**Supplementary material**

The blue water footprint of electricity from hydropower

Mesfin M. Mekonnen[a,b] and Arjen Y. Hoekstra[a]

[a]Department of Water Engineering and Management, University of Twente, P.O. Box 217, 7500 AE Enschede, The Netherlands

[b] corresponding author: m.m.mekonnen@ctw.utwente.nl

## Method

The water footprint of electricity ($WF$, m$^3$/GJ) generated from hydropower is calculated by dividing the amount of water evaporated from the reservoir annually ($WE$, m$^3$/yr) by the amount of energy generated ($EG$, GJ/yr):

$$WF = \frac{WE}{EG} \tag{1}$$

The total volume of evaporated water ($WE$, m$^3$/yr) from the hydropower reservoir over the year is:

$$WE = \left(10 \times \sum_{t=1}^{365} E\right) \times A \tag{2}$$

where $E$ is the daily evaporation (mm/day) and $A$ the area of the reservoir (ha).

There are a number of methods for the measurement or estimation of evaporation. These methods can be grouped into several categories including (Singh and Xu, 1997): (i) empirical, (ii) water budget, (iii) energy budget, (iv) mass transfer and (v) a combination of the previous methods.

Empirical methods relate pan evaporation, actual lake evaporation or lysimeter measurements to meteorological factors using regression analyses. The weakness of these empirical methods is that they have a limited range of applicability. The water budget methods are simple and can potentially provide a more reliable estimate of evaporation, as long as each water budget component is accurately measured. However, owing to difficulties in measuring some of the variables such as the seepage rate in a water system the water budget methods rarely produce reliable results in practice (Lenters et al., 2005, Singh and Xu, 1997). In the energy budget method, the evaporation from a water body is estimated as the difference between energy inputs and outputs measured at a site. Energy budget methods are considered to be the most reliable in theory (Lenters et al., 2005, Singh and Xu, 1997), but require costly instrumentation and a large commitment of personnel for field work and data processing (Winter et al., 1995). The mass-transfer (aerodynamic) based methods utilize the concept of eddy motion transfer of water vapour from an evaporating surface to the atmosphere. The mass-transfer methods normally use easily measurable variables and give satisfactory results in many cases. However, measurement of wind speed and air temperature at inconsistent heights, have resulted in a large number of equations with similar or identical structure (Singh and Xu, 1997). The combination methods combine the mass transfer and energy budget principles in a single equation. Two of the most commonly known combination methods are the Penman equation and the Penman-Monteith equation.

Owing to its limited empirical basis, the Penman-Monteith equation is more readily applicable to a variety of water bodies. In addition, the model takes into account heat storage within water bodies. Therefore, for the purpose of the current study the Penman-Monteith equation with heat storage is considered suitable for the estimation of evaporation from the selected hydropower reservoirs.

The evaporation from the water surface ($E$, mm/day) is estimated using the Penman-Monteith equation with an inclusion of water body heat storage. This equation is written as (McJannet et al., 2008):

$$E = \frac{1}{\lambda} \times \left( \frac{\Delta_w \times (R_n - G) + \gamma \times f(u) \times (e_w - e_a)}{\Delta_w + \gamma} \right) \tag{3}$$

where $E$ is open water evaporation (mm/day); $\lambda$ the latent heat of vaporization (MJ/kg); $\Delta_w$ the slope of the temperature saturation water vapour curve at water temperature (kPa/$^o$C); $R_n$ net radiation (MJ m$^{-2}$day$^{-1}$); $G$ the change in heat storage in the water body (MJ/m$^2$/day); $f(u)$ the wind function (MJ/m$^2$/day/kPa); $e_w$ the saturated vapour pressure at water temperature (kPa); $e_a$ the vapour pressure at air temperature (kPa); and $\gamma$ the psychrometric constant (kPa/$^o$C).

The latent heat of vaporisation ($\lambda$, MJ/kg) at air temperature ($T_a$, $^o$C) is calculated as (McJannet et al., 2008):

$$\lambda = 2.501 - 2.361 \times 10^{-3} T_a \tag{4}$$

The psychrometric constant ($\gamma$, kPa/$^o$C) is calculated from (Allen et al., 1998):

$$\gamma = \frac{c_p \times P}{\varepsilon \times \lambda} = \frac{1.63 \times 10^{-3} P}{\lambda} \tag{5}$$

in which $P$ is the atmospheric pressure (kPa); $c_p$ the specific heat of air at constant pressure (which is equal to 1.013x10$^{-3}$ MJ/kg/$^o$C) and $\varepsilon$ the ratio of molecular weight of water vapour to dry air and is equal to 0.622 (dimensionless).

The atmospheric pressure ($P$, kPa) varies with elevation above sea level ($\psi$, m) and is expressed as (Allen et al., 1998):

$$P = 101.3 \times \left( \frac{293 - 0.0065\psi}{293} \right)^{5.26} \tag{6}$$

The wind function $f(u)$ (MJ/m$^2$/day/kPa) is calculated from wind speed at 10 m ($u_{10}$, m/s) and the so-called equivalent area ($A_e$, km$^2$) (Sweers, 1976):

$$f(u) = \left(\frac{5}{A_e}\right)^{0.05} \times (3.80 + 1.57 u_{10}) \tag{7}$$

The equivalent area ($A_e$, km$^2$) is equal to the total surface area for regularly shaped reservoirs, but for irregularly shaped reservoirs, it can be taken equal to the square of the mean width.

Saturated vapour pressure at air temperature ($e_a$, kPa) is calculated from:

$$e_a = 0.6108 \times \exp\left(\frac{17.27 T_a}{(T_a + 237.3)}\right) \tag{8}$$

Net radiation ($R_n$, MJ m$^{-2}$ d$^{-1}$) is the difference between the net incoming short-wave radiation ($R_{ns}$, MJ m$^{-2}$ d$^{-1}$) and the net outgoing long-wave radiation ($R_{nl}$, MJ/m$^2$/day) (Allen et al., 1998):

$$R_n = R_{ns} - R_{nl} \tag{9}$$

The net incoming short-wave radiation ($R_{ns}$, MJ/m$^2$/day) resulting from the balance between incoming and reflected solar radiation is given by (Allen et al., 1998):

$$R_{ns} = (1 - \alpha) \times R_s \tag{10}$$

where $\alpha$ is the albedo coefficient for open water (dimensionless), which has a value of 0.07 (Lenters et al., 2005), and $R_s$ the incoming solar radiation (MJ/m$^2$/day).

Solar radiation ($R_s$, MJ m$^{-2}$ day$^{-1}$) can be calculated with the Angstrom formula, which relates solar radiation to extraterrestrial radiation and relative sunshine duration:

$$R_s = (a_s + b_s \times \frac{n}{N}) \times R_a \tag{11}$$

where $n$ is the actual duration of sunshine (hours); $N$ the maximum possible duration of sunshine or daylight hours (hours); $n/N$ the relative sunshine duration (which is equal to one minus the cloud cover fraction, dimensionless); $R_a$ extraterrestrial radiation (MJ/m$^2$ /day); $a_s$ a regression constant, expressing the fraction of extraterrestrial radiation reaching the earth on overcast days (n = 0) and $a_s$+$b_s$ the fraction of extraterrestrial radiation reaching the earth on clear days (when n = N).

Depending on atmospheric conditions (humidity, dust) and solar declination (latitude and month), the Angstrom values $a_s$ and $b_s$ will vary. Where no actual solar radiation data are available and no calibration has been carried out for improved $a_s$ and $b_s$ parameters, the values $a_s$ = 0.25 and $b_s$ = 0.50 are taken as recommended by Allen et al. (1998).

The extraterrestrial radiation, $R_a$, for each day of the year and for different latitudes, can be estimated from the solar constant, the solar declination and the time of the year.

$$R_a = \frac{24\times 60}{\pi} G_{sc} \times d_r \left[\omega_s \times \sin(\varphi) \times \sin(\delta) + \cos(\varphi) \times \cos(\delta) \times \sin(\omega_s)\right] \quad (12)$$

where $G_{sc}$ is the solar constant (which is equal to 0.0820 MJ/m$^2$/day); $d_r$ the inverse relative distance Earth-Sun; $\omega_s$ the sunset hour angle (rad); $\varphi$ the latitude (rad) and $\delta$ the solar decimation (rad).

The inverse relative distance Earth-Sun, $d_r$, and the solar declination, $\delta$, are given by:

$$d_r = 1 + 0.033 \cos\left(\frac{2\pi}{365} \times J\right) \quad (13)$$

$$\delta = 0.409 \sin\left(\frac{2\pi}{365} \times J - 1.39\right) \quad (14)$$

where $J$ is the number of the day in the year between 1 (1 January) and 365 or 366 (31 December). The latitude $\varphi$, expressed in radians, is positive for the northern hemisphere and negative for the southern hemisphere.

The sunset hour angle, $\omega_s$, is given by:

$$\omega_s = arccos[-tan(\varphi) \times tan(\delta)] \quad (15)$$

The net outgoing long-wave radiation ($R_{nl}$, MJ/m$^2$/day) is the difference between the outgoing long-wave radiation ($R_l\uparrow$, MJ/m$^2$/day) and the incoming long-wave radiation ($R_l\downarrow$, MJ m$^{-2}$ d$^{-1}$):

$$R_{nl} = R_l \uparrow - R_l \downarrow \quad (16)$$

The incoming long-wave radiation ($R_l\downarrow$, MJ/m$^2$/day) is calculated from (Fischer et al., 1979; Henderson-Sellers, 1986):

$$R_l \downarrow = \varepsilon_a \times \sigma \times (T_a + 273.15)^4 (1 + 0.17 C_f^2)(1 - r_{lw}) \quad (17)$$

where $\varepsilon_a$ is the emissivity of air (dimensionless); $\sigma$ the Stefan-Boltzmann constant (4.903x10$^{-9}$ MJ/K$^4$/m$^2$/day); $C_f$ the fractional cloud cover (dimensionless); and $r_{lw}$ the total reflectivity of the water surface for long wave radiation, taken as a constant with a value of 0.03 (Henderson-Sellers, 1986).

The emissivity of air is calculated as (Swinbank, 1963):

$$\varepsilon_a = C_\varepsilon \times (T_a + 273.15)^2 \tag{18}$$

where $C_\varepsilon = 9.37\times10^{-6}$ K$^{-2}$.

The outgoing long-wave radiation at water temperature ($R_l\uparrow$, MJ/m$^2$/day) is calculated as (Sellers, 1986):

$$R_l \uparrow = \varepsilon_w \times \sigma \times (T_w + 273.15)^4 \tag{19}$$

where $\sigma$ is the Stefan-Boltzmann constant (MJ/m$^2$/K$^4$/day); $T_w$ the water surface temperature ($^o$C); and $\varepsilon_w$ the emissivity of water, equal to 0.97.

The water temperature at day $i$ ($T_{wi}$, $^o$C) is calculated from the following equation (De Bruin, 1982):

$$T_{w,i} = T_e + (T_{w,i-1} - T_e) \times exp(-1/\tau) \tag{20}$$

where $T_{w,i-1}$ is the water temperature at day $i$-1 ($^o$C); $T_e$ the equilibrium temperature ($^o$C); and $\tau$ the time constant (day).

The equilibrium temperature ($T_e$, $^o$C) is calculated as follows (De Bruin, 1982):

$$T_e = T_n + \frac{R_n^*}{4\sigma \times (T_n + 273.15)^3 + f(u) \times (\Delta_n + \gamma)} \tag{21}$$

Wet-bulb temperature ($T_n$, $^o$C) is calculated using vapour pressure ($e_a$, kPa) and dew point temperature ($T_d$, $^o$C) as follows (McJannet et al., 2008):

$$T_n = \frac{0.00066 \times 100 T_a + \left(4098 e_a / (T_d + 237.3)^2\right) \times T_d}{0.00066 \times 100 + \left(4098 e_a / (T_d + 237.3)^2\right)} \tag{22}$$

The slope of the temperature saturation water vapour curve at wet bulb temperature ($\Delta_n$, kPa/K) is:

$$\Delta_n = \frac{4098 \times \left[ 0.6108 \times exp\left( \frac{17.27T_n}{(T_n + 237.3)} \right) \right]}{(T_n + 237.3)^2} \tag{23}$$

Net radiation at wet-bulb temperature ( $R_n^*$ , MJ/m$^2$/day) is calculated using albedo ($\alpha$) as follows:

$$R_n^* = (1-\alpha)\times R_s + \left(R_l \downarrow - R_l \uparrow_n\right) \tag{24}$$

Outgoing long-wave radiation at wet-bulb temperature ($R_l\uparrow_n$, MJ/m$^2$/day) is calculated, based on Finch and Gash (2002):

$$R_l \uparrow_n = C_f \times \left(\sigma \times (T_a + 273.15)^4 + 4\sigma \times (T_a + 273.15)^3 \times (T_n - T_a)\right) \tag{25}$$

where $C_f$ is fractional cloud cover.

The time constant ($\tau$, day) is given as (De Bruin, 1982):

$$\tau = \frac{\rho_w \times c_w \times h}{4\sigma \times (T_n + 273.15)^3 + f(u)\times(\Delta_n + \gamma)} \tag{26}$$

where $\rho_w$ is the density of water (= 1000 kg/m$^3$); $c_w$ the specific heat of water (= 0.0042 MJ/kg/K); and $h$ the depth of water (m), estimated from reservoir volume capacity and area.

Change in the heat storage in the water body ($G$, MJ/m$^2$/day) is calculated from Finch (2001):

$$G = \rho_w \times c_w \times h \times \left(T_{w,i} - T_{w,i-1}\right) \tag{27}$$

Saturated vapour pressure at water temperature ($e_w$, kPa) is calculated from:

$$e_w = 0.6108 \times \exp\left( \frac{17.27T_w}{(T_w + 237.3)} \right) \tag{28}$$

Finally, the slope of the temperature saturation water vapour curve at water temperature ($\Delta_w$, kPa $^oC^{-1}$) is:

$$\Delta_w = \frac{4098 \times \left[ 0.6108 \times \exp\left( \frac{17.27T_w}{(T_w + 237.3)} \right) \right]}{(T_w + 237.3)^2} \tag{29}$$

## References


Allen, R.G., Pereira, L.S., Raes, D. and Smith, M.: Crop evapotranspiration: guidelines for computing crop water requirements, FAO Drainage and Irrigation Paper 56, Food and Agriculture Organization, Rome, 1998.

Chao, B. F., Wu, Y. H. and Li, Y. S.: Impact of artificial reservoir water impoundment on global sea level, Science, 320(5873), 212-214, 2008.

De Bruin, H.A.R.: Temperature and energy balance of a water reservoir determined from standard weather data of a land station, J. Hydrol., 59(3-4), 261-274, 1982.

Finch, J. W. and Gash, J. H. C.: Application of a simple finite difference model for estimating evaporation from open water, J. Hydrol., 255(1-4), 253-259, 2002.

Finch, J. W.: A comparison between measured and modelled open water evaporation from a reservoir in south-east England, Hydrol. Process., 15(14), 2771-2778, 2001.

Fischer, H.B., List, E.J., Koh, R.C.Y., Imberger, J. and Brooks, N.H.: Mixing in inland and coastal waters, Academic Press, San Diego, 1979.

Henderson-Sellers, B.: Calculating the surface energy balance for lake and reservoir modelling: A review. Rev. Geophys. 24(3), 625-649, 1986.

Lenters, J. D., Kratz, T. K. and Bowser, C. J.: Effects of climate variability on lake evaporation: Results from a long-term energy budget study of sparkling lake, northern Wisconsin (USA), J. Hydrol., 308(1-4), 168-195, 2005.

McJannet, D.L., Webster, I.T., Stenson, M.P. and Sherman, B.S.: Estimating open water evaporation for the Murray Darling basin, Report for CSIRO, Australia, 2008.

Sellers, W. D.: Physical climatology, University of Chicago Press, Chicago, Illinois, 1965.

Shiklomanov, I. A.: Appraisal and assessment of world water resources, Water Int., 25(1), 11-32, 2000.

Singh, V. P. and Xu, C.-Y.: Evaluation and generalization of 13 mass-transfer equations for determining free water evaporation, Hydrol. Process., 11(3), 311-323, 1997.

Sweers, H. E.: A monogram to estimate the heat-exchange coefficient at the air-water interface as a function of wind speed and temperature; a critical survey of some literature, J. Hydrol., 30(4), 375-401, 1976.

Swinbank, W. C.: Long-wave radiation from clear skies, Quarterly Journal of the Royal Meteorological Society 89 (381), 339-348, 1963.

Winter, T. C., Rosenberry, D. O. and Sturrock, A. M.: Evaluation of 11 equations for determining evaporation for a small lake in the north central united states, Water Resour. Res., 31(4), 983-993, 1995.