# Chapter 5. WATER BALANCE AND PERCOLATION

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## 5.1 Introduction

The water balance and percolation components of the WEPP model are designed to use input from the climate, infiltration, and crop growth components to estimate soil water content in the root zone and evapotranspiration losses throughout the simulation period. The time step in predicting evapotranspiration and percolation is 24 hours. The WEPP water balance uses many of the algorithms developed for the SWRRB (Simulator for Water Resources in Rural Basins) model by Williams et al. (1985). Some modification has been made to improve estimation of rainfall interception, percolation and soil evaporation parameters.

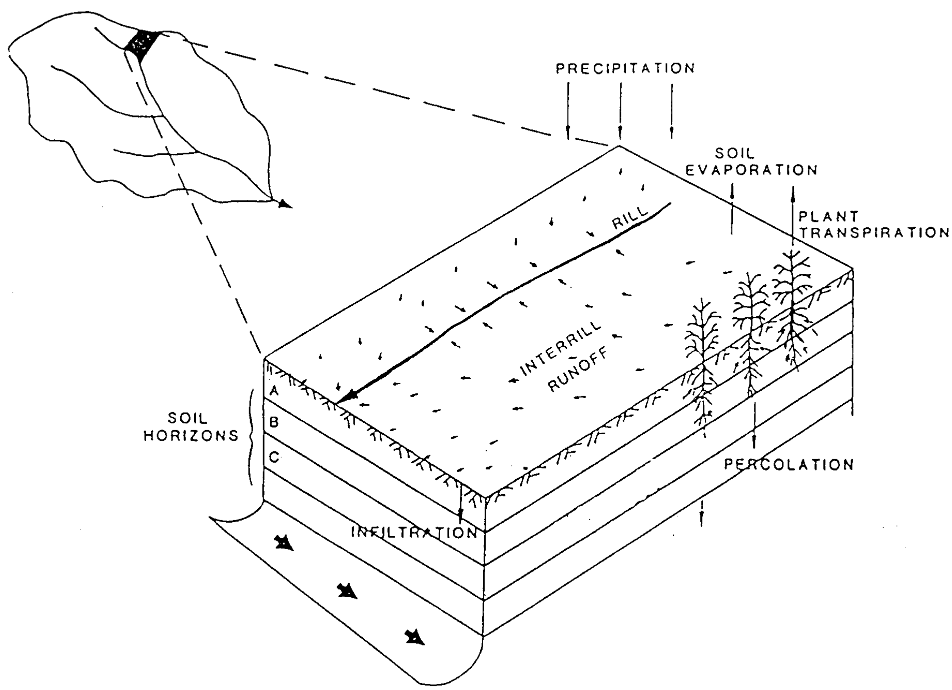


Figure 5.1.1. Processes in WEPP hillslope hydrology include precipitation (rain or snow), infiltration, runoff, plant transpiration, soil evaporation and percolation.

The hydrologic processes in WEPP hillslope model include infiltration, runoff routing, soil evaporation, plant transpiration, snowmelt, and seepage (Fig. 5.1.1). The model maintains a continuous water balance on a daily basis using the equation:

$$\Theta = \Theta_{in} + (P - I) \pm S - Q - ET - D - Q_d \qquad [5.1.1]$$

where $\Theta$ is the soil water content in the root zone in any given day ($m$), $\Theta_{in}$ is the initial soil water in the root zone ($m$), $P$ is the cumulative precipitation ($m$), $I$ is precipitation interception by vegetation ($m$), $S$ is the snow water content ($m$) ( (+) for snowmelt and it equals daily snowmelt, (-) snow accumulation (see Chapter 3), $Q$ is the cumulative amount of surface runoff ($m$), $ET$ is the cumulative amount of evapotranspiration ($m$), $D$ is the cumulative amount of percolation loss below the root zone ($m$), and $Q_d$ is subsurface lateral flow or flow to drain tiles ($m$). Precipitation interception by vegetation is calculated using the method described by Savabi and Stott (1994).

$$I = .001(3.7\ VE - 1.10\ x\ 10^{-4}\ VE^2) \qquad [5.1.2]$$

where VE is above ground biomass in $kg{\cdot}m^{-2}$.

Precipitation is partitioned between rainfall and snowfall using air temperature. For a day on which the maximum temperature is below $0^oC$, all precipitation is assumed to be snow, and for a day on which the minimum temperature is above $0^oC$, all precipitation is assumed to be rain. For the days on which minimum temperature is below $0^oC$ and maximum temperature is above $0^oC$, the time of precipitation occurrence within the day is randomly predicted and used to determine if a particular hour period is experiencing rainfall or snowfall, based upon a constructed diurnal temperature function. If the majority of precipitation is predicted to occur as rainfall on bare soil, the entire storm is assumed to occur as rainfall on bare soil on an individual overland flow element (OFE). An OFE is a region on a hillslope of homogeneous soil, cropping and management.

Accumulated snowpack will be subject to evaporation and melt (see Chapter 3, Winter Hydrology). Soil evaporation is considered first to come from the snowpack, if present, and then from the soil. Snow is melted on days when the maximum temperature exceeds zero degree Celsius. Melted snow is treated in the water balance Eq. [5.1.1] as rainfall for estimating runoff and percolation.

## 5.2 Evapotranspiration

The evapotranspiration component of WEPP is a modified Ritchie's model (Ritchie, 1972). However, depending on meteorological data availability, two options are given to users to estimate reference potential ET.

In the case where daily radiation, temperature, wind and dew point temperature or relative humidity data are available or all generated by the CLIGEN program for the United States, the WEPP model uses the Penman equation with the original wind function method (Penman, 1963; and Jensen 1974):

$$E_u = \frac{\delta}{\delta+\gamma}(R_n - G) + \frac{\gamma}{\gamma+\delta}\ 6.43\ (1.0 + 0.53\ u_{z)}\ (e_z^o - e_z) \qquad [5.2.1]$$

where $E_u$ = daily potential evapotranspiration ($MJ{\cdot}m^{-2}{\cdot}d^{-1}$), $\delta$ = slope of the saturated vapor pressure curve at mean air temperature, $\gamma$ = psychrometric constant, $G$ = soil heat flux ($MJ{\cdot}m^{-2}{\cdot}d^{-1}$), $R_n$ = net radiation ($MJ{\cdot}m^{-2}{\cdot}d^{-1}$), $u_z$ = wind speed ($m{\cdot}s^{-1}$), $e_z^o$ = saturated vapor pressure, ($KPa$), $e_z$ = vapor pressure, ($KPa$). $E_u$ is converted to meter per day by dividing it by $2.501 - 2.361{*}10^{-3}\ T$, where $T$ is average air temperature (Harrison, 1963).

In the case where only solar radiation and temperature data are available, the model uses the Priestly-Taylor (1972) method:

$$E_u = 0.00128\ \frac{R_n l}{58.3}\ \frac{\delta}{\delta+\gamma} \qquad [5.2.2]$$

where $R_n l$ = daily net solar radiation ($ly$). Net radiation in equation 5.2.1 and 5.2.2 are calculated by multiplying the incoming daily solar radiation by $(1 - A)$, where $A$ is albedo (0 - 1.0).

The albedo is evaluated by considering the soil, crop, and snow cover. If a snow cover exists with at least 0.005 m water content, the value of albedo is set to 0.80, otherwise the soil albedo is used. The albedo is estimated during the growing season using the equation:

$$A = 0.23\,(1. - C_f) + (A_s)\,C_f \qquad [5.2.3]$$

where 0.23 is the plant albedo, $C_f$ is the soil cover index (0 - 1.0), and $A_s$ is the soil albedo.

The value of $C_f$ is calculated using the equation:

$$C_f = e^{(-0.000029\,C)} \qquad [5.2.4]$$

where $C$ is the sum of above ground biomass and plant residue ($kg\ ha^{-1}$), determined in the crop growth component.

The value of $\delta$ in Eqs. [5.2.1 and 5.2.2] is determined from:

$$\delta = \frac{5304}{T_k^2}\, e^{\left[21.25 - \frac{5304}{T_k}\right]} \qquad [5.2.5]$$

where $T_k$ is the daily average air temperature, degrees Kelvin. The psychrometric constant is computed with the equation $\gamma = 6.6 * 10^{-4}\, PB$ where $PB$ is barometric pressure ($KPa$). The barometric pressure is calculated by

$$PB = 101 - 0.0115\, he + 5.44 * 10^{-7}\, he^2 \qquad [5.2.6]$$

where $he$ is the elevation of the site ($m$).

The soil heat flux is estimated by using air temperature (deVries, 1963). Saturated vapor pressure, in $KPa$ is calculated using the equation

$$e_z^o = \exp \frac{16.78\, T - 116.9}{T + 237.3} \qquad [5.2.7]$$

where $T$ is the average daily temperature in $^oC$. The vapor pressure is calculated by using the dew point temperature in Eq. 5.2.7.

Potential soil evaporation, $E_{sp}$, and plant transpiration, $E_{tp}$, are predicted (Fig. 5.2.1.b) with the equations:

$$E_{sp} = E_u\, e^{(-0.4\, L)} \qquad [5.2.8]$$

$$E_{tp} = (1 - E_{sp}/E_u) * E_u \qquad [5.2.9]$$

where $L$ is the leaf area index defined as the area of plant leaves relative to the soil surface area.

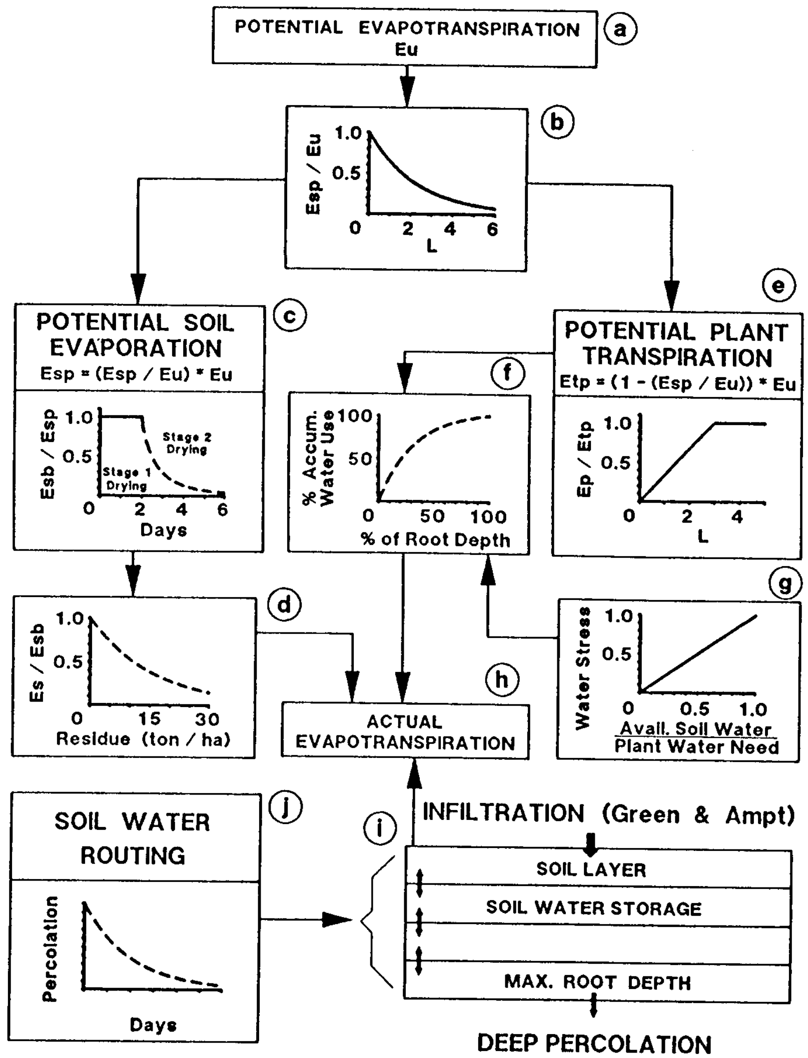


Figure 5.2.1. Schematic computational sequence of the WEPP evapotranspiration and soil water redistribution. $E_u$ is daily potential evapotranspiration.

Bare soil evaporation, $E_{sb}$, is calculated in two stages (Fig. 5.2.1.c). In the first stage, soil evaporation is limited only by the energy available at the soil surface and, therefore, it is equal to potential soil evaporation, $E_{sp}$. The upper limit for the stage one soil evaporation is calculated using the equation (Ritchie, 1972):

$$E_{su} = 0.009\ (T_r - 3.0)^{\,0.42} \qquad [5.2.10]$$

where $E_{su}$ is the upper limit soil evaporation of stage one ($m$), and $T_r$ is the soil transmissivity ($mm \cdot d^{-0.5}$), dependent on soil texture:

$$T_r = 4.165 + 0.02456\, S_a - 0.01703\, C_l - 0.0004\, S_a^2 \qquad [5.2.11]$$

where $S_a$ is the percentage of sand in the bare soil evaporated layer and $C_l$ is the percentage of clay in the bare soil evaporated layer.

When the accumulated soil evaporation exceeds the stage one upper limit, $E_{su}$, stage two evaporation begins. Stage two soil evaporation is estimated using the equation:

$$S_2 = 0.001\, T_r\, [d_2^{1/2} - (d_2 - 1)^{1/2}] \qquad [5.2.12]$$

where $S_2$ is the stage two bare soil evaporation rate for a day ($m \cdot d^{-1}$) and $d_2$ is the number of days since stage two soil evaporation began.

If precipitation is greater than or equal to accumulated stage two soil evaporation, then stage one soil evaporation is assumed. For more details see Ritchie (1972). During a drying cycle, evaporation from the soil continues until the soil water content is at a residual moisture content, $\Theta_r$, a moisture content below which no more water can be evaporated from the bare soil. $\Theta_r$ is calculated using soil organic matter, percent clay, and soil bulk density (see Chapter 7 for more detail). Computed bare soil evaporation, $E_{sb}$, in either stage is reduced with increased plant residue (Fig. 5.2.1.d) using the equation

$$E_s = E_{sb}\, e^{(-0.000064\, C_r)} \qquad [5.2.13]$$

where $E_s$ = actual soil evaporation ($m \cdot d^{-1}$), $E_{sb}$ = bare soil evaporation ($m \cdot d^{-1}$), and $C_r$ = plant residue on soil ($kg \cdot ha^{-1}$) (data from J. L. Steiner, personal communication).

Adjusted potential plant transpiration is computed as a linear function of $L$ and $E_{tp}$ up to $L$ of 3.

$$E_{tp} = \frac{E_{tp}\, L}{3} \qquad L \le 3 \qquad [5.2.14]$$

where $E_{tp}$ is the daily adjusted potential plant transpiration, ($m \cdot d^{-1}$). Beyond $L = 3$, potential plant transpiration is not adjusted for leaf area index (Fig. 5.2.1.e).

## 5.3 Distribution of Evapotranspiration in the Root Zone

The distribution of calculated soil evaporation, $E_s$, in the root zone is determined by considering snow cover and soil water content of the effective depth influenced by bare soil evaporation, $d_x$. If the water content of the snow cover is equal to or greater than $E_s$, all the soil evaporation comes from the snow cover. If $E_s$ exceeds the water content of the snow cover, the difference will be removed from soil water. The depth of the soil where water is evaporated, $d_s$, is predicted with the equation:

$$d_s = d_x \frac{E_s}{\Theta} - \Theta_r\, d_x \qquad E_s \le \Theta - (\Theta_r\, d_x) \qquad [5.3.1]$$

$$d_s = d_x \qquad E_s > \Theta - (\Theta_r\, d_x)$$

where $d_s$ is the soil evaporated depth of any given day ($m$), $d_x$ is the maximum soil evaporated depth influenced by soil evaporation ($m$) (Lane and Stone, 1983), $E_s$ is the predicted daily actual soil evaporation ($m$), $\Theta$ is the soil water content of the soil layers above $d_x$ ($m$), and $\Theta_r$ is the residual moisture content, fraction by volume ($m^3 \cdot m^{-3}$).

The maximum soil evaporation effective depth, $d_x$, is calculated based on soil texture with the equation:

$$d_x = 0.09 - 0.00077\ C_l + 0.000006\ S_a^2 \qquad [5.3.2]$$

If the water content in the $d_x$ depth is not sufficient for calculated soil evaporation ($E_s$), soil evaporation is reduced accordingly.

The potential plant transpiration is distributed in the root zone, *RZ*, with the equation:

$$U_{Pi} = \frac{E_{tp}}{1 - e^{(-V)}} \left( 1 - e^{\left[ -V \frac{h_i}{RZ} \right]} \right) - \sum_{j=1}^{i-1} U_j \qquad [5.3.3]$$

where $U_{Pi}$ is the potential water use rate from layer $i$ ($m \cdot d^{-1}$), and $V$ is a use rate-depth parameter (3.065 is used in WEPP, assuming about 30 percent of the total water use comes from the top 10 percent of the root zones). The details of evaluating $V$ are given by Williams and Hann (1978). $h_i$ is the depth of soil layer $i$ in $m$, $RZ$ is the root zone depth ($m$), and $U$ is the actual water use from the soil layer above layer i ($m$).

The potential water use, $U_{Pi}$, is adjusted for water deficits to obtain the actual water use, $U_i$, for each layer.

$$U_i = U_{Pi} \qquad \Theta_i > \Theta_c\ UL_i \qquad [5.3.4]$$

$$U_i = U_{Pi}\ \frac{\Theta_i}{\Theta_c\ UL_i} \qquad \Theta_i \le \Theta_c\ UL_i$$

where $\Theta_i$ is soil water content of layer $i$ ($m$), and $\Theta_c$ is a critical soil water content below which plant growth is subjected to water stress, fraction by volume ($m^3 \cdot m^{-3}$). $\Theta_c$ is a crop dependent parameter provided by the user. The default value is 0.25. $UL_i$ is the upper limit soil water content for layer $i$, ($m$).

Equation [5.3.4] allows roots to compensate for water deficits in some layers by using more water in another layer having adequate supplies.

### 5.4 Percolation

The percolation component of WEPP uses storage routing techniques to predict flow through each soil layer in the root zone. In addition to percolation, the WEPP model simulates subsurface lateral flow and flow to drainage tile and ditches (See Chapter 6 for more detail). In each layer, water content exceeding the corresponding field capacity is subjected to percolation through the succeeding layer. Water moving below the root zone is considered lost and will not be traced. Saturated hydraulic conductivity is being calculated for each layer based on soil physical properties such as soil texture, organic matter and porosity. Flow through a soil layer may be reduced by coarse fragments in the layer, frozen layer, and saturated or nearly saturated lower layer.

Percolation of water in excess of field capacity from a layer is computed using the equation:

$$pe_i = (\Theta_i - FC_i)\ [1 - e^{\left[ \frac{-\Delta t}{t_i} \right]}] \qquad \Theta_i > FC_i \qquad [5.4.1]$$

$$pe_i = 0 \qquad \Theta_i \le FC_i$$

where $pe_i$ is the percolation rate through layer $i$ ($m \cdot d^{-1}$), $FC_i$ is the field capacity water content (at 33 KPa of tension for many soils) for layer $i$ ($m$), $\Delta t$ is the travel interval ($s$), and $t_i$ is the travel time through layer $i$ ($s$).

The travel time through a particular layer is computed with the linear storage equation:

$$t_i = \frac{\Theta_i - FC_i}{K_{sai}} \qquad [5.4.2]$$

$K_{sai}$ is the adjusted hydraulic conductivity of layer $i$ ($m \cdot s^{-1}$).

The hydraulic conductivity is varied from the saturated conductivity, $K_s$, value at saturation to near zero at field capacity.

$$K_{sai} = K_{si} \left[ \frac{\Theta_i}{UL_i} \right]^{B_i} \qquad [5.4.3]$$

where $K_{si}$ is the saturated hydraulic conductivity for layer $i$ ($m \cdot s^{-1}$) and $B_i$ is a parameter that causes $K_{sai}$ to approach zero as $\Theta_i$ approaches $FC_i$.

$$B_i = \frac{-2.655}{\log \frac{FC_i}{UL_i}} \qquad [5.4.4]$$

The constant -2.655 in Eq. [5.4.4] assures a $K_{sai}$ of $0.002 * K_{si}$ at field capacity.

The computation of saturated hydraulic conductivity of each layer, $K_{si}$, and adjustments for rocks, frozen ground and entrapped air are presented in Chapter 7.

Flow through a soil layer may be restricted by a lower layer which is at or near saturation. The effect of lower layer water content is given in the equation:

$$pe_i = pe_i \sqrt{1 - \frac{\Theta_{i+1}}{UL_{i+1}}} \qquad [5.4.5]$$

where $pe_i$ is the percolation rate adjusted for lower layer ($i+1$) water content ($m \cdot d^{-1}$).

**5.5 Linkage of Water Balance and Percolation Components with the Other WEPP Components**

The infiltration component of WEPP is linked with the evapotranspiration and percolation components (Fig. 5.2.1) to maintain a continuous water balance. Infiltrated water is added to the upper layer's soil water content and routed through the lower soil layers. Soil water in each layer is subjected to percolation and/or evapotranspiration (Fig. 5.2.1). The upper layer soil water content is being used to establish initial moisture conditions for the infiltration component (Green and Ampt model). Percolation below the root zone is considered lost from the WEPP water balance.

Daily leaf area index, root depth, total plant biomass and residue cover are entered as input to the evapotranspiration component from the crop growth component. The plant growth water stress factor is computed by considering supply and demand in the equation:

$$W_s = \frac{\sum_{i=1}^{n} U_i}{E_{tp}} \qquad [5.5.1]$$

where $W_s$ is the plant growth water stress factor (0 - 1.0), $U_i$ is actual water use from layer i ($m \cdot d^{-1}$), $n$ is number of soil layers, and $E_{tp}$ is the potential plant transpiration ($m \cdot d^{-1}$).

The water stress factor, $W_s$, is used in the WEPP plant growth component to adjust daily plant growth.

## 5.6 Model Validation

The water balance component of the WEPP hillslope was evaluated using data from a tall grass prairie watershed, near Manhattan, Kansas (Savabi et al., 1989). The model was tested independently from the WEPP hillslope model, therefore, measured infiltration ($L$), plant biomass, and residue cover were used in the validation. The other input data included daily maximum and minimum temperatures, solar radiation, as well as watershed soil physical properties of the root zone.

### 5.6.1 Watershed Description and Field Measurements

Watershed 1D (37.7 ha) at the Konza prairie near Manhattan, Kansas was selected for this study (Fig. 5.6.1). The soil of the watershed is classified as Benfield-Florence Complex, which consists of Benfield silty clay and Florence cherty silt loam. The soil is well drained and has low available water capacity. Annual rainfall is about 86 cm with about 75 percent of the moisture falling during the growing season (May to August).

The native vegetation, according to Anderson and Fly (1955), are mid-grasses, such as little bluestem (Andropogon scoparius), side oats grama (Bouteloua curtipendula), and Kentucky bluegrass (Poa Pratensis), together with tall grasses including big bluestem (Andropogon furcatus), indiangrass (Sorghastrum nutans), and switchgrass (Panicum virgatum).

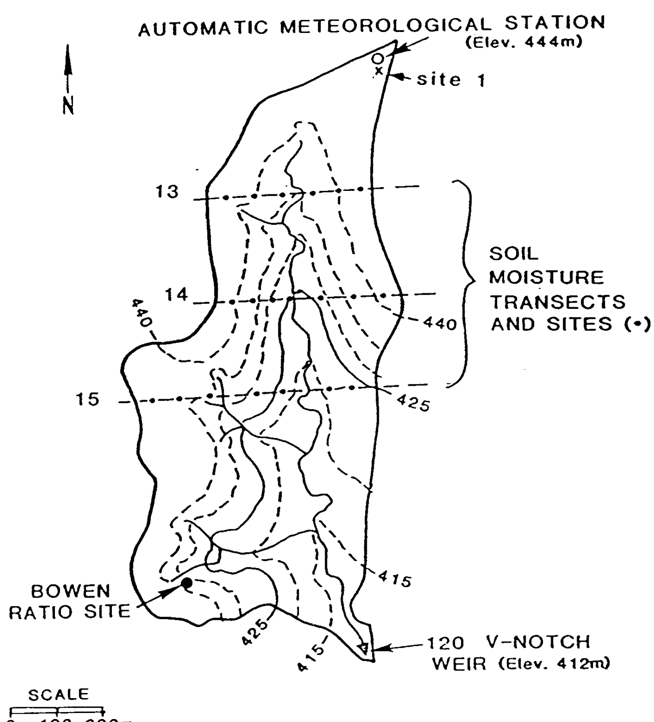


Figure 5.6.1. Watershed 1D, Konza Prairie and location of the automatic meteorological, Bowen ratio stations and soil moisture transects.

Rainfall, maximum and minimum temperatures, and net radiation were measured by automatic meteorological stations, and streamflow was measured with a sharp crested weir and a clock driven analog recorder (Fig. 5.6.1) Near surface (.05 m) soil moisture was measured gravimetrically across three transects (Fig. 5.6.1). Soil moisture content was also measured periodically to a depth of 2 meters during the 1987 growing season using neutron meter techniques. Five neutron access tubes were installed at site 1 near the automatic meteorological station (Fig. 5.6.1).

Actual evapotranspiration was estimated using the Energy Balance Bowen Ratio (EBBR) method (Bowen, 1926; Slatyer and McIlroy, 1961). Several studies suggest that EBBR estimates of *ET* are in good agreement with lysimeter measurement in nonadvective conditions (Tanner, 1960; Pruitt and Lourence, 1968; and Denmead and McIlroy, 1970). The method involves determining the latent heat flux through solution of the energy balance equation,

$$LE = \frac{R_n - G}{1 + \gamma \dfrac{dt}{de}} \qquad [5.6.1]$$

where $LE$ is the latent heat flux ($W \cdot m^{-2}$), $R_n$ is the net radiation ($W \cdot m^{-2}$), $G$ is the soil heat flux ($W \cdot m^{-2}$), $\gamma$ is the psychrometric constant, and $dt$ and $de$ are the air temperature ($^oK$) and vapor pressure ($Pa$) differences at two heights above the plant canopy, respectively.

At the Bowen ratio site (Fig. 5.6.1), $R_n$, $G$, $dt$, and $de$ were measured by an automatic Bowen ratio system every 30 minutes during the 1987 growing season. $LE$ was determined every 30 minutes and integrated over 24 hours to determine daily $LE$. Daily $LE$ was converted to a depth of evaporated water (1 $m$ water equals 676000 $W \cdot m^{-2}$ at 25 $^oC$).

Leaf area index of live vegetation and plant residue ($kg \cdot ha^{-1}$) were among several biophysical measurements made periodically on the watershed.

The model was tested using the measured data of the 1987 growing season. No calibration was performed. The model-simulated *ET* was compared with EBBR-*ET*. In addition, model-simulated and measured soil water contents were compared.

### 5.6.2 Results and Discussion

Daily model-simulated *ET* is compared with EBBR-*ET* using least square analysis (Fig. 5.6.2). The calculated coefficient of determination is 0.67 and is significant at the 0.05 probability level. The intercept and the slope of the regression equation between daily model-simulated and EBBR-*ET* are not different from zero and unity, respectively, at the 0.05 probability level, which indicates statistically a good agreement between the WEPP model *ET* calculations and EBBR-*ET*.

Percent model-simulated and average field measured soil water content for the top 0.05 $m$ of the soil are shown in Fig. 5.6.3. The average field measured soil water content values are the arithmetic mean of all measurements within the three transects (13, 14, and 15, Fig. 5.6.1). The calculated standard error between model simulated and average field measured soil water content is 0.002 m. In the model, infiltrated water is added to the surface soil layer where it is subjected to percolation to lower layers, evaporation from the surface soil layer, and transpiration from the root zone by plants. Good agreement between simulated and measured near surface soil water content indicates that the model is capable of predicting antecedent soil water content for the infiltration component of the WEPP model with reasonable accuracy.

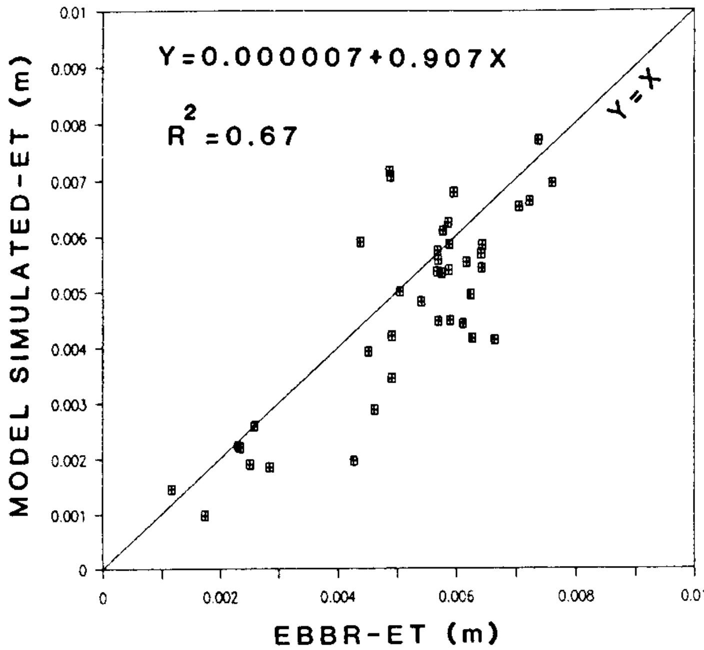


Figure 5.6.2. Least square analysis between WEPP simulation *ET* and estimated *ET* using EBBR method on watershed 1D during 1987 growing season.

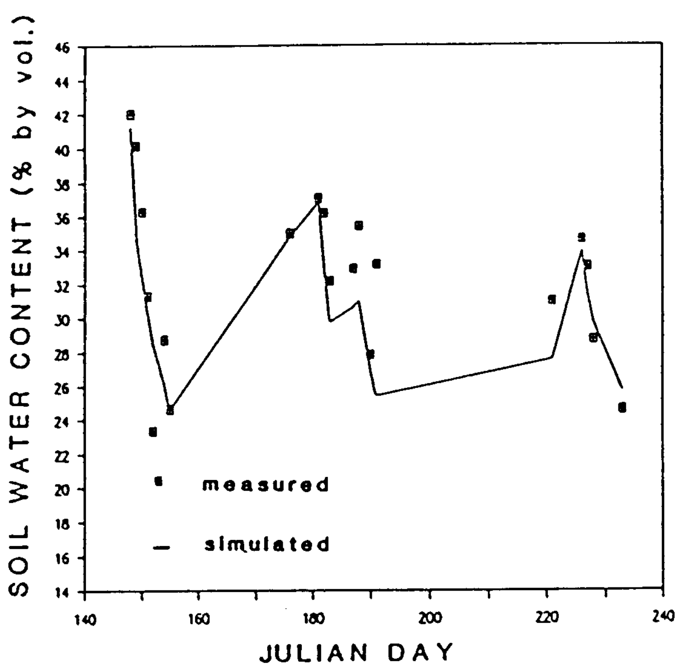


Figure 5.6.3. Comparison of model-simulated and measured soil water content of top 5 cm soil. The measured values are the arithmetic mean of all measurements within the transects 13, 14, and 15.

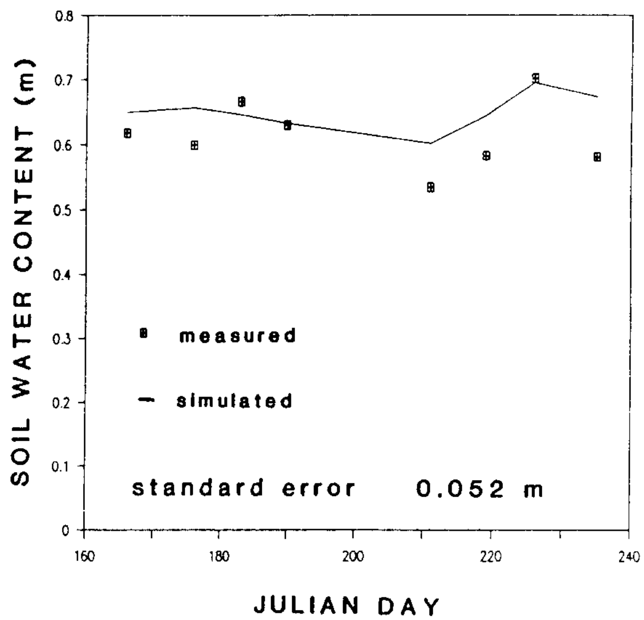


Figure 5.6.4. Comparison of model-simulated and average neutron probe measured soil water content (site 1) for the 200 cm soil depth.

Fig. 5.6.4 shows a comparison between the model-simulated and average measured soil water content from the surface to 2 m depth. The average measured soil water content values are the arithmetic average of the five access tubes at site 1 (Fig. 5.6.1). Calculated standard error is 0.052 m of soil water. Except for three days, measured soil water contents lay below model-simulated water content for the 1987 growing season (Fig. 5.6.4).

The reasons for such deviations can be several including the fact that only five neutron meter access tubes were installed in the vicinity of watershed 1D and were assumed to represent the entire watershed. This assumption may not be valid because the access tubes are located at the north end of the watershed (site 1) where the elevation is greater than the entire watershed 1D (Fig. 5.6.1). The same argument can be used for the EBBR-*ET* measurement; however, there is a good agreement between simulated and measured water content of the top .05 m of soil which is more representative of the watershed soil water content (Fig. 5.6.1). Hence, the *ET* values given by the model are probably representative of the *ET* rate of the watershed.

## 5.7 Summary and Conclusions

The WEPP water balance is designed to simulate soil water evaporation, plant transpiration, and root zone soil water content. The model uses many algorithms from the SWRRB model. The WEPP model was tested using measured data from watershed 1D in the Konza natural prairie for the 1987 growing season. Comparison of model-simulated and measured *ET* and soil water content is presented. The comparison of model-simulated *ET* and EBBR-*ET* indicates that the model estimates are representative of watershed 1D. This is further supported by a good agreement between the model-simulated and the field measured soil water content of the top 0.05 m of the soil profile. In addition, it indicates that the model is able to simulate antecedent soil water content which is used in the infiltration component of WEPP (Chapter 4). The comparison of simulated and measured soil water content of the soil surface to a 2 meter depth was less than desirable. Considering the size and topography of watershed 1D, the set of soil moisture measurements at site 1 may not be well representative of the entire watershed.

## 5.8 Acknowledgements

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## 5.10 List of Symbols

| Symbol | Definition | Units | Variable |
|---|---|---|---|
| $A$ | albedo (0 - 1.0) | $NOD$ | alb |
| $A_s$ | soil albedo (0 - 1.0) | $NOD$ | salb |
| $B_i$ | conductivity adjustment parameter for layer $i$ (0 - 1.0) | $NOD$ | hk |
| $C$ | sum of plant biomass and plant residue | $kg{\cdot}ha^{-1}$ | $c_v$ |
| $C_f$ | soil cover index (0 - 1.0) | $NOD$ | eaj |
| $C_l$ | percent clay content | % | - |
| $C_r$ | plant residue | $kg{\cdot}ha^{-1}$ | resamt |
| $D$ | cumulative amount of percolation loss | $m$ | sep |
| $d_s$ | soil evaporated depth | $m$ | esd |
| $dt$ | air temperature difference at 2 heights above canopy | $^oK$ | - |
| $de$ | vapor pressure difference at 2 heights above canopy | $Pa$ | - |
| $d_x$ | maximum soil evaporated depth influenced by bare soil evaporation. | $m$ | esb |
| $d_2$ | days since stage two soil evaporation began | $d$ | tv |
| $E_s$ | actual daily soil evaporation | $m{\cdot}d^{-1}$ | $e_s$ |
| $E_{sb}$ | bare soil evaporation | $m{\cdot}d^{-1}$ | $e_s$ |
| $E_{sp}$ | potential soil evaporation | $m{\cdot}d^{-1}$ | es |
| $E_{su}$ | upper limit soil evaporation of stage one | $m{\cdot}d^{-1}$ | tu |
| $ET$ | cumulative amount of evapotranspiration | $m$ | estep |
| $E_{tp}$ | potential plant transpiration | $m{\cdot}d^{-1}$ | ep |
| $E_u$ | potential evapotranspiration ($md^{-1}$) | | |
| $FC_i$ | field capacity water content | $m$ | fc |
| $G$ | soil heat flux | $W{\cdot}m^{-2}$ | - |
| $h_i$ | depth of soil layer | $m$ | solthk |
| $he$ | elevation | $m$ | elev |
| $i$ | soil layers | - | i |
| $K_{sai}$ | adjusted hydraulic conductivity for layer $i$ | $m{\cdot}s^{-1}$ | SSC |
| $K_{si}$ | saturated hydraulic conductivity for layer $i$ | $m{\cdot}s^{-1}$ | SSC |

| | | | |
|---|---|---|---|
| $L$ | leaf area index | $m^2 \cdot m^{-2}$ | LAI |
| $LE$ | latent heat flux | $W \cdot m^{-2}$ | - |
| $p$ | cumulative rainfall | $m$ | rain |
| $pe_i$ | percolation rate through layer $i$ | $m \cdot d^{-1}$ | sep |
| $Q$ | cumulative amount of surface runoff | $m$ | runoff |
| $R_l$ | daily solar radiation | $ly$ | radly |
| $R_n$ | net daily solar radiation | $MJ \cdot m^{-2} \cdot d^{-1}$ | radmj |
| $RZ$ | root zone depth | $m$ | rtd |
| $S$ | snow water content | $m$ | sno |
| $S_a$ | percentage of sand | % | - |
| $S_2$ | stage two soil evaporation | $m \cdot d^{-1}$ | S2 |
| $t_i$ | travel time through layer $i$ | $s$ | xx |
| $T_k$ | daily average air temperature | $^oK$ | tk |
| $T_r$ | soil transmissivity parameter | $mm \cdot d^{-0.5}$ | trans |
| $U_i$ | actual water use from layer $i$ | $m \cdot d^{-1}$ | u |
| $UL_i$ | upper limit soil water content for layer $i$ | $m$ | ul |
| $U_{Pi}$ | potential water use rate for layer $i$ | $m \cdot d^{-1}$ | U |
| $V$ | a use rate - depth parameter | | ub |
| $W_s$ | plant growth water stress factor (0 - 1.0) | $NOD$ | watstr |
| $\delta$ | slope of the saturated vapor pressure curve | - | d |
| $\gamma$ | psychrometric constant | $Pa \cdot {}^oK^{-1}$ | gma |
| $\Delta t$ | percolation travel interval | $s$ | 84600 |
| $\Theta$ | water content of the root zone | $m$ | watcon |
| $\Theta_c$ | critical soil water content below which plants growth is subjected to water stress | (0 - 1.0) | pltol |
| $\Theta_i$ | soil water content in each layer | $m$ | soilwa |
| $\Theta_{in}$ | initial water content | $m$ | watcon |
| $\Theta_r$ | residual moisture content | $m^3 \cdot m^{-3}$ | wrd |

Note - *NOD* indicates nondimensional variable.