2. Mathematical Description of the Model Components

In this chapter a mathematical description of all model components is given. First, hydrological processes are described in Section 2.1, followed by vegetation/crop growth processes (Section 2.2), nutrient dynamics processes (Section 2.3), and erosion (Section 2.4). After that a description of the channel routing processes is given in Section 2.5. This chapter is based mostly on the SWAT User Manual (Arnold et al., 1994) and the MATSALU model description (Krysanova et al., 1989a).

2.1 Hydrological Processes

The hydrological submodel in SWIM is based on the following water balance equation

$$SW(t+1) = SW(t) + PRECIP - Q - ET - PERC - SSF \quad (1)$$

where *SW(t)* is the soil water content in the day t, *PRECIP* – precipitation, *Q* – surface runoff, *ET* - evapotranspiration, *PERC* - percolation, and *SSF* – subsurface flow.

All values are the daily amounts in mm. Here the precipitation is an input, assuming that precipitation may differ between sub-basins, but it is uniformly distributed inside the sub-basin. The melted snow is added to precipitation.

The surface runoff, evapotranspiration, percolation below root zone and subsurface flow are described below. Some river basins, especially in the semiarid zone, have alluvial channels that abstract large quantities of stream flow. The transmission losses reduce runoff volumes when the flood wave travels downstream. This reduction is taken into account by a special module that accounts for transmission losses.

2.1.1 Snow Melt

If air temperature is below 0, precipitation occurs as snow, and snow is accumulated. If snow is present on soil, it may be melted when the temperature of the second soil layer exceeds 0°C (according to the model requirements, the depth of the first soil layer must be always set to 10 mm). The approach used is similar to that of CREAMS model (Knisel, 1980). Snow is melted as a function of the snow pack temperature in accordance with the equation

$$SML = 4.57 \cdot TMX$$
$$0. \leq SML \leq SNO \quad (2)$$

where *SML* is the snowmelt rate in mm d^{-1}, *SNO* is the snow in mm of water, *TMX* is the maximum daily air temperature in °C.

Melted snow is treated the same as rainfall for estimating runoff volume and percolation, but rainfall energy is set to 0.

2.1.2 Surface Runoff

The model takes the daily rainfall amounts as input and simulates surface runoff volumes and peak runoff rates. Runoff volume is estimated by using a modification of the Soil Conservation Service (SCS) curve number technique (USDA-SCS, 1972; Arnold *et al.*, 1990). The technique was selected for use in SWIM as well as in SWAT due to several reasons:

(a) it is reliable and has been used for many years in the United States and worldwide;
(b) the required inputs are usually available;
(c) it relates runoff to soil type, land use, and management practices; and
(d) it is computationally efficient.

The use of daily precipitation data is a particularly important feature of the technique because for many locations, and especially at the regional scale, more detailed precipitation data with time increments of less than one day are not available.

Surface runoff is estimated from daily precipitation taking into account a dynamic retention coefficient SMX by using the SCS curve number equation

$$Q = \frac{(PRECIP - 0.2 \cdot SMX)^2}{PRECIP + 0.8 \cdot SMX}, \qquad PRECIP > 0.2 \cdot SMX$$
$$Q = 0, \qquad PRECIP \leq 0.2 \cdot SMX \tag{3}$$

where *Q* is the daily runoff in mm, *PRECIP* is the daily precipitation in mm, and *SMX* is a retention coefficient.

The retention coefficient *SMX* varies a) spatially, because soils, land use, management, and slope vary, and b) temporally, because soil water content is changing. The retention coefficient *SMX* is related to the curve number *CN* by the SCS equation

$$SMX = 254 \cdot \left(\frac{100}{CN} - 1 \right) \tag{4}$$

To illustrate the approach, **Fig. 2.1** shows estimation of surface runoff Q from daily precipitation with equations (3) and (4) assuming different CN values.

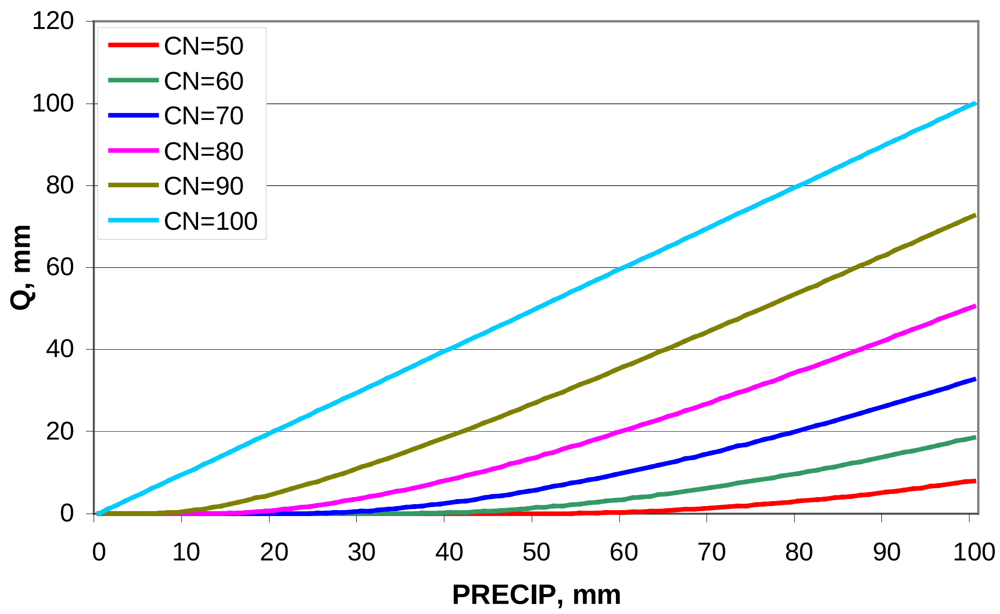

Fig. 2.1 Estimation of surface runoff, Q, from daily precipitation, PRECIP, for different values of CN (equations 3 and 4)

The parameter CN is defined in three variations:

- for moisture condition 1 (or dry conditions) as CN_1
- for moisture condition 2 (or average conditions) as CN_2 and
- for moisture condition 3 (or wet conditions) as CN_3.

CN_2 can be obtained from the SCS hydrology handbook (USDA-SCS, 1972) for a set of land use types, hydrologic soil groups and management practices (see also **Tab. 3.20** in Chapter 3 of the Manual). The corresponding values of CN_1 and CN_3 are also tabulated in the handbook. For computing purposes, CN_1 and CN_3 were related to CN_2 with the equations (see also **Fig. 2.2**)

$$CN_1 = CN_2 - \frac{20 \cdot (100 - CN_2)}{100 - CN_2 + \exp[2.533 - 0.0636 \cdot (100 - CN_2)]} \tag{5}$$

or an approximation of equation 5:

$$CN_1 = -16.911 + 1.3481 \cdot CN_2 - 0.013793 \cdot CN_2^2 + 0.00011772 \cdot CN_2^3 \tag{6}$$

and

$$CN_3 = CN_2 \cdot \exp[0.00673 \cdot (100 - CN_2)] \tag{7}$$

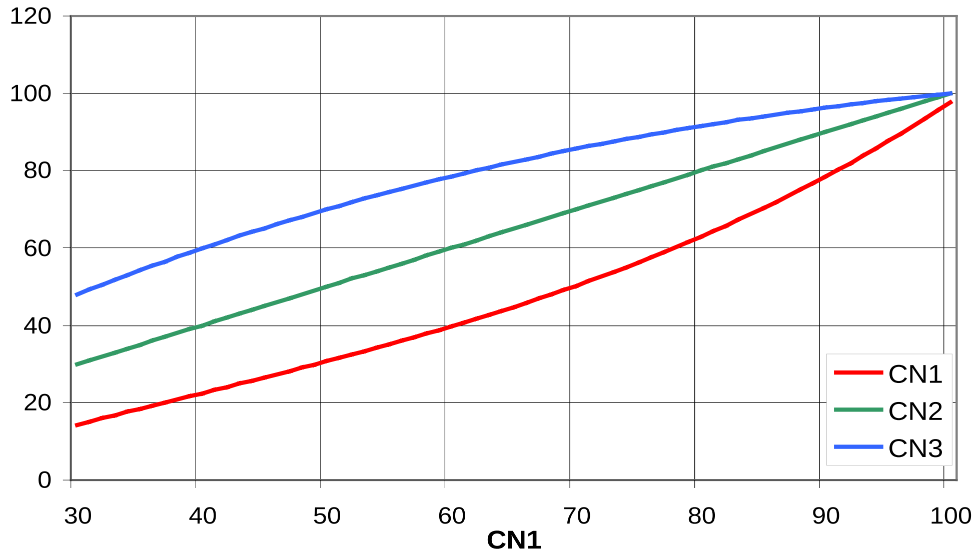

Fig. 2.2 Correspondence between CN_1, CN_2 and CN_3 (equations 6, 7)

The values of CN_1, CN_2 and CN_3 are related to land use types, hydrologic soil groups and management practices. An additional assumption was made to relate curve numbers to slope. Namely, it was assumed that the CN_2 value is appropriate for a 5% slope, the following equation was derived (Arnold et al., 1994) to adjust it for lower and higher slopes (see also **Fig. 2.3**):

$$CN_{2s} = CN_2 + \frac{CN_3 - CN_2}{3} \cdot \left(1 - 2 \cdot \exp(-13.86 \cdot SS)\right) \tag{8}$$

where CN_{2s} is the adjusted CN_2 value, and SS is the slope steepness in m m^{-1}.

The retention coefficient is changing dynamically due to fluctuations in soil water content according to the equation

$$SMX = SMX_1 \cdot \left(1 - \frac{SW}{SW + \exp(WF_1 - WF_2 \cdot SW)}\right) \tag{9}$$

where SMX_1 is the value of SMX associated with CN_1, SW is the soil water content in mm, and WF_1 and WF_2 are shape parameters. **Fig. 2.4** depicts the relationships between the retention coefficient SMX and the curve number CN, on one hand, and the relative soil water content, on the other hand.

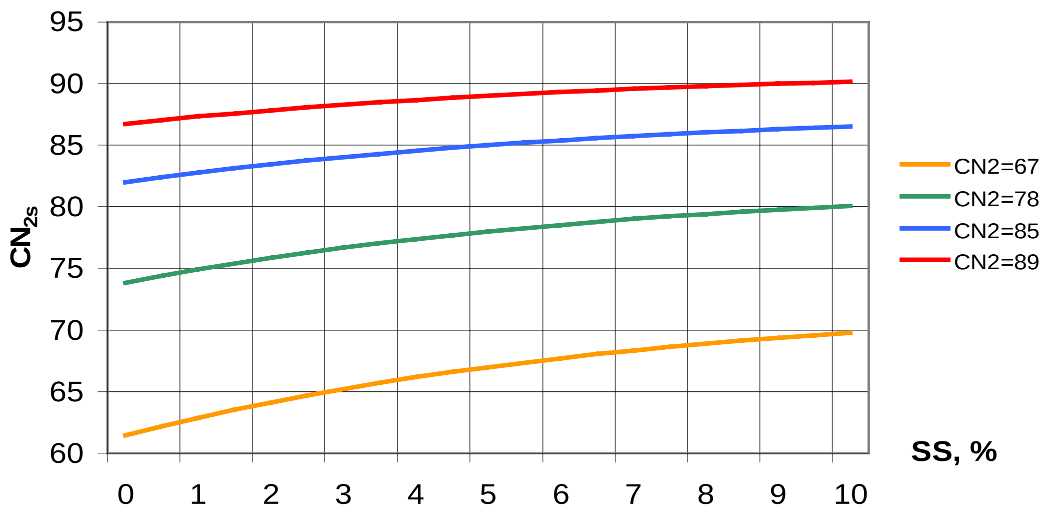

Fig. 2.3 Adjustment of curve number CN_2 to the slope (equation 8) for some typical values of CN_2 = 67, 78, 85, and 89, corresponding to straight row crop and four hydrologic groups A, B, C, and D, respectively

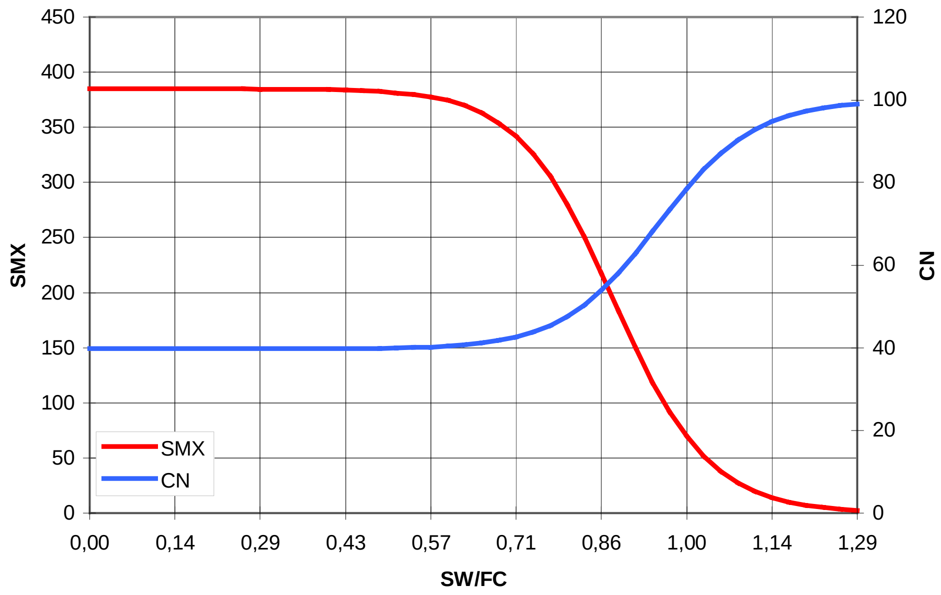

Fig. 2.4 Retention coefficient SMX and curve number CN as functions of soil water content SW (equation 9 and 4) assuming $CN_2 = 60$, $WP = 5$ mm mm^{-1}, $FC = 35$ mm mm^{-1}, $PO = 45$ mm mm^{-1}

The following assumptions are made for the retention coefficient SMX

$$
\begin{aligned}
SMX &= SMX_1 && \text{if} \quad SW = WP, \\
SMX &= SMX_2 && \text{if} \quad FCC = 0.7, \\
SMX &= SMX_3 && \text{if} \quad SW = FC, \\
SMX &= 2.54 && \text{if} \quad SW = PO
\end{aligned}
\tag{10}
$$

where SMX_2 is the retention parameter corresponding to CN_2, SMX_3 is the retention parameter corresponding to CN_3, WP is the wilting point water content in mm mm^{-1}, FC is the field capacity water content in mm mm^{-1}, PO is the soil porosity in mm mm^{-1}, and FFC is the fraction of field capacity defined with the equation

$$FFC = \frac{SW - WP}{FC - WP} \tag{11}$$

The assumption that SMX = 2.54 in (10) means that at full saturation CN = 99 (approaches its maximum).

Values for WF_1 and WF_2 are obtained from a simultaneous solution of equation 8 according to the assumptions (10) as following

$$WF_1 = \ln\left(\frac{FC}{1 - SMX_3/SMX_1} - FC\right) + FC \cdot WF_2 \tag{12}$$

$$WF_2 = \frac{\ln\left(\frac{FC}{1 - SMX_3/SMX_1} - FC\right) - \ln\left(\frac{PO}{1 - 2.54/SMX_1} - PO\right)}{PO - FC} \tag{13}$$

The value of *FFC* defined in equation 11 represents soil water uniformly distributed through the root zone of soil or the upper 1m of soil. Runoff estimates can be improved if the depth distribution of water in soil is known. For example, water distributed near the soil surface results in more runoff than the same volume of water uniformly distributed throughout the soil profile. Since SWIM estimates water content of each soil layer daily, the depth distribution is available. The effect of depth distribution on runoff is expressed in the depth weighting function

$$FFC^* = \frac{\sum_{i=1}^{M}\left(FFC_i \cdot \frac{Z_i - Z_{i-1}}{Z_i}\right)}{\sum_{i=1}^{M}\left(\frac{Z_i - Z_{i-1}}{Z_i}\right)}, \qquad Z_i \le 1.0m \tag{14}$$

where FFC^* is the depth-weighted *FFC* value for use in (9), Z_i is the depth to the bottom of soil layer i in mm, and M is the number of soil layers.

Equation 14 performs two functions:

a) it reduces the influence of lower layers because FFC_i is divided by Z_i and
b) b) it gives proper weight to thick layers relative to thin layers because *FFC* is multiplied by the layer thickness.

There is also a possibility for estimating runoff from frozen soil. If the temperature of the second soil layer is less than 0°C, the retention coefficient is reduced by using the equation

$$SMX_{froz} = SMX \cdot (1. - \exp(-0.000862 \cdot SMX)) \tag{15}$$

where SMX_{froz} is the retention coefficient for frozen ground. Equation 15 increases runoff for frozen soils, but allows significant infiltration when soil is dry.

2.1.3 Peak Runoff Rate

The peak runoff rate is estimated in SWIM for sub-basins using the modified Rational formula (Maidment, 1993; Arnold et al. 1994). A stochastic element is included in the Rational formula to allow a more realistic simulation of peak runoff rates, given only daily rainfall and monthly rainfall intensity. The Rational formula can be written in the form

$$PEAKQ = \frac{RUNC \cdot RI \cdot A}{360} \tag{16}$$

where *PEAKQ* is the peak runoff rate in m^3 s^{-1}, *RUNC* is a dimentionless runoff coefficient expressing the watershed infiltration characteristics, *RI* is the rainfall intensity in mm h^{-1} for the watershed's time of concentration, and *A* is the drainage area in ha.

The runoff coefficient can be calculated for each day from the amounts of precipitation and runoff as following

$$RUNC = \frac{Q}{PRECIP} \tag{17}$$

Since daily precipitation is input and *Q* is calculated with equation (3), *RUNC* can be estimated directly.

Rainfall intensity can be expressed as

$$RI = \frac{PRECIP_{tc}}{TC} \tag{18}$$

where *TC* is the watershed's time of concentration in h, and $PRECIP_{tc}$ is the amount of rainfall in mm during the time of concentration.

The value of $PRECIP_{tc}$ can be estimated by developing a relationship with total daily *PRECIP*. Generally, $PRECIP_{tc}$ and $PRECIP_{24}$ (24-h duration is appropriate for the daily time step model) are proportional for various frequencies.

Thus, a dimensionless parameter α that expresses the proportion of total daily rainfall that occurs during time of concentration can be introduced. Then

$$PRECIP_{tc} = \alpha \cdot PRECIP_{24} \tag{19}$$

The equation for the peak runoff rate is obtained by substituting equations 17, 18, and 19 into equation 16:

$$PEAKQ = \frac{\alpha \cdot Q \cdot A}{360 \cdot TC} \tag{20}$$

The time of concentration can be estimated by adding the surface and channel flow times

$$TC = TC_{ch} + TC_{ov} \tag{21}$$

where TC_{ch} is the time of concentration for channel flow in h, and TC_{ov} is the time of concentration for overland surface flow in h.

The time of concentration for channel flow can be calculated by the equation

$$TC_{ch} = \frac{CHFL}{3.6 \cdot CHV} \tag{22}$$

where *CHFL* is the average channel flow length for the basin in km and *CHV* is the average channel velocity in m s^{-1}.

The average channel flow length can be estimated by the equation

$$CHFL = \sqrt{CHL \cdot CHL_{cen}} \tag{23}$$

where *CHL* is the channel length from the most distant point to the watershed outlet in km and CHL_{cen} is the distance from the outlet along the channel to the watershed centroid in km. We can assume that $CHL_{cen}=0.5\ CHL$.

Average velocity can be estimated by using Manning's equation and assuming a trapezoidal channel with 2:1 side slopes and a 10:1 bottom width to depth ratio. Substitution of these estimated and assumed values, and conversion of units gives the following estimation of the time of concentration for channel

$$TC_{ch} = \frac{0.62 \cdot CHL \cdot CHN^{0.75}}{(QAV \cdot A)^{0.25} \cdot CHS^{0.375}} \tag{24}$$

where *CHN* is Manning's n, *QAV* is the average flow rate in mm h^{-1}, and *CHS* is the average channel slope in m m^{-1}.

The average flow rate is obtained from the estimated average flow rate from a unit source in the watershed (1 ha area) and the relationship

$$QAV = QAV_0 \cdot A^{-0.5} \tag{25}$$

where QAV_0 is the average flow rate from a 1 ha area in mm h^{-1}.

Substitution of equation 25 into equation 24 gives the final equation for TC_{ch}:

$$TC_{ch} = \frac{0.62 \cdot CHL \cdot CHN^{0.75}}{QAV_0^{\ 0.25} \cdot A^{0.125} \cdot CHS^{0.375}} \qquad (26)$$

A similar approach is used to estimate the time of concentration for overland surface flow

$$TC_{ov} = \frac{SL}{3600 \cdot SV} \qquad (27)$$

where *SL* is the surface slope length in m and *SV* is the surface flow velocity in m s^{-1}.

The surface flow velocity is estimated applying Manning's equation to a strip 1 m wide down the slope length, and assuming that flow is concentrated into a small trapezoidal channel with 1:1 side slopes and 5:1 bottom width to depth ratio as following

$$SV = \frac{0.00748 \cdot FD^{0.666} \cdot SS^{0.5}}{SN} \qquad (28)$$

where *SV* is the surface flow velocity in m^3 s^{-1}, *FD* is flow depth in m, *SS* is the land surface slope in m m^{-1}, and *SN* is Manning's roughness coefficient 'n' for the surface.

The average flow depth FD is calculated from Manning's equation as a function of flow rate

$$FD = \left(\frac{QAV_0 \cdot SN}{5.025 \cdot SS^{0.5}} \right)^{0.375} \qquad (29)$$

where AVQ_0 is the average flow rate in m^3 s^{-1}. Substitution of equations 28 and 29 into equation 27 gives

$$TC_{ov} = \frac{0.0556 \cdot SL \cdot SN^{0.75}}{QAV_0^{\ 0.25} \cdot SS^{0.375}} \qquad (30)$$

The average flow rate from a unit source area in the basin is estimated with the equation

$$QAV_0 = \frac{Q}{DUR} \qquad (31)$$

where the rainfall duration *DUR* (in h) is calculated using the equation

$$DUR = \frac{2.303}{-\ln(1.-\alpha_{0.5})} \tag{32}$$

where $\alpha_{0.5}$ is the fraction of rainfall that occurs during 0.5 h. It is calculated with equation 19 using $PRECIP_{0.5}$ instead of $PRECIP_{tc}$.

Equation 32 is derived assuming that rainfall intensity is exponentially distributed. To evaluate α properly, variation in rainfall patterns must be considered. For some short duration storms, most or all the rain occurs during TC causing α to approach its upper limit of 1.0. Other storm events of uniform intensity cause α to approach a minimum value. By substituting the products of intensity and time into equation 19, an expression for the minimum value of α, α_{min}, is obtained

$$\alpha_{\min} = TC/24 \tag{33}$$

Thus, α ranges within the limits

$$TC/24 < \alpha < 1.0 \tag{34}$$

Although confined between limits, the value of α is assigned with considerable uncertainty when only daily rainfall and simulated runoff amounts are given. This can lead to considerable uncertainties in estimating daily runoff and has to be kept in mind. The value of α is estimated in the model from the gamma distribution, taking into account the average monthly rainfall intensity for the basin under study.

2.1.4 Percolation

A storage routing technique (Arnold et al., 1990) is used in SWIM to simulate percolation through each soil layer. The percolation from the bottom soil layer is treated as recharge to the shallow aquifer. The storage routing technique is based on the equation

$$SW(t+1) = SW(t) \cdot \exp\left(\frac{-\Delta t}{TT_i}\right) \tag{35}$$

where *SW(t+1)* and *SW(t)* are the soil water contents at the beginning and end of the day in mm, Δt is the time interval (24 h), and TT_i is the travel time through layer *i* in h. Thus, the percolation can be calculated by subtracting SW_t from SW_{t+1}:

$$PERC_i = SW_i \cdot \left[1.-\exp\left(\frac{-\Delta t}{TT_i}\right)\right] \tag{36}$$

where *PERC* is the percolation rate in mm d^{-1}. The travel time TT_i is calculated for each soil layer with the linear storage equation

$$TT_i = \frac{SW_i - FC_i}{HC_i} \qquad (37)$$

where HC_i is the hydraulic conductivity in mm h^{-1} and *FC* is the field capacity water content for layer *i* in mm. The hydraulic conductivity is varying from the saturated conductivity value at saturation to near zero at field capacity (see also **Fig. 2.5**) as

$$HC_i = SC_i \cdot \left(\frac{SW_i}{UL_i} \right)^{\beta_i} \qquad (38)$$

where SC_i is the saturated conductivity for layer i in mm h^{-1}, UL_i is soil water content at saturation in mm mm^{-1},and β_i is a shape parameter that causes HC_i to approach zero as SW_i approaches FC_i.

The equation for estimating β_i is

$$\beta_i = \frac{-2.655}{\log_{10}\left(\frac{FC_i}{UL_i} \right)} \qquad (39)$$

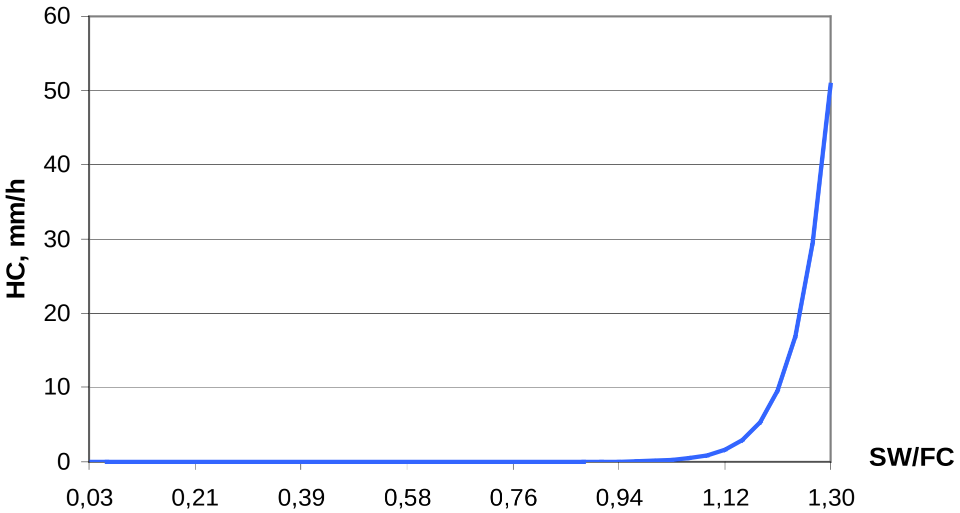

Fig. 2.5 Hydraulic conductivity as a function of soil water content (equation 39) assuming SC = 50.8 mm h^{-1}, FC = 33 mm mm^{-1}, UL = 43 mm mm^{-1}

The constant in equation 39 is set to –2.655 to assure that at field capacity

$$HC_i = 0.002 \cdot SC_i \tag{40}$$

Water flow through a soil layer may occur until the lower layer is not saturated. If the layer below the layer being considered is saturated, no flow can occur. The effect of lower layer water content is expressed by the equation

$$PERC_{ic} = PERC_i \cdot \sqrt{1 - \frac{SW_i + 1}{UL_i + 1}} \tag{41}$$

where $PERC_{ic}$ is the percolation rate for layer i in mm d^{-1} corrected for layer i+1 water content and $PERC_i$ is the percolation calculated with equation 36.

Percolation is also affected by soil temperature. If the temperature in a particular layer is O°C or below, no percolation is allowed from that layer.

Since the one-day time interval is relatively low for routing flow through the soil root zone, the water is divided into several portions for routing through soil. This is necessary because flow rates are dependent upon soil water content, which is continuously changing. For example, if the soil is extremely wet, equations 36, 37, and 38 may overestimate percolation, if only one routing is performed. To overcome this problem, each layer's inflow is divided into 4-mm slugs for routing.

Besides, when the inflow is divided into 4-mm slugs and each slug is routed individually through the layers, the relationship taking into account the lower layer water content (equation 41) works more realistically.

2.1.5 Lateral Subsurface Flow

The kinematic storage model developed by Sloan et al. (1983) uses the mass continuity equation for the entire soil profile, considering it as the control volume. The mass continuity equation in the finite difference form for the kinematic storage model is

$$\frac{SUP_2 - SUP_1}{t_2 - t_1} = WIR \cdot SL - \frac{SSF_1 + SSF_2}{2} \tag{42}$$

where *SUP* is the drainable volume of water stored in the saturated zone m m^{-1} (water above field capacity), t is time in h, *SSF* is the lateral subsurface flow in m^3 h^{-1}, *WIR* is the rate of water input to the saturated zone in m^2 h^{-1}, *SL* is the hillslope length in m, and subscripts 1 and 2 refer to the beginning and end of the time step, respectively. The drainable volume of water stored, *SUP*, is updated daily.

The lateral flow at the hillslope outlet is given by

$$SSF = \frac{2 \cdot SUP \cdot VEL \cdot SLW}{PORD \cdot SL} \tag{43}$$

where *VEL* is the velocity of flow at the outlet in mm h^{-1}, *SLW* is the hillslope width in m, and *PORD* is the drainable porosity of the soil in m m^{-1}. Velocity at the outlet is estimated as

$$VEL = SC \cdot \sin(\nu) \tag{44}$$

where *SC* is the saturated conductivity in mm h^{-1}, and ν is the hillslope steepness in m m^{-1} . Combination of equations 43 and 44 gives

$$SSF = 0.024 \cdot \frac{2 \cdot SUP \cdot SC \cdot \sin(\nu)}{PORD \cdot SL} \tag{45}$$

where *SSF* is in mm d^{-1}, *SUP* in m m^{-1}, γ in m m^{-1}, *PORD* in m m^{-1}, and *SL* in m.

If the saturated zone rises above the soil layer, water is allowed to flow to the layer above. The amount of flow upward is estimated as a function of saturated conductivity *SC* and the saturated slope length

$$QUP = \frac{24 \cdot SC \cdot SL_{sat}}{SL} \tag{46}$$

where *QUP* is the upward flow in mm d^{-1}, and Sl_{sat} is the saturated slope length in m.

To account for multiple layers, the model is applied to each soil layer independently starting at the upper layer to allow for percolation from one soil layer to the next.

2.1.6 Potential Evapotranspiration

The method of Priestley-Taylor (1972) is used in the model for estimation of potential evapotranspiration, which requires only solar radiation, air temperature, and elevation as inputs. Instead, the method of Penman-Monteith (Monteith, 1965) can be used, if additional input data are available. The Penman-Monteith method requires solar radiation, air temperature, wind speed, and relative humidity as input.

The Priestley-Taylor method estimates potential evapotranspiration as a function of net radiation as following

$$EO = 1.28 \cdot \left(\frac{RAD}{HV}\right) \cdot \left(\frac{\delta}{\delta + \gamma}\right) \tag{47}$$

where *EO* is the potential evaporation in mm, *RAD* is the net radiation in MJ m^{-2}, *HV* is the latent heat of vaporization in MJ kg^{-1}, δ is the slope of the saturation vapor pressure curve in kPa C^{-1}, and γ is a psychrometer constant in kPa C^{-1}.

The latent heat of vaporization is estimated as a function of the mean daily air temperature T in °C

$$HV = 2.5 - 0.0022 \cdot T \tag{48}$$

The saturation vapor pressure *VP* is also estimated as a function of temperature

$$VP = 0.1 \cdot \exp\left[54.88 - 5.03 \cdot \ln(T + 273) - \frac{6791}{T + 273}\right] \tag{49}$$

Then the slope of the saturation vapor pressure curve is calculated with the equation

$$\delta = \left(\frac{VP}{T + 273}\right) \cdot \left(\frac{6791}{T + 273} - 5.03\right) \tag{50}$$

The psychrometer constant γ is calculated as a function of barometric pressure BP (in kPa)

$$\gamma = 6.6 \cdot 10^{-4} \cdot BP \tag{51}$$

The barometric pressure is estimated as a function of elevation ELEV (in m)

$$BP = 101 - 0.0115 \cdot ELEV + 5.44 \cdot 10^{-7} \cdot ELEV^{2} \tag{52}$$

If actual net radiation is not available, in can be estimated from the maximum solar radiation as following. First, the maximum possible solar radiation *RAM* in Ly is calculated as

$$RAM = \\ = \frac{711}{D^2} \cdot \left(\phi \cdot \sin\left(\frac{2 \cdot \pi \cdot LAT}{360}\right) \cdot \sin(\theta) + \cos\left(\frac{2 \cdot \pi \cdot LAT}{360}\right) \cdot \cos(\theta) \cdot \sin(\phi)\right) \tag{53}$$

where *D* is the earth's radius vector in km, ϕ is the sun's half day length in radians, *LAT* is the latitude of the site in degrees, and θ is the sun's declination angle in radians.

The earth’s radius vector D can be calculated for any day t as

$$D = \frac{1}{\sqrt{1.+0.0335 \cdot \sin\left[\frac{2 \cdot \pi \cdot (t+88.2)}{365}\right]}} \tag{54}$$

The sun's declination angle is calculated with the equation

$$\theta = 0.4102 \cdot \sin\left[\frac{2 \cdot \pi \cdot (t-80.25)}{365}\right] \tag{55}$$

The sun’s half day length is calculated as

$$\begin{aligned} \phi &= \cos^{-1}\left[\tan\left(\frac{2 \cdot \pi \cdot LAT}{360}\right) \cdot \tan(\theta)\right], & -1 \le \theta \le 1 \\ \phi &= 0 \ , & \theta > 1 \\ \phi &= \pi \ , & \theta \le -1 \end{aligned} \tag{56}$$

Then the net radiation is estimated with the equation

$$RAD = RAM \cdot (1.-ALB) \tag{57}$$

where RAD is the solar radiation in MJ m^{-2} and ALB is albedo.

The albedo is estimated by considering the soil, crop/vegetation cover, and snow cover. When crops are growing, albedo is determined by using the equation

$$ALB = 0.23 \cdot (1.-SCOV) + ALB_{soil} \cdot SCOV \tag{58}$$

where 0.23 is the albedo for plants, ALB_{soil} is the soil albedo, and $SCOV$ is a soil cover index.

The value of the soil cover index $SCOV$ ranges from 0 to 1.0 according to the equation

$$SCOV = \exp(-0.05 \cdot BMR) \tag{59}$$

where BMR is the sum of the above ground biomass and crop residue in t ha^{-1}.

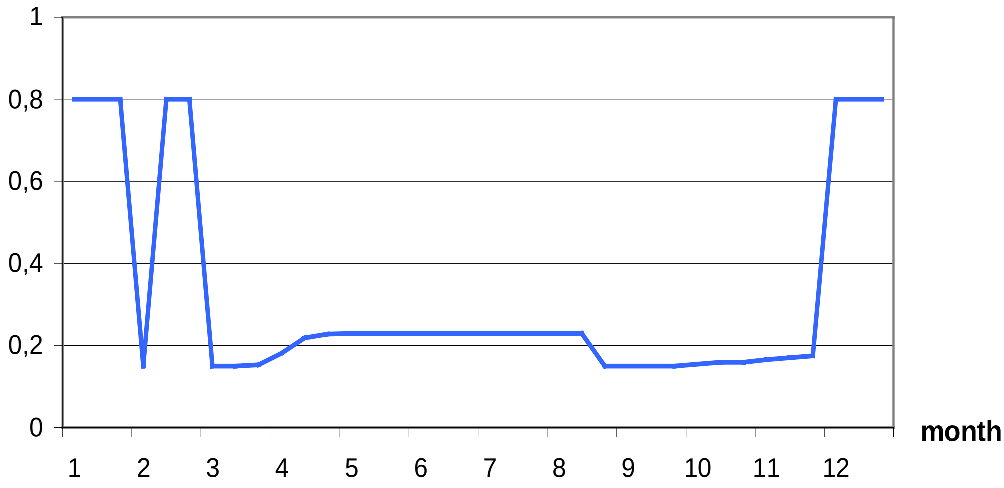

Fig. 2.6 An example of the annual dynamics of soil albedo (equations 58, 59)

If a snow cover exists with 5 mm or greater water content, the value of albedo is set to 0.8. If the snow cover is less than 5 mm and no crop is growing, the soil albedo is set to the input value (default value = 0.15). An example on **Fig. 2.6** shows possible seasonal dynamics of albedo in a temperate zone with a maximum 0.8 in winter (snow cover), minimum in march and september (equal to the bare soil albedo), and increasing up to 0.23 in summer (crop growth).

2.1.7 Soil Evaporation and Plant Transpiration.

The model calculates evaporation from soils and transpiration by plants separately using an approach similar to that of Ritchie (1972). The plant transpiration is calculated as

$$EP = \frac{EO \cdot LAI}{3}, \qquad 0 \le LAI \le 3.0$$
$$EP = EO, \qquad LAI > 3.0 \tag{60}$$

where *EO* is the potential evapotranspiration in mm d^{-1} estimated by equation (47), *EP* is the plant water transpiration rate in mm d^{-1} and *LAI* is the leaf area index (area of plant leaves relative to the soil surface area).

If soil water is limited, plant water transpiration is reduced. The approach is described in section 2.2.2 about water stress.

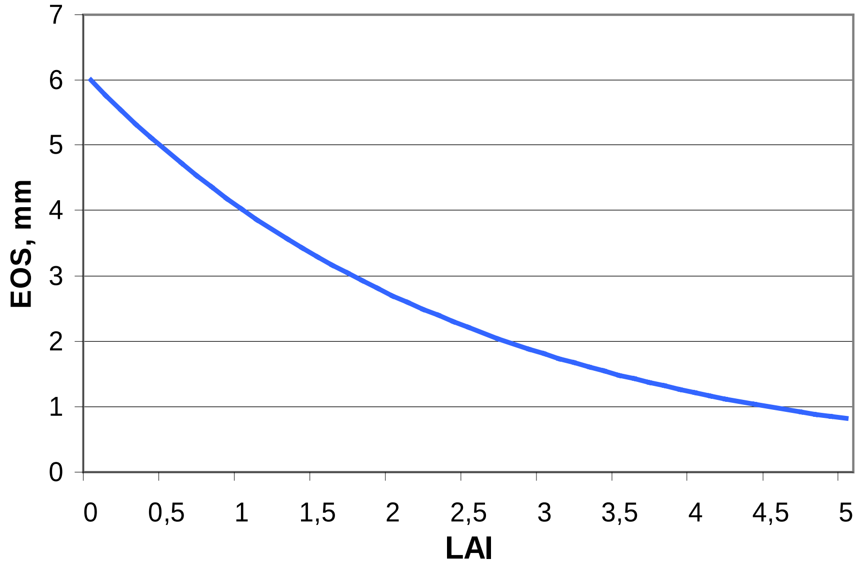

Fig. 2.7 Potential soil evaporation, ESO, as a function of leaf area index, LAI (equation 61) under assumption that EO = 6 mm d^{-1}

Potential soil evaporation *ESO* in mm d^{-1} is simulated by an exponential function of leaf area index *LAI* according to the equation of Richardson and Richie (1973) (see also **Fig. 2.7**):

$$ESO = EO \cdot \exp(-0.4 \cdot LAI) \tag{61}$$

Actual soil evaporation is calculated in two stages. In the first stage, soil evaporation is limited only by the energy available at the surface, and is equal to the potential soil evaporation. When the accumulated soil evaporation exceeds the first stage threshold (equal to 6 mm), the second stage begins. Then soil evaporation is estimated with the equation

$$ES = 3.5 \cdot \left(\sqrt{TST} - \sqrt{TST - 1}\right) \tag{62}$$

where *ES* is the soil evaporation for day t in mm d^{-1} and *TST* is the number of days since stage two evaporation began.

Actual soil water evaporation is estimated on the basis of the top 30 cm of soil and snow cover, if any. If the water content of the snow cover is greater or equal to ES, the soil evaporation comes from the snow cover. If ES exceeds the water content of the snow cover, water is removed from the upper soil layers if available.

2.1.8 Groundwater Flow

The groundwater submodel in the integrated river basin model like SWIM is intended for general use in regions where extensive field measurements are not available. Thus, the groundwater component has to be parameterized using readily available inputs. Also, it must have the level of sophistication similar to those of the other components. Therefore a detailed numerical model is not justified for this case, and a relatively simple yet realistic approach was chosen for use in SWAT and SWIM.

The simulated hydrological system consists of four control volumes that include:

- the soil surface,
- the soil profile or root zone,
- the shallow aquifer, and
- the deep aquifer.

The percolation from the soil profile is assumed to recharge the shallow aquifer. The surface runoff, the lateral subsurface flow from the soil profile, and return flow from the shallow aquifer contribute to the stream flow. The water balance equation for the shallow aquifer is

$$SAW(t+1) = SAW(t) + RCH - REVAP - GWQ - SEEP \tag{63}$$

where *SAW(t)* is the shallow aquifer storage in the day *t*, *RCH* is the recharge, *REVAP* is the water flow from the shallow aquifer back to the soil profile, *GWQ* is the return flow or groundwater contribution to streamflow, *SEEP* is the percolation or seepage to the deep aquifer (all – in mm d^{-1}), and *t* is the day.

REVAP is defined as water that raises from the shallow aquifer to the soil profile and is lost to the atmosphere by soil evaporation or plant root uptake.

The approach of Smedema and Rycroft (1983), who derived the non-steady-state response of groundwater flow to periodic recharge from Hooghoudt's (1940) steady-state formula, is used

$$GWQ = 8 \cdot \frac{KD \cdot GWH}{DS^2} \tag{64}$$

where *KD* is the hydraulic conductivity of groundwater in mm d^{-1}, *DS* is the drain spacing in m, and *GWH* is the water table height in m.

Assuming that the shallow aquifer is recharged by seepage from stream channels, reservoirs, or the soil profile (rainfall and irrigation), and is depleted by the return flow to the stream, fluctuations of water table can be estimated using the equation of Smedema and Rycroft (1983)

$$\frac{d(GWH)}{dt} = \frac{RCH - GWQ}{0.8 \cdot SY} \tag{65}$$

where *SY* is the specific yield.

The return flow can be estimated assuming that its variation with time is also linearly related to the rate of change of the water table height:

$$\frac{d(GWQ)}{dt} = 10 \cdot \frac{KD \cdot (RCH - GWQ)}{SY \cdot DS^2} = RF \cdot (RCH - GWQ) \qquad (66)$$

where *RF* is the constant of proportionality or the reaction factor for groundwater.

Integration of equation 66 gives

$$GWQ(t+1) = GWQ(t) \cdot \exp(-RF \cdot \Delta t) + RCH \cdot [1 - \exp(-RF \cdot \Delta t)] \qquad (67)$$

The relationship for the water table height is derived combining equations 64 and 67. It results in the following relationship

$$\begin{aligned} & GWH(t+1) = \\ & = GWH(t) \cdot \exp(-RF \cdot \Delta t) + \frac{RCH}{0.8 \cdot SY \cdot RF} \cdot (1. - \exp(-RF \cdot \Delta t)) \end{aligned} \qquad (68)$$

The percolation from the soil profile is assumed to recharge the shallow aquifer. The delay time or drainage time of the aquifer is used to correct the recharge. Sangrey et al. (1984) used an exponential decay weighting function proposed by Venetis (1969) to estimate the delay time for return flow in their precipitation / groundwater response model

$$\begin{aligned} & RCH(t+1) = \\ & = \left(1. - \exp\left(-\frac{1}{DEL}\right)\right) \cdot RCH(t+1) + \exp\left(-\frac{1}{DEL}\right) \cdot RCH(t) \end{aligned} \qquad (69)$$

where *DEL* is the delay time or drainage time of the aquifer in days (Sangrey et al., 1984). This equation will affect only the timing of the return flow and not the total volume. The equation (69) is used in SWIM to correct the recharge.

The volume of water flow from the shallow aquifer back to the soil profile, REVAP, is estimated with the equations

$$\begin{aligned} & REVAP = CR \cdot ET \ , \qquad && REVAP > RST \\ & REVAP = 0 \ , \qquad && REVAP \le RST \end{aligned} \qquad (70)$$

where *ET* is the actual evapotranspiration occurring in the soil profile, CR is the *revap* coefficient, and *RST* is the *revap* storage in mm.

The amount of percolation or seepage from the shallow aquifer (recharge to the deep aquifer) is estimated as a linear function

$$SEEP = CS \cdot RCH \tag{71}$$

where *CS* is the seepage coefficient.

2.1.9 Transmission Losses

Many watersheds, especially in semiarid areas, have alluvial channels that abstract large quantities of stream flow (Lane, 1982). The abstractions, or transmission losses, reduce runoff volumes because water is lost when the flood wave travels downstream.

A procedure for estimating transmission losses for ephemeral streams is described by Lane in the SCS Hydrology Handbook (USDA, 1983, chapter 19). The procedure is based on derived regression equations for estimation of transmission losses in the absence of observed inflow-outflow data. It enables the user to estimate transmission losses for similar channels of arbitrary length and width using channel geometry parameters (width and depth) and Manning’s "n". This procedure is used in SWIM as well as in SWAT to estimate transmission losses.

The unit channel intercept and slope, and the decay factor are estimated with regression equations obtained from the analysis of observed data in different conditions:

$$AR = -0.001831 \cdot CHK \cdot DU \tag{72}$$

$$DEC = -1.09 \cdot \ln\left(1. - \frac{0.2649 \cdot CHK \cdot DU}{VOLQ_{in}}\right) \tag{73}$$

$$BR = \exp(-DEC) \tag{74}$$

where *AR* is the unit channel intercept in m^3, *CHK* is the effective hydraulic conductivity of the channel alluvium in mm h^{-1} (Lane, 1982; USDA, 1983 update), *DU* is the duration of streamflow in h, *DEC* is the decay factor in m km^{-1}, $VOLQ_{in}$ is inflow volume of m^3, and *BR* is the unit channel regression slope.

The inflow volume is assumed to be equal to the surface runoff from the sub-basin. The flow duration DU in h is estimated from

$$DU = \frac{Q \cdot A}{1.8 \cdot PEAKQ} \tag{75}$$

where *Q* is the surface runoff volume in mm, *A* is the drainage area in ha, and *PEAKQ* is the peak flow rate in $m^3\ s^{-1}$.

The regression parameters are estimated as

$$AX = [AR \cdot (1 - BR)] \cdot (1 - BR \cdot CHL) \qquad (76)$$

$$BX = CHL \cdot CHW \cdot \exp(-2.04 \cdot DEC) \qquad (77)$$

$$TH_0 = -\frac{AX}{BX} \qquad (78)$$

where *AX* is the regression intercept in m km^{-1}, *BX* is the regression slope, *CHW* is average width of flow in m, *CHL* is length of channel in km, and TH_o is the threshold volume for a unit channel in m^3.

Then the final equation for runoff volume after losses, $VOLQ_{tr}$, is

$$\begin{array}{ll} VOLQ_{tr} = -AX + BX \cdot VOLQ_{in} & VOLQ_{in} > TH_0 \\ VOLQ_{tr} = 0 & VOLQ_{in} < TH_0 \end{array} \qquad (79)$$

The final equation for peak discharge after losses $PEAKQ_{tr}$, is

$$PEAKQ_{tr} = \frac{12.1 \cdot AX}{DU - (1 - BX) \cdot VOLQ_{in} + BX \cdot PEAKQ_{in}}, \quad VOLQ_{in} > 0 \qquad (80)$$

where $PEAKQ_{in}$ is the initial peak runoff rate.

2.2 Crop / Vegetation Growth

2.2.1 Crop Growth

The crop model in SWIM and SWAT is a simplification of the EPIC crop model (Williams et al., 1984). The SWIM model uses

- a concept of phenological crop development based on daily accumulated heat units,
- Monteith's approach (1977) for potential biomass,
- water, temperature, and nutrients stress factors, and
- harvest index for partitioning grain yield.

However, the more detailed EPIC root growth and nutrient cycling modules are not included.

A single model is used for simulating all the crops and natural vegetation considered (see **Table 3.14** in Chapter 3). The model is capable of simulating crop growth for both annual and perennial plants. Annual crops grow from planting date to harvest date or until the accumulated heat units equal the potential heat units for the crop. Perennial crops maintain their root systems throughout the year, although the plant may become dormant after frost. Later the term 'crop' will be used instead of 'crop or natural vegetation'.

Phenological development of the crop is based on accumulation of daily heat units. The value of heat units accumulated in the day t, *HUNA*, is calculated as

$$HUNA(t) = \left(\frac{TMX + TMN}{2}\right) - TB \ , \qquad HUNA \geq 0 \tag{81}$$

where *TMX* and *TMN* are the maximum and minimum temperature in °C, and *TB* is the crop-specific base temperature in °C assuming that no growth occurs at or below TB.

Then the heat unit index *IHUN* ranging from 0 at planting to 1 at physiological maturity is calculated as

$$IHUN = \frac{\sum_t HUNA(t)}{PHUN} \tag{82}$$

where *PHUN* is the value of potential heat units required for the maturity of the crop. The values of *PHUN* for different crops are provided in the crop database supplemented with the model.

Interception of solar radiation is estimated with Beer's law equation (Monsi and Saeki, 1953) as a function of photosynthetic active radiation and leaf area index (see **Fig. 2.8**)

$$PAR = 0.02092 \cdot RAD \cdot \left[1 - \exp(-0.65 \cdot LAI)\right] \tag{83}$$

where *PAR* is the photosynthetic active radiation in MJ m^{-2}, *RAD* is solar radiation in Ly, and *LAI* is the leaf area index.

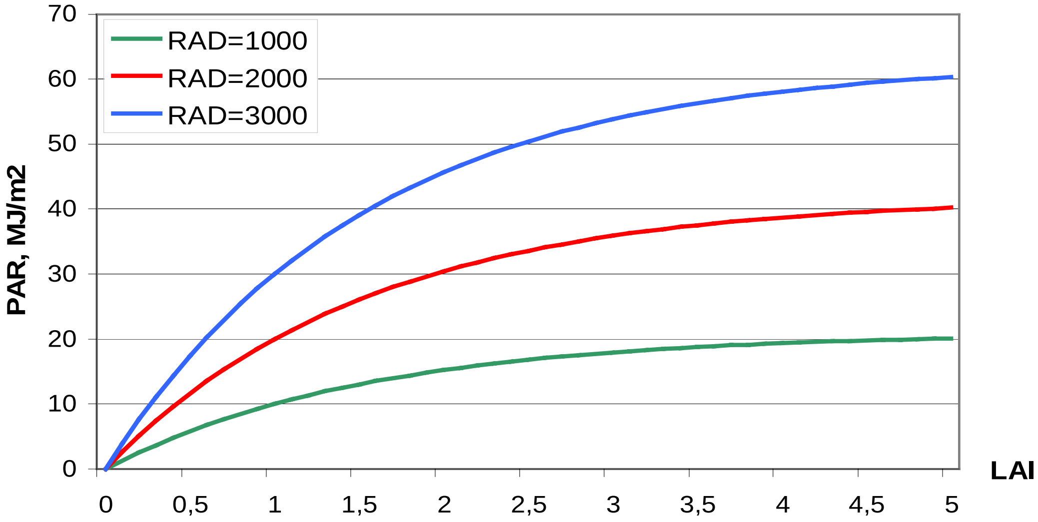

Fig. 2.8 Photosynthetic active radiation, PAR as a function of leaf area index, LAI for RAD= 1000, 2000 and 3000 Ly (equation 83)

Potential increase in biomass for a day is calculated using the approach of Monteith (1977) with the equation

$$\Delta BP = BE \cdot PAR \tag{84}$$

where ΔBP is the daily potential increase in total biomass in kg h^{-1} a^{-1}, and BE is the crop-specific parameter for converting energy to biomass in kg m^2 MJ^{-1} ha^{-1} d^{-1}. The latter one is taken from the crop database.

The potential increase in biomass estimated with equation 84 is adjusted daily if one of the plant stress factors is less than 1.0. The model considers stresses caused by water, nutrients, and temperature. The following equation is used to estimate the daily increase in biomass ΔB (in kg ha^{-1})

$$\Delta B = \Delta BP \cdot REGF \tag{85}$$

where *REGF* is the crop growth regulating factor estimated as the minimum stress factor:

$$REGF = \min(WS, TS, NS, PS) \tag{86}$$

where *WS, TS, NS, PS* are stress factors caused by water, temperature, nitrogen and phosphorus, all varying between 0 and 1.

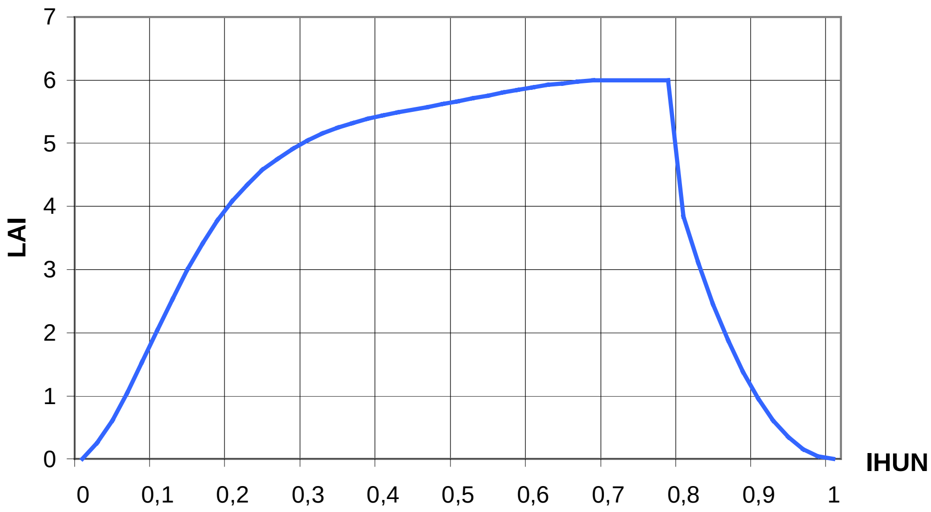

Fig. 2.9 Leaf area index as a function of the heat unit index (equation 87)

The leaf area index LAI is simulated as a function of heat units and biomass, differently for two phases of the growing season:

$$LAI = \frac{LAIMX \cdot BAG}{BAG + \exp(9.5 - 0.0006 \cdot BAG)}, \qquad IHUN \le DLAI$$
$$LAI = 16 \cdot LAIMX \cdot (1 - IHUN)^2, \qquad IHUN > DLAI \tag{87}$$

where *LAIMX* is the maximum potential LAI for the specific crop, *BAG* is aboveground biomass in kg ha^{-1}, and *DLAI* is the fraction of the growing season before *LAI* starts declining (crop-specific parameter). An example of LAI dynamics is shown in **Fig. 2.9**).

The aboveground biomass is estimated as

$$BAG = (1 - RWT) \cdot BT \tag{88}$$

where *RWT* is the fraction of total biomass partitioned to the root system, and *BT* is total biomass in kg ha^{-1}.

The fraction of total biomass partitioned to the root system normally decreases from 0.3 to 0.5 in the seedling to 0.05 to 0.20 at maturity (Jones, 1985). The model estimates the root fraction to range linearly from 0.4 at emergence to 0.2 at maturity using the equation

$$RWT = (0.4 - 0.2 \cdot IHUN) \tag{89}$$

2.2.2 Growth Constraint: Water Stress

The water stress factor is calculated by considering water supply and water demand with the following equation

$$WS = \frac{\sum_{i=1}^{M} WU_i}{EP} \tag{90}$$

where WU_i is plant water use in layer i in mm. The value of potential plant transpiration *EP* is calculated in the evapotranspiration module.

The plant water use is estimated using the approach of Williams and Hann (1978) for simulating plant water uptake. First, the root depth is calculated with the equation

$$RD = 2.5 \cdot IHUN \cdot RDMX \tag{91}$$

where *RD* is the fraction of the root zone that contains roots and *RDMX* is the maximum root depth in m (crop-specific parameter).

Then the potential water use in each soil layer is estimated with the equation

$$WUP_i = \frac{EP}{1 - \exp(RDP)} \cdot \left(1 - \exp\left(-\frac{RDP \cdot RZD_i}{RD}\right)\right) \tag{92}$$

where WUP_i is the potential water use rate from layer i in mm d^{-1}, *RDP* is the rate-depth parameter, and RZD_i is the root zone depth parameter for the layer i in mm.

The latter one is defined as

$$RZD_i = \left\{ \begin{array}{ll} Z_i, & RD > Z_i \\ RD, & RD \leq Z_i \end{array} \right\} \tag{93}$$

The value of *RDP* used in the model (3.065) was determined assuming that about 30% of the total water use comes from the top 10% of the root zone. The details of evaluating *RDP* are given in Williams and Hann (1978). Equation 92 allows roots to compensate for water deficits in certain layers by using more water in layers with adequate supply.

Then the potential water use must be adjusted for water deficits to obtain the actual water use *WU* for each layer:

$$WU_i = WUP_i \cdot \frac{SW_i}{0.25 \cdot FC_i} \ , \qquad SW_i \le 0.25 \cdot FC_i \tag{94}$$

$$WU_i = WUP_i \ , \qquad SW_i > 0.25 \cdot FC_i \tag{95}$$

After the calculation of actual water use by plants, the plant transpiration EP is adjusted.

2.2.3 Growth Constraint: Temperature Stress.

The temperature stress factor is calculated as an asymmetrical function, differently for temperature below the optimal temperature TO, and above it. The equation for the temperature stress factor *TS* for temperatures below *TO* is

$$TS = \exp\left(\ln(0.9) \cdot \left(\frac{CTSL \cdot (TO - T)}{T + 1. \cdot 10^{-6}} \right)^2 \right) \qquad T \le TO \tag{96}$$

where *CTSL* is the temperature stress parameter for temperatures below TO, and T is the daily average air temperature in °C. The temperature stress parameter *CTSL* is evaluated as

$$CTSL = \frac{TO + TB}{TO - TB} \tag{97}$$

where *TB* is the base temperature for the crop in °C. Equation 96 assures that TS=0.9 when the air temperature is (TO+TB)/2.

For the temperatures higher than TO

$$TS = \exp\left(\ln(0.9) \cdot \left(\frac{TO - T}{CTSH + 1. \cdot 10^{-6}} \right)^2 \right) \qquad T > TO, \tag{98}$$

where the temperature stress parameter for temperatures higher than TO, *CTSH,* is evaluated as

$$CTSH = 2 \cdot TO - T - TB \tag{99}$$

An example of the temperature stress factor calculated with equations 96 and 98 is shown in **Fig. 2.10**.

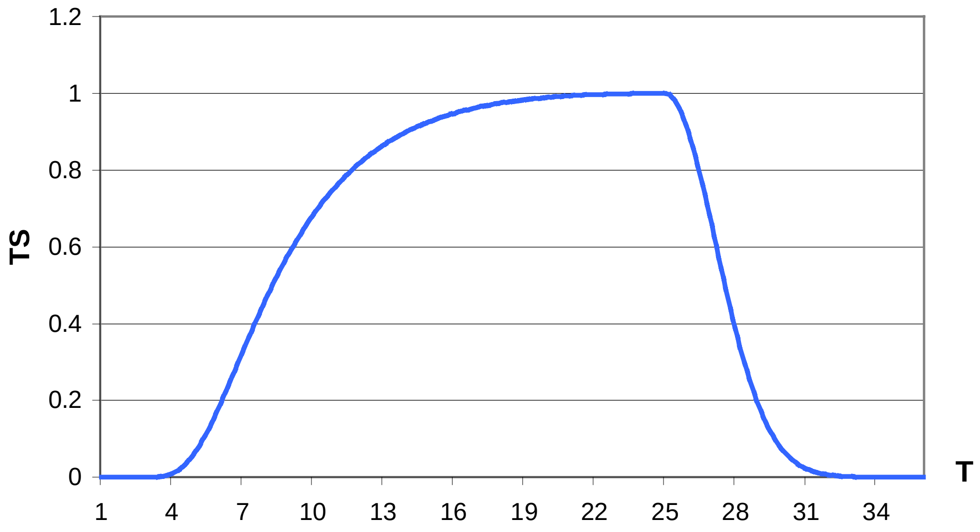

Fig. 2.10 Temperature stress factor as a function of average daily air temperature (equations 96 and 98), assuming TO = 25° C and TB = 3° C

2.2.4 Growth Constraints: Nutrient Stress

Estimation of nutrient stress factors is based on the ratio of simulated plant N and P contents to the optimal values of nutrient content. The stress factors vary non-linearly from 0 when N or P is half the optimal level to 1.0 at optimal N and P contents (Jones et al., 1984).

Let us consider the N stress factor first. As an initial step, the scaling factor SFN is calculated as

$$SFN = 200 \cdot \left(\frac{\sum UN(t)}{CNB \cdot BT} - 0.5 \right) \tag{100}$$

where *UN(t)* is the crop N uptake on day t in kg ha^{-1}, *CNB* is the optimal N concentration for the crop, *BT* is the accumulated total biomass in kg ha^{-1}.

Then the N stress factor is calculated with the equation (see also **Fig. 2.11**)

$$NS = \frac{SFN}{SFN + \exp(3.52 - 0.026 \cdot SFN)}, \qquad if \;\; SFN > 0 \tag{101}$$

The P stress factor, PS, is calculated analogously, using the optimal P concentration, COP, instead.

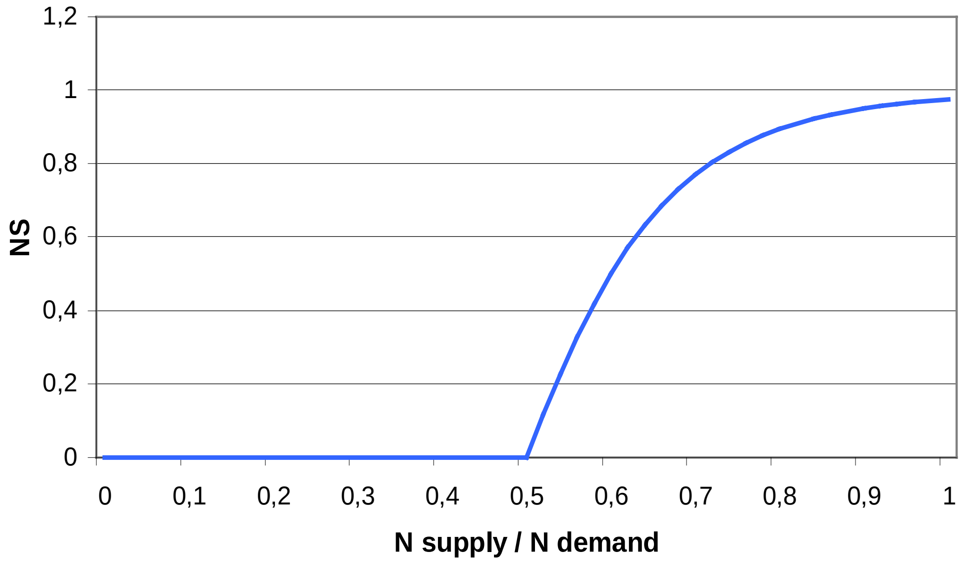

Fig. 2.11 Nitrogen stress factor as a function of N supply and N demand (equations 100 – 101)

2.2.5 Crop Yield and Residue

The economic yield of most crops is a reproductive organ. Harvest index (economic yield divided by aboveground biomass) is often a relatively stable value across a range of environmental conditions. Crop yield is estimated in the model using the harvest index concept

$$YLD = HI \cdot BAG \tag{102}$$

where *YLD* is the crop yield removed from the field in kg ha^{-1}, *HI* is the harvest index at harvest, and *BAG* is the above-ground biomass in kg ha^{-1}.

Harvest index *HIA* increases non-linearly during the growth season and can be estimated as the function of the accumulated heat units

$$\begin{aligned} HIA &= HVSTI \cdot HIC_1 = \\ &= HVSTI \cdot \frac{100 \cdot IHUN}{100 \cdot IHUN + \exp(11.1 - 10 \cdot IHUN)} \end{aligned} \tag{103}$$

where *HVSTI* is the crop-specific harvest index under favourable growing conditions, and HIC_1 is a factor depending on *IHUN* (see **Fig. 2.12**).

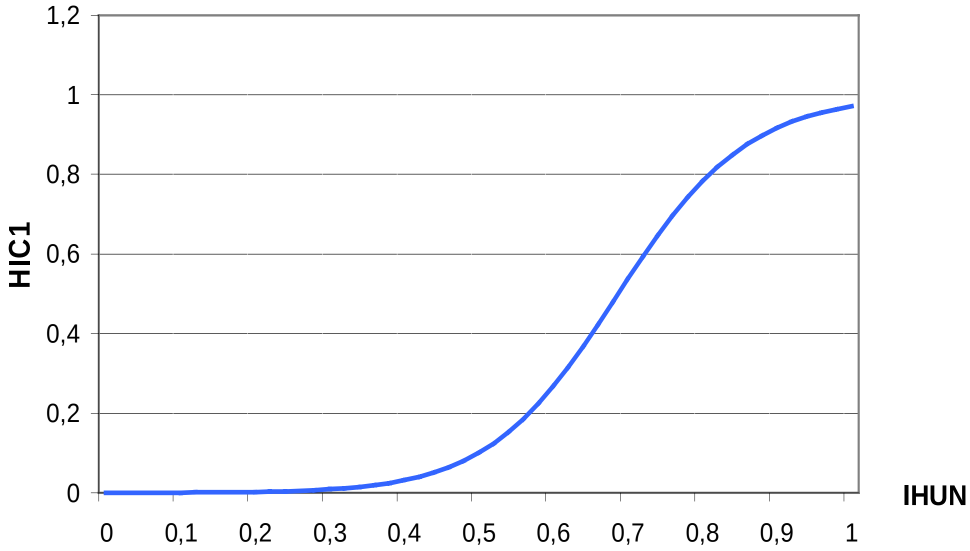

Fig. 2.12 Harvest index as a function of heat unit index (factor HIC_1, equation 103)

The constants in equation 103 are set to allow *HIA* to increase from 0.1 at *IHUN=0.5* to 0.92 at *IHUN=0.9*. This is consistent with economic yield development of crops, which produce most economic yield in the second half of the growing season.

Most crops are particularly sensitive to water stress, especially in the second half of the growing season, when major yield components are determined (Doorenbos and Kassam, 1979). The effect of water stress on the harvest index is described by the following two equations

$$HIAD = HIA \cdot HIC_2 = \\ = HIA \cdot \frac{WSF}{WSF + \exp(6.117 - 0.086 \cdot WSF)} \tag{104}$$

$$WSF = 100 \cdot \frac{SWU}{SWP + 1.e^{-6}} \tag{105}$$

where *HIAD* is the adjusted harvest index, *WSF* is a parameter expressing water supply conditions for crop, HIC_2 is a factor depending on *WSF* (see also factor HIC_2 at **Fig. 2.13**), *SWU* is accumulated actual plant transpiration in the second half of the growing season (*IHUN>0.5*), and SWP is accumulated potential plant transpiration in the second half of the growing season. The harvest index at harvest, *HI* is equal to *HIAD*.

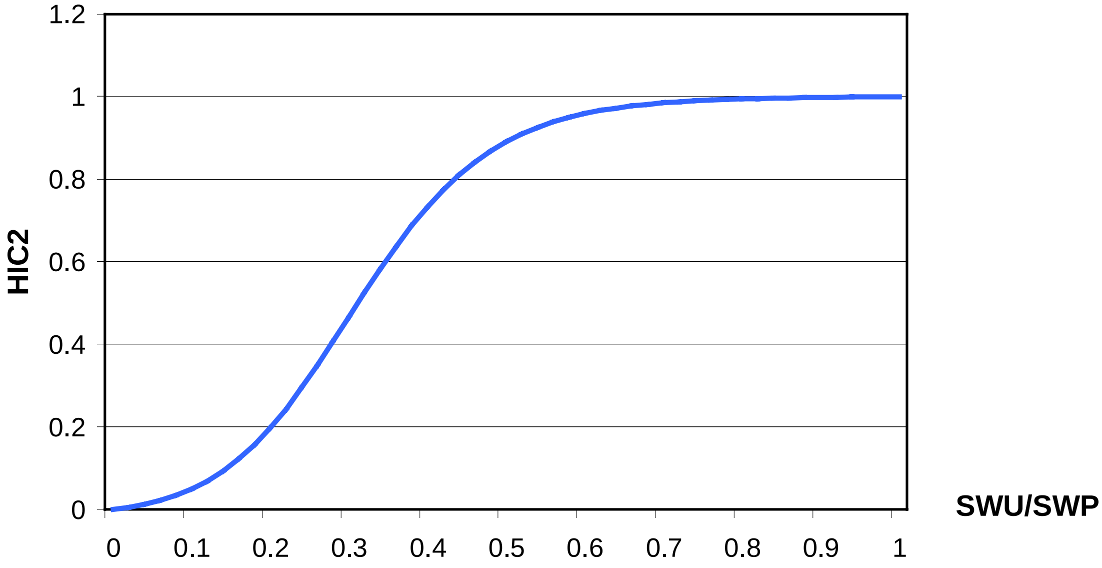

Fig. 2.13 Harvest index as a function of soil water content (factor HIC_2, equation 104)

The residue *RSD* is estimated at harvest as

$$RSD = (1 - RWT) \cdot BT \cdot HI \tag{106}$$

where *RWT* is the fraction of roots, and *BT* is the total biomass. This relationship can be modified for some crops if residue come from the roots.

All processes described in Sections 2.2.1 – 2.2.5 are presented graphically in **Fig. 2.14**. There are three basic blocks in the crop module (depicted by the grey coloured boxes) that are used to estimate the crop yield: accumulated heat units (top middle), stress factors (lower half), and harvest index (top left). The stress factors include temperature stress, nutrient stress (nitrogen and phosphorus), and water stress. The crop growth regulating factor is estimated as the minimum of these four factors. Nutrient stress is determined from the actual and potential nutrient uptake. Water stress is induced from water use and plant transpiration. The heat units accumulation is estimated from the crop specific minimum growth temperature, the daily minimum and maximum air temperatures and the assumed accumulated heat units. The adjusted harvest index is evaluated from the actual and potential transpiration and the crop specific harvest index. The small rectangles denote dependent variables, whereas the coloured ovals refer to model parameters independent from the others computed within the module. They describe the specifications of crop (green), climate (blue) and soil (brown).

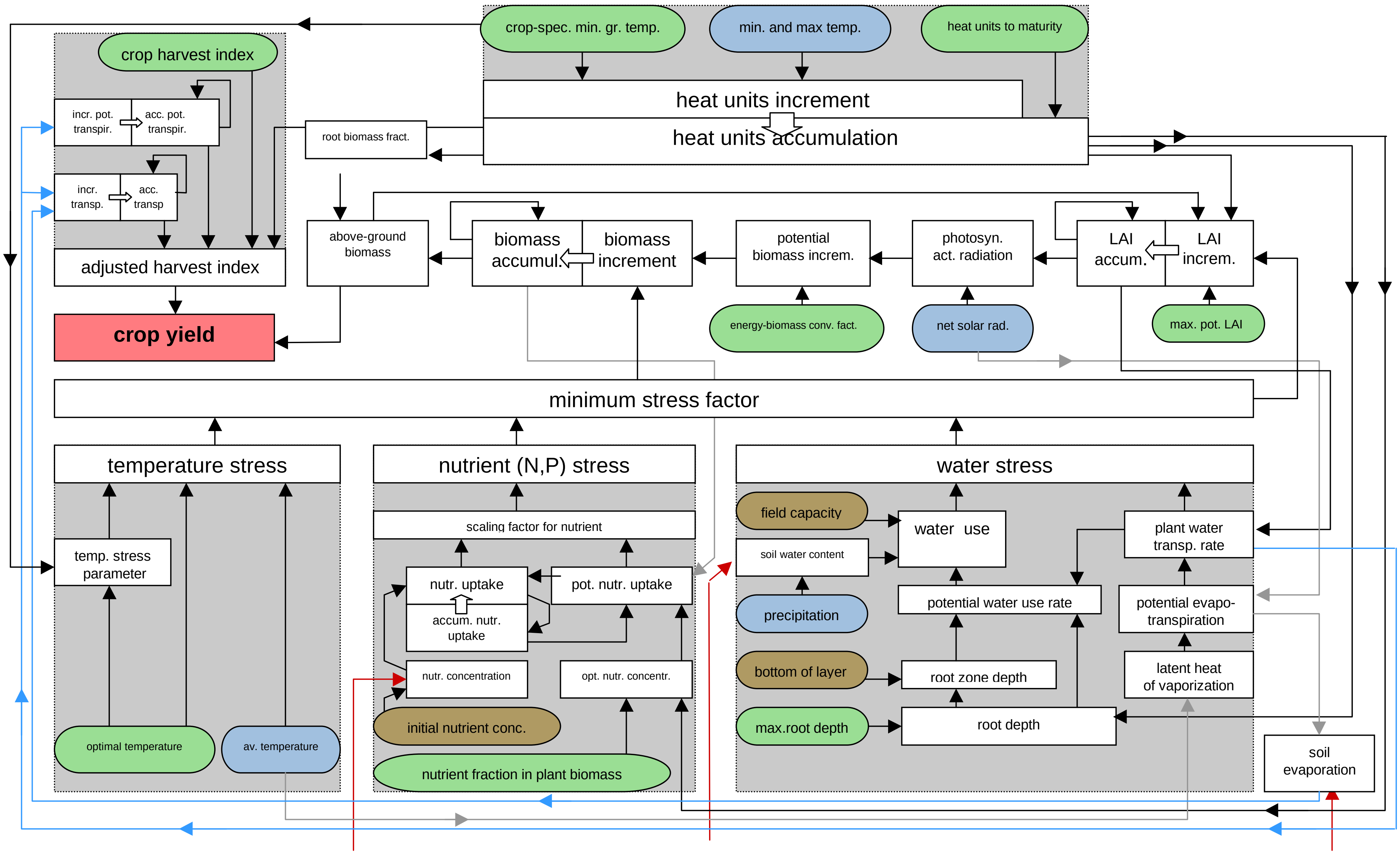

Fig. 2.14 Scheme of operations included in SWIM crop module

2.2.6 Adjustment of Net Photosynthesis to Altered CO_2

Different approaches for the adjustment of net photosynthesis and evapotranspiration to altered atmospheric CO_2 concentration have been used in modelling studies (Goudrian et al., 1984; Rotmans et al., 1993). Detailed results about the interaction of higher CO_2 and water use efficiency are described in (Easmus, 1991; Grossman et al., 1995; Kimball et al., in press).

Two different approaches can be used in SWIM for the adjustment of net photosynthesis (factor *ALFA*):

1) an empirical approach based on adjustment of the biomass-energy factor as suggested in EPIC and SWAT models (Arnold et al., 1994), and
2) a new semi-mechanistic approach derived by F. Wechsung from a mechanistic model for leaf net assimilation (Harley et al., 1992), which takes into account the interaction between CO_2 and temperature.

The second method and its application for climate change impact study with SWIM is described in Krysanova, Wechsung et al., 1999)

The factor ALFA is defined as

$$ALFA = \frac{AS_2}{AS_1} \tag{107}$$

where AS_1 and AS_2 are net leaf assimilation rates (μmol m^{-2} s^{-1}) in two periods, corresponding to two different CO_2 concentrations.

In the first method ALFA is estimated as

$$ALFA = \frac{100 \cdot CA}{BE \cdot \left(CA + \exp\left(SHP_1 - CA \cdot SHP_2\right)\right)} \tag{108}$$

where *BE* is the biomass-energy factor as in equation (83), *CA* is the current atmospheric CO_2 concentration (μmol mol^{-1}), and SHP_1 and SHP_2 are the coefficients of the S-shape curve, describing the assumed change in BE for two different CO_2 concentrations.

For the CO_2 doubling, 1.1 times increase in BE is assumed for maize, and 1.3 times increase for wheat and barley (see **Fig. 2.15**).

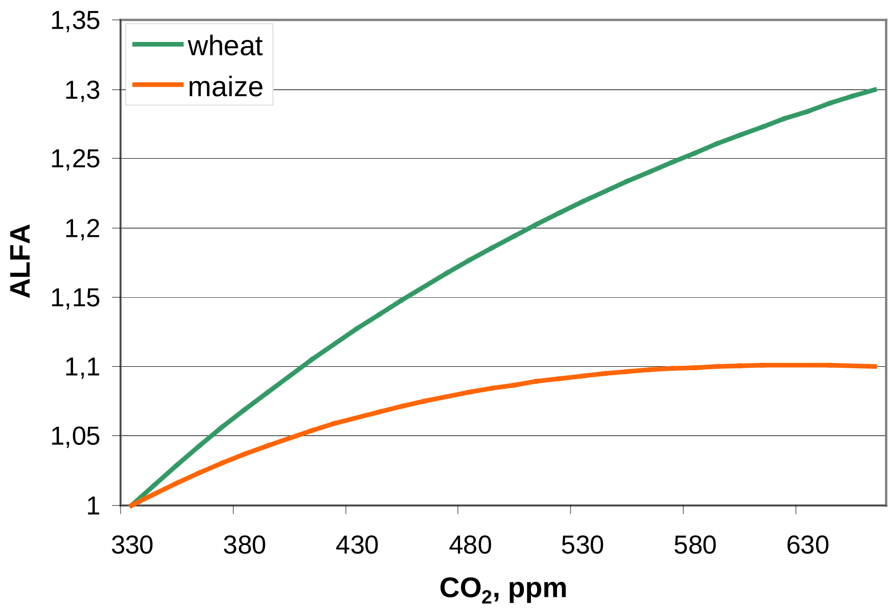

Fig. 2.15 Factor ALFA as a function of CO_2 concentration for wheat and maize estimated using the first method (equations 108, 109, 110) and assuming BE = 30 kg m^2 MJ^{-1} ha^{-1} d^{-1} for wheat and BE = 40 kg m^2 MJ^{-1} ha^{-1} d^{-1} for maize. The CO_2 concentration is changing from 330 to 660 ppm

If CO_2 concentration *CA* is changing from CA_1 to CA_2, and *BE* is changing from BE_1 to BE_2, the coefficients SHP_1 and SHP_2 can be estimated as following:

$$SHP_2 = \frac{\log(100 \cdot CA_1 / BE_1 - CA_1) - \log(100 \cdot CA_2 / BE_2 - CA_2)}{CA_2 - CA_1} \tag{109}$$

$$SHP_1 = \log(100 \cdot CA_1 / BE_1 - CA_1) + CA_1 \cdot SHP_2 \tag{110}$$

In the second method a temperature-dependent enhancement factor α was derived from Harley et al., 1992 for cotton

$$ALFA_{cot} = \\ = \exp\left[P_1 \cdot (CL_2 - CL_1) - P_2 \cdot \left((CL_2)^2 - (CL_1)^2\right) + P_3 \cdot TL \cdot (CL_2 - CL_1)\right] \tag{111}$$

where *TL* is the leaf temperature (°C), CL_1 and CL_2 are the current and future CO_2 concentration inside leaves (μmol mol^{-1}), and coefficients $P_1 = 0.3898 \cdot 10^{-2}$, $P_2 = 0.3769 \cdot 10^{-5}$, and $P_3 = 0.3697 \cdot 10^{-4}$.

It is assumed in the model that the leaf temperature *TL* coincides with the air temperature *TX,* and that the CO_2 concentration inside leaves is a linear function of the atmospheric CO_2 concentration:

$$CL = 0.7 \cdot CA \tag{112}$$

Then the cotton-specific factor ALFA was adjusted for wheat, barley and maize according to the latest crop-specific results reported in the literature (Peart et al., 1989; Kimball et al., in press)

$$ALFA_{wheat} = \left(ALFA_{\mathrm{cot}}\right)^{0.6} \tag{113}$$

$$ALFA_{barley} = \left(ALFA_{\mathrm{cot}}\right)^{0.6} \tag{114}$$

$$ALFA_{maize} = \left(ALFA_{\mathrm{cot}}\right)^{0.36} \tag{115}$$

which imply an increase in leaf net photosynthesis of 31, 31 and 10% for wheat, barley and maize, respectively, if the atmospheric CO_2 increases from 360 to 720 ppm at 20°C and corresponds to the analogous assumption made in the first method. **Fig. 2.16** shows the temperature-dependent ALFA factor for cotton, wheat and maize in the case of CO_2 doubling (a) and in the case of 50% increase in CO_2 (b) assuming CA_1 = 330 ppm estimated with the second method (equations 111,112, 113, 115).

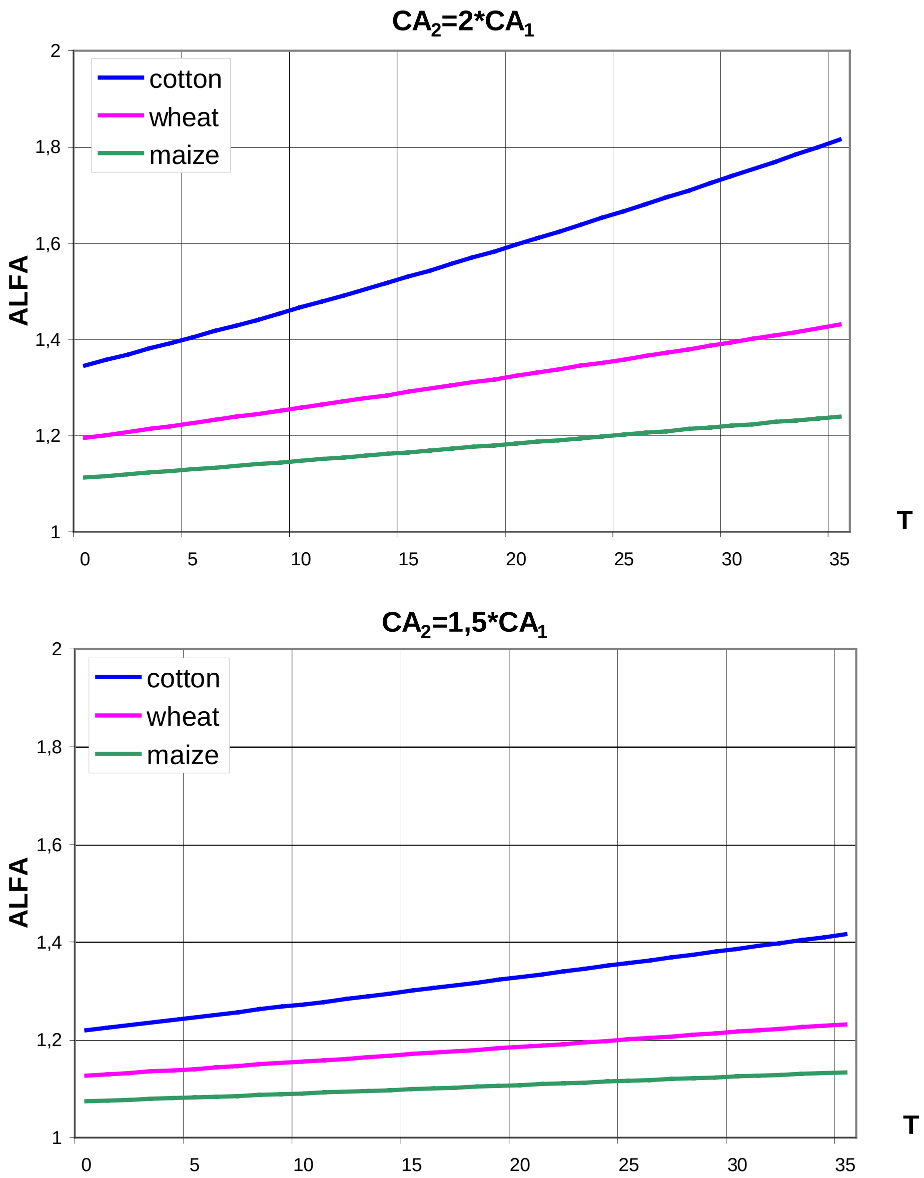

Fig. 2.16 ALFA factor for cotton, wheat and maize as dependent on temperature in the case of CO_2 doubling (a) and in the case of 50% increase in CO_2 (b) assuming initial CO_2 concentration 330 ppm (equations 111 - 115 and 121)

2.2.7 Adjustment of Evapotranspiration to Altered CO_2

Additionally, a possible reduction of potential leaf transpiration due to higher CO_2 (factor *BETA*) derived directly from the enhancement of photosynthesis (factor ALF) was taken into account in combination with both methods for the adjustment of net photosynthesis. The method was suggested by F. Wechsung.

The factor *BETA* is defined as

$$BETA = \frac{EPO_2}{EPO_1} \tag{116}$$

where EPO_1 and EPO_2 are potential plant transpiration rates (mol m^{-2} s^{-1}) in two periods, corresponding to two different CO_2 concentrations.

Assuming that

$$\frac{AS}{EPO} = \frac{CA - CL}{VPD} \cdot \frac{RESW}{RESC} \tag{117}$$

where VPD is the vapour pressure deficit (kPa), RESC is the total leaf resistance to CO_2 transfer (m^2 s mol^{-1}), RESW is the total leaf resistance to water vapour transfer (m^2 s mol^{-1}).

From definitions 107 and116 and equation 117 the ratio can be estimated

$$\frac{ALFA}{BETA} = \frac{CA_2 - CL_2}{CA_1 - CL_1} \cdot \frac{VPD_1}{VPD_2} \cdot \frac{RESW_2}{RESC_2} \cdot \frac{RESC_1}{RESW_1} \tag{118}$$

The following assumptions can be accepted for a given plant (see, e.g. Morrison, 1993)

$$\frac{RESW_2}{RESC_2} \approx \frac{RESW_1}{RESC_1} \tag{119}$$

and

$$VPD_2 \approx VPD_1 \tag{120}$$

Then the following estimation is derived for BETA from equations 112, 118, 119 and 120

$$BETA = ALFA \cdot \frac{CA_1}{CA_2} \tag{121}$$

Jarvis and McNaughton (1986) postulate that on the regional scale there is no control of stomatal resistance on evapotranspiration, because the humidity profiles are adjusted within the planetary boundary layer. This response would counter stomatal closure as a negative feedback. On the other hand, recent model studies suggest that stomata have far more control on regional and global evapotranspiration than postulated by Jarvis and McNaughton (Kimball et al., 1995).

Simulation runs with SWIM, which included the CO_2 fertilization effect on crops, have been carried out (Krysanova, Wechsung et al., 1999) applying both methods for ALFA factor in two variants: without and with factor BETA. In this way it is possible to account for current uncertainty regarding significance of stomatal effects on higher CO_2 for regional evapotranspiration. The comparison of two methods for estimation of ALFA factor is shown in **Fig. 2.17.**

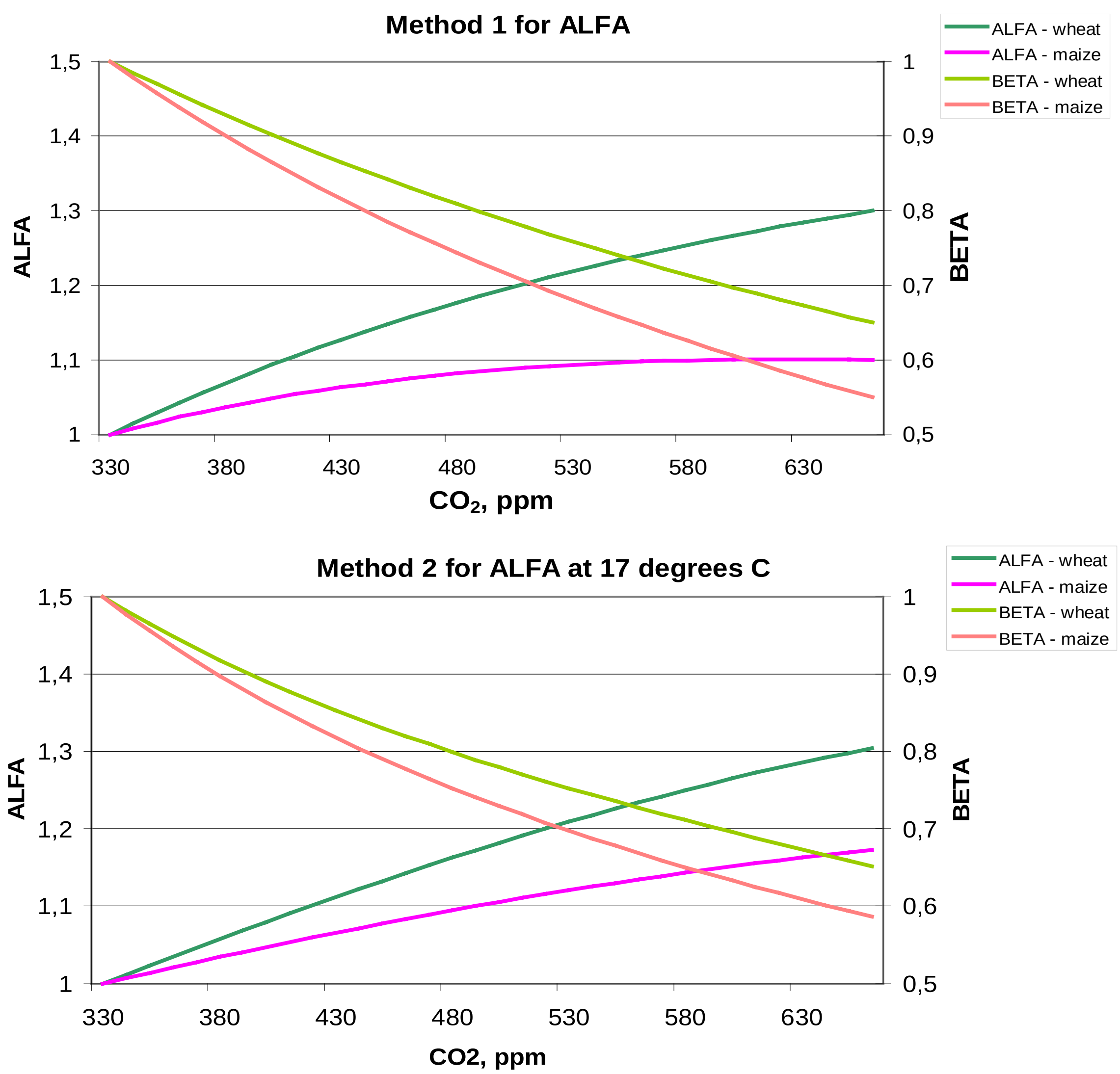

Fig. 2.17 Comparison of two methods for the estimation of ALFA and BETA factors: ALFA and BETA as functions of CO_2 concentration under assumption that temperature is 17° C for the second method

2.3 Nutrient Dynamics

Sub-basin nutrient cycling modules were taken from MATSALU and SWAT, and modified where necessary. The approach used in SWAT was modified from the EPIC model (Williams et al., 1984). The model simulates water, sediment and nutrients dynamics in every hydrotope, aggregates results for sub-basins, and then routes the water, sediment, and nutrients with lateral flow from the sub-basin outlet to the basin outlet.

2.3.1 Soil Temperature

Several processes of nutrient transformation, like mineralisation, are of microbial character, therefore estimation of soil temperature is necessary. Daily average soil temperature is defined at the center of each soil layer. The basic soil temperature equation is

$$TSO(Z,t) = TAV + \frac{AMP}{2} \cdot \cos\left(\frac{2 \cdot \pi}{365} \cdot (t - 200) - \frac{Z}{DD}\right) \cdot \exp\left(-\frac{Z}{DD}\right) \tag{122}$$

where *TSO(Z,t)* is the soil temperature at the depth *Z* in the day *t* in °C, *Z* is depth from the soil surface in mm, *t* is time d, *TAV* is the average annual air temperature in °C, *AMP* is the annual amplitude in daily average temperature in °C, and *DD* is the damping depth for the soil in mm.

The damping depth *DD* can be defined as a function of soil bulk density *BD* and water content *SW* as expressed in the following equations

$$DD = DP \cdot \exp\left[\ln\left(\frac{500}{DP}\right) \cdot \left(\frac{1 - SPD}{1 + SPD}\right)^2\right] \tag{123}$$

$$DP = 1000 + \frac{2500 \cdot BD}{BD + 686 \cdot \exp(-5.63 \cdot BD)} \tag{124}$$

$$SPD = \frac{SW}{(0.356 - 0.144 \cdot BD) \cdot ZM} \tag{125}$$

where *DP* is the maximum damping depth for the soil in mm, *BD* is the soil bulk density in t m^{-3}, *ZM* is the distance from the bottom of the lowest soil layer to the surface in mm, and *SPD* is a scaling parameter.

Equation (122) reflects average conditions, if only TAV and AMP parameters are used. Since air temperature is provided as input, the soil temperature module can use the air temperature as driver to correct equation 122.

First, the bare soil surface temperature is estimated as

$$TGB(t) = WFT \cdot (TMX - T) + T \ , \qquad PRECIP = 0 \tag{126}$$

$$TGB(t) = WFT \cdot (T - TMN) + TMN \ , \qquad PRECIP > 0 \tag{127}$$

where *TGB(t)* is the bare soil surface temperature in °C in the day t, *TMX, T,* and *TMN* are the maximum, average and minimum daily air temperature in °C, and *WFT* is a proportion of rainy days in a month.

Equation 127 uses the minimum air temperature as a base to estimate surface temperature on rainy days. Higher temperatures are estimated on dry days using equation 126. The value of *WFT* is determined by considering the number of rainy days in this month:

$$WFT = \frac{NRD}{NDD} \tag{128}$$

where *NDD* is the number of days in a month, and *NRD* is the number of rainy days in a month.

The soil surface temperature is also affected by residue and snow cover. This effect is introduced by lagging the predicted base surface temperature with the equation

$$TG(t) = BCV \cdot TGB(t-1) + (1 - BCV) \cdot TGB(t) \tag{129}$$

where *BCV* is a lagging factor for simulating residue and snow cover effects on surface temperature. The value of BCV is 0 for bare soil and approaches 1.0 as cover increased as expressed in the equation

$$BCV = \max\left\{ \begin{matrix} \dfrac{COV}{COV + \exp\left(7.563 - 1.297 \cdot 10^{-4} \cdot COV\right)}, \\ \dfrac{SNO}{SNO + \exp\left(6.055 - 0.3022 \cdot SNO\right)} \end{matrix} \right\} \tag{130}$$

where *COV* is the land cover, or the sum of above ground biomass and crop residue in kg ha^{-1} and *SNO* is the water content of the snow cover in mm.

Then the soil temperature at any depth is estimated with equation 122 by substituting *TG(t)* for *TS(0,t). TG(t)* is a better estimate of the surface temperature than *T(0,t),* because current weather and cover conditions are considered. At the soil surface (*Z*=0), the proper substitution can be accomplished by adding *TG(t)* and subtracting *TS(0,t)* from equation 122. Differences between *TG(t)* and *TS(0,t)* are damped as *Z* increases. So, the final equation for estimating soil temperature at any depth is

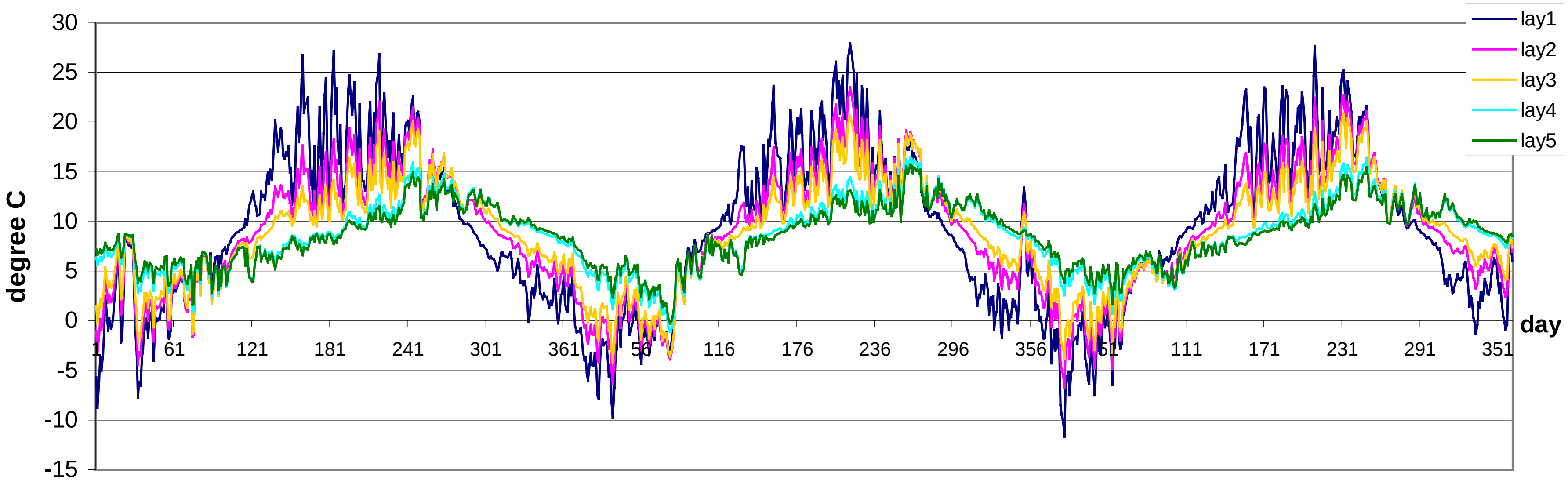

Fig. 2.18 An example of soil temperature dynamics in five soil layers simulated with SWIM using equation 131

$$TSO(Z,t) = TAV + \\ + \left(\frac{AMP}{2} \cdot \cos\left(\frac{2 \cdot \pi}{365} \cdot (t - 200) - \frac{Z}{DD} \right) + TG(t) - TS(0,t) \right) \cdot \exp\left(- \frac{Z}{DD} \right) \quad (131)$$

An example of soil temperature dynamics as simulated by SWIM using equation 131 is shown in **Fig. 2.18**.

2.3.2 Fertilization and Input with Precipitation

Fertilization in form of mineral and active organic N and P is treated as input information in SWIM. The amounts and dates should be specified in advance. The amounts of fertilizers applied can be either strict or calculated values, depending on whether the strict or flexible fertilization scheme is applied. In the latter case the amounts of applied N and P depend on the actual concentration of mineral N and P in soil.

To estimate the N contribution from rainfall, SWIM uses an average rainfall N and P concentration, specific for the region. The amount of N and P in precipitation is estimated as the product of rainfall amount and concentration.

2.3.3 Nitrogen Mineralisation

The nitrogen mineralisation model is a modification of the PAPRAN mineralisation model (Seligman and van Keulen, 1981). The model considers two sources of mineralisation:
(a) fresh organic N pool, associated with crop residue, and
(b) the active organic N pool, associated with the soil humus.

Step 1. When the model is initialized, organic N associated with humus is divided into two pools: active or readily mineralisable organic nitrogen *ANOR* and stable organic nitrogen *SNOR* (in kg ha^{-1}) by using the equation

$$ANOR = ANFR \cdot NOR \quad (132)$$

where *ANFR* is the active pool fraction (set to 0.15), *NOR* is the total organic N in kg ha^{-1} estimated from the initial soil data.

Organic N flow between the active and stable pools is described with the equilibrium equation

$$ASNFL = CASN \cdot \left[\frac{ANOR}{ANFR} - SNOR \right] \quad (133)$$

where *ASNFL* is the flow in kg ha^{-1} d^{-1} between the active and stable organic N pools, *CASN* is the rate constant (10^{-4} d^{-1}). The daily flow of humus-related organic N, *ASNFL*, is added to the stable organic N pool and subtracted from the active organic N pool.

Step 2. The residue is decomposed daily in accordance with the equation

$$RSD = RSD \cdot (1 - DECR) \tag{134}$$

where *DECR* is the decomposition rate. Fresh organic N pool *FON* is associated with residue. It is recalculated with the same equation daily:

$$FON = FON \cdot (1 - DECR) \tag{135}$$

and N mineralisation flow from fresh organic N in kg ha^{-1} d^{-1}, *FOMN*, is estimated as

$$FOMN = DECR \cdot FON \tag{136}$$

The decomposition rate *DECR* is a function of C:N ratio, C:P ratio, temperature, and water content in soil

$$DECR = 0.05 \cdot \min(CNRF, CPRF) \cdot \sqrt{TFM_2 \cdot WFM} \tag{137}$$

where *CNRF* and *CPRF* are the C:N and C:P ratio factors of mineralisation, respectively, and TFM_2 and *WFM* are the temperature and soil water factors of mineralisation, respectively. The values of CNRF and CPRF are calculated with the equations

$$CNRF = \exp\left[-\frac{0.693 \cdot (CNR - 25)}{25}\right] \tag{138}$$

$$CPRF = \exp\left[-\frac{0.693 \cdot (CPR - 200)}{200}\right] \tag{139}$$

where *CNR* is the C:N ratio and *CPR* is the C:P ratio.
The CNR and CPR are calculated with the equations

$$CNR = \frac{0.58 \cdot RSD}{FON + NMIN} \tag{140}$$

$$CPR = \frac{0.58 \cdot RSD}{FOP + PLAB} \tag{141}$$

where *FON* is the amount of fresh organic N in kg ha^{-1} , *FOP* is the amount of fresh organic P in kg ha^{-1}, *NMIN* is the amount of mineral nitrogen (or nitrate nitrogen plus ammonium nitrogen) in kg ha^{-1}, and *PLAB* is the amount of labile P in kg ha^{-1}.

The temperature factor in 137 is expressed by the equation (see also **Fig. 2.19**)

$$TFM_2 = \frac{TSO(2,t)}{TSO(2,t) + \exp[6.82 - 0.232 \cdot TSO(2,t)]} \tag{142}$$

where *TSO(2,t)* is soil temperature in the second soil layer in °C (the depth of first layer is 10 mm). The soil water factor considers the relation of total soil water to field capacity

$$WFM = \frac{SW}{FC} \tag{143}$$

The N mineralisation flow from residue, *FOMN*, calculated by equation 136 is distributed between mineral nitrogen and active organic nitrogen pools in the proportion 4:1.

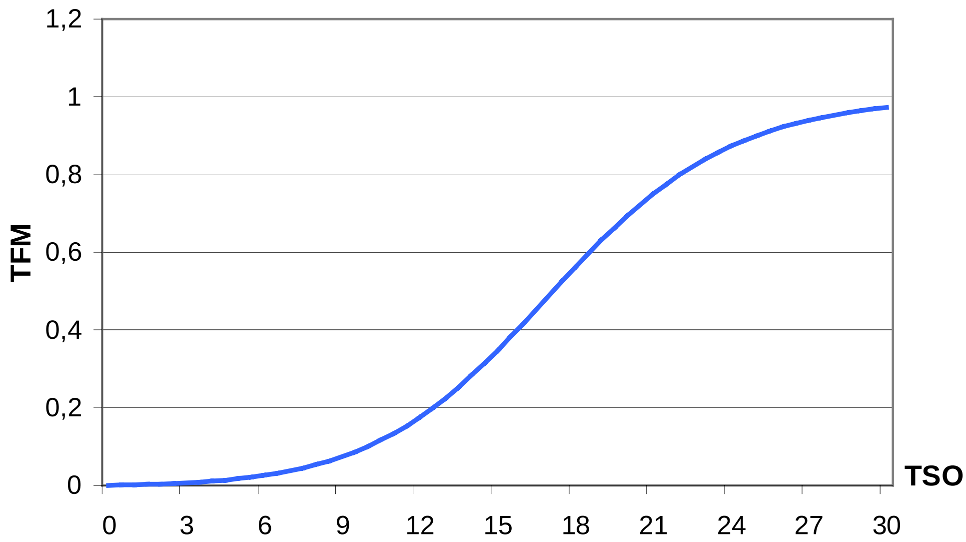

Fig. 2.19 Temperature factor of mineralisation, TFM (equation 142)

Step 3. The stable organic N pool is not subjected to mineralisation. Only the active pool of organic N in soil is exposed to mineralisation. The mineralisation from the active organic N is expressed by the equation

$$HUMN_i = COMN \cdot \sqrt{TFM_i \cdot WFM_i} \cdot ANOR_i \qquad (144)$$

where $HUMN_i$ is the mineralisation rate in kg ha^{-1} d^{-1} for the active organic N pool in layer i, *COMN* is the humus rate constant for N (0.0003 d^{-1}), and TFM_i and WFM_i are the temperature and water factors of mineralisation for the layer i.

The temperature and water factors are calculated for any soil layer the same as for residue decomposition using equations 142 and 143. At the end of the day, the humus mineralisation is subtracted from the active organic N pool and added to the mineral N pool.

2.3.4 Phosphorus Mineralisation

The phosphorus mineralisation model is structurally similar to the nitrogen mineralisation model, with some differences as explained below.

Step 1. Fresh organic P pool *FOP* is associated with residue. It is recalculated daily as

$$FOP = FOP \cdot (1 - DECR) \qquad (145)$$

Then the P mineralisation flow from fresh organic P in kg ha^{-1} d^{-1}, *FOMP*, is estimated as

$$FOMP = DECR \cdot FOP \qquad (146)$$

where the rate DECR is calculated the same as for nitrogen using equation 137.

Step 2. Mineralisation of organic P associated with humus is estimated for each soil layer with the following equation

$$HUMP_i = COMP \cdot \sqrt{TFM_i \cdot WFM_i} \cdot POR_i \qquad (147)$$

where $HUMP_i$ is the mineralisation rate in kg ha^{-1} d^{-1} i, *COMP* is the humus mineralisation rate constant for P, and POR is the P organic pool in soil layer i.

To maintain the P balance at the end of a day, the mineralized humus is subtracted from the organic P pool and added to the mineral P pool, and the mineralized residue is subtracted from the FOP pool. Then 1/5 of *FOMP* is added to the *POR* pool, and 4/5 of *FOMP* is added to labile P pool, *PLAB*.

2.3.5 Phosphorus Sorption / Adsorption

Mineral phosphorus is distributed between three pools: labile phosphorus, PLAB, active mineral phosphorus, PMA and stabile mineral phosphorus, PMS. Mineral P flow between the active and stable mineral pools is governed by the equilibrium equation

$$ASPFL = CASP \cdot (4 \cdot PMA - PMS) \tag{148}$$

where *ASPFL* is the flow in kg ha^{-1} d^{-1} between the active and stable mineral P pools, *CASP* is the rate constant (0.0006 d^{-1}). The daily flow *ASPFL* is added to the stable mineral pool and subtracted from the active mineral pool.

Mineral P flow between the active and labile mineral pools is governed by the equilibrium equation

$$ALPFL = PLAB - CALP \cdot PMA \tag{149}$$

where *ALPFL* is the flow in kg ha^{-1} d^{-1} between the active and labile mineral P pools, *CALP* is the equilibrium constant (default: 1.). The daily flow *ALPFL* is added to the active mineral pool and subtracted from the labile mineral pool.

2.3.6 Denitrification

Denitrification causes NO_3 to be volatilized from soil. The denitrification occurs only in the conditions of oxygen deficit, which usually is associated with high water content. Besides, as one of the microbial processes, denitrification is a function of temperature and carbon content. The equation used to estimate the denitrification rate is

$$\begin{aligned} DENIT_i &= WFD_i \cdot TCFD_i \cdot NIT_i, & SW_i / FC_i \geq 0.9 \\ DENIT &= 0. & SW_i / FC_i < 0.9 \end{aligned} \tag{150}$$

where *DENIT* is the denitrification flow in layer *i* in kg ha^{-1} d^{-1}, *WFD* is the soil water factor of denitrification, and *TCFD* is the combined temperature-carbon factor.

The soil water factor considers total soil water and is represented by the exponential equation (see **Fig. 2.20**)

$$WFD_i = 0.06 \cdot \exp\left(\frac{3 \cdot SW_i}{FC}\right) \tag{151}$$

where SW_i is the soil water content in layer i in mm and FC_i is the field capacity in mm mm^{-1}.

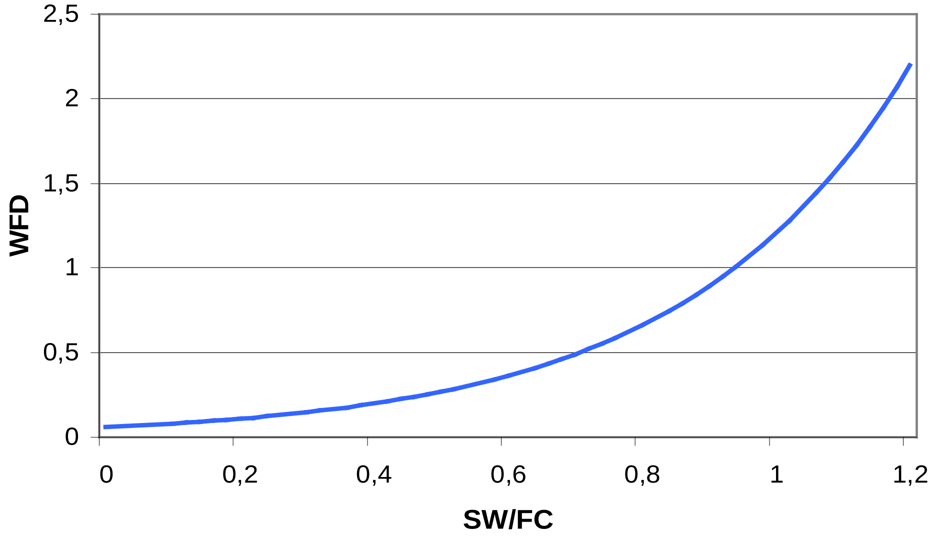

Fig. 2.20 Soil water factor of denitrification (equation 151)

The combined temperature and carbon factor is expressed by the equation

$$TCFD_i = 1 - \exp\left(CDN \cdot TFM_i \cdot CBN_i\right) \tag{152}$$

where *CDN* is a shape coefficient, TFM_i coincides with the temperature factor of mineralisation, and CBN_i is the carbon content, and subscript i refers to the layers.

2.3.7 Nutrient Uptake by Crops

Nitrogen uptake by crop is estimated using a supply and demand approach. The daily (day t) crop N demand can be computed using the equation

$$NDEM(t) = CNB(t) \cdot BT(t) - CNB(t-1) \cdot BT(t-1) \tag{153}$$

where *NDEM(t)* is the N demand of the crop in kg ha^{-1}, *CNB(t)* is the optimal N concentration in the crop biomass, and BT(t) is the accumulated biomass in kg ha^{-1}. Three parameters BN_1, BN_2, and BN_3 are specified for every crop in the crop database, which describe: BN_1 - normal fraction of nitrogen in plant biomass excluding seed at emergence, BN_2 – at 0.5 maturity, and BN_3 - at maturity.

Then the optimal crop N concentration is calculated as a function of growth stage using the equation (see also **Fig. 2.21**)

$$CNB = (BN_1 - BN_3) \cdot \left[1 - \frac{IHUN}{IHUN + \exp(SP_1 - SP_2 \cdot IHUN)}\right] + BN_3 \qquad (154)$$

where SP1 and SP2 are shape parameters assuring the definition above, and IHUN(t) is the heat unit index expressing the fraction of the growing season as calculated in equation 81. The crop is allowed to take nitrogen from any soil layer that has roots. Uptake starts at the upper layer and proceeds downward until the daily demand is met or until all N has been depleted.

The same approach is used to estimate P uptake, differing only by the parameter values.

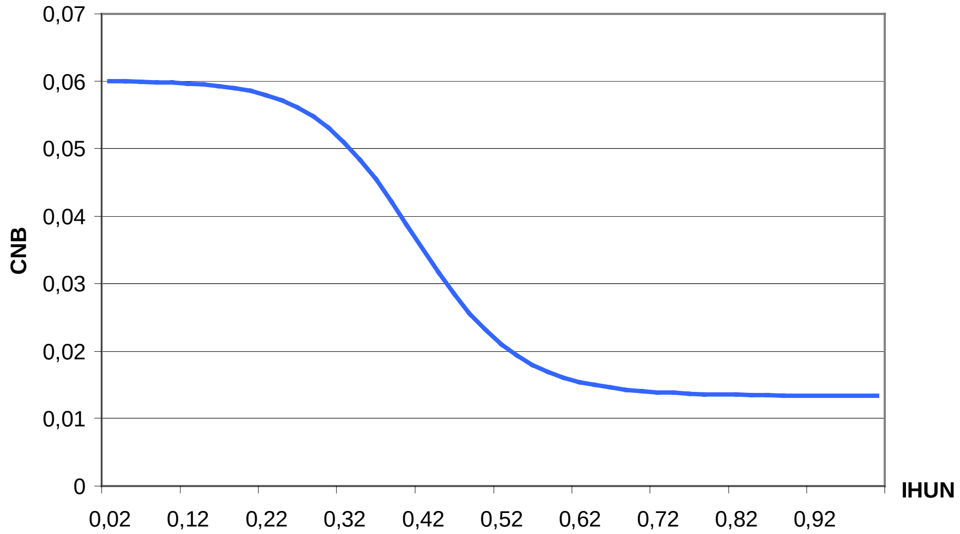

Fig. 2.21 The optimal crop N concentration, CNB, as a function of growth stage IHUN (equation 154) assuming BN1 = 0.06 g g^{-1}, BN2 = 0.0231 g g^{-1} and BN3 = 0.0134 g g^{-1}

2.3.8 Nitrate Loss in Surface Runoff and Leaching to Groundwater

The total amount of water lost from the soil layer i is the sum of surface runoff, lateral subsurface flow (or interflow), and percolation from this layer:

$$WTOT_i = Q_i + SSF_i + PERC_i \tag{155}$$

where *WTOT* is the total water lost from the soil layer in mm, Q is the surface runoff in mm, *SSF* is the lateral subsurface flow in mm, and *PERC* is the percolation in mm, and i is the layer.

The amount of nitrate nitrogen lost with $WTOT_i$ is the product of NO_3-N concentration and water loss as expressed by the equation

$$NFL_i = WTOT_i \cdot CON_i \tag{156}$$

where NFL_i is the amount NO_3-N lost from the layer *i* in kg ha^{-1} and CON_i is the concentration of NO_3-N in the layer *i* in kg ha^{-1}.

The amount of NO_3-N left in the layer is adjusted daily as

$$NMIN(t) = NMIN(t-1) - WTOT_i \cdot CON_i \tag{157}$$

where *NMIN(t-1)* and *NMIN(t)* are the amounts of NO_3-N contained in the layer at the beginning and end of the day (in kg ha^{-1}).

Then the NO_3-N concentration can be estimated by dividing the weight of NO_3-N by the water storage in the layer:

$$CON_i(t) = CON_i(t-1) - CON_i(t-1) \cdot \left(- \frac{WTOT_i}{PO_i - WP_i} \right) \tag{158}$$

where $CON_i(t)$ is the concentration of NO_3-N at the end of the day in kg ha^{-1}, *PO* is the soil porosity in mm mm^{-1}, and *WP* is the wilting point water content for soil layer in mm mm^{-1}.

Equation 158 is a finite different approximation of the exponential equation

$$CON_i(t) = CON_i(t-1) - \exp\left(- \frac{WTOT_i}{PO_i - WP_i} \right) \tag{159}$$

Then the integration of equation 159 allows to calculate *NFL* for any WTOT value

$$NFL_i = NMIN_i \cdot CW_i = NMIN_i \cdot \left(1 - \exp\left(- \frac{WTOT}{PO_i - WP_i} \right) \right) \qquad (160)$$

The coefficient CW as the function of relative water content is depicted in **Fig. 2.22**. The average concentration for the day is

$$CON_i = \frac{NFL_i}{WTOT_i} \qquad (161)$$

Amounts of NO_3-N contained in surface runoff, lateral subsurface flow, and percolation are estimated as the products of the volume of water and the concentration with equation 161.

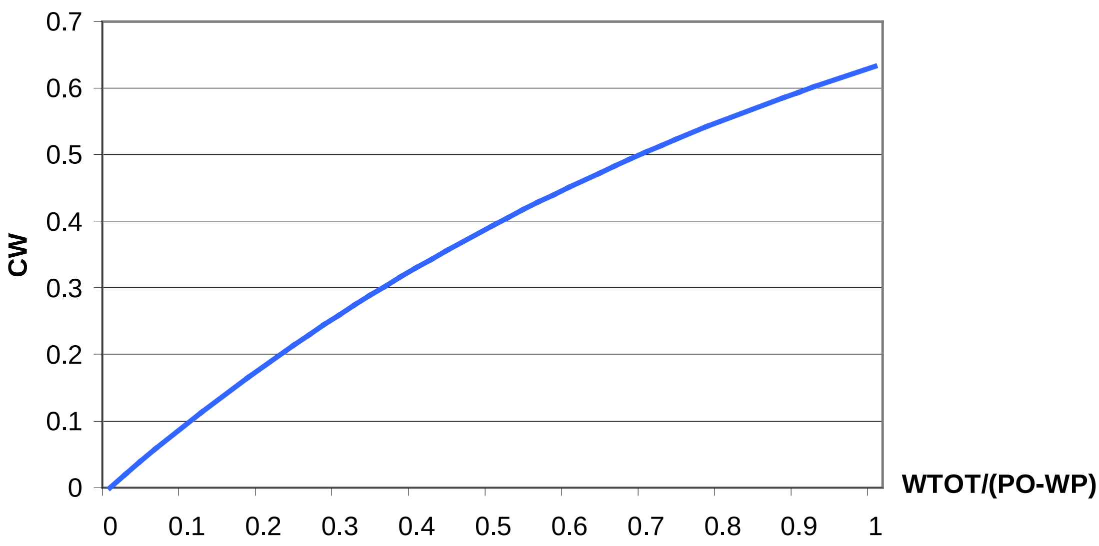

Fig. 2.22 Coefficient CW to calculate the amount NO_3-N lost from the layer as a function of water content (equation 160)

2.3.9 Soluble Phosphorus Loss in Surface Runoff

Phosphorus in soil is mostly associated with the sediment phase. Therefore the soluble P runoff equation can be expressed in the simple form

$$PFL = \frac{0.01 \cdot COP \cdot Q}{CSW} \tag{162}$$

where *PFL* is the soluble P in kg ha^{-1} d^{-1} lost with surface runoff, *Q* is the surface runoff in mm, *COP* is the concentration of labile phosphorus in soil layer in g t^{-1}, and *CSW* is the P concentration in the sediment divided by that of the water in m^3 t^{-1}. The value of COP is input to the model and remains constant. The default value of CSW used in the model is 175.

All processes described in Sections 2.3.1 – 2.3.9 are presented graphically in **Figs. 2.23** (for nitrogen cycle) and **2.24** (for phosphorus).

The nitrogen module operates with four main pools depicted by the blue rectangles in **Fig. 2.23**: nitrate, stable organic N, mineralisable organic N and fresh organic N (crop residue). The nitrate pool is influenced by the following flows (depicted as flags): N fertilizer application, N precipitation input, N leaching, potential N uptake by plants and denitrification. The latter one is subject to the impact of the following variables and parameters: soil water content, field capacity, shape coefficient, temperature factor of mineralisation and carbon content. The exchange between stable and mineralisable organic nitrogen pools, whose intensity depends on the size of these pools and the rate constant, is shown on the right-hand side. The mineralisation is a function of soil temperature, soil water content, field capacity and the humus rate constant.

The phosphorus module (**Fig. 2.24**) consists of five pools, namely fresh organic P (crop residue), organic P, labile P, active and stable mineral P. Labile P is influenced by the following five flows: decomposition, mineralisation, potential nutrient uptake, P loss by leaching and P exchange with the active mineral phosphorus pool. The size of the latter two flows is modulated by the amount of P in the concerned pools. The two-directional influence we meet in the case of the exchange flow between active and stable mineral P, and the mineralisation and decomposition flows (also pictured as flags). Mineralisation, decomposition, and soil erosion control the amount of organic P. The same as for the nitrogen cycle, mineralisation is influenced by soil temperature, soil water content, field capacity and the humus rate constant, whereas the decomposition rate essentially depends on the C-N-ratio, C-P-ratio and soil temperature.

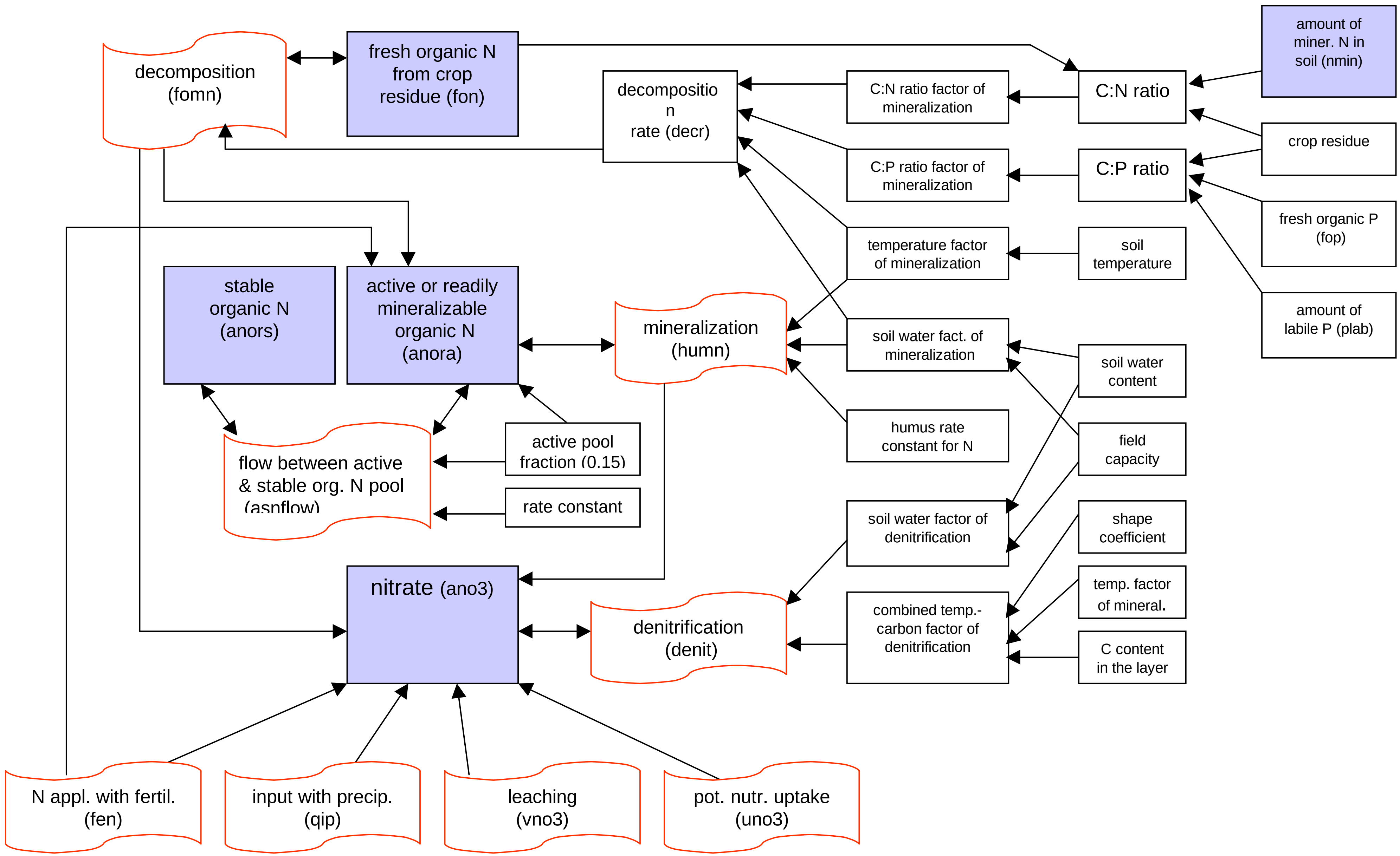

Fig. 2.23 Scheme of operations included in SWIM nitrogen module

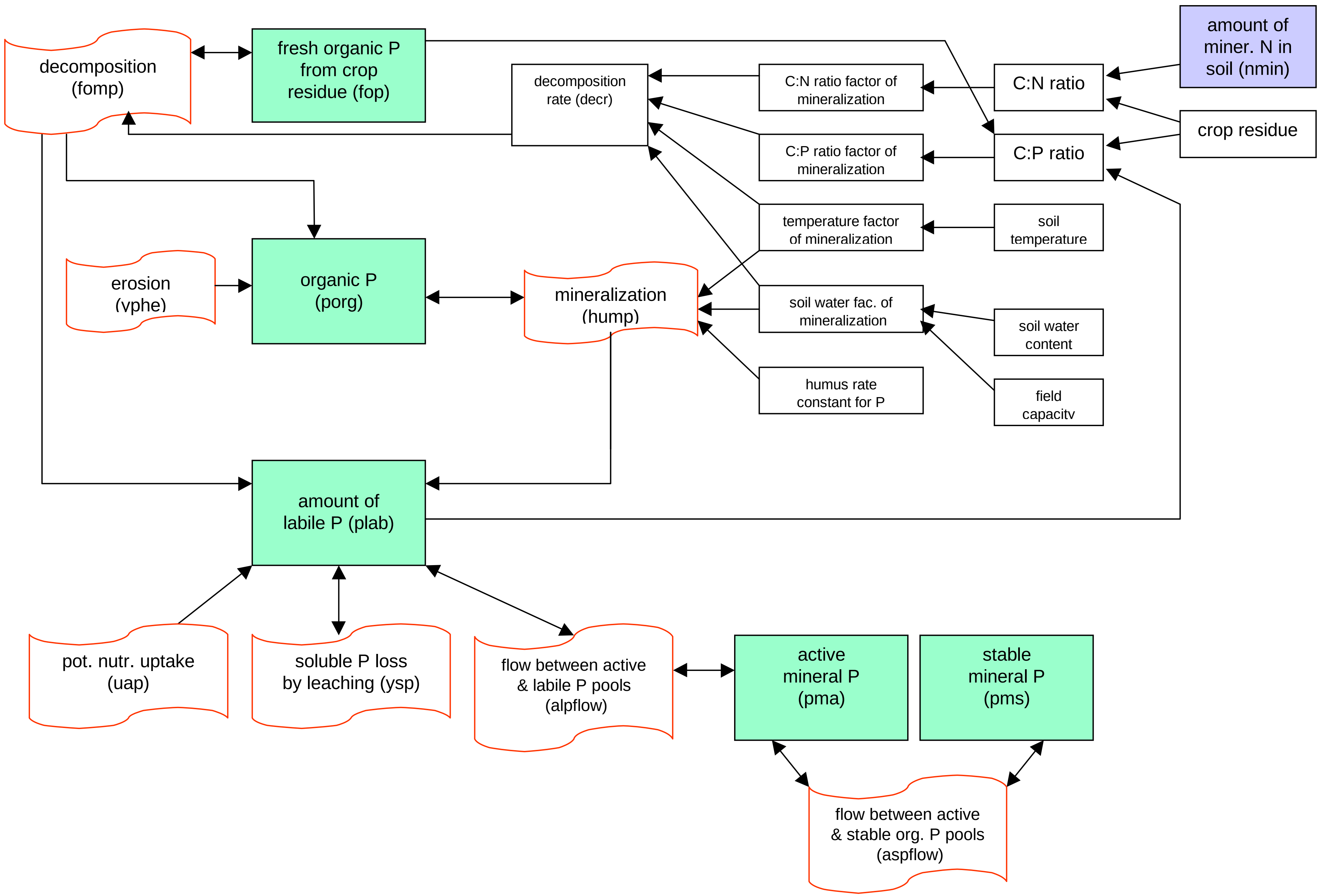

Fig. 2.24 Scheme of operations included in SWIM phosphorus module

2.4 Erosion

2.4.1 Sediment Yield

Sediment yield is calculated for each sub-basin with the Modified Universal Soil Loss Equation (MUSLE) (Williams and Berndt, 1977), practically the same as in SWAT:

$$YSED = 11.8 \cdot (VOLQ \cdot PEAKQ)^{0.56} \cdot K \cdot C \cdot ECP \cdot LS \tag{163}$$

where *YSED* is the sediment yield from the sub-basin in t, *VOLQ* is the surface runoff column for the sub-basin in m^3, *PEAKQ* is the peak flow rate for the sub-basin in $m^3\,s^{-1}$, K is the soil erodibility factor, *C* is the crop management factor, *ECP* is the erosion control practice factor, and *LS* is the slope length and steepness factor.

The only difference between SWAT and SWIM in the erosion module is that the surface runoff, the soil erodibility factor K and the crop management factor C are estimated in SWIM for every hydrotope, and then averaged for the sub-basin (weighted areal average). In SWAT there are two options: option 1 based on two-level disaggregation “basin – sub-basins”, when the above mentioned factors are first estimated for the sub-basins, and option 2 similar to that of SWIM, when the factors are estimated first for HRUs (Hydrologic Response Units).

The soil erodibility factor K is estimated from the texture of the upper soil layer or is taken from a database.

The crop management factor, C, is evaluated with the equation,

$$C = \exp\left[CMN + (-0.2231 - CMN) \cdot \exp(-0.00115 \cdot COV)\right] \tag{164}$$

where *COV* is the soil cover (above ground biomass + residue) in kg ha^{-1} and *CMN* is the minimum value of C.

The value of *CMN* is estimated from the average annual value of C factor, *CAV*, using the equation

$$CMN = 1.463 \cdot \ln(CAV) + 0.1034 \tag{165}$$

The value of *CAV* for each crop is determined from tables prepared by Wischmeier and Smith (1978).

The erosion control practice is estimated as default value of 0.5, if no other data are available (which is usually the case for mesoscale basins and regional case studies).

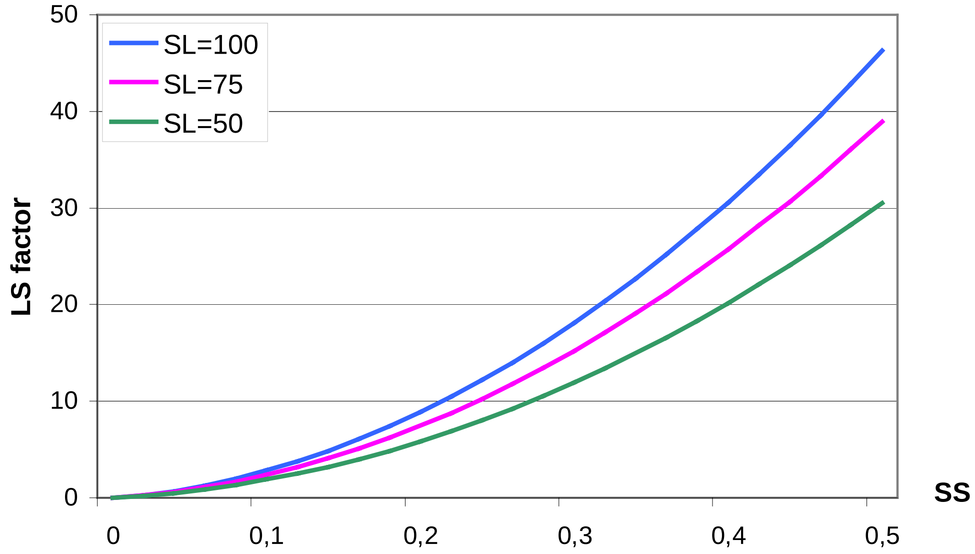

Fig. 2.25 The LS factor calculated as a function of slope steepness SS for different slope lengths SL (equations 166-167)

The LS factor is estimated with the equation (Wischmeier and Smith, 1978) (see **Fig. 2.25**)

$$LS = \left(\frac{SL}{22.1}\right)^{\xi} \cdot \left(65.41 \cdot SS^2 + 4.565 \cdot SS + 0.065\right) \tag{166}$$

where *SL* is the slope length, *SS* is the slope steepness, and the exponent ξ varies with slope and is computed with the equation

$$\xi = 0.6 \cdot \left[1 - \exp(-35.835 \cdot SS)\right] \tag{167}$$

The slope length and slope steepness are calculated in SWIM/GRASS interface for every sub-basin.

2.4.2 Organic Nitrogen Transport by Sediment

A loading function developed by McElroy et al. (1976) and modified by Williams and Hann (1978) for application to individual runoff events is used to estimate organic N loss for each sub-basin. The loading function is

$$YON = 0.001 \cdot YSED \cdot CNOR \cdot ER \tag{168}$$

where *YON* is the organic N runoff loss at the sub-basin outlet in kg ha^{-1}, *CNOR* is the concentration of organic N in the top soil layer in g t^{-1}, and *ER* is the enrichment ratio. The value of CNOR is input to the model and is constant throughout the simulation.

The enrichment ratio is the concentration of organic N in sediment divided by that of the soil. Enrichment ratios are logarithmically related to sediment concentration as described by Menzel (1980). An individual event enrichment sediment concentration relationship was developed considering upper and lower bounds. The upper bound of the enrichment ratio is the inverse of the sediment delivery ratio *DR* (sub-basin sediment yield divided by gross sheet erosion): *ER<1/DR*. Exceeding the inverse of the delivery ratio implied that more organic N leaves the watershed than is dislodged from the soil.

The delivery ratio is estimated for each runoff event using the equation

$$DR = \left(\frac{PEAKQ}{PRER} \right)^{0.56} \tag{169}$$

where *DR* is the sediment delivery ratio, *PEAKQ* is the peak runoff rate in mm h^{-1}, and *PRER* is the peak rainfall excess rate in mm h^{-1}.

Equation 169 is based on sediment yield estimated using MUSLE (Williams, 1975). The rainfall excess rate cannot be evaluated directly because the model uses only the total daily runoff volume, and not the event rainfall. An estimation of PRER can be obtained, however, using the equation

$$PRER = PRR - AIR \tag{170}$$

where *PRR* is the peak rainfall rate in mm h^{-1} and *AIR* is the average infiltration rate in mm h^{-1}.

The peak rainfall rate can be calculated with the equation

$$PRR = 2 \cdot PRECIP \cdot \log\left(\frac{1}{1 - \alpha_{0.5}} \right) \tag{171}$$

The average infiltration rate can be calculated with the equation

$$AIR = \frac{PRECIP - Q}{DUR} \tag{172}$$

where *DUR* is the rainfall duration in h, and *PRECIP* is rainfall in mm.

The rainfall duration is estimated the same as in equation 32 according to Williams et al. (1984).

$$DUR = \frac{4.605 \cdot PRECIP}{PRR} = -\frac{2.3025}{\log((1 - \alpha_{0.5})} \tag{173}$$

The enrichment ratio is estimated with the logarithmic equation

$$ER = PCON \cdot SEDC^{PEXP} \tag{174}$$

where *SEDC* is the sediment concentration in g m^{-3}, and *PCON* and *PEXP* are parameters set by the upper and lower limits.

To approach the lower limit for the enrichment ratio, 1.0, the sediment concentration should be extremely high. Conversely, a very low sediment concentration would cause the enrichment ratio to approach 1/DR. The simultaneous solution of equation 174 at the boundaries assuming that sediment concentrations range from 500 to 250000 g m^{-3} gives the following estimations for *PEXP* and *PCON*

$$PEXP = \frac{\log\left(\frac{1}{DR}\right)}{2.699} \tag{175}$$

$$PCON = \frac{1}{0.25^{PEXP}} \tag{176}$$

2.4.3 Phosphorus Transport by Sediment

Phosphorus transport with sediments is simulated with a loading function similar to that described in 2.4.2 for the organic N transport. The loading function for phosphorus is

$$YP = 0.01 \cdot YSED \cdot POR_1 \cdot ER \tag{177}$$

where *YP* is the sediment phase of P loss in runoff in kg ha^{-1}, and POR_1 is the concentration of organic P in the top soil layer in g t^{-1}.

2.5 River Routing

2.5.1 Flow Routing

The model uses the Muskingum flow routing method (see Maidment, 1993 and Schulze, 1995). For a given reach, the continuity equation may be expressed as:

$$\frac{d(STOR)}{dt} = QIN(t) - QOUT(t) \tag{178}$$

where *d(STOR)/dt* is the rate of change of storage within the reach ($m^3\ s^{-1}$), *QIN(t)* is the inflow rate ($m^3\ s^{-1}$) at time t, and *QOUT(t)* is the outflow rate ($m^3\ s^{-1}$) at time t.

The Muskingum method assumes a variable discharge storage equation:

$$STOR(t) = KST \cdot [X \cdot QIN(t) + (1 - X) \cdot QOUT(t)] \tag{179}$$

where *STOR(t)* is the storage (m^3) in river reach at time t, *KST* is the storage time constant for the reach (s), and X is the dimentionless weighting factor in river reach routing.

Here *KST* is the ratio of storage to discharge and has the dimention of time. In physical terms, *KST* is considered to be an average reach travel time for a flood wave, and X indicates the relative importance of the input *QIN* and outflow *QOUT* in determining the storage in a reach. The lower and upper limits for *X* are 0 and 0.5, respectively. Typical values of *X* for a river reach range between 0.0 and 0.3, with a mean value near 0.2.

Thus, from 179 the change in storage over time Δt is given as

$$\begin{aligned} &STOR(t+1) - STOR(t) = \\ &= KST \cdot [X \cdot QIN(t+1) + (1 - X) \cdot QOUT(t+1)] - \\ &- KST \cdot [X \cdot QIN(t) + (1 - X) \cdot QOUT(t)] \end{aligned} \tag{180}$$

The Muskingum equation is derived from the finite difference form of the continuity equation 178 and equation 180 as the following:

$$QOUT(t+1) = C_1 \cdot QIN(t+1) + C_2 \cdot QIN(t) + C_3 \cdot QOUT(t) \tag{181}$$

where the parameters C1, C2 and C3 are determined as (see also **Fig. 2.26**)

$$C_1 = \frac{-KST \cdot X + 0.5 \cdot \Delta t}{KST - KST \cdot X + 0.5 \cdot \Delta t} \tag{182}$$

$$C_2 = \frac{KST \cdot X + 0.5 \cdot \Delta t}{KST - KST \cdot X + 0.5 \cdot \Delta t} \tag{183}$$

$$C_3 = \frac{KST - KST \cdot X - 0.5 \cdot \Delta t}{KST - KST \cdot X + 0.5 \cdot \Delta t} \tag{184}$$

Here *KST* and *Δt* must have the same time units and the three coefficients C_1, C_2 and C_3 sum to 1.0. Numerical stability is attained and the computation of negative outflows is avoided if the following condition is fulfilled

$$2 \cdot KST \cdot X < \Delta t < 2 \cdot KST \cdot (1 - X) \tag{185}$$

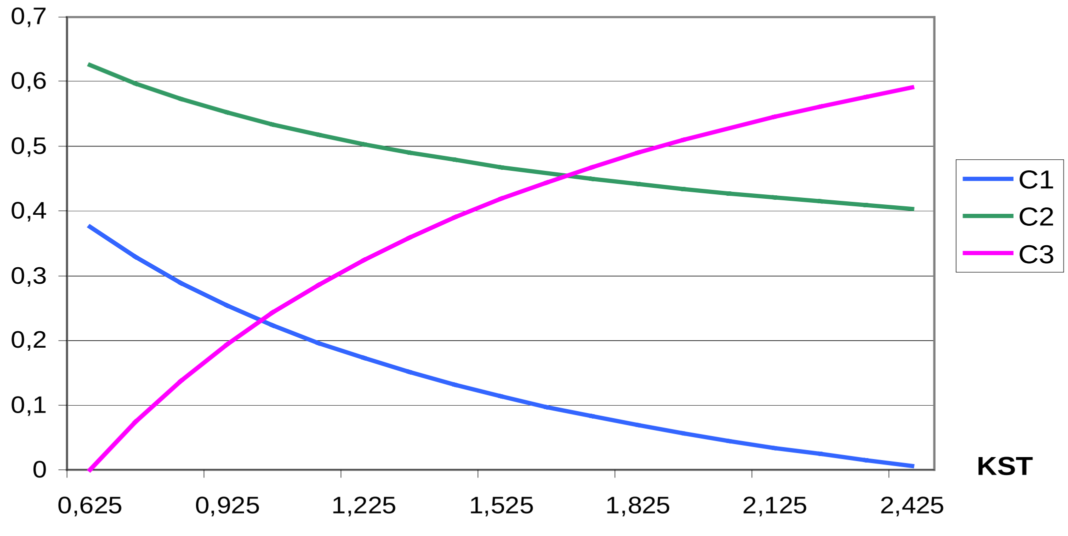

Fig. 2.26 Coefficients C_1, C_2 and C_3 as functions of parameter KST as used to calculate flow routing with the Muskingum equations 182-184 assuming that X = 0.2 and *Δt = 1*

Estimation of *KST* is based on the reach geometry

$$KST = \frac{CHL}{CLR} \tag{186}$$

where *CHL* is the reach length, and *CLR* is the wave celerity.

The celerity may be estimated by using the Manning formula with an adjusting coefficient for a certain reach shape. For the wide rectangular reach the celerity may be estimated (Schulze, p. AT13-9) as

$$CLR = \frac{5 \cdot CHV}{3} \tag{187}$$

where *CHV* is the average stream velocity in m s^{-1}. The average velocity is estimated from the Manning formula as

$$CHV = \frac{HR^{2/3} \cdot \sqrt{CHS}}{CHN} \tag{188}$$

where *HR* is hydraulic radius, *CHS* is channel bottom slope, *CHN* is the Manning's roughness N. The value of X is set in the model to 0.2.

2.5.2 Sediment Routing

The sediment routing model consists of two components operating simultaneously – deposition and degradation in the streams. Deposition in the stream channel is based on the stream velocity in the channel, which is estimated as a function of the peak flow rate, the flow depth, and the average channel width with the equation

$$CHV = \frac{PEAKQ}{FD \cdot CHW} \tag{189}$$

where *CHV* is the stream velocity in the channel in m s^{-1}, *PEAKQ* is the peak flow rate in m^3 s^{-1}, *FD* is the flow depth in m, and *CHW* is the average channel width in m.

The flow depth is calculated using the Manning's formula as

$$FD = \left(\frac{PEAKQ \cdot CHN}{CHW \cdot \sqrt{CHS}} \right)^{0.6} \tag{190}$$

where *CHN* is the channel roughness, N, and *CHS* is the channel slope in m m^{-1}.

The sediment delivery ratio *DELR* through the reach is described by the logarithmic equation suggested by J. Williams (similar to equation 174)

$$DELR = \frac{Q}{YSED_{in}} \cdot SPCON \cdot CHV^{sp\exp} \tag{191}$$

where $YSED_{in}$ is the sediment amount entering the reach, and the parameters SPCON (between 0.0001 and 0.01) and SPEXP (between 1.0 and 1.5) can be used for calibration. The power function in 191 is shown in **Fig. 2.27** for different combinations of SPCON and CHV.

If $DELR < 1.0$, the degradation is zero, and deposition is estimated from the sediment input as

$$\begin{aligned} &DEP = YSED_{in} \cdot (1 - DELR), \\ &DEGR = 0. \end{aligned} \tag{192}$$

Otherwise, if $DELR \geq 1$. the deposition is zero, and the degradation is calculated from the sediment input as

$$\begin{aligned} &DEP = 0. \\ &DEGR = YSED_{in} \cdot (DELR - 1) \cdot CHK \cdot CHC \end{aligned} \tag{193}$$

where *CHK* is the channel K factor or the effective hydraulic conductivity of the channel alluvium (see also equation 71), and *CHC* is the channel C factor.

Finally, the amount of sediment reaching the sub-basin outlet, $YSED_{out}$, is

$$YSED_{out} = YSED_{in} - DEP + DEGR \tag{194}$$

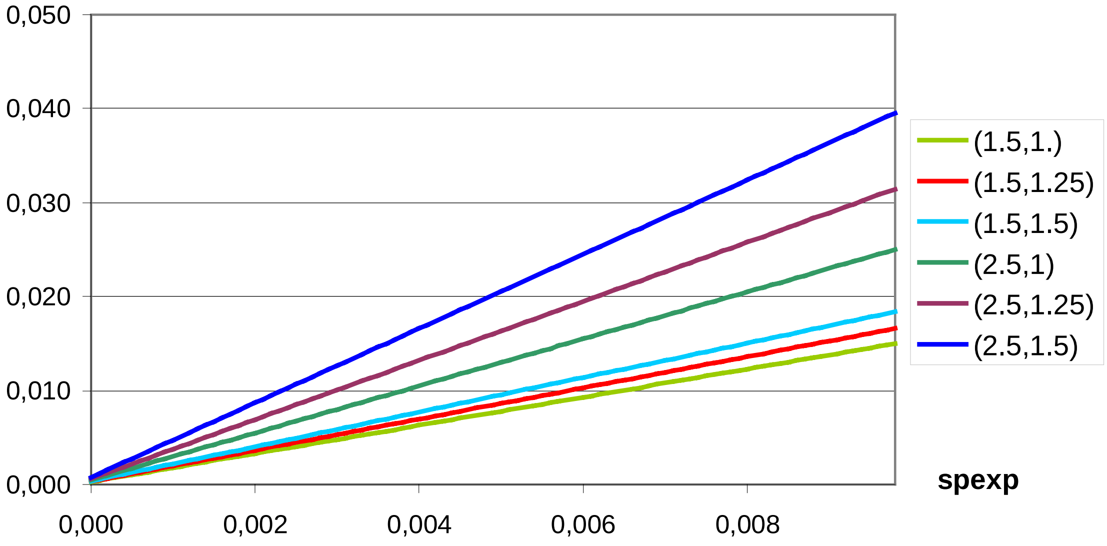

Fig. 2.27 Function $SPCON \cdot CHV^{spexp}$ to estimate the sediment delivery ratio DELR (equation 191) for different combinations of (CHV, SPCON)

2.5.3 Nutrient Routing

Nitrate nitrogen and soluble phosphorus are considered in the model as conservative materials for the duration of an individual runoff event (Williams, 1980). Thus they are routed by adding contributions from all sub-basins to determine the basin load.

Table 2.1 Abbreviations to Equations 1 – 194

α	the dimensionless parameter that expresses the proportion of total rainfall that occurs during time of concentration	-	19, 20, 32, 33, 34
$\alpha_{0.5}$	the fraction of rainfall that occurs during 0.5 h	-	32
α_{min}	minimum value of $\alpha_{.0.5}$, the fraction of rainfall that occurs during 0.5 h	-	33
β_i	a shape parameter to estimate the hydraulic conductivity for the layer i	-	38, 39
ν	the hillslope steepness	radian, or $m\ m^{-1}$	44, 45
δ	the slope of the saturation vapor pressure curve	$kPa\ C^{-1}$	47, 50
γ	a psychrometer constant	$kPa\ C^{-1}$	47
ϕ	the sun's half day length	radians	53
θ	the sun's declination angle	radians	53
ξ	the exponent to calculate the slope length and steepness factor of erosion, LS	-	166
ΔB	the daily increase in biomass	$kg\ ha^{-1}$	85
ΔBP	the daily potential increase in total biomass	$kg\ ha^{-1}\ d^{-1}$	84
Δt	the time interval (24 h)	h	35
A	the drainage area	ha	16, 20, 24, 25, 26
AIR	the average infiltration rate	$mm\ h^{-1}$	170
ALB	the albedo	-	57
ALB_{soil}	the bare soil albedo	-	58
ALFA	a factor to adjust net photosynthesis to altered CO2 concentration	-	107
$ALFA_{barley}$ $ALFA_{cot}$ $ALFA_{maize}$ $ALFA_{wheat}$	a factor to adjust net photosynthesis to altered CO_2 concentration for barley, cotton, maize and wheat (temperature dependent)	-	114
ALPFL	the flow from the active to the labile mineral P pool	$kg\ ha^{-1}\ d^{-1}$	149
AMP	the annual amplitude in daily average temperature	°C	122
ANFR	the active pool fraction (default: set to 0.15)	-	132
ANOR	active or readily mineralisable organic nitrogen	$kg\ ha^{-1}$	132
AR	the unit channel intercept	m^3	72
AS	the net leaf assimilation rate	$\mu mol\ m^{-2}\ s^{-1}$	107

ASNFL	the flow from the active to the stable organic N pool	kg ha^{-1} d^{-1}	133
ASPFL	the flow from the active to the stable mineral P pool	kg ha^{-1} d^{-1}	148
AX	the regression intercept to estimate the threshold volume for a unit channel	m km^{-1}	76
BAG	the aboveground biomass	kg ha^{-1}	87
BCV	a lagging factor for simulating residue and snow cover effects on surface temperature	-	129
BD	the soil bulk density	t m^{-3}	124
BE	the crop-specific parameter for converting energy to biomass	kg m^2 MJ^{-1} ha^{-1} d^{-1}	84
BETA	a factor to adjust potential leaf transpiration to CO_2 concentration		116
BMR	the sum of the above ground biomass and crop residue	t ha^{-1}	59
BN_1	the normal fraction of nitrogen in plant biomass at emergence (excluding seeds)	g g^{-1}	154
BN_2	the normal fraction of nitrogen in plant biomass at 0.5 maturity.	g g^{-1}	154
BN_3	the normal fraction of nitrogen in plant biomass at maturity	g g^{-1}	154
BP	the barometric pressure	kPa	51, 52
BR	the unit channel regression slope	-	74
BT	the accumulated total biomass	kg ha^{-1}	88
BX	the regression slope	-	77
C	the crop management factor,	-	163
C_1	a parameter to calculate river routing	-	181
C_2	a parameter to calculate river routing	-	181
C_3	a parameter to calculate river routing	-	181
CA	the current atmospheric CO_2 concentration	µmol mol^{-1}	108
CALP	the equilibrium constant between the active and labile mineral P pools (default: 1.)	-	149
CASN	the rate constant for the flow from the active to stable organic N pool	d^{-1}	133
CASP	the rate constant for the flow between the stable and active mineral P pools (default: 0.006)	d^{-1}	148
CAV	the average annual value of C factor	-	165
CBN_i	the carbon content in the layer i	kg ha^{-1}	152
CDN	a shape coefficient to estimate the combined temperature-carbon factor of denitrification, TCFD	-	152

CHC	the channel C factor	-	193
CHFL	the average channel flow length for the basin	km	22, 23
CHK	the effective hydraulic conductivity of the channel alluvium in	mm h^{-1}	72
CHL	the channel length from the most distant point to the watershed outlet	km	23, 24, 26
CHL_{cen}	the distance from the outlet along the channel to the watershed centroid	km	23
CHN	Manning's roughness coefficient n for the channel	-	24, 26
CHS	the average channel slope	m m^{-1}	24, 26
CHV	the average stream velocity in the channel	m s^{-1}	22
CHW	the average channel width	m	77
CL_1	the current CO_2 concentration inside leaves	μmol mol^{-1}	111
CL_2	the future CO_2 concentration inside leaves	μmol mol^{-1}	111
CLR	the stream wave celerity	m s^{-1}	186
CMN	the minimum value of C	-	164
CN	the curve number	-	4
CN_1	the curve number for soil moisture condition 1 (dry)	-	5, 6
CN_2	the curve number for soil moisture condition 2 (average)	-	5, 7, 8
CN_{2s}	the curve number for soil moisture condition 2 adjusted for slope	-	8
CN_3	the curve number for soil moisture condition 3 (wet)	-	7, 8
CNB	the optimal N concentration for the crop	g g^{-1}	100
CNOR	the concentration of organic N in the top soil layer	g t^{-1}	168
CNR	the C:N ratio	-	138, 140
CNRF	the C:N ratio factor of mineralisation	-	137
COMN	the humus rate constant for N (default: 0.0003)	d^{-1}	144
COMP	the humus mineralisation rate constant for P	kg ha^{-1} d^{-1}	147
$CON_i(t)$	the concentration of NO_3-N in the layer i on the day t	kg ha^{-1} mm^{-1}	156
COP	the concentration of labile phosphorus in soil layer	g t^{-1}	162
COV	the land cover, or the sum of above ground biomass and crop residue	kg ha^{-1}	130
CPR	the C:P ratio	-	139, 141
CPRF	the C:P ratio factor of mineralisation	-	137
CR	the revap coefficient	-	70
CS	the seepage coefficient	-	71

CSW	the P concentration in the sediment divided by that of the water	$m^3\ t^{-1}$	162
CTSH	the temperature stress parameter for the crop for temperatures above TO	-	98, 99
CTSL	the temperature stress parameter for the crop for temperatures below TO	-	96, 97
D	the earth's radius vector	km	53
DD	the damping depth for the soil	mm	122
DEC	the decay factor	$m\ km^{-1}$	73
DECR	the residue decomposition rate	-	134
DEGR	the degradation in stream	t	192, 193
DEL	the delay time or drainage time of the aquifer	day	69
DELR	the sediment delivery ratio through the reach	-	191
DENIT	the denitrification flow in layer i	$kg\ ha^{-1}\ d^{-1}$	150
DEP	the deposition in stream	t	192, 193
DLAI	the fraction of the growing season before LAI starts declining	-	87
DP	the maximum damping depth for the soil	mm	123
DR	the sediment delivery ratio	-	169
DS	the drain spacing	m	64
DU	the duration of streamflow	h	72
DUR	the rainfall duration	h	31, 32
ECP	the erosion control practice factor,	-	163
ELEV	the elevation of the site	m	52
EO	the potential evaporation	mm	47
EP	the plant water transpiration rate	$mm\ d^{-1}$	60
EPO	the potential plant transpiration rates	$mol\ m^{-2}\ s^{-1}$	116
ER	the enrichment ratio	-	168
ES	the soil evaporation for day t	$mm\ d^{-1}$	62
ESO	the potential soil evaporation	$mm\ d^{-1}$	61
ET	the evapotranspiration	$mm\ d^{-1}$	1
FC	the field capacity water content	vol % or $mm\ mm^{-1}$	10, 11, 12, 13, 37, 39
FD	the flow depth	m	28, 29
FFC	the fraction of field capacity	-	10, 11, 14
FFC^*	the depth-weighted FFC value	-	14
FOMN	N mineralisation flow from fresh organic N	$kg\ ha^{-1}\ d^{-1}$	136

FOMP	P mineralisation flow from fresh organic P	kg ha^{-1} d^{-1}	146
FON	the fresh organic N pool	kg ha^{-1}	135
FOP	the amount of fresh organic P	kg ha^{-1}	141
GWH	the water table height	m	64
GWQ	the return flow	mm d^{-1}	63
HC_i	the hydraulic conductivity	mm h^{-1}	37, 38, 40
HI	the harvest index at harvest	-	102
HIAD	the adjusted harvest index	-	104
HIC_1	the factor to estimate harvest index as depending on IHUN		103
HIC_2	the factor to estimate harvest index as depending on WSF		104
HR	the hydraulic radius	m	188
$HUMN_i$	the mineralisation rate for the active organic N pool in layer i,	kg ha^{-1} d^{-1}	144
$HUMP_i$	the mineralisation rate in the layer i	kg ha^{-1} d^{-1}	147
HUNA	the value of heat units accumulated in the day t	°C	81
HV	the latent heat of vaporization	MJ kg^{-1}	47, 48
HVSTI	the crop-specific harvest index	-	103
i	the soil layer	-	14
IHUN	the heat unit index	-	82
K	the soil erodibility factor	-	163
KD	the hydraulic conductivity in shallow aquifer	mm d^{-1}	64
KST	the storage time constant for the reach	s	179
LAI	the leaf area index	-	60
LAIMX	the maximum potential LAI for the specific crop	-	87
LAT	the latitude of the site	degrees	53
LS	the slope length and steepness factor	-	163
M	the number of soil layers	-	14
NDD	the number of days in a month	d	128
NDEM(t)	the N demand of the crop	kg ha^{-1}	153
NFL_i	the amount NO_3-N lost from the layer i	kg ha^{-1}	156
NMIN	the amount of mineral nitrogen (or nitrate nitrogen plus ammonium nitrogen) in soil	kg ha^{-1}	140
NOR	the total organic N pool	kg ha^{-1}	132
NRD	the number of rainy days in a month	d	128

NS	the stress factors caused by nitrogen	-	86
PAR	photosynthetic active radiation	$MJ\ m^{-2}$	83
PCON	a shape parameter to calculate the enrichment ratio ER	-	174
PEAKQ	the peak runoff rate	$m^3\ s^{-1}$	16, 20
$PEAKQ_{in}$	the initial peak discharge rate	$m^3\ s^{-1}$	80
$PEAKQ_{tr}$	the peak discharge rate after losses	$m^3\ s^{-1}$	80
PERC	the percolation, the percolation in layer i	$mm\ d^{-1}$	1
$PERC_{ic}$	the percolation rate for layer i corrected for layer i+I water content	$mm\ d^{-1}$	41
PEXP	a shape parameter to calculate the enrichment ratio ER	-	174
PFL	the soluble P lost with surface runoff	$kg\ ha^{-1}$	162
PHUN	the value of potential heat units required for the maturity of crop	°C	82
PLAB	the amount of labile P	$kg\ ha^{-1}$	141
PMA	the active mineral P pool	$kg\ ha^{-1}$	148
PMS	the stabile mineral P pool	$kg\ ha^{-1}$	148
PO	the soil porosity, the soil porosity for the layer i	vol % or $mm\ mm^{-1}$	10, 13
POR	the P organic pool in soil layer i	$kg\ ha^{-1}$ or $g\ t^{-1}$	147
PORD	the drainable porosity of the soil	$m\ m^{-1}$	43, 45
PRECIP	the precipitation	$mm\ d^{-1}$	1, 3, 17
$PRECIP_{24}$	the amount of rainfall during 24 hours	mm	19
$PRECIP_{tc}$	the amount of rainfall during the watershed's time of concentration	mm	18, 19
PRER	the peak rainfall excess rate	$mm\ h^{-1}$	169
PRR	the peak rainfall rate	$mm\ h^{-1}$	170
PS	the stress factors caused phosphorus	-	86
Q	the surface runoff	$mm\ d^{-1}$	1, 3, 17, 20, 31
QAV	the average flow rate	$mm\ h^{-1}$	24, 25
QAV_0	the average flow rate from a 1 ha area	$mm\ h^{-1}$	25, 26, 29, 30, 31
QIN(t)	the inflow rate at time t,	$m^3\ s^{-1}$	178
QOUT(t)	the outflow rate at time t	$m^3\ s^{-1}$	178
QUP	the upward flow	$mm\ d^{-1}$	46
RAD	the net solar radiation	$MJ\ m^{-2}$, or Ly	47
RAM	the maximum possible solar radiation	$MJ\ m^{-2}$, or Ly	53

RCH	the recharge	mm d^{-1}	63
RD	the fraction of the root zone that contains roots	-	91
RDMX	the maximum root depth (crop-specific parameter)	m	91
RDP	the rate-depth parameter	mm	92
REGF	the crop growth regulating factor estimated as the minimum stress factor	-	85, 86
RESC	the total leaf resistance to CO_2 transfer	m^2 s mol^{-1}	117
RESW	is the total leaf resistance to water vapour transfer	m^2 s mol^{-1}	117
REVAP	the water flow from the shallow aquifer back to the soil profile	mm d^{-1}	63
RF	the constant of proportionality or the reaction factor for groundwater	-	67
RI	the rainfall intensity for the watershed's time of concentration	mm h^{-1}	16, 18
RSD	the crop residue	kg ha^{-1}	106
RST	the revap storage	mm	70
RUNC	a runoff coefficient expressing the watershed infiltration characteristics	-	16, 17
RWT	the fraction of total biomass partitioned to the root system	-	88
RZD_i	the root zone depth parameter for the layer i	mm	92, 93
SAW_t	the shallow aquifer storage	mm	63
SC	the saturated conductivity	mm h^{-1}	38, 40, 44, 45, 46
SCOV	the soil cover index	-	58
SEDC	the sediment concentration	g m^{-3}	174
SEEP	the percolation or seepage to the deep aquifer	mm d^{-1}	63
SFN	the scaling factor to estimate the N stress factor	-	100
SHP_1	the coefficient of the S-shape curve, describing the assumed change in BE for two different CO_2 concentrations	-	108
SHP_2	the coefficient of the S-shape curve, describing the assumed change in BE for two different CO_2 concentrations	-	108
SL	the surface slope length (or hillslope length)	m	27, 30, 42, 43, 45, 46
SL_{sat}	the saturated slope length	m	46
SLW	the hillslope width	m	43
SML	the snowmelt rate	mm d^{-1}	2

SMX	the retention parameter for estimation of daily runoff	-	3, 4, 9, 10, 15
SMX_1	the value of SMX associated with CN_1, corresponding to moisture conditions 1	-	9, 10, 12, 13
SMX_2	SMX = SMX_2 when FCC = 0.7	-	10
SMX_3	SMX = SMX_3, when SW = FC	-	10, 12, 13
SMX_{froz}	the retention parameter for frozen ground	-	15
SN	Manning's roughness coefficient for the surface	-	28, 29, 30
SNO	the water content of snow cover	mm	2
SP_1	a shape parameter to estimate the optimal N concentration in the crop biomass	-	154
SP_2	a shape parameter the optimal N concentration in the crop biomass	-	154
SPCON	a shape parameter to estimate the sediment delivery ratio through the reach, DELR (between 0.0001 and 0.01)	-	191
SPD	a scaling parameter to estimate damping depth	-	123
SPEXP	a shape parameter to estimate the sediment delivery ratio through the reach, DELR (between 1.0 and 1.5)	-	191
SS	the land surface slope	$m\ m^{-1}$	8, 28, 29, 30
SSF	the subsurface flow	$mm\ d^{-1}$	1, 43, 45
STOR	the storage within the reach	$m^3\ s^{-1}$	178
SUP	the drainable volume of water stored in the saturated zone (water above field capacity)	$m\ m^{-1}$	42, 43, 45
SV	the surface flow velocity	$m\ s^{-1}$ or $m^3\ s^{-1}$	27, 28
SW(t)	the soil water content in day t	mm	1, 9, 10, 11, 35, 37, 38, 41
SWP	the accumulated potential plant transpiration in the second half of the growing season	mm	105
SWU	the accumulated actual plant transpiration in the second half of the growing season (IHUN>0.5)	mm	105
SY	the specific yield	-	65
t	the time	day	1
T	the mean daily air temperature	°C	48, 49, 50
TAV	the average annual air temperature	°C	122
TB	the crop-specific base temperature	°C	81
TC	the time of concentration	h	18, 20, 21, 33, 34
TC_{ch}	the time of concentration for channel flow	h	21, 22, 24, 26

TCFD	the combined temperature-carbon factor of denitrification	-	150
TC_{ov}	the time of concentration for surface flow	h	21, 27, 30
TFM_i	the temperature factor of mineralisation for the layer i	-	137
TGB(t)	the bare soil surface temperature in the day t,	°C	126, 127
TH_o	the threshold volume for a unit channel	m^3	78
TL	the leaf temperature	°C	111
TMN	the minimum temperature	°C	81
TMX	the maximum daily air temperature	°C	2
TO	the optimal temperature for the crop	°C	96
TS	the stress factors caused by temperature	-	86
TSO(Z,t)	the soil temperature at the depth Z in the day t	°C	122
TST	the number of days since stage two evaporation began	day	62
TT_i	the travel time through layer i	h	35
UL_i	the soil water content at saturation	mm mm^{-1}	38, 39, 41
UN(t)	the crop N uptake on day t	kg ha^{-1}	100
VEL	the velocity of flow at the outlet	mm h^{-1}	43, 44
VOLQ	the surface runoff column for the sub-basin	m^3	163
$VOLQ_{in}$	inflow volume	m^3	73
$VOLQ_{tr}$	the runoff volume after losses	m^3	79
VP	the saturation vapor pressure	kPa	49, 50
VPD	the vapour pressure deficit	kPa	117
WF_1	the shape parameter for estimation of the retention parameter SMX	-	9, 12
WF_2	the shape parameter for estimation of the retention parameter SMX	-	9, 12, 13
WFD	the water factor of denitrification	-	150
WFM	the water factor of mineralisation	-	137
WFT	a proportion of rainy days in a month	-	126, 127
WIR	the rate of water input to the saturated zone	m^2 h^{-1}	42
WP	the wilting point water content	vol % or mm mm^{-1}	10, 11
WS	the stress factors caused by water	-	86
WSF	a parameter expressing water supply conditions for crop	-	104, 105
WTOT	the total water lost from the soil layer	mm	155
WU_i	the plant water use in layer	mm	90

WUP_i	the potential water use rate from layer i	mm d^{-1}	92
X	a dimentionless weighting factor in river reach routing	-	179
YLD	the crop yield removed from the field	kg ha^{-1}	102
YON	the organic N runoff loss at the sub-basin outlet	kg ha^{-1}	168
YP	the sediment phase of P loss in runoff	kg ha^{-1}	177
YSED	the sediment yield from the sub-basin	t	163
$YSED_{in}$	the sediment amount entering the reach	t	191
$YSED_{out}$	the amount of sediment reaching the sub-basin outlet	t	194
Z_i	the depth to the bottom of soil layer i	mm	14
ZM	the distance from the bottom of the lowest soil layer to the surface	mm	125