**Aerodynamic mass-transfer open-water evaporation**

Reservoir evaporation at each site was calculated following the aerodynamic mass transfer approach (Quinn, 1979; Subrahamanyam, 2002; Tanny 2008; Verburg, 2010). The aerodynamic method estimates vapor flux based on differential specific humidity and turbulent transfer theory where

|  |  |
| --- | --- |
|  | (1) |

and is evaporation (mm -1), is a time step conversion (s -1) is the density of moist air (kg m-3), is the bulk transfer coefficient, is the windspeed at 2 m (m s-1), is the specific humidity at 2 m (kg kg-1), and is the saturated specific humidity at the water surface. values for each reservoir were based on Monin-Obukhov Similarity Theory (MOST) following equations developed by Brutsaert, 1982 (see Bulk mass-transfer coefficient section below).

In-situ atmospheric measurements were converted to the appropriate variables following psychrometric relationships. Saturated specific humidity at the water surface, , was calculated by

|  |  |
| --- | --- |
|  | (2) |

where (kPa) is the vapor pressure at the surface and is atmospheric pressure (kPa) at the surface. Barometric pressure was not available at all sites (i.e. Stampede, CA) and was estimated from the site elevation according to the hypsometric equation when missing by

|  |  |
| --- | --- |
|  | (3) |

where *P* (kPa) is the pressure at the given elevation and *h* (m) is the reservoir elevation. Vapor pressure at the surface, esat, was calculated from the water surface skin temperature by

|  |  |
| --- | --- |
|  | (4) |

where is the air temperature (°C), and is the skin temperature adjusted for both emissivity and reflected radiation (°C; see Skin Temperature Correction section below). The specific humidity at 2 m,, was calculated using barometric pressure and vapor pressure

|  |  |
| --- | --- |
|  | (5) |

where (kPa) is the saturated vapor pressure at 2 m, found using steps similar to equations 2 and 4 above. The density of moist air,, was calculated according to (Brutsaert, 2005)

|  |  |
| --- | --- |
|  | (6) |

where *P* is the atmospheric pressure (Pa), is the universal gas constant (286.9 J kg-1 K-1), and is the air temperature (K).

**Bulk mass-transfer coefficient, CE**

The bulk mass-transfer coefficient, CE, was calculated for each time-step using an iterative approach based on MOST. MOST applies stability corrections to the near surface transfer coefficients based on wind speed and atmospheric stability. The Monin-Obukhov length, L, can be used to describe atmospheric stability where, , , and correspond to neutral, stable, and unstable conditions, respectively.

The iterative process relies on values of surface temperature, air temperature, wind speed, atmospheric pressure, and specific humidity. This iterative approach has been applied in various forms to estimate bulk transfer coefficients over water bodies including both oceans and reservoirs (Quinn, 1979; Croley, 1989; Tanny, 2008; Verburg, 2010; Subrahamanyam, 2002). This study follows stability functions and roughness length equations developed by Brutsaert, 1982. The general approach is presented below:

Friction velocity, , can be solved by

(7)

where is average wind speed at the reference height (m s-1), is von Karman’s constant (0.41), is the measurement height (2 m in this study), is the roughness length of momentum (m) and is the stability function of momentum. Stability parameters (wind/momentum) and (humidity) are solved for based on atmospheric stability as follows:

Neutral Conditions (z/L=0)

(8a)

Stable Conditions (z/L ≥ 0)

(8b)

Unstable Conditions (z/L ≤ 0)

(8c)

where .The Monin-Obukhov length,, can be represented by

, (9)

where is the virtual temperature of the atmosphere, is the scaling temperature, is the acceleration due to gravity (9.8 m s-2). can be solved by

and can be solved by

(10)

The roughness length of momentum was estimated by

(11)

Where is the kinematic viscosity

The roughness length of humidity,, was estimated by

(12)

The above system of equations can be solved iteratively starting with equations 11, 12, and 10 using the initial conditions of =0.1 m s-1, and and = 0. The iteration then continues solving equations 9, 8, 10, 11, 12, and 7 until the values converge.

The final values are used to solve for by

This final value can be used in equation 1 to solve for evaporation.

**Skin Temperature Correction, Ts**

Following the approach for correcting skin temperature given by Apogee Instruments (2015), the longwave radiation measured by the IRT can be expressed as

|  |  |
| --- | --- |
|  | (13) |

where (W m-2) is the outgoing longwave radiation measured by the sensor, (W m-2) is the outgoing longwave radiation emitted by the water, and (W m-2) is the incoming longwave radiation. Equation 13 can then be reduced using the Stephan-Boltzman equation to the form

|  |  |
| --- | --- |
|  | (14) |

where (K) is the uncorrected skin temperature reading from the sensor, (K) is the corrected skin temperature, (K) is the background sky temperature , is the emissivity of the water, and is the Stephan-Boltzmann constant, . In this study, the IRT was positioned at approximately 45° to the normal, therefore an emissivity of 0.97 was assumed based on findings by Robinson and Davies (1972). The corrected skin temperature can be found by rearranging equation 14 to

|  |  |
| --- | --- |
| . | (15) |

Using the Stephan-Boltzman equation, the background temperature of the sky can be expressed as

|  |  |
| --- | --- |
|  | (16) |

where (W m-2) is the incoming longwave radiation measured with the upward facing sensors of the CNR4 pyrgeometer. Equation 16 can be rearranged to estimate as

|  |  |
| --- | --- |
| . | (17) |

Substituting equation 17 in to equation 15 gives

|  |  |
| --- | --- |
| . | (18) |

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