Surface and Upper Weather Pre-processor (WeatherPrep) Methods for i-Tree Tools Eco and HydroPlus

Version 1.1

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1. Introduction

US Forest Service's i-Tree tools utilize field-surveyed urban forest information, location specific data, weather data, and air pollutant measurements to quantify urban forest structure and numerous forest-related effects such as carbon sequestration, energy savings, and pollution removals. Key input is weather data; hourly global surface weather data (NCDC 2013a) and upper air data (NOAA 2013b). The document describes the methods to process hourly weather parameters required to run i-Tree tools.

2. Materials

Surface weather data contain measurements at a single weather station for a year in the Integrated Surface Hourly (ISH) format. Upper air data contain measurements in the morning and afternoon at a single site for a year in the FSL format. Both can be downloaded from Internet or extracted from archive DVDs (NOAA 2013a, 2013b). Tables 1 and 2 summarize the parameter contained in these input data and created and stored in output MS-Access database.

Table 1 Surface weather variables

Input		Output	
Variables	Units/Values	Variables	Units/Values
wind speed	[miles h^{-1}]	wind speed	[Knot, m s^{-1}, miles h^{-1}]
cloud ceiling height	[100 ft]	cloud ceiling height	[100 ft]
sky cover	[clear]	total cloud cover	[0/3.75/7.5/10]
	[scattered]	opaque cloud cover	[0/3.75/5/7.5/10]
	[broken]	translucent cloud cover	[0/2.5/3.75/5/7.5/10]
	[overcast]		
	[obscured]		
	[partial obscuration]		
temperature	[F]	temperature	[F, C, K]
dew point temperature	[F]	dew point temperature	[F, C]
station altimeter setting	[in]		
station pressure	[mb]	station pressure	[in, kPa, mb]
1-hour liquid precipitation	[in]	1-hour liquid precipitation	[in, m]
		solar zenith angle	[deg]
		air mass	[kg m^{-2}]
		direct normal solar radiation	[W m^{-2}]
		diffuse horizontal solar radiation	[W m^{-2}]
		global horizontal solar radiation	[W m^{-2}]
		photosynthetically active radiation (PAR)	[W m^{-2}, uE m^{-2} s^{-1}]
		net radiation	[W m^{-2}]
		relative humidity	
		vapor pressure	[kPa]
		saturated vapor pressure	[kPa]
		potential evaporation tree	[m h^{-1}]
		potential evaporation ground	[m h^{-1}]
		potential evaporation snow tree	[m h^{-1}]
		potential evaporation snow ground	[m h^{-1}]
		potential evapotranspiration tree	[m h^{-1}]

Table 2 Upper air variables

Input		Output	
Variables	Units/Values	Variables	Units
height	[m]	rural mixing height	[m]
pressure	[0.1 mb]	urban mixing height	[m]
temperature	[0.1 °C]		

3. Methods

3.1. Solar Zenith Angle

Solar position is calculated based on algorithms reported by Iqbal (1983). The solar zenith angle, θ_z [deg] is calculated as:

$$\cos\theta_z = \sin\delta\sin\phi + \cos\delta\cos\phi\cos\omega \tag{1}$$

δ [deg] is the solar declination:

$$\delta = (0.006918 - 0.399912\cos\psi + 0.070257\sin\psi - 0.006758\cos 2\psi + 0.000907\sin 2\psi - 0.002697\cos 3\psi + 0.00148\sin 3\psi)(180/\pi) \tag{2}$$

ψ [deg] is the day angle:

$$\psi = \frac{360(d-1)}{365} \tag{3}$$

d is the day number of a year (1 to 365, assuming no leap year).
ϕ[deg] is the geographic latitude (north positive).
ω[deg] is the hour angle:

$$\omega = 15\mathrm{TST} - 180 \tag{4}$$

TST is the true solar time:

$$TST = LST + 4(|L_s| - |L_e|) + E_t \tag{5}$$

L_s [deg] is the standard meridian, occurring every 15° of longitudinal displacement from the prime meridian that runs through Greenwich, England:

$$\mathrm{L_s} = 15\mathrm{T_z} \tag{6}$$

T_z is the time difference in hour from Greenwich to a certain location (east positive). L_e [deg] is the local longitude (east positive), 4 [min] is the longitude correction for every degree of difference between L_s and L_e, and E_t [min] is a function of time, variation in length of a solar day (the duration of the sun to complete one cycle about a stationary observer on earth) throughout a year:

$$\mathrm{E_t} = (0.000075 + 0.001868\cos\psi - 0.032077\sin\psi - 0.014615\cos 2\psi - 0.040849\sin 2\psi)(229.18) \tag{7}$$

Atmosphere-corrected solar zenith angle ($\theta_z{}'$) is calculated as:

$$\theta_z' = \theta_z - C_r \tag{8}$$

C_r is the refraction correction on the solar elevation angle (α=90- θ_z), and calculated as (Astronomical Almanac 1992): if $-1 \leq \alpha < 15$

$$C_r = \frac{P_a}{T_a} \cdot \frac{0.1594+0.0196\alpha+0.00002\alpha^2}{1+0.505\alpha+0.0845\alpha^2} \tag{9}$$

if $15 \leq \alpha < 90$

$$C_r = \frac{0.00452 \cdot \frac{P_a}{T_a}}{\tan\alpha} \tag{10}$$

P_a is the absolute pressure (=1013.25 mb) and T_a is the absolute temperature (=288.15 K).

3.2. Solar Radiation

Accounting for extraterrestrial, and direct and diffuse components for clear and cloudy, solar radiation parameters (direct normal solar radiation, diffuse horizontal solar radiation, global horizontal solar radiation, PAR, and net radiation) are calculated based on National Renewable Energy Laboratory's METSTAT (Maxwell 1998, NSRDB 1995) and the Bird Clear Sky model (Bird & Hulstrom 1981).
The extraterrestrial global horizontal radiation, *ETR* [Wm^{-2}] is calculated as (Maxwell 1998):

$$ETR = I_o \cos\theta_z \tag{11}$$

I_o [Wm^{-2}] is the extraterrestrial normal radiation (i.e., solar radiation on a plane normal to the solar beam) (Maxwell 1998):

$$I_o = e_o I_c \tag{12}$$

I_c is the solar constant, 1367 [W m^{-2}] (Fröhlich and Brusa 1981). e_o is the eccentricity of the Earth's orbit (Spencer 1971; Bird and Riordan 1986; Iqbal 1983):

$$e_o = (r_o/r)^2 = 1.00011 + 0.034221cos\psi + 0.00128sin\psi$$

$$+0.000719cos2\psi + 0.000077sin2\psi \tag{13}$$

r is the sun-earth distance and r_o is the mean sun-earth distance.

Molecules of the air, ozone, uniformly mixed gas, water vapors, aerosols, translucent and opaque clouds impact the direct component through scattering and absorption. The direct solar radiation after passing through the Earth's atmosphere, S_n [W m^{-2}]. i.e., solar radiation adjusted for Earth's curvature and latitude (not on a plane normal to the solar beam but referred to by some as horizontal when adjusted for Earth's curvature), is calculated as:

$$S_{dir} = K_n ETR \tag{14}$$

K_n is the transmission value (Maxwell 1998):

$$K_n = 0.9751 T_R T_O T_{UM} T_W T_A T_{OPQ} T_{TRN} \tag{15}$$

T_R is the Rayleigh scattering transmittance (Bird and Hulstrom 1981):

$$T_R = exp\{-0.0903(M')^{0.84}[1 + M' - (M')^{1.01}]\} \tag{16}$$

M' [kg m^{-2}] is the pressure-corrected relative optical air mass (Bird and Riordan 1986):

$$M' = M\frac{P}{P_o} \tag{17}$$

M [kg m^{-2}] is relative optical air mass, which is a measure of the length of the path through the atmosphere to sea level traversed by light rays from a celestial body, expressed as a multiple of the path length for a light source at the zenith. P [mb] is measured surface pressure and P_o=1013mb. Optical air mass is calculated as (Kasten and Young 1989):

$$M = \left[sin\,\alpha + 0.50572(\alpha + 6.07995^{\circ})^{-1.6364}\right]^{-1} \tag{18}$$

T_O is the ozone absorption transmittance (Bird and Hulstrom 1981):

$$T_O = 1 - 0.1611X_O(1 + 139.48X_O)^{-0.3035} - \frac{0.002715X_O}{1+0.044X_O+0.0003X_O{}^2} \tag{19}$$

X_O [cm] is the total amount of ozone in a slanted path (Bird and Hulstrom 1981):

$$X_O = U_O M \tag{20}$$

U_O [cm] is the amount of ozone in a vertical column from surface.
T_{UM} is the uniformly mixed gas absorption transmittance (Bird and Hulstrom 1981):

$$T_{UM} = exp[-0.0127(M')^{0.26}] \quad (21)$$

T_W is the water vapor absorption transmittance and calculated as (Maxwell 1998):

$$T_W = 1.0 - 1.668X_W/[(1.0 + 54.6X_W)^{0.637} + 4.042X_W] \quad (22)$$

X_W [cm] is the precipitable water vapor in a slant path through the atmosphere (NSRDB 1995):

$$X_W = U_W M \quad (23)$$

T_A is the aerosol absorption and scattering transmittance (Maxwell 1998):

$$T_A = exp(-\tau_A M) \quad (24)$$

τ_A is the broadband aerosol optical depth in a vertical path from the surface (broadband turbidity) (Maxwell 1998):

$$\tau_A = a\ sin[(360d/365) - b] + c \quad (25)$$

Coefficients, *a*, *b*, and *c* for a study site are provided by CoeffA, CoeffPHI and CoeffC, respectively in the i-Tree's location database.

T_{OPQ} is the opaque cloud cover transmittance (Maxwell 1998):

$$T_{OPQ} = [10.0 - (OPQ + N)]/10.0 \quad (26)$$

OPQ represents observed opaque cloud cover (0 -10 tenths). *N* is added to obtain effective opaque cloud cover (Maxwell 1998):

$$N = A_1\, sin(18.0OPQ) + B_1 sin\,(36.0OPQ) \quad (27)$$

The coefficients A_1 and B_1 were determined empirically as (NSRDB 1995):

$$A_1 = 4.955[1.0 - exp(-0.454M)] - 3.4 \quad (28)$$

$$B_1 = -0.2A_1\ \ if\ A_1 \le 0.0\ or$$

$$= 0.1A_1\ if A_1 > 0.0 \quad (29)$$

T_{TRN} is the translucent cloud transmittance (NSRDB 1995).

$$T_{TRN} = A_{TRN} - B_{TRN}M \quad (30)$$

Coefficients A_{TRN} and B_{TRN} can be determined empirically using measurements (NSRDB 1995).

Molecules of the air, aerosols, opaque and translucent clouds impact the diffuse component through scattering and reflection. The diffuse solar radiation after passing through the Earth's atmosphere, S_d, (not on a plane normal to the solar beam but referred to by some as horizontal when adjusted for Earth's curvature) is calculated as:

$$S_{dif} = K_d ETR \tag{31}$$

K_d is the transmission value (Maxwell 1998):

$$K_d = K_{d0} + K_{S_{GRF}} \tag{32}$$

where

$$K_{d0} = \left[f(M)\left(K_{S_R}+K_{S_A}\right) + K_{S_{OPQ}} + K_{S_{TRN}}\right] PSW \tag{33}$$

f(M) is the empirical air mass function accounting for an air mass dependency of K_{S_R} and K_{S_A} (NSRDB 1995):

$$f(M) = 0.38 + 0.925 exp(-0.851M) \tag{34}$$

K_{S_R} is the diffuse radiation from Rayleigh scattering (NSRDB 1995):

$$K_{S_R} = 0.5(1.0 - T_R)T_O T_{UM} T_{AA} \tag{35}$$

T_{AA} is aerosol absorption transmittance (Bird and Hulstrom 1981):

$$T_{AA} = 1.0 - K_1(1.0 - M + M^{1.06})(1.0 - T_A) \tag{36}$$

The coefficient K_1=0.10.

K_{S_A} is the diffuse radiation from aerosol scattering (NSRDB1995):

$$K_{S_A} = B_A(1.0 - T_A)T_O T_{UM} T_{AA} \tag{37}$$

The coefficient B_A=0.84.

$K_{S_{OPQ}}$is the diffuse radiation from opaque cloud scattering (NSRDB 1995):

$$K_{S_{OPQ}} = -0.06 + B_2 T_A + C_2 T_A^2 \tag{38}$$

$$B_2 = 0.0953 + 0.137OPQD - 0.0409OPQD^2 + 0.00579OPQD^3 - 0.000328OPQD^4 \quad (39)$$

$$C_2 = -0.109 - 0.02OPQD + 0.011OPQD^2 - 0.00156OPQD^3 + 0.000121OPQD^4 \quad (40)$$

$$OPQD = OPQ + D \quad (41)$$

D is used to adjust the observed *OPQ* data for calculating the diffuse radiation from clouds (NSRDB 1995):

$$D = 0.5A_1 \, sin(18.0OPQ) \quad (42)$$

$K_{S_{TRN}}$ is the diffuse radiation from translucent cloud scattering (NSRDB 1995):

$$K_{S_{TRN}} = -0.00235 + 0.00689TRN + 0.000209TRN^2 \quad (43)$$

TRN is observed translucent cloud cover (0-10 tenths), calculated as the difference of total cloud cover, *TOTAL* (0-10 tenths) and *OPQ*.

$$TRN = TOTAL - OPQ \quad (44)$$

Precipitation switch, *PSW*=0.06 when *OPQ* is greater than or equal to 8 tenths and the occurrence of rain is reported. *PSW*=1.0 otherwise.

K_{SGRF} is diffuse radiation from ground reflectance, and the sum of K_{SGRF1} and K_{SGRF2}, normalized diffuse radiation from multiple reflections between surface and clouds, and between surface and the atmosphere, respectively (NSRDB 1995):

$$K_{S_{GRF}} = K_{S_{GRF_1}} + K_{S_{GRF_2}} \quad (45)$$

$$K_{S_{GRF_1}} = (K_n + K_{d0})[R_{CLD}(ALB - 0.2)] \quad (46)$$

$$K_{S_{GRF_2}} = (K_n + K_{d0})(R_{ATM}ALB) \quad (47)$$

ALB is the monthly surface albedo defined for a study site.
R_{CLD} is the broadband cloud reflectance (NSRDB 1995):

$$R_{CLD} = 0.06OPQ + 0.02TRN \quad (48)$$

R_{ATM} is the broadband atmospheric reflectance (NSRDB 1995):

$$R_{ATM} = [0.0685 + 0.16(1.0 - T_{AS})][(10 - OPQ)/10] \quad (49)$$

T_{AS} is aerosol scattering transmittance (Bird and Hulstrom 1981):

$$R_{AS} = T_A/T_{AA} \tag{50}$$

The global total solar radiation G [W/m^2] is the sum of the direct solar radiation S_{dir} and the diffuse solar radiation S_{dif} (both not on a plane normal to the solar beam but referred to by some as horizontal when adjusted for Earth's curvature).

$$G = S_{dir} + S_{dif} = K_n ETR + K_d ETR = (K_n + K_d)ETR \tag{51}$$

It is assumed that 46% of G [W m^{-2}] is in the visible portion or PAR (Norman 1982).

$$PAR = 0.46G \tag{52}$$

Net radiation at Earth's surface R_n [W m^{-2}] is calculated with the net short and long wave radiations:

$$R_n = S_{net} + L_{net} \tag{53}$$

S_n is the net short wave radiation:

$$\mathrm{S_{net}} = (1 - \mathrm{ALB})\mathrm{G} \tag{54}$$

L_n is the net long wave radiation:

$$L_{net} = E_s(L_{sky} + L_{cld} - L_{sfc}) \tag{55}$$

L_{sky} and L_{cld} is the downwelling long wave radiation from the sky and cloud, respectively:

$$L_{sky} = E\,\frac{1-TOTAL}{10}\,\sigma T^4 \tag{56}$$

$$L_{cld} = E\,\frac{TOTAL}{10}\,\sigma T^4 \tag{57}$$

E is the emissivity of the clear sky:

$$E = 0.741 + 0.0062T \tag{58}$$

σ is the Stefan-Bolzmann constant (=5.67×10^{-8} W m^{-2} K^{-4}).

L_{sfc} is the upwelling long wave radiation from the surface:

$$L_{sfc} = \sigma T^4 \tag{59}$$

3.3. Potential Evaporation

It is important for users of these methods to recognize uncertainty surrounding estimation of evaporation, transpiration, and sublimation. The science and engineering research continues, and scholars and practitioners admit to not fully representing the real-world process with any equation, nor having sufficient insight details on the best parameter values and constraining conditions. In our effort to proceed, we have consulted directly with experts Richard Allen and Steven Fassnacht, cited below, to create reasonable yet simple algorithms representing these processes. We have been assured the practice is and will remain imperfect for some time, and what we have created is surely imperfect, and we accept all the errors as ours. The modified Penman-Monteith approach was chosen due to its global familiarity, ability to work across a range of situations, its analytical origins and process-based physics, and its successes. The approach was primarily intended for irrigated agricultural crops, yet its parameters can be modified to represent a range of natural and artificial situations within reasonable ranges of uncertainty.

Potential evaporation, PE [m s^{-1}] below is used to represent potential evaporation, transpiration, and sublimation from different land cover types using a single form of the modified Penman-Monteith equation from Eq 13.1 of Chin (2021). This equation is also described in Shuttleworth (1992) in the Handbook of Hydrology, but with unit confusion. It is critical to note, Equation 13.1 of Chin (2021) requires that all terms are in fundamental SI units (J, Pa, kg, s, m, C, etc.). Note, Issues with Eq 4.2.27 of Shuttleworth (1993) include not dividing by density of water, using mixed units MJ day^{-1}, kPa $°C^{-1}$, etc.

$$PE = \frac{1}{\rho_w \lambda}\left[\frac{\Delta(R_n - G) + \rho_a c_p \frac{e_s - e_a}{r_a}}{\Delta + \gamma\left(1 + \frac{r_s}{r_a}\right)}\right] \tag{60}$$

The potential evaporation equation above is used for all surfaces, soil, vegetation, water, and snow, as explained below with calculation of coefficients. For snow, on the ground or on vegetation canopy, the use of the above equation is based on Eq 6 and 7 of Lundberg et al. (1998), as discussed in personal communication with Fassnacht (2022).

λ [J kg^{-1}] is the latent heat of vaporization, but note it computed by Eq 13.36 Chin (2021) in units of MJ kg^{-1}. Please note, although some terms are computed in mixed units, for use in the potential evapotranspiration equation they need to be converted to fundamental units.

$$\lambda = 2.501 - 0.002361T \tag{61}$$

where T is air temperature [°C] given water temperature often unknown.

ρ_w [kg m^{-3}] is the density of water based on the temperature of water, T_w [°C] from equation within Figure 11.1.1 McCutcheon (1993):

$$\rho_w = 1000 \cdot \left(1 - \left(\frac{T_w + 288.9414}{508929.2 \cdot (T_w + 68.12963)}\right) \cdot (T_w - 3.9863)^2\right) \tag{62}$$

where T_w if unknown is taken as air temperature as a best available estimate.

G [W m^{-2}] is heat ground flux, defined for two vegetation heights after Eq 13.32 by Chin (2021); also Jensen and Allen (2015) Chapter 5. Note, Jensen and Allen (2015) "Standardized Reference Evapotranspiration for Short Reference ET_{os}: Reference ET calculated for a short crop having height of 0.12 m (similar to grass), albedo of 0.23, surface resistance of 70 sm^{-1} for 24-h calculation time steps, and 50 sm^{-1} for hourly or shorter periods during daytime and 200 sm^{-1} during nighttime". Note, Jensen and Allen (2015) "Standardized Reference Evapotranspiration for Tall Reference ET_{rs}: Reference ET calculated for a tall crop having height of 0.50m (similar to alfalfa), albedo of 0.23, surface resistance of 45 s m^{-1} for 24-h calculation time steps, and 30 s m^{-1} for hourly or shorter periods during daytime and 200 sm^{-1} during nighttime with daytime G=R_n =0.04 and nighttime G=R_n =0.2.". Note, short crop coefficients 0.1 during day, 0.5 during night. Tall crop coefficients 0.04 during day, 0.2 during night. Note, Jensen and Allen (2015) generally suggest using reference crop coefficients and then adjusting to actual conditions. If R_n and net radiation (W/m^2) are positive, then presume during day and ground heat flux takes one form, otherwise presume during night and it takes another form, shown below.

For land without water or snow covering the surface use:

$$G = 0.04(R_n) \text{ if } R_n > 0 \tag{63a}$$

$$G = 0.2(R_n) \text{ if } R_n <= 0 \tag{63b}$$

For water or snow use:

$$G = 0.25(R_{sw}) - 0.05(R_{lw}) \tag{63c}$$

where R_n [W m^{-2}] is net radiation, the balance of shortwave and longwave, with positive downwards, R_{sw} [W m^{-2}] is shortwave radiation (direct and diffuse combined) and R_{lw} [W m^{-2}] is longwave radiation. Ground heat flux for water is essentially a heat storage term. Note, this approximates theory of Jensen and Allen (2015) Chapter 6, Evaporation from Water Surfaces, "An important distinction between R_n for a water body and R_n for vegetation or soil is that with soil and vegetation, essentially all of the R_n quantity is captured at the “opaque” surface and is immediately available solar radiation, R_s, for conversion to λE or H or conduction into the surface as G. With water, however, much of the penetrates to some depth in the water body, depending on the turbidity of the water, where it is converted to Q_t, heat storage." This approach is modified for smaller waters below. The adjustment is from Eq 6-14a and 6-14b of Jensen and Allen (2015) modified from large lakes. Note, large lakes used C_{Ga} = 0.5 and C_{Gb} = 0.8; reduced for smaller waters. Large lakes increased loss from long wave radiation at Julian Day = 180; not simulated for smaller waters.

Δ [kPa °C^{-1}] is the slope of vapor pressure temperature curve:

$$\Delta = \frac{4098 e_s}{(237.3+T)^2} \tag{64}$$

ρ_a [kg m^{-3}] is the density of air from Eq 4.2.4 from Shuttleworth (1993), but note Shuttleworth incorrectly used 275 in denominator rather than the correct value of 273.15, as noted in Chin Eq 13.51 which unfortunately uses 3.45 in place of the correct 3.486. The equation was tested against values of air density from the EngineeringToolbox.com for air temperature from 0 to 50 °C at standard atmospheric pressure.

$$\rho_a = 3.486 \cdot \frac{P}{273.15+T} \tag{65}$$

P [kPa] is atmospheric pressure

T [°C] is air temperature

cp [kJ kg^{-1} °C^{-1}] is the specific heat of moist air, =1.013 kJ kg^{-1} °C^{-1}

e_s [kPa] is the saturated vapor pressure, using Eq 4.2.2 of Shuttleworth (1993) using air dry bulb temperature T (°C). Note, 0.6108 kPa is the gas ratio used for units of kPa:

$$e_s = 0.6108 exp\left(\frac{17.27T}{237.3+T}\right) \tag{66}$$

e_a [kPa] is the actual vapor pressure, using Eq 4.2.2 of Shuttleworth (1993) using dew point temperature T_{dew} (°C). Note, 0.6108 kPa is the gas ratio used for units of kPa:

$$e_a = 0.6108 exp\left(\frac{17.27T_{dew}}{237.3+T_{dew}}\right) \tag{67}$$

γ [kPa °C^{-1}] is the psychrometric constant from Eq 13.37 Chin (2021). Note: Eq 13.37 should use specific heat with units of MJ kg^{-1} °C^{-1}, but incorrectly states units are kJ kg^{-1} °C^{-1}.:

$$\gamma = \frac{c_p p}{E * \lambda} \tag{68}$$

P [mb] is the measured surface pressure.

c_p [MJ kg^{-1} °C^{-1}] is specific heat of moist air at constant pressure, 1013.0 E^{-6} [MJ kg^{-1} °C^{-1}].

E is the ratio of the molecular weight of water vapor to the molecular weight of dry air, 0.622.

λ [MJ kg^{-1}] is latent heat of vaporization, calculated above with a value that averages 2257000.0 E^{-6} MJ/kg

r_s [s m^{-1}] is the surface resistance or stomatal resistance, which for liquid water or snow is set to 0. For vegetation it is based on a bulk stomatal resistance value of around 200 s/m according to Shuttleworth (1993) Eq 4.2.22 and interpretation of Allen (1998), and from Chin (2021) around Eq 13.9. Note, r_s is derived from bulk stomatal resistance by division with LAI. Note, minimum values of r_s ~ 100 s m^{-1} as given in Table 2 of Liang et al. (1994) for the VIC model. Note, Box 5 of Allen et al. (1998) and the minimum values of r_s ~ 40 s m^{-1} found in Chpt 11 of Jensen and Allen (2015).

The parameter B_f used here is based on suggestions of Jensen and Allen (2015) Chapter 4 Energy Balance, "In addition, a recommendation by Allen et al. (2006) to use the same 50 s m^{-1} surface resistance for hourly or shorter periods during daytime and 200 s m^{-1} during nighttime with the FAO-56 Penman-Monteith method has made the FAO ET_o and ET_{os} references equivalent for hourly time steps and for 24-h periods.". Note, Allen (personal communication) suggests larger stomatal resistance for vegetation in a landscape, i.e., not irrigated reference crop, stating it evolved to survive by restricting water loss. Note, the equation derived here and coefficient 0.9 was adjusted to fit expected trends in evapotranspiration by utilizing information about humidity from vapor pressure.

$$r_s = \left(\frac{200}{B_f}\right)/LAI \tag{69}$$

where LAI [m^2 m^{-2}] is the leaf area index; if the LAI < 1 then it is set to 1, LAI = 1.

B_f [fraction] is the bulk stomatal resistance fraction, defined as

$$B_f = [1 - ((e_s - e_a)/e_x)]^{0.9} \tag{70}$$

where the e_s is the saturated vapor pressure, e_a is actual.

The aerodynamic resistance r_a [s m^{-1}] for the vegetation is from Eq 13.2 of Chin (2021) or Eq 4.2.25 of Shuttleworth (1993). Note, Eq 4.2.25 should use the station windspeed, not a windspeed estimated at the height of any lower object. Note, a height of wind speed measurement must be greater than the zero plane displacement height to avoid failure of the log functions.

$$r_a = \frac{ln\left[\frac{z_m - d}{z_{om}}\right] ln\left[\frac{z_h - d}{z_{oh}}\right]}{k^2 u_z} \tag{71}$$

z_m [m] is the measurement height of the wind speed.

z_h [m] is the measurement height of the air temperature and humidity.

d [m] is zero plane displacement height from Eq 13.3 of Chin (2021) equal to 2/3 h, where h [m] is the height of the vegetation.

z_{om} [m] is the roughness height for momentum transfer from Eq 13.3 of Chin (2021) equal to 0.123 h [m] is the height of the vegetation.

z_{oh} [m] is the roughness height for heat transfer from Eq 13.4 of Chin (2021) equal to 0.1(z_{om}) [m] or 0.0123 h.

k is the von Karman constant of 0.4 or 0.41

U_z [m s^{-1}] is the wind speed at the measurement height.

The aerodynamic resistance r_a [s m^{-1}] for water surfaces is different than for vegetation, based on Eq 13.8 Chin (2021) or Eq 4.2.29 from Shuttleworth (1993), and uses an aerodynamic resistance ratio of snow to rain, R_{s2r}, set to 10 according to Lundberg et al. (1998) Eq 6 and 7. For open water it is calculated as:

$$r_a = \frac{4.72 \cdot ln\left(\frac{z_m}{z_w}\right)}{1 + 0.536 U_z} R_{s2r} \tag{72}$$

z_m [m] is the measurement height of the wind speed.

z_w [m] is the roughness height of water = 0.00137m as given by Chin (2021).

R_{s2r} is the aerodynamic resistance ratio of snow to rain = 10 from Lundberg et al. (1998) derived from Eq 6 and 7.

The aerodynamic resistance r_a [s m^{-1}] for snow on the ground is based on Eq 4 of Fassnacht (2004), who cites Eq 3 of Light (1941). Note the Fassnacht (2004) equation had an error and did not square the numerator, which is corrected below:

$$r_a = \frac{ln\left(\frac{z_m}{z_s}\right)ln\left(\frac{z_m}{z_s}\right)}{k^2\ U_z} R_{s2r} \tag{73}$$

z_m [m] is the measurement height of the wind speed.

z_s [m] is the roughness height of snow = 0.005 m as given by Fassnacht (2004) and Stull (2000) in Table 4.1.
R_{s2r} is the aerodynamic resistance ratio of snow to rain = 10 from Lundberg et al. (1998) derived from Eq 6 and 7.

The aerodynamic resistance r_a [s m^{-1}] for snow on in the tree canopy is based on Eq 4 of Fassnacht (2004), who cites Eq 3 of Light (1941). Note the Fassnacht (2004) equation had an error and did not square the numerator, which is corrected below:

$$r_a = \frac{ln\left(\frac{z_t}{z_s}\right)ln\left(\frac{z_t}{z_s}\right)}{k^2\ U_t} \tag{74}$$

z_t [m] is the measurement height of the wind speed at tree height.

z_s [m] is the roughness height of snow = 0.005 m as given by Fassnacht (2004) and Stull (2000) in Table 4.1.

U_t [m/s] is the wind speed at tree height.

Wind speed at tree height is computed with Eq 4.14b of Stull (2000):

$$U_t = U_z \frac{ln\left(\frac{Z_t}{z_{om}}\right)}{ln\left(\frac{Z_m}{z_{om}}\right)} \tag{75}$$

U_z is the measured wind speed, z_t is the wind estimate height for the tree, z_m is the actual wind measurement height (e.g., 10m), and z_{om} is the roughness height for the measured wind speed location, typically the airport and given as = 0.03 m.

3.4. Mixing Height

3.4.1. Twice-Daily Mixing Height

Hourly mixing height is calculated based on US EPA's mixing height program (US EPA 1998) and PCRAMMET (US EPA 1999). Twice-daily (i.e. morning and afternoon) mixing heights are first calculated with surface weather and upper air data, they are then interpolated hourly. The results will be used in i-Tree Eco to quantify the air quality improvements due to air pollutant removal by vegetation.

The twice-daily mixing height is computed as the height where the surface temperature raised at the dry adiabatic lapse rate intersects with the observed 12:00 GMT sounding temperatures. For morning and afternoon mixing height, the minimum surface temperature from 02:00 to 06:00 local standard time (LST) and the maximum from 12:00 to 16:00 LST is used, respectively. Surface and sounding temperatures are converted to the potential temperature (θ) [K] as:

$$\theta = T\left(\frac{1000}{P}\right)^{0.286} \tag{76}$$

where T is temperature (K) and P is pressure (mb). Five is added to the minimum surface temperature to account for some initial surface heating just after sunrise (Holzworth 1967). Two heights in the sounding layers can be identified so that θ at the surface is found between θ in the two layers. Based on the two heights identified, the mixing height can be calculated by linear interpolation. When the height is missing for the layer, the mixing height is determined by interpolating from θ to P and then from P to height. When the surface θ is smaller than θ at the first layer, or upper air data are missing for a specific day, it is impossible to calculate the mixing heights. In such cases, twice-daily mixing heights are interpolated from those calculated.

3.4.2. Hourly Mixing Height

Computation of hourly mixing heights requires: 1) morning mixing heights for the computation day (i) and the following day (i+1) and afternoon mixing heights for the days (i-1), (i), and (i+1), 2) sunrise and sunset LST, and 3) hourly stability class (neutral or not neutral). Table 3 presents interpolation methods defined based on hours of estimation, stability (neutral or not), and sites (urban or rural). AM_i and PM_i are mixing heights estimated for the morning and afternoon, respectively on the computation day, *i*. Letters from (a) to (e) indicate interpolation methods illustrated in Figures 1 (a) to (e).

Table 3 Hourly mixing height interpolation methods for neutral and not neutral stability conditions in urban and rural sites

Hours (h)	Urban		Rural	
	Neutral	Not Neutral	Neutral	Not Neutral
0:00 ≤ h ≤ sunrise[a]	(a)	AM_i	(a)	(a)
sunrise < h < 13:00[b]	(a)	(b)	(a)	(e)
13:00 ≤ h ≤ sunset[b]	PM_i	PM_i	PM_i	PM_i
sunset < h ≤ 23:00[b]	(c)	(d)	(c)	(c)

[a]: stability class at sunrise hour is used. [b]: stability class at each hour is used.

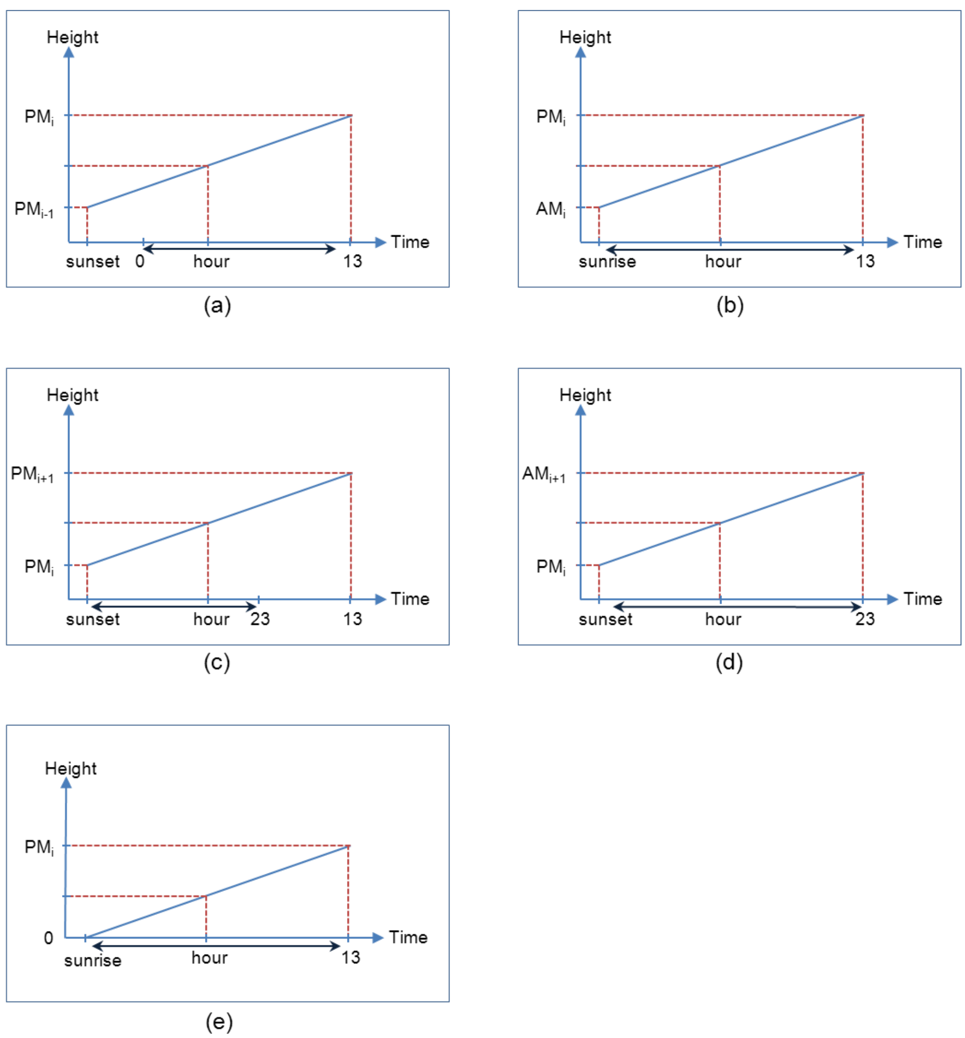

Figure 1 Hourly mixing height interpolation methods (a) interpolate with PM_{i-1} and PM_i for 00:00 to 13:00 LST

(b) interpolate with AM_i and PM_i for sunrise to 13:00 LST (c) interpolate with PM_i and PM_{i+1} after sunset to 23:00 LST

(d) interpolate with PM_i and AM_{i+1} after sunset to 23:00 LST (e) interpolate with 0 and PM_i after sunrise to 13:00 LST

3.4.3. Sunrise and Sunset LST

Following algorithms implemented in PCRAMMET, this function calculates hourly solar elevation angles as well as sunrise and sunset hours for the location specified with latitude, longitude, and time zone for a Julian day. The PCRAMMET uses known earth-sun relationships (e.g., Sellers (1965)).

3.4.4. Atmospheric stability class

Seven atmospheric stability classes (1 to 7) are determined based upon day/nighttime, wind speed, and insolation classes for daytime. The insolation is classified into five classes (Strong, Moderate, Slight, Weak, and Overcast) by means of the Turner (1964) objective method using solar elevation angle, ceiling height, and cloud cover as indicators (Table 4).

Table 4 Insolation classes as a function of solar elevation angle, ceiling height and cloud cover

Solar elevation angle (a)	Ceiling height (h)	Cloud cover (c)		
		c≤5/10	5/10<c<10/10	c=10/10
0° < a≤ 15°	h<7000 ft	Weak	Weak	Overcast
	7000 ft ≤h≤16000 ft			Weak
	16000 ft < h			
15° < a≤ 35°	h<7000 ft	Slight	Weak	Overcast
	7000 ft ≤h≤16000 ft			Weak
	16000 ft < h			
35° < a≤ °60	h<7000 ft	Moderate	Weak	Overcast
	7000 ft ≤h≤16000 ft		Slight	Weak
	16000 ft < h			Slight
60° < a	h<7000 ft	Strong	Slight	Overcast
	7000 ft ≤h≤16000 ft		Moderate	Slight
	16000 ft < h			Moderate

The atmospheric stability classes are selected from Table 5 based upon wind speed, day/nighttime, and insolation classes for daytime determined above. Stability smoothing is implemented as standard EPA practice in regulatory dispersion modeling is to restrict temporal changes in stability class to no more than one per hour.

Table 5 Stability classification criteria

Wind speed	Daytime	Day/ Nighttime	Nighttime

(knots)	Strong	Moderate	Slight	Weak	Overcast	≥ 5/10 cloud	< 5/10 cloud
≤ 1	1	1	2	3	4	6	7
2	1	2	2	3	4	6	7
3	1	2	2	3	4	6	7
4	1	2	3	4	4	5	6
5	1	2	3	4	4	5	6
6	2	2	3	4	4	5	6
7	2	2	3	4	4	4	5
8	2	3	3	4	4	4	5
9	2	3	3	4	4	4	5
10	3	3	4	4	4	4	5
11	3	3	4	4	4	4	4
≥ 12	3	4	4	4	4	4	4

The first six of the seven classes correspond to Pasquill's (1974) classifications (1: strongly unstable, 2: moderately unstable, 3: slightly unstable, 4: neutral, 5: slightly stable, 6: moderately stable). The seventh category corresponds to the 'dashes' in the original classification by Pasquill and indicates a strong, ground-based nocturnal temperature inversion with non-definable wind flow conditions.

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