3 Atmosphere – Plant and Soil interaction

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3.1 Rainfall and snowfall

Precipitation and irrigation are the main incoming water fluxes. Irrigation will be discussed in chapter 10. For most model applications data of daily rainfall amounts will suffice. In such a case SWAP will distribute the daily rainfall amount equally over the day.

For studies with fast reacting components, e.g. runoff (Par. 4.1) or macro pore flow (Par. 6.5), actual rain intensities are important. In that case extra options are available to specify the mean rain intensity (cm d^{-1}) for each season or to give the duration of rainfall for each day. When the mean rainfall intensities are specified, the period of rainfall during a day is calculated by dividing the total amount of rainfall by the intensity. SWAP will schedule the rainfall at the beginning of a day.

Optionally the precipitation can be subdivided in rain and snowfall. With this option the snowfall accumulates in a snow pack, which will be discussed in Par. 3.2. The subdivision in rain and snow is based on the air temperature. Above 0.5 ^oC the precipitation is rain and below 0.5 ^oC the precipitation is snow. It is obvious that for this option the daily air temperatures are necessary.

3.2 The snowpack

In case of snowfall, the water accumulates in a snowpack. The water will be released by snowmelt, during which a large volume of water becomes available for runoff or infiltration into the soil.

To use SWAP for cold regions it is necessary to expand the model with snow and frost conditions. Numerous ways exist to do so, from a delay in precipitation till a complete water and energy balance of the snowpack. The more complex the method, the more data will be needed. The method implemented in SWAP requires just the daily weather data, which are usually available to the model.

With the option to calculate snow accumulation and snowmelt for each day, the water balance of the snowpack on the soil surface will be calculated. This balance consists of several fluxes and a storage change in the snow layer (Figure 8).

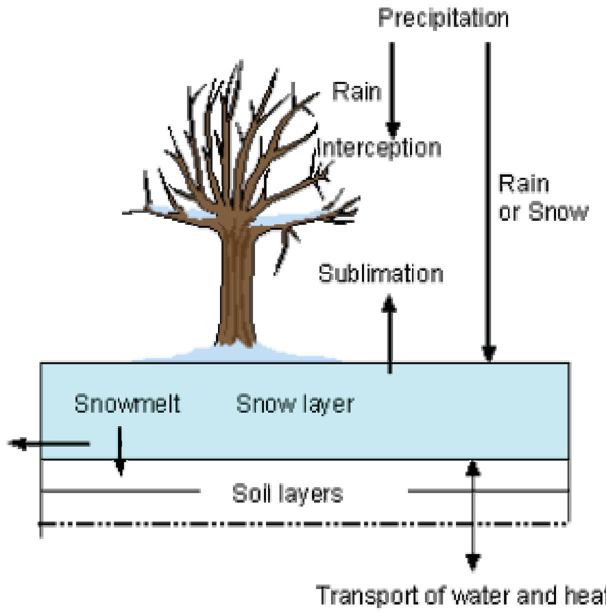

Figure 8 The water fluxes to and from the snow layer

The incoming fluxes are the rain and snowfall. The outgoing fluxes are the snowmelt and sublimation. The snowmelt q_{melt} (cm d^{-1}) is calculated when the air temperature rises above 0 ^{o}C (Kustas & Rango, 1994) with:

$$q_{melt} = C\,(T_{av} - T_s) \quad (3.1)$$

where C is a constant which can be specified by the user (d ^{o}C cm^{-1}) , T_{av} is the daily average air temperature (^{o}C) and T_s is the temperature of the snow (^{o}C). The assumption is made that the maximum snow temperature is 0 ^{o}C when the air temperature is above 0 ^{o}C.

In case of rainfall on the snow pack P_r (cm·d^{-1}) additional melt will occur due to heat released by splashing raindrops. This amount of snowmelt $q_{melt,r}$, (cm·d^{-1}) is calculated with (Fernández, 1998; Singh et al., 1997):

$$q_{melt,r} = \frac{P_r \cdot C_m \cdot (T_{av} - T_s)}{L_m} \quad (3.2)$$

where C_m is the specific heat of water (4180 J kg^{-1} k^{-1}) and L_m is the latent heat of melting (333580 J kg^{-1}). The melt fluxes leave the snow pack as runoff or infiltrate into the soil.

The snow can also evaporate directly into the air, a process called sublimation. The sublimation rate is taken equal to the potential evaporation rate (Par. 3.4). When a snow pack exists, the evapotranspiration from the soil and vegetation is set to zero.

The snow storage (S_{snow}) is calculated as the storage of the previous day plus the precipitation (P_r and P_s) minus the melt (q_{melt} and $q_{melt,r}$) and sublimation (E_s) amounts:

$$S^{t+1}_{snow} = S^{t}_{snow} + (P_r + P_s - q_{melt} - q_{melt,r} - E_s)\,\Delta t \quad (3.3)$$

Model input			
Variable	*Code*	*Description*	*Default*
S_{snow}	SNOWINCO	initial soil water equivalent (cm)	0.0
C	SNOWCOEF	snowmelt calibration factor (d °C cm^{-1})	0.2

3.3 Interception of rainfall

For the interception of rainfall two methods are available in SWAP, one for agricultural crops and one for trees and forests.

3.3.1 Agricultural crops

Von Hoyningen-Hüne (1983) and Braden (1985) measured interception of precipitation for various crops. They proposed the following general formula for canopy interception (Figure 9):

$$P_i = a \cdot LAI \left(1 - \frac{1}{1 + \frac{b \cdot P_{gross}}{a \cdot LAI}} \right) \tag{3.4}$$

where P_i is intercepted precipitation (cm d^{-1}), *LAI* is leaf area index, P_{gross} is gross precipitation (cm d^{-1}), *a* is an empirical coefficient (cm d^{-1}) and *b* is the soil cover fraction (≈ LAI/3.0) (-). For increasing precipitation amounts, the amount of intercepted precipitation asymptotically reaches the saturation amount *a LAI*. In principle *a* must be determined experimentally and should be specified in the input file. In case of ordinary agricultural crops we may, generally, assume *a* = 0.25 cm d^{-1}.

In case irrigation water is applied through sprinklers, total intercepted precipitation must be divided into a rain part and an irrigation part, as the solute concentration of both water sources may be different. Observed rainfall P_{gross} minus intercepted rainfall P_i is called net rainfall P_{net}. Likewise, applied irrigation depth I_{gross} minus intercepted irrigation water is called net irrigation depth I_{net}.

The method of Von Hoyningen-Hüne and Braden is based on daily precipitation values, so daily rainfall must be specified in the meteo input file. Additionally, rainfall may be specified in SWAP in smaller time steps. In this case the daily fraction P_{net}/P_{gross} is used to correct small time step rainfall for interception losses.

Model input			
Variable	*Code*	*Description*	*Default*
P_{gross}	RAIN	gross precipitation as a daily value (mm)	
a	COFAB	empirical coefficient Von Hoyningen-Hüne and Braden (cm d^{-1})	0.25
LAI	LAI	Leaf Area Index as a function of crop development stage (-)	

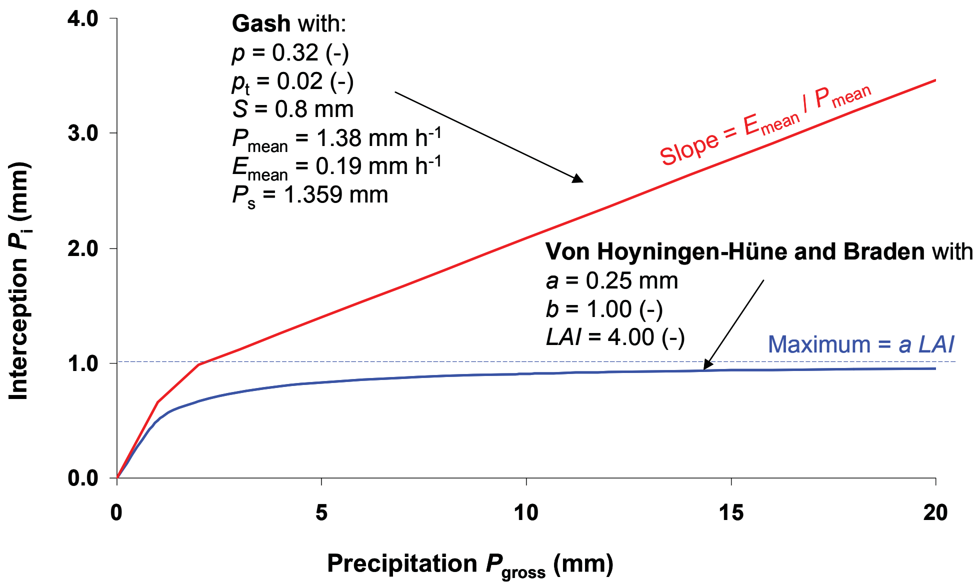

Figure 9 Interception for agricultural crops (Von Hoyningen-Hüne, 1983; Braden, 1985) and forests (Gash, 1979; 1985)

3.3.2 Forests

An important drawback of Eq. (3.4) is that the effect of rain duration and evaporation during the rain event is not explicitly taken into account. In case of interception by trees the effect of evaporation during rainfall can not be neglected. Gash (1979, 1985) formulated a physically based and widely used interception formula for forests. He considered rainfall to occur as a series of discrete events, each comprising a period of wetting up, a period of saturation and a period of drying out after rainfall ceases. The canopy is assumed to have sufficient time to dry out between storms. During wetting up, the increase of intercepted amount is described by:

$$\frac{\partial P_i}{\partial t} = (1 - p - p_t) P_{mean} - \frac{P_i}{S} E_{mean} \qquad (3.5)$$

where p is a free throughfall coefficient (-), p_t is the proportion of rainfall diverted to stemflow (-), P_{mean} is the mean rainfall rate (mm h^{-1}), E_{mean} is the mean evaporation rate of intercepted water when the canopy is saturated (mm h^{-1}) and S is the maximum storage of intercepted water in the canopy (mm). Integration of Eq. (3.5) yields the amount of rainfall which saturates the canopy, P_s (mm):

$$P_s = -\frac{P_{mean} S}{E_{mean}} \ln\left(1 - \frac{E_{mean}}{P_{mean}(1 - p - p_t)}\right) \quad \text{with} \quad 1 - \frac{E_{mean}}{P_{mean}(1 - p - p_t)} \geq 0 \qquad (3.6)$$

For small storms ($P_{gross} < P_s$) the interception can be calculated from:

$$P_i = (1 - p - p_t) P_{gross} \qquad (3.7)$$

For large storms ($P_{gross} > P_s$) the interception according to Gash (1979) follows from:

$$P_{\mathrm{i}} = \left(1 - p - p_{\mathrm{t}}\right) P_{\mathrm{s}} + \frac{E_{\mathrm{mean}}}{P_{\mathrm{mean}}} \left(P_{\mathrm{gross}} - P_{\mathrm{s}}\right) \tag{3.8}$$

Figure 9 shows the relation of Gash for typical values of a pine forest as function of rainfall amounts. The slope $\partial P_{\mathrm{i}} / \partial P_{\mathrm{gross}}$ before saturation of the canopy equals $(1 - p - p_{\mathrm{t}})$, after saturation of the canopy this slope equals $E_{\mathrm{mean}} / P_{\mathrm{mean}}$.

Model input

Variable	*Code*	*Description*	*Default*
P_{gross}	RAIN	gross precipitation as a daily value (mm)	
S	SCANOPY	storage capacity of the canopy (cm)	
p	PFREE	free throughfall coefficient (-)	
p_{t}	PSTEM	stemflow coefficient (-)	
P_{mean}	AVPREC	average rainfall intensity (cm d^{-1})	
E_{mean}	AVEVAP	average evaporation intensity during rainfall from a wet canopy (cm d^{-1})	

3.4 Potential evapotranspiration

Evapotranspiration covers both transpiration of the plants and evaporation of the soil or of water intercepted by vegetation or ponding on the soil surface. In the past, many empirical equations have been derived to calculate potential evapotranspiration which refers to evapotranspiration of cropped soils with an optimum water supply. These empirical equations are valid for the local conditions under which they were derived; they are hardly transferable to other areas. Nowadays, therefore, the focus is mainly on physically-based approaches, which have a wider applicability (Feddes and Lenselink, 1994).

For the process of evapotranspiration, three conditions in the soil-plant-atmosphere continuum must be met (Jensen et al., 1990):

a) A continuous supply of water;
b) Energy available to change liquid water into vapour;
c) A vapour pressure gradient to maintain a flux from the evaporating surface to the atmosphere.

The various methods of determining evapotranspiration are based on one or more of these requirements. For example, the soil water balance approach is based on (a), the energy balance approach on (b), and the combination method (energy balance plus heat and mass transfer) on parts of (b) and (c). Penman (1948) was the first to introduce the combination method. He estimated the evaporation from an open water surface, and then used that as a reference evaporation. Multiplied by a crop factor, this provided an estimate of the potential evapotranspiration from a cropped surface. The combination method requires measured climatic data on temperature, humidity, solar radiation and wind speed. Since the combination method retains a number of empirical relationships, numerous modifications to adjust it to local conditions have been proposed.

Analyzing a range of lysimeter data worldwide, Doorenbos and Pruitt (1977) proposed the FAO Modified Penman method, which has found worldwide application in irrigation and drainage projects. These authors adopted the same two-step approach as Penman to estimate crop water requirements (i.e. estimating a reference evapotranspiration, selecting crop coefficients per crop and per growth stage, and then multiplying the two to find the crop water requirements, in this way accounting for incomplete soil cover and different surface roughness). They replaced Penman's open water evaporation by the evapotranspiration from a reference crop. The reference crop of Doorenbos and Pruitt was defined as 'an extended surface of a tall green grass cover of uniform height (8 - 15 cm), actively growing, completely shading the ground, and not short of water'. There was evidence, however, that the method sometimes over-predicted the crop water requirements (Allen, 1991).

Using similar physics as Penman (1948), Monteith (1965) derived an equation that describes the evapotranspiration from a dry, extensive, horizontally-uniform vegetated surface, which is optimally supplied with water. This equation is known as the Penman-Monteith equation. Jensen et al. (1990) analyzed the performance of 20 different evapotranspiration formula against lysimeter data for 11 stations around the world under different climatic conditions. The Penman-Monteith formula ranked as the best for all climatic conditions. This equation has become an international standard for calculation of potential evapotranspiration.

Potential and even actual evapotranspiration estimates are possible with the Penman-Monteith equation, through the introduction of canopy and air resistances to water vapour diffusion. This direct, or one-step, approach is increasingly being followed nowadays, especially in research environments. Nevertheless, since accepted canopy and air resistances may not yet be available for many crops, a two-step approach is still recommended under field conditions. The first step is the calculation of the potential evapotranspiration, using the minimum value of the canopy resistance and the actual air resistance. In the second step the actual evapotranspiration is calculated using the root water uptake reduction due to water and/or salinity stress and evaporation reduction (Par. 2.2.3). This two-step approach is followed in SWAP.

3.4.1 Penman-Monteith equation

The original form of the Penman-Monteith equation can be written as (Monteith, 1965, 1981):

$$ET_p = \frac{\dfrac{\Delta_v}{\lambda_w}(R_n - G) + \dfrac{p_1 \rho_{air} C_{air}}{\lambda_w} \dfrac{e_{sat} - e_a}{r_{air}}}{\Delta_v + \gamma_{air}\left(1 + \dfrac{r_{crop}}{r_{air}}\right)} \tag{3.9}$$

where ET_p is the potential transpiration rate of the canopy (mm d^{-1}), Δ_v is the slope of the vapour pressure curve (kPa °C^{-1}), λ_w is the latent heat of vaporization (J kg^{-1}), R_n is the net radiation flux at the canopy surface (J m^{-2} d^{-1}), G is the soil heat flux (J m^{-2} d^{-1}), p_1 accounts for unit conversion (=86400 s d^{-1}), ρ_{air} is the air density (kg m^{-3}), C_{air} is the heat capacity of moist air (J kg^{-1} °C^{-1}), e_{sat} is the saturation vapour pressure (kPa), e_a is the actual vapour pressure (kPa), γ_{air} is the psychrometric constant (kPa °C^{-1}), r_{crop} is the crop resistance (s m^{-1}) and r_{air} is the aerodynamic resistance (s m^{-1}).

To facilitate analysis of the combination equation, an aerodynamic and radiation term are defined:

$$ET_p = ET_{rad} + ET_{aero} \tag{3.10}$$

where ET_p is potential transpiration rate of crop canopy (cm d^{-1}), ET_rad is the radiation term (cm d^{-1}) and ET_aero is the aerodynamic term (cm d^{-1}).

The radiation term equals:

$$ET_{rad} = \frac{\Delta_v (R_n - G)}{\lambda_w (\Delta_v + \gamma^*_{air})} \tag{3.11}$$

where the modified psychrometric constant (kPa °C^{-1}) is:

$$\gamma^*_{air} = \gamma_{air} \left(1 + \frac{r_{crop}}{r_{air}}\right) \tag{3.12}$$

The aerodynamic term equals:

$$ET_{aero} = \frac{p_1 \rho_{air} C_{air} (e_{sat} - e_a)}{\lambda_w (\Delta_v + \gamma^*_{air}) r_{air}} \tag{3.13}$$

Many meteorological stations provide mean daily values of air temperature T_air (°C), global solar radiation R_s (J m^{-2} d^{-1}), wind speed u_0 (m s^{-1}) and air humidity e_act (kPa). The Food and Agricultural Organisation of the UN has proposed a clearly defined and well established methodology to apply the Penman-Monteith equation using above 4 weather data. (Allen et al., 1998). This methodology is applied in SWAP and is described in Par. 3.4.1.1 and 3.4.1.2.

3.4.1.1 Radiation term

The net radiation R_n (J m^{-2} d^{-1}) is the difference between incoming and outgoing radiation of both short and long wavelengths. It is the balance between the energy adsorbed, reflected and emitted by the earth's surface:

$$R_\mathrm{n} = \left(1 - \alpha_\mathrm{r}\right) R_\mathrm{s} - R_\mathrm{nl} \tag{3.14}$$

where α_r is the reflection coefficient or albedo (-) and R_nl is the net longwave radiation (J m^{-2} d^{-1}). The albedo is highly variable for different surfaces and for the angle of incidence or slope of the ground surface. It may be as large as 0.95 for freshly fallen snow and as small as 0.05 for a wet bare soil. A green vegetation cover has an albedo of about 0.20-0.25 (De Bruin, 1998). SWAP will assume in case of a crop $\alpha_r = 0.23$, in case of bare soil $\alpha_r = 0.15$.

The earth emits longwave radiation, which increases with temperature and which is adsorbed by the atmosphere or lost into space. The longwave radiation received by the atmosphere increases its temperature and, as a consequence, the atmosphere radiates energy of its own. Part of this radiation finds its way back to the earth's surface. As the outgoing longwave radiation is almost always greater than the incoming longwave radiation, the net longwave radiation R_nl represents an energy loss. Allen et al. (1998) recommend the following formula for the net longwave radiation:

$$R_{\mathrm{nl}} = \sigma_{\mathrm{sb}} \left[\frac{T_{\max}^4 + T_{\min}^4}{2} \right] \left(0.34 - 0.14\sqrt{e_{\mathrm{act}}} \right) \left(0.1 + 0.9 N_{\mathrm{rel}} \right) \tag{3.15}$$

where σ_{sb} is the Stefan-Boltzmann constant (4.903 10^{-3} J K^{-4} m^{-2} d^{-1}), $T_{\min}$ and $T_{\max}$ are the minimum and maximum absolute temperatures during the day (K), respectively, e_{act} is the actual vapour pressure (kPa), and N_{rel} is the relative sunshien duration. The latter can be derived from the measured global solar radiation R_n and the extraterrestrial radiation R_a (J m^{-2} d^{-1}), which is received at the top of the Earth's atmosphere on a horizontal surface:

$$N_{\mathrm{rel}} = \frac{R_{\mathrm{s}}}{b\, R_{\mathrm{a}}} - a \tag{3.16}$$

where a and b are empirical coefficients which depend on the local climate. For international use Allen et al. (1998) recommend a = 0.25 and b = 0.50.

The extraterrestrial radiation R_a depends on the latitude and the day of the year. R_a is calculated with:

$$R_{\mathrm{a}} = \frac{G_{\mathrm{sc}}}{\pi} d_{\mathrm{r}} \left[\omega_{\mathrm{s}} \sin(\varphi) \sin(\delta) + \cos(\varphi) \cos(\delta) \sin(\omega_{\mathrm{s}}) \right] \tag{3.17}$$

where d_r is the inverse relative distance Earth-Sun (-), ω_s is the sunset hour angle (rad), φ is the latitude (rad) and δ is the solar declination (rad). The inverse relative distance Earth-Sun and the solar declination are given by:

$$d_{\mathrm{r}} = 1 + 0.033 \cos\left(\frac{2\pi}{365} J \right) \tag{3.18}$$

$$\delta = 0.409 \sin\left(\frac{2\pi}{365} J - 1.39 \right) \tag{3.19}$$

where J is the number of the day in the year (1-365 or 366, starting January 1). The sunset hour angle expresses the day length and is given by:

$$\omega_{\mathrm{s}} = \arccos\left[-\tan(\varphi) \tan(\delta) \right] \tag{3.20}$$

3.4.1.2 Aerodynamic term

Latent heat of vaporization, λ_w (J g^{-1}), depends on the air temperature T_{air} (°C) (Harrison, 1963):

$$\lambda_w = 2501 - 2.361 T_{air} \tag{3.21}$$

Saturation vapour pressure, e_{sat} (kPa), also can be calculated from air temperature (Tetens, 1930):

$$e_{sat} = 0.611 \exp\left(\frac{17.27\, T_{air}}{T_{air} + 237.3} \right) \tag{3.22}$$

The slope of the vapour pressure curve, Δ_v (kPa °C^{-1}), is calculated as (Murray, 1967):

$$\Delta_v = \frac{4098\; e_{sat}}{(T_{air} + 237.3)^2} \tag{3.23}$$

The psychrometric constant, γ_{air} (kPa °C^{-1}), follows from (Brunt, 1952):

$$\gamma_{\text{air}} = 0.00163 \frac{p_{\text{air}}}{\lambda_{\text{w}}} \tag{3.24}$$

with p_{air} the atmospheric pressure (kPa) at elevation z_0 (m), which is calculated from (Burman et al., 1987):

$$p_{\text{air}} = 101.3 \left(\frac{T_{\text{air,K}} - 0.0065\, z_0}{T_{\text{air,K}}} \right)^{5.256} \tag{3.25}$$

Employing the ideal gas law, the atmospheric density, ρ_a (g cm^{-3}), can be shown to depend on p and the virtual temperature T_{vir} (K):

$$\rho_{\text{air}} = 3.486\, 10^{-3} \frac{p_{\text{air}}}{T_{\text{vir}}} \tag{3.26}$$

where the virtual temperature is derived from:

$$T_{\text{vir}} = \frac{T_{\text{air,K}}}{1 - 0.378 \dfrac{e_{\text{act}}}{p_{\text{air}}}} \tag{3.27}$$

The heat capacity of moist air, C_{air} (J g^{-1} °C^{-1}), follows from:

$$C_{\text{air}} = 622 \frac{\gamma_{\text{air}} \lambda_{\text{w}}}{p_{\text{air}}} \tag{3.28}$$

Aerodynamic resistance

The aerodynamic resistance r_{air} depends on the wind speed profile and the roughness of the canopy and is calculated as (Allen et al., 1998):

$$r_{air} = \frac{\ln\left(\dfrac{z_m - d}{z_{om}} \right) \cdot \ln\left(\dfrac{z_h - d}{z_{oh}} \right)}{\kappa_{vk}^2 \cdot u} \tag{3.29}$$

where z_m is height of wind speed measurements (m), z_h is height of temperature and humidity measurements (m), d is zero plane displacement of wind profile (m), z_{om} is roughness parameter for momentum (m) and z_{oh} is roughness parameter for heat and vapour (m), κ_{vk} is von Karman constant = 0.41 (-), u is wind speed measurement at height z_m (m s^{-1}),

The parameters d, z_{om} and z_{oh} are defined as:

$$d = \tfrac{2}{3}\, h_{crop} \tag{3.30}$$

$$z_{om} = 0.123 \ h_{crop} \qquad (3.31)$$

$$z_{oh} = 0.1 \ z_{om} \qquad (3.32)$$

with h_{crop} the crop height (cm)

A default height of 2 m is assumed for wind speed measurements (z_m) and height of temperature and humidity measurements (z_h).

Meteorological stations generally provide 24 hour averages of wind speed measurements, according to international standards, at an altitude of 10 meter.

To calculate r_{air}, the average daytime wind (7.00 - 19.00 h) should be used. For ordinary conditions we assume (Smith, 1991) for the average daytime windspeed ($u_{0,day}$):

$$u_{0,day} = 1.33 \ u_0 \qquad (3.33)$$

where u_0 is the measured average wind speed over 24 hours (m s^{-1}).

When crop height (h_{crop}) reaches below or above measurement height ($z_{m,meas}$), the wind speed is corrected with the following assumptions:

- a uniform wind pattern at an altitude of 100 meter;
- wind speed measurements are carried out above grassland;
- a logarithmic wind profile is assumed;
- below 2 meter wind speed is assumed to be unchanged with respect to a value at an altitude of 2 meter; applying a logarithmic wind profile at low altitudes is not carried out due to the high variation below 2 meter.

These assumptions result in the following equation for wind speed correction:

$$u = \frac{\ln\left(\dfrac{z_{act} - d_{act}}{z_{om,act}}\right)}{\ln\left(\dfrac{z_{100} - d_{act}}{z_{om,act}}\right)} \ \frac{\ln\left(\dfrac{z_{100} - d_{grass}}{z_{om,grass}}\right)}{\ln\left(\dfrac{z_{m,meas} - d_{grass}}{z_{om,grass}}\right)} \ u_{0,day} \qquad (3.34)$$

where: u wind speed at crop height (m s^{-1}), z_{act} is the actual crop heigh with a minimum value of 2 m, d_{act} and d_{grass} are zero plane displacement of actual crop and grass (m), $z_{om,act}$ and $z_{om,grass}$ are roughness parameter for momentum actual crop and grass (m).

3.4.1.3 Fluxes above homogeneous surfaces

SWAP calculates three quantities with the Penman-Monteith equation (eq. (3.9)):

- ET_{w0} (cm d^{-1}), potential evapotranspiration rate of a wet canopy, completely covering the soil;
- ET_{p0} (cm d^{-1}), potential evapotranspiration rate of a dry canopy, completely covering the soil;

- E_{p0} (cm d^{-1}), potential evaporation rate of a wet, bare soil.

These quantities are obtained by varying the values for crop resistance, crop height and the reflection coefficient. In case of a wet canopy, the crop resistance r_{crop} is set to zero. In case of a dry crop with optimal water supply in the soil, r_{crop} is minimal and varies between 30 s m^{-1} for arable crop to 150 s m^{-1} for trees in a forest (Allen et al., 1986, 1989). In case of the bare wet soil, the program takes r_{crop} = 0 and 'crop height' h_{crop} = 0.1 cm. Reflection coefficient α_r in case of a (wet or dry) crop equals 0.23, while for a bare soil α_r = 0.15 is assumed.

Model input

Variable	*Code*	*Description*	*Default*
L_g	LAT	geographical latitude (degrees, North positive)	
z_0	ALT	altitude above mean sea level (m)	
$z_{m,meas}$	ALTW	altitude of wind speed measurement above mean soil surface (m)	
h_{crop}	CH	crop height as a function of crop development stage (cm)	
r_{crop}	RSC	minimum crop resistance (s m^{-1})	70
Daily (average 0-24 hrs) values of:			
$T_{air,min}$	TMIN	minimum air temperature at 2 m height (°C)	
$T_{air,max}$	TMAX	maximum air temperature at 2 m height (°C)	
R_s	RAD	global solar radiation (kJ m^{-2} d^{-1})	
u_0	WIND	wind speed at 2 m height (m s^{-1})	
e_{act}	HUM	air humidity as vapour pressure at 2 m height (kPa)	

3.4.2 Reference evapotranspiration and crop factors

Application of the Penman-Monteith equation requires daily values of air temperature, net radiation, wind speed and air humidity, which data might not be available. Also in some studies other methods than Penman-Monteith might be needed. For instance in The Netherlands the Makkink equation is widely used (Makkink, 1957; Feddes, 1987). Therefore SWAP allows the use of a reference potential evapotranspiration rate ET_{ref} (cm d^{-1}). In that case ET_{p0} is calculated by:

$$ET_{p0} = k_c ET_{ref} \tag{3.35}$$

where k_c is the so called crop factor, which depends on the crop type and the method employed to obtain ET_{ref}. The crop factor converts the reference evapotranspiration rate into the potential evapotranspiration rate of a dry canopy that completely covers the soil: k_c is thus taken to be constant from crop emergence up to maturity.

This approach, however, does not allow differentiation between a dry crop and wet crop. Therefore SWAP assumes: ET_{w0} = ET_{ref}. SWAP allows the use of a 'crop factor' to translate ET_{ref} into E_{p0}:

$$E_{p0} = k_{soil}\, ET_{ref} \tag{3.36}$$

If this option is not used, SWAP will assume ET_{p0} = ET_{ref}.

The reference evapotranspiration rate can be determined in several ways, such as pan evaporation, the Penman open water evaporation (Penman, 1948), the FAO modified Penman equation (Doorenbos and Pruitt, 1977), the Penman-Monteith equation applied for a reference crop (Allen et al., 1998), Priestly-Taylor (1972), Makkink (Makkink, 1957; Feddes, 1987) or Hargreaves et al. (1985). In order to transform all these reference evapotranspiration rates to the potential transpiration of the considered crop, the crop factors are needed.

Model input			
Variable	*Code*	*Description*	*Default*
k_c	CF	crop factor as function of crop development stage (-)	
k_{soil}	CFBS	'crop factor' for bare soil (-)	1.0

Programs like CROPWAT (Smith, 1992) and CRIWAR (Bos et al., 1996) use crop factors that are a function of the crop development stage. After multiplication with a reference *potential* evapotranspiration rate, a kind of evapotranspiration rate is obtained that is representative for a potentially transpiring crop that is well supplied with water in the root zone and that partly covers the soil. Because the soil has generally a dry top layer, soil evaporation is usually below the potential evaporation rate. Hence, the crop factor combines the effect of an incomplete soil cover and reduced soil evaporation. It enables effective extraction of the potential crop transpiration rate from the reference potential evapotranspiration rate, under the assumption that soil evaporation is constant and relatively small. Significant errors however may be expected when the soil is regularly rewetted and the soil cover fraction is low.

SWAP firstly separates potential plant transpiration rate T_p and potential soil evaporation rate E_p and subsequently calculates the reduction of potential plant transpiration rate and potential soil evaporation rate (Figure 10) according to a physically based approach (Par. 2.2.3). In order to partition potential evapotranspiration rate into potential transpiration rate and potential soil evaporation rate, either the leaf area index, LAI (m^2 m^{-2}) or the soil cover fraction, SC (-), both as a function of crop development, are used.

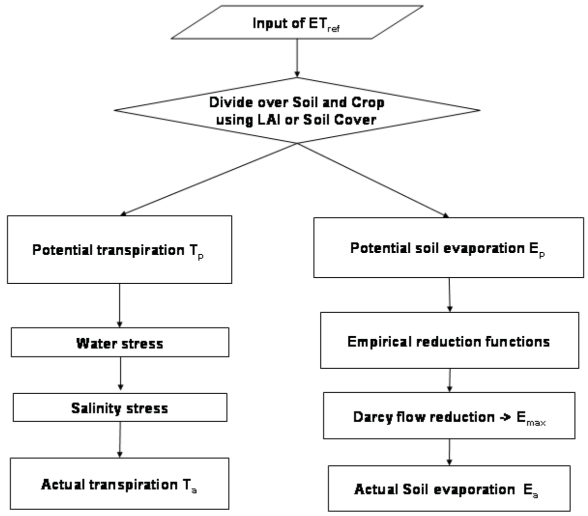

Figure 10 Partitioning of evapotranspiration over crop and soil

3.4.3 Partitioning of potential evapotranspiration

3.4.3.1 Use of leaf area index

The potential evaporation rate of a soil under a standing crop is derived from the Penman Monteith equation by neglecting the aerodynamic term. The aerodynamic term will be small because the wind velocity near the soil surface is relatively small, which makes the aerodynamic resistance r_{air} very large (Ritchie, 1972). Thus, the only source for soil evaporation is net radiation that reaches the soil surface. Assuming that the net radiation inside the canopy decreases according to an exponential function, and that the soil heat flux can be neglected, we can derive (Goudriaan, 1977; Belmans, 1983):

$$E_{p} = E_{p0}\, e^{-\kappa_{gr} LAI} \qquad (3.37)$$

where κ_{gr} (-) is the extinction coefficient for global solar radiation. Ritchie (1972) and Feddes (1978) used κ_{gr} = 0.39 for common crops. More recent approaches estimate κ_{gr} as the product of the extinction coefficient for diffuse visible light, κ_{df} (-), which varies with crop type from 0.4 to 1.1, and the extinction coefficient for direct visible light, κ_{dir} (-):

$$\kappa_{gr} = \kappa_{df}\kappa_{dir} \qquad (3.38)$$

SWAP assumes that the evaporation rate of the water intercepted by the vegetation is equal to ET_{w0}, independent of the soil cover fraction. Then the fraction of the day that the crop is wet, W_{frac} (-), follows from the ratio of the daily amount of intercepted precipitation P_i (Par. 3.3) and ET_{w0}:

$$W_{frac} = \frac{P_i}{ET_{w0}} \quad \text{with} \quad W_{frac} \leq 1.0 \qquad (3.39)$$

During evaporation of intercepted water, the transpiration rate through the leaf stomata is assumed to be negligible. After the canopy has become dry, the transpiration through the leaf stomata starts again at a rate ET_{p0}. SWAP calculates a daily average of the potential transpiration rate, T_p (cm d^{-1}), taking into account the fraction of the day W_{frac} during which the intercepted water evaporates as well as the potential soil evaporation rate E_p:

$$T_p = (1.0 - W_{frac}) ET_{p0} - E_p \qquad \text{with} \qquad T_p \geq 0 \tag{3.40}$$

Model input			
Variable	*Code*	*Description*	*Default*
κ_{df}	KDIF	extinction coefficient for diffuse visible light (-)	0.60
κ_{dir}	KDIR	extinction coefficient for direct visible light (-)	0.75

3.4.3.2 Use of soil cover fraction

As the soil cover is only specified in case of the simple crop growth model, only in that case this option can be used. Taking into account the fraction of the day that the crop is wet (Eq. (3.39)), the potential soil transpiration rate T_p follows straight from:

$$T_p = (1.0 - W_{frac}) SC\ ET_{p0} \tag{3.41}$$

The potential soil evaporation rate is calculated as:

$$E_p = (1.0 - SC)(1 - W_{frac}) E_{p0} \tag{3.42}$$

Model input			
Variable	*Code*	*Description*	*Default*
SC	SCF	soil cover as function of crop development stage (-)	