PART 1

MODEL DESCRIPTION

CLIMATE

The climatic inputs to the model are reviewed first because it is these inputs that provide the moisture and energy that drive all other processes simulated in the watershed. The climatic processes modeled in SWAT consist of precipitation, air temperature, soil temperature and solar radiation. Depending on the method used to calculate potential evapotranspiration, wind speed and relative humidity may also be modeled.

CHAPTER 2

EQUATIONS: ENERGY

Once water is introduced to the system as precipitation, the available energy, specifically solar radiation, exerts a major control on the movement of water in the land phase of the hydrologic cycle. Processes that are greatly affected by temperature and solar radiation include snow fall, snow melt and evaporation. Since evaporation is the primary water removal mechanism in the watershed, the energy inputs become very important in reproducing or simulating an accurate water balance.

2.1 SUN-EARTH RELATIONSHIPS

A number of basic concepts related to the earth's orbit around the sun are required by the model to make solar radiation calculations. This section summarizes these concepts. Iqbal (1983) provides a detailed discussion of these and other topics related to solar radiation for users who require more information.

2.1.1 DISTANCE BETWEEN EARTH AND SUN

The mean distance between the earth and the sun is 1.496 x 10^8 km and is called one astronomical unit (AU). The earth revolves around the sun in an elliptical orbit and the distance from the earth to the sun on a given day will vary from a maximum of 1.017 AU to a minimum of 0.983 AU.

An accurate value of the earth-sun distance is important because the solar radiation reaching the earth is inversely proportional to the square of its distance from the sun. The distance is traditionally expressed in mathematical form as a Fourier series type of expansion with a number of coefficients. For most engineering applications a simple expression used by Duffie and Beckman (1980) is adequate for calculating the reciprocal of the square of the radius vector of the earth, also called the eccentricity correction factor, E_0, of the earth's orbit:

$$E_0 = (r_0/r)^2 = 1 + 0.033\cos[(2\pi d_n/365)] \qquad 2.1.1$$

where r_0 is the mean earth-sun distance (1 AU), r is the earth-sun distance for any given day of the year (AU), and d_n is the day number of the year, ranging from 1 on January 1 to 365 on December 31. February is always assumed to have 28 days, making the accuracy of the equation vary due to the leap year cycle.

2.1.2 SOLAR DECLINATION

The solar declination is the earth's latitude at which incoming solar rays are normal to the earth's surface. The solar declination is zero at the spring and fall equinoxes, approximately +23½° at the summer solstice and approximately -23½° at the winter solstice.

A simple formula to calculate solar declination from Perrin de Brichambaut (1975) is:

$$\delta = \sin^{-1}\left\{0.4\sin\left[\frac{2\pi}{365}\left(d_n - 82\right)\right]\right\} \qquad 2.1.2$$

where δ is the solar declination reported in radians, and d_n is the day number of the year.

2.1.3 SOLAR NOON, SUNRISE, SUNSET AND DAYLENGTH

The angle between the line from an observer on the earth to the sun and a vertical line extending upward from the observer is called the zenith angle, θ_z (Figure 2-1). Solar noon occurs when this angle is at its minimum value for the day.

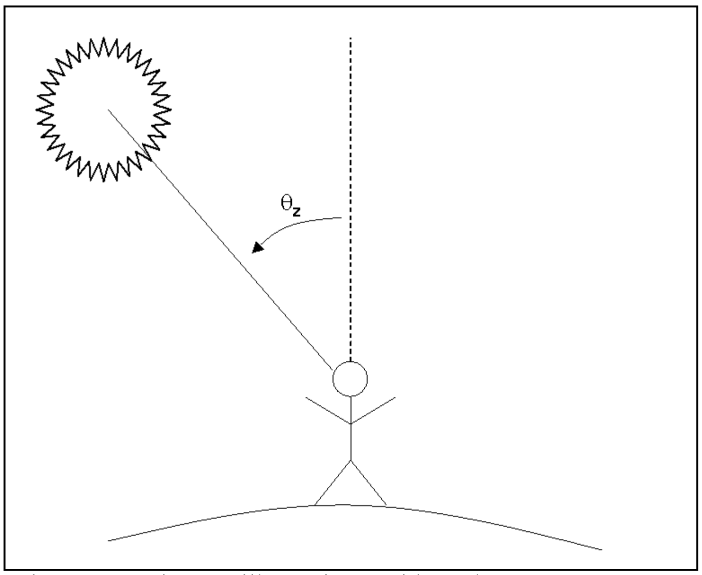

Figure 2-1: Diagram illustrating zenith angle

For a given geographical position, the relationship between the sun and a horizontal surface on the earth's surface is:

$$\cos\theta_z = \sin\delta\sin\phi + \cos\delta\cos\phi\cos\omega t \qquad 2.1.3$$

where δ is the solar declination in radians, ϕ is the geographic latitude in radians, ω is the angular velocity of the earth's rotation (0.2618 rad h^{-1} or 15° h^{-1}), and t is the solar hour. t equals zero at solar noon, is a positive value in the morning and is a negative value in the evening. The combined term ωt is referred to as the hour angle.

Sunrise, T_{SR}, and sunset, T_{SS}, occur at equal times before and after solar noon. These times can be determined by rearranging the above equation as:

$$T_{SR} = +\frac{\cos^{-1}[-\tan\delta\tan\phi]}{\omega} \qquad 2.1.4$$

and

$$T_{SS} = -\frac{\cos^{-1}[-\tan\delta\tan\phi]}{\omega} \qquad 2.1.5$$

Total daylength, T_{DL}, is calculated:

$$T_{DL} = \frac{2\cos^{-1}[-\tan\delta\tan\phi]}{\omega} \qquad 2.1.6$$

At latitudes above 66.5° or below -66.5°, the absolute value of [-tanδ tanϕ] can exceed 1 and the above equation cannot be used. When this happens, there is either no sunrise (winter) or no sunset (summer) and T_{DL} must be assigned a value of 0 or 24 hours, respectively.

To determine the minimum daylength that will occur during the year, equation 2.1.6 is solved with the solar declination set to -23.5° (-0.4102 radians) for the northern hemisphere or 23.5° (0.4102 radians) for the southern hemisphere.

The only SWAT input variable used in the calculations reviewed in Section 2.1 is given in Table 2-1.

Table 2-1: SWAT input variables that used in earth-sun relationship calculations.

Variable name	Definition	File Name
LATITUDE	Latitude of the subbasin (degrees).	.sub

2.2 SOLAR RADIATION

2.2.1 EXTRATERRESTRIAL RADIATION

The radiant energy from the sun is practically the only source of energy that impacts climatic processes on earth. The solar constant, I_{SC}, is the rate of total solar energy at all wavelengths incident on a unit area exposed normally to rays of the sun at a distance of 1 AU from the sun. Quantifying this value has been the object of numerous studies through the years. The value officially adopted by the Commission for Instruments and Methods of Observation in October 1981 is

I_{SC} = 1367 W m^{-2} = 4.921 MJ m^{-2} h^{-1}

On any given day, the extraterrestrial irradiance (rate of energy) on a surface normal to the rays of the sun, I_{0n}, is:

$$I_{0n} = I_{SC} E_0 \qquad 2.2.1$$

where E_0 is the eccentricity correction factor of the earth's orbit, and I_{0n} has the same units as the solar constant, I_{SC}.

To calculate the irradiance on a horizontal surface, I_0,

$$I_0 = I_{0n} \cos\theta_z = I_{SC} E_0 \cos\theta_z \qquad 2.2.2$$

where cos θ_z is defined in equation 2.1.3.

The amount of energy falling on a horizontal surface during a day is given by

$$H_0 = \int_{sr}^{ss} I_0 dt = 2\int_0^{ss} I_0 dt \qquad 2.2.3$$

where H_0 is the extraterrestrial daily irradiation (MJ m^{-2} d^{-1}), *sr* is sunrise, and *ss* is sunset. Assuming that E_0 remains constant during the one day time step and converting the time *dt* to the hour angle, the equation can be written

$$H_0 = \frac{24}{\pi} I_{SC} E_0 \int_0^{\omega T_{SR}} (\sin\delta \sin\phi + \cos\delta \cos\phi \cos\omega t) d\omega t \qquad 2.2.4$$

or

$$H_0 = \frac{24}{\pi} I_{SC} E_0 [\omega T_{SR} (\sin\delta \sin\phi) + (\cos\delta \cos\phi \sin(\omega T_{SR}))] \qquad 2.2.5$$

where I_{SC} is the solar constant (4.921 MJ m^{-2} h^{-1}), E_0 is the eccentricity correction factor of the earth's orbit, ω is the angular velocity of the earth's rotation (0.2618 rad h^{-1}), the hour of sunrise, T_{SR}, is defined by equation 2.1.4, δ is the solar declination in radians, and ϕ is the geographic latitude in radians. Multiplying all the constants together gives

$$H_0 = 37.59 E_0 [\omega T_{SR} \sin\delta \sin\phi + \cos\delta \cos\phi \sin(\omega T_{SR})] \qquad 2.2.6$$

2.2.2 SOLAR RADIATION UNDER CLOUDLESS SKIES

When solar radiation enters the earth's atmosphere, a portion of the energy is removed by scattering and adsorption. The amount of energy lost is a function of the transmittance of the atmosphere, the composition and concentration of the constituents of air at the location, the path length the radiation travels through the air column, and the radiation wavelength.

Due to the complexity of the process and the detail of the information required to accurately predict the amount of radiant energy lost while passing through the atmosphere, SWAT makes a broad assumption that roughly 20% of the extraterrestrial radiation is lost while passing through the atmosphere under cloudless skies. Using this assumption, the maximum possible solar radiation, H_{MX}, at a particular location on the earth's surface is calculated as:

$$H_{MX} = 30.0 E_0 \left[\omega T_{SR} \sin\delta \sin\phi + \cos\delta \cos\phi \sin(\omega T_{SR})\right] \qquad 2.2.7$$

where the maximum possible solar radiation, H_{MX}, is the amount of radiation reaching the earth's surface under a clear sky (MJ m^{-2} d^{-1}).

2.2.3 DAILY SOLAR RADIATION

The solar radiation reaching the earth's surface on a given day, H_{day}, may be less than H_{MX} due to the presence of cloud cover. The daily solar radiation data required by SWAT may be read from an input file or generated by the model.

The variable SLRSIM in the input control code (.cod) file identifies the method used to obtain solar radiation data. To read in daily solar radiation data, the variable is set to 1 and the name of the solar radiation data file and the number of solar radiation records stored in the file are set in the control input/output (file.cio) file. To generate daily solar radiation values, SLRSIM is set to 2. The equations used to generate solar radiation data in SWAT are reviewed in Chapter 4. SWAT input variables that pertain to solar radiation are summarized in Table 2-2.

Table 2-2: SWAT input variables used in solar radiation calculations.

Variable name	Definition	File Name
LATITUDE	Latitude of the subbasin (degrees).	.sub
SLRSIM	Solar radiation input code: 1-measured, 2-generated	.cod
NSTOT	Number of solar radiation records within the .slr file (required if SLRSIM = 1)	file.cio
SLRFILE	Name of measured solar radiation input file (.slr) (required if SLRSIM = 1)	file.cio
ISGAGE	Number of solar radiation record used within the subbasin (required if SLRSIM = 1)	file.cio
see description of .slr file in the User's Manual for input and format requirements if measured daily solar radiation data is being used		

2.2.4 HOURLY SOLAR RADIATION

The extraterrestrial radiation falling on a horizontal surface during one hour is given by the equation:

$$I_0 = I_{SC} E_0 \left(\sin\delta \sin\phi + \cos\delta \cos\phi \cos\omega t \right) \qquad 2.2.8$$

where I_0 is the extraterrestrial radiation for 1 hour centered around the hour angle ωt.

An accurate calculation of the radiation for each hour of the day requires a knowledge of the difference between standard time and solar time for the location. SWAT simplifies the hourly solar radiation calculation by assuming that solar noon occurs at 12:00pm local standard time.

When the values of I_0 calculated for every hour between sunrise and sunset are summed, they will equal the value of H_0. Because of the relationship between I_0 and H_0, it is possible to calculate the hourly radiation values by multiplying H_0 by the fraction of radiation that falls within the different hours of the day. The benefit of this alternative method is that assumptions used to estimate the difference between maximum and actual solar radiation reaching the earth's surface can be automatically incorporated in calculations of hourly solar radiation at the earth's surface.

SWAT calculates hourly solar radiation at the earth's surface with the equation:

$$I_{hr} = I_{frac} \cdot H_{day} \qquad 2.2.9$$

where I_{hr} is the solar radiation reaching the earth's surface during a specific hour of the day (MJ m^{-2} hr^{-1}), I_{frac} is the fraction of total daily radiation falling during that hour, and H_{day} is the total solar radiation reaching the earth's surface on that day.

The fraction of total daily radiation falling during an hour is calculated

$$I_{frac} = \frac{\left(\sin\delta \sin\phi + \cos\delta \cos\phi \cos\omega t_i \right)}{\sum_{t=SR}^{SS} \left(\sin\delta \sin\phi + \cos\delta \cos\phi \cos\omega t \right)} \qquad 2.2.10$$

where t_i is the solar time at the midpoint of hour i.

2.2.5 DAILY NET RADIATION

Net radiation requires the determination of both incoming and reflected short-wave radiation and net long-wave or thermal radiation. Expressing net radiation in terms of the net short-wave and long-wave components gives:

$$H_{net} = H_{day} \downarrow - \alpha \cdot H_{day} \uparrow + H_L \downarrow - H_L \uparrow \qquad 2.2.11$$

or

$$H_{net} = (1-\alpha) \cdot H_{day} + H_b \qquad 2.2.12$$

where H_{net} is the net radiation (MJ m^{-2} d^{-1}), H_{day} is the short-wave solar radiation reaching the ground (MJ m^{-2} d^{-1}), α is the short-wave reflectance or albedo, H_L is the long-wave radiation (MJ m^{-2} d^{-1}), H_b is the net incoming long-wave radiation (MJ m^{-2} d^{-1}) and the arrows indicate the direction of the radiation flux.

2.2.5.1 NET SHORT-WAVE RADIATION

Net short-wave radiation is defined as $(1-\alpha) \cdot H_{day}$. SWAT calculates a daily value for albedo as a function of the soil type, plant cover, and snow cover. When the snow water equivalent is greater than 0.5 mm,

$$\alpha = 0.8 \qquad 2.2.13$$

When the snow water equivalent is less than 0.5 mm and no plants are growing in the HRU,

$$\alpha = \alpha_{soil} \qquad 2.2.14$$

where α_{soil} is the soil albedo. When plants are growing and the snow water equivalent is less than 0.5 mm,

$$\alpha = \alpha_{plant} \cdot (1 - cov_{sol}) + \alpha_{soil} \cdot cov_{sol} \qquad 2.2.15$$

where α_{plant} is the plant albedo (set at 0.23), and cov_{sol} is the soil cover index. The soil cover index is calculated

$$cov_{sol} = \exp(-5.0 \times 10^{-5} \cdot CV) \qquad 2.2.16$$

where CV is the aboveground biomass and residue (kg ha^{-1}).

2.2.5.2 NET LONG-WAVE RADIATION

Long-wave radiation is emitted from an object according to the radiation law:

$$H_R = \varepsilon \cdot \sigma \cdot T_K^{\ 4} \qquad 2.2.17$$

where H_R is the radiant energy (MJ m^{-2} d^{-1}), ε is the emissivity, σ is the Stefan-Boltzmann constant (4.903×10^{-9} MJ m^{-2} K^{-4} d^{-1}), and T_K is the mean air temperature in Kelvin (273.15 + °C).

Net long-wave radiation is calculated using a modified form of equation 2.2.17 (Jensen et al., 1990):

$$H_b = f_{cld} \cdot (\varepsilon_a - \varepsilon_{vs}) \cdot \sigma \cdot T_K^{\ 4} \qquad 2.2.18$$

where H_b is the net long-wave radiation (MJ m^{-2} d^{-1}), f_{cld} is a factor to adjust for cloud cover, ε_a is the atmospheric emittance, and ε_{vs} is the vegetative or soil emittance.

Wright and Jensen (1972) developed the following expression for the cloud cover adjustment factor, f_{cld}:

$$f_{cld} = a \cdot \frac{H_{day}}{H_{MX}} - b \qquad 2.2.19$$

where a and b are constants, H_{day} is the solar radiation reaching the ground surface on a given day (MJ m^{-2} d^{-1}), and H_{MX} is the maximum possible solar radiation to reach the ground surface on a given day (MJ m^{-2} d^{-1}).

The two emittances in equation 2.2.18 may be combined into a single term, the net emittance ε'. The net emittance is calculated using an equation developed by Brunt (1932):

$$\varepsilon' = \varepsilon_a - \varepsilon_{vs} = -\left(a_1 + b_1 \cdot \sqrt{e}\right) \qquad 2.2.20$$

where a_1 and b_1 are constants and e is the vapor pressure on a given day (kPa). The calculation of e is given in Chapter 3.

Combining equations 2.2.18, 2.2.19, and 2.2.20 results in a general equation for net long-wave radiation:

$$H_b = -\left[a \cdot \frac{H_{day}}{H_{MX}} - b\right] \cdot \left[a_1 + b_1\sqrt{e}\right] \cdot \sigma \cdot T_K^{\ 4} \qquad 2.2.21$$

Experimental values for the coefficients a, b, a_1 and b_1 are presented in Table 2.3. The default equation in SWAT uses coefficent values proposed by Doorenbos and Pruitt (1977):

$$H_b = -\left[0.9 \cdot \frac{H_{day}}{H_{MX}} + 0.1\right] \cdot \left[0.34 - 0.139\sqrt{e}\right] \cdot \sigma \cdot T_K^{\ 4} \qquad 2.2.22$$

Table 2-3: Experimental coefficients for net long-wave radiation equations (from Jensen et al., 1990)

Region	**(a,**	**b)**	**(a_1,**	**b_1)**
Davis, California	(1.35,	-0.35)	(0.35,	-0.145)
Southern Idaho	(1.22,	-0.18)	(0.325,	-0.139)
England	not available		(0.47,	-0.206)
England	not available		(0.44,	-0.253)
Australia	not available		(0.35,	-0.133)
General	(1.2,	-0.2)	(0.39,	-0.158)
General-humid areas	(1.0,	0.0)		
General-semihumid areas	(1.1,	-0.1)		

Table 2-4: SWAT input variables used in net radiation calculations.

Variable name	**Definition**	**File Name**
SOL_ALB	α_{soil}: moist soil albedo	.sol
MAX TEMP	T_{mx}: Daily maximum temperature (°C)	.tmp
MIN TEMP	T_{mn}: Daily minimum temperature (°C)	.tmp
SOL_RAD	H_{day}: Daily solar radiation reaching the earth's surface (MJ m^{-2} d^{-1})	.slr

2.3 TEMPERATURE

Temperature influences a number of physical, chemical and biological processes. Plant production is strongly temperature dependent, as are organic matter decomposition and mineralization. Daily air temperature may be input to the model or generated from average monthly values. Soil and water temperatures are derived from air temperature.

2.3.1 DAILY AIR TEMPERATURE

SWAT requires daily maximum and minimum air temperature. This data may be read from an input file or generated by the model. The user is strongly recommended to obtain measured daily temperature records from gages in or near the watershed if at all possible. The accuracy of model results is significantly improved by the use of measured temperature data.

The variable TMPSIM in the input control code (.cod) file identifies the method used to obtain air temperature data. To read in daily maximum and minimum air temperature data, the variable is set to 1 and the name of the temperature data file(s) and the number of temperature records stored in the file are set in the control input/output (file.cio) file. To generate daily air temperature values, TMPSIM is set to 2. The equations used to generate air temperature data in SWAT are reviewed in Chapter 4. SWAT input variables that pertain to air temperature are summarized in Table 2-5.

Table 2-5: SWAT input variables that pertain to daily air temperature.

Variable name	Definition	File Name
TMPSIM	Air temperature input code: 1-measured, 2-generated	.cod
NTGAGE	Number of temperature gage (.tmp) files used in simulation file (required if TMPSIM = 1)	file.cio
NTTOT	Number of temperature records used in simulation (required if TMPSIM = 1)	file.cio
NTFIL	Number of temperature records within each .tmp file file (required if TMPSIM = 1)	file.cio
TFILE	Name of measured temperature input file (.tmp) Up to 18 files may be used. (required if TMPSIM = 1)	file.cio
ITGAGE	Number of temperature record used within the subbasin (required if TMPSIM = 1)	file.cio
see description of .tmp file in the User's Manual for input and format requirements if measured temperature data is being used		

2.3.2 HOURLY AIR TEMPERATURE

Air temperature data are usually provided in the form of daily maximum and minimum temperature. A reasonable approximation for converting these to hourly temperatures is to assume a sinusoidal interpolation function between the minimum and maximum daily temperatures. The maximum daily temperature is assumed to occur at 1500 hours and the minimum daily temperature at 300 hours (Campbell, 1985). The temperature for the hour is then calculated with the equation:

$$T_{hr} = \overline{T}_{av} + \frac{(T_{mx} - T_{mn})}{2} \cdot \cos(0.2618 \cdot (hr - 15)) \qquad 2.3.1$$

where T_{hr} is the air temperature during hour hr of the day (°C), $\overline{T}_{av}$ is the average temperature on the day (°C), T_{mx} is the daily maximum temperature (°C), and T_{mn} is the daily minimum temperature (°C).

Table 2-6: SWAT input variables that pertain to hourly air temperature.

Variable name	Definition	File Name
MAX TEMP	T_{mx}: Daily maximum temperature (°C)	.tmp
MIN TEMP	T_{mn}: Daily minimum temperature (°C)	.tmp

2.3.3 SOIL TEMPERATURE

Soil temperature will fluctuate due to seasonal and diurnal variations in temperature at the surface. Figure 2-2 plots air temperature and soil temperature at 5 cm and 300 cm below bare soil at College Station, Texas.

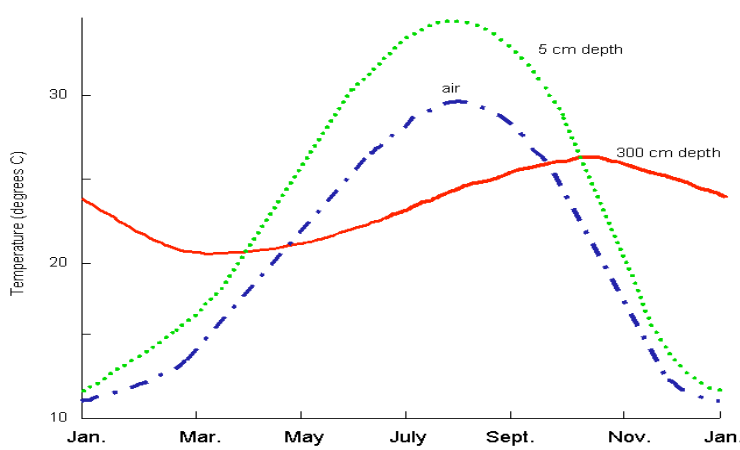

Figure 2-2: Four-year average air and soil temperature at College Station, Texas.

This figure illustrates several important attributes of temperature variation in the soil. First, the annual variation in soil temperature follows a sinusoidal function. Second, the fluctuation in temperature during the year (the amplitude of the sine wave) decreases with depth until, at some depth in the soil, the temperature remains constant throughout the year. Finally, the timing of maximum and minimum temperatures varies with depth. Note in the above graph that there is a three month difference between the recording of the minimum temperature at the surface (January) and the minimum temperature at 300 cm (March).

Carslaw and Jaeger (1959) developed an equation to quantify the seasonal variation in temperature:

$$T_{soil}(z, d_n) = \overline{T}_{AA} + A_{surf} \exp(-z/dd)\sin(\omega_{tmp} d_n - z/dd) \qquad 2.3.2$$

where $T_{soil}(z,d_n)$ is the soil temperature (°C) at depth z (mm) and day of the year d_n, $\overline{T}_{AA}$ is the average annual soil temperature (°C), A_{surf} is the amplitude of the surface fluctuations (°C), dd is the damping depth (mm), and ω_{tmp} is the angular frequency. When z = 0 (soil surface), equation 2.3.2 reduces to $T_{soil}(0, d_n) = \overline{T}_{AA} + A_{surf} \sin(\omega_{tmp} d_n)$. As $z \rightarrow \infty$, equation 2.3.2 becomes $T_{soil}(\infty, d_n) = \overline{T}_{AA}$.

In order to calculate values for some of the variables in this equation, the heat capacity and thermal conductivity of the soil must be known. These are properties not commonly measured in soils and attempts at estimating values from other soil properties have not proven very effective. Consequently, an equation has been adopted in SWAT that calculates the temperature in the soil as a function of the previous day's soil temperature, the average annual air temperature, the current day's soil surface temperature, and the depth in the profile.

The equation used to calculate daily average soil temperature at the center of each layer is:

$$T_{soil}(z, d_n) = \ell \cdot T_{soil}(z, d_n - 1) + [1.0 - \ell] \cdot \left[df \cdot \left[\overline{T}_{AAair} - T_{ssurf}\right] + T_{ssurf}\right] \qquad 2.3.3$$

where $T_{soil}(z,d_n)$ is the soil temperature (°C) at depth z (mm) and day of the year d_n, ℓ is the lag coefficient (ranging from 0.0 to 1.0) that controls the influence of the previous day's temperature on the current day's temperature, $T_{soil}(z,d_n\text{-}1)$ is the soil temperature (°C) in the layer from the previous day, df is the depth factor that quantifies the influence of depth below surface on soil temperature, $\overline{T}_{AAair}$ is the average annual air temperature (°C), and T_{ssurf} is the soil surface temperature on the day. SWAT sets the lag coefficient, ℓ, to 0.80. The soil temperature from the previous day is known and the average annual air temperature is calculated from the long-term monthly maximum and minimum temperatures reported in the weather generator input (.wgn) file. This leaves the depth factor, df, and the soil surface temperature, T_{ssurf}, to be defined.

The depth factor is calculated using the equation:

$$df = \frac{zd}{zd + \exp(-0.867 - 2.078 \cdot zd)} \qquad 2.3.4$$

where zd is the ratio of the depth at the center of the soil layer to the damping depth:

$$zd = \frac{z}{dd} \qquad 2.3.5$$

where z is the depth at the center of the soil layer (mm) and dd is the damping depth (mm).

From the previous three equations (2.3.3, 2.3.4 and 2.3.5) one can see that at depths close to the soil surface, the soil temperature is a function of the soil surface temperature. As the depth increases, soil temperature is increasingly influenced by the average annual air temperature, until at the damping depth, the soil temperature is within 5% of $\overline{T}_{AAair}$.

The damping depth, dd, is calculated daily and is a function of the maximum damping depth, bulk density and soil water. The maximum damping depth, dd_{max}, is calculated:

$$dd_{\max} = 1000 + \frac{2500\rho_b}{\rho_b + 686\exp(-5.63\rho_b)} \qquad 2.3.6$$

where dd_{max} is the maximum damping depth (mm), and ρ_b is the soil bulk density (Mg/m^3).

The impact of soil water content on the damping depth is incorporated via a scaling factor, φ, that is calculated with the equation:

$$\varphi = \frac{SW}{(0.356 - 0.144\rho_b)\cdot z_{tot}} \qquad 2.3.7$$

where SW is the amount of water in the soil profile expressed as depth of water in the profile (mm H_2O), ρ_b is the soil bulk density (Mg/m^3), and z_{tot} is the depth from the soil surface to the bottom of the soil profile (mm).

The daily value for the damping depth, dd, is calculated:

$$dd = dd_{\max}\cdot\exp\left[\ln\left(\frac{500}{dd_{\max}}\right)\cdot\left(\frac{1-\varphi}{1+\varphi}\right)^2\right] \qquad 2.3.8$$

where dd_{max} is the maximum damping depth (mm), and φ is the scaling factor for soil water.

The soil surface temperature is a function of the previous day's temperature, the amount of ground cover and the temperature of the surface when no cover is present. The temperature of a bare soil surface is calculated with the equation:

$$T_{bare} = \overline{T}_{av} + \varepsilon_{sr}\frac{(T_{mx} - T_{mn})}{2} \qquad 2.3.9$$

where T_{bare} is the temperature of the soil surface with no cover (°C), $\overline{T}_{av}$ is the average temperature on the day (°C), T_{mx} is the daily maximum temperature (°C), T_{mn} is the daily minimum temperature (°C), and ε_{sr} is a radiation term. The radiation term is calculated with the equation:

$$\varepsilon_{sr} = \frac{H_{day}\cdot(1-\alpha) - 14}{20} \qquad 2.3.10$$

where H_{day} is the solar radiation reaching the ground on the current day (MJ m^{-2} d^{-1}), and α is the albedo for the day.

Any cover present will significantly impact the soil surface temperature. The influence of plant canopy or snow cover on soil temperature is incorporated with a weighting factor, *bcv*, calculated as:

$$bcv = \max\left\{\begin{array}{c}\dfrac{CV}{CV + \exp\left(7.563 - 1.297\times 10^{-4}\cdot CV\right)} \\ \\ \dfrac{SNO}{SNO + \exp\left(6.055 - 0.3002\cdot SNO\right)}\end{array}\right\} \quad 2.3.11$$

where CV is the total aboveground biomass and residue present on the current day (kg ha^{-1}) and SNO is the water content of the snow cover on the current day (mm H_2O). The weighting factor, *bcv*, is 0.0 for a bare soil and approaches 1.0 as cover increases.

The equation used to calculate the soil surface temperature is:

$$T_{ssurf} = bcv\cdot T_{soil}\left(1, d_n - 1\right) + \left(1 - bcv\right)\cdot T_{bare} \quad 2.3.12$$

where T_{ssurf} is the soil surface temperature for the current day (°C), *bcv* is the weighting factor for soil cover impacts, $T_{soil}(1,d_n\text{-}1)$ is the soil temperature of the first soil layer on the previous day (°C), and T_{bare} is the temperature of the bare soil surface (°C). The influence of ground cover is to place more emphasis on the previous day's temperature near the surface.

SWAT input variables that directly impact soil temperature calculations are listed in Table 2-7. There are several other variables that initialize residue and snow cover in the subbasins or HRUs (SNO_SUB and SNOEB in .sub; RSDIN in .hru). The influence of these variables will be limited to the first few months of simulation. Finally, the timing of management operations in the .mgt file will affect ground cover and consequently soil temperature.

Table 2-7: SWAT input variables that pertain to soil temperature.

Variable name	Definition	File Name
TMPMX	Average maximum air temperature for month (°C)	.wgn
TMPMN	Average minimum air temperature for month (°C)	.wgn
SOL_Z	z: Depth from soil surface to bottom of layer (mm)	.sol
SOL_BD	ρ_b: Moist bulk density (Mg m^{-3} or g cm^{-3})	.sol
SOL_ALB	Moist soil albedo.	.sol
MAX TEMP	T_{mx}: Daily maximum temperature (°C)	.tmp
MIN TEMP	T_{mn}: Daily minimum temperature (°C)	.tmp

2.3.4 WATER TEMPERATURE

Water temperature is required to model in-stream biological and water quality processes. SWAT uses an equation developed by Stefan and Preud'homme (1993) to calculate average daily water temperature for a well-mixed stream:

$$T_{water} = 5.0 + 0.75\overline{T}_{av} \qquad 2.3.13$$

where T_{water} is the water temperature for the day (°C), and $\overline{T}_{av}$ is the average temperature on the day (°C).

Due to thermal inertia of the water, the response of water temperature to a change in air temperature is dampened and delayed. When water and air temperature are plotted for a stream or river, the peaks in the water temperature plots usually lag 3-7 hours behind the peaks in air temperature. As the depth of the river increases, the lag time can increase beyond this typical interval. For very large rivers, the lag time can extend up to a week. Equation 2.3.13 assumes that the lag time between air and water temperatures is less than 1 day.

In addition to air temperature, water temperature is influenced by solar radiation, relative humidity, wind speed, water depth, ground water inflow, artificial heat inputs, thermal conductivity of the sediments and the presence of impoundments along the stream network. SWAT assumes that the impact of these other variables on water temperature is not significant.

Table 2-8: SWAT input variables that pertain to water temperature.

Variable name	Definition	File Name
MAX TEMP	T_{mx}: Daily maximum temperature (°C)	.tmp
MIN TEMP	T_{mn}: Daily minimum temperature (°C)	.tmp

2.4 WIND SPEED

Wind speed is required by SWAT if the Penman-Monteith equation is used to estimate potential evapotranspiration and transpiration. SWAT assumes wind speed information is collected from gages positioned 1.7 meters above the ground surface.

When using the Penman-Monteith equation to estimate transpiration, the wind measurement used in the equation must be above the canopy. In SWAT, a minimum difference of 1 meter is specified for canopy height and wind speed measurements. When the canopy height exceeds 1 meter, the original wind measurements is adjusted to:

$$z_w = h_c + 100 \qquad 2.4.1$$

where z_w is the height of the wind speed measurement (cm), and h_c is the canopy height (cm).

The variation of wind speed with elevation near the ground surface is estimated with the equation (Haltiner and Martin, 1957):

$$u_{z2} = u_{z1} \cdot \left[\frac{z_2}{z_1}\right]^{aa} \qquad 2.4.2$$

where u_{z1} is the wind speed (m s^{-1}) at height z_1 (cm), u_{z2} is the wind speed (m s^{-1}) at height z_2 (cm), and *aa* is an exponent between 0 and 1 that varies with atmospheric stability and surface roughness. Jensen (1974) recommended a value of 0.2 for *aa* and this is the value used in SWAT.

The daily wind speed data required by SWAT may be read from an input file or generated by the model. The variable WNDSIM in the input control code (.cod) file identifies the method used to obtain wind speed data. To read in daily wind speed data, the variable is set to 1 and the name of the wind speed data file and the number of different records stored in the file are set in the control input/output (file.cio) file. To generate daily wind speed values, WNDSIM is set to 2. The equations used to generate wind speed data in SWAT are reviewed in Chapter 4.

Table 2-9: SWAT input variables used in wind speed calculations.

Variable name	Definition	File Name
WNDSIM	Wind speed input code: 1-measured, 2-generated	.cod
NWTOT	Number of wind speed records within the .wnd file (required if WNDSIM = 1)	file.cio
WNDFILE	Name of measured wind speed input file (.wnd) (required if WNDSIM = 1)	file.cio
IWGAGE	Number of wind speed record used within the subbasin (required if WNDSIM = 1)	file.cio
see description of .wnd file in the User's Manual for input and format requirements if measured daily wind speed data is being used		

2.5 NOMENCLATURE

A_{surf}	Amplitude of the surface fluctuations in soil temperature (°C)
AU	Astronomical unit (1 AU = 1.496 x 10^8 km)
CV	Total aboveground biomass and residue present on current day (kg ha^{-1})
E_0	Eccentricity correction factor of earth $(r_0/r)^2$
H_0	Extraterrestrial daily irradiation (MJ m^{-2} d^{-1})
H_b	Net outgoing long-wave radiation (MJ m^{-2} d^{-1})
H_{day}	Solar radiation reaching ground on current day of simulation (MJ m^{-2} d^{-1})
H_L	Long-wave radiation (MJ m^{-2} d^{-1})
H_{MX}	Maximum possible solar radiation (MJ m^{-2} d^{-1})
H_{net}	Net radiation on day (MJ m^{-2} d^{-1})
H_R	Radiant energy (MJ m^{-2} d^{-1})
I_{frac}	Fraction of daily solar radiation falling during specific hour on current day of simulation
I_{hr}	Solar radiation reaching ground during specific hour on current day of simulation (MJ m^{-2} h^{-1})
I_{SC}	Solar constant (4.921 MJ m^{-2} h^{-1})
I_0	Extraterrestrial daily irradiance incident on a horizontal surface (MJ m^{-2} h^{-1})
I_{0n}	Extraterrestrial daily irradiance incident on a normal surface (MJ m^{-2} h^{-1})
SNO	Water content of snow cover on current day (mm H_2O)
SW	Amount of water in soil profile (mm H_2O)
T_{bare}	Temperature of soil surface with no cover (°C)
T_{DL}	Daylength (h)
T_{hr}	Air temperature during hour (°C)
T_K	Mean air temperature in Kelvin (273.15 + °C)
T_{mn}	Minimum air temperature for day (°C)
T_{mx}	Maximum air temperature for day (°C)
T_{soil}	Soil temperature (°C)
T_{ssurf}	Soil surface temperature (°C)
T_{SR}	Time of sunrise in solar day (h)
T_{SS}	Time of sunset in solar day (h)
T_{water}	Average daily water temperature (°C)
$\overline{T}_{AA}$	Average annual soil temperature (°C)

$\overline{T}_{AAair}$	Average annual air temperature (°C)
$\overline{T}_{av}$	Average air temperature for day (°C)
a	Constant in equation used to calculate the cloud cover adjustment factor
a_1	Constant in equation used to calculate net emissivity
aa	Exponent between 0 and 1 that varies with atmospheric stability and surface roughness that is used in calculating wind speed at different heights
b	Constant in equation used to calculate the cloud cover adjustment factor
b_1	Constant in equation used to calculate net emissivity
bcv	weighting factor for impact of ground cover on soil surface temperature
cov_{sol}	Soil cover index for albedo determination
d_n	Day number of year, 1 on January 1 and 365 on December 31
dd	Damping depth (mm)
dd_{max}	Maximum damping depth (mm)
df	Depth factor used in soil temperature calculations
e	Vapor pressure (actual) on a given day (kPa)
f_{cld}	Factor to adjust for cloud cover in net long-wave radiation calculation
h_c	Canopy height (cm)
hr	Hour of day (1-24)
r	Actual earth-sun distance (AU)
r_0	Mean earth-sun distance, 1 AU
t	Number of hours before (+) or after (-) solar noon
t_i	Solar time at the midpoint of the hour i
u_{z1}	Wind speed (m s^{-1}) at height z_1 (cm)
u_{z2}	Wind speed (m s^{-1}) at height z_2 (cm)
z	Depth below soil surface (mm)
z_1	Height of wind speed measurement (cm)
z_2	Height of wind speed measurement (cm)
z_{tot}	Depth to bottom of soil profile (mm)
z_w	Height of the wind speed measurement (cm)
zd	Ratio of depth in soil to damping depth
α	Short-wave reflectance or albedo
α_{plant}	Plant albedo (set at 0.23)
α_{soil}	Soil albedo
δ	Solar declination (radians)
ε	Emissivity
ε'	Net emittance
ε_a	Atmospheric emittance
ε_{sr}	Radiation term for bare soil surface temperature calculation
ε_{vs}	Vegetative or soil emittance
ℓ	Lag coefficient that controls influence of previous day's temperature on current days temperature
σ	Stefan-Boltzmann constant (4.903×10^{-9} MJ m^{-2} K^{-4} d^{-1})
θ_z	Zenith angle (radians)

ϕ	Latitude in radians
ρ_b	Soil bulk density (Mg m^{-3})
φ	Scaling factor for impact of soil water on damping depth
ω	Angular velocity of the earth's rotation (0.2618 radians h^{-1})
ω_{tmp}	Angular frequency in soil temperature variation

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CHAPTER 3

EQUATIONS: ATMOSPHERIC WATER

Precipitation is the mechanism by which water enters the land phase of the hydrologic cycle. Because precipitation controls the water balance, it is critical that the amount and distribution of precipitation in space and time is accurately simulated by the model.

3.1 PRECIPITATION

The precipitation reaching the earth's surface on a given day, R_{day}, may be read from an input file or generated by the model. Users are strongly recommended to incorporate measured precipitation into their simulations any time the data is available. The ability of SWAT to reproduce observed stream hydrographs is greatly improved by the use of measured precipitation data.

Unfortunately, even with the use of measured precipitation the model user can expect some error due to inaccuracy in precipitation data. Measurement of precipitation at individual gages is subject to error from a number of causes and additional error is introduced when regional precipitation is estimated from point values. Typically, total or average areal precipitation estimates for periods of a year or longer have relative uncertainties of 10% (Winter, 1981).

Point measurements of precipitation generally capture only a fraction of the true precipitation. The inability of a gage to capture a true reading is primarily caused by wind eddies created by the gage. These wind eddies reduce the catch of the smaller raindrops and snowflakes. Larson and Peck (1974) found that deficiencies of 10% for rain and 30% for snow are common for gages projecting above the ground surface that are not designed to shield wind effects. Even when the gage is designed to shield for wind effects, this source of error will not be eliminated. For an in-depth discussion of this and other sources of error as well as methods for dealing with the error, please refer to Dingman (1994).

The variable PCPSIM in the input control code (.cod) file identifies the method used to obtain precipitation data. To read in daily precipitation data, the variable is set to 1 and the names of the precipitation data files and the number of precipitation records stored in the files are defined in the control input/output (file.cio) file. To generate daily precipitation values, PCPSIM is set to 2. The equations used to generate precipitation data in SWAT are reviewed in Chapter 4. SWAT input variables that pertain to precipitation are summarized in Table 3-1.

Table 3-1: SWAT input variables used in precipitation calculations.

Variable name	Definition	File Name
PCPSIM	Precipitation input code: 1-measured, 2-generated	.cod
NRGAGE	Number of precipitation gage files (.pcp) used (required if PCPSIM = 1)	file.cio
NRTOT	Total number of precipitation records used in simulation (required if PCPSIM = 1)	file.cio
NRGFIL	Number of precipitation records in each .pcp file (required if PCPSIM = 1)	file.cio
RFILE	Name of measured precipitation input file(s) (.pcp) (required if PCPSIM = 1)	file.cio
IRGAGE	Number of precipitation record used within the subbasin (required if PCPSIM = 1)	file.cio
see description of .pcp file in the User's Manual for input and format requirements if measured daily precipitation data is being used		

3.2 MAXIMUM HALF-HOUR RAINFALL

The maximum half-hour rainfall is required by SWAT to calculate the peak runoff rate. The maximum half-hour rainfall is reported as a fraction of the total daily rainfall, $\alpha_{0.5}$. If sub-daily precipitation data is used in the model, SWAT will calculate the maximum half-hour rainfall fraction directly from the precipitation data. If daily precipitation data is used, SWAT generates a value for $\alpha_{0.5}$ using the equations summarized in Chapter 4.

3.3 WATER VAPOR

Relative humidity is required by SWAT if the Penman-Monteith or Priestley-Taylor equation is used to estimate potential evapotranspiration. The Penman-Monteith equation includes terms that quantify the effect of the amount of water vapor in the air near the evaporative surface on evaporation. Both Penman-Monteith and Priestley-Taylor require the actual vapor pressure, which is calculated from the relative humidity.

Relative humidity is the ratio of an air volume's actual vapor pressure to its saturation vapor pressure:

$$R_h = \frac{e}{e^o} \qquad 3.3.1$$

where R_h is the relative humidity on a given day, e is the actual vapor pressure on a given day (kPa), and e^o is the saturation vapor pressure on a given day (kPa).

The saturation vapor pressure is the maximum vapor pressure that is thermodynamically stable and is a function of the air temperature. SWAT calculates saturation vapor pressure using an equation presented by Tetens (1930) and Murray (1967):

$$e^o = \exp\left[\frac{16.78 \cdot \overline{T}_{av} - 116.9}{\overline{T}_{av} + 237.3}\right] \qquad 3.3.2$$

where e^o is the saturation vapor pressure on a given day (kPa) and $\overline{T}_{av}$ is the mean daily air temperature (°C). When relative humidity is known, the actual vapor pressure can be calculated by rearranging equation 3.3.1:

$$e = R_h \cdot e^o \qquad 3.3.3$$

The saturation vapor pressure curve is obtained by plotting equation 3.3.2. The slope of the saturation vapor pressure curve can be calculated by differentiating equation 3.3.2:

$$\Delta = \frac{4098 \cdot e^o}{\left(\overline{T}_{av} + 237.3\right)^2} \qquad 3.3.4$$

where Δ is the slope of the saturation vapor pressure curve (kPa °C^{-1}), e^o is the saturation vapor pressure on a given day (kPa) and $\overline{T}_{av}$ is the mean daily air temperature (°C).

The rate of evaporation is proportional to the difference between the vapor pressure of the surface layer and the vapor pressure of the overlying air. This difference is termed the vapor pressure deficit:

$$vpd = e^o - e \qquad 3.3.5$$

where *vpd* is the vapor pressure deficit (kPa), e^o is the saturation vapor pressure on a given day (kPa), and *e* is the actual vapor pressure on a given day (kPa). The greater the value of *vpd* the higher the rate of evaporation.

The latent heat of vaporization, λ, is the quantity of heat energy that must be absorbed to break the hydrogen bonds between water molecules in the liquid state to convert them to gas. The latent heat of vaporization is a function of temperature and can be calculated with the equation (Harrison, 1963):

$$\lambda = 2.501 - 2.361\times10^{-3} \cdot \overline{T}_{av} \qquad 3.3.6$$

where λ is the latent heat of vaporization (MJ kg^{-1}) and $\overline{T}_{av}$ is the mean daily air temperature (°C).

Evaporation involves the exchange of both latent heat and sensible heat between the evaporating body and the air. The psychrometric constant, γ, represents a balance between the sensible heat gained from air flowing past a wet bulb thermometer and the sensible heat converted to latent heat (Brunt, 1952) and is calculated:

$$\gamma = \frac{c_p \cdot P}{0.622 \cdot \lambda} \qquad 3.3.7$$

where γ is the psychrometric constant (kPa °C^{-1}), c_p is the specific heat of moist air at constant pressure (1.013×10^{-3} MJ kg^{-1} °C^{-1}), P is the atmospheric pressure (kPa), and λ is the latent heat of vaporization (MJ kg^{-1}).

Calculation of the psychrometric constant requires a value for atmospheric pressure. SWAT estimates atmospheric pressure using an equation developed by Doorenbos and Pruitt (1977) from mean barometric pressure data at a number of East African sites:

$$P = 101.3 - 0.01152 \cdot EL + 0.544\times10^{-6} \cdot EL^2 \qquad 3.3.8$$

where P is the atmospheric pressure (kPa) and EL is the elevation (m).

The daily relative humidity data required by SWAT may be read from an input file or generated by the model. The variable RHSIM in the input control code (.cod) file identifies the method used to obtain relative humidity data. To read in daily relative humidity data, the variable is set to 1 and the name of the relative humidity data file and the number of different records stored in the file are set in the control input/output (file.cio) file. To generate daily relative humidity values, RHSIM is set to 2. The equations used to generate relative humidity data in SWAT are reviewed in Chapter 4.

Table 3-2: SWAT input variables used in relative humidity calculations.

Variable name	Definition	File Name
RHD	R_h: daily average relative humidity	.hmd
TMP_MX	T_{mx}: maximum temperature for day (°C)	.tmp
TMP_MN	T_{mn}: minimum temperature for day (°C)	.tmp
ELEV	*EL*: elevation (m)	.sub
RHSIM	Relative humidity input code: 1-measured, 2-generated	.cod
NHTOT	Number of relative humidity records within the .hmd file (required if RHSIM = 1)	file.cio
RHFILE	Name of measured relative humidity input file (.hmd) (required if RHSIM = 1)	file.cio
IHGAGE	Number of relative humidity record used within the subbasin (required if RHSIM = 1)	file.cio
see description of .hmd file in the User's Manual for input and format requirements if measured relative humidity data is being used		

3.4 SNOW COVER

SWAT classifies precipitation as rain or freezing rain/snow by the mean daily air temperature. The boundary temperature, T_{s-r}, used to categorize precipitation as rain or snow is defined by the user. If the mean daily air temperature is less than the boundary temperature, then the precipitation within the HRU is classified as snow and the water equivalent of the snow precipitation is added to the snow pack.

Snowfall is stored at the ground surface in the form of a snow pack. The amount of water stored in the snow pack is reported as a snow water equivalent. The snow pack will increase with additional snowfall or decrease with snow melt or sublimation. The mass balance for the snow pack is:

$$SNO = SNO + R_{day} - E_{sub} - SNO_{mlt} \tag{3.4.1}$$

where SNO is the water content of the snow pack on a given day (mm H_2O), R_{day} is the amount of precipitation on a given day (added only if $\bar{T}_{av} \le T_{s-r}$) (mm H_2O), E_{sub} is the amount of sublimation on a given day (mm H_2O), and SNO_{mlt} is the amount of snow melt on a given day (mm H_2O). The amount of snow is expressed as depth over the total HRU area.

Due to variables such as drifting, shading and topography, the snow pack in a subbasin will rarely be uniformly distributed over the total area. This results in a fraction of the subbasin area that is bare of snow. This fraction must be quantified to accurately compute snow melt in the subbasin.

The factors that contribute to variable snow coverage are usually similar from year to year, making it possible to correlate the areal coverage of snow with the amount of snow present in the subbasin at a given time. This correlation is expressed as an areal depletion curve, which is used to describe the seasonal growth and recession of the snow pack as a function of the amount of snow present in the subbasin (Anderson, 1976).

The areal depletion curve requires a threshold depth of snow, SNO_{100}, to be defined above which there will always be 100% cover. The threshold depth will depend on factors such as vegetation distribution, wind loading of snow, wind scouring of snow, interception and aspect, and will be unique to the watershed of interest.

The areal depletion curve is based on a natural logarithm. The areal depletion curve equation is:

$$sno_{\mathrm{cov}} = \frac{SNO}{SNO_{100}} \cdot \left(\frac{SNO}{SNO_{100}} + \exp\left(cov_1 - cov_2 \cdot \frac{SNO}{SNO_{100}} \right) \right)^{-1} \qquad 3.4.2$$

where sno_{cov} is the fraction of the HRU area covered by snow, SNO is the water content of the snow pack on a given day (mm H_2O), SNO_{100} is the threshold depth of snow at 100% coverage (mm H_2O), cov_1 and cov_2 are coefficients that define the shape of the curve. The values used for cov_1 and cov_2 are determined by solving equation 3.4.2 using two known points: 95% coverage at 95% SNO_{100}; and 50% coverage at a user specified fraction of SNO_{100}.

Example areal depletion curves for various fractions of SNO_{100} at 50% coverage are shown in the following figures.

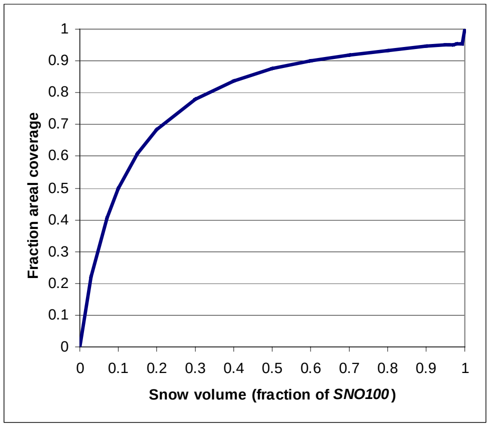

Figure 3-1:10% SNO_{100} = 50% coverage

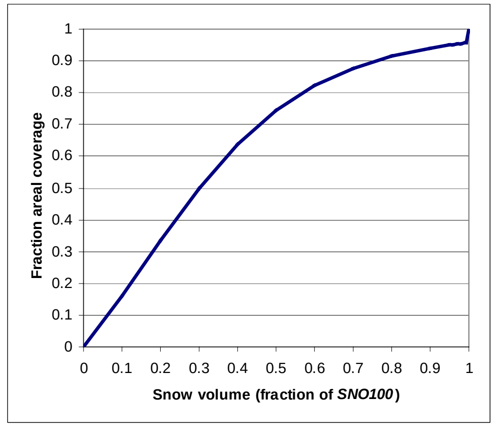

Figure 3-2: 30% SNO_{100} = 50% coverage

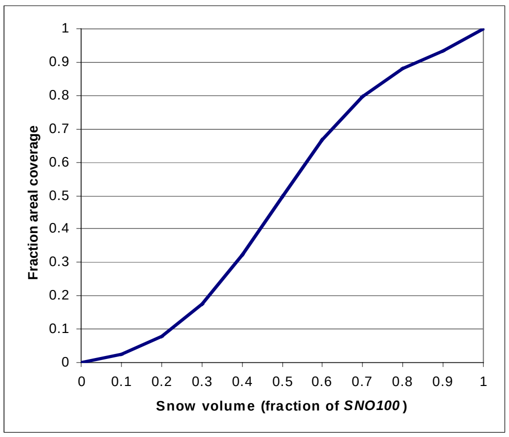

Figure 3-3: 50% SNO_{100} = 50% coverage

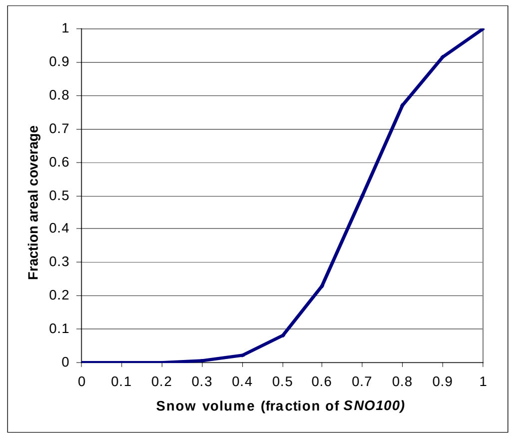

Figure 3-4: 70% SNO_{100} = 50% coverage

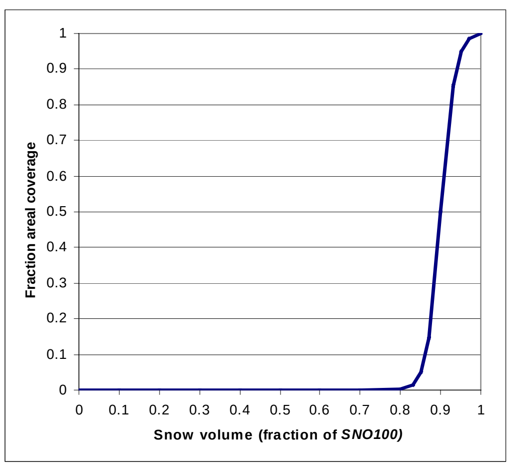

Figure 3-5: 90% SNO_{100} = 50% coverage

It is important to remember that once the volume of water held in the snow pack exceeds SNO_{100} the depth of snow over the HRU is assumed to be uniform, i.e. sno_{cov} = 1.0. The areal depletion curve affects snow melt only when the snow pack water content is between 0.0 and SNO_{100}. Consequently if SNO_{100} is set to a very small value, the impact of the areal depletion curve on snow melt will be minimal. As the value for SNO_{100} increases, the influence of the areal depletion curve will assume more importance in snow melt processes.

Table 3-3: SWAT input variables used in snow cover calculations.

Variable name	Definition	File Name
SFTMP	$T_{s\text{-}r}$: Mean air temperature at which precipitation is equally likely to be rain as snow/freezing rain (°C)	.bsn
SNOCOVMX	SNO_{100}: Threshold depth of snow, above which there is 100% cover	.bsn
SNO50COV	Fraction of SNOCOVMX that provides 50% cover	.bsn
SNO_SUB	Initial snow water content in subbasin (mm H_2O)	.sub
SNOEB	Initial snow water content in subbasin elevation band (mm H_2O)	.sub

3.5 SNOW MELT

Snow melt is controlled by the air and snow pack temperature, the melting rate, and the areal coverage of snow.

Snow melt is included with rainfall in the calculations of runoff and percolation. When SWAT calculates erosion, the rainfall energy of the snow melt fraction of the water is set to zero. The water released from snow melt is assumed to be evenly distributed over the 24 hours of the day.

3.5.1 SNOW PACK TEMPERATURE

The snow pack temperature is a function of the mean daily temperature during the preceding days and varies as a dampened function of air temperature (Anderson, 1976). The influence of the previous day's snow pack temperature on the current day's snow pack temperature is controlled by a lagging factor, ℓ_{sno}. The lagging factor inherently accounts for snow pack density, snow pack depth, exposure and other factors affecting snow pack temperature. The equation used to calculate the snow pack temperature is:

$$T_{snow(d_n)} = T_{snow(d_n-1)} \cdot \left(1 - \ell_{sno}\right) + \overline{T}_{av} \cdot \ell_{sno} \qquad 3.5.1$$

where $T_{snow(d_n)}$ is the snow pack temperature on a given day (°C), $T_{snow(d_n-1)}$ is the snow pack temperature on the previous day (°C), ℓ_{sno} is the snow temperature lag factor, and $\overline{T}_{av}$ is the mean air temperature on the current day (°C). As ℓ_{sno} approaches 1.0, the mean air temperature on the current day exerts an increasingly greater influence on the snow pack temperature and the snow pack temperature from the previous day exerts less and less influence.

The snow pack will not melt until the snow pack temperature exceeds a threshold value, T_{mlt}. This threshold value is specified by the user.

3.5.2 SNOW MELT EQUATION

The snow melt in SWAT is calculated as a linear function of the difference between the average snow pack-maximum air temperature and the base or threshold temperature for snow melt:

$$SNO_{mlt} = b_{mlt} \cdot sno_{\text{cov}} \cdot \left[\frac{T_{snow} + T_{mx}}{2} - T_{mlt}\right] \qquad 3.5.2$$

where SNO_{mlt} is the amount of snow melt on a given day (mm H_2O), b_{mlt} is the melt factor for the day (mm H_2O/day-°C), sno_{cov} is the fraction of the HRU area covered by snow, T_{snow} is the snow pack temperature on a given day (°C), T_{mx} is the maximum air temperature on a give day (°C), and T_{mlt} is the base temperature above which snow melt is allowed (°C).

The melt factor is allowed a seasonal variation with maximum and minimum values occurring on summer and winter solstices:

$$b_{mlt} = \frac{(b_{mlt6} + b_{mlt12})}{2} + \frac{(b_{mlt6} - b_{mlt12})}{2} \cdot \sin\left(\frac{2\pi}{365} \cdot (d_n - 81)\right) \qquad 3.5.3$$

where b_{mlt} is the melt factor for the day (mm H_2O/day-°C), b_{mlt6} is the melt factor for June 21 (mm H_2O/day-°C), b_{mlt12} is the melt factor for December 21 (mm H_2O/day-°C), and d_n is the day number of the year.

In rural areas, the melt factor will vary from 1.4 to 6.9 mm H_2O/day-°C (Huber and Dickinson, 1988). In urban areas, values will fall in the higher end of the range due to compression of the snow pack by vehicles, pedestrians, etc.

Urban snow melt studies in Sweden (Bengston, 1981; Westerstrom, 1981) reported melt factors ranging from 3.0 to 8.0 mm H_2O/day-°C. Studies of snow melt on asphalt (Westerstrom, 1984) gave melt factors of 1.7 to 6.5 mm H_2O/day-°C.

Table 3-4: SWAT input variables used in snow melt calculations.

Variable name	Definition	File Name
TIMP	ℓ_{sno} : Snow temperature lag factor	.bsn
SMTMP	T_{mlt}: Threshold temperature for snow melt (°C)	.bsn
SMFMX	b_{mlt6}: Melt factor on June 21 (mm H_2O/day-°C)	.bsn
SMFMN	b_{mlt12}: Melt factor on December 21 (mm H_2O/day-°C)	.bsn

3.6 NOMENCLATURE

E_{sub} Amount of sublimation on a given day (mm H_2O)
EL Elevation (m)
P Atmospheric pressure (kPa)
R_{day} Amount of rainfall on a given day (mm H_2O)
R_h Average relative humidity for the day
SNO Water content of snow cover on current day (mm H_2O)
SNO_{100} Amount of snow above which there is 100% cover (mm H_2O)
SNO_{mlt} Amount of snow melt on a given day (mm H_2O)
T_{mlt} Threshold temperature for snow melt (°C)
T_{mx} Maximum air temperature for day (°C)
$T_{s\text{-}r}$ Rain/snow boundary temperature (°C)
T_{snow} Snow pack temperature on a given day (°C)
$\overline{T}_{av}$ Average air temperature for day (°C)

b_{mlt} Melt factor for the day (mm H_2O/day-°C)
b_{mlt6} Melt factor for June 21 (mm H_2O/day-°C)
b_{mlt12} Melt factor for December 21 (mm H_2O/day-°C)
c_p Specific heat of moist air at constant pressure (1.013×10^{-3} MJ kg^{-1} °C^{-1})
cov_1 Snow cover areal depletion curve shape coefficient
cov_2 Snow cover areal depletion curve shape coefficient
d_n Day number of year, 1 on January 1 and 365 on December 31
e Actual vapor pressure on a given day (kPa)
e^o Saturation vapor pressure on a given day (kPa)
sno_{cov} Fraction of the HRU area covered by snow
vpd Vapor pressure deficit (kPa)

$\alpha_{0.5}$ Maximum half-hour rainfall expressed as a fraction of daily rainfall
Δ Slope of the saturation vapor pressure curve (kPa °C^{-1})
γ Psychrometric constant (kPa °C^{-1})
λ Latent heat of vaporization (MJ kg^{-1})
ℓ_{sno} Snow temperature lag factor

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CHAPTER **4**

EQUATIONS: WEATHER GENERATOR

SWAT requires daily values of precipitation, maximum and minimum temperature, solar radiation, relative humidity and wind speed. The user may choose to read these inputs from a file or generate the values using monthly average data summarized over a number of years.

SWAT includes the WXGEN weather generator model (Sharpley and Williams, 1990) to generate climatic data or to fill in gaps in measured records. This weather generator was developed for the contiguous U.S. If the user prefers a different weather generator, daily input values for the different weather parameters may be generated with an alternative model and formatted for input to SWAT.

The occurrence of rain on a given day has a major impact on relative humidity, temperature and solar radiation for the day. The weather generator first independently generates precipitation for the day. Maximum temperature, minimum temperature, solar radiation and relative humidity are then generated based on the presence or absence of rain for the day. Finally, wind speed is generated independently.

4.1 PRECIPITATION

The precipitation generator is a Markov chain-skewed (Nicks, 1974) or Markov chain-exponential model (Williams, 1995). A first-order Markov chain is used to define the day as wet or dry. When a wet day is generated, a skewed distribution or exponential distribution is used to generate the precipitation amount. Table 4.1 lists SWAT input variables that are used in the precipitation generator.

4.1.1 OCCURRENCE OF WET OR DRY DAY

With the first-order Markov-chain model, the probability of rain on a given day is conditioned on the wet or dry status of the previous day. A wet day is defined as a day with 0.1 mm of rain or more.

The user is required to input the probability of a wet day on day i given a wet day on day $i - 1$, $P_i(W/W)$, and the probability of a wet day on day i given a dry day on day $i - 1$, $P_i(W/D)$, for each month of the year. From these inputs the remaining transition probabilities can be derived:

$$P_i(D/W) = 1 - P_i(W/W) \qquad 4.1.1$$

$$P_i(D/D) = 1 - P_i(W/D) \qquad 4.1.2$$

where $P_i(D/W)$ is the probability of a dry day on day i given a wet day on day i – 1 and $P_i(D/D)$ is the probability of a dry day on day i given a dry day on day $i - 1$.

To define a day as wet or dry, SWAT generates a random number between 0.0 and 1.0. This random number is compared to the appropriate wet-dry probability, $P_i(W/W)$ or $P_i(W/D)$. If the random number is equal to or less than the wet-dry probability, the day is defined as wet. If the random number is greater than the wet-dry probability, the day is defined as dry.

4.1.2 AMOUNT OF PRECIPITATION

Numerous probability distribution functions have been used to describe the distribution of rainfall amounts. SWAT provides the user with two options: a skewed distribution and an exponential distribution.

The skewed distribution was proposed by Nicks (1974) and is based on a skewed distribution used by Fiering (1967) to generate representative streamflow. The equation used to calculate the amount of precipitation on a wet day is:

$$R_{day} = \mu_{mon} + 2 \cdot \sigma_{mon} \cdot \left(\frac{\left[\left(SND_{day} - \frac{g_{mon}}{6} \right) \cdot \left(\frac{g_{mon}}{6} \right) + 1 \right]^3 - 1}{g_{mon}} \right) \qquad 4.1.3$$

where R_{day} is the amount of rainfall on a given day (mm H_2O), μ_{mon} is the mean daily rainfall (mm H_2O) for the month, σ_{mon} is the standard deviation of daily rainfall (mm H_2O) for the month, SND_{day} is the standard normal deviate calculated for the day, and g_{mon} is the skew coefficient for daily precipitation in the month.

The standard normal deviate for the day is calculated:

$$SND_{day} = \cos(6.283 \cdot rnd_2) \cdot \sqrt{-2 \ln(rnd_1)} \qquad 4.1.4$$

where rnd_1 and rnd_2 are random numbers between 0.0 and 1.0.

The exponential distribution is provided as an alternative to the skewed distribution. This distribution requires fewer inputs and is most commonly used in areas where limited data on precipitation events is available. Daily precipitation is calculated with the exponential distribution using the equation:

$$R_{day} = \mu_{mon} \cdot (-\ln(rnd_1))^{r\exp} \qquad 4.1.5$$

where R_{day} is the amount of rainfall on a given day (mm H_2O), μ_{mon} is the mean daily rainfall (mm H_2O) for the month, rnd_1 is a random number between 0.0 and 1.0, and *rexp* is an exponent that should be set between 1.0 and 2.0. As the value of *rexp* is increased, the number of extreme rainfall events during the year will increase. Testing of this equation at locations across the U.S. have shown that a value of 1.3 gives satisfactory results.

Table 4-1: SWAT input variables that pertain to generation of precipitation.

Variable name	Definition	File Name
PCPSIM	Precipitation input code: 1-measured, 2-generated	.cod
PR_W(1,mon)	$P_i(W/D)$: probability of a wet day following a dry day in month	.wgn
PR_W(2,mon)	$P_i(W/W)$: probability of a wet day following a wet day in month	.wgn
IDIST	Rainfall distribution code: 0-skewed, 1-exponential	.cod
REXP	*rexp*: value of exponent (required if IDIST = 1)	.cod
PCPMM(mon)	average amount of precipitation falling in month (mm H_2O)	.wgn
PCPD(mon)	average number of days of precipitation in month (μ_{mon} = PCPMM / PCPD)	.wgn
PCPSTD(mon)	σ_{mon}: standard deviation for daily precipitation in month (mm H_2O)	.wgn
PCPSKW(mon)	g_{mon}: skew coefficient for daily precipitation in month	.wgn

4.2 SOLAR RADIATION & TEMPERATURE

The procedure used to generate daily values for maximum temperature, minimum temperature and solar radiation (Richardson, 1981; Richardson and Wright, 1984) is based on the weakly stationary generating process presented by Matalas (1967).

4.2.1 DAILY RESIDUALS

Residuals for maximum temperature, minimum temperature and solar radiation are required for calculation of daily values. The residuals must be serially correlated and cross-correlated with the correlations being constant at all locations. The equation used to calculate residuals is:

$$\chi_i(j) = A\chi_{i-1}(j) + B\varepsilon_i(j) \qquad 4.2.1$$

where $\chi_i(j)$ is a 3×1 matrix for day *i* whose elements are residuals of maximum temperature (j = 1), minimum temperature (j = 2) and solar radiation (j = 3), $\chi_{i-1}(j)$ is a 3×1 matrix of the previous day's residuals, ε_i is a 3×1 matrix of independent random components, and *A* and *B* are 3×3 matrices whose elements are defined such that the new sequences have the desired serial-correlation and cross-correlation coefficients. The *A* and *B* matrices are given by

$$A = M_1 \cdot M_0^{-1} \qquad 4.2.2$$

$$B \cdot B^T = M_0 - M_1 \cdot M_0^{-1} \cdot M_1^T \qquad 4.2.3$$

where the superscript –1 denotes the inverse of the matrix and the superscript T denotes the transpose of the matrix. M_0 and M_1 are defined as

$$M_0 = \begin{bmatrix} 1 & \rho_0(1,2) & \rho_0(1,3) \\ \rho_0(1,2) & 1 & \rho_0(2,3) \\ \rho_0(1,3) & \rho_0(2,3) & 1 \end{bmatrix} \qquad 4.2.4$$

$$M_1 = \begin{bmatrix} \rho_1(1,1) & \rho_1(1,2) & \rho_1(1,3) \\ \rho_1(2,1) & \rho_1(2,2) & \rho_1(2,3) \\ \rho_1(3,1) & \rho_1(3,2) & \rho_1(3,3) \end{bmatrix} \qquad 4.2.5$$

$\rho_0(j,k)$ is the correlation coefficient between variables j and k on the same day where j and k may be set to 1 (maximum temperature), 2 (minimum temperature) or 3 (solar radiation) and $\rho_1(j,k)$ is the correlation coefficient between variable j and k with variable k lagged one day with respect to variable j.

Correlation coefficients were determined for 31 locations in the United States using 20 years of temperature and solar radiation data (Richardson, 1982). Using the average values of these coefficients, the M_0 and M_1 matrices become

$$M_0 = \begin{bmatrix} 1.000 & 0.633 & 0.186 \\ 0.633 & 1.000 & -0.193 \\ 0.186 & -0.193 & 1.000 \end{bmatrix} \qquad 4.2.6$$

$$M_1 = \begin{bmatrix} 0.621 & 0.445 & 0.087 \\ 0.563 & 0.674 & -0.100 \\ 0.015 & -0.091 & 0.251 \end{bmatrix} \qquad 4.2.7$$

Using equations 4.2.2 and 4.2.3, the A and B matrices become

$$A = \begin{bmatrix} 0.567 & 0.086 & -0.002 \\ 0.253 & 0.504 & -0.050 \\ -0.006 & -0.039 & 0.244 \end{bmatrix} \qquad 4.2.8$$

$$B = \begin{bmatrix} 0.781 & 0 & 0 \\ 0.328 & 0.637 & 0 \\ 0.238 & -0.341 & 0.873 \end{bmatrix} \qquad 4.2.9$$

The *A* and *B* matrices defined in equations 4.2.8 and 4.2.9 are used in conjunction with equation 4.2.1 to generate daily sequences of residuals of maximum temperature, minimum temperature and solar radiation.

4.2.2 GENERATED VALUES

The daily generated values are determined by multiplying the residual elements generated with equation 4.2.1 by the monthly standard deviation and adding the monthly average value.

$$T_{mx} = \mu mx_{mon} + \chi_i(1) \cdot \sigma mx_{mon} \qquad 4.2.10$$

$$T_{mn} = \mu mn_{mon} + \chi_i(2) \cdot \sigma mn_{mon} \qquad 4.2.11$$

$$H_{day} = \mu rad_{mon} + \chi_i(3) \cdot \sigma rad_{mon} \qquad 4.2.12$$

where T_{mx} is the maximum temperature for the day (°C), μmx_{mon} is the average daily maximum temperature for the month (°C), $\chi_i(1)$ is the residual for maximum temperature on the given day, σmx_{mon} is the standard deviation for daily maximum temperature during the month (°C), T_{mn} is the minimum temperature for the day (°C), μmn_{mon} is the average daily minimum temperature for the month (°C), $\chi_i(2)$ is the residual for minimum temperature on the given day, σmn_{mon} is the standard deviation for daily minimum temperature during the month (°C), H_{day} is the solar radiation for the day (MJ m^{-2}), μrad_{mon} is the average daily solar radiation for the month (MJ m^{-2}), $\chi_i(3)$ is the residual for solar radiation on the given day, and σrad_{mon} is the standard deviation for daily solar radiation during the month (MJ m^{-2}).

The user is required to input standard deviation for maximum and minimum temperature. For solar radiation the standard deviation is estimated as ¼ of the difference between the extreme and mean value for each month.

$$\sigma rad_{mon} = \frac{H_{mx} - \mu rad_{mon}}{4} \qquad 4.2.13$$

where σrad_{mon} is the standard deviation for daily solar radiation during the month (MJ m^{-2}), H_{mx} is the maximum solar radiation that can reach the earth's surface on a given day (MJ m^{-2}), and μrad_{mon} is the average daily solar radiation for the month (MJ m^{-2}).

4.2.3 ADJUSTMENT FOR CLEAR/OVERCAST CONDITIONS

Maximum temperature and solar radiation will be lower on overcast days than on clear days. To incorporate the influence of wet/dry days on generated values of maximum temperature and solar radiation, the average daily maximum temperature, μmx_{mon}, and average daily solar radiation, μrad_{mon}, in equations 4.2.10 and 4.2.12 are adjusted for wet or dry conditions.

4.2.3.1 MAXIMUM TEMPERATURE

The continuity equation relates average daily maximum temperature adjusted for wet or dry conditions to the average daily maximum temperature for the month:

$$\mu mx_{mon} \cdot days_{tot} = \mu Wmx_{mon} \cdot days_{wet} + \mu Dmx_{mon} \cdot days_{dry} \qquad 4.2.14$$

where μmx_{mon} is the average daily maximum temperature for the month (°C), $days_{tot}$ are the total number of days in the month, μWmx_{mon} is the average daily maximum temperature of the month on wet days (°C), $days_{wet}$ are the number of wet days in the month, μDmx_{mon} is the average daily maximum temperature of the month on dry days (°C), and $days_{dry}$ are the number of dry days in the month.

The wet day average maximum temperature is assumed to be less than the dry day average maximum temperature by some fraction of (μmx_{mon} - μmn_{mon}):

$$\mu Wmx_{mon} = \mu Dmx_{mon} - b_T \cdot \left(\mu mx_{mon} - \mu mn_{mon}\right) \qquad 4.2.15$$

where μWmx_{mon} is the average daily maximum temperature of the month on wet days (°C), μDmx_{mon} is the average daily maximum temperature of the month on dry days (°C), b_T is a scaling factor that controls the degree of deviation in temperature caused by the presence or absence of precipitation, μmx_{mon} is the average daily maximum temperature for the month (°C), and μmn_{mon} is the average daily minimum temperature for the month (°C). The scaling factor, b_T, is set to 0.5 in SWAT.

To calculate the dry day average maximum temperature, equations 4.2.14 and 4.2.15 are combined and solved for μDmx_{mon}:

$$\mu Dmx_{mon} = \mu mx_{mon} + b_T \cdot \frac{days_{wet}}{days_{tot}} \cdot \left(\mu mx_{mon} - \mu mn_{mon}\right) \qquad 4.2.16$$

Incorporating the modified values into equation 4.2.10, SWAT calculates the maximum temperature for a wet day using the equation:

$$T_{mx} = \mu Wmx_{mon} + \chi_i(1) \cdot \sigma mx_{mon} \qquad 4.2.17$$

and the maximum temperature for a dry day using the equation:

$$T_{mx} = \mu Dmx_{mon} + \chi_i(1) \cdot \sigma mx_{mon} \qquad 4.2.18$$

4.2.3.2 SOLAR RADIATION

The continuity equation relates average daily solar radiation adjusted for wet or dry conditions to the average daily solar radiation for the month:

$$\mu rad_{mon} \cdot days_{tot} = \mu Wrad_{mon} \cdot days_{wet} + \mu Drad_{mon} \cdot days_{dry} \qquad 4.2.19$$

where μrad_{mon} is the average daily solar radiation for the month (MJ m^{-2}), $days_{tot}$ are the total number of days in the month, $\mu Wrad_{mon}$ is the average daily solar radiation of the month on wet days (MJ m^{-2}), $days_{wet}$ are the number of wet days in the month, $\mu Drad_{mon}$ is the average daily solar radiation of the month on dry days (MJ m^{-2}), and $days_{dry}$ are the number of dry days in the month.

The wet day average solar radiation is assumed to be less than the dry day average solar radiation by some fraction:

$$\mu Wrad_{mon} = b_R \cdot \mu Drad_{mon} \qquad 4.2.20$$

where $\mu Wrad_{mon}$ is the average daily solar radiation of the month on wet days (MJ m^{-2}), $\mu Drad_{mon}$ is the average daily solar radiation of the month on dry days (MJ m^{-2}), and b_R is a scaling factor that controls the degree of deviation in solar radiation caused by the presence or absence of precipitation. The scaling factor, b_R, is set to 0.5 in SWAT.

To calculate the dry day average solar radiation, equations 4.2.19 and 4.2.20 are combined and solved for $\mu Drad_{mon}$:

$$\mu Drad_{mon} = \frac{\mu rad_{mon} \cdot days_{tot}}{b_R \cdot days_{wet} + days_{dry}} \qquad 4.2.21$$

Incorporating the modified values into equation 4.2.12, SWAT calculated the solar radiation on a wet day using the equation:

$$H_{day} = \mu Wrad_{mon} + \chi_i(3) \cdot \sigma rad_{mon} \qquad 4.2.22$$

and the solar radiation on a dry day using the equation:

$$H_{day} = \mu Drad_{mon} + \chi_i(3) \cdot \sigma rad_{mon} \qquad 4.2.23$$

Table 4-2: SWAT input variables that pertain to generation of temperature and solar radiation.

Variable name	Definition	File Name
TMPSIM	Temperature input code: 1-measured, 2-generated	.cod
SLRSIM	Solar radiation input code: 1-measured, 2-generated	.cod
TMPMX(mon)	μmx_{mon}: average maximum air temperature for month (°C)	.wgn
TMPSTDMX(mon)	σmx_{mon}: standard deviation for maximum air temperature in month (°C)	.wgn
TMPMN(mon)	μmn_{mon}: average minimum air temperature for month (°C)	.wgn
TMPSTDMN(mon)	σmn_{mon}: standard deviation for minimum air temperature in month (°C)	.wgn
SOLARAV(mon)	μrad_{mon}: average daily solar radiation for month (MJ m^{-2})	.wgn
PCPD(mon)	$days_{wet}$: average number of days of precipitation in month	.wgn

4.3 RELATIVE HUMIDITY

Relative humidity is required by SWAT when the Penman-Monteith equation is used to calculate potential evapotranspiration. Daily average relative humidity values are calculated from a triangular distribution using average monthly relative humidity.

4.3.1 MEAN MONTHLY RELATIVE HUMIDITY

Relative humidity is defined as the ratio of the actual vapor pressure to the saturation vapor pressure at a given temperature:

$$R_{hmon} = \frac{e_{mon}}{e^o_{mon}} \qquad 4.3.1$$

where R_{hmon} is the average relative humidity for the month, e_{mon} is the actual vapor pressure at the mean monthly temperature (kPa), and e^o_{mon} is the saturation vapor pressure at the mean monthly temperature (kPa). The saturation vapor pressure, e^o_{mon}, is related to the mean monthly air temperature with the equation:

$$e^o_{mon} = \exp\left[\frac{16.78 \cdot \mu tmp_{mon} - 116.9}{\mu tmp_{mon} + 237.3}\right] \qquad 4.3.2$$

where e^o_{mon} is the saturation vapor pressure at the mean monthly temperature (kPa), and μtmp_{mon} is the mean air temperature for the month (°C). The mean air temperature for the month is calculated by averaging the mean maximum monthly temperature, μmx_{mon}, and the mean minimum monthly temperature, μmn_{mon}.

The dew point temperature is the temperature at which the actual vapor pressure present in the atmosphere is equal to the saturation vapor pressure. Therefore, by substituting the dew point temperature in place of the average monthly temperature in equation 4.3.2, the actual vapor pressure may be calculated:

$$e_{mon} = \exp\left[\frac{16.78 \cdot \mu dew_{mon} - 116.9}{\mu dew_{mon} + 237.3}\right] \qquad 4.3.3$$

where e_{mon} is the actual vapor pressure at the mean month temperature (kPa), and μdew_{mon} is the average dew point temperature for the month (°C).

4.3.2 GENERATED DAILY VALUE

The triangular distribution used to generate daily relative humidity values requires four inputs: mean monthly relative humidity, maximum relative humidity value allowed in month, minimum relative humidity value allowed in month, and a random number between 0.0 and 1.0.

The maximum relative humidity value, or upper limit of the triangular distribution, is calculated from the mean monthly relative humidity with the equation:

$$R_{hUmon} = R_{hmon} + \left(1 - R_{hmon}\right) \cdot \exp\left(R_{hmon} - 1\right) \qquad 4.3.4$$

where R_{hUmon} is the largest relative humidity value that can be generated on a given day in the month, and R_{hmon} is the average relative humidity for the month.

The minimum relative humidity value, or lower limit of the triangular distribution, is calculated from the mean monthly relative humidity with the equation:

$$R_{hLmon} = R_{hmon} \cdot \left(1 - \exp\left(-R_{hmon}\right)\right) \tag{4.3.5}$$

where R_{hLmon} is the smallest relative humidity value that can be generated on a given day in the month, and R_{hmon} is the average relative humidity for the month.

The triangular distribution uses one of two sets of equations to generate a relative humidity value for the day. If $rnd_1 \leq \left(\dfrac{R_{hmon} - R_{hLmon}}{R_{hUmon} - R_{hLmon}}\right)$ then

$$R_h = R_{hLmon} + \left[rnd_1 \cdot \left(R_{hUmon} - R_{hLmon}\right) \cdot \left(R_{hmon} - R_{hLmon}\right)\right]^{0.5} \tag{4.3.6}$$

If $rnd_1 > \left(\dfrac{R_{hmon} - R_{hLmon}}{R_{hUmon} - R_{hLmon}}\right)$ then

$$R_h = R_{hUmon} - \left(R_{hUmon} - R_{hmon}\right) \cdot \left[\frac{R_{hUmon}\left(1 - rnd_1\right) - R_{hLmon}\left(1 - rnd_1\right)}{R_{hUmon} - R_{hmon}}\right]^{0.5} \tag{4.3.7}$$

where R_h is the average relative humidity calculated for the day, rnd_1 is a random number generated by the model each day, R_{hmon} is the average relative humidity for the month, R_{hLmon} is the smallest relative humidity value that can be generated on a given day in the month, and R_{hUmon} is the largest relative humidity value that can be generated on a given day in the month.

4.3.3 ADJUSTMENT FOR CLEAR/OVERCAST CONDITIONS

To incorporate the effect of clear and overcast weather on generated values of relative humidity, monthly average relative humidity values can be adjusted for wet or dry conditions.

The continuity equation relates average relative humidity adjusted for wet or dry conditions to the average relative humidity for the month:

$$R_{hmon} \cdot days_{tot} = R_{hWmon} \cdot days_{wet} + R_{hDmon} \cdot days_{dry} \tag{4.3.8}$$

where R_{hmon} is the average relative humidity for the month, $days_{tot}$ are the total number of days in the month, R_{hWmon} is the average relative humidity for the month on wet days, $days_{wet}$ are the number of wet days in the month, R_{hDmon} is the average relative humidity of the month on dry days, and $days_{dry}$ are the number of dry days in the month.

The wet day average relative humidity is assumed to be greater than the dry day average relative humidity by some fraction:

$$R_{hWmon} = R_{hDmon} + b_H \cdot (1 - R_{hDmon}) \qquad 4.3.9$$

where R_{hWmon} is the average relative humidity of the month on wet days, R_{hDmon} is the average relative humidity of the month on dry days, and b_H is a scaling factor that controls the degree of deviation in relative humidity caused by the presence or absence of precipitation. The scaling factor, b_H, is set to 0.9 in SWAT.

To calculate the dry day relative humidity, equations 4.3.8 and 4.3.9 are combined and solved for R_{hDmon}:

$$R_{hDmon} = \left(R_{hmon} - b_H \cdot \frac{days_{wet}}{days_{tot}} \right) \cdot \left(1.0 - b_H \cdot \frac{days_{wet}}{days_{tot}} \right)^{-1} \qquad 4.3.10$$

To reflect the impact of wet or dry conditions, SWAT will replace R_{hmon} with R_{hWmon} on wet days or R_{hDmon} on dry days in equations 4.3.4 through 4.3.7.

Table 4-3: SWAT input variables that pertain to generation of relative humidity.

Variable name	Definition	File Name
RHSIM	Relative humidity input code: 1-measured, 2-generated	.cod
TMPMN(mon)	μmn_{mon}: average minimum air temperature for month (°C)	.wgn
TMPMX(mon)	μmx_{mon}: average maximum air temperature for month (°C)	.wgn
DEWPT(mon)	μdew_{mon}: average dew point temperature for month (°C)	.wgn
PCPD(mon)	$days_{wet}$: average number of days of precipitation in month	.wgn

4.4 MAXIMUM HALF-HOUR RAINFALL

Maximum half-hour rainfall is required by SWAT to calculate the peak flow rate for runoff. When daily precipitation data is used by the model, the maximum half-hour rainfall is calculated from a triangular distribution using monthly maximum half-hour rainfall data. The maximum half-hour rainfall is calculated only on days where surface runoff has been generated.

4.4.1 MONTHLY MAXIMUM HALF-HOUR RAIN

For each month, users provide the maximum half-hour rain observed over the entire period of record. These extreme values are used to calculate representative monthly maximum half-hour rainfall fractions.

Prior to calculating the representative maximum half-hour rainfall fraction for each month, the extreme half-hour rainfall values are smoothed by calculating three month average values:

$$R_{0.5sm(mon)} = \frac{R_{0.5x(mon-1)} + R_{0.5x(mon)} + R_{0.5x(mon+1)}}{3} \qquad 4.4.1$$

where $R_{0.5sm(mon)}$ is the smoothed maximum half-hour rainfall for a given month (mm H_2O) and $R_{0.5x}$ is the extreme maximum half-hour rainfall for the specified month (mm H_2O). Once the smoothed maximum half-hour rainfall is known, the representative half-hour rainfall fraction is calculated using the equation:

$$\alpha_{0.5mon} = adj_{0.5\alpha} \cdot \left[1 - \exp\left(\frac{R_{0.5sm(mon)}}{\mu_{mon} \cdot \ln\left(\frac{0.5}{yrs \cdot days_{wet}} \right)} \right) \right] \qquad 4.4.2$$

where $\alpha_{0.5mon}$ is the average half-hour rainfall fraction for the month, $adj_{0.5\alpha}$ is an adjustment factor, $R_{0.5sm}$ is the smoothed half-hour rainfall amount for the month (mm H_2O), μ_{mon} is the mean daily rainfall (mm H_2O) for the month, yrs is the number of years of rainfall data used to obtain values for monthly extreme half-hour rainfalls, and $days_{wet}$ are the number of wet days in the month. The adjustment factor is included to allow users to modify estimations of half-hour rainfall fractions and peak flow rates for runoff.

4.4.2 GENERATED DAILY VALUE

The triangular distribution used to generate the maximum half-hour rainfall fraction requires four inputs: average monthly half-hour rainfall fraction, maximum value for half-hour rainfall fraction allowed in month, minimum value for half-hour rainfall fraction allowed in month, and a random number between 0.0 and 1.0.

The maximum half-hour rainfall fraction, or upper limit of the triangular distribution, is calculated from the daily amount of rainfall with the equation:

$$\alpha_{0.5U} = 1 - \exp\left(\frac{-125}{R_{day} + 5}\right) \qquad 4.4.3$$

where $\alpha_{0.5U}$ is the largest half-hour fraction that can be generated on a given day, and R_{day} is the precipitation on a given day (mm H_2O).

The minimum half-hour fraction, or lower limit of the triangular distribution, $\alpha_{0.5L}$, is set at 0.02083.

The triangular distribution uses one of two sets of equations to generate a maximum half-hour rainfall fraction for the day. If $rnd_1 \leq \left(\frac{\alpha_{0.5mon} - \alpha_{0.5L}}{\alpha_{0.5U} - \alpha_{0.5L}}\right)$ then

$$\alpha_{0.5} = \alpha_{0.5L} + \left[rnd_1 \cdot (\alpha_{0.5U} - \alpha_{0.5L}) \cdot (\alpha_{0.5mon} - \alpha_{0.5L})\right]^{0.5} \qquad 4.4.4$$

If $rnd_1 > \left(\frac{\alpha_{0.5mon} - \alpha_{0.5L}}{\alpha_{0.5U} - \alpha_{0.5L}}\right)$ then

$$\alpha_{0.5} = \alpha_{0.5U} - (\alpha_{0.5U} - \alpha_{0.5mon}) \cdot \left[\frac{\alpha_{0.5U}(1 - rnd_1) - \alpha_{0.5L}(1 - rnd_1)}{\alpha_{0.5U} - \alpha_{0.5mon}}\right]^{0.5} \qquad 4.4.5$$

where $\alpha_{0.5}$ is the maximum half-hour rainfall fraction for the day, $\alpha_{0.5mon}$ is the average maximum half-hour rainfall fraction for the month, rnd_1 is a random number generated by the model each day, $\alpha_{0.5L}$ is the smallest half-hour rainfall fraction that can be generated, and $\alpha_{0.5U}$ is the largest half-hour fraction that can be generated.

Table 4-4: SWAT input variables that pertain to generation of maximum half-hour rainfall.

Variable name	**Definition**	**File Name**
RAINHHMX(mon)	$R_{0.5x}$: extreme half-hour rainfall for month (mm H_2O)	.wgn
APM	$adj_{0.5\alpha}$: peak rate adjustment factor	.bsn
PCPMM(mon)	average amount of precipitation falling in month (mm H_2O)	.wgn
PCPD(mon)	$days_{wet}$: average number of days of precipitation in month (μ_{mon} = PCPMM / PCPD)	.wgn
RAIN_YRS	*yrs*: number of years of data used to obtain values for RAINHHMX	.wgn
PRECIPITATION	R_{day}: amount of rain falling on a given day (mm H_2O)	.pcp

4.5 WIND SPEED

Wind speed is required by SWAT when the Penman-Monteith equation is used to calculate potential evapotranspiration. Mean daily wind speed is generated in SWAT using a modified exponential equation:

$$\mu_{10m} = \mu wnd_{mon} \cdot \left(-\ln\left(rnd_1\right)\right)^{0.3} \qquad 4.5.1$$

where μ_{10m} is the mean wind speed for the day (m s^{-1}), μwnd_{mon} is the average wind speed for the month (m s^{-1}), and rnd_1 is a random number between 0.0 and 1.0.

Table 4-5: SWAT input variables that pertain to generation of wind speed.

Variable name	Definition	File Name
WNDSIM	Wind speed input code: 1-measured, 2-generated	.cod
WNDAV(mon)	μwnd_{mon}: Average wind speed in month (m/s)	.wgn

4.6 NOMENCLATURE

A 3×3 matrix of elements defined to ensure serial and cross correlation of generated temperature and radiation values $A = M_1 \cdot M_0^{-1}$

B 3×3 matrix of elements defined to ensure serial and cross correlation of generated temperature and radiation values $B \cdot B^T = M_0 - M_1 \cdot M_0^{-1} \cdot M_1^T$

H_{day} Solar radiation reaching ground on current day of simulation (MJ m^{-2} d^{-1})

H_{MX} Maximum possible solar radiation (MJ m^{-2} d^{-1})

M_0 3×3 matrix of correlation coefficients between maximum temperature, minimum temperature and solar radiation on same day

M_1 3×3 matrix of correlation coefficients between maximum temperature, minimum temperature and solar radiation on consecutive days

$P_i(D/D)$ Probability of a dry day on day i given a dry day on day $i - 1$

$P_i(D/W)$ Probability of a dry day on day i given a wet day on day $i - 1$

$P_i(W/D)$ Probability of a wet day on day i given a dry day on day $i - 1$

$P_i(W/W)$ Probability of a wet day on day i given a wet day on day $i - 1$

$R_{0.5sm}$ Smoothed maximum half-hour rainfall for a given month (mm H_2O)

$R_{0.5x}$ Extreme maximum half-hour rainfall for the specified month (mm H_2O)

R_{day} Amount of rainfall on a given day (mm H_2O)

R_h Average relative humidity for the day

R_{hDmon} Average relative humidity of the month on dry days

R_{hLmon} Smallest relative humidity value that can be generated on a given day in the month
R_{hUmon} Largest relative humidity value that can be generated on a given day in the month
R_{hWmon} Average relative humidity for the month on wet days
R_{hmon} Average relative humidity for the month
SND_{day} Standard normal deviate for the day
T_{mn} Minimum air temperature for day (°C)
T_{mx} Maximum air temperature for day (°C)

$adj_{0.5\alpha}$ Peak rate adjustment factor
b_H Scaling factor that controls the degree of deviation in relative humidity caused by the presence or absence of precipitation
b_R Scaling factor that controls the degree of deviation in solar radiation caused by the presence or absence of precipitation
b_T Scaling factor that controls the degree of deviation in temperature caused by the presence or absence of precipitation
$days_{dry}$ Number of dry days in the month
$days_{tot}$ Total number of days in the month
$days_{wet}$ Number of wet days in the month
e_{mon} Actual vapor pressure at the mean monthly temperature (kPa)
e^o_{mon} Saturation vapor pressure at the mean monthly temperature (kPa)
g_{mon} Skew coefficient for daily precipitation in the month
$rexp$ Exponent for exponential precipitation distribution
rnd_1 Random number between 0.0 and 1.0
rnd_2 Random number between 0.0 and 1.0
yrs Number of years of rainfall data used to obtain values for monthly extreme half-hour rainfalls

$\alpha_{0.5}$ Maximum half-hour rainfall expressed as a fraction of daily rainfall
$\alpha_{0.5L}$ Smallest half-hour rainfall fraction that can be generated on a given day
$\alpha_{0.5mon}$ Average maximum half-hour rainfall fraction for the month
$\alpha_{0.5U}$ Largest half-hour rainfall fraction that can be generated on a given day
ε_i 3×1 matrix of independent random components
σ_{mon} Standard deviation of daily rainfall (mm H_2O) for the month
σmn_{mon} Standard deviation for daily minimum temperature during the month (°C)
σmx_{mon} Standard deviation for daily maximum temperature during the month (°C)
σrad_{mon} Standard deviation for daily solar radiation during the month (MJ m^{-2})
$\rho_0(j,k)$ Correlation coefficient between variables j and k on the same day where j and k may be set to 1 (maximum temperature), 2 (minimum temperature) or 3 (solar radiation)
$\rho_1(j,k)$ Correlation coefficient between variable j and k with variable k lagged one day with respect to variable j
μ_{mon} Mean daily rainfall (mm H_2O) for the month
μDmx_{mon} Average daily maximum temperature of the month on dry days (°C)
$\mu Drad_{mon}$ Average daily solar radiation of the month on dry days (MJ m^{-2})

μWmx_{mon} Average daily maximum temperature of the month on wet days (°C)
$\mu Wrad_{mon}$ Average daily solar radiation of the month on wet days (MJ m^{-2})
μdew_{mon} Average dew point temperature for the month (°C)
μmn_{mon} Average daily minimum temperature for the month (°C)
μmx_{mon} Average daily maximum temperature for the month (°C)
μrad_{mon} Average daily solar radiation for the month (MJ m^{-2})
μtmp_{mon} Mean air temperature for the month (°C)
μwnd_{mon} Average wind speed for the month (m s^{-1})
μ_{10m} Mean wind speed for the day at height of 10 meters (m s^{-1})
$\chi_i(j)$ 3×1 matrix for day i whose elements are residuals of maximum temperature (j = 1), minimum temperature (j = 2) and solar radiation (j = 3),

4.7 REFERENCES

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CHAPTER 5

EQUATIONS: CLIMATE CUSTOMIZATION

SWAT is capable of simulating orographic impacts on temperature and precipitation for watersheds in mountainous regions. The model will also modify climate inputs for simulations that are looking at the impact of climatic change in a given watershed.

5.1 ELEVATION BANDS

Orographic precipitation is a significant phenomenon in certain areas of the world. To account for orographic effects on both precipitation and temperature, SWAT allows up to 10 elevation bands to be defined in each subbasin. Precipitation and maximum and minimum temperatures are calculated for each band as a function of the respective lapse rate and the difference between the gage elevation and the average elevation specified for the band. For precipitation,

$$R_{band} = R_{day} + \left(EL_{band} - EL_{gage}\right) \cdot \frac{plaps}{1000} \qquad \text{when } R_{day} > 0.01 \qquad 5.1.1$$

where R_{band} is the precipitation falling in the elevation band (mm H_2O), R_{day} is the precipitation recorded at the gage or generated from gage data (mm H_2O), EL_{band} is the mean elevation in the elevation band (m), EL_{gage} is the elevation at the recording gage (m), $plaps$ is the precipitation lapse rate (mm H_2O/km), and 1000 is a factor needed to convert meters to kilometers. For temperature,

$$T_{mx,band} = T_{mx} + \left(EL_{band} - EL_{gage}\right) \cdot \frac{tlaps}{1000} \qquad 5.1.2$$

$$T_{mn,band} = T_{mn} + \left(EL_{band} - EL_{gage}\right) \cdot \frac{tlaps}{1000} \qquad 5.1.3$$

$$\bar{T}_{av,band} = \bar{T}_{av} + \left(EL_{band} - EL_{gage}\right) \cdot \frac{tlaps}{1000} \qquad 5.1.4$$

where $T_{mx,band}$ is the maximum daily temperature in the elevation band (°C), $T_{mn,band}$ is the minimum daily temperature in the elevation band (°C), $\bar{T}_{av,band}$ is the mean daily temperature in the elevation band (°C), T_{mx} is the maximum daily temperature recorded at the gage or generated from gage data (°C), T_{mn} is the minimum daily temperature recorded at the gage or generated from gage data (°C), $\bar{T}_{av}$ is the mean daily temperature recorded at the gage or generated from gage data (°C), EL_{band} is the mean elevation in the elevation band (m), EL_{gage} is the elevation at the recording gage (m), $tlaps$ is the temperature lapse rate (°C/km), and 1000 is a factor needed to convert meters to kilometers.

Once the precipitation and temperature values have been calculated for each elevation band in the subbasin, new average subbasin precipitation and temperature values are calculated:

$$R_{day} = \sum_{bnd=1}^{b} R_{band} \cdot fr_{bnd} \qquad 5.1.5$$

$$T_{mx} = \sum_{bnd=1}^{b} T_{mx,band} \cdot fr_{bnd} \qquad 5.1.6$$

$$T_{mn} = \sum_{bnd=1}^{b} T_{mn,band} \cdot fr_{bnd} \qquad 5.1.7$$

$$\overline{T}_{av} = \sum_{bnd=1}^{b} \overline{T}_{av,band} \cdot fr_{bnd} \qquad 5.1.8$$

where R_{day} is the daily average precipitation adjusted for orographic effects (mm H_2O), T_{mx} is the daily maximum temperature adjusted for orographic effects (°C), T_{mn} is the daily minimum temperature adjusted for orographic effects (°C), $\overline{T}_{av}$ is the daily mean temperature adjusted for orographic effects (°C), R_{band} is the precipitation falling in elevation band *bnd* (mm H_2O), $T_{mx,band}$ is the maximum daily temperature in elevation band *bnd* (°C), $T_{mn,band}$ is the minimum daily temperature in elevation band *bnd* (°C), $\overline{T}_{av,band}$ is the mean daily temperature in elevation band *bnd* (°C), fr_{bnd} is the fraction of subbasin area within the elevation band, and b is the total number of elevation bands in the subbasin.

The only processes modeled separately for each individual elevation band are the accumulation, sublimation and melting of snow. As with the initial precipitation and temperature data, after amounts of sublimation and snow melt are determined for each elevation band, subbasin average values are calculated. These average values are the values that are used in the remainder of the simulation and reported in the output files.

Table 5-1: SWAT input variables that pertain to orographic effects.

Variable Name	Definition	Input File
ELEVB	EL_{band}: Elevation at center of the elevation band (m)	.sub
ELEVB_FR	fr_{bnd}: Fraction of subbasin area within the elevation band.	.sub
WELEV	EL_{gage}: Elevation of recording gage whose data is used to calculate values in .wgn file (m)	.wgn
ELEVATION	EL_{gage}: Elevation of precipitation recording gage (m)	.pcp
ELEVATION	EL_{gage}: Elevation of temperature recording gage (m)	.tmp
PLAPS	*plaps*: Precipitation lapse rate (mm H_2O/km)	.sub
TLAPS	*tlaps*: Temperature lapse rate (°C/km)	.sub
PRECIPITATION	R_{day}: Daily precipitation (mm H_2O)	.pcp
MAX TEMP	T_{mx}: Daily maximum temperature (°C)	.tmp
MIN TEMP	T_{mn}: Daily minimum temperature (°C)	.tmp

5.2 CLIMATE CHANGE

The impact of global climate change on water supply is a major area of research. Climate change can be simulated with SWAT by manipulating the climatic input that is read into the model (precipitation, temperature, solar radiation, relative humidity, wind speed, potential evapotranspiration and weather generator parameters). A less time-consuming method is to set adjustment factors for the various climatic inputs.

SWAT will allow users to adjust precipitation, temperature, solar radiation, relative humidity, and carbon dioxide levels in each subbasin. The alteration of precipitation, temperature, solar radiation and relative humidity are straightforward:

$$R_{day} = R_{day} \cdot \left(1 + \frac{adj_{pcp}}{100}\right) \qquad 5.2.1$$

where R_{day} is the precipitation falling in the subbasin on a given day (mm H_2O), and adj_{pcp} is the % change in rainfall.

$$T_{mx} = T_{mx} + adj_{tmp} \qquad 5.2.2$$

where T_{mx} is the daily maximum temperature (°C), and adj_{tmp} is the change in temperature (°C).

$$T_{mn} = T_{mn} + adj_{tmp} \qquad 5.2.3$$

where T_{mn} is the daily minimum temperature (°C), and adj_{tmp} is the change in temperature (°C).

$$\overline{T}_{av} = \overline{T}_{av} + adj_{tmp} \qquad 5.2.4$$

where $\overline{T}_{av}$ is the daily mean temperature (°C), and adj_{tmp} is the change in temperature (°C).

$$H_{day} = H_{day} + adj_{rad} \qquad 5.2.5$$

where H_{day} is the daily solar radiation reaching the earth's surface (MJ m^{-2}), and adj_{rad} is the change in radiation (MJ m^{-2} d^{-1}).

$$R_h = R_h + adj_{hmd} \qquad 5.2.6$$

where R_h is the relative humidity for the day expressed as a fraction, and adj_{hmd} is the change in relative humidity expressed as a fraction.

SWAT allows the adjustment terms to vary from month to month so that the user is able to simulate seasonal changes in climatic conditions.

Changes in carbon dioxide levels impact plant growth. As carbon dioxide levels increase, plant productivity increases and plant water requirements go down. The equations used to account for the impact of carbon dioxide levels on plant water requirements are reviewed in Chapters 7 and 18. When carbon dioxide climate change effects are being simulated, the Penman-Monteith equation must be used to calculate potential evapotranspiration. This method has been modified to account for CO_2 impacts on potential evapotranspiration levels.

Table 5-2: SWAT input variables that pertain to climate change.

Variable Name	Definition	Input File
RFINC(mon)	adj_{pcp}: % change in rainfall for month	.sub
TMPINC(mon)	adj_{tmp}: increase or decrease in temperature for month (°C)	.sub
RADINC(mon)	adj_{rad}: increase or decrease in solar radiation reaching earth's surface for month (MJ m^{-2})	.sub
HUMINC(mon)	adj_{hmd}: increase or decrease in relative humidity for month	.sub
CO2	CO_2: carbon dioxide level in subbasin (ppmv)	.sub
IPET	Potential evapotranspiration method	.cod

5.3 NOMENCLATURE

CO_2 Concentration of carbon dioxide in the atmosphere (ppmv)
EL_{band} Mean elevation in the elevation band (m)
EL_{gage} Elevation at the precipitation, temperature, or weather generator data recording gage (m)
H_{day} Solar radiation reaching ground on current day of simulation (MJ m^{-2} d^{-1})
R_{band} Precipitation falling in the elevation band (mm H_2O)
R_{day} Amount of rainfall on a given day (mm H_2O)
R_h Average relative humidity for the day
T_{mn} Minimum air temperature for day (°C)
$T_{mn,band}$ Minimum daily temperature in the elevation band (°C)
T_{mx} Maximum air temperature for day (°C)
$T_{mx,band}$ Maximum daily temperature in the elevation band (°C)
$\bar{T}_{av}$ Mean air temperature for day (°C)
$\bar{T}_{av,band}$ Mean daily temperature in the elevation band (°C)

adj_{hmd} Change in relative humidity expressed as a fraction
adj_{pcp} % change in rainfall
adj_{rad} Change in radiation (MJ m^{-2} d^{-1})
adj_{tmp} Change in temperature (°C)
fr_{bnd} Fraction of subbasin area within the elevation band
$plaps$ Precipitation lapse rate (mm H_2O/km)
$tlaps$ Temperature lapse rate (°C/km)

HYDROLOGY

The land phase of the hydrologic cycle is based on the water balance equation:

$$SW_t = SW_0 + \sum_{i=1}^{t}(R_{day} - Q_{surf} - E_a - w_{seep} - Q_{gw})$$

where SW_t is the final soil water content (mm H_2O), SW_0 is the initial soil water content (mm H_2O), t is the time (days), R_{day} is the amount of precipitation on day i (mm H_2O), Q_{surf} is the amount of surface runoff on day i (mm H_2O), E_a is the amount of evapotranspiration on day i (mm H_2O), w_{seep} is the amount of percolation and bypass flow exiting the soil profile bottom on day i (mm H_2O), and Q_{gw} is the amount of return flow on day i (mm H_2O).

CHAPTER 6

EQUATIONS: SURFACE RUNOFF

Surface runoff occurs whenever the rate of water application to the ground surface exceeds the rate of infiltration. When water is initially applied to a dry soil, the application rate and infiltration rates may be similar. However, the infiltration rate will decrease as the soil becomes wetter. When the application rate is higher than the infiltration rate, surface depressions begin to fill. If the application rate continues to be higher than the infiltration rate once all surface depressions have filled, surface runoff will commence.

SWAT provides two methods for estimating surface runoff: the SCS curve number procedure (SCS, 1972) and the Green & Ampt infiltration method (1911).

6.1 RUNOFF VOLUME: SCS CURVE NUMBER PROCEDURE

The SCS runoff equation is an empirical model that came into common use in the 1950s. It was the product of more than 20 years of studies involving rainfall-runoff relationships from small rural watersheds across the U.S. The model was developed to provide a consistent basis for estimating the amounts of runoff under varying land use and soil types (Rallison and Miller, 1981).

The SCS curve number equation is (SCS, 1972):

$$Q_{surf} = \frac{(R_{day} - I_a)^2}{(R_{day} - I_a + S)} \qquad 6.1.1$$

where Q_{surf} is the accumulated runoff or rainfall excess (mm H_2O), R_{day} is the rainfall depth for the day (mm H_2O), I_a is the initial abstractions which includes surface storage, interception and infiltration prior to runoff (mm H_2O), and S is the retention parameter (mm H_2O). The retention parameter varies spatially due to changes in soils, land use, management and slope and temporally due to changes in soil water content. The retention parameter is defined as:

$$S = 25.4\left(\frac{1000}{CN} - 10\right) \qquad 6.1.2$$

where CN is the curve number for the day. The initial abstractions, I_a, is commonly approximated as $0.2S$ and equation 6.1.1 becomes

$$Q_{surf} = \frac{(R_{day} - 0.2S)^2}{(R_{day} + 0.8S)} \qquad 6.1.3$$

Runoff will only occur when $R_{day} > I_a$. A graphical solution of equation 6.1.3 for different curve number values is presented in Figure 6-1.

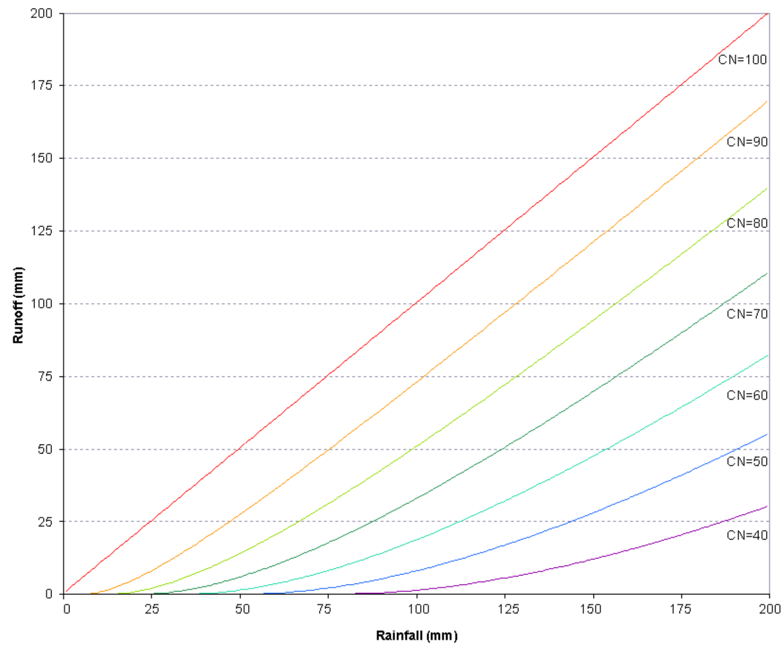

Figure 6-1: Relationship of runoff to rainfall in SCS curve number method.

6.1.1 SCS CURVE NUMBER

The SCS curve number is a function of the soil's permeability, land use and antecedent soil water conditions. Typical curve numbers for moisture condition II are listed in tables 6-1, 6-2 and 6-3 for various land covers and soil types (SCS Engineering Division, 1986). These values are appropriate for a 5% slope.

Table 6-1: Runoff curve numbers for cultivated agricultural lands

Cover			Hydrologic Soil Group			
Land Use	**Treatment or practice**	**Hydrologic condition**	**A**	**B**	**C**	**D**
Fallow	Bare soil	- - - -	77	86	91	94
	Crop residue cover*	Poor	76	85	90	93
		Good	74	83	88	90
Row crops	Straight row	Poor	72	81	88	91
		Good	67	78	85	89
	Straight row w/ residue	Poor	71	80	87	90
		Good	64	75	82	85
	Contoured	Poor	70	79	84	88
		Good	65	75	82	86
	Contoured w/ residue	Poor	69	78	83	87
		Good	64	74	81	85
	Contoured & terraced	Poor	66	74	80	82
		Good	62	71	78	81
	Contoured & terraced w/ residue	Poor	65	73	79	81
		Good	61	70	77	80
Small grains	Straight row	Poor	65	76	84	88
		Good	63	75	83	87
	Straight row w/ residue	Poor	64	75	83	86

* Crop residue cover applies only if residue is on at least 5% of the surface throughout the year.

Table 6-1, cont.: Runoff curve numbers for cultivated agricultural lands

Cover			Hydrologic Soil Group			
Land Use	**Treatment or practice**	**Hydrologic condition**	**A**	**B**	**C**	**D**
		Good	60	72	80	84
	Contoured	Poor	63	74	82	85
		Good	61	73	81	84
	Contoured w/ residue	Poor	62	73	81	84
		Good	60	72	80	83
	Contoured & terraced	Poor	61	72	79	82
		Good	59	70	78	81
	Contoured & terraced w/ residue	Poor	60	71	78	81
		Good	58	69	77	80
Close-seeded or broadcast legumes or rotation	Straight row	Poor	66	77	85	89
		Good	58	72	81	85
	Contoured	Poor	64	75	83	85
		Good	55	69	78	83
	Contoured & terraced	Poor	63	73	80	83
		Good	51	67	76	80

Table 6-2: Runoff curve numbers for other agricultural lands

Cover		Hydrologic Soil Group			
Cover Type	**Hydrologic condition**	**A**	**B**	**C**	**D**
Pasture, grassland, or range—continuous forage for grazing[1]	Poor	68	79	86	89
	Fair	49	69	79	84
	Good	39	61	74	80
Meadow—continuous grass, protected from grazing and generally mowed for hay	- - - -	30	58	71	78
Brush—brush-weed-grass mixture with brush the major element[2]	Poor	48	67	77	83
	Fair	35	56	70	77
	Good	30	48	65	73
Woods—grass combination (orchard or tree farm)	Poor	57	73	82	86
	Fair	43	65	76	82
	Good	32	58	72	79
Woods[3]	Poor	45	66	77	83
	Fair	36	60	73	79
	Good	30	55	70	77
Farmsteads—buildings, lanes, driveways, and surrounding lots.	- - - -	59	74	82	86

Table 6-3: Runoff curve numbers for urban areas[§]

Cover			Hydrologic Soil Group			
Cover Type	**Hydrologic condition**	**Average % impervious area**	**A**	**B**	**C**	**D**
Fully developed urban areas						
Open spaces (lawns, parks, golf courses, cemeteries, etc.)[†]	Poor		68	79	86	89
	Fair		49	69	79	84
	Good		39	61	74	80
Impervious areas:						
Paved parking lots, roofs, driveways, etc. (excl. right-of-way)	- - - -		98	98	98	98
Paved streets and roads; open ditches (incl. right-of-way)	- - - -		83	89	92	93
Gravel streets and roads (including right-of-way)	- - - -		76	85	89	91
Dirt streets and roads (including right-of way)	- - - -		72	82	87	89

[1] *Poor:* < 50% ground cover or heavily grazed with no mulch; *Fair:* 50 to 75% ground cover and not heavily grazed; *Good*: > 75% ground cover and lightly or only occasionally grazed

[2] *Poor:* < 50% ground cover; *Fair:* 50 to 75% ground cover; Good: > 75% ground cover

[3] *Poor:* Forest litter, small trees, and brush are destroyed by heavy grazing or regular burning; *Fair:* Woods are grazed but not burned, and some forest litter covers the soil; *Good:* Woods are protected from grazing, and litter and brush adequately cover the soil.

[§] SWAT will automatically adjust curve numbers for impervious areas when IURBAN and URBLU are defined in the .hru file. Curve numbers from Table 6-3 should ***not*** be used in this instance.

[†] *Poor:* grass cover < 50%; *Fair:* grass cover 50 to 75%; *Good:* grass cover > 75%

Table 6-3, continued: Runoff curve number for urban areas

Cover			Hydrologic Soil Group			
Cover Type	**Hydrologic condition**	**Average % impervious area**	**A**	**B**	**C**	**D**
Urban districts:						
Commercial and business		85%	89	92	94	95
Industrial		72%	81	88	91	93
Residential Districts by average lot size:						
1/8 acre (0.05 ha) or less (town houses)		65%	77	85	90	92
1/4 acre (0.10 ha)		38%	61	75	83	87
1/3 acre (0.13 ha)		30%	57	72	81	86
1/2 acre (0.20 ha)		25%	54	70	80	85
1 acre (0.40 ha)		20%	51	68	79	84
2 acres (0.81 ha)		12%	46	65	77	82
Developing urban areas:						
Newly graded areas (pervious areas only, no vegetation)			77	86	91	94

6.1.1.1 SOIL HYDROLOGIC GROUPS

The U.S. Natural Resource Conservation Service (NRCS) classifies soils into four hydrologic groups based on infiltration characteristics of the soils. NRCS Soil Survey Staff (1996) defines a hydrologic group as a group of soils having similar runoff potential under similar storm and cover conditions. Soil properties that influence runoff potential are those that impact the minimum rate of infiltration for a bare soil after prolonged wetting and when not frozen. These properties are depth to seasonally high water table, saturated hydraulic conductivity, and depth to a very slowly permeable layer. Soil may be placed in one of four groups, A, B, C, and D, or three dual classes, A/D, B/D, and C/D. Definitions of the classes are:

A: (Low runoff potential). The soils have a high infiltration rate even when thoroughly wetted. They chiefly consist of deep, well drained to excessively drained sands or gravels. They have a high rate of water transmission.

B: The soils have a moderate infiltration rate when thoroughly wetted. They chiefly are moderately deep to deep, moderately well-drained to well-drained soils that have moderately fine to moderately coarse textures. They have a moderate rate of water transmission.

C: The soils have a slow infiltration rate when thoroughly wetted. They chiefly have a layer that impedes downward movement of water or have moderately fine to fine texture. They have a slow rate of water transmission.

D. (High runoff potential). The soils have a very slow infiltration rate when thoroughly wetted. They chiefly consist of clay soils that have a high swelling potential, soils that have a permanent water table, soils that have a claypan or clay layer at or near the surface, and shallow soils over nearly impervious material. They have a very slow rate of water transmission.

Dual hydrologic groups are given for certain wet soils that can be adequately drained. The first letter applies to the drained condition, the second to the undrained. Only soils that are rated D in their natural condition are assigned to dual classes. A summary of U.S. soils and their hydrologic group is given in Appendix D.

6.1.1.2 ANTECEDENT SOIL MOISTURE CONDITION

SCS defines three antecedent moisture conditions: I—dry (wilting point), II—average moisture, and III—wet (field capacity). The moisture condition I curve number is the lowest value the daily curve number can assume in dry conditions. The curve numbers for moisture conditions I and III are calculated with the equations:

$$CN_1 = CN_2 - \frac{20 \cdot (100 - CN_2)}{(100 - CN_2 + \exp[2.533 - 0.0636 \cdot (100 - CN_2)])} \qquad 6.1.4$$

$$CN_3 = CN_2 \cdot \exp[0.00673 \cdot (100 - CN_2)] \qquad 6.1.5$$

where CN_1 is the moisture condition I curve number, CN_2 is the moisture condition II curve number, and CN_3 is the moisture condition III curve number.

The retention parameter varies with soil profile water content according to the following equation:

$$S = S_{\max} \cdot \left(1 - \frac{SW}{[SW + \exp(w_1 - w_2 \cdot SW)]}\right) \qquad 6.1.6$$

where S is the retention parameter for a given moisture content (mm), S_{max} is the maximum value the retention parameter can achieve on any given day (mm), SW is the soil water content of the entire profile excluding the amount of water held in the profile at wilting point (mm H_2O), and w_1 and

w_2 are shape coefficients. The maximum retention parameter value, S_{max}, is calculated by solving equation 6.1.2 using CN_1.

The shape coefficients are determined by solving equation 6.1.6 assuming that

1) the retention parameter for moisture condition I curve number corresponds to wilting point soil profile water content,
2) the retention parameter for moisture condition III curve number corresponds to field capacity soil profile water content, and
3) the soil has a curve number of 99 (S = 2.54) when completely saturated.

$$w_1 = \ln\left[\frac{FC}{1 - S_3 \cdot S_{\max}{}^{-1}} - FC\right] + w_2 \cdot FC \qquad 6.1.7$$

$$w_2 = \frac{\left(\ln\left[\frac{FC}{1 - S_3 \cdot S_{\max}{}^{-1}} - FC\right] - \ln\left[\frac{SAT}{1 - 2.54 \cdot S_{\max}{}^{-1}} - SAT\right]\right)}{(SAT - FC)} \qquad 6.1.8$$

where w_1 is the first shape coefficient, w_2 is the second shape coefficient, FC is the amount of water in the soil profile at field capacity (mm H_2O), S_3 is the retention parameter for the moisture condition III curve number, S_{max} is the retention parameter for the moisture condition I curve number, SAT is the amount of water in the soil profile when completely saturated (mm H_2O), and 2.54 is the retention parameter value for a curve number of 99.

When the top layer of the soil is frozen, the retention parameter is modified using the following equation:

$$S_{frz} = S_{\max} \cdot [1 - \exp(-0.000862 \cdot S)] \qquad 6.1.9$$

where S_{frz} is the retention parameter adjusted for frozen conditions (mm), S_{max} is the maximum value the retention parameter can achieve on any given day (mm), and S is the retention parameter for a given moisture content calculated with equation 6.1.6 (mm).

The daily curve number value adjusted for moisture content is calculated by rearranging equation 6.1.2 and inserting the retention parameter calculated for that moisture content:

$$CN = \frac{25400}{(S + 254)} \qquad 6.1.10$$

where CN is the curve number on a given day and S is the retention parameter calculated for the moisture content of the soil on that day.

6.1.1.3 SLOPE ADJUSTMENTS

The moisture condition II curve numbers provided in the tables are assumed to be appropriate for 5% slopes. Williams (1995) developed an equation to adjust the curve number to a different slope:

$$CN_{2s} = \frac{(CN_3 - CN_2)}{3} \cdot [1 - 2 \cdot \exp(-13.86 \cdot slp)] + CN_2 \qquad 6.1.11$$

where CN_{2s} is the moisture condition II curve number adjusted for slope, CN_3 is the moisture condition III curve number for the default 5% slope, CN_2 is the moisture condition II curve number for the default 5% slope, and slp is the average percent slope of the subbasin. SWAT does not adjust curve numbers for slope. If the user wishes to adjust the curve numbers for slope effects, the adjustment must be done prior to entering the curve numbers in the management input file.

Table 6-1: SWAT input variables that pertain to surface runoff calculated with the SCS curve number method.

Variable Name	Definition	Input File
IEVENT	Rainfall, runoff, routing option.	.cod
PRECIPITATION	R_{day}: Daily precipitation (mm H_2O)	.pcp
CN2	CN_2: Moisture condition II curve number	.mgt
CNOP	CN_2: Moisture condition II curve number	.mgt

6.2 RUNOFF VOLUME: GREEN & AMPT INFILTRATION METHOD

The Green & Ampt equation was developed to predict infiltration assuming excess water at the surface at all times (Green & Ampt, 1911). The equation assumes that the soil profile is homogenous and antecedent moisture is uniformly distributed in the profile. As water infiltrates into the soil, the model assumes the soil above the wetting front is completely saturated and there is a sharp break in moisture content at the wetting front. Figure 6-2 graphically illustrates the difference between the moisture distribution with depth modeled by the Green & Ampt equation and what occurs in reality.

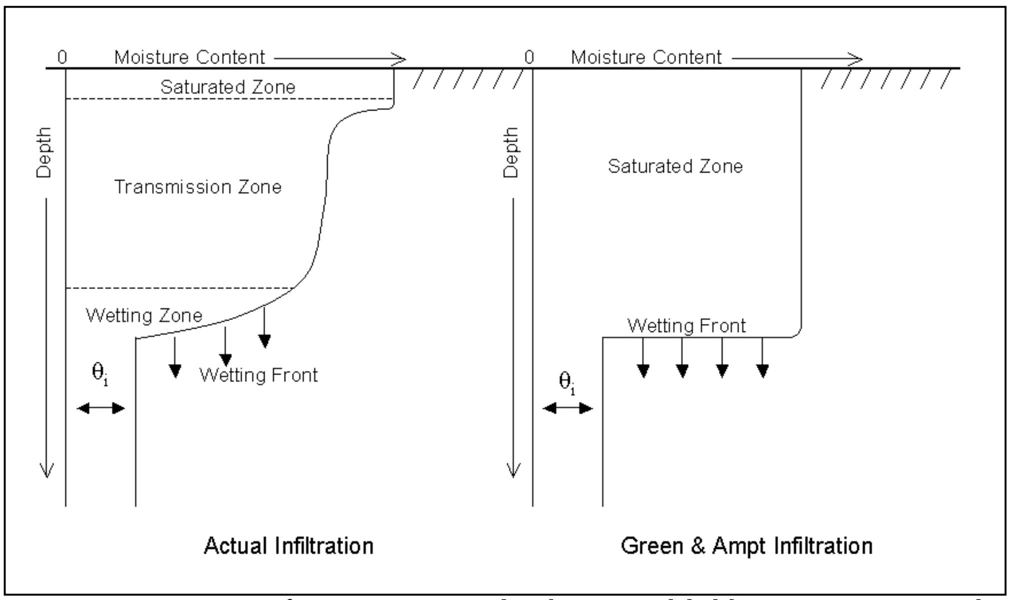

Figure 6-2: Comparison of moisture content distribution modeled by Green & Ampt and a typical observed distribution.

Mein and Larson (1973) developed a methodology for determining ponding time with infiltration using the Green & Ampt equation. The Green-Ampt Mein-Larson excess rainfall method was incorporated into SWAT to provide an alternative option for determining surface runoff. This method requires sub-daily precipitation data supplied by the user.

The Green-Ampt Mein-Larson infiltration rate is defined as:

$$f_{inf,t} = K_e \cdot \left(1 + \frac{\Psi_{wf} \cdot \Delta\theta_v}{F_{inf,t}}\right) \qquad 6.2.1$$

where f_{inf} is the infiltration rate at time t (mm/hr), K_e is the effective hydraulic conductivity (mm/hr), Ψ_{wf} is the wetting front matric potential (mm), $\Delta\theta_v$ is the change in volumetric moisture content across the wetting front (mm/mm) and F_{inf} is the cumulative infiltration at time t (mm H_2O).

When the rainfall intensity is less than the infiltration rate, all the rainfall will infiltrate during the time period and the cumulative infiltration for that time period is calculated:

$$F_{inf,t} = F_{inf,t-1} + R_{\Delta t} \qquad 6.2.2$$

where and $F_{inf,t}$ is the cumulative infiltration for a given time step (mm H_2O), $F_{inf,t-1}$ is the cumulative infiltration for the previous time step (mm H_2O), and $R_{\Delta t}$ is the amount of rain falling during the time step (mm H_2O).

The infiltration rate defined by equation 6.2.1 is a function of the infiltrated volume, which in turn is a function of the infiltration rates in previous time steps. To avoid numerical errors over long time steps, f_{inf} is replaced by dF_{inf}/dt in equation 6.2.1 and integrated to obtain

$$F_{inf,t} = F_{inf,t-1} + K_e \cdot \Delta t + \Psi_{wf} \cdot \Delta\theta_v \cdot \ln\left[\frac{F_{inf,t} + \Psi_{wf} \cdot \Delta\theta_v}{F_{inf,t-1} + \Psi_{wf} \cdot \Delta\theta_v}\right] \qquad 6.2.3$$

Equation 6.2.3 must be solved iteratively for $F_{inf,t}$, the cumulative infiltration at the end of the time step. A successive substitution technique is used.

The Green-Ampt effective hydraulic conductivity parameter, K_e, is approximately equivalent to one-half the saturated hydraulic conductivity of the soil, K_{sat} (Bouwer, 1969). Nearing et al. (1996) developed an equation to calculate the effective hydraulic conductivity as a function of saturated hydraulic conductivity and curve number. This equation incorporates land cover impacts into the calculated effective hydraulic conductivity. The equation for effective hydraulic conductivity is:

$$K_e = \frac{56.82 \cdot K_{sat}^{0.286}}{1 + 0.051 \cdot \exp(0.062 \cdot CN)} - 2 \qquad 6.2.4$$

where K_e is the effective hydraulic conductivity (mm/hr), K_{sat} is the saturated hydraulic conductivity (mm/hr), and CN is the curve number.

Wetting front matric potential, Ψ_{wf}, is calculated as a function of porosity, percent sand and percent clay (Rawls and Brakensiek, 1985):

$$\Psi_{wf} = 10 \cdot \exp\left[6.5309 - 7.32561 \cdot \phi_{soil} + 0.001583 \cdot m_c^{\ 2} + 3.809479 \cdot \phi_{soil}^{\ 2} + 0.000344 \cdot m_s \cdot m_c - 0.049837 \cdot m_s \cdot \phi_{soil} + 0.001608 \cdot m_s^{\ 2} \cdot \phi_{soil}^{\ 2} + 0.001602 \cdot m_c^{\ 2} \cdot \phi_{soil}^{\ 2} - 0.0000136 \cdot m_s^{\ 2} \cdot m_c - 0.003479 \cdot m_c^{\ 2} \cdot \phi_{soil} - 0.000799 \cdot m_s^{\ 2} \cdot \phi_{soil}\right] \quad 6.2.5$$

where ϕ_{soil} is the porosity of the soil (mm/mm), m_c is the percent clay content, and m_s is the percent sand content.

For each time step, SWAT calculates the amount of water entering the soil. The water that does not infiltrate into the soil becomes surface runoff.

Table 6-2: SWAT input variables that pertain to Green & Ampt infiltration calculations.

Variable Name	Definition	Input File
IEVENT	Rainfall, runoff, routing option.	.cod
IDT	Length of time step (min): Δt=IDT/60	.cod
PRECIPITATION	$R_{\Delta t}$: Precipitation during time step (mm H_2O)	.pcp
SOL_K	K_{sat}: Saturated hydraulic conductivity of first layer (mm/hr)	.sol
CN2	CN: Moisture condition II curve number	.mgt
CNOP	CN: Moisture condition II curve number	.mgt
SOL_BD	ρ_b: Moist bulk density (Mg/m^3): ϕ_{soil}=1 - ρ_b / 2.65	.sol
CLAY	m_c: % clay content	.sol
SAND	m_s: % sand content	.sol

6.3 PEAK RUNOFF RATE

The peak runoff rate is the maximum runoff flow rate that occurs with a given rainfall event. The peak runoff rate is an indicator of the erosive power of a storm and is used to predict sediment loss. SWAT calculates the peak runoff rate with a modified rational method.

The rational method is widely used in the design of ditches, channels and storm water control systems. The rational method is based on the assumption that if a rainfall of intensity i begins at time $t = 0$ and continues indefinitely, the rate of runoff will increase until the time of concentration, $t = t_{conc}$, when the entire subbasin area is contributing to flow at the outlet. The rational formula is:

$$q_{peak} = \frac{C \cdot i \cdot Area}{3.6} \qquad 6.3.1$$

where q_{peak} is the peak runoff rate (m^3 s^{-1}), C is the runoff coefficient, i is the rainfall intensity (mm/hr), $Area$ is the subbasin area (km^2) and 3.6 is a unit conversion factor.

6.3.1 TIME OF CONCENTRATION

The time of concentration is the amount of time from the beginning of a rainfall event until the entire subbasin area is contributing to flow at the outlet. In other words, the time of concentration is the time for a drop of water to flow from the remotest point in the subbasin to the subbasin outlet. The time of concentration is calculated by summing the overland flow time (the time it takes for flow from the remotest point in the subbasin to reach the channel) and the channel flow time (the time it takes for flow in the upstream channels to reach the outlet):

$$t_{conc} = t_{ov} + t_{ch} \qquad 6.3.2$$

where t_{conc} is the time of concentration for a subbasin (hr), t_{ov} is the time of concentration for overland flow (hr), and t_{ch} is the time of concentration for channel flow (hr).

6.3.1.1 OVERLAND FLOW TIME OF CONCENTRATION

The overland flow time of concentration, t_{ov}, can be computed using the equation

$$t_{ov} = \frac{L_{slp}}{3600 \cdot v_{ov}} \qquad 6.3.3$$

where L_{slp} is the subbasin slope length (m), v_{ov} is the overland flow velocity (m s^{-1}) and 3600 is a unit conversion factor.

The overland flow velocity can be estimated from Manning's equation by considering a strip 1 meter wide down the sloping surface:

$$v_{ov} = \frac{q_{ov}^{\,0.4} \cdot slp^{0.3}}{n^{0.6}} \qquad 6.3.4$$

where q_{ov} is the average overland flow rate (m^3 s^{-1}), slp is the average slope in the subbasin (m m^{-1}), and n is Manning's roughness coefficient for the subbasin. Assuming an average flow rate of 6.35 mm/hr and converting units

$$v_{ov} = \frac{0.005 \cdot L_{slp}^{\;0.4} \cdot slp^{0.3}}{n^{0.6}} \qquad 6.3.5$$

Substituting equation 6.3.5 into equation 6.3.3 gives

$$t_{ov} = \frac{L_{slp}^{\;0.6} \cdot n^{0.6}}{18 \cdot slp^{0.3}} \qquad 6.3.6$$

Table 6-3: Values of Manning's roughness coefficient, n, for overland flow (Engman, 1983).

Characteristics of Land Surface	**Median**	**Range**
Fallow, no residue	0.010	0.008-0.012
Conventional tillage, no residue	0.090	0.060-0.120
Conventional tillage, residue	0.190	0.160-0.220
Chisel plow, no residue	0.090	0.060-0.120
Chisel plow, residue	0.130	0.100-0.160
Fall disking, residue	0.400	0.300-0.500
No till, no residue	0.070	0.040-0.100
No till, 0.5-1 t/ha residue	0.120	0.070-0.170
No till, 2-9 t/ha residue	0.300	0.170-0.470
Rangeland, 20% cover	0.600	
Short grass prairie	0.150	0.100-0.200
Dense grass	0.240	0.170-0.300
Bermudagrass	0.410	0.300-0.480

6.3.1.2 CHANNEL FLOW TIME OF CONCENTRATION

The channel flow time of concentration, t_{ch}, can be computed using the equation:

$$t_{ch} = \frac{L_c}{3.6 \cdot v_c} \qquad 6.3.7$$

where L_c is the average flow channel length for the subbasin (km), v_c is the average channel velocity (m s^{-1}), and 3.6 is a unit conversion factor.

The average channel flow length can be estimated using the equation

$$L_c = \sqrt{L \cdot L_{cen}} \qquad 6.3.8$$

where L is the channel length from the most distant point to the subbasin outlet (km), and L_{cen} is the distance along the channel to the subbasin centroid (km). Assuming $L_{cen} = 0.5 \cdot L$, the average channel flow length is

$$L_c = 0.71 \cdot L \qquad 6.3.9$$

The average velocity can be estimated from Manning's equation assuming a trapezoidal channel with 2:1 side slopes and a 10:1 bottom width-depth ratio.

$$v_c = \frac{0.489 \cdot {q_{ch}}^{0.25} \cdot {slp_{ch}}^{0.375}}{n^{0.75}} \qquad 6.3.10$$

where v_c is the average channel velocity (m s^{-1}), q_{ch} is the average channel flow rate (m^3 s^{-1}), slp_{ch} is the channel slope (m m^{-1}), and n is Manning's roughness coefficient for the channel. To express the average channel flow rate in units of mm/hr, the following expression is used

$$q_{ch} = \frac{q_{ch}^{*} \cdot Area}{3.6} \qquad 6.3.11$$

where q_{ch}^{*} is the average channel flow rate (mm hr^{-1}), $Area$ is the subbasin area (km^2), and 3.6 is a unit conversion factor. The average channel flow rate is related to the unit source area flow rate (unit source area = 1 ha)

$$q_{ch}^{*} = q_0^{*} \cdot (100 \cdot Area)^{-0.5} \qquad 6.3.12$$

where q_0^{*} is the unit source area flow rate (mm hr^{-1}), $Area$ is the subbasin area (km^2), and 100 is a unit conversion factor. Assuming the unit source area flow rate is 6.35 mm/hr and substituting equations 6.3.11 and 6.3.12 into 6.3.10 gives

$$v_c = \frac{0.317 \cdot Area^{0.125} \cdot {slp_{ch}}^{0.375}}{n^{0.75}} \qquad 6.3.13$$

Substituting equations 6.3.9 and 6.3.13 into 6.3.7 gives

$$t_{ch} = \frac{0.62 \cdot L \cdot n^{0.75}}{Area^{0.125} \cdot {slp_{ch}}^{0.375}} \qquad 6.3.14$$

where t_{ch} is the time of concentration for channel flow (hr), L is the channel length from the most distant point to the subbasin outlet (km), n is

Manning's roughness coefficient for the channel, *Area* is the subbasin area (km^2), and slp_{ch} is the channel slope ($m\ m^{-1}$).

Table 6-4: Values of Manning's roughness coefficient, *n*, for channel flow (Chow, 1959).[1]

Characteristics of Channel	Median	Range
Excavated or dredged		
Earth, straight and uniform	0.025	0.016-0.033
Earth, winding and sluggish	0.035	0.023-0.050
Not maintained, weeds and brush	0.075	0.040-0.140
Natural streams		
Few trees, stones or brush	0.050	0.025-0.065
Heavy timber and brush	0.100	0.050-0.150

[1] Chow (1959) has a very extensive list of Manning's roughness coefficients. These values represent only a small portion of those he lists in his book.

Although some of the assumptions used in developing equations 6.3.6 and 6.3.14 may appear liberal, the time of concentration values obtained generally give satisfactory results for homogeneous subbasins. Since equations 6.3.6 and 6.3.14 are based on hydraulic considerations, they are more reliable than purely empirical equations.

6.3.2 RUNOFF COEFFICIENT

The runoff coefficient is the ratio of the inflow rate, $i \cdot Area$, to the peak discharge rate, q_{peak}. The coefficient will vary from storm to storm and is calculated with the equation:

$$C = \frac{Q_{surf}}{R_{day}} \qquad 6.3.15$$

where Q_{surf} is the surface runoff (mm H_2O) and R_{day} is the rainfall for the day (mm H_2O).

6.3.3 RAINFALL INTENSITY

The rainfall intensity is the average rainfall rate during the time of concentration. Based on this definition, it can be calculated with the equation:

$$i = \frac{R_{tc}}{t_{conc}} \qquad 6.3.16$$

where i is the rainfall intensity (mm/hr), R_{tc} is the amount of rain falling during the time of concentration (mm H_2O), and t_{conc} is the time of concentration for the subbasin (hr).

An analysis of rainfall data collected by Hershfield (1961) for different durations and frequencies showed that the amount of rain falling during the time of concentration was proportional to the amount of rain falling during the 24-hr period.

$$R_{tc} = \alpha_{tc} \cdot R_{day} \qquad 6.3.17$$

where R_{tc} is the amount of rain falling during the time of concentration (mm H_2O), α_{tc} is the fraction of daily rainfall that occurs during the time of concentration, and R_{day} is the amount of rain falling during the day (mm H_2O).

For short duration storms, all or most of the rain will fall during the time of concentration, causing α_{tc} to approach its upper limit of 1.0. The minimum value of α_{tc} would be seen in storms of uniform intensity ($i_{24} = i$). This minimum value can be defined by substituting the products of time and rainfall intensity into equation 6.3.17

$$\alpha_{tc,\min} = \frac{R_{tc}}{R_{day}} = \frac{i \cdot t_{conc}}{i_{24} \cdot 24} = \frac{t_{conc}}{24} \qquad 6.3.18$$

Thus, α_{tc} falls in the range $t_{conc}/24 \le \alpha_{tc} \le 1.0$.

SWAT estimates the fraction of rain falling in the time of concentration as a function of the fraction of daily rain falling in the half-hour of highest intensity rainfall.

$$\alpha_{tc} = 1 - \exp\left[2 \cdot t_{conc} \cdot \ln\left(1 - \alpha_{0.5}\right)\right] \qquad 6.3.19$$

where $\alpha_{0.5}$ is the fraction of daily rain falling in the half-hour highest intensity rainfall, and t_{conc} is the time of concentration for the subbasin (hr). The determination of a value for $\alpha_{0.5}$ is discussed in Chapters 3 and 4.

6.3.4 MODIFIED RATIONAL FORMULA

The modified rational formula used to estimate peak flow rate is obtained by substituting equations 6.3.15, 6.3.16, and 6.3.17 into equation 6.3.1

$$q_{peak} = \frac{\alpha_{tc} \cdot Q_{surf} \cdot Area}{3.6 \cdot t_{conc}} \qquad 6.3.20$$

where q_{peak} is the peak runoff rate ($m^3\ s^{-1}$), α_{tc} is the fraction of daily rainfall that occurs during the time of concentration, Q_{surf} is the surface runoff (mm H_2O), *Area* is the subbasin area (km^2), t_{conc} is the time of concentration for the subbasin (hr) and 3.6 is a unit conversion factor.

Table 6-5: SWAT input variables that pertain to peak rate calculations.

Variable Name	Definition	Input File
DA_KM	Area of the watershed (km^2)	.bsn
HRU_FR	Fraction of total watershed area contained in HRU	.hru
SLSUBBSN	L_{slp}: Average slope length (m)	.hru
SLOPE	*slp*: Average slope steepness (m/m)	.hru
OV_N	*n*: Manning's "n" value for overland flow	.hru
CH_L(1)	*L*: Longest tributary channel length in subbasin (km)	.sub
CH_S(1)	slp_{ch}: Average slope of tributary channels (m/m)	.sub
CH_N(1)	*n*: Manning's "n" value for tributary channels	.sub

6.4 SURFACE RUNOFF LAG

In large subbasins with a time of concentration greater than 1 day, only a portion of the surface runoff will reach the main channel on the day it is generated. SWAT incorporates a surface runoff storage feature to lag a portion of the surface runoff release to the main channel.

Once surface runoff is calculated with the curve number or Green & Ampt method, the amount of surface runoff released to the main channel is calculated:

$$Q_{surf} = \left(Q'_{surf} + Q_{stor,i-1}\right)\cdot\left(1 - \exp\left[\frac{-surlag}{t_{conc}}\right]\right) \qquad 6.4.1$$

where Q_{surf} is the amount of surface runoff discharged to the main channel on a given day (mm H_2O), Q'_{surf} is the amount of surface runoff generated in the subbasin on a given day (mm H_2O), $Q_{stor,i-1}$ is the surface runoff stored or lagged from the previous day (mm H_2O), *surlag* is the surface runoff lag coefficient, and t_{conc} is the time of concentration for the subbasin (hrs).

The expression $\left(1-\exp\left[\frac{-surlag}{t_{conc}}\right]\right)$ in equation 6.4.1 represents the fraction of the total available water that will be allowed to enter the reach on any one day. Figure 6-3 plots values for this expression at different values for *surlag* and t_{conc}.

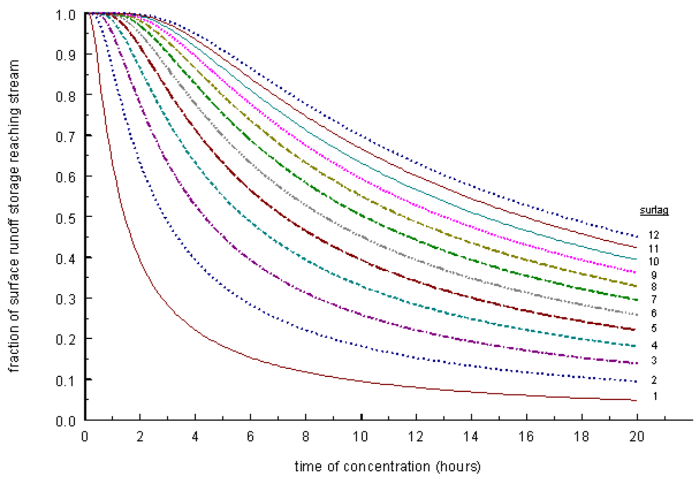

Figure 6-3: Influence of *surlag* and t_{conc} on fraction of surface runoff released.

Note that for a given time of concentration, as *surlag* decreases in value more water is held in storage. The delay in release of surface runoff will smooth the streamflow hydrograph simulated in the reach.

Table 6-6: SWAT input variables that pertain to surface runoff lag calculations.

Variable Name	Definition	Input File
SURLAG	*surlag*: surface runoff lag coefficient	.bsn

6.5 TRANSMISSION LOSSES

Many semiarid and arid watersheds have ephemeral channels that abstract large quantities of streamflow (Lane, 1982). The abstractions, or transmission losses, reduce runoff volume as the flood wave travels downstream. Chapter 19 of the SCS Hydrology Handbook (Lane, 1983) describes a procedure for estimating transmission losses for ephemeral streams which has been incorporated into SWAT. This method was developed to estimate transmission losses in the absence of observed inflow-outflow data and assumes no lateral inflow or out-of-bank flow contributions to runoff.

The prediction equation for runoff volume after transmission losses is

$$vol_{Qsurf,f} = \begin{cases} 0 & vol_{Qsurf,i} \le vol_{thr} \\ a_x + b_x \cdot vol_{Qsurf,i} & vol_{Qsurf,i} > vol_{thr} \end{cases} \qquad 6.5.1$$

where $vol_{Qsurf,f}$ is the volume of runoff after transmission losses (m^3), a_x is the regression intercept for a channel of length L and width W (m^3), b_x is the regression slope for a channel of length L and width W, $vol_{Qsurf,i}$ is the volume of runoff prior to transmission losses (m^3), and vol_{thr} is the threshold volume for a channel of length L and width W (m^3). The threshold volume is

$$vol_{thr} = -\frac{a_x}{b_x} \qquad 6.5.2$$

The corresponding equation for peak runoff rate is

$$q_{peak,f} = \frac{1}{(3600 \cdot dur_{flw})} \cdot \left[a_x - (1 - b_x) \cdot vol_{Qsurf,i}\right] + b_x \cdot q_{peak,i} \qquad 6.5.3$$

where $q_{peak,f}$ is the peak rate after transmission losses (m^3/s), dur_{flw} is the duration of flow (hr), a_x is the regression intercept for a channel of length L and width W (m^3), b_x is the regression slope for a channel of length L and width W, $vol_{Qsurf,i}$ is the volume of runoff prior to transmission losses (m^3), $q_{peak,i}$ is the peak rate before accounting for transmission losses (m^3/s). The duration of flow is calculated with the equation:

$$dur_{flw} = \frac{Q_{surf} \cdot Area}{3.6 \cdot q_{peak}} \qquad 6.5.4$$

where dur_{flw} is the duration of runoff flow (hr), Q_{surf} is the surface runoff (mm H_2O), *Area* is the area of the subbasin (km^2), q_{peak} is the peak runoff rate (m^3/s), and 3.6 is a conversion factor.

In order to calculate the regression parameters for channels of differing lengths and widths, the parameters of a unit channel are needed. A unit channel is defined as a channel of length L = 1 km and width W = 1 m. The unit channel parameters are calculated with the equations:

$$k_r = -2.22 \cdot \ln\left[1 - 2.6466 \cdot \frac{K_{ch} \cdot dur_{flw}}{vol_{Qsurf,i}}\right] \qquad 6.5.5$$

$$a_r = -0.2258 \cdot K_{ch} \cdot dur_{flw} \qquad 6.5.6$$

$$b_r = \exp\left[-0.4905 \cdot k_r\right] \qquad 6.5.7$$

where k_r is the decay factor (m^{-1} km^{-1}), a_r is the unit channel regression intercept (m^3), b_r is the unit channel regression slope, K_{ch} is the effective hydraulic conductivity of the channel alluvium (mm/hr), dur_{flw} is the duration of runoff flow (hr), and $vol_{Qsurf,i}$ is the initial volume of runoff (m^3). The regression parameters are

$$b_x = \exp\left[-k_r \cdot L \cdot W\right] \qquad 6.5.8$$

$$a_x = \frac{a_r}{(1 - b_r)} \cdot (1 - b_x) \qquad 6.5.9$$

where a_x is the regression intercept for a channel of length L and width W (m^3), b_x is the regression slope for a channel of length L and width W, k_r is the decay factor (m^{-1} km^{-1}), L is the channel length from the most distant point to the subbasin outlet (km), W is the average width of flow, i.e. channel width (m) a_r is the unit channel regression intercept (m^3), and b_r is the unit channel regression slope.

Transmission losses from surface runoff are assumed to percolate into the shallow aquifer.

Table 6-7: SWAT input variables that pertain to transmission loss calculations.

Variable Name	Definition	Input File
DA_KM	Area of the watershed (km^2)	.bsn
HRU_FR	Fraction of total watershed area contained in HRU	.hru
CH_K(1)	K_{ch}: effective hydraulic conductivity (mm/hr)	.sub
CH_W(1)	W: average width of tributary channel (m)	.sub
CH_L(1)	L: Longest tributary channel length in subbasin (km)	.sub

6.6 NOMENCLATURE

$Area$ Subbasin area (km^2)
C Runoff coefficient in peak runoff rate calculation
CN Curve number
CN_1 Moisture condition I curve number
CN_2 Moisture condition II curve number
CN_{2s} Moisture condition II curve number adjusted for slope
CN_3 Moisture condition III curve number
F_{inf} Cumulative infiltration at time t (mm H_2O)
FC Water content of soil profile at field capacity (mm H_2O)
I_a Initial abstractions which includes surface storage, interception and infiltration prior to runoff (mm H_2O)
K_{ch} Effective hydraulic conductivity of the channel alluvium (mm/hr)
K_e Effective hydraulic conductivity (mm/hr)
K_{sat} Saturated hydraulic conductivity (mm/hr)
L Channel length from the most distant point to the subbasin outlet (km)
L_c Average flow channel length for the subbasin (km)
L_{cen} Distance along the channel to the subbasin centroid (km)
L_{slp} Subbasin slope length (m)
Q_{stor} Surface runoff stored or lagged (mm H_2O)
Q_{surf} Accumulated runoff or rainfall excess (mm H_2O)
$R_{\Delta t}$ Amount of rain falling during the time step (mm H_2O)
R_{day} Amount of rainfall on a given day (mm H_2O)
R_{tc} Amount of rain falling during the time of concentration (mm H_2O)
S Retention parameter in SCS curve number equation (mm)
S_3 Retention parameter for the moisture condition III curve number
S_{frz} Retention parameter adjusted for frozen conditions (mm)
S_{max} Maximum value the retention parameter can achieve on any given day (mm)
SAT Amount of water in the soil profile when completely saturated (mm H_2O),
SW Amount of water in soil profile (mm H_2O)
W Average width of flow, i.e. channel width (m)

a_r Unit channel regression intercept (m^3)
a_x Regression intercept for a channel of length L and width W (m^3)
b_r Unit channel regression slope
b_x Regression slope for a channel of length L and width W
dur_{flw} Duration of flow (hr)
f_{inf} Infiltration rate (mm/hr)
i Rainfall intensity (mm/hr)
k_r Decay factor (m^{-1} km^{-1})
m_c Percent clay content
m_s Percent sand content
n Manning's roughness coefficient for the subbasin or channel
q_0^* Unit source area flow rate (mm hr^{-1})

q_{ch} Average channel flow rate (m^3 s^{-1})
q^*_{ch} Average channel flow rate (mm hr^{-1})
q_{ov} Average overland flow rate (m^3 s^{-1})
q_{peak} Peak runoff rate (m^3/s)
$q_{peak,f}$ Peak rate after transmission losses (m^3/s)
$q_{peak,i}$ Peak rate before accounting for transmission losses (m^3/s)
slp Average slope of the subbasin (% or m/m)
slp_{ch} Average channel slope (m m^{-1})
$surlag$ Surface runoff lag coefficient
t_{ch} Time of concentration for channel flow (hr)
t_{conc} Time of concentration for a subbasin (hr)
t_{ov} Time of concentration for overland flow (hr)
v_c Average channel velocity (m s^{-1})
v_{ov} Overland flow velocity (m s^{-1})
$vol_{Qsurf,f}$ Volume of runoff after transmission losses (m^3)
$vol_{Qsurf,i}$ Volume of runoff prior to transmission losses (m^3)
vol_{thr} Threshold volume for a channel of length L and width W (m^3)
w_1 Shape coefficient in retention parameter adjustments for soil moisture content
w_2 Shape coefficient in retention parameter adjustments for soil moisture content

$\alpha_{0.5}$ Fraction of daily rain falling in the half-hour highest intensity rainfall,
α_{tc} Fraction of daily rainfall that occurs during the time of concentration
ϕ_{soil} Porosity of the soil (mm/mm)
Ψ_{wf} Wetting front matric potential (mm)
θ_v Volumetric moisture content (mm/mm)

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CHAPTER 7

EQUATIONS: EVAPOTRANSPIRATION

Evapotranspiration is a collective term that includes all processes by which water at the earth's surface is converted to water vapor. It includes evaporation from the plant canopy, transpiration, sublimation and evaporation from the soil.

Evapotranspiration is the primary mechanism by which water is removed from a watershed. Roughly 62% of the precipitation that falls on the continents is evapotranspired. Evapotranspiration exceeds runoff in most river basins and on all continents except Antarctica (Dingman, 1994).

The difference between precipitation and evapotranspiration is the water available for human use and management. An accurate estimation of evapotranspiration is critical in the asssessment of water resources and the impact of climate and land use change on those resources.

7.1 CANOPY STORAGE

The plant canopy can significantly affect infiltration, surface runoff and evapotranspiration. As rain falls, canopy interception reduces the erosive energy of droplets and traps a portion of the rainfall within the canopy. The influence the canopy exerts on these processes is a function of the density of plant cover and the morphology of the plant species.

When calculating surface runoff, the SCS curve number method lumps canopy interception in the term for initial abstractions. This variable also includes surface storage and infiltration prior to runoff and is estimated as 20% of the retention parameter value for a given day (see Chapter 6). When the Green and Ampt infiltration equation is used to calculate surface runoff and infiltration, the interception of rainfall by the canopy must be calculated separately.

SWAT allows the maximum amount of water that can be held in canopy storage to vary from day to day as a function of the leaf area index:

$$can_{day} = can_{mx} \cdot \frac{LAI}{LAI_{mx}} \qquad 7.1.1$$

where can_{day} is the maximum amount of water that can be trapped in the canopy on a given day (mm H_2O), can_{mx} is the maximum amount of water that can be trapped in the canopy when the canopy is fully developed (mm H_2O), LAI is the leaf area index for a given day, and LAI_{mx} is the maximum leaf area index for the plant.

When precipitation falls on any given day, the canopy storage is filled before any water is allowed to reach the ground:

$$R_{INT(f)} = R_{INT(i)} + R'_{day} \quad \text{and} \quad R_{day} = 0 \qquad \text{when } R'_{day} \leq can_{day} - R_{INT(i)} \qquad 7.1.2$$

$$R_{INT(f)} = can_{day} \quad \text{and} \quad R_{day} = R'_{day} - \left(can_{day} - R_{INT(i)}\right) \qquad \text{when } R'_{day} > can_{day} - R_{INT(i)} \qquad 7.1.3$$

where $R_{INT(i)}$ is the initial amount of free water held in the canopy on a given day (mm H_2O), $R_{INT(f)}$ is the final amount of free water held in the canopy on a given day (mm H_2O), R'_{day} is the amount of precipitation on a given day before canopy interception is removed (mm H_2O), R_{day} is the amount of precipitation on a given day that reaches the soil surface (mm H_2O), and can_{day} is the maximum amount of water that can be trapped in the canopy on a given day (mm H_2O).

Table 7-1: SWAT input variables used in canopy storage calculations.

Variable name	Definition	File Name
CANMX	can_{mx}: maximum canopy storage	.hru

7.2 POTENTIAL EVAPOTRANSPIRATION

Potential evapotranspiration (PET) was a concept originally introduced by Thornthwaite (1948) as part of a climate classification scheme. He defined PET is the rate at which evapotranspiration would occur from a large area uniformly covered with growing vegetation that has access to an unlimited supply of soil water and that was not exposed to advection or heat storage effects. Because the evapotranspiration rate is strongly influenced by a number of vegetative surface characteristics, Penman (1956) redefined PET as “the amount of water transpired ... by a short green crop, completely shading the ground, of uniform height and never short of water”. Penman used grass as his reference crop, but later researchers (Jensen, et al., 1990) have suggested that alfalfa at a height of 30 to 50 cm may be a more appropriate choice.

Numerous methods have been developed to estimate PET. Three of these methods have been incorporated into SWAT: the Penman-Monteith method (Monteith, 1965; Allen, 1986; Allen et al., 1989), the Priestley-Taylor method (Priestley and Taylor, 1972) and the Hargreaves method (Hargreaves et al., 1985). The model will also read in daily PET values if the user prefers to apply a different potential evapotranspiration method.

The three PET methods included in SWAT vary in the amount of required inputs. The Penman-Monteith method requires solar radiation, air temperature, relative humidity and wind speed. The Priestley-Taylor method requires solar radiation, air temperature and relative humidity. The Hargreaves method requires air temperature only.

7.2.1 PENMAN-MONTEITH METHOD

The Penman-Monteith equation combines components that account for energy needed to sustain evaporation, the strength of the mechanism required to remove the water vapor and aerodynamic and surface resistance terms. The Penman-Monteith equation is:

$$\lambda E = \frac{\Delta \cdot (H_{net} - G) + \rho_{air} \cdot c_p \cdot \left[e_z^o - e_z\right]/r_a}{\Delta + \gamma \cdot (1 + r_c/r_a)} \qquad 7.2.1$$

where λE is the latent heat flux density (MJ m^{-2} d^{-1}), E is the depth rate evaporation (mm d^{-1}), Δ is the slope of the saturation vapor pressure-temperature curve, de/dT (kPa ˚C^{-1}), H_{net} is the net radiation (MJ m^{-2} d^{-1}), G is the heat flux density to the ground (MJ m^{-2} d^{-1}), ρ_{air} is the air density (kg m^{-3}), c_p is the specific heat at constant pressure (MJ kg^{-1} ˚C^{-1}), e_z^o is the saturation vapor pressure of air at height z (kPa), e_z is the water vapor pressure of air at height z (kPa), γ is the psychrometric constant (kPa ˚C^{-1}), r_c is the plant canopy resistance (s m^{-1}), and r_a is the diffusion resistance of the air layer (aerodynamic resistance) (s m^{-1}).

For well-watered plants under neutral atmospheric stability and assuming logarithmic wind profiles, the Penman-Monteith equation may be written (Jensen et al., 1990):

$$\lambda E_t = \frac{\Delta \cdot (H_{net} - G) + \gamma \cdot K_1 \cdot (0.622 \cdot \lambda \cdot \rho_{air}/P) \cdot (e_z^o - e_z)/r_a}{\Delta + \gamma \cdot (1 + r_c/r_a)} \qquad 7.2.2$$

where λ is the latent heat of vaporization (MJ kg^{-1}), E_t is the maximum transpiration rate (mm d^{-1}), K_1 is a dimension coefficient needed to ensure the two terms in the numerator have the same units (for u_z in m s^{-1}, K_1 = 8.64 x 10^4), and P is the atmospheric pressure (kPa).

The calculation of net radiation, H_{net}, is reviewed in Chapter 2. The calculations for the latent heat of vaporization, λ, the slope of the saturation vapor pressure-temperature curve, Δ, the psychrometric constant, γ, and the saturation and actual vapor pressure, e_z^o and e_z, are reviewed in Chapter 3. The remaining undefined terms are the soil heat flux, G, the combined term $K_1 0.622\lambda\rho/P$, the aerodynamic resistance, r_a, and the canopy resistance, r_c.

7.2.1.1 SOIL HEAT FLUX

Soil heat storage or release can be significant over a few hours, but is usually small from day to day because heat stored as the soil warms early in the day is lost when the soil cools late in the day or at night. Since the magnitude of daily soil heat flux over a 10- to 30-day period is small when the soil is under a crop cover, it can normally be ignored for most energy balance estimates. SWAT assumes the daily soil heat flux, G, is always equal to zero.

7.2.1.2 AERODYNAMIC RESISTANCE

The aerodynamic resistance to sensible heat and vapor transfer, r_a, is calculated:

$$r_a = \frac{\ln\left[\left(z_w - d\right)/z_{om}\right]\ln\left[\left(z_p - d\right)/z_{ov}\right]}{k^2 u_z} \qquad 7.2.3$$

where z_w is the height of the wind speed measurement (cm), z_p is the height of the humidity (psychrometer) and temperature measurements (cm), d is the zero plane displacement of the wind profile (cm), z_{om} is the roughness length for momentum transfer (cm), z_{ov} is the roughness length for vapor transfer (cm), k is the von Kármán constant, and u_z is the wind speed at height z_w (m s^{-1}).

The von Kármán constant is considered to be a universal constant in turbulent flow. Its value has been calculated to be near 0.4 with a range of 0.36 to 0.43 (Jensen et al., 1990). A value of 0.41 is used by SWAT for the von Kármán constant.

Brutsaert (1975) determined that the surface roughness parameter, z_o, is related to the mean height (h_c) of the plant canopy by the relationship

h_c/z_o = $3e$ or 8.15 where e is the natural log base. Based on this relationship, the roughness length for momentum transfer is estimated as:

$$z_{om} = h_c / 8.15 = 0.123 \cdot h_c \qquad h_c \leq 200cm \qquad 7.2.4$$

$$z_{om} = 0.058 \cdot (h_c)^{1.19} \qquad h_c > 200cm \qquad 7.2.5$$

where mean height of the plant canopy (h_c) is reported in centimeters.

The roughness length for momentum transfer includes the effects of bluff-body forces. These forces have no impact on heat and vapor transfer, and the roughness length for vapor transfer is only a fraction of that for momentum transfer. To estimate the roughness length for vapor transfer, Stricker and Brutsaert (1978) recommended using:

$$z_{ov} = 0.1 \cdot z_{om} \qquad 7.2.6$$

The displacement height for a plant can be estimated using the following relationship (Monteith, 1981; Plate, 1971):

$$d = 2/3 \cdot h_c \qquad 7.2.7$$

The height of the wind speed measurement, z_w, and the height of the humidity (psychrometer) and temperature measurements, z_p, are always assumed to be 170 cm.

7.2.1.3 CANOPY RESISTANCE

Studies in canopy resistance have shown that the canopy resistance for a well-watered reference crop can be estimated by dividing the minimum surface resistance for a single leaf by one-half of the canopy leaf area index (Jensen et. al, 1990):

$$r_c = r_\ell / (0.5 \cdot LAI) \qquad 7.2.8$$

where r_c is the canopy resistance (s m^{-1}), r_ℓ is the minimum effective stomatal resistance of a single leaf (s m^{-1}), and LAI is the leaf area index of the canopy.

The distribution of stomates on a plant leaf will vary between species. Typically, stomates are distributed unequally on the top and bottom of plant leaves. Plants with stomates located on only one side are classified as hypostomatous while plants with an equal number of stomates on both sides of the leaf are termed amphistomatous. The effective leaf stomatal resistance is determined by considering the stomatal resistance of the top (adaxial) and bottom (abaxial) sides to be connected in parallel (Rosenburg, et al., 1983). When there are unequal numbers of stomates on the top and bottom, the effective stomatal resistance is calculated:

$$r_\ell = \frac{r_{\ell-ad} \cdot r_{\ell-ab}}{r_{\ell-ab} + r_{\ell-ad}} \qquad 7.2.9$$

where r_ℓ is the minimum effective stomatal resistance of a single leaf (s m^{-1}), $r_{\ell-ad}$ is the minimum adaxial stomatal leaf resistance (s m^{-1}), and $r_{\ell-ab}$ is the minimum abaxial stomatal leaf resistance (s m^{-1}). For amphistomatous leaves, the effective stomatal resistance is:

$$r_\ell = \frac{r_{\ell-ad}}{2} = \frac{r_{\ell-ab}}{2} \qquad 7.2.10$$

For hypostomatous leaves the effective stomatal resistance is:

$$r_\ell = r_{\ell-ad} = r_{\ell-ab} \qquad 7.2.11$$

Leaf conductance is defined as the inverse of the leaf resistance:

$$g_\ell = \frac{1}{r_\ell} \qquad 7.2.12$$

where g_ℓ is the maximum effective leaf conductance (m s^{-1}). When the canopy resistance is expressed as a function of leaf conductance instead of leaf resistance, equation 7.2.8 becomes:

$$r_c = (0.5 \cdot g_\ell \cdot LAI)^{-1} \qquad 7.2.13$$

where r_c is the canopy resistance (s m^{-1}), g_ℓ is the maximum conductance of a single leaf (m s^{-1}), and *LAI* is the leaf area index of the canopy.

For climate change simulations, the canopy resistance term can be modified to reflect the impact of change in CO_2 concentration on leaf conductance. The influence of increasing CO_2 concentrations on leaf conductance was reviewed by Morison (1987). Morison found that at CO_2 concentrations between 330 and 660 ppmv, a doubling in CO_2 concentration resulted in a 40% reduction in leaf conductance. Within the specified range, the reduction in conductance is linear (Morison and Gifford, 1983). Easterling et al. (1992) proposed the following modification to the leaf conductance term for simulating carbon dioxide concentration effects on evapotranspiration:

$$g_{\ell,CO_2} = g_\ell \cdot \left[1.4 - 0.4 \cdot \left(CO_2/330\right)\right] \qquad 7.2.14$$

where g_{ℓ,CO_2} is the leaf conductance modified to reflect CO_2 effects (m s^{-1}) and CO_2 is the concentration of carbon dioxide in the atmosphere (ppmv).

Incorporating this modification into equation 7.2.8 gives

$$r_c = r_\ell \cdot \left[\left(0.5 \cdot LAI\right) \cdot \left(1.4 - 0.4 \cdot \frac{CO_2}{330}\right)\right]^{-1} \qquad 7.2.15$$

SWAT will default the value of CO_2 concentration to 330 ppmv if no value is entered by the user. With this default, the term $\left(1.4 - 0.4 \cdot \frac{CO_2}{330}\right)$ reduces to 1.0 and the canopy resistance equation becomes equation 7.2.8.

When calculating actual evapotranspiration, the canopy resistance term is modified to reflect the impact of high vapor pressure deficit on leaf conductance (Stockle et al, 1992). For a plant species, a threshold vapor pressure deficit is defined at which the plant's leaf conductance begins to drop in response to the vapor pressure deficit. The adjusted leaf conductance is calculated:

$$g_\ell = g_{\ell,mx} \cdot \left[1 - \Delta g_{\ell,dcl}\left(vpd - vpd_{thr}\right)\right] \quad \text{if } vpd > vpd_{thr} \qquad 7.2.16$$

$$g_\ell = g_{\ell,mx} \quad \text{if } vpd \le vpd_{thr} \qquad 7.2.17$$

where g_ℓ is the conductance of a single leaf (m s^{-1}), $g_{\ell,mx}$ is the maximum conductance of a single leaf (m s^{-1}), $\Delta g_{\ell,dcl}$ is the rate of decline in leaf conductance per unit increase in vapor pressure deficit (m s^{-1} kPa^{-1}), vpd is the vapor pressure deficit (kPa), and vpd_{thr} is the threshold vapor pressure deficit above which a plant will exhibit reduced leaf conductance (kPa). The rate of decline in leaf conductance per unit increase in vapor pressure deficit is calculated by solving equation 7.2.16 using measured values for stomatal conductance at two different vapor pressure deficits:

$$\Delta g_{\ell,dcl} = \frac{\left(1 - fr_{g,mx}\right)}{\left(vpd_{fr} - vpd_{thr}\right)} \qquad 7.2.18$$

where $\Delta g_{\ell,dcl}$ is the rate of decline in leaf conductance per unit increase in vapor pressure deficit (m s^{-1} kPa^{-1}), $fr_{g,mx}$ is the fraction of the maximum stomatal conductance, $g_{\ell,mx}$, achieved at the vapor pressure deficit vpd_{fr}, and vpd_{thr} is the threshold vapor pressure deficit above which a plant will exhibit reduced leaf conductance (kPa). The threshold vapor pressure deficit is assumed to be 1.0 kPa for all plant species.

7.2.1.4 COMBINED TERM

For wind speed in m s^{-1}, Jensen et al. (1990) provided the following relationship to calculate $K_1 0.622 \lambda \rho / P$:

$$K_1 \cdot 0.622 \cdot \lambda \cdot \rho / P = 1710 - 6.85 \cdot \overline{T}_{av} \qquad 7.2.19$$

where $\overline{T}_{av}$ is the mean air temperature for the day (˚C).

To calculate potential evapotranspiration, the Penman-Monteith equation must be solved for a reference crop. SWAT uses alfalfa at a height of 40 cm with a minimum leaf resistance of 100 s m^{-1} for the reference crop. Using this canopy height, the equation for aerodynamic resistance (7.2.3) simplifies to:

$$r_a = \frac{114.}{u_z} \qquad 7.2.20$$

The equation for canopy resistance requires the leaf area index. The leaf area index for the reference crop is estimated using an equation developed by Allen et al. (1989) to calculate *LAI* as a function of canopy height. For nonclipped grass and alfalfa greater than 3 cm in height:

$$LAI = 1.5 \cdot \ln(h_c) - 1.4 \qquad 7.2.21$$

where LAI is the leaf area index and h_c is the canopy height (cm). For alfalfa with a 40 cm canopy height, the leaf area index is 4.1. Using this value, the equation for canopy resistance simplifies to:

$$r_c = 49 \bigg/ \left(1.4 - 0.4 \cdot \frac{CO_2}{330}\right) \qquad 7.2.22$$

The most accurate estimates of evapotranspiration with the Penman-Monteith equation are made when evapotranspiration is calculated on an hourly basis and summed to obtain the daily values. Mean daily parameter values have been shown to provide reliable estimates of daily evapotranspiration values and this is the approach used in SWAT. However, the user should be aware that calculating evapotranspiration with the Penman-Monteith equation using mean daily values can potentially lead to significant errors. These errors result from diurnal distributions of wind speed, humidity, and net radiation that in combination create conditions which the daily averages do not replicate.

7.2.2 PRIESTLEY-TAYLOR METHOD

Priestley and Taylor (1972) developed a simplified version of the combination equation for use when surface areas are wet. The aerodynamic component was removed and the energy component was multiplied by a coefficient, α_{pet} = 1.28, when the general surroundings are wet or under humid conditions

$$\lambda E_o = \alpha_{pet} \cdot \frac{\Delta}{\Delta + \gamma} \cdot (H_{net} - G) \qquad 7.2.23$$

where λ is the latent heat of vaporization (MJ kg^{-1}), E_o is the potential evapotranspiration (mm d^{-1}), α_{pet} is a coefficient, Δ is the slope of the saturation vapor pressure-temperature curve, de/dT (kPa $^{\circ}C^{-1}$), γ is the psychrometric constant (kPa $^{\circ}C^{-1}$), H_{net} is the net radiation (MJ m^{-2} d^{-1}), and G is the heat flux density to the ground (MJ m^{-2} d^{-1}).

The Priestley-Taylor equation provides potential evapotranspiration estimates for low advective conditions. In semiarid or arid areas where the advection component of the energy balance is significant, the Priestley-Taylor equation will underestimate potential evapotranspiration.

7.2.3 HARGREAVES METHOD

The Hargreaves method was originally derived from eight years of cool-season Alta fescue grass lysimeter data from Davis, California (Hargreaves, 1975). Several improvements were made to the original equation (Hargreaves and Samani, 1982 and 1985) and the form used in SWAT was published in 1985 (Hargreaves et al., 1985):

$$\lambda E_o = 0.0023 \cdot H_0 \cdot \left(T_{mx} - T_{mn}\right)^{0.5} \cdot \left(\overline{T}_{av} + 17.8\right) \qquad 7.2.24$$

where λ is the latent heat of vaporization (MJ kg^{-1}), E_o is the potential evapotranspiration (mm d^{-1}), H_0 is the extraterrestrial radiation (MJ m^{-2} d^{-1}), T_{mx} is the maximum air temperature for a given day (°C), T_{mn} is the minimum air temperature for a given day (°C), and $\overline{T}_{av}$ is the mean air temperature for a given day (°C).

Table 7-2: SWAT input variables used in potential evapotranspiration calculations summarized in this section.

Variable name	Definition	File Name
IPET	Potential evapotranspiration method	.cod
WND_SP	u_z: Daily wind speed (m/s)	.wnd
CO2	CO_2: Carbon dioxide concentration (ppmv)	.sub
MAX TEMP	T_{mx}: Daily maximum temperature (°C)	.tmp
MIN TEMP	T_{mn}: Daily minimum temperature (°C)	.tmp
GSI	$g_{\ell,mx}$: maximum leaf conductance (m s^{-1})	crop.dat
FRGMAX	$fr_{g,mx}$: Fraction of maximum leaf conductance achieved at the vapor pressure deficit specified by vpd_{fr}	crop.dat
VPDFR	vpd_{fr}: Vapor pressure deficit corresponding to value given for $fr_{g,mx}$ (kPa)	crop.dat

7.3 ACTUAL EVAPOTRANSPIRATION

Once total potential evapotranspiration is determined, actual evaporation must be calculated. SWAT first evaporates any rainfall intercepted by the plant canopy. Next, SWAT calculates the maximum amount of transpiration and the maximum amount of sublimation/soil evaporation using an approach similar to that of Richtie (1972). The actual amount of sublimation and evaporation from the soil is then calculated. If snow is present in the HRU, sublimation will occur. Only when no snow is present will evaporation from the soil take place.

7.3.1 EVAPORATION OF INTERCEPTED RAINFALL

Any free water present in the canopy is readily available for removal by evapotranspiration. The amount of actual evapotranspiration contributed by intercepted rainfall is especially significant in forests where in some instances evaporation of intercepted rainfall is greater than transpiration.

SWAT removes as much water as possible from canopy storage when calculating actual evaporation. If potential evapotranspiration, E_o, is less than the amount of free water held in the canopy, R_{INT}, then

$$E_a = E_{can} = E_o \qquad 7.3.1$$

$$R_{INT(f)} = R_{INT(i)} - E_{can} \qquad 7.3.2$$

where E_a is the actual amount of evapotranspiration occurring in the watershed on a given day (mm H_2O), E_{can} is the amount of evaporation from free water in the canopy on a given day (mm H_2O), E_o is the potential evapotranspiration on a given day (mm H_2O), $R_{INT(i)}$ is the initial amount of free water held in the canopy on a given day (mm H_2O), and $R_{INT(f)}$ is the final amount of free water held in the canopy on a given day (mm H_2O). If potential evapotranspiration, E_o, is greater than the amount of free water held in the canopy, R_{INT}, then

$$E_{can} = R_{INT(i)} \qquad 7.3.3$$

$$R_{INT(f)} = 0 \qquad 7.3.4$$

Once any free water in the canopy has been evaporated, the remaining evaporative water demand ($E'_o = E_o - E_{can}$) is partitioned between the vegetation and snow/soil.

7.3.2 TRANSPIRATION

If the Penman-Monteith equation is selected as the potential evapotranspiration method, transpiration is also calculated with the equations summarized in Section 7.2.1. For the other potential evapotranspiration methods, transpiration is calculated as:

$$E_t = \frac{E'_o \cdot LAI}{3.0} \qquad 0 \le LAI \le 3.0 \qquad 7.3.5$$

$$E_t = E'_o \qquad LAI > 3.0 \qquad 7.3.6$$

where E_t is the maximum transpiration on a given day (mm H_2O), E'_o is the potential evapotranspiration adjusted for evaporation of free water in the canopy (mm H_2O), and LAI is the leaf area index. The value for transpiration calculated by equations 7.3.5 and 7.3.6 is the amount of transpiration that will occur on a given day when the plant is growing under ideal conditions. The actual amount of transpiration may be less than this due to lack of available water in the soil profile. Calculation of actual plant water uptake and transpiration is reviewed in Chapters 18 and 19.

7.3.3 SUBLIMATION AND EVAPORATION FROM THE SOIL

The amount of sublimation and soil evaporation will be impacted by the degree of shading. The maximum amount of sublimation/soil evaporation on a given day is calculated as:

$$E_s = E'_o \cdot cov_{sol} \qquad 7.3.7$$

where E_s is the maximum sublimation/soil evaporation on a given day (mm H_2O), E'_o is the potential evapotranspiration adjusted for evaporation of free water in the canopy (mm H_2O), and cov_{sol} is the soil cover index. The soil cover index is calculated

$$cov_{sol} = \exp\left(-5.0 \times 10^{-5} \cdot CV\right) \qquad 7.3.8$$

where CV is the aboveground biomass and residue (kg ha^{-1}). If the snow water content is greater than 0.5 mm H_2O, the soil cover index is set to 0.5.

The maximum amount of sublimation/soil evaporation is reduced during periods of high plant water use with the relationship:

$$E'_s = \min\left[E_s,\quad \frac{E_s \cdot E'_o}{E_s + E_t}\right] \qquad 7.3.9$$

where E'_s is the maximum sublimation/soil evaporation adjusted for plant water use on a given day (mm H_2O), E_s is the maximum sublimation/soil evaporation on a given day (mm H_2O), E'_o is the potential evapotranspiration adjusted for evaporation of free water in the canopy (mm H_2O), and E_t is the transpiration on a given day (mm H_2O). When E_t is low $E'_s \rightarrow E_s$. However, as E_t approaches E'_o, $E'_s \rightarrow \dfrac{E_s}{1 + cov_{sol}}$.

7.3.3.1 SUBLIMATION

Once the maximum amount of sublimation/soil evaporation for the day is calculated, SWAT will first remove water from the snow pack to meet the evaporative demand. If the water content of the snow pack is greater than the maximum sublimation/soil evaporation demand, then

$$E_{sub} = E'_s \qquad 7.3.10$$

$$SNO_{(f)} = SNO_{(i)} - E'_s \qquad 7.3.11$$

$$E''_s = 0. \qquad 7.3.12$$

where E_{sub} is the amount of sublimation on a given day (mm H_2O), E'_s is the maximum sublimation/soil evaporation adjusted for plant water use on a given day (mm H_2O), $SNO_{(i)}$ is the amount of water in the snow pack on a given day prior to accounting for sublimation (mm H_2O), $SNO_{(f)}$ is the amount of water in the snow pack on a given day after accounting for sublimation (mm H_2O), and E''_s is the maximum soil water evaporation on a given day (mm H_2O). If the water content of the snow pack is less than the maximum sublimation/soil evaporation demand, then

$$E_{sub} = SNO_{(i)} \quad 7.3.13$$

$$SNO_{(f)} = 0. \quad 7.3.14$$

$$E''_s = E'_s - E_{sub} \quad 7.3.15$$

7.3.3.2 SOIL WATER EVAPORATION

When an evaporation demand for soil water exists, SWAT must first partition the evaporative demand between the different layers. The depth distribution used to determine the maximum amount of water allowed to be evaporated is:

$$E_{soil,z} = E''_s \cdot \frac{z}{z + \exp(2.374 - 0.00713 \cdot z)} \quad 7.3.16$$

where $E_{soil,z}$ is the evaporative demand at depth z (mm H_2O), E''_s is the maximum soil water evaporation on a given day (mm H_2O), and z is the depth below the surface. The coefficients in this equation were selected so that 50% of the evaporative demand is extracted from the top 10 mm of soil and 95% of the evaporative demand is extracted from the top 100 mm of soil.

The amount of evaporative demand for a soil layer is determined by taking the difference between the evaporative demands calculated at the upper and lower boundaries of the soil layer:

$$E_{soil,ly} = E_{soil,zl} - E_{soil,zu} \quad 7.3.16$$

where $E_{soil,ly}$ is the evaporative demand for layer ly (mm H_2O), $E_{soil,zl}$ is the evaporative demand at the lower boundary of the soil layer (mm H_2O), and $E_{soil,zu}$ is the evaporative demand at the upper boundary of the soil layer (mm H_2O).

Figure 7-1 graphs the depth distribution of the evaporative demand for a soil that has been partitioned into 1 mm layers assuming a total soil evaporation demand of 100 mm.

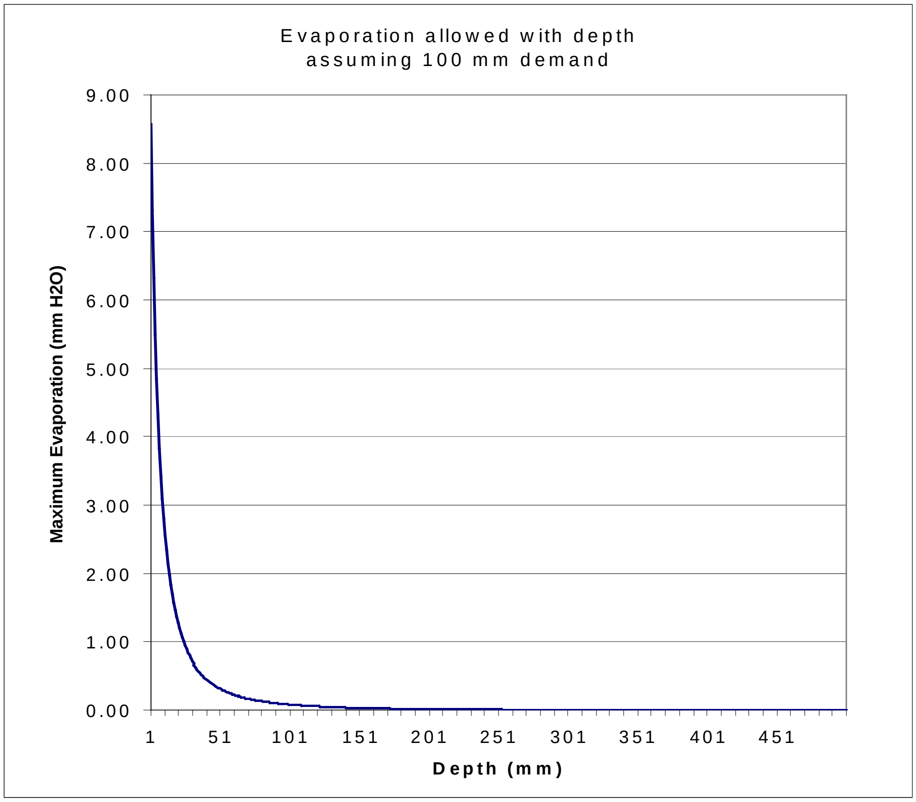

Figure 7-1: Soil evaporative demand distribution with depth.

As mentioned previously, the depth distribution assumes 50% of the evaporative demand is met by soil water stored in the top 10 mm of the soil profile. With our example of a 100 mm total evaporative demand, 50 mm of water is 50%. This is a demand that the top layer cannot satisfy.

SWAT does not allow a different layer to compensate for the inability of another layer to meet its evaporative demand. The evaporative demand not met by a soil layer results in a reduction in actual evapotranspiration for the HRU.

A coefficient has been incorporated into equation 7.3.16 to allow the user to modify the depth distribution used to meet the soil evaporative demand. The modified equation is:

$$E_{soil,ly} = E_{soil,zl} - E_{soil,zu} \cdot esco \qquad 7.3.17$$

where $E_{soil,ly}$ is the evaporative demand for layer ly (mm H_2O), $E_{soil,zl}$ is the evaporative demand at the lower boundary of the soil layer (mm H_2O), $E_{soil,zu}$ is the evaporative demand at the upper boundary of the soil layer (mm H_2O), and *esco* is the soil evaporation compensation coefficient. Solutions to this equation for different values of *esco* are graphed in Figure 7-2. The plot for *esco* = 1.0 is that shown in Figure 7-1.

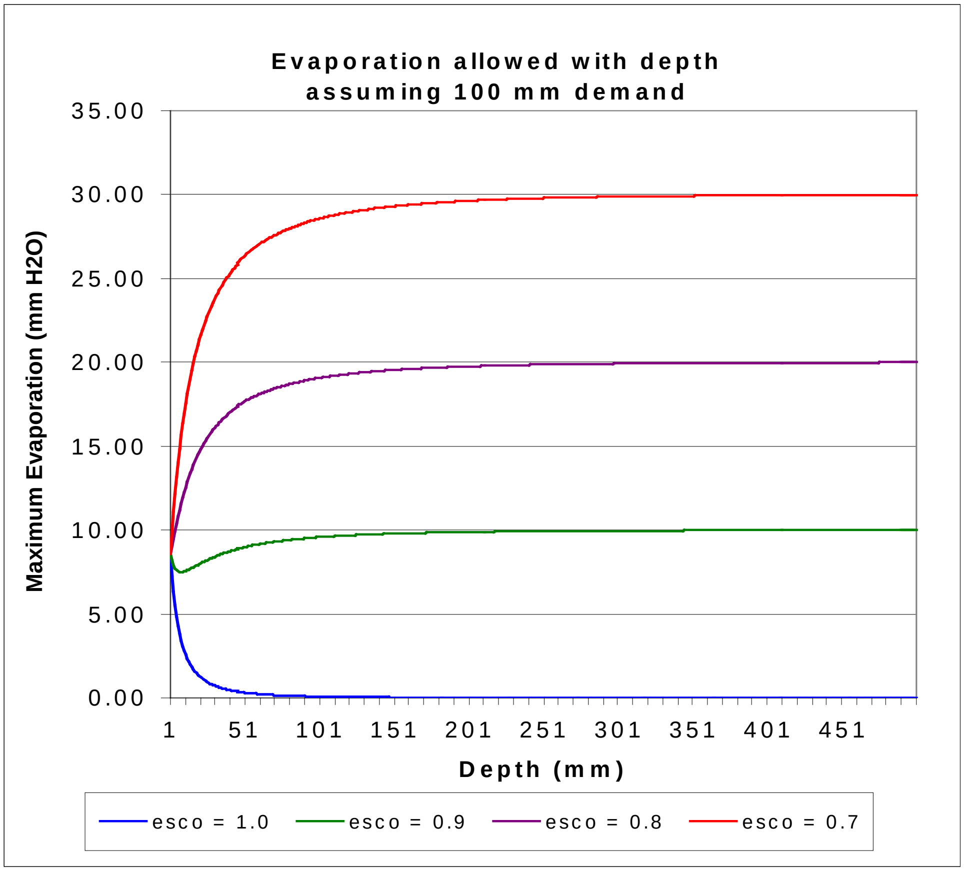

Figure 7-2: Soil evaporative demand distribution with depth

As the value for *esco* is reduced, the model is able to extract more of the evaporative demand from lower levels.

When the water content of a soil layer is below field capacity, the evaporative demand for the layer is reduced according to the following equations:

$$E'_{soil,ly} = E_{soil,ly} \cdot \exp\left(\frac{2.5 \cdot \left(SW_{ly} - FC_{ly}\right)}{FC_{ly} - WP_{ly}}\right) \quad \text{when } SW_{ly} < FC_{ly} \qquad 7.3.18$$

$$E'_{soil,ly} = E_{soil,ly} \quad \text{when } SW_{ly} \geq FC_{ly} \qquad 7.3.19$$

where $E'_{soil,ly}$ is the evaporative demand for layer *ly* adjusted for water content (mm H_2O), $E_{soil,ly}$ is the evaporative demand for layer *ly* (mm

H_2O), SW_{ly} is the soil water content of layer ly (mm H_2O), FC_{ly} is the water content of layer ly at field capacity (mm H_2O), and WP_{ly} is the water content of layer ly at wilting point (mm H_2O).

In addition to limiting the amount of water removed by evaporation in dry conditions, SWAT defines a maximum value of water that can be removed at any time. This maximum value is 80% of the plant available water on a given day where the plant available water is defined as the total water content of the soil layer minus the water content of the soil layer at wilting point (-1.5 MPa).

$$E''_{soil,ly} = \min\left(E'_{soil,ly} \quad 0.8 \cdot (SW_{ly} - WP_{ly})\right) \qquad 7.3.20$$

where $E''_{soil,ly}$ is the amount of water removed from layer ly by evaporation (mm H_2O), $E'_{soil,ly}$ is the evaporative demand for layer ly adjusted for water content (mm H_2O), SW_{ly} is the soil water content of layer ly (mm H_2O), and WP_{ly} is the water content of layer ly at wilting point (mm H_2O).

Table 7-3: SWAT input variables used in soil evaporation calculations.

Variable name	Definition	File Name
ESCO	*esco*: soil evaporation compensation coefficient	.bsn, .hru

7.4 NOMENCLATURE

CO_2 Concentration of carbon dioxide in the atmosphere (ppmv)
CV Total aboveground biomass and residue present on current day (kg ha^{-1})
E Depth rate evaporation (mm d^{-1})
E_a Actual amount of evapotranspiration on a given day (mm H_2O)
E_{can} Amount of evaporation from free water in the canopy on a given day (mm H_2O)
E_o Potential evapotranspiration (mm d^{-1})
E'_o Potential evapotranspiration adjusted for evaporation of free water in the canopy (mm H_2O)
E_s Maximum sublimation/soil evaporation on a given day (mm H_2O)
E'_s Maximum sublimation/soil evaporation adjusted for plant water use on a given day (mm H_2O)
E''_s Maximum soil water evaporation on a given day (mm H_2O)
$E_{soil,ly}$ Evaporative demand for layer ly (mm H_2O)

$E'_{soil,ly}$	Evaporative demand for layer *ly* adjusted for water content (mm H_2O)
$E''_{soil,ly}$	Amount of water removed from layer *ly* by evaporation (mm H_2O)
$E_{soil,z}$	Evaporative demand at depth *z* (mm H_2O)
E_{sub}	Amount of sublimation on a given day (mm H_2O)
E_t	Transpiration rate (mm d^{-1})
FC_{ly}	Water content of layer *ly* at field capacity (mm H_2O)
G	Heat flux density to the ground (MJ m^{-2} d^{-1})
H_0	Extraterrestrial daily irradiation (MJ m^{-2} d^{-1})
H_{net}	Net radiation on day (MJ m^{-2} d^{-1})
K_1	Dimension coefficient in Penman-Monteith equation
LAI	Leaf area index of the canopy
LAI_{mx}	Maximum leaf area index for the plant
P	Atmospheric pressure (kPa)
R_{day}	Amount of rainfall on a given day (mm H_2O)
R'_{day}	Amount of precipitation on a given day before canopy interception is removed (mm H_2O)
R_{INT}	Amount of free water held in the canopy on a given day (mm H_2O)
SNO	Water content of snow cover on current day (mm H_2O)
SW_{ly}	Soil water content of layer *ly* (mm H_2O)
T_{mn}	Minimum air temperature for day (°C)
T_{mx}	Maximum air temperature for day (°C)
$\overline{T}_{av}$	Mean air temperature for day (°C)
WP_{ly}	Water content of layer *ly* at wilting point (mm H_2O).

c_p	Specific heat of moist air at constant pressure (1.013×10^{-3} MJ kg^{-1} $°C^{-1}$)
can_{day}	Maximum amount of water that can be trapped in the canopy on a given day (mm H_2O)
can_{mx}	Maximum amount of water that can be trapped in the canopy when the canopy is fully developed (mm H_2O)
cov_{sol}	Soil cover index
d	Zero plane displacement of the wind profile (cm)
e	Actual vapor pressure on a given day (kPa)
e^o	Saturation vapor pressure on a given day (kPa)
$esco$	Soil evaporation compensation coefficient
$fr_{g,mx}$	Fraction of the maximum stomatal conductance, $g_{\ell,mx}$, achieved at the vapor pressure deficit, vpd_{fr}
g_ℓ	Leaf conductance (m s^{-1})
$g_{\ell,mx}$	Maximum conductance of a single leaf (m s^{-1})
h_c	Canopy height (cm)
k	Von Kármán constant
r_a	Diffusion resistance of the air layer (aerodynamic resistance) (s m^{-1})
r_c	Plant canopy resistance (s m^{-1})
r_ℓ	Minimum effective resistance of a single leaf (s m^{-1})

$r_{\ell-ab}$ Minimum abaxial stomatal leaf resistance (s m^{-1})
$r_{\ell-ad}$ Minimum adaxial stomatal leaf resistance (s m^{-1})
u_z Wind speed at height z_w (m s^{-1})
vpd Vapor pressure deficit (kPa)
vpd_{fr} Vapor pressure deficit corresponding to $fr_{g,mx}$ (kPa)
vpd_{thr} Threshold vapor pressure deficit above which a plant will exhibit reduced leaf conductance (kPa)
z Depth below soil surface (mm)
z_{om} Roughness length for momentum transfer (cm)
z_{ov} Roughness length for vapor transfer (cm)
z_p Height of the humidity (psychrometer) and temperature measurements (cm)
z_w Height of the wind speed measurement (cm)

α_{pet} Coefficient in Priestley-Taylor equation
Δ Slope of the saturation vapor pressure curve (kPa °C^{-1})
$\Delta g_{\ell,dcl}$ Rate of decline in leaf conductance per unit increase in vapor pressure deficit (m s^{-1} kPa^{-1})
ρ_{air} Air density (kg m^{-3})
γ Psychrometric constant (kPa °C^{-1})
λ Latent heat of vaporization (MJ kg^{-1})

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CHAPTER **8**

EQUATIONS: SOIL WATER

Water that enters the soil may move along one of several different pathways. The water may be removed from the soil by plant uptake or evaporation. It can percolate past the bottom of the soil profile and ultimately become aquifer recharge. A final option is that water may move laterally in the profile and contribute to streamflow. Of these different pathways, plant uptake of water removes the majority of water that enters the soil profile.

8.1 SOIL STRUCTURE

Soil is comprised of three phases—solid, liquid and gas. The solid phase consists of minerals and/or organic matter that forms the matrix or skeleton of the soil. Between the solid particles, soil pores are formed that hold the liquid and gas phases. The soil solution may fill the soil pores completely (saturated) or partially (unsaturated). When the soil is unsaturated, the soil solution is found as thin films along particle surfaces, as annular wedges around contact points of particles and as isolated bodies in narrow pore passages.

The soil's bulk density defines the relative amounts of pore space and soil matrix. Bulk density is calculated:

$$\rho_b = \frac{M_S}{V_T} \qquad 8.1.1$$

where ρ_b is the bulk density (Mg m^{-3}), M_S is the mass of the solids (Mg), and V_T is the total volume (m^3). The total volume is defined as

$$V_T = V_A + V_W + V_S \qquad 8.1.2$$

where V_A is the volume of air (m^3), V_W is the volume of water (m^3), and V_S is the volume of solids (m^3). The relationship between soil porosity and soil bulk density is

$$\phi_{soil} = 1 - \frac{\rho_b}{\rho_s} \qquad 8.1.3$$

where ϕ_{soil} is the soil porosity expressed as a fraction of the total soil volume, ρ_b is the bulk density (Mg m^{-3}), and ρ_s is the particle density (Mg m^{-3}). The particle density, or density of the solid fraction, is a function of the mineral composition of the soil matrix. Based on research, a default value of 2.65 Mg m^{-3} is used for particle density.

Storage, transport and availability of soil solution and soil air are not nearly as dependent on the total amount of porosity as they are on the arrangement of pore space. Soil pores vary in size and shape due to textural and structural arrangement. Based on the diameter of the pore at the narrowest point, the pores may be classified as macropores (narrowest diameter > 100 μm), mesopores (narrowest diameter 30-100 μm), and micropores (narrowest diameter < 30 μm) (Koorevaar et al, 1983). Macropores conduct water only during flooding or ponding rain and drainage of water from these pores is complete soon after cessation of the water supply. Macropores control aeration and drainage processes in the soil. Mesopores conduct water even after macropores have emptied, e.g. during non-ponding rain and redistribution. Micropores retain soil solution or conduct it very slowly.

When comparing soils of different texture, clay soils contain a greater fraction of mesopores and micropores while sand soils contain mostly macropores. This is evident when the hydraulic conductivities of clay and sand soils are compared. The conductivity of a sand soil can be several orders of magnitude greater than that for a clay soil.

The water content of a soil can range from zero when the soil is oven dried to a maximum value (ϕ_{soil}) when the soil is saturated. For plant-soil interactions, two intermediate stages are recognized: field capacity and permanent wilting point. Field capacity is the water content found when a thoroughly wetted soil has drained for approximately two days. Permanent wilting point is the water content found when plants growing in the soil wilt and do not recover if their leaves are kept in a humid atmosphere overnight. To allow these two stages to be quantified more easily, they have been redefined in terms of tensions at which water is held by the soil. Field capacity is the amount of water held in the soil at a tension of 0.033 MPa and the permanent wilting point is the amount of water held in the soil at a tension of 1.5 MPa. The amount of water held in the soil between field capacity and permanent wilting point is considered to be the water available for plant extraction.

Table 8-1: Water contents for various soils at different moisture conditions.

		Water content (fraction total soil volume)		
Texture	Clay Content (% Solids)	Saturation	Field capacity	Permanent wilting point
sand	3 %	0.40	0.06	0.02
loam	22 %	0.50	0.29	0.05
clay	47 %	0.60	0.41	0.20

Table 8-1 lists the water content for three soils as a fraction of the total volume for different moisture conditions. Note that the total porosity, given by the water content at saturation, is lowest for the sand soil and highest for the clay soil.

The sand soil drains more quickly than the loam and clay. Only 15% of the water present in the sand soil at saturation remains at field capacity. 58% of the water present at saturation in the loam remains at field capacity while 68% of the water present at saturation in the clay soil remains at field capacity. The reduction of water loss with increase in clay content is cause by two factors. As mentioned previously, clay soils contain more mesopores and micropores than sand soils. Also, unlike sand and silt particles, clay particles possess a net negative charge. Due to the polar nature of water molecules, clay particles are able to attract and retain water molecules. The higher water retention of clay soils is also seen in the fraction of water present at permanent wilting point. In the soils listed in Table 8-1, the volumetric water content of the clay is 0.20 at the wilting point while the sand and loam have a volumetric water content of 0.02 and 0.05 respectively.

The plant available water, also referred to as the available water capacity, is calculated by subtracting the fraction of water present at permanent wilting point from that present at field capacity.

$$AWC = FC - WP \qquad 8.1.4$$

where AWC is the plant available water content, FC is the water content at field capacity, and WP is the water content at permanent wilting point. For the three soil textures listed in Table 8-1, the sand has an available water capacity of 0.04, the loam has an available water capacity of 0.24 and the clay has an available water capacity of 0.21. Even though the clay contains a greater amount of water than the loam at all three tensions, the loam has a larger amount of water available for plant uptake than the clay. This characteristic is true in general.

SWAT estimates the permanent wilting point volumetric water content for each soil layer as:

$$WP_{ly} = 0.40 \cdot \frac{m_c \cdot \rho_b}{100} \qquad 8.1.5$$

where WP_{ly} is the water content at wilting point expressed as a fraction of the total soil volume, m_c is the percent clay content of the layer (%), and ρ_b is the bulk density for the soil layer (Mg m^{-3}). Field capacity water content is estimated

$$FC_{ly} = WP_{ly} + AWC_{ly} \qquad 8.1.6$$

where FC_{ly} is the water content at field capacity expressed as a fraction of the total soil volume, WP_{ly} is the water content at wilting point expressed as a fraction of the total soil volume, and AWC_{ly} is the available water capacity of the soil layer expressed as a fraction of the total soil volume. AWC_{ly} is input by the user.

Water in the soil can flow under saturated or unsaturated conditions. In saturated soils, flow is driven by gravity and usually occurs in the downward direction. Unsaturated flow is caused by gradients arising due to adjacent areas of high and low water content. Unsaturated flow may occur in any direction.

SWAT directly simulates saturated flow only. The model records the water contents of the different soil layers but assumes that the water is uniformly distributed within a given layer. This assumption eliminates the need to model unsaturated flow in the horizontal direction. Unsaturated flow between layers is indirectly modeled with the depth distribution of plant water uptake (equation 18.2.1) and the depth distribution of soil water evaporation (equation 7.3.16).

Saturated flow occurs when the water content of a soil layer surpasses the field capacity for the layer. Water in excess of the field capacity water content is available for percolation, lateral flow or tile flow drainage unless the temperature of the soil layer is below 0°C. When the soil layer is frozen, no water movement is calculated.

Table 8-1: SWAT input variables used in percolation calculations.

Variable name	Definition	File Name
CLAY	m_c: Percent clay content	.sol
SOL_BD	ρ_b: Bulk density (Mg m^{-3})	.sol
SOL_AWC	AWC_{ly}: available water capacity	.sol

8.2 PERCOLATION

Percolation is calculated for each soil layer in the profile. Water is allowed to percolate if the water content exceeds the field capacity water content for that layer. When the soil layer is frozen, no water flow out of the layer is calculated.

The volume of water available for percolation in the soil layer is calculated:

$$SW_{ly,excess} = SW_{ly} - FC_{ly} \quad \text{if} \quad SW_{ly} > FC_{ly} \qquad 8.2.1$$

$$SW_{ly,excess} = 0 \quad \text{if} \quad SW_{ly} \leq FC_{ly} \qquad 8.2.2$$

where $SW_{ly,excess}$ is the drainable volume of water in the soil layer on a given day (mm H_2O), SW_{ly} is the water content of the soil layer on a given day (mm H_2O) and FC_{ly} is the water content of the soil layer at field capacity (mm H_2O).

The amount of water that moves from one layer to the underlying layer is calculated using storage routing methodology. The equation used to calculate the amount of water that percolates to the next layer is:

$$w_{perc,ly} = SW_{ly,excess} \cdot \left(1 - \exp\left[\frac{-\Delta t}{TT_{perc}}\right]\right) \qquad 8.2.3$$

where $w_{perc,ly}$ is the amount of water percolating to the underlying soil layer on a given day (mm H_2O), $SW_{ly,excess}$ is the drainable volume of water in the soil layer on a given day (mm H_2O), Δt is the length of the time step (hrs), and TT_{perc} is the travel time for percolation (hrs).

The travel time for percolation is unique for each layer. It is calculate

$$TT_{perc} = \frac{SAT_{ly} - FC_{ly}}{K_{sat}} \qquad 8.2.4$$

where TT_{perc} is the travel time for percolation (hrs), SAT_{ly} is the amount of water in the soil layer when completely saturated (mm H_2O), FC_{ly} is the water content of the soil layer at field capacity (mm H_2O), and K_{sat} is the saturated hydraulic conductivity for the layer (mm·h^{-1}).

Water that percolates out of the lowest soil layer enters the vadose zone. The vadose zone is the unsaturated zone between the bottom of the soil profile and the top of the aquifer. Movement of water through the vadose zone and into the aquifers is reviewed in Chapter 9.

Table 8-2: SWAT input variables used in percolation calculations.

Variable name	Definition	File Name
SOL_K	K_{sat}: Saturated hydraulic conductivity (mm/hr)	.sol

8.3 BYPASS FLOW

One of the most unique soil orders is the Vertisols. These soils are characterized by a propensity to shrink when dried and swell when moistened. When the soil is dry, large cracks form at the soil surface. This behavior is a result of the type of soil material present and the climate. Vertisols contain at least 30% clay with the clay fraction dominated by smectitic mineralogy and occur in areas with cyclical wet and dry periods.

Vertisols are found worldwide (Figure 8-1). They have a number of local names, some of which are listed in Table 8-3.

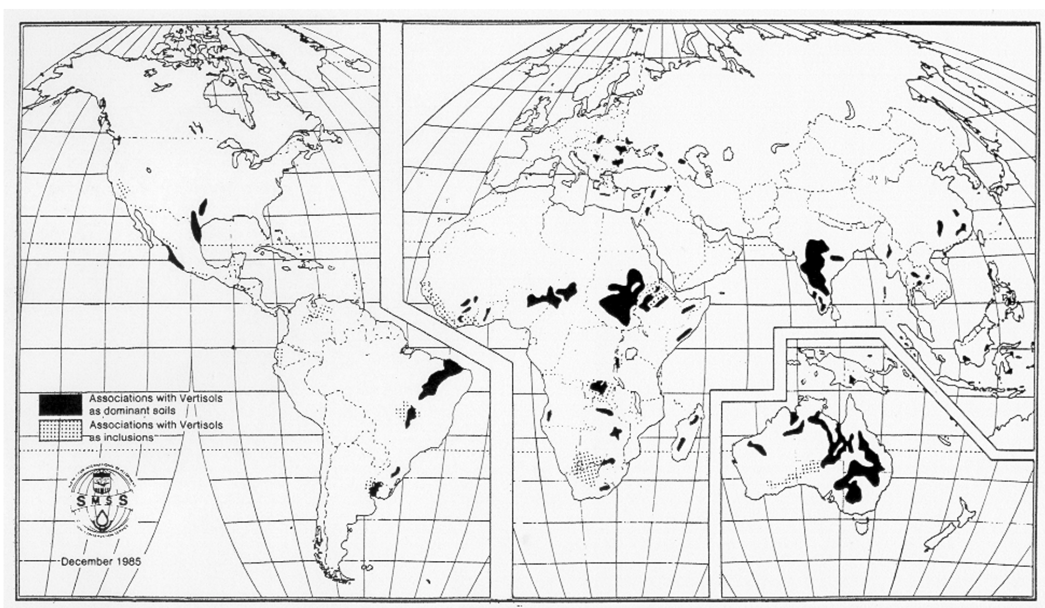

Figure 8-1: Soil associations of Vertisols (From Dudal and Eswaran, 1988)

Table 8-3: Alternative names for Vertisols or soils with Vertic properties (Dudal and Eswaran, 1988).

Names	Countries
Names that include the word "black"	
Barros pretos	Portugal
Black clays	South Africa, Australia
Black cotton soils	Africa, India
Black cracking clays	Uganda
Black earths	Australia, Africa
Black turf soils	South Africa
Dark clay soils	United States
Subtropical black clays	Africa
Sols noirs tropicaux	Africa
Terra nera	Italy
Terres noires tropicales	Africa
Terras negras tropicais	Mozambique
Tierras negras de Andalucia	Spain
Tropical black earths	Angola, Ghana
Tropical black clays	Africa
Names that reflect the black color	
Karail	India
Melanites	Ghana
Teen Suda	Sudan
Tropical Chernozems	Africa, India
Impact Chernozems	Russia
Vernacular names	
Adobe soils	United States, Philippines
Badobes	Sudan
Dian Pere	French West Africa
Gilgai soils	Australia
Firki	Nigeria
Mbuga	Tanzania
Kahamba	Congo
Makande	Malawi
Morogan	Romania
Regur	India
Rendzina	United States
Shachiang soils	China
Smolnitza	Bulgaria, Romania
Smonitza	Austria, Yugoslavia
Sols de paluds	France
Tirs	Morocco, North Africa
Vlei grond	South Africa
Sonsocuite	Nicaragua
Coined names	
Densinigra soils	Angola
Gravinigra soils	Angola
Grumusols	United States
Margalite soils	Indonesia
Vertisols	United States

One criteria used to classify a soil as a Vertisol is the formation of shrinkage cracks in the dry season that penetrate to a depth of more than 50 cm and are at least 1 cm wide at 50 cm depth. The cracks can be considerably wider at the surface—30 cm cracks at the surface are not unusual although 6-15 cm cracks are more typical.

To accurately predict surface runoff and infiltration in areas dominated by soils that exhibit Vertic properties, the temporal change in soil volume must be quantified. Bouma and Loveday (1988) identified three soil moisture conditions for which infiltration needs to be defined (Figure 8-2).

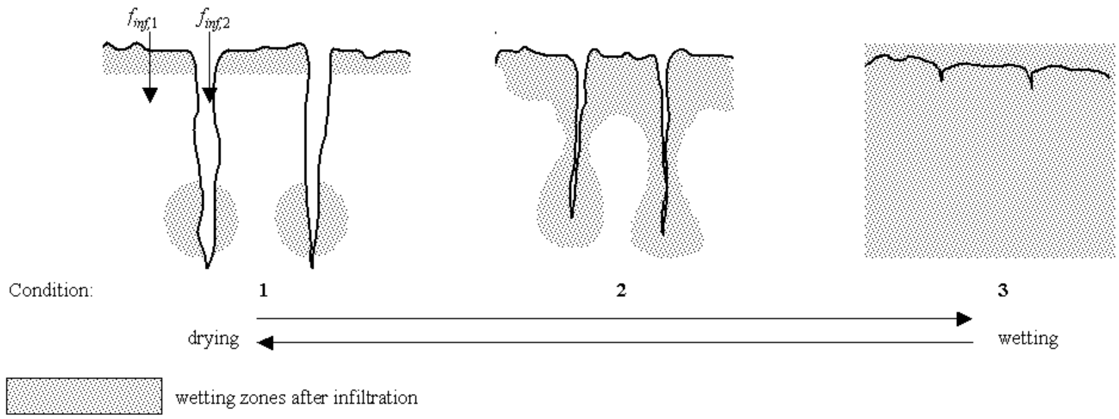

Figure 8-2: Diagram showing the effect of wetting and drying on cracking in Vertisols (from Bouma and Loveday, 1988)

Traditional models of infiltration are applicable to soils in which cracks have been closed by swelling and the soil acts as a relatively homogenous porous medium (Condition 3 in Figure 8-2). Condition 1 in Figure 8-2 represents the driest state with cracks at maximum width, a condition present at the end of the dry season/beginning of the rainy season. Condition 2 in Figure 8-2 represents the crack development typical with an actively growing crop requiring multiple irrigation or rainfall events to sustain growth. Bypass flow, the vertical movement of free water along macropores through unsaturated soil horizons, will occur in conditions 1 and 2. Bypass flow ($f_{inf,2}$ in Figure 8-2) occurs when the rate of rainfall or irrigation exceeds the vertical infiltration rate into the soil peds ($f_{inf,1}$ in Figure 8-2).

When bypass flow is modeled, SWAT calculates the crack volume of the soil matrix for each day of simulation by layer. On days in which precipitation events occur, infiltration and surface runoff is first calculated for the soil peds ($f_{inf,1}$ in Figure 8-2) using the curve number or Green & Ampt method. If any surface runoff is generated, it is allowed to enter the cracks. A volume of water equivalent to the total crack volume for the soil profile may enter the profile as bypass flow. Surface runoff in excess of the crack volume remains overland flow.

Water that enters the cracks fills the soil layers beginning with the lowest layer of crack development. After cracks in one layer are filled, the cracks in the overlying layer are allowed to fill.

The crack volume initially estimated for a layer is calculated:

$$crk_{ly,i} = crk_{max,ly} \cdot \frac{coef_{crk} \cdot FC_{ly} - SW_{ly}}{coef_{crk} \cdot FC_{ly}} \qquad 8.3.1$$

where $crk_{ly,i}$ is the initial crack volume calculated for the soil layer on a given day expressed as a depth (mm), $crk_{max,ly}$ is the maximum crack volume possible for the soil layer (mm), $coef_{crk}$ is an adjustment coefficient for crack flow, FC_{ly} is the water content of the soil layer at field capacity (mm H_2O), and SW_{ly} is the water content of the soil layer on a given day (mm H_2O). The adjustment coefficient for crack flow, $coef_{crk}$, is set to 0.10.

When the moisture content of the entire profile falls below 90% of the field capacity water content for the profile during the drying stage, the crack volume for a given day is a function of the crack volume estimated with equation 8.3.1 and the crack volume of the layer on the previous day. When the soil is wetting and/or when the moisture content of the profile is above 90% of the field capacity water content, the crack volume for a given day is equal to the volume calculated with equation 8.3.1.

$$crk_{ly} = \ell_{crk} \cdot crk_{ly,d-1} + (1.0 - \ell_{crk}) \cdot crk_{ly,i}$$

$$\text{when } SW < 0.90 \cdot FC \text{ and } crk_{ly,i} > crk_{ly,d-1} \qquad 8.3.2$$

$$crk_{ly} = crk_{ly,i}$$

$$\text{when } SW \geq 0.90 \cdot FC \text{ or } crk_{ly,i} \leq crk_{ly,d-1} \qquad 8.3.3$$

where crk_{ly} is the crack volume for the soil layer on a given day expressed as a depth (mm), ℓ_{crk} is the lag factor for crack development during drying, $crk_{ly,d\text{-}1}$ is the crack volume for the soil layer on the previous day (mm), $crk_{ly,i}$ is the initial crack volume calculated for the soil layer on a given day using equation 8.3.1

(mm), SW is the water content of the soil profile on a given day (mm H_2O), and FC is the water content of the soil profile at field capacity (mm H_2O).

As the tension at which water is held by the soil particles increases, the rate of water diffusion slows. Because the rate of water diffusion is analogous to the coefficient of consolidation in classical consolidation theory (Mitchell, 1992), the reduction in diffusion will affect crack formation. The lag factor is introduced during the drying stage to account for the change in moisture redistribution dynamics that occurs as the soil dries. The lag factor, ℓ_{crk}, is set to a value of 0.99.

The maximum crack volume for the layer, $crk_{max,ly}$, is calculated:

$$crk_{max,ly} = 0.916 \cdot crk_{max} \cdot \exp\left[-0.0012 \cdot z_{l,ly}\right] \cdot depth_{ly} \qquad 8.3.4$$

where $crk_{max,ly}$ is the maximum crack volume possible for the soil layer (mm), crk_{max} is the potential crack volume for the soil profile expressed as a fraction of the total volume, $z_{l,ly}$ is the depth from the soil surface to the bottom of the soil layer (mm), and $depth_{ly}$ is the depth of the soil layer (mm). The potential crack volume for the soil profile, crk_{max}, is input by the user. Those needing information on the measurement of this parameter are referred to Bronswijk (1989; 1990).

Once the crack volume for each layer is calculated, the total crack volume for the soil profile is determined.

$$crk = \sum_{ly=1}^{n} crk_{ly} \qquad 8.3.5$$

where crk is the total crack volume for the soil profile on a given day (mm), crk_{ly} is the crack volume for the soil layer on a given day expressed as a depth (mm), ly is the layer, and n is the number of layers in the soil profile.

After surface runoff is calculated for rainfall events using the curve number or Green & Ampt method, the amount of runoff is reduced by the volume of cracks present that day:

$$Q_{surf} = Q_{surf,i} - crk \qquad \text{if } Q_{surf,i} > crk \qquad 8.3.6$$

$$Q_{surf} = 0 \qquad \text{if } Q_{surf,i} \leq crk \qquad 8.3.7$$

where Q_{surf} is the accumulated runoff or rainfall excess for the day (mm H_2O), $Q_{surf,i}$ is the initial accumulated runoff or rainfall excess determined with the Green & Ampt or curve number method (mm H_2O), and crk is the total crack volume for the soil profile on a given day (mm). The total amount of water entering the soil is then calculated:

$$w_{inf} = R_{day} - Q_{surf} \qquad 8.3.8$$

where w_{inf} is the amount of water entering the soil profile on a given day (mm H_2O), R_{day} is the rainfall depth for the day adjusted for canopy interception (mm H_2O), and Q_{surf} is the accumulated runoff or rainfall excess for the day (mm H_2O).

Bypass flow past the bottom of the profile is calculated:

$$w_{crk,btm} = 0.5 \cdot crk \cdot \left(\frac{crk_{ly=nn}}{depth_{ly=nn}} \right) \qquad 8.3.9$$

where $w_{crk,btm}$ is the amount of water flow past the lower boundary of the soil profile due to bypass flow (mm H_2O), crk is the total crack volume for the soil profile on a given day (mm), $crk_{ly=nn}$ is the crack volume for the deepest soil layer ($ly=nn$) on a given day expressed as a depth (mm), and $depth_{ly=nn}$ is the depth of the deepest soil layer ($ly=nn$) (mm).

After $w_{crk,btm}$ is calculated, each soil layer is filled to field capacity water content beginning with the lowest layer and moving upward until the total amount of water entering the soil, w_{inf}, has been accounted for.

Table 8-4: SWAT input variables used in bypass flow calculations.

Variable name	Definition	File Name
ICRK	Bypass flow code: 0-do not model bypass flow; 1-model bypass flow	.cod
SOL_CRK	crk_{max}: Potential crack volume for soil profile	.sol

8.4 LATERAL FLOW

Lateral flow will be significant in areas with soils having high hydraulic conductivities in surface layers and an impermeable or semipermeable layer at a shallow depth. In such a system, rainfall will percolate vertically until it encounters the impermeable layer. The water then ponds above the impermeable layer forming a saturated zone of water, i.e. a perched water table. This saturated zone is the source of water for lateral subsurface flow.

SWAT incorporates a kinematic storage model for subsurface flow developed by Sloan et al. (1983) and summarized by Sloan and Moore (1984). This model simulates subsurface flow in a two-dimensional cross-section along a flow path down a steep hillslope. The kinematic approximation was used in its derivation.

This model is based on the mass continuity equation, or mass water balance, with the entire hillslope segment used as the control volume. The hillslope segment has a permeable soil surface layer of depth D_{perm} and length L_{hill} with an impermeable soil layer or boundary below it as shown in Figure 8-3. The hillslope segment is oriented at an angle α_{hill} to the horizontal.

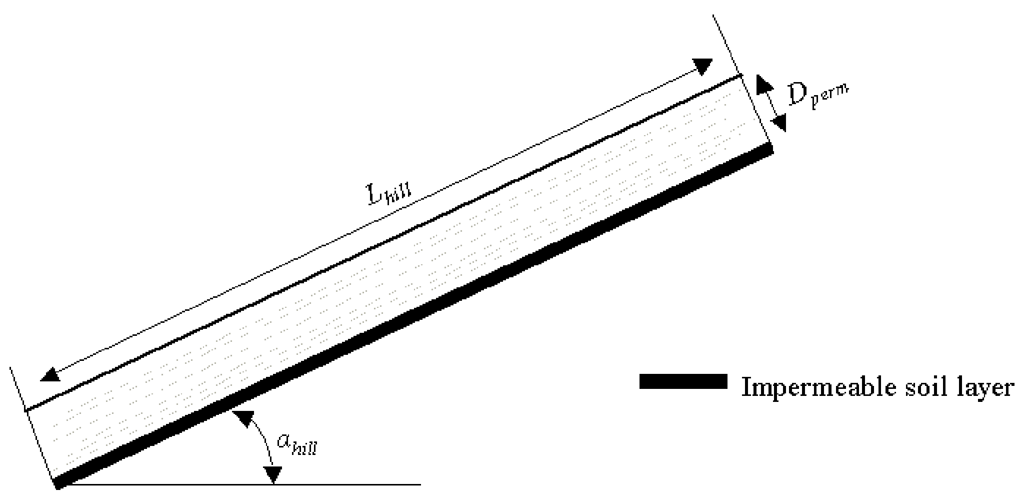

Figure 8-3: Conceptual representation of the hillslope segment.

The kinematic wave approximation of saturated subsurface or lateral flow assumes that the lines of flow in the saturated zone are parallel to the impermeable boundary and the hydraulic gradient equals the slope of the bed.

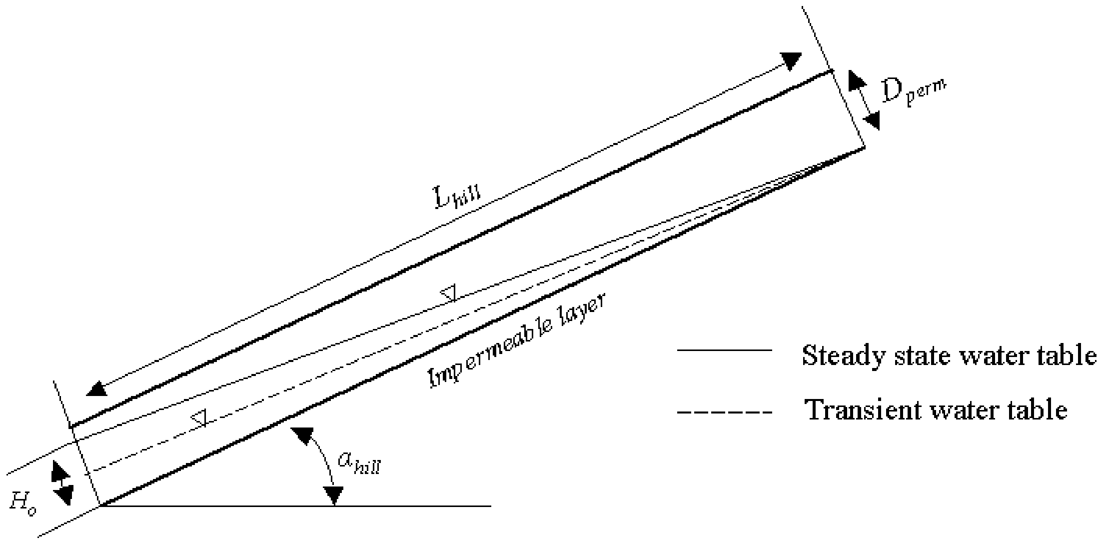

Figure 8-4: Behavior of the water table as assumed in the kinematic storage model.

From Figure 8-4, the drainable volume of water stored in the saturated zone of the hillslope segment per unit area, $SW_{ly,excess}$, is

$$SW_{ly,excess} = \frac{1000 \cdot H_o \cdot \phi_d \cdot L_{hill}}{2} \qquad 8.4.1$$

where $SW_{ly,excess}$ is the drainable volume of water stored in the saturated zone of the hillslope per unit area (mm H_2O), H_o is the saturated thickness normal to the hillslope at the outlet expressed as a fraction of the total thickness (mm/mm), ϕ_d is the drainable porosity of the soil (mm/mm), L_{hill} is the hillslope length (m), and 1000 is a factor needed to convert meters to millimeters. This equation can be rearranged to solve for H_o:

$$H_o = \frac{2 \cdot SW_{ly,excess}}{1000 \cdot \phi_d \cdot L_{hill}} \qquad 8.4.2$$

The drainable porosity of the soil layer is calculated:

$$\phi_d = \phi_{soil} - \phi_{fc} \qquad 8.4.3$$

where ϕ_d is the drainable porosity of the soil layer (mm/mm), ϕ_{soil} is the total porosity of the soil layer (mm/mm), and ϕ_{fc} is the porosity of the soil layer filled with water when the layer is at field capacity water content (mm/mm).

A soil layer is considered to be saturated whenever the water content of the layer exceeds the layer's field capacity water content. The drainable volume of water stored in the saturated layer is calculated:

$$SW_{ly,excess} = SW_{ly} - FC_{ly} \quad \text{if} \quad SW_{ly} > FC_{ly} \qquad 8.4.4$$

$$SW_{ly,excess} = 0 \quad \text{if} \quad SW_{ly} \leq FC_{ly} \qquad 8.4.5$$

where SW_{ly} is the water content of the soil layer on a given day (mm H_2O) and FC_{ly} is the water content of the soil layer at field capacity (mm H_2O).

The net discharge at the hillslope outlet, Q_{lat}, is given by

$$Q_{lat} = 24 \cdot H_o \cdot v_{lat} \qquad 8.4.6$$

where Q_{lat} is the water discharged from the hillslope outlet (mm H_2O/day), H_o is the saturated thickness normal to the hillslope at the outlet expressed as a fraction of the total thickness (mm/mm), v_{lat} is the velocity of flow at the outlet (mm·h^{-1}), and 24 is a factor to convert hours to days.

Velocity of flow at the outlet is defined as

$$v_{lat} = K_{sat} \cdot \sin(\alpha_{hill}) \qquad 8.4.7$$

where K_{sat} is the saturated hydraulic conductivity (mm·h^{-1}) and α_{hill} is the slope of the hillslope segment. The slope is input to SWAT as the increase in elevation per unit distance (*slp*) which is equivalent to $\tan(\alpha_{hill})$. Because $\tan(\alpha_{hill}) \cong \sin(\alpha_{hill})$, equation 8.4.3 is modified to use the value for the slope as input to the model:

$$v_{lat} = K_{sat} \cdot \tan(\alpha_{hill}) = K_{sat} \cdot slp \qquad 8.4.8$$

Combining equations 8.4.2 and 8.4.8 with equation 8.4.6 yields the equation

$$Q_{lat} = 0.024 \cdot \left(\frac{2 \cdot SW_{ly,excess} \cdot K_{sat} \cdot slp}{\phi_d \cdot L_{hill}} \right) \qquad 8.4.9$$

where all terms are previously defined.

8.4.1 LATERAL FLOW LAG

In large subbasins with a time of concentration greater than 1 day, only a portion of the lateral flow will reach the main channel on the day it is generated. SWAT incorporates a lateral flow storage feature to lag a portion of lateral flow release to the main channel.

Once lateral flow is calculated, the amount of lateral flow released to the main channel is calculated:

$$Q_{lat} = \left(Q'_{lat} + Q_{latstor,i-1}\right) \cdot \left(1 - \exp\left[\frac{-1}{TT_{lat}}\right]\right) \qquad 8.4.10$$

where Q_{lat} is the amount of lateral flow discharged to the main channel on a given day (mm H_2O), Q'_{lat} is the amount of lateral flow generated in the subbasin on a given day (mm H_2O), $Q_{latstor,i-1}$ is the lateral flow stored or lagged from the previous day (mm H_2O), and TT_{lag} is the lateral flow travel time (days).

The model will calculate lateral flow travel time or utilize a user-defined travel time. In the majority of cases, the user should allow the model to calculate the travel time. If drainage tiles are present in the HRU, lateral flow travel time is calculated:

$$TT_{lag} = \frac{tile_{lag}}{24} \qquad 8.4.11$$

where TT_{lag} is the lateral flow travel time (days) and $tile_{lag}$ is the drain tile lag time (hrs). In HRUs without drainage tiles, lateral flow travel time is calculated:

$$TT_{lag} = 10.4 \cdot \frac{L_{hill}}{K_{sat,mx}} \qquad 8.4.12$$

where TT_{lag} is the lateral flow travel time (days), L_{hill} is the hillslope length (m), and $K_{sat,mx}$ is the highest layer saturated hydraulic conductivity in the soil profile (mm/hr).

The expression $\left(1 - \exp\left[\frac{-1}{TT_{lag}}\right]\right)$ in equation 8.4.10 represents the fraction of the total available water that will be allowed to enter the reach on any one day. Figure 8-5 plots values for this expression at different values of TT_{lag}.

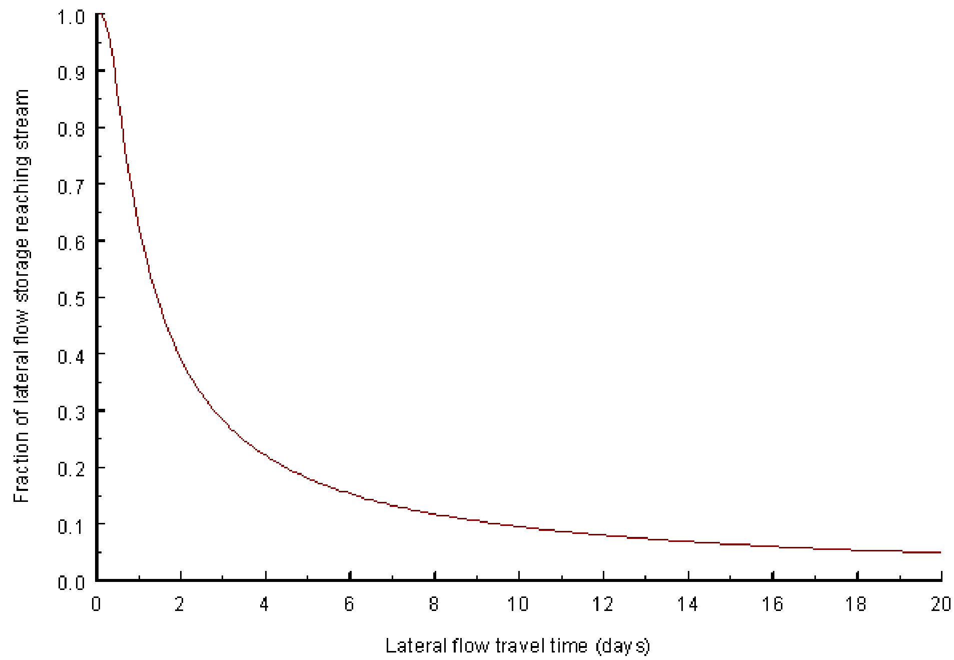

Figure 8-5: Influence of TT_{lag} on fraction of lateral flow released.

The delay in release of lateral flow will smooth the streamflow hydrograph simulated in the reach.

Table 8-5: SWAT input variables used in lateral flow calculations.

Variable name	Definition	File Name
SLSOIL	L_{hill}: Hillslope length (m)	.hru
SOL_K	K_{sat}: Saturated hydraulic conductivity (mm/hr)	.sol
SLOPE	*slp*: Average slope of the subbasin (m/m)	.hru
LAT_TTIME	TT_{lag}: Lateral flow travel time (days)	.hru
GDRAIN	$tile_{lag}$: Drain tile lag time (hrs)	.hru

8.5 NOMENCLATURE

AWC Available water capacity (fraction or mm H_2O)
AWC_{ly} Available water capacity of soil layer (fraction or mm H_2O)
FC Water content of soil profile at field capacity (fraction or mm H_2O)
FC_{ly} Water content of layer *ly* at field capacity (fraction or mm H_2O)
H_o Saturated thickness normal to the hillslope at the outlet expressed as a fraction of the total thickness (mm/mm)
K_{sat} Saturated hydraulic conductivity (mm/hr)
L_{hill} Hillslope length (m)
M_S Mass of the solids (Mg)
Q_{lat} Lateral flow; water discharged from the hillslope outlet (mm H_2O/day)
$Q_{latstor,i-1}$ Lateral flow stored or lagged from the previous day (mm H_2O)
Q_{surf} Accumulated runoff or rainfall excess (mm H_2O)
R_{day} Amount of rainfall on a given day (mm H_2O)
SAT_{ly} Amount of water in the soil layer when completely saturated (mm H_2O)
SW Amount of water in soil profile (mm H_2O)
SW_{ly} Soil water content of layer *ly* (mm H_2O)
$SW_{ly,excess}$ Drainable volume of water stored layer (mm H_2O)
TT_{lag} Lateral flow travel time (days)
TT_{perc} Travel time for percolation (hrs)
V_A Volume of air (m^3)
V_S Volume of solids (m^3)
V_T Total soil volume (m^3)
V_W Volume of water (m^3)
WP Water content at wilting point (fraction or mm H_2O)
WP_{ly} Water content of the soil layer at wilting point (fraction or mm H_2O)

$coef_{crk}$ Adjustment coefficient for crack flow
crk Total crack volume for the soil profile on a given day (mm)
crk_{ly} Crack volume for the soil layer on a given day expressed as a depth (mm)
$crk_{ly,d-1}$ Crack volume for the soil layer on the previous day (mm)
$crk_{ly,i}$ Initial crack volume calculated for the soil layer on a given day expressed as a depth (mm)
crk_{max} Potential crack volume for the soil profile expressed as a fraction of the total volume
$crk_{max,ly}$ Maximum crack volume possible for the soil layer (mm)
$depth_{ly}$ Depth of the soil layer (mm)
m_c Percent clay content
slp Average slope of the subbasin (% or m/m)
$tile_{lag}$ Drain tile lag time (hrs).
v_{lat} *V*elocity of flow at the hillslope outlet ($mm \cdot h^{-1}$)
$w_{crk,btm}$ Amount of water flow past the lower boundary of the soil profile due to bypass flow (mm H_2O)
w_{inf} Amount of water entering the soil profile on a given day (mm H_2O)

$w_{perc,ly}$ Amount of water percolating to the underlying soil layer on a given day (mm H_2O)
$z_{l,ly}$ Depth from the surface to the bottom of the soil layer (mm)

α_{hill} Slope of the hillslope segment (degrees)
Δt Length of the time step (hrs)
ℓ_{crk} Lag factor for crack development during drying
ρ_b Bulk density (Mg m^{-3})
ρ_s Particle density (Mg m^{-3})
ϕ_d Drainable porosity of the soil (mm/mm)
ϕ_{fc} Porosity of the soil layer filled with water when the layer is at field capacity water content (mm/mm)
ϕ_{soil} Porosity of the soil (mm/mm)

8.6 REFERENCES

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CHAPTER 9

EQUATIONS: GROUNDWATER

Groundwater is water in the saturated zone of earth materials under pressure greater than atmospheric, i.e. positive pressure. Water enters groundwater storage primarily by infiltration/percolation, although recharge by seepage from surface water bodies may occur. Water leaves groundwater storage primarily by discharge into rivers or lakes, but it is also possible for water to move upward from the water table into the capillary fringe.

9.1 GROUNDWATER SYSTEMS

Within the saturated zone of groundwater, regions of high conductivity and low conductivity will be found. The regions of high conductivity are made up of coarse-grained particles with a large percentage of macropores that allow water to move easily. The regions of low conductivity are made up of fine-grained particles with a large percentage of mesopores and micropores that restrict the rate of water movement.

An aquifer is "a geologic unit that can store enough water and transmit it at a rate fast enough to be hydrologically significant" (Dingman, 1994). An unconfined aquifer is an aquifer whose upper boundary is the water table. The water table is defined as the depth at which the water pressure equals atmospheric pressure. A confined aquifer is an aquifer bounded above and below by geologic formations whose hydraulic conductivity is significantly lower than that of the aquifer. Figure 9-1 illustrates the two types of aquifers.

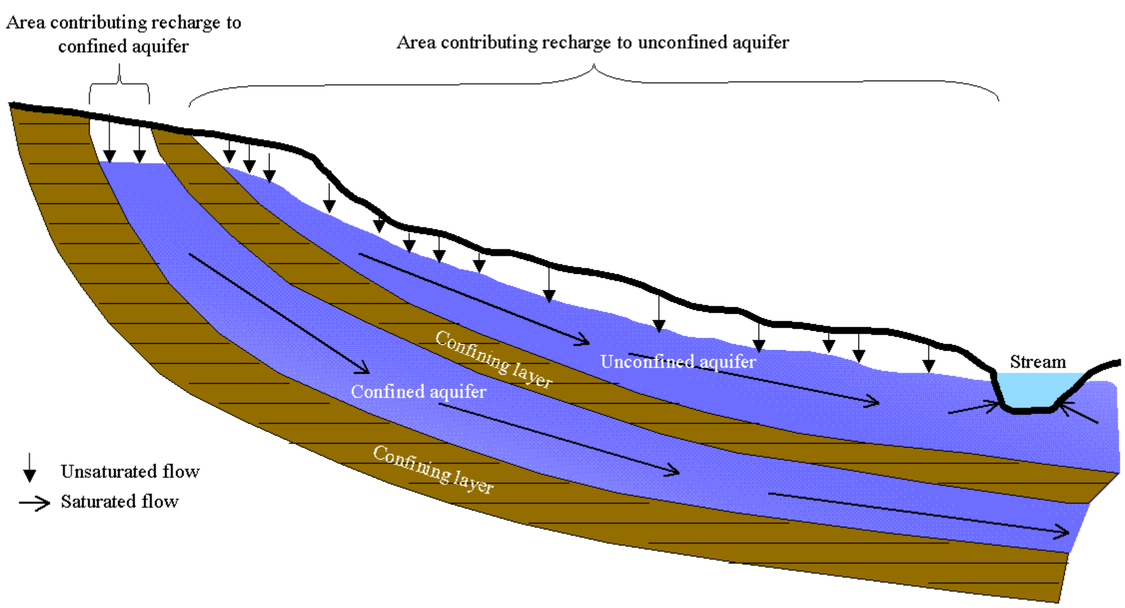

Figure 9-1: Unconfined and confined aquifers (from Dingman, 1994).

Recharge to unconfined aquifers occurs via percolation to the water table from a significant portion of the land surface. In contrast, recharge to confined aquifers by percolation from the surface occurs only at the upstream end of the confined aquifer, where the geologic formation containing the aquifer is exposed at the earth's surface, flow is not confined, and a water table is present.

Topography exerts an important influence on groundwater flow. The flow of groundwater in an idealized hilly upland area is depicted in Figure 9-2. The landscape can be divided into areas of recharge and areas of discharge. A recharge area is defined as a portion of a drainage basin where ground water flow is directed away from the water table. A discharge area is defined as a portion of the drainage basin where ground water flow is directed toward the water table. The water table is at or near the surface in discharge areas and surface water bodies are normally located in discharge areas.

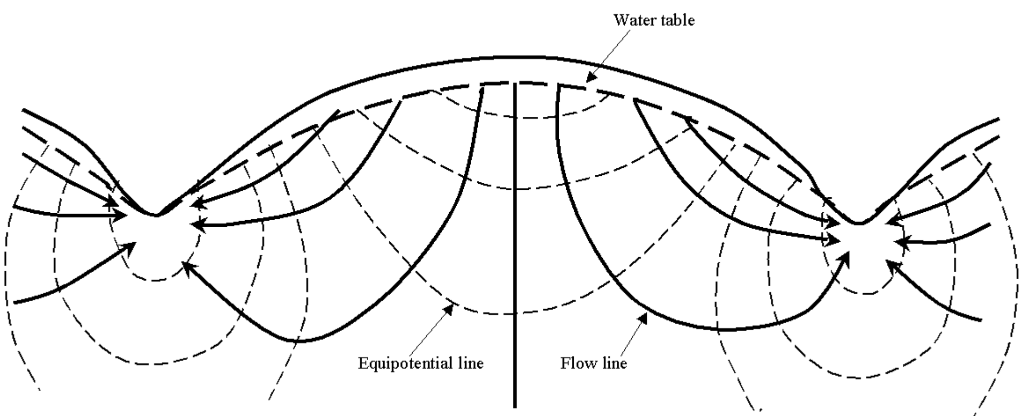

Figure 9-2: Groundwater flow net in an idealized hilly region with homogenous permeable material resting on an impermeable base (from Hubbert, 1940)

Streams may be categorized by their relationship to the groundwater system. A stream located in a discharge area that receives groundwater flow is a gaining or effluent stream (Figure 9-3a). This type of stream is characterized by an increase in discharge downstream. A stream located in a recharge area is a losing or influent stream. This type of stream is characterized by a decrease in discharge downstream. A losing stream may be connected to (Figure 9-3b) or perched above (Figure 9-3c) the groundwater flow area. A stream that simultaneously receives and loses groundwater is a flow-through stream (Figure 9-3d).

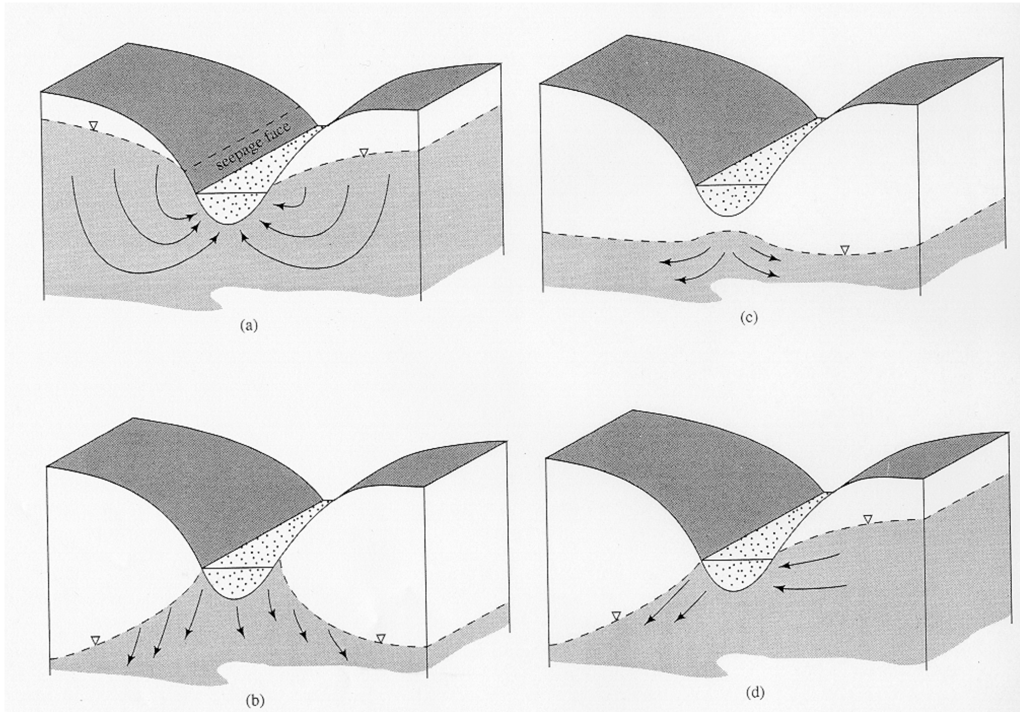

Figure 9-3: Stream-groundwater relationships: a) gaining stream receiving water from groundwater flow; b) losing stream connected to groundwater system; c) losing stream perched above groundwater system; and d) flow-through stream (from Dingman, 1994).

SWAT simulates two aquifers in each subbasin. The shallow aquifer is an unconfined aquifer that contributes to flow in the main channel or reach of the subbasin. The deep aquifer is a confined aquifer. Water that enters the deep aquifer is assumed to contribute to streamflow somewhere outside of the watershed (Arnold et al., 1993).

9.2 SHALLOW AQUIFER

The water balance for the shallow aquifer is:

$$aq_{sh,i} = aq_{sh,i-1} + w_{rchrg} - Q_{gw} - w_{revap} - w_{deep} - w_{pump,sh} \qquad 9.2.1$$

where $aq_{sh,i}$ is the amount of water stored in the shallow aquifer on day i (mm H_2O), $aq_{sh,i\text{-}1}$ is the amount of water stored in the shallow aquifer on day i-1 (mm H_2O), w_{rchrg} is the amount of recharge entering the aquifer on day i (mm H_2O), Q_{gw} is the groundwater flow, or base flow, into the main channel on day i (mm H_2O), w_{revap} is the amount of water moving into the soil zone in response to water deficiencies on day i (mm H_2O), w_{deep} is the amount of water percolating from the shallow aquifer into the deep aquifer on day i (mm H_2O), and $w_{pump,sh}$ is the amount of water removed from the shallow aquifer by pumping on day i (mm H_2O).

9.2.1 RECHARGE

Water that moves past the lowest depth of the soil profile by percolation or bypass flow enters and flows through the vadose zone before becoming shallow aquifer recharge. The lag between the time that water exits the soil profile and enters the shallow aquifer will depend on the depth to the water table and the hydraulic properties of the geologic formations in the vadose and groundwater zones.

An exponential decay weighting function proposed by Venetis (1969) and used by Sangrey et al. (1984) in a precipitation/groundwater response model is utilized in SWAT to account for the time delay in aquifer recharge once the water exits the soil profile. The delay function accommodates situations where the recharge from the soil zone to the aquifer is not instantaneous, i.e. 1 day or less.

The recharge to the aquifer on a given day is calculated:

$$w_{rchrg,i} = \left(1 - \exp\left[-1/\delta_{gw}\right]\right) \cdot w_{seep} + \exp\left[-1/\delta_{gw}\right] \cdot w_{rchrg,i-1} \qquad 9.2.2$$

where $w_{rchrg,i}$ is the amount of recharge entering the aquifer on day i (mm H_2O), δ_{gw} is the delay time or drainage time of the overlying geologic formations (days), w_{seep} is the total amount of water exiting the bottom of the soil profile on day i (mm H_2O), and $w_{rchrg,i-1}$ is the amount of recharge entering the aquifer on day i-1 (mm H_2O). The total amount of water exiting the bottom of the soil profile on day i is calculated:

$$w_{seep} = w_{perc,ly=n} + w_{crk,btm} \qquad 9.2.3$$

where w_{seep} is the total amount of water exiting the bottom of the soil profile on day i (mm H_2O), $w_{perc,ly=n}$ is the amount of water percolating out of the lowest layer, n, in the soil profile on day i (mm H_2O), and $w_{crk,btm}$ is the amount of water flow past the lower boundary of the soil profile due to bypass flow on day i (mm H_2O).

The delay time, δ_{gw}, cannot be directly measured. It can be estimated by simulating aquifer recharge using different values for δ_{gw} and comparing the simulated variations in water table level with observed values. Johnson (1977) developed a simple program to iteratively test and statistically evaluate different delay times for a watershed. Sangrey et al. (1984) noted that monitoring wells in the same area had similar values for δ_{gw}, so once a delay time value for a geomorphic area is defined, similar delay times can be used in adjoining watersheds within the same geomorphic province.

9.2.2 GROUNDWATER/BASE FLOW

The shallow aquifer contributes base flow to the main channel or reach within the subbasin. Base flow is allowed to enter the reach only if the amount of water stored in the shallow aquifer exceeds a threshold value specified by the user, $aq_{shthr,q}$.

The steady-state response of groundwater flow to recharge is (Hooghoudt, 1940):

$$Q_{gw} = \frac{8000 \cdot K_{sat}}{L_{gw}^{\ 2}} \cdot h_{wtbl} \qquad 9.2.4$$

where Q_{gw} is the groundwater flow, or base flow, into the main channel on day i (mm H_2O), K_{sat} is the hydraulic conductivity of the aquifer (mm/day), L_{gw} is the distance from the ridge or subbasin divide for the groundwater system to the main channel (m), and h_{wtbl} is the water table height (m).

Water table fluctuations due to non-steady-state response of groundwater flow to periodic recharge is calculated (Smedema and Rycroft, 1983):

$$\frac{dh_{wtbl}}{dt} = \frac{w_{rchrg} - Q_{gw}}{800 \cdot \mu} \qquad 9.2.5$$

where $\frac{dh_{wtbl}}{dt}$ is the change in water table height with time (mm/day), w_{rchrg} is the amount of recharge entering the aquifer on day i (mm H_2O), Q_{gw} is the groundwater flow into the main channel on day i (mm H_2O), and μ is the specific yield of the shallow aquifer (m/m).

Assuming that variation in groundwater flow is linearly related to the rate of change in water table height, equations 9.2.5 and 9.2.4 can be combined to obtain:

$$\frac{dQ_{gw}}{dt} = 10 \cdot \frac{K_{sat}}{\mu \cdot L_{gw}^{2}} \cdot \left(w_{rchrg} - Q_{gw}\right) = \alpha_{gw} \cdot \left(w_{rchrg} - Q_{gw}\right) \qquad 9.2.6$$

where Q_{gw} is the groundwater flow into the main channel on day i (mm H_2O), K_{sat} is the hydraulic conductivity of the aquifer (mm/day), μ is the specific yield of the shallow aquifer (m/m), L_{gw} is the distance from the ridge or subbasin divide for the groundwater system to the main channel (m), w_{rchrg} is the amount of recharge entering the aquifer on day i (mm H_2O) and α_{gw} is the baseflow recession constant or constant of proportionality. Integration of equation 9.2.6 and rearranging to solve for Q_{gw} yields:

$$Q_{gw,i} = Q_{gw,i-1} \cdot \exp\left[-\alpha_{gw} \cdot \Delta t\right] + w_{rchrg} \cdot \left(1 - \exp\left[-\alpha_{gw} \cdot \Delta t\right]\right) \qquad 9.2.7$$

where $Q_{gw,i}$ is the groundwater flow into the main channel on day i (mm H_2O), $Q_{gw,i\text{-}1}$ is the groundwater flow into the main channel on day i-1 (mm H_2O), α_{gw} is the baseflow recession constant, Δt is the time step (1 day), and w_{rchrg} is the amount of recharge entering the aquifer on day i (mm H_2O).

The baseflow recession constant, α_{gw}, is a direct index of groundwater flow response to changes in recharge (Smedema and Rycroft, 1983). Values vary from 0.1-0.3 for land with slow response to recharge to 0.9-1.0 for land with a rapid response. Although the baseflow recession constant may be calculated, the best estimates are obtained by analyzing measured streamflow during periods of no recharge in the watershed.

When the shallow aquifer receives no recharge, equation 9.2.7 simplifies to:

$$Q_{gw} = Q_{gw,0} \cdot \exp\left[-\alpha_{gw} \cdot t\right] \qquad 9.2.8$$

where Q_{gw} is the groundwater flow into the main channel at time t (mm H_2O), $Q_{gw,0}$ is the groundwater flow into the main channel at the beginning of the recession (time t=0) (mm H_2O), α_{gw} is the baseflow recession constant, and t is the time lapsed since the beginning of the recession (days). The baseflow recession constant is measured by rearranging equation 9.2.8.

$$\alpha_{gw} = \frac{1}{N} \cdot \ln\left[\frac{Q_{gw,N}}{Q_{gw,0}}\right] \qquad 9.2.9$$

where α_{gw} is the baseflow recession constant, N is the time lapsed since the start of the recession (days), $Q_{gw,N}$ is the groundwater flow on day N (mm H_2O), $Q_{gw,0}$ is the groundwater flow at the start of the recession (mm H_2O).

It is common to find the baseflow days reported for a stream gage or watershed. This is the number of days for base flow recession to decline through one log cycle. When baseflow days are used, equation 9.2.9 can be further simplified:

$$\alpha_{gw} = \frac{1}{N} \cdot \ln\left[\frac{Q_{gw,N}}{Q_{gw,0}}\right] = \frac{1}{BFD} \cdot \ln[10] = \frac{2.3}{BFD} \qquad 9.2.10$$

where α_{gw} is the baseflow recession constant, and BFD is the number of baseflow days for the watershed.

9.2.3 REVAP

Water may move from the shallow aquifer into the overlying unsaturated zone. In periods when the material overlying the aquifer is dry, water in the capillary fringe that separates the saturated and unsaturated zones will evaporate and diffuse upward. As water is removed from the capillary fringe by evaporation, it is replaced by water from the underlying aquifer. Water may also be removed from the aquifer by deep-rooted plants which are able to uptake water directly from the aquifer.

SWAT models the movement of water into overlying unsaturated layers as a function of water demand for evapotranspiration. To avoid confusion with soil evaporation and transpiration, this process has been termed ‘revap’. This process is significant in watersheds where the saturated zone is not very far below the surface or where deep-rooted plants are growing. Because the type of plant cover will affect the importance of revap in the water balance, the parameters governing revap are usually varied by land use. Revap is allowed to occur only if the amount of water stored in the shallow aquifer exceeds a threshold value specified by the user, $aq_{shthr,rvp}$.

The maximum amount of water than will be removed from the aquifer via ‘revap’ on a given day is:

$$w_{revap,mx} = \beta_{rev} \cdot E_o \qquad 9.2.11$$

where $w_{revap,mx}$ is the maximum amount of water moving into the soil zone in response to water deficiencies (mm H_2O), β_{rev} is the revap coefficient, and E_o is the potential evapotranspiration for the day (mm H_2O). The actual amount of revap that will occur on a given day is calculated:

$$w_{revap} = 0 \qquad \text{if } aq_{sh} \leq aq_{shthr,rvp} \qquad 9.2.12$$

$$w_{revap} = w_{revap,mx} - aq_{shthr,rvp} \qquad \text{if } aq_{shthr,rvp} < aq_{sh} < \left(aq_{shthr,rvp} + w_{revap,mx}\right) \qquad 9.2.13$$

$$w_{revap} = w_{revap,mx} \qquad \text{if } aq_{sh} \geq \left(aq_{shthr,rvp} + w_{revap,mx}\right) \qquad 9.2.14$$

where w_{revap} is the actual amount of water moving into the soil zone in response to water deficiencies (mm H_2O), $w_{revap,mx}$ is the maximum amount of water moving into the soil zone in response to water deficiencies (mm H_2O), aq_{sh} is the amount of water stored in the shallow aquifer at the beginning of day i (mm H_2O) and $aq_{shthr,rvp}$ is the threshold water level in the shallow aquifer for revap or percolation to deep aquifer to occur (mm H_2O).

9.2.4 PERCOLATION TO DEEP AQUIFER

A fraction of the total daily recharge can be routed to the deep aquifer. Percolation to the deep aquifer is allowed to occur only if the amount of water stored in the shallow aquifer exceeds a threshold value specified by the user, $aq_{shthr,rvp}$.

The maximum amount of water than will be removed from the shallow aquifer via percolation to the deep aquifer on a given day is:

$$w_{deep,mx} = \beta_{deep} \cdot w_{rchrg} \qquad 9.2.15$$

where $w_{deep,mx}$ is the maximum amount of water moving into the deep aquifer on day i (mm H_2O), β_{deep} is the aquifer percolation coefficient, and w_{rchrg} is the amount of recharge entering the aquifer on day i (mm H_2O). The actual amount of percolation to the deep aquifer that will occur on a given day is calculated:

$$w_{deep} = 0 \qquad \text{if } aq_{sh} \le aq_{shthr,rvp} \qquad 9.2.12$$

$$w_{deep} = w_{deep,mx} - aq_{shthr,rvp} \qquad \text{if } aq_{shthr,rvp} < aq_{sh} < \left(aq_{shthr,rvp} + w_{revap,mx}\right) \qquad 9.2.13$$

$$w_{deep} = w_{deep,mx} \qquad \text{if } aq_{sh} \ge \left(aq_{shthr,rvp} + w_{revap,mx}\right) \qquad 9.2.14$$

where w_{deep} is the actual amount of water moving into the deep aquifer on day i (mm H_2O), $w_{deep,mx}$ is the maximum amount of water moving into the deep aquifer on day i (mm H_2O), aq_{sh} is the amount of water stored in the shallow aquifer at the beginning of day i (mm H_2O) and $aq_{shthr,rvp}$ is the threshold water level in the shallow aquifer for revap or percolation to deep aquifer to occur (mm H_2O).

9.2.5 PUMPING

If the shallow aquifer is specified as the source of irrigation water or water removed for use outside the watershed, the model will allow an amount of water up to the total volume of the shallow aquifer to be removed on any given day. Detailed information on water management may be found in Chapter 21.

9.2.6 GROUNDWATER HEIGHT

Although SWAT does not currently print groundwater height in the output files, the water table height is updated daily by the model. Groundwater height is related to groundwater flow by equation 9.2.4.

$$Q_{gw} = \frac{8000 \cdot K_{sat}}{L_{gw}^{2}} \cdot h_{wtbl} = \frac{8000 \cdot \mu}{10} \cdot \frac{10 \cdot K_{sat}}{\mu \cdot L_{gw}^{2}} \cdot h_{wtbl} = 800 \cdot \mu \cdot \alpha_{gw} \cdot h_{wtbl} \qquad 9.2.15$$

where Q_{gw} is the groundwater flow into the main channel on day i (mm H_2O), K_{sat} is the hydraulic conductivity of the aquifer (mm/day), L_{gw} is the distance from the ridge or subbasin divide for the groundwater system to the main channel (m), h_{wtbl} is the water table height (m), μ is the specific yield of the shallow aquifer (m/m), and α_{gw} is the baseflow recession constant. Substituting this definition for Q_{gw} into equation 9.2.7 gives

$$h_{wtbl,i} = h_{wtbl,i-1} \cdot \exp\left[-\alpha_{gw} \cdot \Delta t\right] + \frac{w_{rchrg} \cdot \left(1 - \exp\left[-\alpha_{gw} \cdot \Delta t\right]\right)}{800 \cdot \mu \cdot \alpha_{gw}} \qquad 9.2.16$$

where $h_{wtbl,i}$ is the water table height on day i (m), $h_{wtbl,i-1}$ is the water table height on day i-1 (m), α_{gw} is the baseflow recession constant, Δt is the time step (1 day),

w_{rchrg} is the amount of recharge entering the aquifer on day i (mm H_2O), and μ is the specific yield of the shallow aquifer (m/m).

Table 9-1: SWAT input variables used in shallow aquifer calculations.

Variable name	Definition	File Name
GW_DELAY	δ_{gw}: Delay time for aquifer recharge (days)	.gw
GWQMN	$aq_{shthr,q}$: Threshold water level in shallow aquifer for base flow (mm H_2O)	.gw
ALPHA_BF	α_{gw}: Baseflow recession constant	.gw
REVAPMN	$aq_{shthr,rvp}$: Threshold water level in shallow aquifer for revap or percolation to deep aquifer (mm H_2O)	.gw
GW_REVAP	β_{rev}: Revap coefficient	.gw
RCHRG_DP	β_{deep}: Aquifer percolation coefficient	.gw
GW_SPYLD	μ: Specific yield of the shallow aquifer (m/m)	.gw

9.3 DEEP AQUIFER

The water balance for the deep aquifer is:

$$aq_{dp,i} = aq_{dp,i-1} + w_{deep} - w_{pump,dp} \qquad 9.3.1$$

where $aq_{dp,i}$ is the amount of water stored in the deep aquifer on day i (mm H_2O), $aq_{dp,i\text{-}1}$ is the amount of water stored in the deep aquifer on day i-1 (mm H_2O), w_{deep} is the amount of water percolating from the shallow aquifer into the deep aquifer on day i (mm H_2O), and $w_{pump,dp}$ is the amount of water removed from the deep aquifer by pumping on day i (mm H_2O). The amount of water percolating into the deep aquifer is calculated with the equations reviewed in section 9.2.4. If the deep aquifer is specified as the source of irrigation water or water removed for use outside the watershed, the model will allow an amount of water up to the total volume of the deep aquifer to be removed on any given day.

Water entering the deep aquifer is not considered in future water budget calculations and can be considered to be lost from the system.

9.4 NOMENCLATURE

BFD Number of baseflow days for the watershed
E_o Potential evapotranspiration (mm d^{-1})
K_{sat} Hydraulic conductivity of the aquifer (mm/day)
L_{gw} Distance from the ridge or subbasin divide for the groundwater system to the main channel (m)
N Time lapsed since the start of the recession (days)
Q_{gw} Groundwater flow, or base flow, into the main channel (mm H_2O)
$Q_{gw,0}$ Groundwater flow at the start of the recession (mm H_2O)
$Q_{gw,N}$ Groundwater flow on day N (mm H_2O)

aq_{dp} Amount of water stored in the deep aquifer (mm H_2O)
aq_{sh} Amount of water stored in the shallow aquifer (mm H_2O)
$aq_{shthr,q}$ Threshold water level in shallow aquifer for base flow (mm H_2O)
$aq_{shthr,rvp}$ Threshold water level in shallow aquifer for revap or percolation to deep aquifer (mm H_2O)
h_{wtbl} Water table height (m)
$w_{crk,btm}$ Amount of water flow past the lower boundary of the soil profile due to bypass flow (mm H_2O)
w_{deep} Amount of water percolating from the shallow aquifer into the deep aquifer (mm H_2O)
$w_{deep,mx}$ Maximum amount of water moving into the deep aquifer (mm H_2O)
$w_{pump,dp}$ Amount of water removed from the deep aquifer by pumping (mm H_2O)
$w_{pump,sh}$ Amount of water removed from the shallow aquifer by pumping (mm H_2O)
w_{rchrg} Amount of water entering the aquifer via recharge (mm H_2O)
w_{revap} Amount of water moving into the soil zone in response to water deficiencies (mm H_2O)
$w_{revap,mx}$ Maximum amount of water moving into the soil zone in response to water deficiencies on day i (mm H_2O)
w_{seep} Total amount of water exiting the bottom of the soil profile (mm H_2O)

α_{gw} Baseflow recession constant
β_{deep} Aquifer percolation coefficient
β_{rev} Revap coefficient
δ_{gw} Delay time or drainage time for aquifer recharge (days)
μ Specific yield of the shallow aquifer (m/m)

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