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APPENDIX

APPENDIX A: WATERSHED MODEL PROCESSES

1 WATERSHED MODEL PROCESSES

A review of hydrological and sediment transport process decryptions is informative to illustrate the physics behind individual model process representations. The processes reviews are specific to those needs to formulate a fully distributed watershed chemical transport and fate model framework applicable to contaminants such as metals. The major components of the framework are hydrology, sediment transport, and chemical transport and fate. Each of the major components can be viewed as submodels within the overall framework. The reviews that follow are grouped by submodel.

1.1 Hydrologic Processes

The main processes in the hydrological submodel are: 1) rainfall and interception; 2) infiltration and transmission loss; 3) storage; 4) overland and channel flow.

1.1.1 Rainfall and Interception

The hydrological cycle begins with precipitation reaching the near surface of the land or water. The gross volume of water reaching the near surface can be expressed by equation A-1;

$$\frac{\partial V_g}{\partial t} = i_g A_s \tag{A-1}$$

where: V_g = gross precipitation water volume [L^3]

i_g = gross precipitation rate [L/T]

A_s = surface area over which precipitation occurs [L^3]

t = time [T]

Interception is the reduction in the volume of gross precipitation due to water retention by vegetative cover. As precipitation falls to the surface, portion of the gross precipitation at the surface may contact vegetative canopy and may be held on foliage by surface tension (Eagleson, 1970). Much of the precipitation falling during the earl period of an event may be stored on vegetative surface (Linsley *et al.,* 1982). Intercepted water can return to the atmosphere by evaporation. Alternatively, intercepted water may reach the land surface when the force of gravity acting of water drops exceeds the surface tension force holding water to plant surfaces. Conceptually, interception may be represented as a volume. The net rainfall volume equals to the gross rainfall volume minus the volume lost to interception (Linsley *et al.,* 1982)

$$V_i = (S_i + Et_R)A_s \tag{A-2}$$

$$V_n = V_g - V_i \qquad \text{for: } V_g > V_i \tag{A-3}$$

$$V_n = 0 \qquad \text{for: } V_g \leq V_i \tag{A-4}$$

where: V_i = interception volume [L^3]

S_i = interception capacity of project canopy per unit area [L^3/L^2]

E = evaporation rate [L/T]

t_R = precipitation event duration [T]

V_n = net precipitation volume reaching the surface [L^3]

Note that when the cumulative gross rainfall volume that occurs during an event is less than the interception volume, the net rainfall volume (or depth) reaching the land surface is zero. For single storm events, recovery of interception volume by evaporation can be neglected. The net precipitation volume may also be expressed as a net (effective) precipitation rate.

$$i_n = \frac{1}{A_s}\frac{\partial V_n}{\partial t} \tag{A-5}$$

where: i_n = net (effective) rainfall rate at the surface (L/T)

1.1.2 Infiltration and Transmission Loss

Infiltration is the downward transport of water from the surface to the subsurface. The rate at which infiltration occurs may be affected by several factors including hydraulic conductivity, capillary action and gravity (percolation) as the soil matrix reaches saturation. Many relationships have been used to describe infiltration including expressions presented by Green and Ampt (1911), Richards (1931), Philip (1957), and Smith and Parlange (1978). The Green and Ampt relationship is often used because of its ease of application. This relationship assumes a sharp wetting front exists between the infiltration zone and soil at the initial water content (piston flow) and that the length of the wetted zone increases as infiltration progresses. Neglecting the depth of ponding at the surface (i.e. assuming that the pressure head is much smaller than the suction head), the general equation showing the Green and Ampt relationship can be expressed as (Li *et al.,* 1976; Julien, 2002)

$$f = K_h\left(1 + \frac{H_C(1 - S_e)\theta_e}{F}\right) \qquad \text{(A-6)}$$

where: f = infiltration rate [L/T]

K_h = effective hydraulic conductivity [L/T]

H_c = capillary pressure (suction) head at the wetting front [L]

θ_e = effective soil porosity = $(\varphi - \theta_r)$ [dimensionless]

φ = total soil porosity [dimensionless]

θ_r = residual soil moisture content [dimensionless]

S_e = effective soil saturation [dimensionless]

F = cumulative (total) infiltrated water depth [L]

Similar to infiltration in overland areas, water in stream channels may be lost to the subsurface by transmission loss. The rate at which transmission loss occurs in a channel may be affected by several factors, particularly hydraulic conductivity. For ephemeral streams, capillary suction head may be significant when stream sediments are unsaturated. Relationships to describe the volume of transmission loss are presented by Lane (1983). Abdullrazzak and Morel-Seytoux (1983) and Freyberg

(1983) use the Green and Ampt (1911) relationship to assess transmission loss. Following the form of the Green and Ampt relationship and accounting for the depth of (ponded) water in the stream channel (hydrostatic head), the transmission loss rate may be expressed as;

$$t_l = K_h\left(1 + \frac{(H_w + H_c)(1 - S_e)\theta_e}{T}\right) \tag{A-7}$$

where: t_l = transmission loss rate [L/T]

K_h = effective hydraulic conductivity [L/T]

H_w = hydrostatic pressure head (depth of water in channel) [L]

H_c = capillary pressure (suction) head at the wetting front [L]

θ_e = effective sediment porosity = $(\varphi - \theta_r)$ [dimensionless]

φ = total sediment porosity [dimensionless]

θ_r = residual sediment moisture content [dimensionless]

S_e = effective sediment saturation [dimensionless]

T = cumulative (total) depth of water transported by transmission loss [L]

For single storm events, recovery of infiltration capacity by evapotranspiration and percolation can be neglected. Similarly, recovery of transmission loss capacity by evaporation or other processes can also be neglected for single storm events.

1.1.3 Storage

Water may be stored in depressions on the land surface as small, discontinuous surface pools. Precipitation retained in such small surface depressions is depression storage (Linsley *et al.*, 1982). Water in depression storage may be conceptualized as a volume or, when normalized by surface area, a depth. In effect, the depression storage depth represents a threshold limiting the occurrence of overland flow. When the water depth is below the depression storage threshold, overland flow is zero. Note that water in depression storage is still subject to infiltration and evaporation. Similar to depression storage in overland areas, water in channels may be stored in depressions in the stream bed (as the channel water depth falls below some critical level, flow is zero and the water surface discontinuous but individual pools of water remain). This mechanism is termed dead storage. Note that water in dead storage is still subject to transmission loss and evaporation.

For single storm events, recovery of depression storage volume by evaporation can be neglected. Similarly, recovery of dead storage volume by evaporation can also be neglected for single storm events.

1.1.4 Overland and Channel Flow

Overland flow can occur when the water depth on the overland plane exceeds the depression storage threshold. Overland flow is governed by conservation of mass (continuity) and conservation of momentum. The two-dimensional (vertically integrated) continuity equation for gradually-varied flow over a plane in rectangular (x, y) coordinates is (Julien *et al.,* 1995; Julien, 2002):

$$\frac{\partial h}{dt}+\frac{\partial q_x}{dx}+\frac{\partial q_y}{dy}=i_n-f+W=i_e \tag{A-8}$$

where: h = transmission loss rate [L]

q_x, q_y = unit discharge in the x- or y-direction = $\frac{Q_x}{B_x}, \frac{Q_y}{B_y}$ [L^2/T]

Q_x, Q_y = flow in the x- or y-direction [L^3/T]

B_x, B_y = flow width in the x- or y-direction [L]

W = unit discharge from/to a point source/sink [L^2/T]

i_e = excess precipitation rate [L/T]

Momentum equations for the x- and y-directions may be derived by relating the net forces per unit mass to flow acceleration (Julien *et al.,* 1995; Julien, 2002). In full form, with all terms retained, these equations can be expressed in dimensionless form as the friction slope and are known as the Saint Venant equations. The full Saint Venant equations may be simplified by neglecting small terms that describe the local and convective acceleration components of momentum, resulting in the diffusive wave approximation (of the friction slope) for the x- and y-directions:

$$S_{fx} = S_{0x} - \frac{\partial h}{dx} \tag{A-9}$$

$$S_{fy} = S_{0y} - \frac{\partial h}{dy} \tag{A-10}$$

where: S_{fx}, S_{fy} = friction slope (energy grade line) in the x- or y- direction [dimensionless]

S_{0x}, S_{0y} = ground surface slope in the x- or y- direction [dimensionless]

To solve the overland flow equations for continuity and momentum, five hydraulic variables must be defined in terms of a depth-discharge relationship to describe flow resistance. Assuming that flow is turbulent and resistance can be described using the Manning formulation (in S.I. units), the depth-discharge relationships are (Julien *et al.,* 1995; Julien, 2002):

$$q_x = \alpha_x h^{\beta} \tag{A-11}$$

$$q_y = \alpha_y h^{\beta} \tag{A-12}$$

$$\alpha_x = \frac{S_{fx}^{1/2}}{n} \tag{A-13}$$

$$\alpha_y = \frac{S_{fy}^{1/2}}{n} \tag{A-14}$$

where: α_x, α_y = resistance coefficient for flow in the x- or y-direction [$L^{1/3}/T$]

β = resistance exponent = 5/3 [dimensionless]

n = Manning roughness coefficient [$T/L^{1/3}$]

Similarly, channel flow can occur when the water depth in the channel exceeds the dead storage threshold. Channel flow is also governed by conservation of mass (continuity) and conservation of momentum. At the watershed it is convenient to represent channel flows in a watershed as one-dimensional (along the channel in the down-gradient direction). The one-dimensional (laterally and vertically integrated) continuity equation for gradually-varied flow along a channel is (Julien *et al.*, 1995; Julien, 2002):

$$\frac{\partial A_c}{dt}+\frac{\partial Q}{dx}=q_l \tag{A-15}$$

where: A_c = cross sectional area of flow [L^2]

Q = total discharge [L^3/T]

q_l = lateral unit flow (into or out of the channel) [L^2/T]

W = unit discharge from/to a point source/sink [L^2/T]

Based on the momentum equation for the down-gradient direction and again neglecting terms for local and convective acceleration, the diffusive wave approximation may be used for the friction slope (see Eq. 2.11). To solve the channel flow equations for continuity and momentum, the Manning relationship may be used to describe flow resistance (Julien et al., 1995; Julien, 2002):

$$Q=\frac{1}{n}A_c R_h^{2/3} S_f^{1/2} \tag{A-16}$$

where: R_h = hydraulic radius of flow = A_c / P_c [L]

P_c = wetted perimeter of channel flow [L]

1.2 Sediment Transport Processes

The movement of water across the overland plane or through a channel network can transport and redistribute soil and sediment throughout a watershed. The main processes in the sediment transport submodel are: 1) advection-diffusion;

2) erosion; 3) deposition; and 4) bed processes (bed elevation response to erosion and deposition).

1.2.1 Advection-Diffusion

For the overland plane in two-dimensions (vertically integrated), the concentration of particles is governed by conservation of mass (sediment continuity) (Julien, 1998):

$$\frac{\partial C_s}{dt} + \frac{\partial \hat{q}_{tx}}{dx} + \frac{\partial \hat{q}_{ty}}{dy} = \hat{J}_e - \hat{J}_d + \hat{W}_s = \hat{J}_n \qquad \text{(A-17)}$$

where: C_s = concentration of sediment particles in the flow [M/L^3]

$\hat{q}_{tx}, \hat{q}_{ty}$ = total sediment transport areal flux in the x- or y-direction [M/L^{2}T]

$\hat{J}_e$ = sediment erosion volumetric flux [M/L^{3}T]

$\hat{J}_d$ = sediment deposition volumetric flux [M/L^{3}T]

$\hat{W}_s$ = sediment point source/sink volumetric flux [M/L^{3}T]

$\hat{J}_n$ = net sediment transport volumetric flux [M/L^{3}T]

The total sediment transport flux in any direction has three components namely advective, dispersive (mixing) and diffusive which may be expressed as (Julien, 1998):

$$\hat{q}_{tx} = v_x C_s - (R_x + D)\frac{\partial C_s}{dx} \qquad \text{(A-18)}$$

$$\hat{q}_{ty} = v_y C_s - (R_y + D)\frac{\partial C_s}{dy} \qquad \text{(A-19)}$$

Where: v_x, v_y = flow (advective) velocity in the x- or y-direction [L/T]

R_x, R_y = dispersion (mixing) coefficient the x- or y-direction [L^2/T]

D = diffusion coefficient [L^2/T]

$v_x C_s$ = advective flux in the x-direction = J_x [M/L^{2}T]

$v_y C_s$ = advective flux in the y-direction = J_y [M/L^{2}T]

$R_x \dfrac{\partial C_s}{dx}$ = dispersive flux in the x-direction [M/L^{2}T]

$R_y \dfrac{\partial C_s}{dy}$ = dispersive flux in the y-direction [M/L^{2}T]

$D \dfrac{\partial C_s}{dx}$ = diffusive flux in the x-direction [M/L^{2}T]

$D \dfrac{\partial C_s}{dy}$ = dispersive flux in the y-direction [M/L^{2}T]

The dispersive and diffusive flux terms in Eq. (A-17) and (A-18) are negatively signed to indicate that mass is transported in the direction of decreasing concentration gradient. Note that both dispersion and diffusion are represented in forms that follow Fick's Law. However, dispersion represents a relatively rapid turbulent mixing process while diffusion represents a relatively slow a Brownian motion, random walk process (Holley, 1969). In turbulent flow, dispersive fluxes are typically several orders of magnitude larger than diffusive fluxes. Further, flow conditions for intense precipitation events are usually advectively dominated as dispersive fluxes are typically one to two orders smaller than advective fluxes. As a result, both the dispersive and diffusive terms may be neglected.

Similarly, for the channel plane in one-dimension (laterally and vertically integrated), the concentration of particles is governed by conservation of mass (sediment continuity) (Julien, 1998):

$$\frac{\partial C_s}{dt} + \frac{\partial \hat{q}_{tx}}{dx} = \hat{J}_e - \hat{J}_d + \hat{W}_s = \hat{J}_n \qquad \text{(A-20)}$$

Individual terms for the channel advection-diffusion equation are identical to those defined for the overland plane. Similarly, the diffusive flux term can be neglected. The dispersive flux is expected to be larger than the corresponding term for overland flow. However, the channel dispersive flux still may be negligible relative to the channel advective flux during intense runoff events.

1.2.2 Erosion

Erosion is the entrainment (gain) of material from a bottom boundary into a flow by the action of water. The erosion flux may be expressed as a mass rate of particle removal from the boundary over time and the concentration (bulk density) of particles at the boundary:

$$J_e = v_r C_{sb} \tag{A-21}$$

where: J_e = erosion flux (M/L^{2}T]

v_r = resuspension (erosion) velocity [L/T]

C_{sb} = concentration of sediment at the bottom boundary (in the bed) [M/L^3]

Entrained material may be transported as either bedload or suspended load. However, for overland sheet and rill flows, bedload transport by rolling and sliding may predominate as the occurrence of saltation and full suspension may be limited (Julien and Simons, 1985). Entrainment rates may be estimated from site-specific erosion rate studies or, in general, from the difference between sediment transport capacity and advective fluxes:

$$v_r = \frac{J_c - v_a C_s}{\rho_b} \qquad \text{for } J_c > v_a C_s \tag{A-22}$$

$$v_r = 0 \qquad \text{for } J_c \le v_a C_s \tag{A-23}$$

where: v_r = resuspension (erosion) velocity [L/T]

J_c = sediment transport capacity areal flux [M/L^{2}T]

v_a = advective (flow) velocity (in the x- or y-direction) [L/T]

C_s = concentration of sediment entrained in the flow [M/L^3]

ρ_b = bulk density of bed sediments [M/L^3]

In the overland plane, particles can be detached from the bulk soil matrix by raindrop (splash) impact and entrained into the flow by hydraulic action when the exerted shear stress exceeds the stress required to initiate particle motion (Julien and Frenette, 1985; Julien and Simons, 1985). The overland erosion process is influenced by many factors including precipitation (rainfall) intensity and duration, runoff length, surface slope, soil characteristics, vegetative cover, exerted shear stress, and particle size. Raindrop impact may generally be neglected when flow depths are greater than three times the average raindrop diameter (Julien, 2002). Julien and Simons (1985) summarize numerous relationships to describe the transport capacity of overland flow. Julien (1998, 2002) recommends a modified form of the Kilinc and Richardson (1973) relationship that includes soil erodibility, cover, and management practice terms from the Universal Soil Loss Equation (USLE) (Meyer and Weischmeier, 1969) to estimate the total overland sediment transport capacity (for both the x- and y-directions):

$$q_s = 1.542x10^8 q^{2.035} S_f^{1.66} \hat{K}\hat{C}\hat{P} \tag{A-24}$$

$$J_c = \frac{q_s}{B_e} \tag{A-25}$$

where: q_s = total sediment transport capacity [M/LT]

q = unit flow rate of water= $v_a h$ [M/LT]

S_f = friction slope [dimensionless]

$\hat{K}$ = USLE soil erodibility factor [dimensionless]

$\hat{C}$ = USLE soil cover factor [dimensionless]

$\hat{P}$ = USLE soil management practice factor [dimensionless]

B_e = width of eroding surface in flow direction [L]

In channels, sediment particles can be entrained into the flow when the exerted shear stress exceeds the stress required to initiate particle motion. For non-cohesive particles, the channel erosion process is influenced by factors such as particle size, particle density and bed forms. For cohesive particles, the erosion process is significantly influenced by inter-particle forces (such as surface charges that hold grains together and form cohesive bonds) and consolidation. Total (bed material) load transport capacity relationships account for the both bedload and suspended load components of sediment transport. Yang (1996) and Julien (1998) provide summaries of numerous total load transport relationships. The Engelund and Hansen (1967) relationship is considered a reasonable estimator of the total load:

$$C_w = 0.05\left(\frac{G}{G-1}\right)\frac{v_a S_f}{\left[(G-1)gd_p\right]^{0.5}}\left[\frac{R_h S_f}{(G-1)d_p}\right]^{0.5} \qquad \text{(A-26)}$$

$$J_c = \frac{v_a C_t}{A_c} \qquad \text{(A-27)}$$

where: C_w = concentration of entrained sediment particles by weight at the transport capacity [dimensionless]

G = particle specific gravity [dimensionless]

v_a = advective (flow) velocity (in the down-gradient direction) [L/T]

S_f = friction slope [dimensionless]

R_h = hydraulic radius of flow [L]

g = gravitation acceleration [L/T^2]

d_p = particle diameter [L]

A_c = cross sectional area of flow [L^2]

C_t = concentration of entrained sediment particles at the transport

$$\text{capacity} = \frac{10^6 G C_w}{\left[G + (1-G)C_w\right]} \ [\text{M/L}^3] \qquad \text{(A-28)}$$

It is worth noting that one feature common to both the Kilinc and Richardson (1973) and Engelund and Hansen (1967) relationships are that the implicit threshold for incipient motion is zero. This means that the transport capacity of any particle will always be greater than zero, regardless of particle size or the exerted shear stress, as long as the unit flow or flow velocity and friction slope are non-zero. This can lead to inconsistent results when erosion rates are computed from sediment transport capacities. The inferred erosion rate will almost always be greater than zero because the difference between the transport capacity and advective flux will nearly always be greater than zero. Consequently, a nonzero erosion rate can be computed even when the exerted shear stress is far less than the incipient motion threshold for the material. To address this limitation, an incipient motion threshold can be added to the modified Kilinc and Richardson (1973) and Engelund and Hansen (1967) relationships:

$$q_s = 1.542x10^8 (q - q_c)^{2.035} S_f^{1.66} \hat{K}\hat{C}\hat{P} \tag{A-29}$$

$$C_w = 0.05\left(\frac{G}{G-1}\right)\frac{(v_a - v_c)S_f}{[(G-1)gd_p]^{0.5}}\left[\frac{R_h S_f}{(G-1)d_p}\right]^{0.5} \tag{A-30}$$

where: q_c = critical unit flow for erosion (for aggregate the soil matrix)
= $v_c h$ [L^2/T]
v_c = critical velocity for erosion [L/T]
h = surface water depth [L]

1.2.3 Deposition

Deposition is the sedimentation (loss) of material entrained in a flow to a bottom boundary by gravity. The deposition process is influenced by many factors including particle density, diameter, and shape, and fluid turbulence. The deposition flux may be expressed as a mass rate of particle removal from the water column over time and the concentration of sediment particles that are entrained in the flow:

$$J_d = v_{se} C_s \tag{A-31}$$

where: J_d = deposition flux [M/L^2/T]

v_{se} = effective settling (deposition) velocity [L/T]

C_s = concentration of sediment particles in the flow [M/L^3]

Coarse particles (>62 μm) are typically inorganic and non-cohesive and generally have large settling velocities under quiescent conditions. Numerous empirical relationships to describe the non-cohesive particle settling velocities are available. Summaries of relationships and settling velocities are presented by Yang (1996) and Julien (1998). For non-cohesive (fine sand) particles with diameters from 62 μm to 500 μm, the settling velocity can be computed as (Cheng, 1997):

$$v_s = \frac{\nu}{d_p}\left[\left(25 + 1.2d_*^2\right)^{0.5} - 5\right]^{1.5} \qquad \text{(A-32)}$$

$$d_* = d_p\left[\frac{(G-1)g}{\nu^2}\right]^{1/3} \qquad \text{(A-33)}$$

where: v_s = quiescent settling velocity [L/T]

ν = kinematic viscosity of water [L^2/T]

d_* = dimensionless particle diameter [dimensionless]

Medium particles (10 μm < d_p <62 μm) can vary in character. Inorganic particles may behave in a non-cohesive manner. In contrast, organic particles (potentially including particles with organic coatings) may behave in a cohesive manner. Fine particles (<10 μm) often behave in a cohesive manner. If behavior is largely non-cohesive, settling velocities may be estimated as described by Julien (1998). If the behavior is cohesive, flocculation may occur. Floc size and settling velocity depend on the conditions under which the floc was formed (Burban *et al.,* 1990; Krishnappan, 2000; Haralampides *et al.,* 2003). When flocculation occurs, settling velocities of cohesive particles can be approximated by relationship of the form (Burban *et al.,* 1990):

$$v_{sf} = ad_f^m \tag{A-34}$$

where: v_{sf} = floc settling velocity [L/T]

a = experimentally determined constant = 8.4×10^{-3}

d_f = median floc diameter [L]

m = experimentally determined constant = 0.024

However, depending on fluid shear, particle surface charge, and other conditions, fine particles may not flocculate. Under conditions that limit floc formation, fine particles can have very small, near zero settling velocities.

As a result of turbulence and other factors, not all particles settling through a column of flowing water will necessarily reach the sediment-water interface or be incorporated into the sediment bed (Krone, 1962). Beuselinck *et al.* (1999) suggest this process also occurs for the overland plane. As a result, effective settling velocities in flowing water can be much less than quiescent settling velocities. The effective settling velocity of a particle can be described as a reduction in the quiescent settling velocity by the probability of deposition (Krone 1962; Mehta *et al.*, 1989):

$$v_{se} = P_{dep} v_s \tag{A-35}$$

where: v_{se} = effective settling velocity [L/T]

v_s = quiescent settling velocity [L/T]

P_{dep} = probability of deposition [dimensionless]

The probability of deposition varies with shear stress near the sediment bed and particle size. As particle size decreases or shear stress increases, the probability of deposition decreases. For non-cohesive particles, the probability of deposition has been described as a function of bottom shear stress (Gessler, 1965; 1967; 1971):

$$p_{dep} = p = \frac{1}{\sqrt{2\pi}} \int_{-\infty}^{Y} e^{-0.5x^2}\,dx \tag{A-36}$$

$$Y = \frac{1}{\sigma}\left(\frac{\tau_{cd,n}}{\tau} - 1\right) \tag{A-37}$$

where: P = probability integral for the Gaussian distribution

σ = experimentally determined constant = 0.57

τ_0 = bottom shear stress [M/LT^2]

$\tau_{cd,n}$ = critical shear stress for deposition of non-cohesive particles, defined as the shear stress at which 50% of particles deposit [M/LT^2]

For coarse particles, the critical shear stress for deposition can be computed from a force balance following the method of van Rijn (1984a,b) as summarized by QEA (1999), with the particle diameter equal to the mean diameter for a range of particle size in a class (i.e. $d_p = d_{50}$).

For cohesive particles, the probability of deposition has also been described as a function of bottom shear stress (Partheniades, 1992);

$$p_{dep} = 1 - p = 1 - \frac{1}{\sqrt{2\pi}} \int_{-\infty}^{Y} e^{-0.5x^2}\,dx \tag{A-38}$$

$$Y = \frac{1}{\sigma}\ln\left[0.25\left(\frac{\tau_{cd,c}}{\tau} - 1\right)e^{1.27\tau_{cd}}\right] \tag{A-39}$$

where: σ = experimentally determined constant = 0.49

τ_0 = bottom shear stress [M/LT^2]

$\tau_{cd,c}$ = critical shear stress for deposition of cohesive particles, defined as the shear stress at which 100% of particles deposit [M/LT^2]

The probability integrals in Eq. A-36 and A-38 can be approximated as (Abramowitz and Stegun, 1972);

$$p = 1 - F(Y)(0.4392X - 0.1202X^2 + 0.9373X^3) \qquad \text{for Y>0} \qquad \text{(A-40)}$$

$$p = 1 - P(|Y|) \qquad \text{for Y<0}$$

$$F(Y) = \frac{1}{\sqrt{2\pi}} e^{-0.5Y^2} \qquad \text{(A-41)}$$

$$X = (1 + 0.3327Y)^{-1} \qquad \text{(A-42)}$$

1.2.4 Soil and Sediment Bed Processes

In response to the difference between bedform transport, erosion, and deposition fluxes, the net addition (burial) or net loss (scour) of particles from the bed causes the bed surface elevation to increase or decrease. The rise or fall of the bed surface is governed by the sediment continuity (conservation of mass) equation, various forms of which are attributed to Exner (1925) (see Simons and Sentürk, 1992). Julien (1998) presents a derivation of the bed elevation continuity equation for an elemental control volume that includes vertical and lateral (x- and y-direction) transport terms. Neglecting bed consolidation and compaction processes, and assuming that only vertical mass transport processes (erosion and deposition) occur, the sediment continuity equation for the change in elevation of the soil or sediment bed surface may be expressed as:

$$\rho_b \frac{\partial z}{dt} = v_{se} C_s - v_r C_{sb} \qquad \text{(A-43)}$$

where: z = elevation of the soil surface or sediment bed [L]

ρ_b = bulk density of soil or bed sediments [M/L^3]

v_{se} = effective setting (deposition) velocity [L/T]

C_s = concentration of sediment particles in the water column [M/L^3]

v_r = resuspension (erosion) velocity [L/T]

C_{sb} = concentration of sediment particles in the soil or sediment bed [M/L^3]

1.3 Chemical Transport and Fate Processes

The movement of water and sediment across the overland plane or through a channel network can also transport and redistribute chemicals throughout a watershed. On the land surface and in channel environments, chemical typically exist in three phases: 1) dissolved in water, 2) bound with dissolved organic compounds (DOC) or other binding ligands or complexation agents; and 3) particle-associated. The pathways that affect chemical movements and interactions in the environment depend on the phase in which the chemicals are present. The main processes in the chemical transport and fate submodel are: 1) chemical partitioning and phase distribution; 2) advection-diffusion; 3) erosion; 4) deposition; 5) infiltration; and 6) mass transfer and transformation processes (chemical reactions).

1.3.1 Chemical Partitioning and Phase Distribution

Many chemicals are hydrophobic and readily partition between dissolved, bound, and particle-associated (particulate) phases. Partitioning to bound and particulate phases is a function of chemical affinity for surfaces and ion exchange (ionic chemicals) or organic carbon (neutral chemicals) (Karichoff *et al.*, 1979; Schwarzenbach *et al.*, 1993; Chapra, 1997). The equilibrium distribution of chemicals between phases is described by the partition (distribution) coefficient, concentration and binding effectiveness of binding agents, and the concentration of particles or organic carbon. Mechanistically, partitioning is a function of the equilibrium rate at which chemicals sorb (move out of the dissolved phase) and desorb (move back into the dissolved phase). If the rates at which chemicals partition are much faster than the rates of other mass transfer processes, local equilibrium is assumed to exist and the dissolved, bound and particulate phase chemical concentrations can be expressed in terms of the total chemical concentration (sum of phases) (Thomann and Mueller, 1987; Chapra, 1997).

Chemicals may partition to all particle types (sorbents) present in a solution. The equilibrium partition (distribution) coefficient to any particle is defined as (Thomann and Mueller, 1987):

$$\pi_{p_n} = K_{p_n} = f_{oc_n} K_{oc} \tag{A-44}$$

where:

π_{p_n} = equilibrium partition (distribution) coefficient for particle "n" [L^3/M]

K_{p_n} = equilibrium partition (distribution) coefficient for particle "n" [L^3/M]

f_{oc_n} = fraction organic carbon of particle "n" [dimensionless]

K_{oc} = organic carbon normalized partition coefficient [L^3/M]

For particles phases in the water column, equilibrium partition coefficients vary with the concentration of suspended solids as a result of particle interactions. Particle dependent partition coefficients are described as (DiToro, 1985):

$$\pi_{px_n} = \frac{\pi_{px_n}}{1 + \frac{\sum_{n=1}^{N} m_n \pi_{p_n}}{v_x}} = \frac{f_{oc_n} K_{oc}}{1 + \frac{\sum_{n=1}^{N} m_n f_{oc_n} K_{oc}}{v_x}} \tag{A-45}$$

where:

π_{px_n} = particle dependent partition coefficient [L^3/M]

n = particle index = 1,2,3, ... etc

m_n = particle interaction parameter [dimensionless]

v_x = particle interaction parameter [dimensionless]

For the bound phase, the equilibrium binding coefficient is defined as:

$$\pi_b = D_e f_{oc_D} K_{oc} \tag{A-46}$$

where:

π_b = particle dependent partition coefficient [L^3/M]

f_{oc_D} = fraction organic carbon of particle "n" [dimensionless]

D_e = DOC-binding effectiveness coefficient [dimensionless]

Conceptually, dissolved organic compounds are composed entirely of organic carbon ($f_{oc_D} = 1$). Under those conditions, the equilibrium binding coefficient would equal the organic carbon partition coefficient. However, at least for neutral organic chemical binding in some surface waters (the Great Lakes), observed binding coefficients were up to 100 times smaller than Koc (Eadie *et al.,* 1990; Eadie *et al.,* 1992). Also, in sediment observed binding coefficients were up to 10 times smaller than Koc (Landrum *et al.,* 1985; Landrum *et al.*, 1987; Capel and Eisenreich, 1990). One explanation for decreased binding efficiency is photobleaching of DOC by ultraviolet (UV-B) radiation (Kashian *et al.,* 2004).

The equilibrium partition coefficient can be used to describe the fraction of the total chemical that is associated with each phase as follows (Thomann and Mueller, 1987; Chapra, 1997):

$$f_d = \frac{1}{1 + D_{oc}\pi_b + \sum_{n=1}^{N} m_n \pi_{px_n}} \tag{A-47}$$

$$f_b = \frac{D_{oc}\pi_b}{1 + D_{oc}\pi_b + \sum_{n=1}^{N} m_n \pi_{px_n}} \tag{A-48}$$

$$f_{p_n} = \frac{m_n \pi_{px_n}}{1 + D_{oc}\pi_b + \sum_{n=1}^{N} m_n \pi_{px_n}} \tag{A-49}$$

$$f_d + f_b + \sum_{n=1}^{N} f_{p_n} = 1 \tag{A-50}$$

where:
f_d = fraction of the total chemical in the dissolved phase [dimensionless]
f_b = fraction of the total chemical in the DOC-bound phase [dimensionless]
f_{p_n} = fraction of the total chemical in the particulate phase associated with particle "n" [dimensionless]
n = particle index = 1,2,3, .. etc

Equations A-47 to A-49 are presented for the water column. For the sediment bed, π_{p_n} is used in place of π_{px_n}

Lu and Allen (2001) present extensive assessments of copper partitioning onto suspended particulate matter in river water. Their results suggest that adsorption to organic matter binding sites in aqueous and solid phases plays the biggest role in controlling the extent of copper partitioning and that the most significant environmental factors affecting the value of the partition coefficient were the total suspended solids (TSS) concentration and pH. Holm *et al.,* (2003) found that cadmium partitioning, like copper, was highly correlated with soil cation exchange capacity, which is largely determined by organic carbon and clay content. Also, cadmium partition coefficients were found to decrease by an order of magnitude as soil pH decreased from 6.7 to 5.3. Similar behavior is also expected for zinc because, like copper and cadmium, it is divalent. Sauvé *et al*., (2000, 2003) noted that distribution coefficients for cadmium, copper, and zinc and other divalent metals are sensitive to pH. Sauvé *et al.,* (2003) reported distribution coefficients (log K_d) values for acidic (pH 4.4) soils were low: Cd log $K_d = 3.05$; Cu log $K_d = 2.98$; and Zn log $K_d = 2.75$. While pH may be the most important variable for partitioning, Sauvé *et al.*, (2000, 2003) also noted the importance of organic matter as, after being normalized for pH, sorptive capacities for organic soils were reported to be up to 30 times larger than those observed for mineral soils.

1.3.2 Chemical Advection

Advection transports all chemical phases. For two-dimensional flow in the overland plane, a chemical continuity (conservation of mass) equation analogous the sediment continuity equation can be written as:

$$J_{xc} = v_x\left(f_d + f_b + \sum_{n=1}^{N} f_{p_n} \right)C_c = v_x C_c \tag{A-51}$$

$$J_{yc} = v_y\left(f_d + f_b + \sum_{n=1}^{N} f_{p_n} \right)C_c = v_y C_c \tag{A-52}$$

where: J_{xc}, J_{yc} = chemical advective flux in the x- or y- direction [M/L^{2}T]

v_x, x_y = advective velocity in the x- or y- direction [L/T]

f_d = fraction of the total chemical in the dissolved phase [dimensionless]

f_b = fraction of the total chemical in the DOC-bound phase [dimensionless]

f_{p_n} = fraction of the total chemical in the particulate phase associated with particle "n" [dimensionless]

n = particle index = 1,2,3, .. etc

C_c = total chemical concentration [M/L^3]

Similarly, for one-dimensional flow in the channels a chemical continuity (conservation of mass) equation analogous the sediment continuity equation is identical to Eq.A-51.

1.3.3 Erosion and Deposition of Particulate Phase Chemicals

Chemicals associated with particles in the water column will enter the sediment bed if those particles settle. Similarly, chemicals associated with particles in the sediment bed will return to the water column if those particles are entrained (resuspend). Since particle phase chemicals move with the particles transported, the erosion and deposition fluxes of chemicals are described as (Thomann and Mueller, 1987):

$$J_{ec} = \sum_{n=1}^{N} v_{r_n} f_{p2_n} C_{c2} \tag{A-53}$$

$$J_{dc} = \sum_{n=1}^{N} v_{se_n} f_{p1_n} C_{c1} \tag{A-54}$$

where: J_{ec} = chemical erosion flux [M/L^{2}T]

J_{dc} = chemical deposition flux [M/L^{2}T]

n = particle index = 1,2,3, .. etc

v_{r_n} = resuspension (erosion) velocity of particle "n" [L/T]

v_{se_n} = effective settling velocity of particle "n" [L/T]

f_{p1_n} = fraction of the total chemical in particulate phase associated with particle "n" in the water column [dimensionless]

f_{p2_n} = fraction of the total chemical in particulate phase associated with particle "n" in the sediment column [dimensionless]

C_{c1} = total chemical concentration in the water column [M/L^3]

C_{c2} = total chemical concentration in the soil/sediment column [M/L^3]

1.3.4 Chemical Infiltration and Subsurface Transport

Chemicals associated with the dissolved and bound phase in the water column will enter the soil or sediment bed if the water transporting those chemicals infiltrates. When chemical partition coefficients are low and a significant fraction of the total chemical mass is in a mobile form, chemical infiltration may be significant. To account for this process, the chemical infiltration flux can be computed from the water infiltration flux as:

$$J_{ic} = v_i \left(f_{d1} + f_{b1}\right) C_{c1} = v_1 f_{m1} C_{c1} \tag{A-55}$$

where:

J_{ic} = chemical infiltration flux [M/L^{2}T]

v_i = infiltration rate or transmission loss of water [L/T], previously defined as f in Eq. (2.9) or t_l in Eq. (2.10)

f_{d1} = fraction of the total chemical in dissolved phase in the water column [dimensionless]

f_{b1} = fraction of the total chemical in bound phase in the water column [dimensionless]

f_{m1} = fraction of the total chemical in the mobile phase in the water column [dimensionless]

C_{c1} = total chemical concentration in the water column [M/L^3]

Once in the subsurface, infiltrated chemicals would be subject to repartitioning with the chemical mass associated with porewater and particles in the soil column and transport via groundwater. The flow of groundwater through the soil also has the potential to leach chemicals from the soil column. Due to adsorption and the comparatively high bulk density of particles in the soil, subsurface chemical transport is subject to retardation (Fetter, 2001). Chemicals subject to retardation travel trough the subsurface at rates less than the average linear velocity of water. The retardation factor for a chemical in the subsurface is computed as (Fetter, 2001):

$$R = 1 + \frac{\rho_b}{\theta_e} K_p \qquad \text{(A-56)}$$

where: R = Retardation factor [dimensionless]

ρ_b = soil bulk density [M/L^3]

θ_e = effective soil porosity (volume of voids/total volume of particles and voids) [dimensionless]

K_p = chemical equilibrium partition (distribution) coefficient [L^3/M]

1.3.5 Other Chemical Mass Transfer and Transformation Processes

Beyond partitioning and mass transport processes, the fate of chemicals is potentially influenced by a number of other processes such as biodegradation, hydrolysis, oxidation, photolysis, and volatilization, and dissolution. However, for general simulation of elemental metals such as cadmium, copper, and zinc, volatilization, biodegradation, and photolysis do not occur. Hydrolysis and oxidation can affect the ionic speciation and phase distribution of metallic chemicals but do not affect the total chemical concentration. The effect that possible hydrolysis or oxidation reactions have on phase distributions of metals can be represented in terms of the chemical distribution (partitioning) coefficient.

APPENDIX B: AGNPS MODEL

1. AGNPS MODEL

AGNPS is an event-based, deterministic-analytical water quality model (Young *et al.*, 1987; 1994). This model uses a distributed approach, which divides a watershed into uniform square areas or cells. Input data of approximately 22 parameters are required for each cell. Outputs from AGNPS include volume and peak runoff, sediment yield, sediment-attached and soluble nitrogen and phosphorus, COD (soluble chemical oxygen demand yield). A summary of required input and one of output for each cell are shown in Tables B-1 and B-2.

Table B-1 Input summary

1) Cell number (from)	12) Support practice factor
2) Receiving cell number (to)	13) Surface condition constant
3) SCS curve number (CN)	14) Aspen (direction of drainage)
4) Land slope	15) Soil texture
5) Land slope shape factor	16) Fertilization level
6) Filed slope length	17) Fertilization availability factor
7) Channel slope	18) Point source indicator
8) Channel sideslope	19) Gully source level
9) Manning's roughness coefficient	20) Chemical Oxygen Demand (COD)
10) Soil erodibility factor	21) Impoundment factor
11) Cover and management factor	22) Channel indicator

Details for input are as follows:

1. Cell Number – this data will be obtained within AGNPS itself. The program assigns the cell number is algorithm starting from left to right and top to bottom, whatever comes first.
2. Receiving cell number – this data is obtained from AGNPS, AGNPS will use the Digital Elevation Model data (DEM) which is generated from topography data to obtain the slope for each cell and specify which cell the runoff will flow to due to slope.

3. SCS Curve Number - widely used and efficient method for determining the amount of runoff from a rainfall even in a particular area. The requirements for this method are very low, rainfall amount and curve number. The curve number is based on the area's hydrologic soil group, land use, treatment and hydrologic condition. CN can be obtained either from the CN Curve Number chart or from the Technical Release 55 (TR-55) published by the Soil Conservation Service (SRS, 1986).
4. Land Slope, Land slope shape factor, Channel slope, Channel sideslope, Aspen (direction of drainage), Gully source level, Channel indicator – this is obtained from the DEM of the study area.
5. Manning's roughness coefficient, Soil erodibility factor, Chemical Oxygen Demand (COD), Impoundment factor – this data will be obtained from local research at PSU and related literature review for the study area.
6. Cover and management factor, Support practice factor, Surface condition constant, Point source indicator – this data will be obtained from data obtained from Land Development Department (2005) which is modified by Remote Sensing & GIS-PSU the land use of the study area.
7. Soil texture – this data will be obtained from land type of the study area.
8. Fertilization level, Fertilization availability factor – this data was covered in this study.

Table B-2 Output summary

Hydrology:	Nutrient:
1) Runoff volume (inches)	1) Nitrogen
2) Peak runoff rate (cfs)	2) Sediment associated mass
3) Fraction of runoff within the cell	3) Concentration of soluble material
4) Sediment (by particle size and in total)	4) Mass of soluble material in runoff
5) Sediment yield (tons)	5) Phosphorus
6) Sediment concentration (ppm)	6) Sediment associated mass (lbs/acre)
7) Sediment particle size distribution	7) Concentration of soluble material

8) Upland erosion (tons/acre)	8) Mass of soluble material in runoff
9) Chemical erosion (tons/acre)	9) Chemical Oxygen Demand
10) Amount of deposition	10) Concentration (ppm)
11) Enrichment ratio	11) Mass (lbs/acre)
12) Delivery ratio	

1.1 Components of AGNPS

1. Curve Number Method developed by Soil Conservation Service (USDA-SCS, 1972 and 1986). The method is a simple, widely used and efficient method for determining the approximate amount of runoff from a rainfall even in a particular area. Although the method is designed for a single storm event, it can be scaled to find average annual runoff values. The stat requirements for this method are very low, rainfall amount and curve number. The curve number is based on the area's hydrologic soil group, land use, treatment and hydrologic condition. The two former being of greatest importance.

$$Q = \frac{(P - I_a)^2}{(P - I_a) + S} \tag{B-1}$$

where,
Q = Surface Runoff (in)
P = Precipitation (in)
S = Potential maximum retention after runoff begins
Ia = initial loss

The initial Eq. (B-1) is based on trends observed in data from collected sites; therefore it is an empirical equation instead of a physically based equation.

$$I_a = 0.2S \tag{B-2}$$

$$Q = \frac{(P - 0.2S)^2}{(P + 0.8S)} \tag{B-3}$$

$$S = \frac{1000}{CN} - 10 \tag{B-4}$$

After further empirical evaluation of the trends in the data base, the initial abstractions, I_a, could be defined as a percentage of S Eq. (B-2). With this assumption, the Eq. (B-3) could be written in a more simplified form with only three variables. The parameter CN is a transformation of S, and it is used to make interpolating, averaging, and weighting operations more linear Eq. (B-4).

2. Universal Soil Loss Equation (USLE) developed by Wischmeier *et al.,* 1978. USLE formula

$$A = RK(LS)PC \tag{B-5}$$

where (Renart, *et al.,* 2000),

A = Annual soil loss (tons/acre) computed spatial average soil loss and temporal average soil loss per unit of area, expressed in the units selected for K and for the period selected for R.

R = Rainfall factor (EI units) rainfall-runoff erosivity factor-the rainfall erosion index plus a factor for any significant runoff from snowmelt.

K = Soil erodibility factor-the soil-loss rate per erosion index unit for a specified soil as measured on a standard plot, which is defined as a 72.6-ft (22.1-m) length of uniform 9% slope in continuous clean-tilled fallow.

L = field slope length factor, the ratio of soil loss from the field slope length to soil loss from a 72.6-ft length under identical conditions.

S = field slope factor, the ratio of soil loss from the field slope gradient to soil loss from a 9% slope under otherwise identical conditions.

P = Erosion control practice factor, the ratio of soil loss with a support practice like contouring, strip cropping, or terracing to soil loss with straight-row farming up and down the slope.

C = Cover and management factor, the ratio of soil loss from an area with specified cover and management to soil loss from an identical area in tilled continuous fallow.

The rainfall erosivity factor (R) is an index that characterizes the effect of raindrop impact and rate of runoff associated with the rainstorm. It is determined by calculating the EI for a specified period, usually 1 year or one season within the year. The EI averaged over a number of these periods (n) equals R:

$$R = \frac{\sum_{i}^{n} EI_i}{n} \tag{B-6}$$

The energy of a rainstorm depends on the amount of rain and all the component rainfall intensities of the storm. For any given mass in motion, the energy is proportional to velocity squared; therefore, rainfall energy is related directly to rain intensity by the relationship:

$$E = 916 + 331(\log I_i) \tag{B-7}$$

where:

E = kinetic energy per inch of rainfall in ft-tons/acre

I = rainfall intensity in each rainfall intensity period of the storm (inch/hour)

The total kinetic energy of a storm (k_e) is obtained by multiplying E by the depth in inches or rainfall in each intensity period (n), and summing:

$$k_e = \sum_{i}^{n} E = \sum_{i}^{n} 916 + 331(\log I_i) \tag{B-8}$$

The EI for an individual storm is calculated by multiplying the total kinetic energy (k_e) of the storm by the maximum amount of rain falling within 30 consecutive minutes (I_{30}), multiplying by 2 to obtain in./hr. and dividing the result by 100 (to convert from hundreds of ft-tons/acres to ft-tons./acre):

$$EI(storm) = \frac{2k_e I_{30}}{100} \qquad \text{(B-9)}$$

The EI for a specific period (year or season) is the sum of the individual storms' EI values computed for all significant storms during that period. Usually, only storms > 0.5 inches are selected. The R is the determined as the sum of the EI values for all such storms that occurred during a 20-25 year period, divided by the number of years (Eq. B-6)

The soil erodibility factor (K) indicated the susceptibility of soil to erosion and is expressed as soil loss per unit of area per unit of R for a unit plot. By definition, a unity plot is 72.6 ft long, on a uniform 9% slop, maintained in continuous fallow, with tillage when necessary to break surface crusts and to control weeds. These dimensions are selected because they coincide with the erosion research plots used in early work in the United States. Continuous fallow is selected as a base because no single cropping system is common to all agricultural areas, and soil loss from any other plot conditions would be influenced, to a large extent, by residual and current crop and management effects, both of which carry from one location to another. The K. value can be determined as the slope f a regression line through the origin for source data on soil loss (A) and erosivity (R), once the ratios for L, S, C and P have been adjusted to those of unit conditions. When the K. value was originally determined with natural rainfall data, it covered a range of storm sizes and antecedent soil moisture conditions. Results of later studies conducted with rainfall simulators were used to produce a soil erodibility nomograph based on soil texture and structure (Wischmeier *et al.,* 1971).

The topographic factors L and S indicated the effects of slope length and steepness, respectively, on erosion. Slope length refers to overland flow, from where is originated to where runoff reaches a defined channel or to where deposition

Other calculation used within the application is summarized in table B-3

Table B-3 Calculation algorithm summaries

Algorithms	References
Surface runoff - Curve Number (CN) method	USDA-SCS (1972)
Peak flow	Smith *et al.*, (1980)
Runoff velocity	Manning's equation
Soil loss - Modified USLE (Universal Soil Loss Equation)	Wischmeier *et al.*, (1978)
Sediment transport capacity - modified stream power equation	Bagnold (1966)
Sediment transport - stationary continuity equation	Foster *et al.*, (1981) and Lane (1982)
Nutrient transport (N, P)	Frere *et al.*, (1980)

A watershed is usually composed of fields of different crops, soil types, and land slopes, loading to various ways for sediment runoff. To reflect this spatial distribution, the watershed can be conceptually divided into a number of grid cells, with each of them begin uniform with respect to crop species, soil type, hydrological conditions, and topography. A grid cell structures is a discrete representation of a terrain, based on identical square cells arranged in rows and columns. Grids are used to describe spatially distributed parameters. The number of grid cells in a watershed varies with the watershed's size and the cell's dimensions, but should be large enough to account for the watershed's spatial variability. Output from grid cell at the top of the watershed is routed to cells below it and/or to the stream channels, and finally to the watershed's outlet. In each cell, the spatial data from GIS combined with soil properties data are input to the model for calculating the nutrient loads (Chen *et al.*, 2003).

The cell number is the main identifier for each cell in a watershed. Each cell in the watershed is identified by a number. The cells are numbered consecutively, beginning at the cell in the northwest corner, and sweeping from west to east, north to south. These numbers are automatically put into the spreadsheet. If the user divides a cell, the computer automatically will show the resulting cells of this division on the spreadsheet. The cell number will stay the same, but there will also be a number in the cell division line corresponding to the division. An example of the numbering scheme for a divided cell is shown in Figure B-1.

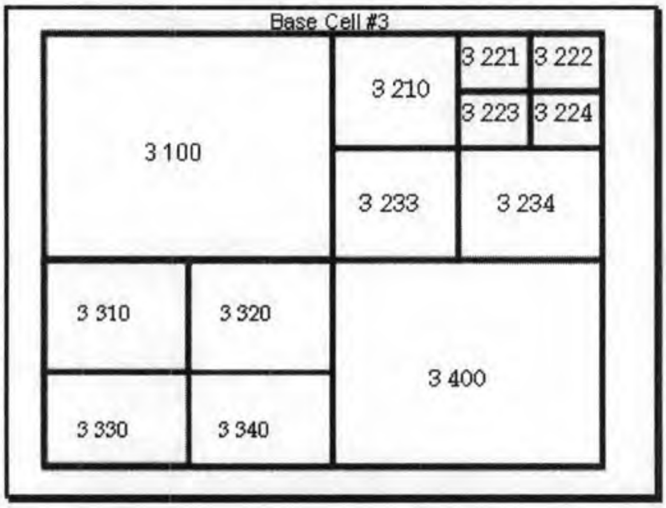

Figure B-1. Cell Division Numbering Scheme (Young *et al.,* 1994)

The flow direction is determined by cell topography and/or channel flow. Flow Direction is a single digit in the range of 1 to 8 indicating the principal direction of drainage from the cell. Each value refers to a direction with 1 representing north and, proceeding clockwise, 8 representing northwest, see Figure B-2.

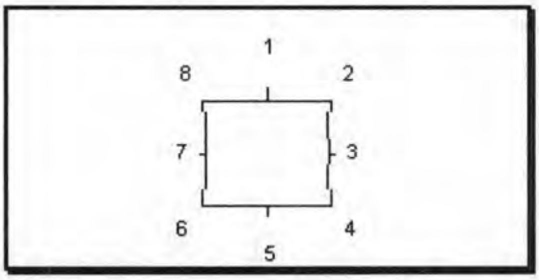

Figure B-2. Flow Direction (Young *et al.,* 1994)

Agricultural Non-Point Source Pollution Modeling System (AGNPS) is a joint USDA Agricultural Research Service and Natural Resources Conservation Service system of computer models developed to predict non point source pollutant loadings within agricultural watersheds. It contains a continuous simulation; surface runoff model designed for risk and cost/benefit analyses. The set of computer programs consist of: (1) input generation & editing as well as associated data bases;

(2) the “annualized” science & technology pollutant loading model (AnnAGNPS); and (3) output reformatting & analysis.

APPENDIX C: SOIL ERODIBILITY FACTORS FOR SOUTHERN THAILAND AND RUNOFF AND SEDIMENT FLOW INTO SONGKHLA LAKE

Table C-1 Soil Erodibility factor (K) of Southern Thailand

Soil Texture	High Land	Low Land
Sand	0.04	0.04
Loamy sand	0.04	0.04
Sandy load	0.02	0.03
Loam	0.33	0.34
Silt loam	0.4	0.34
Silt	-	0.57
Sandy clay loam	0.19	0.21
Clay loam	0.29	0.31
Silt clay loam	0.31	0.21
Clay loam	-	0.81
Silt clay loam	0.22	0.29
Sandy clay	0.11	0.14
Silty clay	0.04	0.04
Clay	0.04	0.04

Source : Office of the National Economic and Social Development Board, 1997

Table C-2 Runoff into Songkhla Lake

Watershed (Million*M^3) \Month	Klong Pa Payom & Thanae	Nathom	Tachiad	Pa Bon	Phru Poh	Rattaphum	U-Tapao
January	57.7	15.5	41.9	8.0	11.4	17.4	51.8
February	52.6	18.6	45.1	7.3	10.2	13.1	23.6
March	34.5	13.6	32.1	5.6	7.7	10.1	36.4
April	33.4	11.2	19.8	3.7	7.7	15.5	53.8
May	29.5	13.6	32.8	6.4	12.1	22.4	67.1
June	22.6	9.7	24.1	5.7	9.9	17.1	53.2
July	29.9	13.7	28.1	6.2	12.5	28.2	75.8
August	35.4	20.1	60.5	7.4	13.1	40.7	115.9
September	46.9	19.5	49.2	8.6	14.9	37.3	109.1
October	84.3	30.9	90.9	17.9	32.5	73.6	178.6
November	248.7	94.0	270.3	44.0	71.7	118.7	286.1
December	270.9	90.3	231.3	40.8	69.4	108.6	318.1
Total	946.28	350.72	925.92	161.44	273.33	502.53	1,369.38

Source: Office of Natural Resources and Environmental Policy and Planning, 2005

Table C-3 Sedimentation Load into Songkhla Lake

Sub watershed	Sedimentation Loss (Tons/Year)	Area (Rai)
Klong Pa Payom	385,676	521,479
Thanae	177,526	219,846
Nathom	648,082	472,538
Tachiad	387,045	479,780
Pa Bon	165,816	204,931
Phru Poh	221,231	314,694
Rattaphum	139,373	389,835
U-Tapao	759,084	1,465,834

Source: Land Development Department, 2005

VITA

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From 1996 to 1997, he worked at the National Petrochemical (Public) Company at Map Ta Phut, Rayong, Thailand, for more than a year as a process control engineer. His work related to process simulation development of the ethylene and propylene process, and making use of it in real plant.

After resignation, he has joined the Department of Chemical Engineering, Faculty of Engineering at Chulalongkorn University as a lecturer teaching Basic Computer for Chemical Engineering and Unit Operation Laboratory. Also, he had been guest lectures at various institutes such as Assumption University, Mahidol University and Fah Luang University in other fields of his interest. He also is an active consultant related to chemical engineering, network security, environmental management and performance management for various organizations both government and private sectors.