EEMT-topo Computations

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Data input

PRISM precipitation PRCP climatology (1981-2010) (spatial resolution 800 m)

Local meteorological data (Temperature, RH, Wind Speed and Pressure) downloaded for the VCNP Headquarters Climate Station from 2003 to 2012. Downloaded from http://www.wrcc.dri.edu/vallescaldera/.

2011 National Agriculture Imagery Program (NAIP) Orthoimagery (multispectral 4-band) for Valles Caldera. Data were collected from May 2011 to August 2011. Downloaded from http://seamless.usgs.gov on 7/5/2012.

Jemez River Basin 2010 LiDAR dataset (Snow-off) available from http://criticalzone.org/catalina-jemez/data/dataset/2613/. 1 m DEM was up-scaled to 10 m DEM.

MODIS Albedo 16-Day L3 Global 500m data (MCD43A3) obtained from (https://lpdaac.usgs.gov/dataset_discovery/modis/modis_products_table/mcd43a3).

Computations

1 Monthly precipitation

Daly, C., W. P. Gibson, G.H. Taylor, G. L. Johnson, P. Pasteris (2002) A knowledge-based approach to the statistical mapping of climate. Climate Research, 22, 99-113.

PRISM Climate Group, Oregon State University, http://www.prismclimate.org, data accessed in August 2012.

Precipitation is in cm

$$P = \frac{Prcp}{1000} \qquad (1)$$

Precipitation data were re-sampled from 800 m grid to 10 m grid using spline interpolation.

2 Solar Radiation

Fu, P., Rich, P.M., 1999. Design and implementation of the Solar Analyst: an ArcView extension for modeling solar radiation at landscape scales. Proceedings of the 19th Annual ESRI User Conference, San Diego, USA. Available from http://proceedings.esri.com/library/userconf/proc99/proceed/papers/pap867/p867.htm.

The radiation term was calculated using the LiDAR elevation data up-scaled to 10 m as

$$S_i = \frac{S_{topo}}{S_{flat}},$$

where S_{topo} is direct shortwave radiation of the topographic surface calculated based on area latitude and topography and S_{flat} is direct radiation for a free flat surface where constant values of zero are used for slope and aspect. Both solar radiation datasets were computed on a monthly basis using an hourly time step, a sky view of 300 pixels, 32 calculation directions, 8 zenith and azimuth divisions, and uniform clear sky conditions.

3 Net Radiation

Net radiation was calculated for each month [MJ m^{-2} $month^{-1}$] as

$$R_n = S_{topo}(1-a) + L_n,$$

where S_{topo} is direct shortwave radiation of the topographic surface, a is albedo over the study area extracted from the MODIS MCD43A3 data product and L_n is net longwave radiation was calculated based on air temperature following Allen et al. (1998) as

$$L_n = \alpha T_i^4(0.34 - 0.14\sqrt{e_a})(1.35\, {}^{R_s}\!/_{R_{so}} - 0.35)$$

where α is Stefan-Boltzmann constant (4.903 10^{-9} MJ K^{-4} m^{-2} d^{-1}), T_i is locally modified temperature, e_a (*VP*) is actual vapor pressure, R_s is solar radiation and R_{so} is clear-sky solar radiation. In computation, we assumed that $R_s = R_{so}$.

4 Leaf Area Index

Leaf area index was derived using a vegetation index approach relating LAI and remotely sensed normalized difference vegetation index (NDVI). The 1-m resolution NAIP 4-band imagery dataset (red, blue, green, and near infrared spectra) was used as the base data for calculating LAI. NDVI was calculated from the NAIP near infrared (NIR) and red bands (Huete et al., 1994):

$$NDVI = \frac{(NIR - Red)}{(NIR + Red)} \, .$$

The polynomial function of Qi et al. (2000) derived for semiarid regions in southern Arizona was used to calculate LAI as:

$$LAI = ax^3 + bx^2 + cx + d,$$

where x is NDVI and a, b, c, and d are equal to 18.99, -15.24, 6.124, and -0.352, respectively. Computed 1 m LAI data were then resampled to the 10 m resolution of the DEM.

5 Locally Modified Temperature

Following Moore et al. (1993) mean monthly air temperature at each pixel (T_i) is calculated using the local lapse rate, topographically modified solar radiation, and leaf area index:

$$T_i = T_b - T_{lapse}\left[\frac{(z_i - z_b)}{1000}\right] + C\left(S_i - \frac{1}{S_i}\right)\left(1 - \frac{LAI_i}{LAI_{max}}\right), \qquad (^{o}\mathrm{C}) \qquad (1)$$

where T_b is temperature (^{o}C) at a base station - VCNP Headquarters Climate Station with the elevation of 2648.4 m, T_{lapse} is the local lapse rate (6.49 ^{o}C 1000 m^{-1}), z_i and z_b are the elevation (m) of the pixel and base station, respectively, C is a constant equal to 1, S_i is the ratio between direct shortwave radiation on the actual surface and direct shortwave radiation on a horizontal surface, LAI_i is pixel leaf area index and LAI_{max} is the maximum value for LAI equal to 10.

6 Vapor Pressure Deficit

Local monthly saturated vapor pressure (Pa) is computed as

$$VP_S = 611.2\, e^{\frac{17.67\, T_i}{T_i + 243.5}}$$

actual local vapor pressure (Pa) is computed as

$$VP_a = RH\, VP_S/100 \, ,$$

and local monthly vapor pressure deficit (Pa) is computed as

$$VP_d = VP_s - VP_a = (100 - RH) VP_S/100 \quad ,$$

where relative humidity *RH* is measured by the reference station and T_i is local temperature.

Relative Humidity and Dewpoint Temperature from Temperature and Wet-Bulb Temperature. Access from the NOAA website at *http://www.srh.noaa.gov/images/epz/wxcalc/rhTdFromWetBulb.pdf*.

Buck, A.L. (1981) New equations for computing vapor pressure and enhancement factor. Journal of Applied Meteorology, 20, 1527 – 1532.

7 Topographic Wetness Index

Topographic wetness index TWI_i (Beven and Kirkby, 1979) is computed for each pixel as

$$TWI_i = \ln(\frac{a_i}{tan\beta_i}) ,$$

where a_i is the upslope contributing area in square meters and β_i is the local slope. The contributing area was calculated using the D-Infinity multiple flow direction approach as described by Tarboton (1997) using a 1 m LiDAR dataset (Guo et al., 2010a) up-scaled to 10 m. Normalized wetness index α_i is computed as

$$\alpha_i = \frac{TWI_i}{\frac{1}{N}\sum TWI_i},$$

where N is number of pixels in catchment or study area. The normalization ensures conservation of mass of the effective precipitation term for a given catchment or area.

8 Evapotranspitation

Potential evapotranspiration was computed using the Penman-Montieth equation (Shuttleworth, 1993) and simplified for calculating potential evapotranspiration from a pan surface such that the surface resistance term (rs) in the denominator is assumed equal to zero

$$PET_{pm} = \frac{\Delta(R_n - G) + \rho_a c_p \frac{VP_d}{r_a}}{\lambda(\Delta + \gamma)},$$

The first term in the numerator is the radiation balance with net radiation R_n and ground heat flux G. The second term in the numerator is the ventilation term that includes vapor pressure deficit VP_d and aerodynamic resistance r_a computed as (Shuttleworth, 1993)

$$r_a = \frac{4.72\left(\ln\left(\frac{z_m}{z_o}\right)\right)^2}{1+0.536U_z},$$

where z_m is the height of meteorological measurements at 2 m, z_o is the aerodynamic roughness of an open water surface set equal to 0.00137 m following Thom and Oliver (1977), and U_z is wind speed. The remaining terms include the slope of the saturated vapor pressure-temperature relationship Δ calculated using mean air temperature as

$$\Delta = 0.04145 e^{0.06088\,T};$$

the psychrometric constant γ determined as

$$\gamma = c_p P/\varepsilon\lambda\,,$$

where c_p is specific heat of moist air at constant pressure 1.013 10^{-3} MJ kg^{-1} $^{\circ}C^{-1}$, ε is the ratio of molar mass of water to that of dry air, P is atmospheric pressure computed from measured values at the base station using elevation z locally estimated lapse rate η determined as

$$P = 101.3\left(\frac{293-\eta z}{293}\right)^{5.26}\,;$$

mean air density ρ_a , and λ the latent heat of evaporation of water.

Actual evapotranspiration AET was estimated using a Budyko curve (Budyko, 1974) describing the partitioning of potential and actual evapotranspiration relative to the aridity index (ratio of annual PET to annual rainfall). Potential evapotranspiration PET_{pm} and precipitation PPT were converted to monthly values of AET using a Zhang– Budyko curve as (Zhang et al., 2001)

$$AET = PPT\left\{1 + \frac{PET_{pm}}{PPT} - \left[1 + \left(\frac{PET_{pm}}{PPT}\right)^{w}\right]^{-1/w}\right\}$$

where w is an empirical constant, here set equal to 2.63.

9 Local Water Balance (Water Redistribution)

The pixel wetting was approximated using a local water balance as (L'Vovich, 1979):

$$W = PPT - SR = AET + F,$$

where W is subsurface pixel wetting, PPT precipitation, AET actual evapotranspiration, F water partitioned to baseflow, and SR surface runoff. The F term quantifies subsurface wetting, a key parameter for calculating $EEMT$, and represents the fraction of water with ability to perform work on the subsurface. The subsurface wetting can be computed as

$$F = P_{eff} - SR$$

where P_{eff} is effective precipitation equivalent to $P_{eff} = PPT - AET$. Using normalized TWI, the value of subsurface wetting was estimated as

$F = \alpha P_{eff}$.

10 Effective energy and mass transfer (EEMT)

Monthly EEMT topo in $MJ.m^{-2}$ is defined as (Rasmussen et al., 2011)

$EEMT = E_{ePPT} + E_{BIO}$.

The monthly heat and mass transfer associated with effective precipitation is computed as

$E_{ePPT} = F c_w \Delta T$

where F is subsurface wetting, c_w is the specific heat of water and $\Delta T = T_{local} - T_{ref}$ with T_{ref} set to 273.15 K.

The net primary productivity energy and mass transfer is computed as

$E_{BIO} = NPP\ h_{BIO}$,

where NPP is the mass flux of C as net primary production and h_{BIO} is the specific biomass enthalpy fixed at a value of 22×10^6 J kg^{-1}. NPP is computed as (Whittaker and Niering, 1975)

$NPP = 0.39z + 346n - 187$,

where z is elevation and n is northness, a unitless parameter computed as the product of the cosine of aspect and the sine of slope.

Yearly EEMT in $MJ.m^{-2}$

$EEMT_{topo} = \sum_{i=1}^{12} EEMTm_i$

For more details about theory and computation, see (Rasmussen et al., 2015):

Rasmussen C., Pelletier J.D., Troch P.A., Swetnam T.L., and Chorover J. (2015): Quantifying Topographic and Vegetation Effects on the Transfer of Energy and Mass to the Critical Zone. Vadose Zone Journal 14 (11). DOI: 10.2136/vzj2014.07.0102

11 References

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