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SUPPLEMENTARY MATERIAL

Simulation of the hydrodynamic behaviour of a mediterranean reservoir under different climate change and management scenarios

Jordi PRATS,[1*] Marie-José SALENÇON,[2] Magali GANT,[2] Pierre-Alain DANIS[3]

[1]Irstea, UR RECOVER, HYNES (Irstea-EDF R&D), Centre d'Aix-en-Provence, 3275 Route Cézanne, 13182 Aix-en-Provence
[2]EDF R&D, HYNES (Irstea-EDF R&D) Laboratoire National d'Hydraulique et Environnement, 6 quai Watier, BP 49, 78401 Chatou
[3]Agence Française pour la Biodiversité, Pôle AFB-Irstea Hydroécologie des plans d'eau, 3275 Route Cézanne, 13182 Aix-en-Provence, France

[*]**Corresponding author**: jordi.prats@irstea.fr

INTRODUCTION

The 1D hydrodynamic model EOLE was developed by the Department of R&D of the French electrical enterprise EdF (Électricité de France) for its application to the enterprise's hydroelectric reservoirs. The first version of the model dates from the 1980s (Enderlé, 1980) and it was described in detail in Salençon and Thébault (1997). EOLE was tested in lakes Pareloup and Rochebout (Salençon, 1994, 1997; Salençon and Thébault, 1997) and has been used by EdF R&D since. In this study we used the latest version of EOLE v.19 (Gant, 2013). Since some of the documents describing the model are of difficult access and/or in French, we describe its main features herein.

DESCRIPTION OF EOLE

EOLE is a one-dimensional hydrodynamic model that considers the most important physical processes that affect the vertical density structure of lakes and reservoirs: exchanges of heat, mass and momentum at the water-air interface, surface and bottom water mixing, and inflow and outflow dynamics. EOLE simulates the temperature in the water column and can be coupled to the water quality model ASTER (Salençon and Thébault, 1996, 1997).

Resolution of the flow equations

The hydrostatic approximation is used (Simons, 1980), together with the fractional step method (Yanenko, 1971), to solve the fluid equations in terms of the resolution of a set of three equations, including a horizontal submodel:

$$\frac{\partial AW}{\partial z} = B\left(U_{in} - U_{out}\right), \tag{1}$$

a vertical submodel:

$$\frac{1}{A}\frac{\partial AT}{\partial t} - B\left(U_{in}T_{in} - U_{out}T_{out}\right) + W\frac{\partial T}{\partial z} = -\frac{1}{A}\frac{\partial}{\partial z}\left(A\overline{T'W'}\right) - \frac{1}{\rho_0 C_p}\frac{\partial Q}{\partial z} \tag{2}$$

and the equation of state (Markofsky and Harleman, 1971):

$$\rho = \rho_0\left[1 - \alpha\left(T - T_0\right)^2\right] \tag{3}$$

In the equations above A is surface area (m^2), W is vertical speed (m s^{-1}), B is the lake width (m), U_{in} is the inflow speed (m s^{-1}), U_{out} is the outflow speed (m s^{-1}), T is the water temperature (ºC), Q is the vertical thermal flux (solar radiation, W m^{-2}), C_p is the specific heat of water ($C_p = 1005$ J kg^{-1} ºC^{-1}), ρ_0 is the maximal density of liquid water at a normal atmospheric pressure ($\rho_0 = 1000$ kg m^{-3}), T_0 is the temperature at which the water density is maximal ($T_0 = 4$ºC) and α is an empirical constant ($\alpha = 6.63 \cdot 10^{-6}$ ºC^{-2}).

EOLE uses a fixed 3-hour time step and a Eulerian finite volume discretization with a fixed layer thickness of 0.25 m, except for the surface layer, with a variable thickness.

Water balance

In EOLE the water balance is calculated as a volume balance:

$$\Delta V = \left(Q_{\inf} - Q_{out} - \frac{H_e A_N}{L_V \rho_N}\right)\Delta t + A_N r_k \tag{4}$$

where ΔV is the variation of volume of the water body (m^3 s^{-1}), Q_{inf} is the sum of the discharges of all the inflows (m^3 s^{-1}), Q_{out} is the sum of the discharges of all the outflows (m^3 s^{-1}), H_e is evaporative heat flux (W m^{-2}), A_N is the area of the surface layer of the reservoir, ρ_N is the density of the surface layer water, r_k is the precipitation in the k-th time step (mm), and L_V is the latent heat of water vaporization (J kg^{-1}):

$$L_V\left(T_w\right) = 1000\left(2500.9 - 2.365T_w\right) \tag{5}$$

Surface heat energy balance

The surface heat energy balance H_T (W·m^{-2}) is calculated as:

$$H_T = H_{sn} + H_{lwn} + H_e + H_c \quad (6)$$

where H_{sn} is the net solar radiation (W m^{-2}), H_{lwn} is the net longwave radiation (W m^{-2}), H_e is the latent heat flux (W m^{-2}) and H_c is the sensible heat flux to the atmosphere (W m^{-2}).

The net solar radiation is an input for EOLE.

EOLE calculates net longwave radiation exchange H_{lwn} as the difference between net atmospheric longwave radiation H_{an} (W m^{-2}) and emitted radiation H_w (W m^{-2}).

$$H_{lwn} = H_{an} + H_w \quad (7)$$

The model calculates longwave atmospheric radiation as a function of air temperature T_a (ºC) and cloud cover C (-) using a variation of the formula of Wunderlich (1972):

$$H_{an} = (1-\beta)\varepsilon_a \sigma (T_a + 273)^4 (1 + kC^2) \quad (8)$$

where k = 0.20 is a coefficient that depends on the type of clouds, $\beta = 0.03$ is longwave radiation albedo, $\sigma = 5.67 \times 10^{-8}$ W m^{-2} ºC^{-4} is the Stefan-Boltzmann constant, and atmospheric emissivity ε_a is calculated as (Swinbank, 1963):

$$\varepsilon_a = 0.937 \cdot 10^{-5} (T_a + 273)^2 \quad (9)$$

EOLE calculates the emitted radiation as:

$$H_w = -\varepsilon_w \sigma (T + 273)^4 \quad (10)$$

where ε_w = 0.97 is the emissivity of water.

EOLE requires wind speed at 10 m (U_{10} in m s^{-1}), air temperature (T_a in ºC), atmospheric pressure (P_a in hPa) and relative humidity (HR in %) as input and calculates the latent heat flux H_e using a mass transfer equation (Singh and Xu, 1997) of the type:

$$H_e = L_E \rho_a f_{EO}(U_2)(q_w - q_a) \quad (11)$$

where U_2 is the wind speed at 2 m above the water surface (m s^{-1}), $L_E = (2500 - 2.36T_w) \cdot 10^3$(J kg^{-1}) is the latent heat of evaporation, and $f_{EO}(U_2) = a + bU_2$ is a wind transfer function where a and b are calibration parameters. Suggested values are $0.0017 \le a \le 0.0035$ and b = 1 (Salençon and Thébault, 1997). The suggested value of a is in accordance with values obtained by Hondzo and Stefan (1993), while the suggested value of b is in accordance with the results obtained by Singh and Xu (1997). The density of air ρ_a (kg m^{-3}) is calculated as:

$$\rho_a = 1.293 + T_a(1.2045 - 1.293)/20 \quad (12)$$

The wind speed at 2 m, U_2, is calculated from the wind speed at 10 m above the water surface (U_{10}) by assuming a logarithmic wind profile so that $U_2 = 0.6U_{10}$.

EOLE calculates the sensible heat flux as:

$$H_s = C_p \rho_a f_{EO}(U_2)(T - T_a) \quad (13)$$

Radiation transfer

The transfer of shortwave radiation into the water body is calculated using the Beer-Lambert law (Henderson-Sellers, 1986):

$$Q(z) = H_{sn} \exp(-\eta z) \quad (14)$$

where z is the depth (m) and η is the light extinction coefficient (m^{-1}), calculated from daily input values of Secchi depth as:

$$\eta = \frac{1.7}{SD} \quad (15)$$

Surface mechanic energy exchange

Mechanic energy is exchanged between the atmosphere and the water surface through friction caused by wind. Shear velocity ($m\ s^{-1}$) is calculated as:

$$u_* = \left(\frac{C_D \rho_a}{\rho_{ws}}\right)^{1/2} U_{10} \quad (16)$$

where ρ_{ws} is surface layer density. Shear stress is calculated as:

$$\tau_a = \rho_a C_D U_{10}^2 \quad (17)$$

where the aerodynamic drag coefficient is usually taken as $C_D = 1.3 \cdot 10^{-3}$. The increase of moment caused by wind stress in the surface mixed layer is calculated as:

$$U_N(t + \Delta t) - U_N(t) = \begin{cases} \dfrac{u_*^2 \Delta t}{h} & t \leq T_c \\ 0 & t > T_c \end{cases} \quad (18)$$

where U_N is the velocity in the surface layer ($m\ s^{-1}$), t is time (s), Δt is the time increment (s), h is the thermocline depth (m), and T_c is the shear period (s). The thermocline depth is calculated according to (Patterson *et al.*, 1984):

$$h = H - \frac{\int_0^H z \frac{d\rho}{dz} dz}{\int_0^H \frac{d\rho}{dz} dz} \quad (19)$$

where z is the level above the bottom (m) and H is the maximum depth (m). The shear period is defined as $T_c = T_i/4$ (s) where T_i is the internal seiche period, calculated as:

$$T_i = \frac{2L}{\sqrt{g \frac{\Delta\rho}{\rho_h} \frac{h(H-h)}{H}}} \quad (20)$$

where L is the fetch (m), ρ_h is hypolimnion density (kg m^{-3}), $\Delta\rho$ is the difference between hypolimnon density and epilimnion density (kg m^{-3}), and g is the acceleration of gravity (m s^{-2}).

Mixing processes

EOLE's turbulence model in the surface mixed layer is based on the hypothesis of Niiler and Kraus (1977). The turbulent kinetic energy balance is stated as:

$$gh \frac{\Delta\rho}{\rho_0} \frac{dh}{dt} = nw_*^3 + 2mu_*^3 f(Ri) \quad (21)$$

where $\rho_0 = 10^3$ kg m^{-3} is a reference density, $m = 0.25$ and $n = 0.124$ are mixing coefficients, w_* is the turbulent velocity scale due to convective overturn (m s^{-1}), and Ri is the Richardson number:

$$Ri = \frac{\Delta\rho g h}{\rho_0 u_*^2} \quad (22)$$

Equation 21 equates the variation of potential energy in the surface layer (left) with the turbulent kinetic energy at the surface created by convective processes (first right term) and transmitted by the wind (second right term). The function $f(Ri)$ takes into account dissipative effects (Bloss and Harleman, 1979; Salençon, 1997; Salençon and Thébault, 1997) and is defined as:

$$f(Ri) = \begin{cases} 0.057 Ri \frac{29.5 - Ri^{1/2}}{14.2 + Ri} & W > 10 \\ \frac{Ri}{14.2 + Ri} & 3 \le W \le 10 \end{cases} \quad (23)$$

where $W = hRi/L$ is the Wedderburn number.

Hypolimnion mixing is parameterised as an eddy diffusion process:

$$\frac{\partial}{\partial z}\left(\overline{T'W'}\right)=\frac{\partial}{\partial z}\left[\left(K_m+K_z\right)\frac{\partial T}{\partial z}\right] \tag{24}$$

where K_m is the coefficient of molecular diffusion and K_z is a constant turbulent diffusion coefficient.

EOLE uses the inflow algorithm proposed by Ryan and Harleman (1971) and assumes that the incoming flows are distributed vertically around the entry level z_{in} according to a Gaussian profile:

$$U_{in}(z)=U_{in,\max}\exp\left(-\frac{\left(z-z_{in}\right)^2}{2\sigma_{in}^2}\right) \tag{25}$$

where U_{in} is the horizontal speed of the inflow (m s^{-1}) at depth z, $U_{in,max}$ is the maximum U_{in} (m s^{-1}), with:

$$\sigma_{in}=\frac{\delta_{in}/2}{1.96} \tag{26}$$

and

$$\delta_{in}=\begin{cases} ch_{in} & \text{surface inflow} \\ h_{in} & \text{subsurface inflow} \end{cases} \tag{27}$$

where c is a calibration parameter between 1 and 1.5 to take entrainment into account, and h_{in} is the characteristic dimension of the inflow. For surface inflows, the entry level z_{in} is the neutral buoyancy level of incoming water. For subsurface inflows, z_{in} is the depth of the inflow. Then, the vertical density profile is stabilised using a convective adjustment algorithm.

Similarly (Ryan and Harleman, 1971), the outflowing water is extracted around the level z_{out} (m) of the outlet according to a Gaussian profile

$$U_{out}(z)=U_{out,\max}\exp\left(-\frac{\left(z-z_{out}\right)^2}{2\sigma_{out}^2}\right) \tag{28}$$

where U_{out} is the horizontal speed of the inflow (m s^{-1}) at depth z, $U_{out,max}$ is the maximum U_{out} (m s^{-1}), and:

$$\sigma_{out} = \frac{\delta_{out}/2}{1.96} \tag{29}$$

where δ_{out} is the characteristic dimension of the outlet.

SENSITIVITY ANALYSIS AND CALIBRATION

Sensitivity analysis

As explained in the main paper, correction coefficients were applied to the meteorological input:

$$T_{a,lake} = T_a + \delta + C_{Ta} \tag{30}$$

$$HS_{lake} = C_{HS} * HS \quad C_{HS} \in [0,1] \tag{31}$$

$$W_{lake} = C_{1,W} + C_{2,W} W \tag{32}$$

$$T_{ni,lake} = T_{ni} + C_{Tni} \tag{33}$$

$$T_{ai,lake} = T_{ai} + C_{Tai} \tag{34}$$

The variables T_a, HS, W, T_{ni} and T_{ai} are, respectively, measured air temperature, solar radiation, wind speed, water temperature of the natural inflow and water temperature of the artificial inflow. $T_{a,lake}$, HS_{lake}, W_{lake}, $T_{ni,lake}$ and $T_{ai,lake}$ are, respectively, air temperature, solar radiation, wind speed, temperature of the natural inflow and temperature of the artificial inflow applied at the lake. C_{Ta}, C_{HS}, $C_{1,W}$, $C_{2,W}$, C_{Tni} and C_{Tai} are correction coefficients and δ=-0.94 ºC is the adiabatic correction.

We carried a local sensitivity analysis to determine the parameters to which EOLE was most sensitive, using data for 2010-2011. The analysis consisted in the modification of the value of the calibration parameters once at a time. The effect of the variation of the value of the parameters was tested on 9 different levels in pre-established ranges. The parameter values of the reference simulation and their ranges of variation are shown in the Tab. S1. A performance indicator was used to compare the different combinations of parameter values. The performance indicator (*mRMSE*) was the average of the volume weighted RMSE of the profiles simulated on the dates when *in situ* data were available:

$$mRMSE = \frac{1}{m}\sum_{i=1}^{m} RMSE_i \tag{35}$$

where m is the number of profiles and $RMSE_i$ is the volume weighted RMSE for the i-th profile:

$$RMSE_i = \left[\frac{\sum_{j=1}^{n_i} V_j \left(T_{s,i,j} - T_{o,i,j} \right)^2}{\sum_{j=1}^{n_i} V_j} \right]^{1/2} \tag{36}$$

where V_j is the volume of the water layer between depths z_j and $z_{j\text{-}1}$, $T_{s,i,j}$ is the simulated water temperature of the i-th profile at the z_j depth, $T_{s,i,j}$ is the observed water temperature of the i-th profile at the z_j depth, and n_i is the number of temperature measurements of the profile.

Tab. S1. List of parameters: reference values and ranges used in the sensitivity analysis and calibration values.

Parameters (units)	Symbols	Reference values	Min.	Max.	Calibrated values
Air temperature correction (ºC)	C_{Ta}	0	-2	2	0.5
Shading coefficient (-)	C_{HS}	1	0.5	1	0.8
Wind corr. coeff. 1 (m s^{-1})	$C_{1,W}$	0	0	2	1.0
Wind corr. coeff. 2 (-)	$C_{2,W}$	1	0.5	3.5	0.81
Nat. inflow temp. corr. (ºC)	C_{Tni}	0	-2	2	0.0
Art. inflow temp. corr. (ºC)	C_{Tai}	0	-2	2	2.5
Depth of infiltration (m)	z_{inf}	275	275	310	285.8
Aerodynamic drag coefficient (-)	C_D	$1.3 \cdot 10^{-3}$	$1 \cdot 10^{-3}$	$3 \cdot 10^{-3}$	$1.3 \cdot 10^{-3}$
Hypolimnetic diffusion (m s^{-2})	K_z	$7 \cdot 10^{-7}$	$1 \cdot 10^{-7}$	$1 \cdot 10^{-5}$	$1 \cdot 10^{-6}$
Stream depth 1 (m) – La Cause	$h_{in,1}$	1	0.5	2	1
Stream depth 2 (m) – Surface artificial inflow	$h_{in,2}$	1	0.5	2	1
Equivalent depth 1 (m)	$c_{in,1}$	1.2	1	1.5	1.2
Equivalent depth 2 (m)	$c_{in,2}$	1.2	1	1.5	1.2
Characteristic dim. outlet 1 (m)	$\delta_{out,1}$	2	0.5	2	2
Characteristic dim. outlet 2 (m)	$\delta_{out,2}$	3	0.5	5	3

The results of the local sensitivity analysis (Fig. S1) showed that the model was most sensitive to the correction coefficients of the meteorological input, especially of wind speed and air temperature. EOLE was also very sensitive to the value of C_D. Since this parameter is correlated with the correction coefficients of wind speed, we kept the default value of $C_D = 1.3 \cdot 10^{-3}$ and adjusted $C_{1,W}$ and $C_{2,W}$ in the calibration process. In a lesser degree, the model was also sensitive to the value of the correction coefficient of the temperature of the artificial inflow, of the depth of infiltration and of the hypolimnetic diffusion coefficient. The hydromorphological parameters of the inflows and the outflows did not have an important effect.

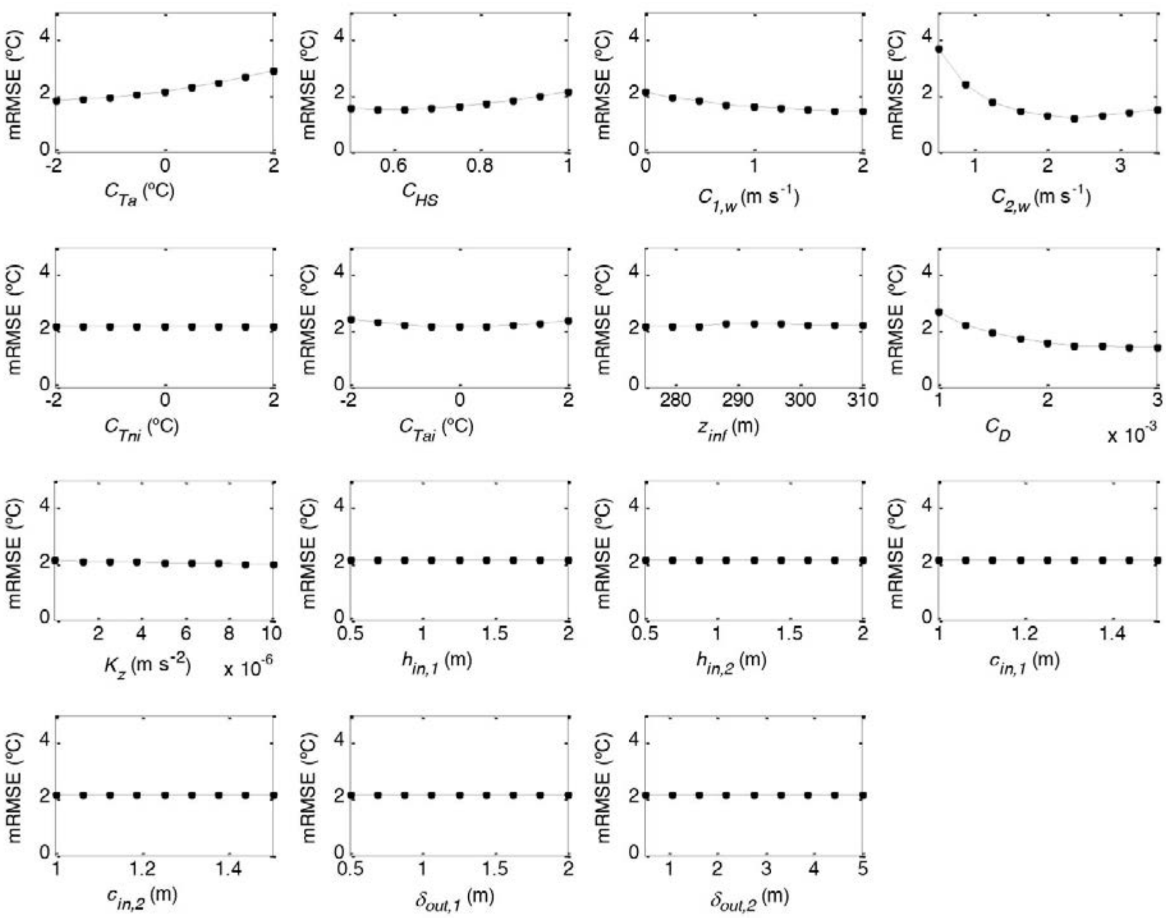


Fig. S1. Results of the local sensitivity analysis. See Tab. 1 for the meaning of the symbols used.

Calibration and validation

Single parameter sensitivity analyses were complemented by logical/physical reasoning to obtain a set of manually calibrated parameter values by trying to minimize *mRMSE*. We calibrated the model using data for 2010-2011 and we validated it for January 2012 - August 2014.

We show simulation results of the calibrated EOLE model in the Fig. S2 for the calibration period and in Fig. S3 for the validation period. The quality of the simulations of the thermal behaviour of Bimont reservoir was good: the RMSE was 1.32ºC for the calibration period and 1.08ºC for the validation period; and mean bias was -0.38ºC in the calibration period and -0.11ºC in the validation period.

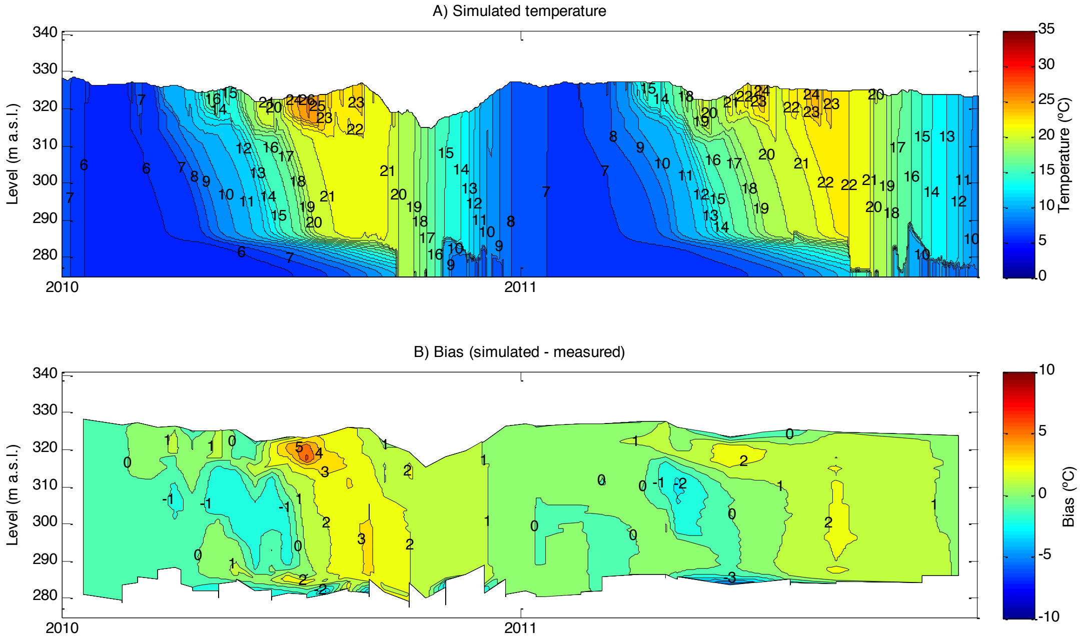
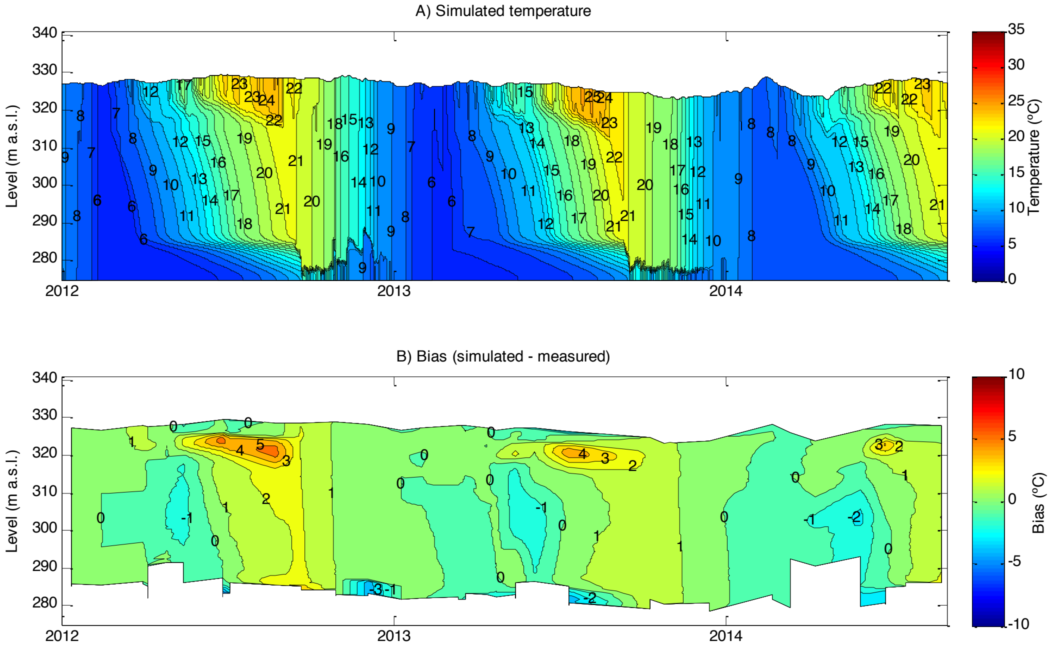


Fig. S2. Simulated water temperature at the reservoir of Bimont (A) and bias (B) during the calibration period.

However, some systematic errors appeared in the simulations. EOLE tended to predict an epilimnion deeper than observed, overestimating water temperature between 5 m and 10 m, especially in the validation period (Fig. S3). In addition, a seasonal pattern appeared, with an underestimation of water temperature at depths of 10-30 m in the winter and spring, and an overestimation during the rest of the year (Figs. S2 and S3). Other authors have observed similar behaviours resulting in more important underestimations of the water temperature in the mixed-water period than in the stratification period for other 1D hydrodynamic models (Stepanenko *et al.*, 2013).

Fig. S3. Simulated water temperature at the reservoir of Bimont (A) and bias (B) during the validation period.



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