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Impact of reservoir evaporation on future water availability in north-eastern Brazil: a multi-scenario assessment

Gláuber Pontes Rodrigues et al.

Correspondence to: Gláuber Pontes Rodrigues (pontesglauber@gmail.com)

S1. AquaSEBS overview

The Surface Energy Balance of Fresh and Saline Waters (AquaSEBS, as in Abdelrady *et al.*, 2016) is an adaptation of the SEBS model (Su, 2002) to estimate evaporation in open water. It consists of a set of tools to determine physical water surface parameters (such as albedo, emissivity, temperature etc.) from spectral reflectance and radiance. It requires three sets of data as input: (1) remote-sensing data including emissivity, surface albedo and water surface temperature; (2) meteorological data, including air pressure, air temperature, relative humidity and wind speed at a reference height; and (3) radiative forcing parameters, such as downward shortwave and long-wave radiations. The algorithm was validated in several water bodies at different environmental conditions (Abdelrady *et al.*, 2016; Losgedaragh and Rahimzadegan, 2018) including Brazilian tropical reservoirs (Rodrigues *et al.*, 2021a).

AquaSEBS uses the energy balance to calculate the instantaneous latent heat flux of evaporation (Equation 1), thus, evaporation is calculated for each pixel of the image.

$$\lambda E_{inst} = R_n - G_{0W} - H \quad \text{(S1)}$$

where λE_{inst} is latent heat flux of evaporation at imaging time (W m^{-2}), R_n is net radiation flux at the surface (W m^{-2}), G_{0W} is the water flux heat (W m^{-2}), and H the sensible heat flux to air.

Afterwards, atmospheric transmissivity is obtained, which is defined as the fraction of incident radiation that is transmitted by the atmosphere and which represents the effects of absorption and reflection occurring within the atmosphere. This effect occurs to incoming radiation and to outgoing radiation and is, thus, squared in Equation 2. The τ_{sw} includes transmissivity of both direct solar beam radiation and diffuse (scattered) radiation to the surface. The term τ_{sw} is calculated using an elevation-based relationship from Waters *et al.* (2002).

$$\tau_{sw} = 0.75 + 2 \cdot 10^{-5} \cdot DEM \quad \text{(S2)}$$

Where DEM is the Digital Elevation Model file. The albedo at the top of the atmosphere (unadjusted for atmospheric transmissivity) was computed through linear combination of the monochromatic reflectance (ρ) of the reflective bands (from 2 to 7, for Landsat 8). It is necessary to estimate the solar constant (ωλ, W m^{-2} μm^{-1}) associated with each one of the OLI reflective bands. Da Silva et al. (2016) according to the methodology proposed by Chander and Markham (2003), found the ωλ values for Landsat 8 and present on the following equation:

$$\alpha_{toa} = 0.3 \times \rho_2 + 0.277 \times \rho_3 + 0.233 \times \rho_4 + 0.143 \times \rho_5 + 0.036 \times \rho_6 + 0.012 \times \rho_7 \quad \text{(S3)}$$

The indexes in each ρ stand for the respective reflectance band. Incoming shortwave radiation (W m^{-2}) is calculated as:

$$R_{s\downarrow} = G_{sc} \cdot \cos(90° - \theta) \cdot d_r \cdot \tau_{sw} \quad \text{(S4)}$$

G_{SC} is the solar constant (1367 W m^{-2}), θ the sun elevation angle and d_r is the inverse squared relative distance between sun and earth, all in conformity with Allen *et al.* (1998). Incoming longwave radiation is the downward thermal radiation flux from the atmosphere (W m^{-2}). It is computed by means of the Stefan-Boltzmann equation:

$$R_{L\downarrow} = \varepsilon_a \cdot \sigma \cdot T_a^4 \quad \text{(S5)}$$

Where T_a is the near surface air temperature (monthly average, in K), σ is the Stefan-Boltzmann constant (5.67×10^{-8} W m^{-2} K^{-4}), and ε_a is the atmospheric emissivity (dimensionless). The following empirical equation by Bastiaanssen (1998) is used to assess ε_a:

$$\varepsilon_a = 0.85 \cdot (-\ln \tau_{sw})^{0.09} \quad \text{(S6)}$$

Water heat flux can be described as the imbalance between solar radiation, thermal radiation, sensible heat and latent heat fluxes. Remote sensing observations only obtain the skin temperature of the water; consequently, the Equilibrium Temperature Model (ETM) was used. The ETM model (Ahmad and Sultan, 1994; Edinger et al., 1968) integrates water surface temperature (T_S) and equilibrium temperature (T_e) through the thermal exchange coefficient (β) to estimate the water heat flux (G_w). In order to derive water heat flux, the following equations (Abdelrady *et al.*, 2016) should be applied:

$$G_w = \beta\,(T_e - T_S) \quad \text{(S7)}$$

$$T_e = T_D + \frac{R_{L\downarrow}}{\beta} \quad \text{(S8)}$$

$$\beta = 4.5 + 0.05T_S + (\eta + 0.47)3.3u \quad \text{(S9)}$$

$$\eta = 0.35 + 0.015T_S + 0.0012\,(T_n)^2 \quad \text{(S10)}$$

$$T_n = 0.5\,(T_S - T_D) \quad \text{(S11)}$$

where T_e is equilibrium temperature (°C), T_D the dew temperature (°C), T_n is the net rate of heat exchange, $R_{S\downarrow}$ the incoming shortwave, u is wind speed at 2m height (m s^{-1}), η represents the predicted percentage of heat loss through the surface[1]. Thus, equation S11 calculates the net rate of heat exchange as half of the difference between the surface temperature and the dew temperature.
The equation to calculate net radiation is given by:

$$R_n = (1 - \alpha)\,R_{S\downarrow} + \varepsilon \cdot R_{L\downarrow} - \varepsilon \cdot \sigma \cdot T_S^4 \quad \text{(S12)}$$

According to Su (2002), the sensible heat flux at the wet-limit is obtained as follows

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[1] In Equation S9, η represents the predicted percentage of heat loss through the surface relative to the maximum possible heat loss. This equation is known as the "Gagge formula", and it is used to estimate the percentage of heat loss through the surface based on the mean surface temperature (T_S) and the net rate of heat exchange (T_n). Thus, η does not directly represent the mean surface temperature. Instead, it's a parameter that quantifies the efficiency of heat loss through the skin under specific thermal conditions.

$$H_{wet} = \frac{\left((R_n - G_{0w}) - \frac{\rho_a C_p}{r_{ew}} \cdot \frac{e_s - e}{\gamma}\right)}{\left(1+\frac{\Delta}{\gamma}\right)} \tag{S13}$$

The term $e_s - e$ represents the vapour pressure deficit, C_p is the specific heat capacity of air (1004 J Kg^{-1} $°C^{-1}$), ρa the specific mass of air (1.184 Kg m^{-3}), γ is the psychrometric parameter (hPa $°C^{-1}$), Δ is the rate of change of saturation vapour pressure with temperature (hPa $°C^{-1}$), while r_{ew} is external resistance and uses the variables wind friction and sensible heat flux.

Remote sensing images can be used to provide evaporation maps with high spatial resolution during overpass, but they are temporarily limited to a definite time during the day. A daily stable term such as the evaporative fraction (EF) can be used together with satellite images to upscale latent heat and the evaporation rate from instantaneous to daily estimation (Waters *et al.*, 2002; Abdelrady *et al.*, 2016). Evaporative fraction is the ratio between latent heat and available energy at the water surface, as follows:

$$EF = \frac{\lambda E}{(R_n - G_w)} \tag{S14}$$

Latent heat is the energy needed for evaporation (equation 1, $\lambda E = R_n - G_{0w} - H_{wet}$). SEBS estimates the total energy used for evaporation in a day-based evaporative fraction term using Equation (Su, 2002). First, latent heat is converted to water depth in (mm) per day, then daily potential evaporation can be calculated as water depth utilising the following equation:

$$E_{daily} = 86400 \cdot EF\ (R_n - G_w) / \lambda Edaily \tag{S15}$$

E_{daily} in Equation S15 is given in water depth (mm) for each pixel in the image.

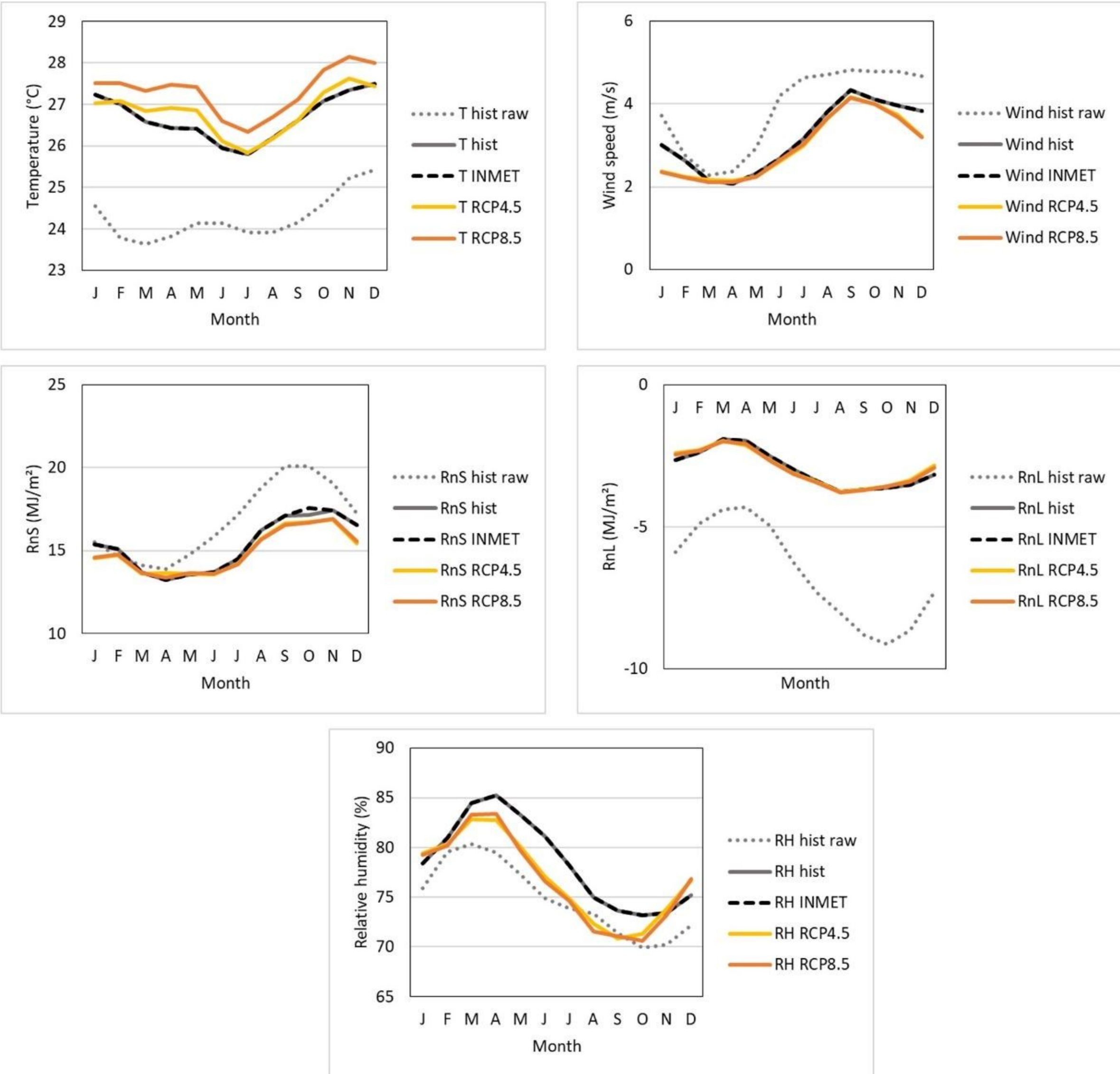

Figure S1. Bias correction of Eta-MIROC5 outputs using QM method

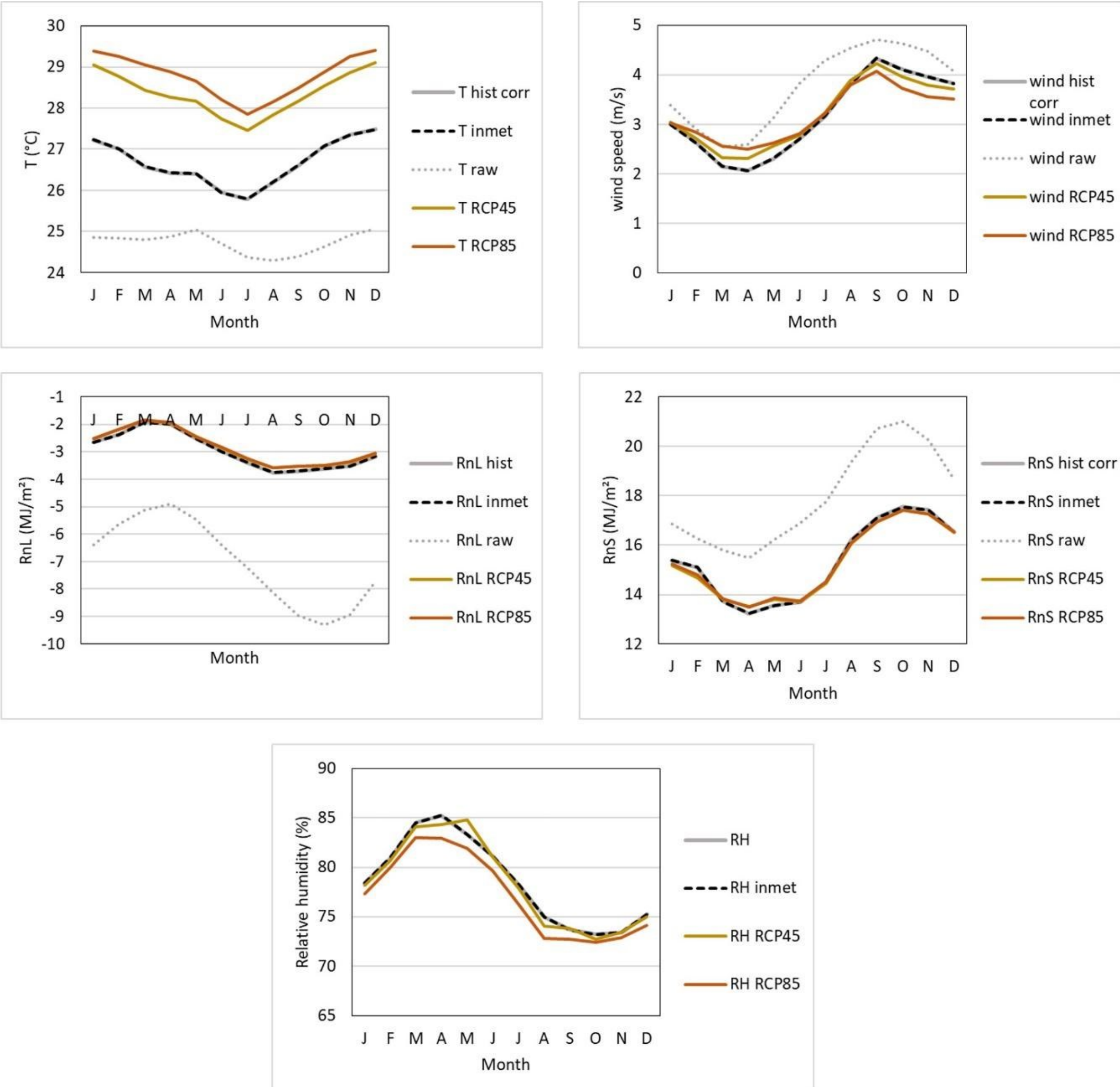

Figure S2. Bias correction of Eta-CanESM2 outputs applying QM method

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