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Modification of the Evapotranspiration Routines in the WEPP Model: Part I

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Abstract. *The physical processes of evaporation and transpiration, collectively termed evapotranspiration, are discussed with respect to the unique conditions specific to forested environments. Forests have significant variations in ET rates due to 1) diurnal, seasonal, and annual climatic fluctuations; 2) spatiotemporal differences in vegetation; 3) evaporation of precipitation intercepted by vegetation, litter, and soil; 4) evaporation from water bodies; and 5) physiographic differences. The earliest methods for computing ET relied on empirical relations between climatic variables and consumptive water use by crops. Later formulations derived potential evaporation by relating solar radiation and temperature to the physical process of latent and sensible heat flux. To generalize Penman's equation for crops that were water-stressed, Monteith incorporated a canopy resistance term to describe the effect that partially closed stomates have on evapotranspiration. Later researchers have modified these equations to account for variable crop density, rainfall interception, bare-soil evaporation, and multiple canopy layers. WEPP primarily uses a modification of Ritchie's method to compute ET. Although WEPP gives the user the option to use either Penman's, Priestly-Taylor's, Hargraves', or Penman-Monteith's equations for calculating ET, the*

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coding for Hargraves' and Penman-Monteith's equations are incomplete, and are therefore turned off. The WEPP model adequately accounts for seasonal and climatic fluctuations, spatiotemporal difference in vegetation, and physiographic differences. Recommended improvements to the WEPP model's ET routine are: 1) completing the coding of the Penman-Monteith equation, 2) computing evaporation of intercepted precipitation, and 3) computing evaporation from water bodies and litter.

Keywords. Evapotranspiration, Penman-Monteith, WEPP, Water Erosion Prediction Project

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Introduction

Evapotranspiration (ET) constitutes an important component of the water fluxes of our hydrosphere and atmosphere. Approximately two-thirds of all rain falling in continental United States is returned to the atmosphere by evapotranspiration processes (Douglass, 1966). ET rates are affected by complex spatial variations in climate, topography, and vegetative cover (Biftu and Gan, 2000). ET estimation is further complicated by complex temporal variations, including the diurnal and seasonal variability of evaporative fluxes (Biftu and Gan, 2000). Recently there has been a growing interest in estimating ET for a combination of land-use classes and for sparsely vegetated surfaces (Biftu and Gan, 2000). Of particular interest here, are the effects of ET (and changes in ET and the water balance) on rates of surface erosion following vegetation modification (or management).

For half a century, there have been many attempts to model evaporation and/or ET for climatological, agronomical, and hydrological purposes (Allen *et al.,* 1998; Biftu and Gan, 2000; Federer *et al.,* 1996; Flerchinger *et al.,* 1996; Murakami *et al.,* 2000; Pereira *et al.,* 1999; Stannard, 1993; Wigmosta *et al.,* 1994; Ziemer, 1979). Historically, the majority of evapotranspiration models were developed for well-watered agricultural crops. The earliest methods relied on empirical relations between climatic variables and consumptive water use by crops (e.g., Thornthwaite, Blaney-Criddle, pan evaporation). Later formulations derived potential (or maximum) evaporation by relating solar radiation and temperature to the physical process of latent and sensible heat flux (e.g., Penman, Jensen-Haise) (Stannard, 1993). To generalize the Penman (1948) equation for crops that were water-stressed, Monteith (1965) incorporated a canopy resistance term to describe the effect that partially closed stomates have on evapotranspiration (Stannard, 1993). Later researchers have modified these equations to account for variable crop density (Ritchie, 1972), rainfall interception (Rutter and Morton, 1977), bare-soil evaporation (Shuttleworth and Wallace, 1985), and multiple canopy layers (Choudhury and Monteith, 1988).

Evaluating ET in forests is complicated by many factors that are not usually relevant for agricultural landscapes. To evaluate ET in forests the following factors should be addressed:

- Variations in ET rates due to diurnal, seasonal, and annual fluctuations;
- Variations in ET rates due to differences in vegetation (e.g., presence of multiple species, multiple canopy layers, annual and perennial vegetation, and variable vegetation density);
- Evaporation of precipitation (both rain and snow) intercepted by vegetation, litter, and soil;
- Evaporation from lakes, ponds, and creeks; and
- Variation in ET rates due to physiographic differences (i.e., soil water properties, topography).

The purposes of this paper are to: 1) define the physical processes of evaporation and transpiration, as they relate to water balances in forests, 2) review relevant literature that describes efforts to model ET processes, and 3) evaluate the current methods for evaluating ET in the WEPP (Water Erosion Prediction Project) model (Flanagan *et al.,* 1995), providing direction for possible improvements to the model. Although, some empirical methods are in current use, the focus of this paper will be on physical models since my objective is to determine which physical ET models are most appropriate for inclusion into the WEPP model.

Physical Processes

Evaporation is a two-stage process where energy changes the state of the water molecules from liquid to vapor, which are then transported away from the evaporative surface. Direct solar radiation and ambient air temperature provide the input energy. Vapor pressure gradient (between the atmosphere and the evaporating surface) and wind speed drive the removal of water vapor from the evaporating surface (Allen *et al.,* 1998). . Transpiration is the loss of water from plants in the form of vapor (Kramer and Kozlowski, 1979). The transpiration loss differs importantly from direct evaporation, in that water obtained from the soil must pass through the plant before it reaches the atmosphere, and this passage is controlled by the plant's physiological processes (Patric, 1967). Evaporation and transpiration occur simultaneously and there is no easy way of distinguishing between the two processes; thus the combined term, evapotranspiration (Dunne and Leopold, 1978). Hereafter, unless the individual processes of evaporation and transpiration are specifically addressed, the term evapotranspiration (ET) will be used to represent all of the evaporation, transpiration, and combined evapotranspiration processes.

Weather, vegetation characteristics, management practices, and physiographic aspects all affect evapotranspiration. The principal weather parameters affecting evapotranspiration are radiation, air temperature, humidity, and wind speed (Allen *et al.,* 1998). The vegetation type, density, volume, and development stage can affect ET rates. Differences in resistance to transpiration, crop height, crop roughness, reflection (albedo), ground cover, and crop rooting characteristics result in different ET levels in different plants under identical environmental conditions (Allen *et al.,* 1998). The physiographic factors that principally affect ET are soil water content, topography, water table depth, and soil properties. The effect of soil water content on ET is conditioned primarily by the magnitude of the water deficit and the type of soil (Allen *et al.,* 1998). Too much water can result in waterlogging, damaging the roots, and limiting root water uptake by inhibiting respiration (Allen *et al.,* 1998). Factors such as soil salinity and the presence of hard or impenetrable soil horizons reduce the evapotranspiration (Allen *et al.,* 1998).

Diurnal and Seasonal Fluctuations of ET

Crops and trees predominately lose their water through stomata during transpiration (Allen *et al.,* 1998). Nearly all water taken up is lost by transpiration and only a tiny fraction is used within the plant (Allen *et al.,* 1998). Stomata are small openings on the plant leaf through which gases and water vapor pass. A prime function of stomata is to prevent leaf desiccation when soil water extraction by the plant has fallen behind the rate of water loss (Ziemer, 1979), or when too much direct solar radiation early in the afternoon causes stomates to close (Kramer and Kozlowski, 1979).

Plants experience variations in temperature associated with diurnal variations in net radiation (Kimmins, 1987). Metabolic and respiratory rates increase with temperature (Kimmins, 1987). Water is lost either as a cooling adaptation or simply because of increased evaporation at higher temperatures (Kimmins, 1987; Kramer and Kozlowski, 1979). Measured values of ET are approximately zero at night (Stannard, 1993) and dramatically increase as temperature and solar energy increases (Kimmins, 1987).

During winter months, incident radiation is lower due to a decreased sun angle, temperatures are lower (due to decreased capture of solar radiation), wind speeds are higher (especially during storms), and relative humidity is generally higher (especially during storms). These factors combine to produce, typically (in the continental U.S.), significantly lower ET rates. Deciduous trees typically do not transpire during the winter without leaves (Kimmins, 1987;

Kramer and Kozlowski, 1979). However, conifers can continue to transpire during the winter, although at a much slower rate, because they maintain their leaves year-round (Kimmins, 1987).

Vegetation Factors Affecting ET

Irregular Vegetation Spacing

Natural forests have canopy openings, changes in vegetation density, variations in vegetation height, and patches of bare mineral soil (Kimmins, 1987). Even uniform forests (e.g., tree farms) and row crops can have irregular vegetation density and spacing. Leaf area index (LAI) is the most important variable for measuring vegetation structure (i.e., density and distribution) over large areas, and is the principal independent variable for calculating canopy interception, transpiration, respiration, and photosynthesis (Running and Coughlan, 1988). LAI is a function of species type, stand density, stand age, canopy height, and succesional stage (Kitteredge, 1948; Patric, 1967; Ziemer, 1979). Irregular vegetation density influences ET, whereas irregular ground cover influences evaporation from bare soil or litter.

Vegetative density affects evapotranspiration rates from forest stands by modifying the area of transpiring surface, net radiation capture, precipitation interception, wind patterns and turbulence, and root distribution (Douglass, 1966). Vegetative surface area, net radiation capture, and precipitation interception are directly proportional to ET rates. However, increased vegetation density reduces wind speed within the canopy, thereby reducing ET rates (by reducing vapor flux). Greater vegetation and root density increases the rates of withdrawal of water from the soil. Reducing vegetative density reduces evapotranspiration, and the greater the density reduction, the greater the evapotranspiration reduction (Douglass, 1966). In some species, an advanced stage of maturity or senescence will also limit evaporation (Ritchie, 1972). Older trees have reduced water demands because they are only using enough water to maintain current vegetation and are not generating new growth (which uses water at a greater rate) (Kramer and Kozlowski, 1979).

Multiple Species and Canopy Layers

Natural forests typically have multiple species (e.g., dominant conifers, sub-dominant hardwoods, shrubs, and forbs), which form multiple canopy layers (i.e., dominant, co-dominant, subdominant, suppressed, and understory) (Kimmins, 1987). The two most important factors affected by multiple vegetative species (in multiple canopy layers) are water use and light capture (Douglass, 1966; Kimmins, 1987; Kramer and Kozlowski, 1979). Conifers and broad-leaved evergreens usually use less water than deciduous trees (Kitteredge, 1948; Ziemer, 1979), and grasses use much greater water than both conifers and hardwoods (Douglass, 1966; Ziemer, 1979). The greatest differences in water usage by species are the result of variations in rooting depth (Douglass, 1966), which are the least for grasses and the greatest for some conifers (e.g., Douglas-fir). Rooting depth has its greatest effect on evapotranspiration in regions characterized by distinct wet and dry seasons (Douglass, 1966).

The light conditions experienced by a forest plant will vary according to average light conditions in the stand and the position of the plant in relation to the rest of the canopy (Kimmins, 1987). A crown dominant will receive full sunlight, while co-dominant, subdominant, suppressed, and understory plants will generally receive progressively less light. Radiation reaching the forest floor varies with crown density, spacing (number of stems per hectare), height above the ground, and time of day (Kimmins, 1987). Light on the forest floor may be only 1 to 5% as intense as that of full sunlight (Kramer and Kozlowski, 1979).

Evaporation of Precipitation

Evaporation From Soil and Litter

Evaporation from the soil surface is known to occur in three stages (Burman and Pochop, 1994; Ritchie, 1972). The first is characterized by a wet soil surface and is usually referred to as the constant stage because soil-surface evaporation occurs at the rate of the climatic potential. The second stage is the falling stage because soil-surface evaporation is related to the square root of time since the stage began and is largely controlled by soil characteristics. The third stage is characterized by very slow rates from a very dry soil profile; where the soil surface is said to control the rate of evaporation (Burman and Pochop, 1994; Ritchie, 1972).

Frequent rains, irrigation, and water transported upwards in a soil from a shallow water table wet the soil surface (Allen *et al.,* 1998). Where the soil is able to supply enough water to satisfy the evaporation demand, the evaporation from the soil is determined only by the meteorological conditions. However, where the interval between rains and irrigation becomes large and the ability of the soil to conduct moisture to near the surface is small, the water content in the topsoil drops and the soil surface dries out (Allen *et al.,* 1998). In the absence of any supply of water to the soil surface, evaporation decreases rapidly and may cease almost completely within a few days (Allen *et al.,* 1998).

Apart from the water availability in the topsoil, evaporation rates from a soil with plant cover are determined by the fraction of the solar radiation reaching the soil surface (Allen *et al.,* 1998). Under a continuous cover of vegetation (as is common in forests), insufficient energy reaches the soil to produce much evaporation, even in the summer (Kimmins, 1987). Plant residue (mulch or litter) can reduce the amount of soil evaporation because the stubble and residue shades the soil surface, reduces the rate of soil temperature increase, and reduces overall soil evaporation (Burman and Pochop, 1994; Kimmins, 1987). Leaf litter can intercept and hold significant volumes of rainfall (Patric, 1967), a portion of which is available for evaporation following the rainstorm.

Evaporation From Vegetative Surfaces

Interception is the process in which rainfall is caught by the vegetative canopy and redistributed as throughfall, stemflow, absorption and evaporation from the vegetation (Zinke, 1966). For an individual storm, there is an initial period when the vegetation canopy is wetted and interception storage capacity is satisfied (Zinke, 1966). This is followed by loss from this storage, which is dependent upon the evaporation opportunity during the remainder of the storm (Rutter and Morton, 1977). The percentage of total precipitation lost decreases as the amount of precipitation per shower increases. Interception is, at most, 100 percent of total rainfall for storms that do not exceed the interception storage capacity of the vegetative cover (Ziemer, 1979); and can be as little as 5 percent for intense, high volume rainstorms (Ziemer, 1979).

Rainfall interception is usually disregarded in croplands, but must be accounted for in forests where a quarter or more of summer rainfall may be lost by interception (Patric, 1967). The amount of precipitation intercepted, and subsequently lost to the atmosphere, is a function of LAI, total precipitation, precipitation intensity, precipitation duration, evaporation rate during precipitation and the length of time between rainstorms (Kitteredge, 1948; Patric, 1967; Ziemer, 1979). Interception losses also vary with species and forest types, within stands (Kitteredge, 1948; Patric, 1967; Ziemer, 1979), and with geographic location (e.g., orographic and topographic influences), time of year (i.e., season), and climate.

Interception losses tend to be much less in the winter than in the summer, due to much lower evaporation rates (Ziemer, 1979). Since vapor pressure and surface-air temperature gradients are very low during typical winter rainstorms, the evaporation rates are also low (especially when compared to summer rates) (Kitteredge, 1948; Patric, 1967; Ziemer, 1979). Monthly or seasonal interception losses vary according to the distribution and timing of the precipitation (Patric, 1967). Longer dry periods between rainstorms result in much greater evaporative losses than shorter periods. So, for small storms that are widely spaced, interception losses are a significant percentage of total rainfall. However, for large storms (> 25 mm) that are in close succession, interception losses are inconsequential (Kitteredge, 1948; Patric, 1967; Ziemer, 1979). Interception losses are usually low in regions where they are compensated by fog or cloud drip (Kitteredge, 1948).

Evaporation from Snow

Approximately 2830 Joules of energy are required to convert water from the solid state to the vapor state. The controls of this process are the same as those for evaporation from a free water surface, but there are some limitations that generally keep the evaporation from snow at a low rate (e.g., cold air temperature, high cloud cover, and low incoming solar radiation). Only under conditions of bright sunshine, warm air temperatures and strong winds does evaporation from snow become significant (Dunne and Leopold, 1978). Of all ET methods evaluated, none were designed for use when snow is present (Stannard, 1993). . It is not surprising that snow is typically ignored, since most of the methods were developed to estimate ET for purposes of accounting for consumptive water use during the crop growing season. All methods need to be adjusted to be used in forest conditions, with perennial vegetation, where interception and snow hydrology may be important (Federer *et al.,* 1996). Additionally, a thorough treatment of sublimation and evaporation from a snowpack that involves a snowmelt model and a method to account for conditions of partial snow cover is needed (Stannard, 1993).

Evaporation from Water Bodies

Globally, evaporation from water bodies is extremely important in the hydrologic cycle. In fact, the development of combination evaporation equations (e.g., Penman 1948, Priestly and Taylor 1972) was derived for calculating evaporation rates from water bodies (Dunne and Leopold, 1978). In forests, water bodies include creeks, rivers, ponds, wetlands, marshes, and small lakes. The extent to which each of these water bodies affects the water balance would be determined by the areal extent and the location (in relation to the forest) of the water body.

Physiographic Factors Affecting ET Rates

Physiographic factors can greatly affect ET rates in forests. Differences in aspect can dramatically influence the interception of incoming solar radiation. Elevation strongly influences vapor pressure, air temperature, water content, and wind speed. Topography strongly influences soil-water redistribution and climate (e.g., rain shadows, orographic lifting). Soil properties (e.g., particle size distribution, water content) influence the amount of plant-available water (Kimmins, 1987) and soil-water redistribution. All of these factors can influence the type, volume, size, and amount of vegetation that occurs on a site (Kimmins, 1987; Kramer and Kozlowski, 1979).

Modeling Evapotranspiration Processes

Penman-type Methods

Penman (1948)

The Penman (1948) combination equation was developed to specifically evaluate evaporation from saturated surfaces, such as open water, bare soil, and recently watered grass. It is considered a combination equation because it combines both energy (sensible heat flux) and atmospheric vapor (latent heat flux) transport to model potential evaporation (Katul and Parlange, 1992), and an empirical wind function (Burman and Pochop, 1994). This method is semi-empirical, but is based on physical properties of water and on microclimatic variables: vapor pressure deficit in air, latent heat of vaporization, net radiation above a surface, wind speed, rate of change of vapor pressure with temperature, water density, and selected constants (e.g., psychrometer, unit conversion) (Federer *et al.,* 1996).

This method is independent of vegetation type, quantity, and structure, which are important factors to consider when estimating rates of ET (Federer *et al.,* 1996), requiring empirical adjustments for varying locations and conditions. It also does not explicitly account for differences in evaporation between wet and dry vegetation, between bare and covered soil (Federer *et al.,* 1996), for advected energy under semi-arid conditions (Patric, 1967). The main theoretical limitation is the use of steady state equations with a process that involves diurnal or shorter period variations (Burman and Pochop, 1994).

McNaughton and Black (1973):

This equation is a simplified form of the Penman equation that excludes wind speed and solar radiation, but includes a canopy resistance coefficient (similar to the ones used in Penman-Monteith-type equations) (Federer *et al.,* 1996). In this formulation of the Penman equation, potential ET varies only with vapor pressure deficit (proportionally) and canopy resistance (inversely proportionally). It is scaled by a combination of the specific heat capacity, density, and psychrometric constant; all of which are functions of temperature, water content, and pressure. This P-M equation reduces to this equation if incoming radiation is ignored, and if the aerodynamic resistance is assumed negligible. The McNaughton-Black method is theoretically only valid for forests (specifically, Douglas-fir forests where it was developed), where aerodynamic resistance is negligible with respect to canopy resistance (Federer *et al.,* 1996). It is, however, easier to compute when microclimatic data are missing or unavailable.

Priestly-Taylor (1972)

The Priestly-Taylor equation was introduced to estimate potential evapotranspiration in conditions of minimal advection (Priestley and Taylor, 1972), as a modification to the Penman equation. It is valid for a horizontally uniform saturated surface (Priestly and Taylor 1972). This method is based on the same physical properties as the Penman equation. It is, however, easier to compute when microclimatic data are missing or unavailable (Federer *et al.,* 1996). The P-T equation is a modified form of the Penman equation, such that the aerodynamic resistance (a term that relates ET to wind speed and vapor pressure deficit) is eliminated, and the heat flux term (first term) is adjusted by a coefficient, $\alpha = 1.26$ (Priestley and Taylor, 1972).

The simplicity and accuracy of the PT equation in well-watered conditions led to the use of modified forms of the equation to estimate latent heat flux for partially dry surfaces (Stannard, 1993). It was reasoned that, as a canopy became water-stressed, α would decrease below

1.26 (Stannard, 1993), and only an adjustment to the alpha coefficient was necessary to estimate ET under a wider range of moisture conditions.

Ritchie (1972)

The challenge in applying the Penman-Monteith equation is in prediction of parameters during periods of partial ground cover and partial soil surface wetness (Pereira *et al.,* 1999). Ritchie recognized that when annual row crops are in early growth stages with little vegetative cover, the evaporation rate from the entire field surface is dominated by the soil evaporation rate (Ritchie, 1972). Ritchie (1972) modified Penman's equation to account for evaporation from vegetation and soil surfaces separately. The evapotranspiration rate from vegetation was scaled by the leaf area index, so that ET changes during the growing season could be accounted for. Wind speed, vapor pressure deficit, and temperature below the canopy (at the soil surface) are reduced in approximate proportion to the canopy density (Ritchie, 1972).

In addition to accounting for variable vegetation cover, Ritchie (1972) also modified Penman's equation to account for variable evaporation rates from bare soil under varying soil moisture conditions. Ritchie developed a two-stage evaporation model to simulate two different soil moisture regimes: constant evaporation rate and falling evaporation rate. In the constant rate stage, the soil is sufficiently wet for the water to be transported to the surface at a rate at least equal to the potential evaporation rate. In the falling rate stage, the surface soil water content has decreased below a threshold value, so that the evaporation rate depends on the flux of water through the upper layer of the soil to the evaporating site near the surface (Ritchie, 1972). The constant rate stage is computed using Ritchie's equation and the falling rate stage is computed using an empirical function of cumulative evaporation (based on local soil hydraulic properties, approximately estimated by the hydraulic conductivity of the soil at –0.1 bar soil matric potential) that is scaled by the square root of the time since the soil surface was wetted (Ritchie, 1972).

Rutter (1975, 1977)

Rutter's modification to Penman's equation was designed to specifically address evaporation of intercepted rainfall from vegetative surfaces (Rutter *et al.,* 1975; Rutter and Morton, 1977). The formulation of the evaporation component of Rutter's equation replaces Penman's empirical wind speed function with an aerodynamic resistance function. Rutter adopted the aerodynamic resistance function developed by Thom (Thom, 1972), which was later adopted by Monteith for his formulation of the Penman-Monteith equation. The resistance function is based on Prandtl's mixing length equation (Burman and Pochop, 1994), which assumes that the wind speed at a fixed distance above the plant surface has a logarithmic velocity profile, assumed is zero at the vegetative surface. Using the evaporation equation, Rutter quantified forest interception losses (following rainstorms) as a function of canopy storage, precipitation volume and rate, and canopy drainage (Rutter *et al.,* 1975).

Penman-Monteith Methods

Penman-Monteith (1965)

The limitations of Penman's equation have been substantively overcome by Monteith (1965) and many subsequent researchers, such that many Penman-type equations are widely used in hydrology today (Katul and Parlange, 1992). This equation is a modified form of the Penman equation that eliminates all of the empirical constants used by Penman; excludes Penman's empirical wind speed function; includes a canopy resistance factor, and includes an additional factor, aerodynamic resistance to account for turbulent mixing, and mass and energy fluxes (Burman and Pochop, 1994; Federer *et al.,* 1996). It is considered the most physically based of all potential ET equations (Federer *et al.,* 1996).

The P-M equation has two general terms: a canopy resistance term and an aerodynamic resistance term. The canopy resistance is a function of stomatal resistance, leaf area index, and the diffusive resistance to water vapor through the canopy air volume (Tan and Black, 1976). In a forest, canopy resistance is much greater than either of the aerodynamic resistances (i.e., resistance to heat and vapor flux), so the aerodynamic resistance term is typically much smaller than the canopy resistance term (Tan and Black, 1976). Modeling of the canopy resistance requires knowledge of how plant and environmental parameters (e.g., light, internal water deficits of the tree, vapor pressure deficit, and temperature) affect the stomatal resistance of each tree species (Tan and Black, 1976). As water availability (from the soil) to a canopy decreases, the value of canopy resistance increases, and latent heat flux decreases. As a canopy approaches well-watered conditions (not including precipitation interception), canopy resistance approaches zero and the Penman-Monteith (P-M) equation approaches the Penman equation (Stannard, 1993).

The P-M model has four main requirements for validity. 1) No local advection occurs over the surface, thus the flux between the two levels is only vertical. Therefore, the conditions at the reference level (solar radiation, wind speed, and vapor pressure deficit) can be considered to be the same as that over a large surrounding area (Pereira *et al.,* 1999). 2) The turbulent exchange coefficients (aerodynamic resistances) are the same for sensible and latent heat and include the corrections for instability or stability (Pereira *et al.,* 1999). In situations where this is not necessarily true, Thom (1972) provides a form of the P-M equation where water-vapor exchange and heat exchange are not equal (Thom, 1972). 3) The evaporative surface level is at the height of the plants or, for a very low crop or a bare soil, the roughness height for momentum (Pereira *et al.,* 1999). 4) The fraction of energy absorbed by the canopy as heat that is transformed into dry matter can be neglected (Pereira *et al.,* 1999).

In addition to the above requirements, the P-M model has several implicit and explicit assumptions. 1) The soil water supply is not limiting. In order to determine actual ET, the level of soil water must be considered (Burman and Pochop, 1994). 2) The weather data collected over the plants are within the fully adjusted boundary layer (having a logarithmic wind profile) and that the surface over which the weather data are collected is the same as the plants for which ET is being predicted by the equation (Pereira *et al.,* 1999). 3) The entire canopy behaves, in terms of resistance, similar to a single "big leaf", such that the leaf surface can be evaluated the same as a free water surface. The big leaf assumption requires that the sources of sensible and latent heat be at the same height and temperature. This requirement is met by a full canopy, or an entirely wet, bare soil surface, but not by a sparse or irregular canopy (Stannard, 1993). 4) The canopy is isothermal (Ziemer, 1979), which is unrealistic in most forest settings. However, Tanner and Fuchs (1968) give an alternative form of the P-M equation for the case of a non-isothermal canopy (Tan and Black, 1976). 5) The canopy resistance is the resistance of all stomata of the leaves acting in parallel (Tan and Black, 1976).

Due to the assumptions stated above, there are some limitations in applying the P-M equation to un-instrumented catchments. It only applies to relatively flat land only where the topographic effect is not taken into consideration (Murakami *et al.,* 2000). Meteorological data, obtained by point measurements, are considered representative of the entire area (Murakami *et al.,* 2000). It does not account for ET rate changes due to partial vegetation cover (Federer *et al.,* 1996) or evaporation from bare soil (explicitly) and leaf litter (Federer *et al.,* 1996) when there is canopy present. However, total evapotranspiration can be considered as mainly transpiration (neglecting soil-water evaporation) with small error for forests with no intercepted water present, since evaporation from the soil is small (Tan and Black, 1976). It does not account for varying soil water contents (Federer *et al.,* 1996) or for evaporation of intercepted rainfall, snowfall, or irrigation water (Ziemer, 1979). Many of these limitations have been subsequently accounted for and adjusted by other researchers (e.g., (Choudhury and Monteith, 1988; Priestley and Taylor, 1972; Ritchie, 1972; Rutter *et al.,* 1975; Shuttleworth and Wallace, 1985)).

Shuttleworth-Wallace (Shuttleworth and Wallace, 1985)

As stated above, Penman and P-M assume that the ground cover, whether it is bare soil, water, grass, or trees, is uniform and complete. Several researchers found that the PM model is less accurate than the Shuttleworth-Wallace (S-W) and P-T models (e.g., (Federer *et al.,* 1996; Pereira *et al.,* 1999; Stannard, 1993) for variable density ground cover. This is not surprising for two reasons. First, for areas with sparse vegetation coverage, the P-M model big leaf assumption does not hold during dry, sunny periods, when a large fraction of sensible heat comes from the hot soil (Shuttleworth and Gurney, 1990; Stannard, 1993). Second, immediately after a rainfall, the P-M model cannot simulate the large values of bare soil evaporation and interception evaporation, because it is exclusively an evapotranspiration model (Stannard, 1993).

Shuttleworth and Wallace (1985) modified the P-M equation by assuming the two asymptotic limits of ground cover (bare substrate and closed canopy) can both be represented by correct representations of aerodynamic resistance in the P-M equation (Shuttleworth and Wallace, 1985). This improvement to the PM equation explicitly accounts for evaporation from both bare soil and sparsely distributed vegetation, separately but simultaneously, allowing for a more complete representation of areas with variable canopy cover (Stannard, 1993). . It is based on the same physical parameters as the PM equation, but includes variables allowing for dramatically different vegetation/canopy heights (Federer *et al.,* 1996). To account for variable surface heights, S-W assumed that aerodynamic mixing within the crop is sufficiently good to allow the hypothetical existence of a 'mean canopy airstream' which can be described by meteorological parameters such as temperature, humidity, and wind speed (Shuttleworth and Wallace, 1985). If reasonable formulations for the resistance terms and an approximate value for the attenuation coefficient (of solar radiation distributed through multiple canopy layers) can be obtained, the data required to estimate potential ET using the S-W or P-M models are substantially the same (Stannard, 1993). The data requirements for both models are, however, large, and both models are computationally demanding (Stannard, 1993).

Four-layer heat budget (by Choudhury and Monteith (1988))

This P-M type equation uses the S-W method for evaporation from a variably covered vegetation canopy and soil surface (Choudhury and Monteith, 1988). This model extends the work of Shuttleworth and Wallace (1985) to include a fourth layer in the system, a soil horizon below the soil surface. The vegetation has two layers, the first extending from a reference height in the atmosphere to the virtual sink for momentum and the second from the virtual sink to the soil surface. Soil between the surface and the damping depth is divided into an upper, completely dry layer and a lower wet layer (Choudhury and Monteith, 1988). Throughout the system, differences of potential per unit flux are specified by soil resistance, assumed proportional to the accumulated loss of water by evaporation from the soil surface, limited by absorption of radiation, the increase of transpiration from foliage as it expands decreases evaporation from the soil. Conversely, as soil dries, transpiration rate per unit of foliage area increases. This interaction of vapor fluxes is governed by the behavior of the saturation vapor pressure deficit with the vegetation (Choudhury and Monteith, 1988).

Recommendations for WEPP improvements

WEPP primarily uses a modification of Ritchie's method (Ritchie, 1972) to compute ET (Savabi and Williams, 1995). In cases where daily radiation, temperature, wind, and dew point temperature or relative humidity data are available, the WEPP model uses the Penman (1963) equation with the original wind function method to compute evaporation. Although Ritchie's method originally used Penman's evaporation equation, in cases where only solar radiation and temperature data are available, the WEPP model substitutes the Priestly-Taylor (1972) equation to compute potential ET. Although WEPP gives the user the option to use either Penman's, Priestly-Taylor's, Hargraves', or Penman-Monteith's equations for calculating ET, the coding for Hargraves' and Penman-Monteith's equations are incomplete, and are therefore turned off. The P-M equation (and subsequent variations) is arguably the state-of-the-science method for computing ET (Allen *et al.,* 1998; Burman and Pochop, 1994). As such, <u>completing the coding of the P-M would be a highly desirable addition to the model.</u> For the factors discussed below, it is assumed (for the purposes of this paper) that the coding of the model is accurately described in the technical documentation. However, when the suggested improvements are being incorporated into the model, it is important to confirm that the coding and parameterization are consistent with the documentation.

How should WEPP Calculate ET for Forests

Variation in potential ET rates due to diurnal, seasonal, and annual fluctuations

The basic ET equations used in WEPP (i.e., Penman, Ritchie, Priestley-Taylor) allow for diurnal, seasonal and annual variation in ET rates. The differences would be due to the parameterization of LAI for varying crop/plant species during the growing season and in the winter. WEPP evaluates ET for the growing season, and has a winter hydrology section that accounts for water loss (evaporation) from snow. The coding may need to be checked to determine whether ET is calculated for conifers or other evergreen species that may transpire (even a small amount) during some dry, warm periods of the winter. <u>No substantive changes are recommended</u>.

Variations in ET rates due to differences in vegetation

WEPP uses Ritchie's method (Ritchie, 1972) to compute ET, which explicitly accounts for vegetation of variable density. WEPP, like Ritchie, uses a variation of Beer's law to account for varying light intensity for multiple species in multiple canopy layers. WEPP uses a model similar to the EPIC crop growth model to evaluate changes in vegetation during the year/growing season (Arnold *et al.,* 1995). LAI is used to account for variable plant density and ground cover density. The model computes temporal changes in plant and residue variables such as canopy cover, canopy height, root development, and biomass production. The model can simulate growth for annual and perennial plants, including agricultural crops and rangeland grasses. For annual or deciduous plants, changes in canopy due to senescence are accounted. For rangeland (or woodland) areas, canopy height is calculated as the weighted average of the tree, shrub, and herbaceous components (Arnold *et al.,* 1995). Crop growth can be reduced by either (or both) temperature or water stress, and herbicide use (Arnold *et al.,* 1995). <u>No substantive changes are recommended</u>.

Evaporation of precipitation (both rain and snow) intercepted by vegetation

In WEPP, precipitation is partitioned between rainfall and snowfall using air temperature. Precipitation interception by vegetation is calculated using the method described by Savabi and Stott (1994), where the interception capacity is a function of the above ground biomass (a second-order, empirical equation) (Savabi and Williams, 1995). WEPP accounts for interception of rain and snow by a vegetative canopy, but it apparently does not do anything with it. The manuals do not explicitly state how intercepted precipitation (that remains on the canopy) is treated. It would be useful to compare the interception equation used in WEPP with Rutter's equation (Rutter *et al.,* 1975), to determine which one is more accurate or physically based. For rough canopies, such as forests, it is recommended to compute precipitation interception losses separately from transpiration and ground evaporation (Federer *et al.,* 1996). Rutter, et al. (1975, 1977) developed a method for computing evaporation of intercepted rainfall (not snow or ice) that accounts for differences in interception due to vegetative differences (e.g., density, LAI, leaf storage), drainage from the canopy and stems, and varying climatic conditions. Since WEPP already uses both of these equations, adding evaporation of intercepted precipitation should not be difficult. <u>This item needs attention and possibly substantive changes</u>.

The code needs to be checked to determine if the U.S. Army Corps of Engineers [*Snow Hydrolgoy*] method for snow hydrology accounts for evaporation, sublimation, and melt from vegetated canopies. This method accounts for climatic conditions (temperature, wind speed, dew point, heat input to snow pack from incoming rainfall), vegetation conditions (canopy height, open soil, stubble height, albedo, and snow depth), physiographic conditions (incoming solar radiation, canopy cover intercepting incoming radiation, longwave radiation from cloud cover); snow pack conditions (temperature, density, depth, albedo, and snow drift). These equations deal with the four major energy components of the snowmelt process: temperature, radiation, vapor transfer, and precipitation (Savabi *et al.,* 1995). The model also accounts for changes in heat, vapor, and water flux when the soil is frozen. The drift and throughfall of intercepted snow may be included already. If not, this needs to be included. <u>This item needs attention and possibly substantive changes</u>.

Soil and litter evaporation

By using Ritchie's method, WEPP accounts for evaporation from bare soil under both wet and drying conditions. Bare soil evaporation is reduced with increasing plant residue using an empirical equation developed by J.L. Steiner (Savabi and Williams, 1995). Although this method is adequate, a more physically based approach is used by Shuttleworth and Wallace (1985). However, since these methods produce comparable results (Stannard, 1993), it may not be worth the effort to make this change. In some forest conditions, especially after harvest, evaporation from litter can become considerable (or litter could be a significant factor in preventing evaporation from the soil). It may be necessary to account for this factor. Litter evaporation is not directly addressed. The four-layer heat budget model (Choudhury and Monteith, 1988) uses a technique that parameterizes heat and vapor flux through a layer of residue (i.e., litter); it may be possible to incorporate these parameters into WEPP if it is determined that litter evaporation can become significant. <u>This problem needs little attention and possibly minor changes</u>.

Water body evaporation (in the cases of lakes, ponds, and creeks)

If WEPP is to be used to evaluate water balances for watersheds, it is possible that some watersheds will have areas with free water surfaces, (e.g., lakes, ponds, marshes, wetlands).

To accurately evaluate the watershed-scale water balance, evaporation from these sources should be evaluated. WEPP already has the equations and the code to do this; both Penman and Priestly-Taylor equations are for evaporation from wet (free-water) surfaces. This problem needs little attention and possibly minor changes.

Variation in actual ET rates due to physiographic differences (i.e., soil water properties, topography)

WEPP accounts for varying soil properties (e.g., water content, hydraulic conductivity, structure, temperature, cover, albedo), topography (aspect), and snow cover. No substantive changes are recommended.

Conclusions

Many micrometerological models (e.g. PASSM, SHAW, Forest BGC), hydrologic models (e.g., DHSVM, SHE), water quality models (e.g., SWAT, ANSWERS), erosion (e.g., WEPP, SHE-SED), and crop models (e.g., ERIN, EPIC) regularly evaluate potential, reference, and/or actual evapotranspiration. However, most models that evaluate ET do not explicitly evaluate surface erosion due to changes in evapotranspiration rates following vegetation changes (e.g., wildfire, fuel management, harvesting). The exceptions are the water quality and erosion models. Of the models that compute both ET and surface erosion, only WEPP uses a physically based approach to compute surface erosion. All other models use variations of the highly empirical USLE (or its variants RUSLE and MUSLE) to estimate erosion rates. Most current models use the Penman-Monteith combination equations (or one of the variants—Ritchie, Priestly-Taylor, Shuttleworth-Wallace) to compute potential and actual evapotranspiration. Therefore, WEPP can prove be a superior watershed erosion, hydrology and water quality model. As such, it is important to ensure that WEPP maintains its physical basis for as many of its sub process as possible.

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