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HAL Authorization

1 **Modelling lake water and isotope mass balance variations under Mediterranean climate: Lake**
2 **Azigza in the Moroccan Middle Atlas**

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9 **Short title:** Modelling lake variations in the Moroccan Middle Atlas

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14 **Abstract**

15 As many Mediterranean headwater catchments, the Moroccan Middle Atlas plays an important role in
16 the highly vulnerable regional water resources. Mountain lakes are numerous in this region, and could
17 be regarded as possible sentinels of hydro-climatic changes, using appropriate modelling tools able to
18 simulate the lake-climate relation. We present a detailed study of Lake Azigza, based on a 4-year (2012-
19 2016) observation period, including lake level measurements, isotope analyses of precipitation, lake and
20 spring waters, and local meteorological data. The approach is based on a step-by-step calibration of a
21 lake water balance model, fed by precipitation and evaporation rates, to estimate the ungauged
22 components of the water balance. The net groundwater balance (inflow *minus* outflow) was quantified
23 during periods deprived of precipitation and further interpolated for the whole period. Diffuse surface
24 runoff was estimated through the adjustment of a rainfall coefficient. Finally, the groundwater fluxes
25 were partitioned between inflow and outflow using the isotope mass balance. The model was able to
26 simulate the continuous lake level decrease (4 m) observed over 2012-2016. Results show the
27 dominance of groundwater exchanges in the lake water balance, with significant interannual variations
28 related to annual precipitation, suggesting that a threshold effect probably limits the seepage when the
29 lake level decreases. This study underlines the importance of groundwater fluxes in the lake level
30 variations for Lake Azigza, probably representative of many similar lakes in the Middle Atlas.

1
2
3 31 **Keywords**
4
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6 32 Lake level, stable isotopes, lake-groundwater exchanges, Middle Atlas
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33 **Length of the manuscript:** 8759 word

34 **Introduction**

1
2
3 35 The Mediterranean region is considered by the latest IPCC report to be a “Hot-Spot” for global climate
4 36 change (IPCC 2013). In the last decades, temperature rise was above the global average and model
5
6 37 projections indicate a warming and drying trend in the Mediterranean basin (Lionello et al. 2014). In the
7
8 38 southern part of the Mediterranean, Morocco constitutes a key area for assessing the impact and
9
10 39 sensitivity of the water cycle to global climate change. Moroccan climate is influenced by air masses
11 40 with various origin (Atlantic Ocean, Mediterranean Sea and Sahara), in addition to the effect of
12
13 41 orography in the Atlas region, which leads to a marked spatial and inter-annual variability in
14
15 42 precipitation. Several studies have already emphasized that precipitation records of the last decades
16 43 showed a decrease in precipitation totals and wet days in Morocco (Driouech et al. 2010; Trambly et
17
18 44 al. 2013a) although heterogeneous behavior can be found at local scale (Khomsi et al. 2016). Future
19
20 45 projections based on high resolution regional climate models (RCM) have forecasted a decrease in
21 46 average precipitation following a north to south gradient for the end of the century (Trambly et al.
22
23 47 2013b; Filahi et al. 2017).

24
25 48 Quantifying the impact of these future precipitation changes on water resources requires long-term
26
27 49 hydrological data to validate model scenarios. An amplified response of runoff to precipitation decrease
28
29 50 is expected (Trambly et al. 2013b), but further studies are needed to explore the link between water
30
31 51 cycle and climate variations. Lakes have often been regarded as representative indicators of the effect
32
33 52 of climate variations on the water cycle at different times scales (Legesse et al. 2004; Vallet-Coulomb
34 53 et al. 2006; Troin et al. 2010; Lyons et al. 2011; Haghghi and Klove 2015; Troin et al. 2016), providing
35
36 54 an appropriate quantification of lake-groundwater exchanges, which have a great impact on lake
37
38 55 behavior (Rosenberry et al. 2015; Jones et al. 2016). Stable isotopes tracers, combined with hydrological
39 56 modelling approaches can be used to quantify the lake water balance (Krabbenhoft et al. 1990; Gibson
40
41 57 and Edwards 2002; Sacks et al. 2014; Bouchez et al. 2016; Arnoux et al. 2017).

42
43 58 In the Middle-Atlas Mountains, considered as the “Moroccan water tower” (Bentayeb and Leclerc
44
45 59 1977), the upper part of the Oum-Er-Rbia” (OER) basin is the major water resource for the downstream
46
47 60 irrigated dry plains and is essential to their economic development (Chehbouni et al. 2008). In this
48
49 61 region, several natural lakes have documented significant lake level variations in response to climate
50
51 62 (Sayad and Chakiri 2010; Sayad et al. 2011; Etebaai et al. 2012; Abba et al. 2012), although the relation
52
53 63 between lake behavior and climate change has never been quantitatively assessed. This paper aims at
54
55 64 implementing the hydrological modelling of the Azigza lake, in the Middle Atlas. Previous studies have
56
57 65 shown that Lake Azigza has recorded significant historical lake level changes (Gayral and Panouse
58
59 66 1954; Flower et al. 1989; Benkaddour et al. 2008; Vidal et al. 2016; Jouve et al. 2018). It was suggested
60
61 67 that water level fluctuations are linked to rainfall variations (Flower and Foster 1992) although no
62
63 68 quantification of this relation was established. We present the climatic, hydrologic and isotopic data

69 collected during a 4-year observation period (2012-2016). Then, in order to calibrate the lake model, the
70 ungauged components of the lake water balance are estimated using a step-by-step approach, the last
71 step relying on an isotope mass balance to quantify groundwater inflows and outflows. In the discussion
72 section, the influence of the lake stratification on the isotope mass balance modelling is first assessed.
73 Secondly, based on a 4-year lake water balance simulation, the role of groundwater processes in
74 modulating the lake response to contrasted climatic conditions is discussed, and finally the lake
75 sensitivity to persistent dry conditions is estimated.

76 **Hydrogeological and climatic settings**

77 Belonging to the Oum-Er-Rbia (OER) catchment, Lake Azigza (32° 58' N, 5° 26' W, 1540 m a.s.l.), is
78 a small natural mountain lake located in the Moroccan Middle Atlas (Fig. 1). The lake has a tectono-
79 karstic origin (Hinaje and Ait Brahim 2002) and is located in a relatively undisturbed and forested
80 region, dominated by Cedar (*Cedrus Atlantica*) and Oak (*Quercus*) woodland formed on calcareous red
81 soils (Flower and Foster 1992). The study area is situated in the Tabular Middle Atlas, generally made
82 up of Jurassic limestone and dolomite plateau (Lepoutre and Martin 1967). The Azigza lake catchment
83 belongs to the locally designated “High OER (HOER) basin”, delineated upstream to the Hansali Dam
84 (Fig. 1). The HOER basin is known for its important hydrological potential and includes the emblematic
85 Oum-Er-Rbia spring system, with an average discharge of 260 m³ s⁻¹. In the different nearby gauged
86 subcatchments, average specific discharges tend to increase with increasing drainage area, from ≈ 170
87 mm yr⁻¹ to ≈ 470 mm yr⁻¹ (Fig. 1; Table 1), evidencing the impact of groundwater circulation in this
88 karstic environment. For comparison, the whole OER catchment (48000 km²) displays a specific
89 discharge of only 25 mm yr⁻¹ (Q = 38 m³ s⁻¹) (Hammani et al. 2005). The climate is of Mediterranean
90 subhumid type, characterized by wet winters and dry summers. Mean annual rainfall approaches 900
91 mm yr⁻¹, most of which falls between October and April, and snow may persist for one or two months;
92 mean annual air temperature is about 12 °C (Martin 1981).

93 **Data and methods**

94 **Lake bathymetry and catchment morphology**

95 A precise lake bathymetry was established using an echo-sounder in April 2013 (Fig. 1). The deepest
96 point of the lake (42 m in April 2013) is located at the center of the southeast part of the lake,
97 characterized by steep slopes compared to the western part. In order to account for possible higher lake
98 levels, elevation data over the emerged part of the lake basin were also collected with a Real-Time
99 Kinematic GPS (RTK GPS). Data were incorporated into a 30 m resolution DEM
100 (earthexplorer.usgs.com) covering the remaining part of the lake watershed. A Geographic Information
101 System (GIS) software (ArcGIS 10.3) was used for DEM quality assessment, data conversion, geo-

referencing, profile extraction, interpretation, visualization and calculation of volume and area corresponding to different lake levels. The lake catchment has a surface of $10.2 \cdot 10^6 \text{ m}^2$, about twenty times the lake surface, with elevation up to 1794 m a.s.l.. It has no outlet and no permanent creek towards the lake. In April 2013, the lake level was 1546 m a.s.l. the lake area was $0.56 \cdot 10^6 \text{ m}^2$ and the volume was $7.9 \cdot 10^6 \text{ m}^3$.

Climate data

A meteorological station was settled in November 2014, located 3.7 m above the ground surface, about 30 m above the current lake level and 700 m away from the lake shoreline. Local climatic parameters were measured hourly from November 2014 to May 2016: precipitation (P), air temperature (T_a), relative humidity (rh), atmospheric pressure (P_a), solar radiation (R_s) and wind speed (U). In addition, to cover the remaining observation period, from October 2012 to September 2016, we used daily precipitation from the Tamchachate station, located about 20 km from the lake, at 1685 m a.s.l. (data from the OER Hydraulic Basin Agency), and reanalyzed data (P, T_a , rh, R_s , and incident longwave radiation, R_i) from the European Centre for Medium Range Weather Forecast (ECMWF) ERA-Interim database (Dee et al. 2011). The good correlations found between locally measured T_a , rh, and R_s and ERA Interim data over the common period ($r^2 = 0.97, 0.77$ and 0.89 respectively) were used to complete the local dataset. Based on daily precipitation, similar rainfall trends were found between the local station and the Tamchachate station, but the common period was too short to establish a robust statistic. We thus used the total rainfall ratio to complete the local dataset: $P_{Azizga} = 0.79 \times P_{Tamchachate}$.

Lake level and water temperature

In April 2013, a reference level was set, and the water level was manually measured approximately each month during the sampling period (April 2013-November 2014). The initial lake level (October 2012) was estimated from a photographic benchmark, compared to the reference level. This lake level time series was complemented by a CTD DIVER, anchored in the eastern part of the lake, that recorded water pressure, temperature and conductivity at an hourly time step from November 2014 to May 2016.

A strong declining trend was observed during the study period (4 m). Over this interannual trend, the lake level showed seasonal variations, with a slight increase in winter (October to April) and a stronger decrease from May to September (Fig. 2a). Based on daily data, a rapid response of lake level variations to precipitation events (≈ 1 day) was observed.

Three temperature profiles were established in January 2013, April 2014 and September 2015. The water column was homogeneous in January, with temperature ranging between 7.3 and 7.5 °C. During April and September, the lake was stratified with an epilimnion characterized by an average temperature of

134 17 °C and 21 °C, respectively, while the hypolimnion remained at an average temperature of 7 °C. These
135 layers were separated by a thermocline between 6 and 9 m below the lake surface. These 3 profiles are
136 consistent with previous data (Gayral and Panouse 1954, Benkaddour et al. 2008).

137 **Isotopic data**

138 A monthly sampling of precipitation, lake water and groundwater (well and springs) was performed
139 between October 2012 and November 2014. In addition, a more extensive field campaign was performed
140 in April 2013, including the collection of deep-water samples. Water samples were analyzed for their
141 oxygen ($\delta^{18}\text{O}$) and hydrogen ($\delta^2\text{H}$) isotopic compositions in the Stable Isotope Laboratory at CEREGE.
142 We used either laser spectrometry (Cavity-Ring-Down Laser Spectrometer, WS-CRDS, Picarro L1102-
143 i) or IRMS analysis instruments. Using IRMS, water samples were equilibrated with CO_2 (10h at 291K)
144 and H_2 (2h at 291K with a platinum catalyst) – for $\delta^{18}\text{O}$ and $\delta^2\text{H}$, respectively – in an automated HDO
145 Thermo-Finnigan equilibrating unit before measurement on a dual inlet Delta Plus mass spectrometer.
146 All the samples were replicated. Calibration of measurements was performed following the IAEA
147 reference sheet (IAEA 2009), using three liquid laboratory standards normalized beforehand against V-
148 SMOW, GISP2 and SLAP2 international standards. The values of the isotopic results are presented in
149 the standard notation δ permil (‰) and referenced to Vienna Standard Mean Ocean Water (V-SMOW).
150 The 1σ measurement precision is 0.05 ‰ for $\delta^{18}\text{O}$ and 1 ‰ for $\delta^2\text{H}$.

151 The weighted average precipitation composition was -7.15 ‰ and -42.6 ‰ for $\delta^{18}\text{O}$ and $\delta^2\text{H}$
152 respectively, and falls slightly above the Moroccan Meteoric Water Line (MMWL) established by Ait
153 Brahim et al. (2016) ($^2\text{H} = 7.7 \times \delta^{18}\text{O} + 9.2$, $r^2 = 0.93$, Fig. 3). Temporary springs encountered on the
154 southern shore displayed a very stable composition ($\delta^{18}\text{O} = -7.70 \pm 0.13$ ‰ and $\delta^2\text{H} = -46.1 \pm 0.8$ ‰),
155 while groundwater sampled monthly in the well presented a slight variation, from -7.56 ‰ to -6.33 ‰
156 for $\delta^{18}\text{O}$, and from -46.0 ‰ to -38.8 ‰ for $\delta^2\text{H}$. Following the Moroccan altitude gradient (Ait Brahim
157 et al. 2016), the average precipitation and springwater compositions would correspond to elevations of
158 1900 m a.s.l. and 2050 m a.s.l. respectively, slightly higher than the catchment maximum elevation.

159 The lake water isotopic compositions (δ_{L}) ranged from -7.56 ‰ to -2.30 ‰ and from -46.3 ‰ to -20.7
160 ‰ for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ respectively, and plot along a well-defined evaporation line ($\delta^2\text{H} = 4.8 \times \delta^{18}\text{O} - 9.9$;
161 $r^2 = 0.97$, $n = 80$), which crosses the composition of springwater and local precipitation (Fig. 3). The
162 average value calculated from the three sampling locations was -3.43 ‰ for $\delta^{18}\text{O}$ and -26.3 ‰ for $\delta^2\text{H}$
163 over the whole studied period.

164 Time evolution (Fig. 2b) shows smoothed seasonal variations at the L1 sampling location, with a
165 maximum during October and a minimum at the beginning of spring. For the other two sampling sites
166 located on the southern shore of the lake, depleted compositions were recorded during rainfall periods,

167 especially for L2 ($\delta^{18}\text{O} = -5.77 \text{‰}$ in February 2013, -7.56‰ in March 2013 and -4.82‰ in January
 168 2014). In the absence of visible surface runoff during the sampling, these depleted compositions are
 169 attributed to the influence of subsurface inflows. Based on the smoothed variations recorded at L1,
 170 seasonal minima were encountered in March-May 2013 ($\delta^{18}\text{O} = -3.93 \text{‰}$ and $\delta^2\text{H} = -28.0 \text{‰}$), and in
 171 January-April 2014 ($\delta^{18}\text{O} = -3.70 \text{‰}$ and $\delta^2\text{H} = -27.2 \text{‰}$), while the most enriched compositions were
 172 found in October. The maximum reached in October 2013 ($\delta^{18}\text{O} = -3.12 \text{‰}$ and $\delta^2\text{H} = -20.7 \text{‰}$), is lower
 173 than the maximum reached in October 2014 ($\delta^{18}\text{O} = -2.43 \text{‰}$ and $\delta^2\text{H} = -23.9 \text{‰}$).

174 The homogeneity of the lake water body was evaluated from a more extensive sampling performed in
 175 April 2013. Based on 8 surface water and 7 bottom water samples, well distributed, the water
 176 composition was found to be homogeneous, with an average value ($\delta^{18}\text{O} = -3.90 \pm 0.12 \text{‰}$ and $\delta^2\text{H} = -$
 177 $27.9 \pm 0.4 \text{‰}$) very similar to the L1 sampling location ($\delta^{18}\text{O} = -3.94 \text{‰}$ and $\delta^2\text{H} = -28.2 \text{‰}$). No bottom
 178 lake water samples are available at the end of the summer season.

179 **Lake water and isotope mass balance model**

180 The dynamic lake water balance equation, associated with quantified area–volume–depth relationships,
 181 is expressed as follows:

$$182 \quad \frac{dV}{dt} = S \times (P - E) + Ri + Gi - Go \quad (1)$$

183 where, for a daily time step (dt), dV is the lake volume variation (m^3), $S = f(V)$ is the lake surface (m^2)
 184 as a function of lake volume, P is the precipitation on the lake surface (m day^{-1}), E is the evaporation
 185 from the lake surface (m day^{-1}), and Gi and Go are the groundwater inflows and outflows respectively
 186 ($\text{m}^3 \text{day}^{-1}$) and Ri is the diffuse surface runoff ($\text{m}^3 \text{day}^{-1}$). The lake level simulation (h, in m a.s.l.) was
 187 then derived from the $h = f(V)$ relationship (Fig. 1).

188 The dynamic isotopic mass balance equation is:

$$189 \quad \frac{d(V \cdot \delta_L)}{dt} = S \times (P \cdot \delta_P - E \cdot \delta_E) + Ri \cdot \delta_P + Gi \cdot \delta_{Gi} - Go \cdot \delta_L \quad (2)$$

190 where δ is either the $\delta^{18}\text{O}$ or $\delta^2\text{H}$ isotopic composition of each of the water balance components: δ_L for
 191 the lake water, which also accounts for the composition of Go, assuming a homogeneous lake water
 192 body, δ_P for precipitation water, which also accounts for the composition of Ri, δ_{Gi} for the composition
 193 of Gi, taken as the springwater composition. The composition of the water vapor evaporated from the
 194 lake, δ_E , was estimated from the Craig and Gordon equation (Craig and Gordon 1965):

$$195 \quad \delta_E = \frac{(\delta_L - \epsilon^*) / \alpha - (rh \cdot \delta_A) - \epsilon_k}{1 - rh + \epsilon_k} \quad (3)$$

196 $\varepsilon_k = (1 - rh) \cdot \theta \cdot n \cdot CD$ (4)

197 where α is the equilibrium fractionation factor calculated at T_w , the surface water temperature (Horita
 198 and Wesolowski 1994), ε^* is the equilibrium isotopic separation, related to the fractionation factor by
 199 $\varepsilon^* = (\alpha - 1)$, ε_K is the kinetic separation, CD a kinetic constant established experimentally ($28.5 \cdot 10^{-3}$ and
 200 $25.1 \cdot 10^{-3}$ for ^{18}O and 2H , Merlivat 1978), $n = 0.5$ for open water bodies, θ is a transport resistance
 201 parameter (Gonfiantini 1986; Gat 1996), rh is the relative humidity normalized to T_w , and δ_A the isotopic
 202 composition of the ambient moisture. δ_A was estimated from the assumption of an isotopic equilibrium
 203 between precipitation and atmospheric vapor ($\delta_A = \delta_P - \varepsilon^*$). All terms are expressed in the decimal
 204 notation. The value of θ is generally lower than 1 for a water body whose strong evaporation flux
 205 influences the atmospheric boundary layer (Horita et al. 2008). A value of 0.5 has been determined for
 206 the eastern Mediterranean Sea (Gat et al. 1996), and for Lake Chad (Bouchez et al. 2016). Here, the θ
 207 value was chosen to match the observed slope of the evaporation line (leading to $\theta = 0.5$).

208 To calculate an annual lake water balance, a steady state is often assumed. The water and isotope mass
 209 balance equations (Eq. 1 and 2) are then simplified to express the evaporation-to-input ratio (E_V/I),
 210 which represents the degree of closure of the lake:

211
$$E_V/I = \frac{\delta_I - \delta_L}{\delta_E - \delta_L}$$
 (5)

212 where I is the total annual water inputs ($I = P_v + R_i + G_i$), E_v and P_v the volumetric terms of evaporation
 213 and precipitation, accounting for lake surface variations at the daily time step, δ_i is the isotopic
 214 composition of inflows, and δ_L is the average annual lake water isotopic composition. When dealing
 215 with strongly seasonal systems, all the annual averages have to be weighted by the corresponding fluxes
 216 (Gibson and Edwards 2002). For imbalanced annual lake water budgets ($\Delta V/\Delta t \neq 0$), Eq. (5) becomes:

217
$$I = \frac{\Delta(V \cdot \delta_L)/\Delta t + E_V(\delta_E - \delta_L) - \delta_L \Delta V/\Delta t}{\delta_I - \delta_L}$$
 (6)

218 When available, the estimate of $\Delta(V \cdot \delta_L) / \Delta t$ can be based on seasonal variations of V and δ_L , at a given
 219 (i) time step, as follows:

220
$$\frac{\Delta(V \cdot \delta_L)}{\Delta t} = \Sigma(V_{(i)} \cdot \delta_{L(i)} - V_{(i-1)} \cdot \delta_{L(i-1)})$$
 (7)

221 All calculations were coded in the PYTHON language.

222 **Evaporation estimate**

223 The Penman combination method (Penman 1948) can be applied with synoptic climate data to estimate
 224 open-water evaporation (Brutsaert 1982; Jensen et al. 1990; Shuttleworth 1992). It combines the energy
 225 balance with an aerodynamic formulation as follows:

$$\lambda_E = (R_n - \Delta S) \frac{D}{D + \gamma} + E_a \frac{\lambda \gamma}{D + \gamma} \quad (8)$$

$$\text{with } E_a = 0.26 (1 + 0.54 U_2) (e_{sw} - e_a) \quad (9)$$

228 where R_n is the net input of energy at the lake surface (W m^{-2}), ΔS is the change of energy storage in the
 229 water body (W m^{-2}), D is the slope of the saturation vapor pressure curve (Pa K^{-1}) at T_a , γ the
 230 psychrometric constant (Pa K^{-1}), U_2 is the wind speed at 2 m above the ground surface (m s^{-1}), e_a the
 231 actual vapor pressure of the atmosphere (Pa), e_{sw} the saturated vapor pressure (Pa) at T_w . The first term
 232 in the Penman approach relies on available energy, and for lake evaporation estimates, involves
 233 accounting for the difference between R_n and ΔS . This has no impact on the total annual rate but it
 234 introduces a shift in the seasonal evaporation variation (Giadrossich et al. 2015). The net radiation R_n is
 235 the sum of the short wave and long wave energy balances:

$$R_n = R_s (1 - a) + R_{nli} (1 - a') - R_{nle} \quad (10)$$

237 where R_s is the incoming (shortwave) solar radiation (W m^{-2}), R_{nli} and R_{nle} the longwave incoming and
 238 emitted radiation respectively (W m^{-2}), a is the albedo or shortwave reflectance, and a' the longwave
 239 reflectivity. According to the work of Cogley (1979) that computes the albedo as a function of the
 240 latitude, a value of 0.08 was taken for a , while a' is assumed to have a constant value equal to 0.03
 241 (Parker et al. 1970). The emitted longwave radiation (R_{nle} in W m^{-2}) is based on the Stefan-Boltzmann
 242 law:

$$R_{nle} = \epsilon_w \times \sigma \times T_w^4 \quad (11)$$

244 where ϵ_w is the emissivity of water, Anderson (1954) gives this value as 0.97, and σ is the Stefan-
 245 Boltzman constant, $5.6697 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$. The change in the energy storage term over a given time
 246 step Δt is:

$$\Delta S = Z \times C_w \times \rho_w \times \frac{\Delta T_w}{\Delta t} \quad (12)$$

248 where C_w is the specific heat capacity of water ($\text{J kg}^{-1} \text{ K}^{-1}$), ρ_w the water density (kg m^{-3}), and Z the layer
 249 thickness (m) affected by the temperature variation.

250 **Results**

251 **Evaporation rate**

252 Daily lake evaporation was computed between 17 November 2014 and 15 July 2016, using the locally
253 measured climatic parameters (Fig. 4a), completed by the incident longwave radiation, R_{nl} obtained
254 from the ECMWF. Considering the thermal stratification and the constant temperature in the lower layer,
255 ΔS was calculated for the upper layer ($Z = 8$ m), and smoothed using a 30-day moving average to avoid
256 the instabilities due to the daily time step calculation. The period of energy accumulation ($\Delta S > 0$) spans
257 between February and July, with a maximum ($\approx 115 \text{ W m}^{-2}$) at the beginning of May, while the stored
258 energy is released between August and January, with a minimum ($\approx -90 \text{ W m}^{-2}$) at the beginning of
259 November (Fig. 4b). This seasonal behavior is comparable in magnitude to that reported for lakes
260 located under a similar climatic regime: Vegoritis Lake in Greece (510 m a.s.l. and mean depth of 20 m;
261 Gianniou and Antonopoulos 2007), and Baratz Lake in Italy (27 m a.s.l. and mean depth of 5 m;
262 Giadrossich et al. 2015). The computed annual evaporation rate (Fig. 4b) gave an annual value of 1217
263 mm over the complete 2015 annual cycle.

264 **Annual water balance framework over the 2012-2016 period**

265 The lake volume and surface were computed daily using the bathymetric relations, in order to calculate
266 the water balance of the four annual cycles monitored (October-September periods), and to quantify the
267 sum of ungauged fluxes: $R_i + G_i - G_o$ (Eq. 1; Table 2). The evaporation rate was extrapolated over the
268 whole studied period based on the daily ERA Interim climate data, with the correlation found with local
269 data, an average wind speed value, and using an annual ΔS time series approximated by a sinusoidal
270 function, with a 70 W m^{-2} amplitude (Fig. 4b). Precipitation data were based on local measurements for
271 the November 2014 – May 2016 period, and on the Tamchachate station for the rest of the period. ΔV
272 was slightly positive during the first annual cycle, and then remained strongly negative (Table 2). Over
273 the 4-year period studied, the lake level progressively decreased from 1545.7 m a.s.l. in October 2012
274 to 1542 m a.s.l. in October 2016, which corresponds to a 23 % loss of lake volume.

275 **Calibration of net groundwater flow and diffuse surface runoff using daily lake level**

276 The net groundwater flow ($NG = G_i - G_o$), and R_i were calibrated using the daily lake level record
277 (2014-2016). Considering the rapid reaction of the lake to rainfall (Fig. 2b), and the small size of the
278 catchment, we assume that during dry periods, the lake level variations are only driven by evaporation
279 and groundwater exchanges. Assuming $P = R_i = 0$, it comes:

$$280 \quad NG = \frac{dV}{dt} - S(V) \times E \quad (13)$$

281 The following criteria were used to select appropriate dry periods: a beginning two days after the last
282 rainfall event, a total duration of at least 4 days with a total precipitation lower than 0.2 mm, and a robust
283 linear trend ($r^2 > 0.8$ for the linear regression applied to $\Delta V/\Delta t$). The latter criterion enables measurement
284 artefacts, or lake level instabilities for which the interpretation is unclear, to be ruled out. A total of 10
285 periods were isolated, for which an average NG was determined. The resulting time series, referred to
286 as $NG_{ref(1)}$, showed a mainly negative balance, with a minimum in November 2014, and a maximum
287 slightly positive in April 2015 (Fig. 5b).

288 The subsequent step was to run the lake level simulation, with a linearly interpolated daily NG time
289 series, and to introduce a runoff coefficient (k):

$$Ri = k \times A \times P \quad (14)$$

290 with A the lake catchment area. An average runoff coefficient ($k = 0.06$) was obtained by adjusting
291 simulated and measured lake level. A water balance closure criteria was used: starting with the measured
292 lake level, the simulation fitted the level measured at the end of the calibration period. Given the
293 respective sizes of the catchment (10.2 km^2) and the lake surface (0.5 km^2 on average), the resulting Ri
294 is of the same order of magnitude as the direct precipitation on the lake surface (Table 2).

296 **Extrapolation of NG over 2012-2016**

297 To estimate NG over the whole studied period, Eq. (1) was applied for each interval between two manual
298 lake level data, using the previously calibrated runoff coefficient (k). This led to a complementary time
299 series referred to as $NG_{ref(2)}$, which, combined with $NG_{ref(1)}$, covered the whole 2012-2016 period.
300 Significant seasonal variations were evidenced (Fig. 5b). The maximum was reached between April 19th
301 and June 1st 2013 ($5200 \text{ m}^3 \text{ day}^{-1}$), and the minimum occurred in November 2014 ($-7000 \text{ m}^3 \text{ day}^{-1}$, 4-
302 day average), and in November 2013 ($-5600 \text{ m}^3 \text{ day}^{-1}$, 31-day average).

303 Then, in order to produce a monthly time series, which is more convenient for rainfall-discharge
304 analysis, a monthly adjustment of NG was performed, based on the lake level simulation, and respecting
305 the quantitative framework given by the combination of $NG_{ref(1)}$ and $NG_{ref(2)}$. The resulting NG time
306 series (Fig. 5b), can be considered as the best evaluation, given the available data. Its seasonal behavior
307 is smoothed and delayed compared to monthly rainfall, with seasonal maxima during spring, that only
308 became positive in 2013 and 2015. Interestingly, NG was almost null between May and September 2016.

309 At the annual time step, NG values were always negative (Table 2), even when the lake water balance
310 was slightly positive (cycle 1), evidencing the dominance of groundwater seepage. The maximum
311 annual NG was found during the first and the fourth annual cycles ($-0.45 \cdot 10^6 \text{ m}^3 \text{ yr}^{-1}$), and the minimum

312 values during the second and the third annual cycles (-0.62 and $-0.60 \cdot 10^6 \text{ m}^3 \text{ yr}^{-1}$, respectively). This
313 trend does not follow the annual variations in precipitation, since the maximum and minimum
314 precipitation occurred during the first and the fourth cycles respectively.

315 **Groundwater flows partitioning and lake water residence time**

316 The partitioning of NG between G_i and G_o is based on the isotope mass balance applied at the annual
317 time step for the two first annual cycles, using daily data (Eq. 6 and 7). Results show the importance of
318 groundwater fluxes, which dominated all other components of the lake water balance (Table 2). G_i was
319 higher than $P+R_i$, with a strong difference between these two years. Considering the lake catchment
320 area, the G_i values correspond to 261 and 71 mm yr^{-1} , respectively. The G_o values were also very
321 different between the two years, and correspond to averages of 8500 and $3700 \text{ m}^3 \text{ day}^{-1}$, respectively,
322 which is of the same magnitude as the minimum NG reached in November 2013 and 2014 (5600 and
323 $7000 \text{ m}^3 \text{ day}^{-1}$). These pronounced differences between the two years can be attributed to the decrease
324 in annual precipitation, but the decrease in G_i (-73%) was stronger than for G_o (-56%), and for P (-48
325 $\%$). For the 2012-2013 hydrological cycle, which is almost at equilibrium, an average residence time of
326 2.1 year can be estimated.

327 **Discussion**

328 **Impact of the lake stratification on the isotope mass balance**

329 The isotopic homogeneity of the lake water body was assessed in April 2013, but not during fall.
330 Nevertheless, our results suggest that the thermal stratification of the lake water column lead to a
331 progressive isotopic stratification during summer, with a more enriched isotopic composition in the
332 surface layer due to the effect of evaporation. This explains the inability of the model to simulate the
333 observed lake isotopic seasonality, with smaller seasonal amplitudes and delayed maxima compared to
334 the measured values (Fig. 5c, simulation #1), while a lower initial isotopic composition would lead to a
335 better simulation of observed isotopic data (Fig. 5c, simulation #2). To illustrate the effect of lake
336 stratification, simple mass balance calculation allowed us to evaluate the magnitude of the observed
337 seasonal enrichment ($+0.85\%$ between June 1st and September 7th 2013, and $+1.26 \%$ between April
338 14th and September 7th 2014). Assuming the whole water body is affected, and neglecting groundwater
339 fluxes, the evaporation rates able to explain observed isotopic variations were much higher than the
340 actual lake evaporation ($+48 \%$ and $+67 \%$, for 2013 and 2014, respectively). Integrating the water
341 column stratification into the simulation would require much more detailed isotopic data, to characterize
342 the seasonal lake isotopic behavior of the water column, and the depth of sub-aquatic sources (G_i) and
343 sinks (G_o), which are probably diverse and localized in such a complex karstic environment.

344 The impact of the lake stratification on the annual isotope mass balance needs to be assessed, since it
345 leads to overestimating the annual average δ_L compared to that of a homogeneous water body.
346 Considering that the thermocline (≈ 8 m depth) divides the lake water body into two approximately
347 equivalent volumes, the amplitude of the seasonal isotopic enrichment is twice that of the conceptual
348 “equivalent homogeneous lake volume”. In addition, the discrepancy between surface and bottom
349 compositions due to the lake stratification is limited to half a year. These simple considerations allowed
350 us to simulate the theoretical impact of the lake stratification on the seasonal behavior of δ_L . The
351 overestimation of annual average δ_L based on surface measurements, compared to the corresponding δ_L
352 of an “equivalent homogeneous lake volume”, represents approximately only 15 % of the amplitude of
353 observed seasonal enrichment, i.e. 0.13 ‰ and 0.19‰ for $\delta^{18}\text{O}$ during cycle #1 and #2, respectively.
354 This sensitivity analysis showed that despite the amplification of the seasonal lake water isotopic
355 enrichment, the water column stratification only slightly affects the annual mass balance by
356 underestimating groundwater inflows by less than 10 %.

357 **Impact of non-steady state on the isotope mass balance**

358 The steady state isotopic mass balance, expressed as the E_v/I ratio (Eq. 5), has been widely used for
359 quantifying an annual lake water balance (e.g. Gibson and Edwards 2002; Yi et al. 2008; Gibson and
360 Reid 2014; Gibson et al. 2017; Cui et al. 2018). This approach, designed for steady state situations, may
361 lead to wrong estimates in the case of interannual trends. We took the opportunity of our seasonal
362 isotopic sampling to evaluate the error associated with the application of the steady state isotope mass
363 balance to non-steady state situation. During the year 2013-2014 (cycle #2), the lake volume decreased
364 by around 10 %. The steady state application of the isotope mass balance (Eq. 5) gives $E_v / I = 0.29$
365 (using both $\delta^{18}\text{O}$ and $\delta^2\text{H}$), compared to the reference ratio: $E_v / I = 0.52$ and 0.51 (using $\delta^{18}\text{O}$ and $\delta^2\text{H}$
366 mass balances, respectively), and would lead to overestimating G_i by 140 %. This comparison points to
367 the importance of a seasonal survey for lake water balance studies.

368 **Magnitude and variations of groundwater exchanges**

369 The shape of monthly NG variations seems to roughly respond to precipitation seasonality, with a time
370 lag of several months after the rainy winter (Fig. 5b). However, annual NG does not follow the
371 precipitation trend, with the driest year (cycle #4) characterized by an annual NG similar to that of the
372 wettest year (cycle #1). Therefore, understanding the relation between rainfall and the groundwater
373 control on the lake water balance requires the partitioning of NG between inflow and outflow.

374 A major result of the isotopic groundwater partitioning is that groundwater components, both G_i and
375 G_o , largely dominate all the other water balance components, even though their difference remains
376 relatively low. Lake Azigza water balance is thus mainly driven by subsurface circulation. In addition,

377 groundwater fluxes are very different between cycles #1 and #2, for both G_i and G_o , while annual
378 precipitations are also strongly contrasted. These G_o variations contradicts the previous assumption of
379 a constant seepage rate (Flower and Foster 1992), and suggests that both groundwater fluxes are linked
380 to the amount of precipitation.

381 Another important result of the isotope water balance is that the magnitude of G_i is consistent with a
382 recharge area equivalent to the lake catchment area, leading to annual values (261 mm and 71 mm) in
383 line with the magnitude of average runoff measured for the smallest gauged catchment in the region
384 (Tamchachate station, 166 mm yr⁻¹, Table 1), and with the corresponding relation between annual
385 precipitation and runoff (Fig. 6). The apparent similarity between the $P - G_i$ relation and the rainfall-
386 runoff relation of the neighboring Tamchachate watershed is consistent with the karstic environment of
387 these upstream catchments, where rivers are typically fed by rapid groundwater circulation. Based on
388 this similarity, the Tamchachate empirical rainfall-runoff relation was used to assess the magnitude of
389 G_i for the two remaining annual cycles. Corresponding G_o values were then deduced from annual NG
390 (Table 2) and four estimates of the Azigza annual lake water balance are proposed.

391 The strongly contrasted climatic situations covered by our studied period provide interesting information
392 on the lake water balance response to precipitation variations. At the annual time scale, a high range of
393 variation was obtained for G_i (1–12 range), while G_o also varied along with annual precipitation, but
394 with a lower magnitude (1–5 range). Contrasting with these strong variations, the difference between
395 the two fluxes remained relatively stable (1–1.4 range for NG). The groundwater partitioning suggests
396 that the higher annual NG during 2015-2016 was due to a very low G_o and G_i . This is consistent with
397 monthly NG, which became almost null between May and September 2016, suggesting a disconnection
398 of the lake from groundwater circulation. Therefore, except for this particular situation, the concomitant
399 variations of annual G_i and G_o together with annual rainfall, indicate that they are probably driven by
400 the same hydraulic gradient.

401 **Lake sensitivity to persistent dry climate conditions**

402 The strong and continuous decreasing trend observed during the studied period raises the question of
403 the future evolution of the lake. Climatic projections suggest an intensification of dry conditions in the
404 Mediterranean area (Tramblay et al. 2013b), and a decrease in rainfall would greatly impact hydrological
405 fluxes. In order to evaluate such impacts, the lake model was used to perform sensitivity analysis.

406 A first test was performed to simulate the lake level behavior, using the 2015-2016 period as
407 representative of dry conditions: the model was run with the hydroclimatic data of cycle #4, by iterating
408 continuously the same hydrological year. Results led to a complete lake drying after 12 years.

409 Nevertheless, by maintaining a constant G_o , this sensitivity test represents an extreme situation. The
410 lake behavior greatly depends on whether groundwater seepage is permanent or not, and the last part of
411 our study period suggest a disconnection of the lake from groundwater flows. In this karstic
412 environment, groundwater circulation occurs through preferential flowpaths, and groundwater seepage
413 could possibly decrease, or even stop, if the lake level fell below the active water routing structures. A
414 progressive decreasing G_o , until the disconnection observed after May 2016, could explain why the
415 seasonal minimum in NG remained relatively high during the end of 2015, compared to the situations
416 of November 2012, 2013 and 2014 (Fig. 5b). This scenario was tested by running the lake model with
417 $NG = 0$, while keeping the climatic conditions of cycle #4. A lake level decrease of 5.5 m during the
418 same 12-year period was obtained.

419 Finally, additional sensitivity analysis was performed to evaluate the impact of an increase in
420 evaporation. Its influence remains almost negligible compared to the influence of groundwater flow
421 variations (1-meter difference in 12 years with a 10 % increase of evaporation).

422 **Conclusion**

423 The Azigza lake water balance variations were determined over a 4-year period, which covered
424 contrasted climatic situations and documented strong hydrological differences. A major result is that
425 Lake Azigza water balance is mainly driven by subsurface circulation, with significant interannual
426 variations in groundwater fluxes.

427 The quantification of groundwater inflow and outflow was achieved by the detailed seasonal water
428 sampling, which allowed us to determine an isotopic mass balance. It was not possible to decipher
429 seasonal variations of groundwater flows because of the influence of the summer thermal stratification,
430 which prevents mixing of the water body. Nevertheless, we showed that the isotopic mass balance is a
431 robust way to partition the groundwater contribution between inflow and outflow, at the annual
432 timescale, given that a sub-annual sampling is available. The short water residence time, estimated at 2
433 years for 2012-2013, explains the rapid lake level response to climate variations.

434 The magnitude of groundwater inflow indicates a recharge area consistent with the topographic
435 catchment area, with interannual variations comparable to runoff variations measured in a nearby
436 catchment. Strong variations of groundwater outflow are also evidenced, which contradicts the previous
437 assumption of a constant seepage (Flower and Foster 1992). Moreover, our results suggest that
438 groundwater seepage may be strongly reduced, or even stopped, when the lake level decreases. Monthly
439 NG and groundwater partitioning point to closure of the lake after May 2016, corresponding to threshold
440 level of 1542.5 m a.s.l.. Such a decrease of G_o associated to the lake level drop could give the lake a
441 resilience capacity in a context of persistent dry climatic conditions.

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630 **Figure captions**

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3 631 **Fig. 1** a) Location of the Oum Er Rbia (OER) basin in Morocco; b) High OER basin delineated at
4 632 Hansali dam; c) Azigza Lake bathymetry (modified from Vidal et al. 2016) with the location of water
5 633 sampling (L1, L2, L3 for lake water, W for well water and S for springs) and meteorological station (red
6 634 star); d) Relation between lake water level, area, and volume (observed lake level range in grey)

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10 635 **Fig. 2** a) Lake level and precipitation time series between October 2012 and October 2016, with
11 636 manual measurements (black crosses) and continuous lake level record (red line). Daily precipitation is
12 637 either measured locally (blue) or derived from Tamchachate station (purple; see text); b) Lake water
13 638 isotopic composition ($\delta^{18}\text{O}$) at the three sampling locations

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18 639 **Fig. 3** $\delta^{18}\text{O}$ - $\delta^2\text{H}$ cross plots of well, springs, and lake data, with the corresponding evaporation line
19 640 (dash-dotted line), weighted average precipitation (red) and the Moroccan Meteoric Water Line (black
20 641 line, Ait Brahim et al. (2016))

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24 642 **Fig. 4** a) Meteorological variables measured at the Azigza station; b) Daily evaporation (blue), with
25 643 corresponding ΔS (red) and its sinusoidal approximation (dotted black line)

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29 644 **Fig. 5** a) Simulated (green) and measured (black crosses and red line) lake level, with daily
30 645 precipitation (same as Fig. 2) and dry periods used for $\text{NG}_{\text{ref}(1)}$ calibration (grey shadow); b) $\text{NG}_{\text{ref}(1)}$
31 646 (black), $\text{NG}_{\text{ref}(2)}$ (purple) and simulated NG time series (blue-red histogram) and monthly precipitation;
32 647 c) Simulation of lake water $\delta^{18}\text{O}$, with monthly values of G_i calculated from the adjusted NG ($G_i =$
33 648 $\text{NG} + G_o$) using two initial isotopic composition (#1 green line and #2 blue line)

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39 649 **Fig. 6** Relation between annual rainfall and runoff at the Tamchachate catchment, compared to Azigza
40 650 P and G_i values for 2012-2013 and 2013-2014

Fig. 1

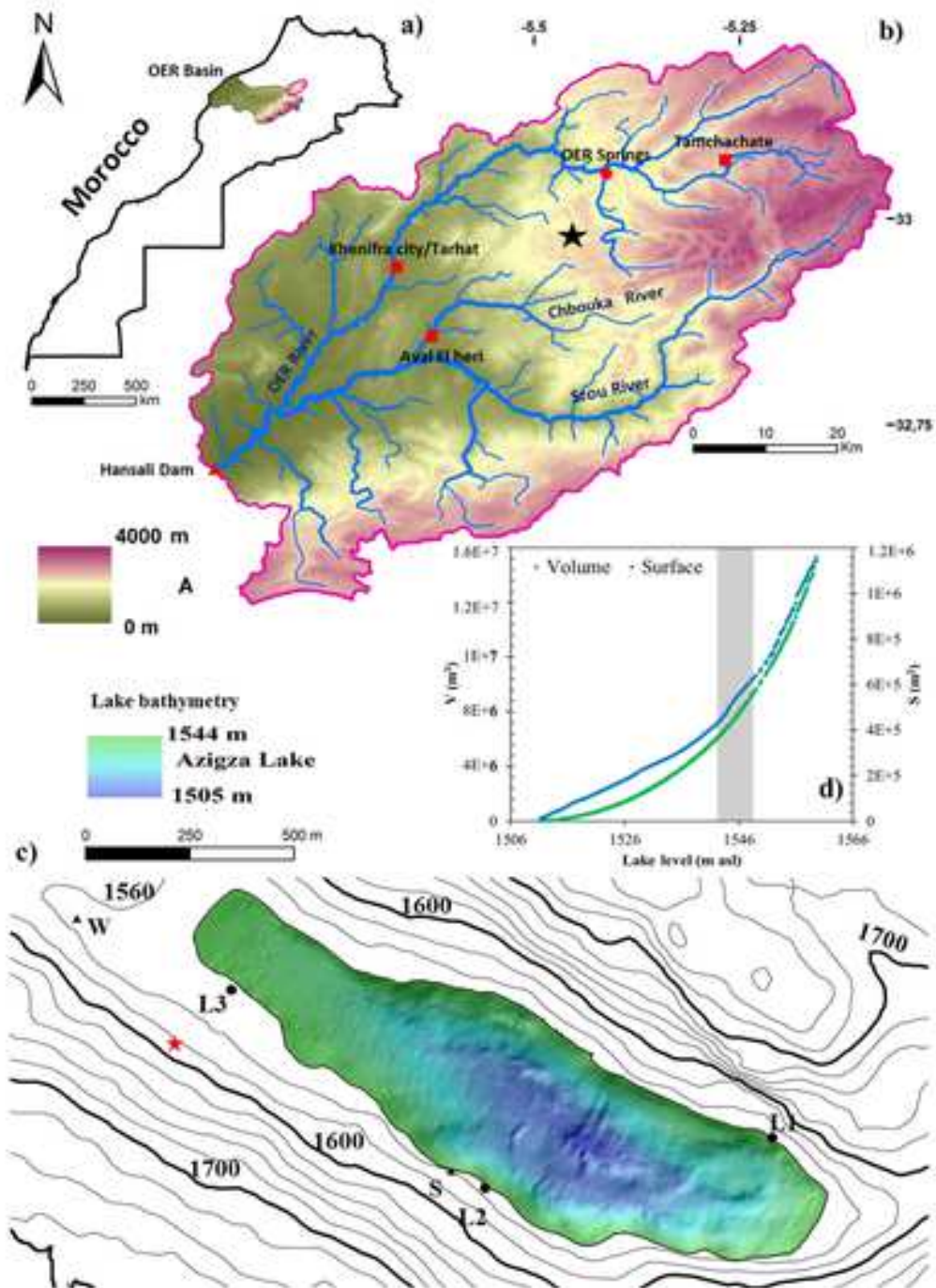


Fig. 2

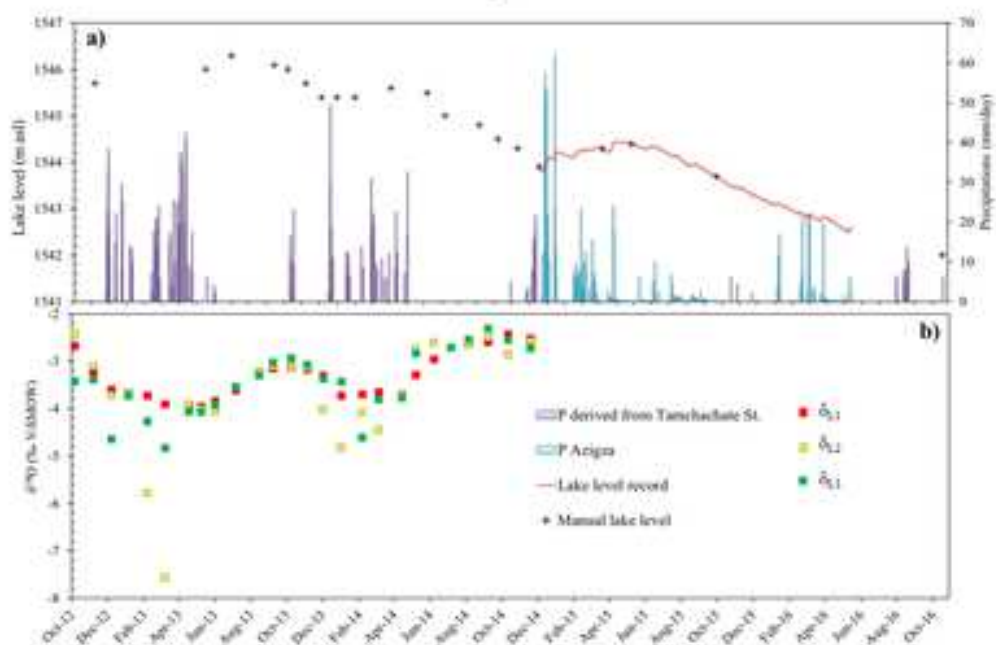


Fig. 3

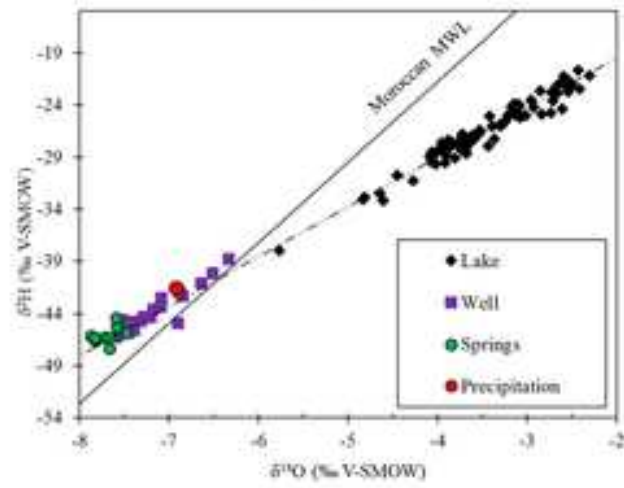


Fig. 4

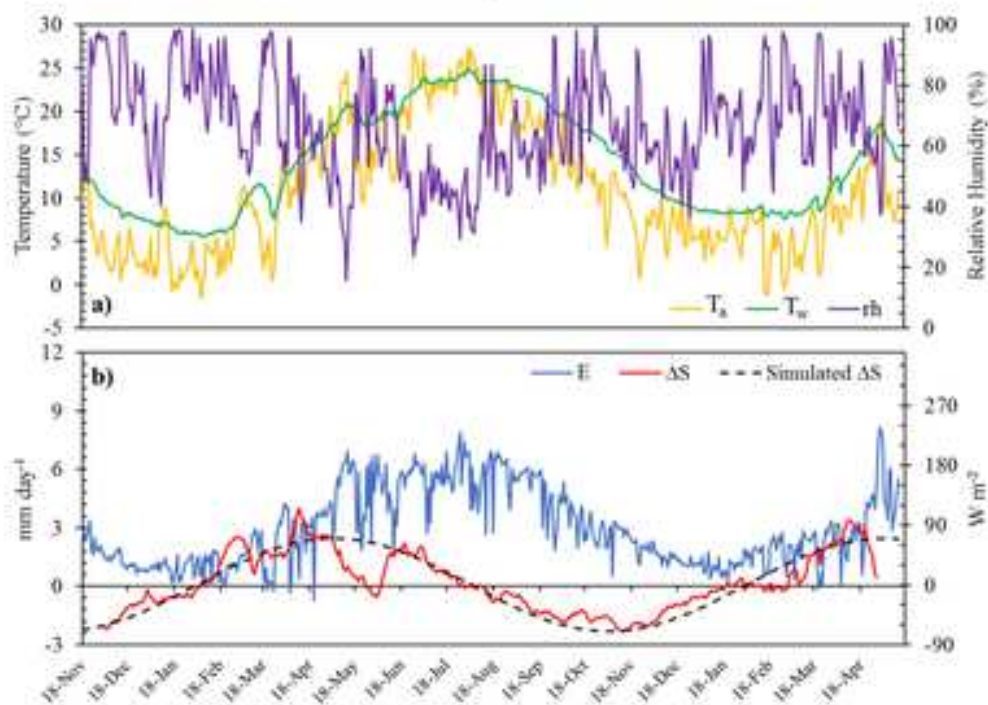


Fig. 5

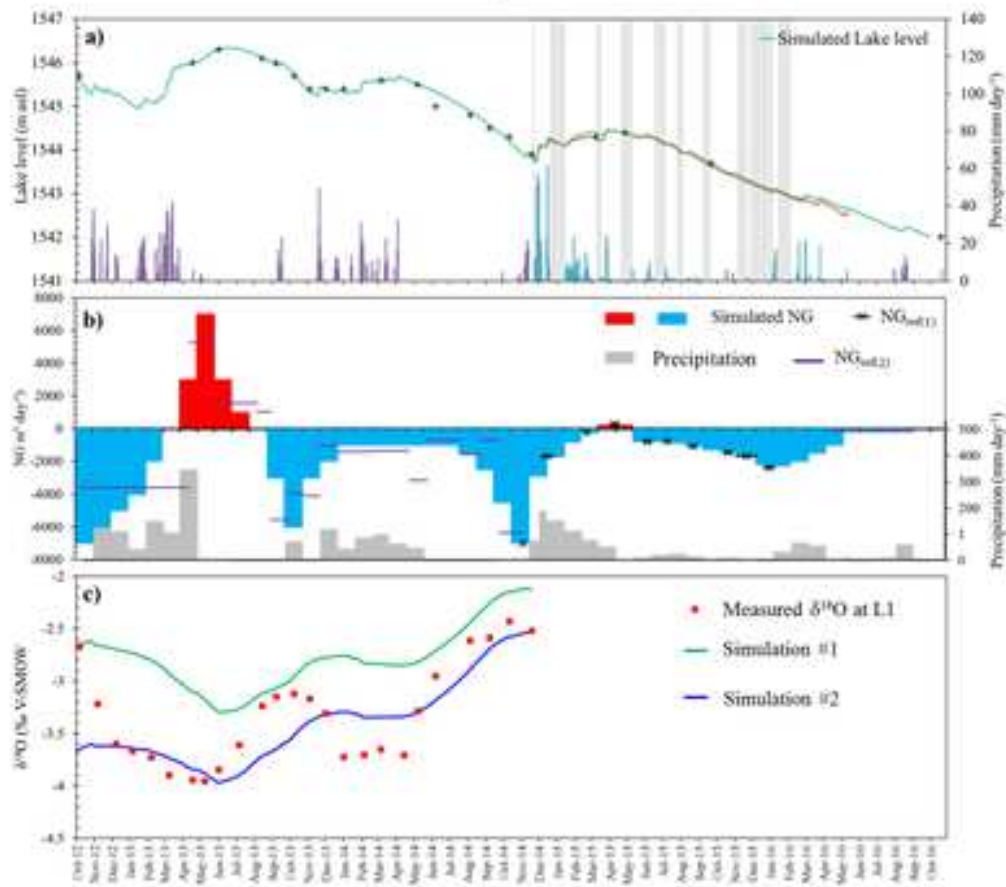
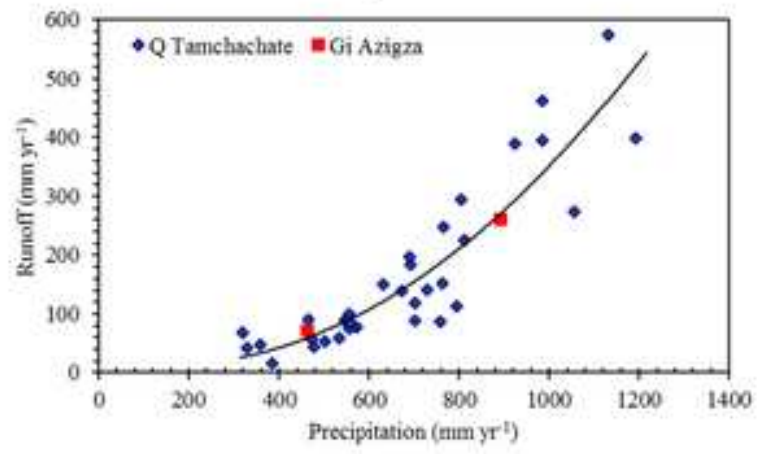


Fig. 6



Station	altitude (m a.s.l.)	drainage area (km²)	specific discharge (mm yr⁻¹)	Q mean (m³ s⁻¹)	recording period
Tamchachate	1685	132	166	0.7	1975-2011
			(173)	(0.72)	(1975-1990)
Aval El heri	830	308	245	2.4	1975-2011
			(254)	(2.48)	(1975-1990)
Tarhat	873	1113	468	16.5	1975-1990

Table 1 Hydrological characteristics of three sub-basins belonging to the HOER basin

	P	E	ΔV	$R_i + G_i - G_o$	R_i	NG	G_i	G_o	E_v/I
Cycle 1 2012-2013	$0.54 \cdot 10^6$ (892 mm)	$0.66 \cdot 10^6$	$+0.04 \cdot 10^6$	$0.16 \cdot 10^6$	$0.60 \cdot 10^6$	$-0.45 \cdot 10^6$	$2.65 \cdot 10^6$ (261 mm)	$3.10 \cdot 10^6$	0.17
Cycle 2 2013-2014	$0.26 \cdot 10^6$ (462 mm)	$0.65 \cdot 10^6$	$-0.73 \cdot 10^6$	$-0.33 \cdot 10^6$	$0.29 \cdot 10^6$	$-0.62 \cdot 10^6$	$0.71 \cdot 10^6$ (71 mm)	$1.34 \cdot 10^6$	0.52
Cycle 3 2014-2015	$0.37 \cdot 10^6$ (781 mm)	$0.59 \cdot 10^6$	$-0.36 \cdot 10^6$	$-0.14 \cdot 10^6$	$0.46 \cdot 10^6$	$-0.60 \cdot 10^6$	$2.11 \cdot 10^6$ (199 mm)	$2.71 \cdot 10^6$	
Cycle 4 2015-2016	$0.12 \cdot 10^6$ (272 mm)	$0.54 \cdot 10^6$	$-0.71 \cdot 10^6$	$-0.29 \cdot 10^6$	$0.17 \cdot 10^6$	$-0.45 \cdot 10^6$	$0.21 \cdot 10^6$ (20 mm)	$0.66 \cdot 10^6$	

Table 2 Annual water balance components ($m^3 \text{ yr}^{-1}$) resulting from the step-by-step calibration procedure (see text for details)