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Estimating evaporation in semi-arid areas facing data scarcity: Example of the El Haouareb dam (Merguellil catchment, Central Tunisia)

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ABSTRACT

Study region: The El Haouareb dam (Merguellil catchment) in central Tunisia, which is typical of semi-arid environments.
Study focus: Most estimates of evaporation from water bodies located in semi-arid environments suffer from the lack of data, or biased field measurements. It is thus important for hydrologists to assess the relative performance of the various available methods used to estimate this water loss, as well as their uncertainties. We confronted physical approaches based on contrasted theoretical formulae (Dalton, simplified BREB, Penman) and geochemical approaches based on mass conservation (stable isotopes and chloride). We compared the results with Colorado pan measurements, and tested the methods' sensitivity to various physical parameters and data gaps.
New hydrological insights for the region: In this region, where mean annual rainfall is 300 mm, estimates of evaporation of the El Haouareb Dam lake ranged from 1400 to 1900 mm a^{-1}, depending on the method and the year. The Penman approach was found to be the most robust and gave an annual mean of 1600 mm a^{-1}. Evaporation values were refined by combining results from the different methods. Mean interannual evaporation was estimated to be 1700 mm a^{-1}, with an uncertainty of 15%. From this work, we propose an annual Colorado pan conversion coefficient of 0.8 which can be adjusted, 1.0 for spring and 0.76 for the rest of the year.

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1. Introduction

Semi-arid environments are well known for the strong spatial and temporal variability of climatic conditions and hydrological processes, as well as for rapid changes caused by human activities (e.g. Cudennec et al., 2007; Leduc et al., 2007). This particular sensitivity and the increased environmental fragility are of global concern because semi-arid and arid areas cover one third of the continents and are home to more than 20% of the world population. In these regions, evaporation implies a complete loss of water resources at the basin scale. For both scientific and social reasons, a reliable estimate of loss due to evaporation is thus needed for improved management of water resources (e.g. Martínez-Granados et al., 2011; Massuel et al., 2014).

Direct measurements of lake evaporation are very rare and mostly limited to very short periods. The Eddy CoVariance (ECV) method allows to measure the vertical turbulent fluxes within the surface atmospheric boundary layer and in turn the latent heat flux. It has been led, for example 51 days in Japan in a study by Ikebuchi et al. (1988), 42 days in Indonesia by Sene et al. (1991), 20 and 44 days in Israel by Assouline and Mahrer (1993) and also in Israel, 21 days then 104 days by Tanny et al. (2008, 2011). More recently, scintillometers have been used to help measure the latent heat flux above the entire open water surface and not only a small area as for the ECV method, thus taking into account potential edge or advection effects in the measurements. McJannet et al. (2013a) used the scintillometry method and obtained reliable estimates of evaporation for a period of 18 months over Logan's dam in Australia. They compared evaporation derived from scintillometer data over a transect covering the whole dam to ECV data obtained on a floating platform at the centre of the dam, with a very good consistency between both measurements. However, this work focused on a lake with a homogeneous surface. Computation of available energy for a heterogenous surface (open water and vegetated or bare soils) is more complicated. Latent heat is indeed retrieved as the residual term of the energy balance from sensible heat measured by the scintillometer. Some authors have also shown the importance of edge effects (Webster and Sherman, 1995; Condie and Webster, 1997) which affect the equilibrium layer over the lake surface over relatively short distances only. For lakes with large surface changes over relatively short time periods, one can apply similar methods as those derived for vegetated surfaces (Chehbouni et al., 2000) but this is clearly beyond the scope of this study.

In the absence of a major experiment involving direct measurements such as ECV or scintillometry methods, indirect estimations based on theoretical formulae or geochemical tracers influenced by the evaporation process have been exploited for decades. Often derived from temperature data like Dalton, Penman and Bowen Ratio Energy Balance (BREB), they have been used in different environmental conditions and varying data availability. Chemical approaches have been less widely used but generally give good results. At the annual time scale, Vallet-Coulomb et al. (2001), and using 50-day intervals, Gibson et al. (1996), found less than 10% difference from the results of physics-based approaches.

In a thorough review, McMahon et al. (2013) discussed the theoretical background of many indirect approaches in detail. But generally speaking, authors do not agree which indirect method is the most reliable, and estimated evaporation rates can vary as much as 30%, as shown by Tanny et al. (2008) who compared five combined methods (Penman, Penman–Monteith–Unsworth, Penman–Brutsaert, Priestley–Taylor and Penman–Doorenbos–Pruitt) with direct measurements. Daily evaporation estimated with the BREB method can differ by a factor of 4 from daily evaporation measured with eddy correlation (Ikebuchi et al., 1988). McJannet et al. (2013b) obtained quite different results with combined models, for example, a difference of 15% between modelled and measured evaporation using the De Bruin–Keijman and Priestley–Taylor methods but of only 5% with the Penman–Monteith approach. Kashyap and Panda (2001) found discrepancies from −1.36 to +22.32% between the estimation of evapotranspiration with ten different methods (combinated, radiation and temperature based) and direct measurement by lysimeter in India.

Comparisons between several indirect methods also showed significant discrepancies. McMahon et al. (2013) pointed to differences of 20% between evaporation estimated using the Penman equation with two different wind functions for 68 sites in Australia. Elsawwaf et al. (2010a) showed a very bad linear correlation ($R^2 < 0.3$) between monthly evaporation rates estimated with BREB and five other traditional methods (Priestley–Taylor, De Bruin-Keijman, Dalton derived mass transfer method, Papadakis, and Penman).

Evaporation pan data are used worldwide because measurements are simple and long historical data series are available. Such data are generally a good proxy for the climatic variations around a lake, even if actual evaporation is overestimated particularly because of the difference in water quality, thermal inertia, advection and edge effects (Riou, 1975; Morton, 1986; Oroud, 1995; Fu et al., 2004; Tanny et al., 2008; Lowe et al., 2009). Morton (1986) addressed the size effect of the water body in his comprehensive study. Fu et al. (2004) tested different sizes and shapes of pans (from 20 m^2 to less than 1 m^2) depending on the season, but comparisons between direct measurement of lake evaporation and pan evaporation are extremely rare: only Tanny et al. (2008) had seven days of overlapping data at their disposal. During this very short period, pan evaporation overestimated actual evaporation by about 65%. Conversion coefficients (C_C) proposed to transform pan observation into evaporation from the lake, are generally based on indirect methods of calculation (Riou, 1975; Allen et al., 1998; Craig, 2006).

Indirect methods require good quality data. Otherwise they suffer from major uncertainties identified by many authors (e.g. Morton, 1986; Lowe et al., 2009; Elsawwaf et al., 2010b; McMahon et al., 2013; McJannet et al., 2013b). This is especially important in semi-arid regions where data are often less frequent, less reliable and less representative (e.g. Kashyap and Panda, 2001; Sivapragasam et al., 2009), while environmental variability is higher. In a semi-arid context, extrapolating local measurements to a larger area or even to a neighbouring catchment is risky, as emphasized by the PUB (Predictions in Ungauged Basins) initiative (e.g. Sivapalan et al., 2003). Identifying the most suitable indirect method is difficult and, based on the wide range of discrepancy observed in the literature, a single approach can lead to unreliable results. The easiest way to reduce uncertainties is thus to combine several approaches.

The aim of this paper is not to discuss the theoretical validity of the different methods, but to deal with the uncertainty that can be expected in a typical case. Beyond enabling the improvement of the local water budget, our analysis of the uncertainty inherent in each approach and of the discrepancy between the results obtained with the different methods will be useful for scientists working in similar imperfect conditions, i.e. intermittent meteorological measurements, short data time series, gaps in data sets. We used a pragmatic approach, the most appropriate for situations when data are scarce, discontinuous or of limited reliability. We considered a typical large dam, the El Haouareb dam in the Merguellil catchment (central Tunisia) where the water balance is mainly determined by evaporation and infiltration through fissures in the limestone bedrock. Like in many previous studies (e.g. Rosenberry et al., 2007; Ali et al., 2008; Elsawwaf et al., 2010a), we first compared the results of different physical approaches: the mass transfer method (Dalton), the energy budget (BREB), and a combined approach (Penman). In the second stage, we calculated evaporation from temporal variations in chloride and ^{18}O, and compared the different results with Colorado pan measurements made at the dam site. Then we ran different sensitivity tests by analysing the influence of individual and combined inputs on evaporation calculation for each method by varying their values. To cross validate the different approaches, we compared the evaporation amounts and the results of sensitivity tests and finally obtained annual and monthly evaporation amounts. We finally proposed Cc values for the Colorado pan at the annual and seasonal scale.

2. Study area

The Merguellil catchment in central Tunisia (Fig. 1) has been the subject of many studies, and many describe the study area and the main problems of regional water management (e.g. Leduc et al., 2007; Le Goulven et al., 2009). This catchment is emblematic of hydrological processes which have been profoundly modified by human activities, including soil and water conservation works, large and small dams, and agricultural development. The Wadi Merguellil plays an important role in the area because it recharges the Kairouan plain aquifer, and this plain has the biggest potential for agricultural development in Tunisia (Le Goulven et al., 2009).

Wadi Merguellil is a typical Mediterranean ephemeral river. Its upper sub-catchment (1200 km^2) is mountainous (altitude between 200 and 1200 m a.s.l.) and is characterized by very diverse geology, morphology, vegetation and land use (e.g. Lacombe et al., 2008). The lower sub-catchment is part

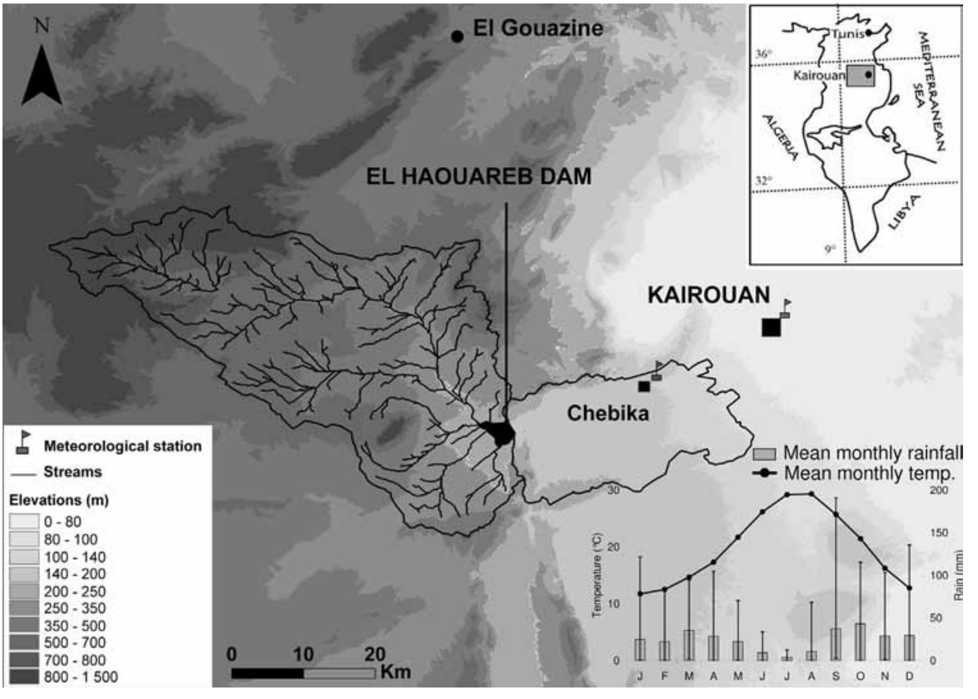

Fig. 1. Limits of the upstream and downstream Merguellil sub-catchments and location of the meteorological stations. The El Haouareb reservoir is the black shape. Chart of climatic characteristic of Kairouan: curve of mean monthly temperatures for 1957–2012 and bars of mean monthly rainfall for 1977–2009 with error bars showing extreme monthly rainfall.

of the large, flat, Kairouan alluvial plain (altitude between 50 and 200 m a.s.l.), which covers an area of about 3000 km^2. The climate is semi-arid, with very marked seasonal fluctuations (Fig. 1); annual mean rainfall ranges from 265 mm in the plain to 515 mm in the highest part of the catchment. In the higher area, there is an average of 70 rainy days per year, 90% of which are between September and May. Monthly temperature ranges from 10 °C in January to 30 °C in July and August (Fig. 1). Penman potential evapotranspiration is 1600 mm a^{-1}.

The El Haouareb dam was built in 1989 over a limestone sill (altitude 200 m) between the upstream and the downstream sub-catchments. The main aim was to protect the city of Kairouan from flooding, but also to enable the development of irrigation schemes downstream (Le Goulven et al., 2009).

Before the dam was built, Merguellil flood water was the main source of recharge to the Kairouan plain aquifer. The Wadi water is now stopped by the dam which loses much of its water by infiltration through fissures in the Mesozoic limestone and by evaporation. Since the construction of the dam, there has been no surface runoff downstream, except for very rare dam releases. The total reservoir water budget between 1989 and 2006 (Leduc et al., 2007) comprised 52% infiltration, 30% evaporation, 12% pumping and 6% dam releases. The theoretical maximum active storage (217 m, $104 * 10^6$ m^3) has never been reached: the highest level of the reservoir (211.5 m) was observed in February 2006 but, because of the progressive silting up of the reservoir, the depth of the water at that time (6.5 m) was shallower than the maximum (10 m) observed in 1993. In fact, the El Haouareb reservoir has been less than 3 m deep for 60% of the last 23 years, and dried up completely for 37 months.

Measurements of temperature, electrical conductivity (EC) and isotope content revealed vertical and horizontal homogeneity of the water in the reservoir. The water is generally dominated by SO_4 and Cl balanced by Ca and Na; mineralization may vary slightly depending on the geographical origin of tributaries and their local geological context (mainly sandstone, limestone, and marl with evaporites).

3. Available data

Regular hydrological surveys of the Merguellil catchment have been carried out by the Tunisian Ministry of Agriculture for the last 40 years. The level of the El Haouareb reservoir is measured every morning (accuracy of 1 cm) and pumped volumes are estimated from the rated capacity of the pumps and the pumping time. Other cooperating institutes (e.g. the French Institute for Research and Development – IRD, the University of Sfax, IAEA) occasionally perform physical and chemical measurements (EC, temperature, pH, and alkalinity) and geochemical analyses (mainly major ions, ^{2}H, ^{18}O and rarely ^{3}H and ^{14}C).

Capacity curves (height, surface area, volume) of the reservoir were originally defined by a precise topographic survey before the dam was built. These curves were updated when the reservoir dried up for a long time in 1994, 1997, 2002 and 2008. In 2008, silting reached a depth of 7 m at the foot of the embankment.

Since 1989, meteorological observations by the Tunisian Ministry of Agriculture at El Haouareb consist in daily measurements of Colorado pan evaporation and rainfall; and, since 2003, measurements of the temperature of the air, the pan water, and water in the reservoir (at 8 AM). The temperature of the reservoir water varies in space and over time, with daily differences of up to 1.5 °C depending on the season and the depth of the water in the reservoir. Other meteorological data came from the airport meteorological station at Kairouan, which is located in the plain at an altitude of 70 m a.s.l., 40 km east of the dam (Fig. 1); hourly air temperature, relative humidity dew point, 2-m wind speed, and atmospheric pressure have been measured since 1997. Additional solar data were obtained from the NASA solar worldwide database (SSE 6.0) with a resolution of one degree square from January 2000 to June 2006. We also used data from the CESBIO experimental station at Chebika, 20 km downstream from the dam at an altitude of 120 m a.s.l. (Fig. 1): air temperature and relative humidity were measured at 30-min intervals from November 2010 to May 2012. The barometric pressure recorded at the El Haouareb dam since 2007 is in good agreement with that recorded at the Kairouan meteorological station.

From 2005 to 2011, the geochemical survey of the water in the reservoir consisted in analyses of 37 isotopic contents (^{2}H and ^{18}O by mass spectrometry) and 22 major ions (ionic chromatography), combined with EC measurements three times a month after 2002, routinely performed as part of overall monitoring by the technical staff of the dam. Isotopic contents of the dam water from 1997 to 2000 (Ben Ammar et al., 2006) were used as a complement, as were five isotope analyses of water from Wadi Merguellil. Isotope contents of the rain at Tunis, 150 km to the north, and Sfax, 130 km to the south-east, were taken from the IAEA databank of the Global Network of Isotopes in Precipitation (GNIP). We also used the isotope contents of the atmospheric moisture and precipitation from a survey conducted by Gay (2004) at El Gouazine, 60 km to the north, in a similar environmental context (Fig. 1). δ^{18}O and δ^2H are reported relative to the Standard Mean Ocean Water (SMOW). The estimated accuracy of isotopic analyses was ±0.15‰ for δ^{18}O and ±1‰ for δ^2H.

4. Methods

Evaporation was first calculated using three widely used physical methods based on contrasted theoretical considerations: Dalton's Law (Dalton, 1802), the simplified BREB (e.g. Ali et al., 2008) and the combined Penman approach applied to free water surfaces (Riou, 1975) with a data set covering 2003–2006 at least, and 2003–2010 for Dalton. We also performed geochemical calculations based on the mass conservation law (such as the isotopic budget) for 16 periods between 2005 and 2010, and the chemical budget for four periods between 2005 and 2011. These two geochemical approaches are described in many studies (e.g. Gonfiantini, 1986; Rozanski et al., 2001; Vallet-Coulomb et al., 2001; Jones and Imbers, 2010). Results of physical and geochemical theoretical methods were then compared to measured Colorado pan evaporation, and the sensitivity of each method to various parameters was tested.

4.1. Dalton's Law

The Dalton approach assumes that evaporation is a function of wind speed and of the difference between the vapour pressure of the water surface and the atmosphere. It uses an empirical mass transfer coefficient to explain the relation between the different parameters through the wind function. It is expressed as:

$$E = f(u) * (e_s - e_a) \tag{1}$$

where $f(u)$ is the wind function depending on 2-m wind speed u (m s^{-1}), e_s is the pressure of saturated water vapour at the temperature of the water surface (mbar) and e_a is the water vapour pressure above the evaporating surface in mbars as a function of relative humidity (RH), and air temperature (T_{air}).

Several authors highlighted the complex determination of the theoretical wind function (e.g. Condie and Webster, 1997; Chu et al., 2010; McJannet et al., 2013a; McMahon et al., 2013), notably because it involves estimating the turbulence over a water body (e.g. Vercauteren et al., 2009; McJannet et al., 2011). Here, we chose the simplest method. Like Ali et al. (2008), we used the term ($b * u$) as a linear wind function where b is the Dalton empirical constant estimated for each study site multiplied by u which is the 2-m wind speed (m s^{-1}). The empirical constant b was estimated based on local pan information: we determined b as the slope of a linear regression intercepting the origin between the measured evaporation of a pan and ($u^*(e_{\text{span}} - e_a)$), where e_{span} is the pressure of saturated water vapour at the temperature of the pan water surface as in Webb (1966) and Riou (1975).

We used wind, atmospheric pressure, relative humidity and air temperature data from the airport meteorological station at Kairouan (40 km from the site) measured on an hourly basis. The actual temperature measured at the surface of the reservoir each morning at the site was also used.

4.2. The simplified Bowen ratio energy balance method

The BREB method estimates net available energy based on the surface energy balance and on the flux gradient relationship between latent heat of water evaporation and sensible heat conducted and convected from the surface of the water to the overlying air. We used the simplified BREB method which neglects the heat flux to the bottom of the water body and the energy advection into the water body (Ali et al., 2008).

$$\mathrm{E}_{\mathrm{BR}} = \frac{Rn - G}{\ell(1 + \beta)} \tag{2}$$

where Rn is net radiation (MJ m^{-2} d^{-1}), ℓ=2.45 MJ.kg^{-1} is the latent heat of evaporation, β is the Bowen ratio, i.e. the ratio of latent heat to sensible heat, G is the heat gained or lost by the water mass proportional to the depth of the water body and temperature changes (MJ m^{-2} d^{-1}) according to the relationship:

$$G = \rho * C \cdot d * (\Delta T/\Delta t) \tag{3}$$

where ρ is the density of water (1000 kg m^{-3}), C is the heat capacity of the water (4.186 * 10^{-3} MJ kg^{-1} °C^{-1}; d is the mean depth of the reservoir (m), and $\Delta T/\Delta t$ is the change in water temperature during the period Δt (°C d^{-1}).

Equations determining Rn and β are detailed in Ali et al. (2008) and Gianniou and Antonopoulos (2007). We used solar radiation data from NASA, atmospheric pressure, air temperature and relative humidity data from the weather station at Kairouan and the temperature of the water in the reservoir measured at the study site to determine Rn and to calculate β. The water depth of the reservoir was calculated from the geometry of the reservoir and daily measurements of the water level. The daily water temperature at the dam surface was used as a proxy for the temperature of the whole reservoir.

4.3. The Penman combined approach applied to free water surfaces

This method (Penman, 1948) combines an aerodynamic and a radiative component. In the Penman approach, the evaporating surface temperature is assumed to be equal to water surface temperature. Unlike the BREB method, heat storage, the exchange of heat with the reservoir, and advected energy are neglected and so the actual evaporation does not affect the overpassing air. As the cumulative heat gained or lost by the water mass (G in the BREB equation) was about 50 times lower than Rn (not shown), G was not included in the available energy used in the Penman equation below:

$$E = \frac{\frac{Rn}{\ell} + \frac{\gamma}{\Delta} \cdot Ea}{1 + \frac{\gamma}{\Delta}} \tag{4}$$

where E_a is evaporation calculated according to Dalton's Law (Eq. (1)), Δ is the slope of the curve of saturation vapour pressure as a function of temperature (kPa °C^{-1}), and γ is the psychrometric constant.

4.4. The isotopic and chemical budgets

The mass conservation law also applies to any part of the water molecule (hydrogen or oxygen) or dissolved elements (e.g. chloride). The mass balance equation written for the chosen tracer has the following general form (Rozanski et al., 2001):

$$C_L \cdot \frac{dV_L}{dt} + V_L \cdot \frac{dC_L}{dt} = C_{Qi} \cdot Qi + C_P.P - C_{SWo} \cdot SWo - C_{GWo} \cdot GWo - C_E \cdot E \tag{5}$$

where, for a chosen period (dt), V_L is the volume of the reservoir, P is rainfall on the surface of the water, Qi is volume of water entering the reservoir, E is the volume of evaporation, GWo are losses due to seepage, and SWo are outputs by pumping or releases. Each term of the water balance is weighted by its chosen tracer concentration C (e.g. ratio $\delta^{18}O$ ‰ or chloride concentration) written with its corresponding index term.

Given the homogeneity of the water in the reservoir, the isotope content of infiltrated and pumped waters was considered to be identical to that of the reservoir water. The evaporated water vapour contains no chloride and the isotope content of the evaporated vapour (δE) has to be estimated. The Craig and Gordon model (1965) allows estimation of the isotope content of the evaporated vapour (δE) according to the relationship:

$$\delta E = \frac{\alpha_{V/L} \cdot \delta_L - h_N \cdot \delta A + \varepsilon_{V/L} + \varepsilon_{diff}}{1 - h_N - \varepsilon_{diff}} \tag{6}$$

where δ_L is the measured isotope content of the reservoir water, h_N is the relative humidity normalized to saturation vapour pressure at the temperature at the air–water interface, δA is the isotope composition of the atmospheric moisture, $\varepsilon_{V/L} = \alpha_{V/L} - 1$ is the equilibrium fraction factor, ε_{diff} is the kinetic fraction factor (for $\delta^{18}O$ it has been shown to be approximately $14.2(1 - h)$ ‰ by Gonfiantini (1986)), $\alpha_{V/L}$ is the equilibrium fraction factor between the vapour and the water at the temperature of the surface of the reservoir.

A simplified equation neglecting δA and derived from the alternative equation of Benson and White (1994) was given by Jones and Imbers (2010). This simplification, based on the same evaporation theory as the Craig and Gordon equation, assumes that the atmospheric water overlying the lake is derived from evaporation rather than from atmospheric moisture:

$$\delta E = \frac{\delta_L \cdot \alpha_{V/L}}{\left(\frac{1 - RH}{\alpha_{kin}}\right) + RH} \tag{7}$$

where α_{kin} is the kinetic fractionation factor depending on wind speed (=0.994 for wind speed under 6.8 m s^{-1}) and RH is the relative humidity.

We used the atmospheric pressure, air temperature and relative humidity data from the meteorological station at Kairouan and the temperature of the water in the reservoir measured at the study

Table 1
Annual means determined with BREB, Penman and Dalton, and annual and monthly extremes for 2003–2006 compared with Colorado pan measurements.

	BREB	Penman	Dalton	Colorado pan
Mean (mm a^{-1})	1400	1600	1750	2049
Min (mm a^{-1})	1370	1590	1670	2003
Max (mm a^{-1})	1440	1640	1880	2110
Monthly min (mm d^{-1})	0.4	0.4	1.5	1.6
Corresponding month	*December*	*January*	*December*	*December*
Monthly max (mm d^{-1})	8.6	9.9	11.0	11.6
Corresponding month	*July*	*July*	*July*	*July*

site. The total volume of the reservoir was calculated from the topographic survey and the water level measured each day. The chloride and ^{18}O data used for each term of the balance equation (5) are discussed below.

5. Results

Because data were available, the BREB and Penman theoretical methods were implemented using the period 2003–2006, the period 2003–2010 was used for Dalton, and the period 2005–2011 was used for the ^{18}O balance and the chloride balance. Because the physical methods work better with time steps of more than ten days (Riou, 1975; Xu and Singh, 1998; Gianniou and Antonopoulos, 2007), evaporation was calculated at monthly and annual scales. Isotopic and chemical balances were modelled using varying periods, but always exceeding 10 days.

5.1. Physical approaches

Annual Colorado pan evaporation ranged from 2000 to 2100 mm over the period 2003–2006 (Table 1), very close to the average of 2037 mm since 1989 (Fig. 2). In the last two decades, (1989–2009) the first and third quartiles were 1920 mm a^{-1} and 2104 mm a^{-1} respectively, and the extreme values 1758 and 2351 mm a^{-1}. Evaporation is highest in July and lowest in December (Fig. 2).

For the period 2003–2006, the Penman, BREB and Dalton theoretical methods produced an average annual evaporation of 1600 mm, ranging from 1370 mm to 1880 mm (Table 1). The monthly averages, minima and maxima are shown in Fig. 3.

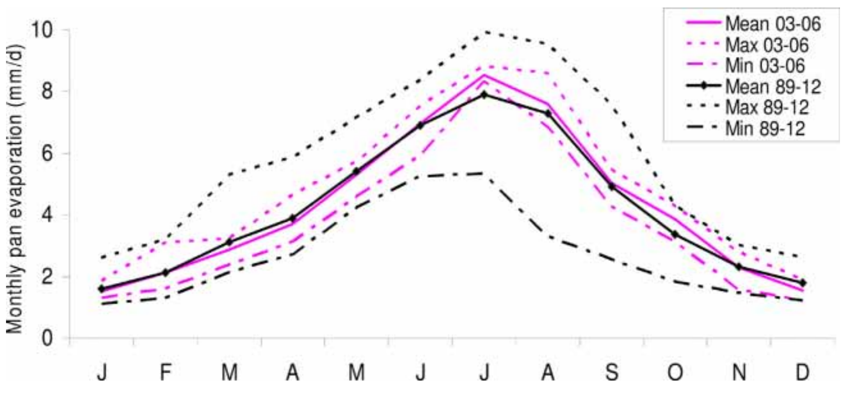

Fig. 2. Monthly means and extremes of evaporation measured in the El Haouareb Colorado pan.

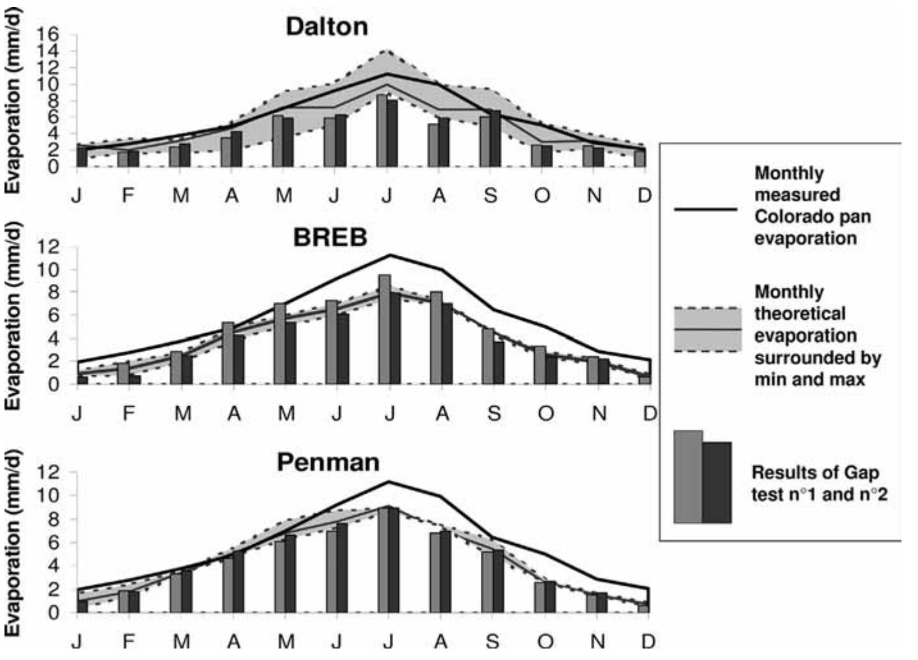

Fig. 3. Monthly evaporation determined with the BREB, Dalton and Penman methods (thin black line), bounded by the extreme values (grey area) and monthly Colorado pan measurements (thick black line) between 2003 and 2006; results of the two tests of robustness (Gap test n°1 and n°2).

With the BREB method, the calculated Bowen ratio (β) was 0.1, in good agreement with other studies in semi-arid climates (e.g. Ali et al., 2008; Gallego-Elvira et al., 2009). The Dalton empirical constant (b) estimated using Colorado pan data was 0.37.

The meteorological data required by the different physical approaches are rarely available in semi-arid areas, so hydrologists substitute information from another site or another period. In our case, some data were taken from Kairouan meteorological station (40 km away). Relative humidity is quite similar in Chebika (20 km away) and Kairouan, with an average monthly deviation of less than 2%, but in Chebika the temperature is about 2 °C lower in winter and 1 °C lower in summer. This can lead to significant biases in calculations and the sensitivity of theoretical approaches could be a major criterion in the choice of a method. For each method, we changed the values of input data within a range defined by the accuracy of field data and compared the resulting differences in monthly averages. Like the authors of similar studies (e.g. Vallet-Coulomb et al., 2001; Gianniou and Antonopoulos, 2007), we tested a deviation of 10% in input data such as relative humidity (*RH*), net radiation (*Rn*) and wind speed. We also tested a deviation of 1 °C in air and water temperature to avoid seasonal bias. Because water temperature was only measured at 8 AM, we tested a deviation of 2 °C in water temperature and of 20% in the integrative parameters (G, β and E_a) linked to water temperature. The main results are presented in Table 2.

Table 2
Percentage of prediction errors in monthly evaporation resulting from errors in the input parameters (values in brackets) estimated with the Dalton, BREB and Penman physical methods.

	Air temp. (±1 °C)	Water temp. (±1 °C)	Water temp. (±2 °C)	*RH* (+10%)	*RH* (−10%)	Wind (±10%)	*Rn* (±10%)	*G* (±20%)	β (±20%)	*Ea* (±20%)
Dalton	±13	±19	±38	−9	+42	±10				
BREB							±10	±5	±2	
Penman							±8			±7

In the Dalton equation, evaporation is proportional to wind speed and water temperature, and inversely proportional to air temperature. Fluctuations in relative humidity do not influence evaporation symmetrically. We also performed estimates of cumulative uncertainties linked to several parameters (not shown). Assuming high uncertainties in the climate data with a drier and colder climate in Kairouan than at the El Haouareb dam (relative humidity −10%, wind +10% and air temperature −2 °C) or a wetter and warmer climate in Kairouan (relative humidity +10%, wind −10% and air temperature +2 °C) the cumulative differences would be +56% and −54%, respectively. Assuming uncertainties in air temperatures (+1 °C), *RH* (−5%) and assuming that the temperature of the reservoir water measured at 8 AM underestimates the average daily temperature by 1 °C, the cumulative uncertainty would be 17%.

With the BREB method, a cumulative change of 10% in net radiation, 20% in the stock of heat (G) and 20% in the Bowen ratio would lead to a cumulative difference in evaporation of less than 20%.

The Penman approach is sensitive to changes in net radiation and evaporation estimated using the Dalton approach (E_a). With a cumulative error of 20% in the E_a term and of 10% in net radiation, the variation in evaporation would be 13% (corresponding to 0.6 mm d^{-1}). With highly inaccurate estimations (±50%) of the advective term and of net radiation (±20%), the cumulative error would be 30%, corresponding to 1.3 mm d^{-1}.

As physical time series are often incomplete in semi-arid areas, we also assessed the possible bias caused by missing data or data sets covering too short periods. The robustness of the methods was explored using two successive tests (Gap tests No. 1 and 2) in which we randomly removed 40% of daily data and used the modified monthly averaged data for calculation (Fig. 3).

With the Dalton method, the monthly mean errors in the two tests of robustness against data gaps were 13% and 16%, reaching 1.9 mm d^{-1} in August (Fig. 3). With the BREB method, monthly mean errors were 15% and 21%, reaching 1.6 mm d^{-1} in July. With the Penman method, monthly mean errors were 4% and 6% and reached 0.4 mm d^{-1} in August.

5.2. Isotopic and chemical budgets

The chloride budget (Eq. (5)) was calculated for four periods of emptying for which we disposed of eight major ion contents between April 2005 and February 2011, each of them lasting between 27 and 58 days. Results are shown in Fig. 4 and Table 3.

In order to better exploit the field EC measurements, we investigated the correlation between 22 chloride contents (from analyses of major ions) and field EC measurements. As expected, the correlation exists for many samples but one third of them shows a deviation of more than 30% without any obvious geochemical explanation.

Similarly, the EC calculated by the software Diagrammes (Simler, 2004) from major ions showed some clear differences with the 22 EC measured in the field: the deviation was higher than 0.3 mS cm^{-1} in five samples, and reached 0.6 mS cm^{-1} in two samples.

We extended the chloride budget calculation to 13 other periods lasting from 20 to 44 days between April 2003 and August 2007, for which we had 66 EC field measurements for the reservoir, ranging from 1.9 to 3 mS cm^{-1}. The results were disappointing: field EC measurements lead to a negative calculation of evaporation and for three periods calculated evaporation was almost zero when pan evaporation was about 10 mm d^{-1}. Others ranged from 3.3 to 10.3 mm d^{-1} when pan evaporation ranged from 3.2 to 11.7 mm d^{-1} with ratios versus pan evaporation varying from 0.4 to 1.6. The difference between evaporation calculated from reconstructed EC and chloride contents was up to 2.5 mm d^{-1} (Table 3).

The isotopic budget (Eq. (5)) was calculated using 23 samples selected from three main episodes of lake emptying between 2005 and 2011, during which the water in the reservoir was regularly sampled, and four pairs of samples of distinct episodes of lake emptying (Fig. 4).

The isotope contents of the reservoir formed a clear evaporation line that deviates from the Tunisian meteoric water line defined by Celle-Jeanton et al. (2001) as $\delta^2H = 8.\delta^{18}O + 11$ (Fig. 5). The average slope of the evaporation line (4.7) varied slightly (from 4.4 to 4.8) between the reservoir emptying episodes.

The monthly isotope contents of the rain recorded by the GNIP stations were available from 1967 to 2006 for Tunis and from 1992 to 2008 for Sfax. The main features recorded by the two meteorological stations are quite similar, in particular, high $\delta^{18}O$ variability, with extreme values of 6.70‰ and

Table 3
Comparison of average Colorado pan evaporation (Epan) and evaporation (E) estimated using the chemical budget based on chloride contents, EC field measurements, and EC calculated with Diagrammes for four periods between 2005 and 2011.

		Chloride contents			EC field measurements			Calculated EC		
Period from... to...	Epan ($mm\,d^{-1}$)	Cl ($mg\,L^{-1}$)	E ($mm\,d^{-1}$)	Ratio E/Epan	EC ($mS\,cm^{-1}$)	E ($mm\,d^{-1}$)	Ratio E/Epan	EC ($mS\,cm^{-1}$)	E ($mm\,d^{-1}$)	Ratio E/Epan
12/04/05	5.8	174.9	7.7	1.39	1.83	6.8	1.22	1.81	5.2	0.94
16/05/05		193.2			1.99			1.93		
25/06/07	11.5	336.6	7.8	0.70	2.34	4.6	0.41	2.46	5.5	0.49
25/07/07		384.5			2.53			2.71		
17/10/08	3.7	124.0	4.7	1.32	1.42	(−4.5)	(−1.22)	1.49	3.3	0.98
13/11/08		134.0			1.32			1.58		
06/12/10	1.6	72.4	1.1	0.74				1.56	2.6	0.84
02/02/11		74.2						1.60		

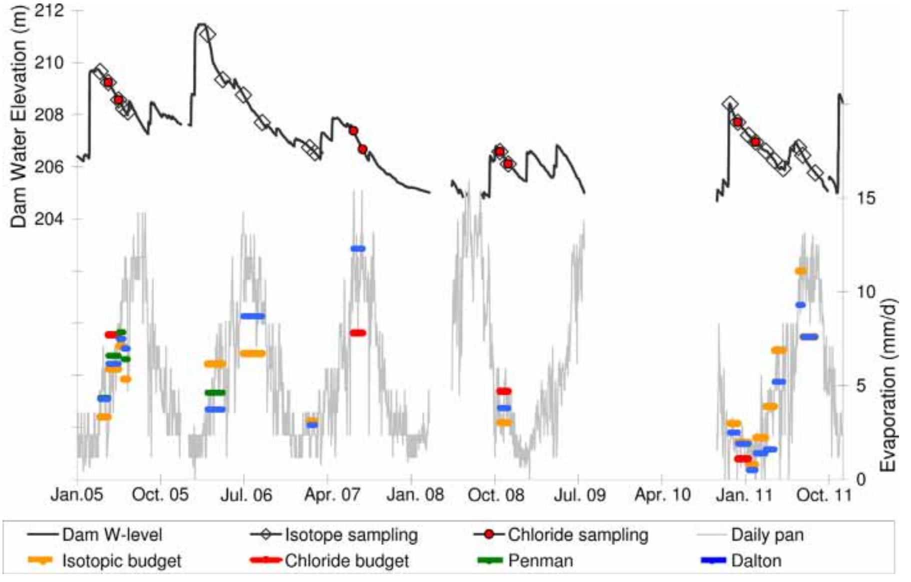

Fig. 4. Daily water level in the dam (black line) and measured evaporation at the Colorado pan (pale grey curve) from 2005 to 2011 with the dates of the isotope and chloride sampling (diamonds and red dots). Evaporation rates from the isotopic budget (orange symbols) chloride budget (red symbols) compared to evaporation rates calculated using the physical methods of Penman and Dalton (green and blue symbols).

−10.60‰ (average −3.47‰). Other records were kept by Gay (2004) at El Gouazine from January 2000 to May 2005: the isotope contents of the 32 rainfall events sampled showed marked variability (from 3.32‰ to −9.69‰) with no clear seasonal trend. Because the Tunis meteorological station record was more complete, we used the monthly average δ^{18}O of Tunis rainfall as the precipitation signal for the reservoir in our isotopic budget, with the altitudinal correction of −0.2‰ per 100 m for δ^{18}O proposed by Ben Ammar et al. (2006).

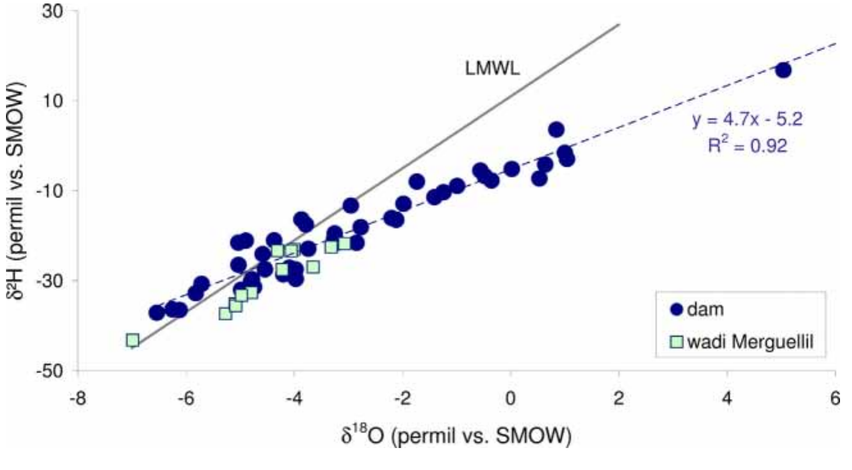

Fig. 5. δ^2H and δ^{18}O (‰ vs. Standard Mean Ocean Water – SMOW) of the reservoir (2005–2011) and of Wadi Merguellil (1997–2011) from this study and the study by Ben Ammar et al. (2006). The black line represents the Local Meteoric Water Line (LMWL) of Tunisia from Celle-Jeanton et al. (2001), and the blue dashed line represents the evaporation line of the dam reservoir.

Table 4
Averaged Colorado pan measurements and values calculated from the 2005 to 2011 isotope budgets, compared with the Penman and Dalton methods and the chloride budget (mm d^{-1}). The isotopic budget was applied for δE derived from Eq. (7) then for δE derived from Eq. (6) with the monthly values of δA estimated by Gay (2004).

Periods (from... to ...)		Pan (mm d^{-1})	Physical approaches		Isotope budget		Chloride budget
			Penman	Dalton	δE from Eq. (7)	δE from Eq. (6) and monthly δA	
16/03/05	12/04/05	3.9	4.4	4.3	3.4	5.0	
12/04/05	16/05/05	5.5	6.6	6.2	5.9	8.9	7.7
16/05/05	01/06/05	8.2	7.8	7.5	7.1	10.8	
01/06/05	15/06/05	9.3	6.4	7.0	5.4	8.5	
Mean spring 2005		6.4	6.1	5.9	5.3	8.0	
04/03/06	22/04/06	5.3	4.6	3.7	6.2	9.3	
29/06/06	30/08/06	11.0		8.7	6.8	11.7	
01/02/07	19/02/07	2.5		2.9	3.1	4.5	
25/06/07	25/07/07	11.5		12.3			7.8
17/10/08	14/11/08	3.5		3.8	3.0		4.7
11/11/10	06/12/10	2.0		2.5	3.0	2.9	
06/12/10	10/01/11	1.6		1.9	1.9	2.0	1.1
10/01/11	02/02/11	1.4		0.5	0.8	0.8	
02/02/11	07/03/11	2.0		1.4	2.1	2.3	
07/03/11	05/04/11	3.6		1.6	3.6	4.2	
05/04/11	04/05/11	4.8		5.2	6.1	7.5	
Mean winter 2010–2011		2.6		2.2	2.9	3.3	
23/06/11	07/07/11	9.6		9.3	11.1	25.9	
07/07/11	17/08/11	10.3		7.6	7.5	14.6	
Mean summer 2011		10.2		8.0	8.4	17.5	

Incoming Wadi Merguellil water showed a wide range of $\delta^{18}O$, between −7‰ and −3‰ (Fig. 5). For our calculations, we used the mean $\delta^{18}O$ of the runoff water across the upstream catchment of −5‰ (Ben Ammar et al., 2006).

Direct measurements of the $\delta^{18}O$ of atmospheric moisture (δA) require complex instrumentation and are very rare in the literature. But it can be deduced by monitoring changes in $\delta^{18}O$ of water in tanks. Monthly averages of δA were obtained by Gay (2004) based on one year of measurements made in evaporation tanks on the embankment of the small El Gouazine reservoir, not far from our study area (Fig. 1). Standard deviations in monthly averages of δA were ±1.8‰ and measurements made on the embankment were up to 2.5‰ more depleted than occasional measurements made in another tank located near the reservoir.

We first used the monthly δA estimated by Gay to calculate $\delta^{18}O$ of the evaporated water vapour (δE) with the Craig and Gordon model (Eq. (6)). Values ranged from −18.7‰ to −5.4 ‰ with an average of −11‰. We then calculated $\delta E^{18}O$ from the simplification of Jones and Imbers (2010) (Eq. (7)). Results ranged from −15‰ to −17.5‰.

Finally, we calculated the evaporation rates using the 2005–2011 isotope budgets. Table 4 and Fig. 4 compare these results with the Colorado pan measurements. When possible, they were also compared with the Penman and Dalton physical approaches.

The sensitivity of the isotopic budget to ^{18}O variability in inputs was tested with the maximum and minimum values observed for Wadi Merguellil (−3‰ and −7‰) and rain (6.70‰ and −10.6‰). Induced variations in evaporation were smaller than 0.5 mm d^{-1}.

6. Discussion

6.1. Physical approaches

As expected, evaporation calculated with the Penman, Dalton and BREB methods was generally lower than pan evaporation, and the annual amplitude was also lower. Dalton monthly rates were generally higher than those determined by Penman and BREB but very variable, sometimes lower than BREB and Penman or higher than pan evaporation. Dalton interannual seasonal means were very close to pan evaporation in autumn, winter and spring (but about 2 mm d^{-1} lower in summer); this is not surprising because the Dalton coefficient is determined using pan data. Penman and BREB reproduced the seasonal trend of pan evaporation rates quite well. BREB evaporation was always lower than pan evaporation (1–3 mm d^{-1} in summer). Penman monthly deviations remained relatively stable over the year (between 1.5 and 2 mm d^{-1}) but spring evaporation was high, close to pan evaporation. These results are consistent with other studies under various climates: Rosenberry et al. (2007) on Mirror Lake (northeastern temperate USA), Sene et al. (1991) on Toba Lake in tropical Indonesia, and Sadek et al. (1997) on Lake Nasser in Egypt, reported a higher value calculated with Penman than with BREB. In several studies conducted in western and central Africa (Riou, 1975; Pouyaud, 1990) the Dalton calculation was higher than Penman.

BREB is sometimes considered to be the most accurate method (e.g. Rosenberry et al., 2007; Gianniou and Antonopoulos, 2007; Ali et al., 2008) but this is discussed. Comparing BREB with direct measurements in Lake Tiberias (Jordan Valley), Assouline and Mahrer (1993) showed that BREB underestimated evaporation for some periods, probably because of heat advection from the areas surrounding the lake. Sene et al. (1991) working in Indonesia, Xu and Singh (1998) in Switzerland and Tanny et al. (2008) in Israel compared with their own direct measurements, and considered the Penman approach gave better results than the BREB method. Despite the fact the Penman method neglects the heat storage effect and heat exchange with the ground, it is widely used for estimating open water (McMahon et al., 2013). The modified Penman–Monteith method taking into account the heat storage term G and aerodynamic resistance is sometimes preferred (e.g. Craig, 2006; Gallego-Elvira et al., 2012; McJannet et al., 2013b). In our case, the monthly G was negligible compared with net radiation (Rn), and it was thus possible to use the simpler Penman equation.

Because of the limited quality of primary data, it is unreasonable to claim an exact uncertainty value, but sensitivity tests can predict an order of magnitude depending on the author's confidence in available input data. With 10–20% confidence in the input data, uncertainties are of about 20% for each method, but Dalton is particularly sensitive to parameters such as temperature and relative humidity. The need for accurate temperature data is a major weakness of the indirect methods, since the evaporating surface temperature used in the BREB, Penman and Dalton equations is actually replaced by one-off temperature measurements of the surface layer of the water (McMahon et al., 2013). Moreover, the overlying air temperature is actually replaced by the temperature measured at Kairouan. As spatial transposition of data is known to be one of the largest contributors to uncertainty in semi-arid environments (Sivapalan et al., 2003), robustness against input data is probably the most important criterion that should determine the choice of method. The comparison of data from the two nearby meteorological stations in the Kairouan plain (Kairouan and Chebika) showed that temperature is the parameter which can differ the most (by about 2 °C in winter). Temperature uncertainty has serious consequences for the reliability of calculations based on temperature data. BREB is much more sensitive to the surface temperature than the other two methods: the Bowen ratio is directly proportional to the difference between the temperature at the surface of the water and the temperature of the air, since in the other methods, temperature mainly affects net radiation. The Penman method appears to be more robust against temperature uncertainty; our results showed that a very poor estimate (50%) of the adiabatic term related to the gradient of saturation vapour led to an error of 30%.

As far as the surface temperature of a lake or reservoir is concerned, the use of remotely sensing is an interesting option. But daily products, such as those acquired by the MODIS platform (MOD11A1) are available only at large (1 km) spatial resolution, which, given the size of the El Haouareb dam, is too large. Devices such as scintillometers (e.g. McJannet et al., 2011, 2013a) should help future estimation of sensible heat but indirect methods will still be necessary, and these are subject to significant uncertainties, notably the estimation of the net radiation. The need for further investigation to estimate evaporation at the field site was underlined by several authors (e.g. Lowe et al., 2009; McJannet et al., 2013b).

Sensitivity tests results resemble those cited in the literature, with a discrepancy of less than 20%. Vallet-Coulomb et al. (2001) reported cumulative changes of less than 20% for BREB and Penman with a 10% error in each input term in Ethiopia, and Elsawwaf et al. (2010b) reported less than 15%

error for BREB, Penman, and Dalton on Lake Nasser in Egypt. Because of uncertainties in water and air temperatures, particular care should be taken especially using the BREB equation, as illustrated by previous works (e.g. Craig, 2006; Gianniou and Antonopoulos, 2007; McJannet et al., 2013a)

Elsawwaf et al. (2010a) reported that temperature uncertainty on Lake Nasser led to major errors in calculating the Bowen ratio, i.e. a deviation of about 20% in mean annual evaporation and of about 40% in monthly evaporation. Sene et al. (1991) and Singh and Xu (1997) reported that the Dalton approach was particularly sensitive to variations in the vapour pressure gradient calculated using air and water temperatures.

Despite the simplicity of the meteorological data required, the Dalton method is also weakened by the determination of the empirical factor (b), which is commonly derived from pan measurements, with the bias inherent to this method. Using the value of b calculated with a more accurate method and reported in the literature is not a valid alternative because several authors strongly advised against extrapolating a Dalton empirical factor even if the sites concerned are located in the same region or have a similar climate (Pouyaud, 1990; Singh and Xu, 1997; Kashyap and Panda, 2001; Tanny et al., 2008).

Assessments of a method's robustness against data gaps are rare in the literature. We observed that Dalton and BREB are very sensitive to gaps in the input data: deviations can reach 2 mm d^{-1} whereas they did not exceed 0.4 mm d^{-1} with the Penman method. This underlines the robustness of the Penman approach compared with both Dalton and BREB.

6.2. Geochemical approaches

The physical approaches mainly take the climatic characteristics into account to the detriment of hydrological, geomorphological or physical–chemical features, which can also play a significant role in evaporation (Lowe et al., 2009). The chemical approaches account for some of these aspects and the results may provide additional information on temporal variability, which depends on non-climatic features.

The hydrological functioning of the El Haouareb dam is rapid filling and slow emptying, which is suitable for calculating the chemical or isotope budget. The chloride budget was calculated for four periods, based on variations in chloride contents ranging from 2.5% to 12.5%. Given this range of variations, this approach is very sensitive. However, four results are not enough to estimate uncertainty. And in our case, the field EC data were not sufficiently precise for a chemical budget. These uncertainties prevented us from performing more calculations based on EC measurements to consolidate the results.

The slope of the evaporation line for the isotope contents of the lake water was in agreement with that identified in other studies based on evaporation tanks in Tunisia (Gay, 2004) and for endorheic systems in semi-arid Ethiopia (Kebede et al., 2009) and Australia (Van den Akker et al., 2011). This suggests that the system is not influenced by inputs (Rozanski et al., 2001) and has the advantage of not being really sensitive to uncertainties in the isotope contents of rivers and of the rain, which was confirmed by sensitivity tests.

The isotope budget requires a minimum of meteorological data for the calculation of δE, but using the Craig and Gordon model (Eq. (6)), estimation of δE is weakened by the uncertain isotope content of atmospheric moisture (δA). Based on tank measurements, Gay (2004) showed a standard deviation of the monthly δE of 1.8‰ and marked spatial variability (2.5‰ at a scale of a few hundred metres). Our estimated evaporation rates using the δA obtained by Gay (2004) were higher than Colorado pan measurements and are therefore most likely overestimated, since pan evaporation is considered to be higher than evaporation from the reservoir (Riou, 1975; Fu et al., 2004; Lowe et al., 2009): the isotopic evaporation rate reached 26 mm d^{-1} during the summer of 2011 when pan evaporation was 11 mm d^{-1}.

The simplification proposed by Jones and Imbers (2010) appears to give more realistic results: with pan evaporation and physical approaches, the mean deviations were 1.3 and 0.9 mm d^{-1} respectively. Even if their theoretical approximation is counterintuitive because the air above the lake has to be renewed to enable evaporation to occur, Benson and White (1994) showed experimentally that in the case of the Pyramid Lake (Nevada, USA) their data was consistent with the moisture immediately above

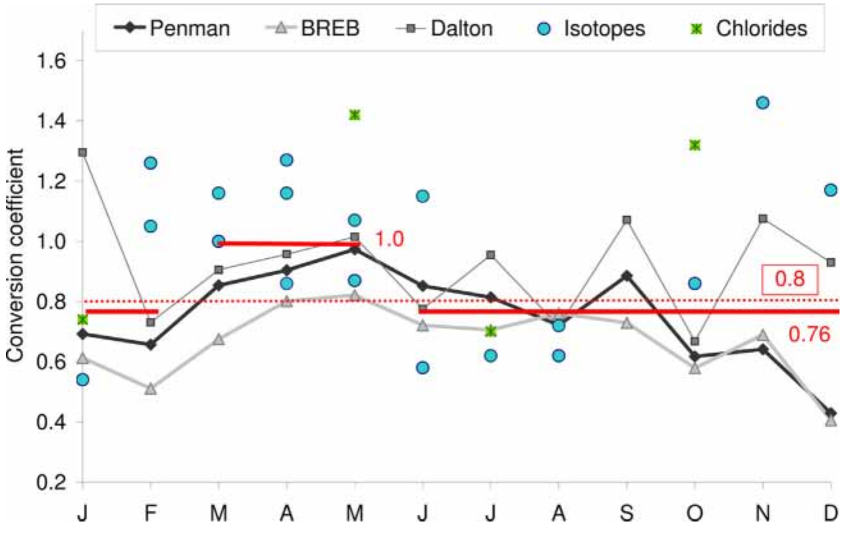

Fig. 6. Penman, Dalton and BREB monthly conversion coefficients estimated for the period 2003–2006 (thick dark, thick pale grey and thin medium grey lines) and occasional conversion coefficients from the isotopic budget for 2005–2011 using Eq. (7) (blue circles) and the chloride budget (green squares). The annual average C_C for the four approaches is represented by the red dotted line with the value in the red box; the seasonal averages are represented by red solid lines with two values written in red.

the laminar layer, which was completely lake derived. For Jones and Imbers (2010), this hypothesis was the most suitable one to model the isotope content of Mediterranean lakes.

Given the uncertainty related to the isotope content of the evaporated vapour and the sensitivity of the chloride budget used alone, the geochemical approach would not produce reliable results. In our study, we were able to compare the results with pan measurements and with the physical approaches; isotopic and chemical approaches thus provided valuable complementary information.

Reducing uncertainty requires comparing several approaches based on different concepts. However, physical and geochemical calculations can only be made over limited periods. The only homogeneous information available in our case was pan evaporation, measured daily from 1989 on. In the following section, we compare the results obtained with theoretical approaches for the period 2003–2010 with pan evaporation records in order to define the conversion coefficient between the Colorado pan and the lake evaporation.

7. Calculation of the Colorado pan conversion coefficient

Here we estimate a pan conversion coefficient (C_C) to convert pan measurement into evaporation from the dam. Monthly C_C was calculated for the period 2003–2006 for the physical approaches, and for the period 2005–2011 for the isotopic approach (using the simplification of Jones and Imbers; Eq. (7)) and the chloride budget (Fig. 6). Dalton C_C oscillates around a median value of 0.9. The Dalton C_C is actually equal to $(e_{\text{slake}} - e_a)/(e_{\text{span}} - e_a)$.

The Penman and BREB methods showed a global trend in which monthly C_C were higher during spring (0.7–1.0) and lower during winter (0.4–0.7). The C_C resulting from the isotopic approach also showed a general trend: high from February to May (0.9–1.6) and low from June to August (0.6–0.7).

The chloride C_c was high in spring and autumn (1.42 and 1.32) and low in winter and summer (0.74 and 0.70). The chemical and isotopic methods generally produced higher coefficients for autumn and winter than the Penman and BREB physical methods.

Even though calculations covered only short periods, a seasonal pattern appeared: monthly C_C was higher in spring and lower in summer. The trends revealed by the isotopic and chloride budgets were

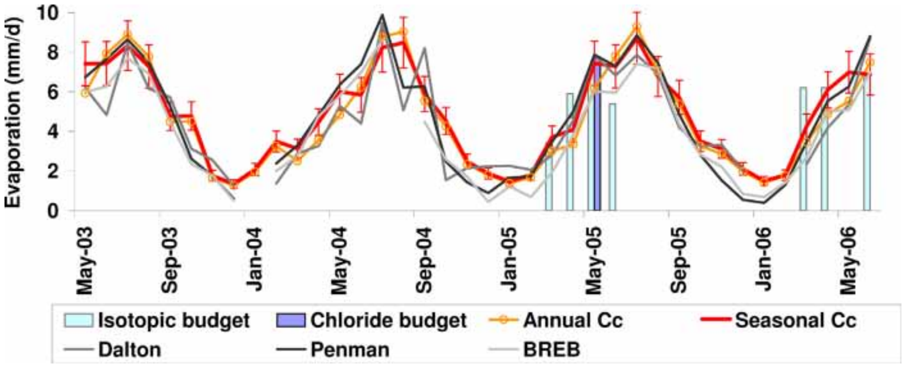

Fig. 7. Monthly evaporation determined using the Dalton, Penman and BREB physical approaches, occasional isotope and chloride budget, and the application of annual and seasonal C_C (with 15% error bars) for the period May 2003–June 2006 (mm d^{-1}).

consistent with trends revealed by the physical approaches. We were thus able to use an average value, but we discarded the Dalton results because of their direct dependence on the Colorado pan.

The seasonal averages of both isotopic and physical approaches were a C_C of 1.0 from March to May and of 0.76 for the rest of the year, i.e. an annual average C_C of 0.8.

The monthly evaporation rates calculated from the seasonal C_C differed by about 0.8 mm d^{-1} from rates calculated using the physical approaches. The overall trends and agreement of the results confirm the reliability of the order of magnitude obtained (Fig. 7). However, the variability of results obliged us to consider an uncertainty of 15%.

The use of seasonal coefficients gave an annual average of 1700 mm for the period 1989–2011. Using an annual C_C of 0.8 led to a deviation of less than 3% for the yearly total. At the monthly scale, the deviation may reach 12% for the summer months, but is insignificant for the winter months.

The problem of a conversion coefficient has been studied for many years, but no definitive solution has been found: previous studies underlined the difficulty in determining C_C and in identifying its temporal variability. Spatial transposition is also one of the main contributors to uncertainty (e.g. Kohler, 1957; Lowe et al., 2009; McJannet et al., 2013b).

C_C obviously depends on the type of pan. Equivalences between classes of pans have frequently been investigated: e.g. the correlation between Class A and Colorado pans defined by the comparison of several studies by Kohler (1957): $C_{C\,\text{class A}} = 0.86 * C_{\text{Colorado}}$. Fu et al. (2004) investigated the influence of size, depth and shape on C_c and its temporal variability using 15 types of pans. The transposition between a reservoir and a pan is much more problematic, C_C appears to be highly dependent on climate and relative humidity is the factor most correlated with C_C (e.g. Martínez Alvarez et al., 2007; Lowe et al., 2009), and also with other parameters: dissimilar weather conditions over the reservoir and the pan; the position; the advection effect; the size and depth of the reservoir; the quality of the water in the reservoir; etc.

We found no comparison between actual measurement of pan and lake evaporation in the literature, so the only C_C values reported in papers must be based on theoretical calculations, with their associated uncertainties. For instance Témez Peláez (2007) proposed values for 38 Class A pans in Spain using the Penman equation, ranging from 0.7 in southern Spain (semi-arid) to more than 0.8 in the cooler and wetter north (C_C ranging from 0.8 to 0.95 for a Colorado pan). Brunel and Bouron (1992) proposed C_C for a Colorado pan for 33 western and central African lakes based on lake water balance, according to a climatic gradient from sub-desert (0.65 to 0.70), Sahelian (0.70 to 0.80) to equatorial climates (0.85 to 0.95). Working on Logan's dam in Australia, McJannet et al. (2013b) compared actual evaporation from the dam using scintillometry with evaporation of a class-A pan located 54 km away from the site over a period of 18 months. They obtained only 5% difference in raw pan evaporation and measured evaporation over the entire 18 month study period, which could be expressed in a $C_{c\,\text{class A}}$ close to 1 and $C_{C\,\text{Colorado}}$ close to 1.16. This result appears to be about 20% higher than previous studies cited in McJannet et al. (2013b) and underlined the combined problems of high temporal variability and the weakness of spatial transposition.

Concerning Kairouan climate features (57% mean relative humidity, and a mean wind speed of 1.5 m s^{-1}), the FAO guidelines provides a correction factor between a Colorado pan and reference evapotranspiration (ET_0) of 0.6 in an arid climate (Allen et al., 1998). ET_0 is generally lower than evaporation from open water. Riou (1975) proposed a factor of 0.8 between ET_0 and evaporation (ET_0 = 0.8 * Evaporation) for three sites in central Africa. The C_C for open water evaporation would then be 0.75.

Our annual mean value of 0.8 is therefore close to the FAO56 value, and in the same order of magnitude as values proposed for a similar climate in southern Spain (0.8) or in the Sahel (0.7–0.8). However, in addition, our study provides an estimation of the confidence interval and temporal variation of C_C.

At the seasonal scale, our C_c reached minimum in the hot dry summer and maximum in spring. This differs slightly from several studies in very different climates: Riou (1975) and Fu et al. (2004) showed that in equatorial and temperate climates, C_C reached maximum in the cold wet season and minimum in the hot dry season. Monthly values of C_C obtained by McJannet et al. (2013b) ranged from 0.67 to 1.17 and the minimum was reached in the coolest month (July) in a tropical sub-humid climate.

Seasonal heat storage does not appear to be the main factor involved in our study since variation in the monthly amount of heat storage (G) was in the same order of magnitude in the reservoir and in the pan and both BREB and Penman results showed the same trend. Advection effects, which are particularly significant in arid environments, could be responsible for this seasonal variation (Oroud, 1998). The difference between advection effects on the reservoir and on the pan is hard to quantify, but can vary considerably depending on the properties of the direct environment.

8. Conclusion

Evaporation is a major component of the hydrological budget of water bodies in semi-arid areas. Because direct measurements are difficult, they are rare and often replaced by indirect methods. We compared the results of different physical, chemical and isotopic approaches and assessed Colorado pan coefficient values, as pan measurements are the most frequent information. Calculations were performed to assess the sensitivity of different parameters and the uncertainty in the indirect estimation of evaporation, based on the El Haouareb dam.

In this part of Central Tunisia where annual precipitation is about 300 mm, annual evaporation has varied by more than 25% over the last 20 years. Four different approaches and three years calculation gave results ranging from 1400 to 1900 mm. The agreement between physical and isotopic methods led to an estimation of the annual evaporation of 1700 mm for the period 1989–2011. The annual Colorado pan conversion coefficient was 0.8 and varied from 1.0 in spring to 0.76 for the rest of the year.

Rarely considered, the scarcity and the representativeness of data for estimating evaporation were tested. An error of 10% in relative humidity can lead to 40% uncertainty in evaporation values calculated with the Dalton equation. Removing 40% of the daily input data led to an additional uncertainty of about 20% using the Dalton and BREB equations. We finally estimated the overall uncertainty in evaporation values around 15%. The Penman approach appears the most robust. It gave an annual evaporation rate of 1600 mm for the period 2003–2006.

Our comparative field study of widely used methods to estimate evaporation aimed to provide practical guidelines for current applications. We developed a method to profitably exploit each data set by combining several independent methods, resulting in significant improvements. Implemented in a typical semi-arid context (i.e. involving data scarcity and high variability of climate and hydrological processes), this method provides valuable information for other works in similar data-poor systems.

Conflicts of interest

The authors declare that there are no conflicts of interest.

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