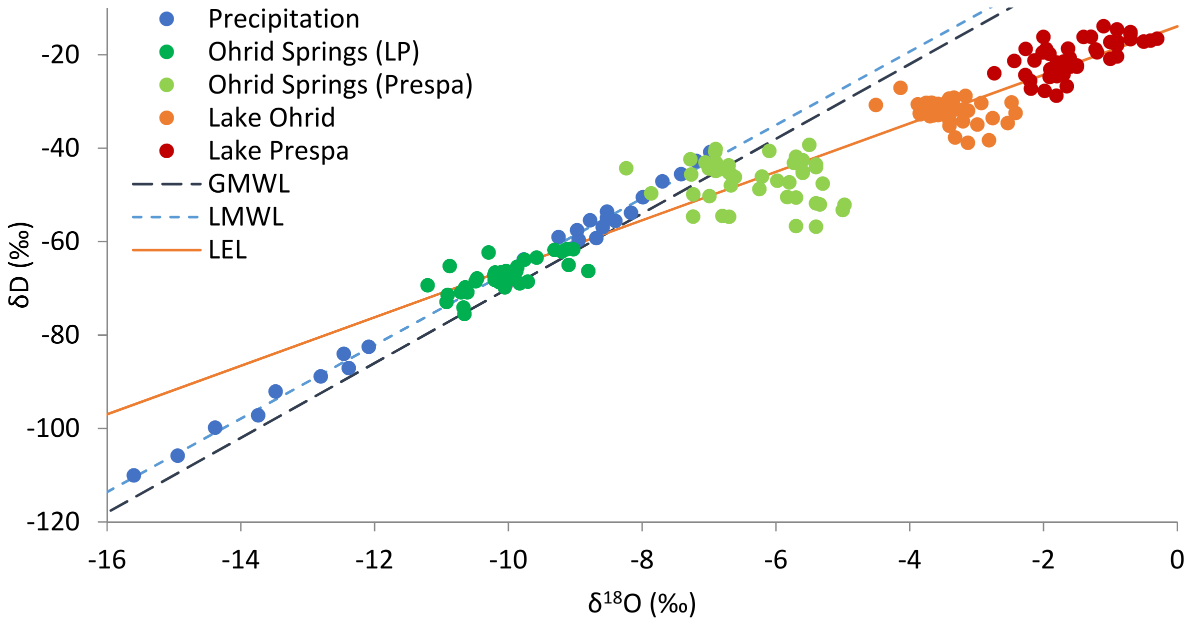

Figure 4 Iterative calculation of evaporation by using variable f and simultaneously evaluating hydrological and isotope mass balance equations to solve for G_iX and G_q (recent/Late Holocene), and P_L and Qq (Early Holocene and last glacial), which are balanced for both $\delta^{18}O$ and δD.

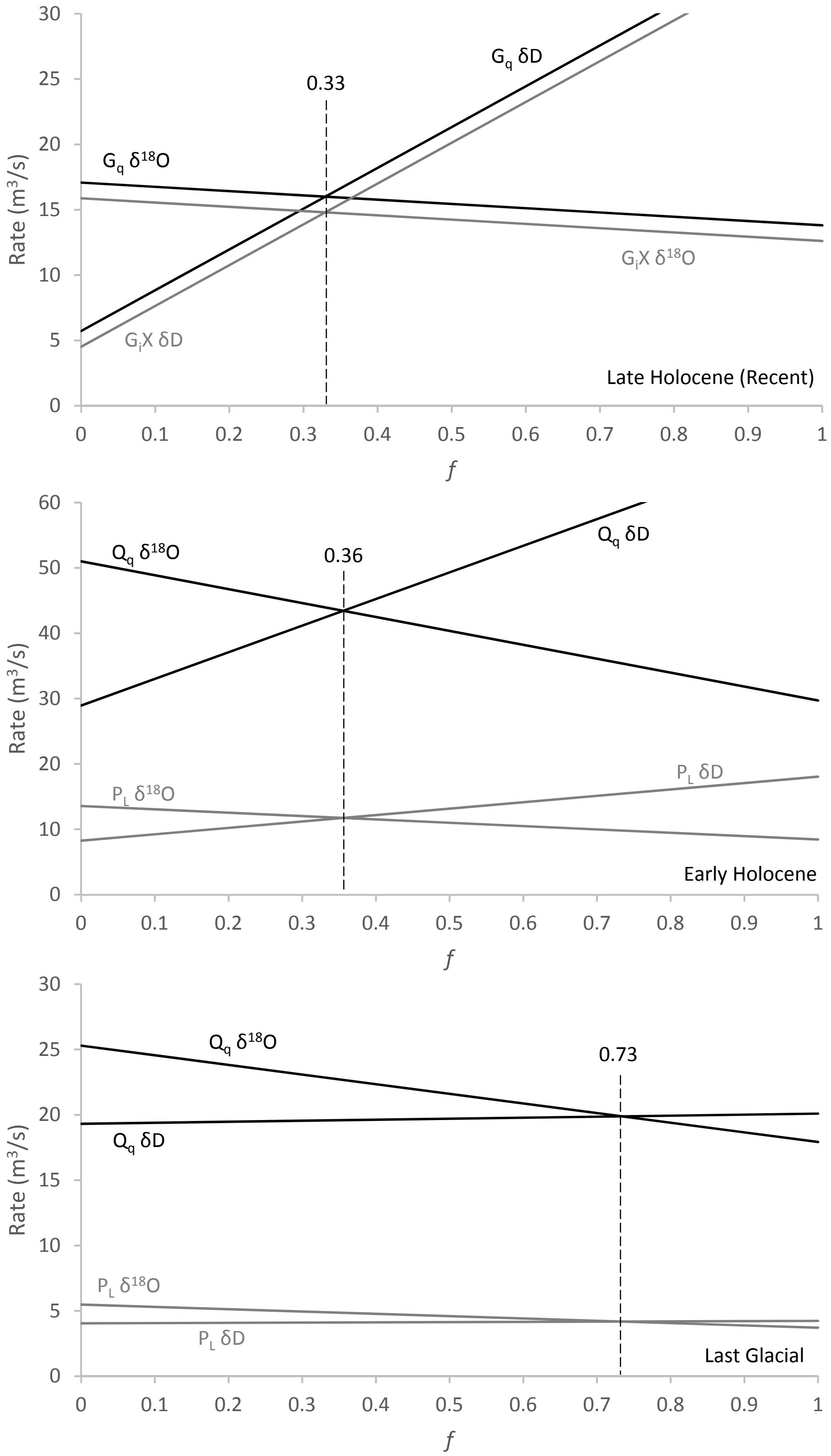

Figure 5 Reconstructed oxygen isotope composition ($\delta^{18}O$) of lakewater from Lake Ohrid cores Co1262 and 5045-1 (Lacey et al., 2015, 2016) and Lake Prespa core Co1215 (Leng et al., 2013). $\delta^{18}O$ lakewater is calculated from calcite and siderite isotope data (see text for calculation).

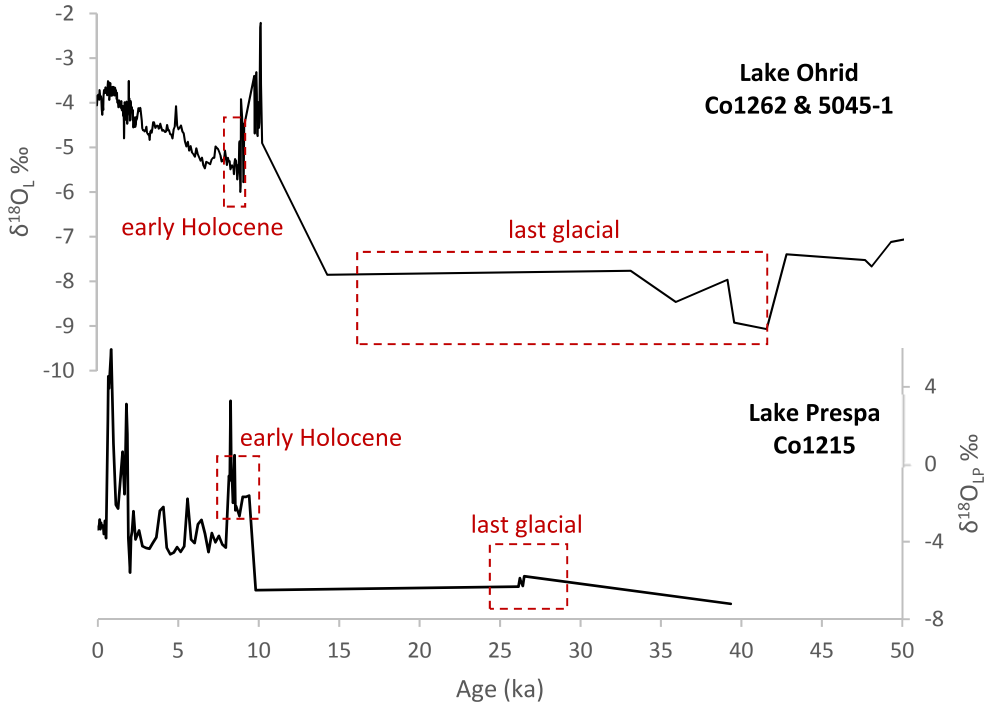

Figure 6 Sensitivity of A) evaporation to changing wind speed at different temperatures for the early Holocene and last glacial, and B) precipitation over the lake (P_L) to changing $\delta^{18}O_{IN}$ at different $\delta^{18}O_{PL}$ during the last glacial (all other parameters remain constant).

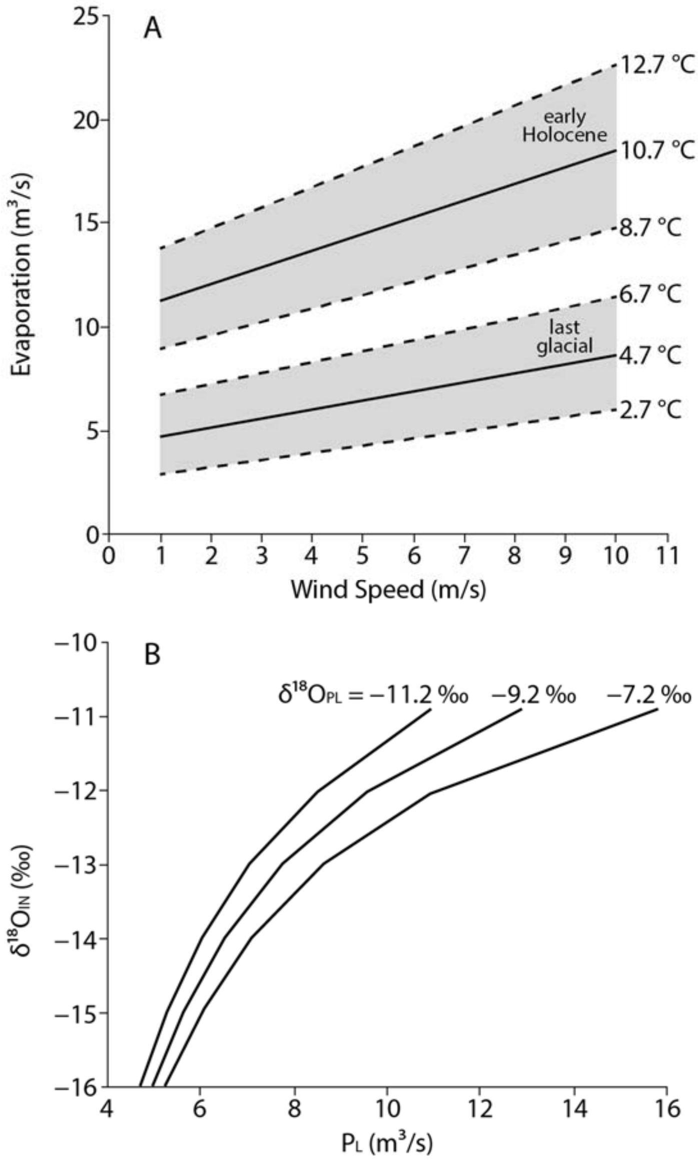

Table 1 Relative proportions of meteoric precipitation and outflow from Lake Prespa comprising spring inflow to Lake Ohrid.

Spring Complex	Component of Spring Inflow (m^3/s)	
	Meteoric Precipitation	Lake Prespa Outflow
St. Naum	4.3 (58%)	3.2 (42%)
Tushemisht	1.2 (47%)	1.3 (53%)
Biljana	0.3 (100%)	0 (0%)
Total	5.8 (56%)	4.5 (44%)

Table 2 Revised water balance for Lake Ohrid, including estimates for groundwater fluxes into and out of the lake. G_iX is the component of groundwater inflow sourced from meteoric precipitation, and GQ is groundwater outflow.

Source	**Flow rate (m^3/s)**
Inputs	
Precipitation (P_L)	8.8
Surface inflow (S_i)	7.2
Groundwater inflow (G_i)	
− Prespa-fed (G_iP)	7.7
− Surface springs (G_iS)	5.8
− Sublacustrine springs (G_iX)	G_iX
	29.5 + G_iX
Outputs	
Evaporation (E)	13.7
Surface outflow (S_q)	14.8
Groundwater outflow (G_q)	G_q
	28.5 + G_Q

Table 3 New water balance for Lake Ohrid based on coupled hydrological and isotope mass balance modelling.

Source	**Flow rate (m^3/s)**		**$\delta^{18}O$ (‰)**	**δD (‰)**
Inputs				
Precipitation (P_L)	8.8	(20%)	−8.4	−52.9
Surface inflow (S_i)	7.2	(16%)	−10.1	−67.4
Groundwater inflow (G_i)				
− Prespa-fed (G_iP)	7.7	(17%)	−1.5	−20.5
− Surface springs (G_iS)	5.8	(13%)	−10.1	−67.4
− Sublacustrine springs (G_iX)	15.3	(34%)	−10.1	−67.4
Outputs				
Evaporation (E)	13.7	(31%)	−19.1	−112.7
Surface outflow (S_q)	14.8	(33%)	−3.5	−31.7
Groundwater outflow (G_q)	16.3	(36%)	−3.5	−31.7

Table 4 Estimate of past hydrological balance of Lake Ohrid during the early Holocene and last glacial.

Source	Flow rate (m^3/s)	$\delta^{18}O$ (‰)	δD (‰)
Early Holocene			
Inputs			
Precipitation (P_L)	11.1	−8.7	−59.6
Inflow (I_i)	35.1	−10.4	−68.8
Prespa inflow (G_iP)	7.7	−2.1	−24.1
Outputs			
Evaporation (E)	12.3	−20.9	−125.3
Outflow (Q_q)	41.6	−5.3	−41.3
Last Glacial			
Inputs			
Precipitation (P_L)	4.9	−9.2	−62.3
Inflow (I_i)	15.4	−16.0	−98.9
Prespa inflow (G_iP)	7.7	−5.8	−44.0
Outputs			
Evaporation (E)	6.3	−25.5	−150.2
Outflow (Q_q)	21.7	−8.1	−56.4

Table 5 Sensitivity of precipitation over the lake (P_L) to changing temperature during the early Holocene and last glacial. Evaporative flux varies with changing temperature, all other parameters remain constant.

Early Holocene

T_{air} (°C)	T_{lake} (°C)	E (m^3/s)	P_L (m^3/s)	Q_q (m^3/s)	*f*
8.7	11.0	9.8	9.2	36.3	0.35
10.7	13.0	12.3	11.1	41.6	0.36
12.7	15.0	15.1	13.1	47.2	0.36

Last glacial

T_{air} (°C)	T_{lake} (°C)	E (m^3/s)	P_L (m^3/s)	Q_q (m^3/s)	*f*
2.7	5.0	4.1	3.5	18.1	0.71
4.7	7.0	6.3	4.9	21.7	0.73
6.7	9.0	8.7	6.4	25.5	0.75