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Drainage-water travel times as a key factor for surface water contamination

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Abstract

The importance of the unsaturated zone as an inextricable part of the hydrologic cycle has long been recognized. The root zone and the unsaturated sub-surface domain are chemically and biologically the most active zones. The interrelationships between soil, subsoil and surface waters make it unrealistic to treat the saturated and unsaturated zones and the discharge to surface waters separately. Point models describe vertical water flow in the saturated zone and possibly lateral flow by defining a sink term. To account for the influence of two- and three-dimensional water flow on the travel-time distribution a conceptualization of the flow field is required. A formulation for upscaling the groundwater flow field is presented which yields the average vertical flux as a key factor for describing the travel time implicitly. Analytical solutions are given for the upscaled description for the transport; they are applied to a simple model consisting of a cascade of reservoirs. The analytical approach given, which includes the main properties of the soil system as well as the drainage system, proves to be useful for the prediction of solute concentrations in exfiltrating groundwater. The use and significance of conceptual models is discussed as well as the opportunities of detailed mechanistic integrated models that treat the unsaturated/saturated zone, overland flow and surface water flow comprehensively. Some results of experimental field work on the assessment of drainage water quality impacts of agricultural land management are summarized. Strengths, weaknesses, opportunities and threats of different modeling approaches are discussed. The unsaturated zone is an essential link in the chain between land management practices and the ecological status of freshwaters.
Keywords: groundwater; pollution; travel time; surface water; interaction; analytical solution

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Notation

a	equivalent time	T
A	area of the catchment	L^2
B	width of drainage canal	L
$c_i(t)$	concentration in soil layer i	$M\ L^{-3}$
$\bar{c}(t)$	concentration in exfiltration water	$M\ L^{-3}$
c_{inp}	input concentration	$M\ L^{-3}$
c_0	initial concentration	$M\ L^{-3}$
D	dispersion coefficient	$L^2\ T^{-1}$
H	thickness of aquifer	L
J_s	solute flux	$M\ L^{-2}\ T^{-1}$
h	hydraulic head	L
h_a	aquifer head	L
h_r	river head	L
L	drain spacing	L
l_i	length of drainage means of system i	L
$q_{d,i}$	drain flux density of system i	$L\ T^{-1}$
$q_y(x,y)$	vertical flux density in a vertical cross-section	$L\ T^{-1}$
$\bar{q}_y(y)$	vertical flux density in a vertical soil column	$L\ T^{-1}$
Q_d	drain discharge	$L^2\ T^{-1}$
n	iteration counter	—
t	travel time	T
Q_d	discharge originating from $0 \le x \le \frac{L}{2}$	$L^2\ T^{-1}$
R	recharge	L
R_{dr}	lateral sink term of vertical column model	T^{-1}
V_i	volume occupied by of drainage flow of system i	L^2
x	horizontal co-ordinate, distance to drain	L
y	vertical co-ordinate, pos. upward	L
(x_d, y_d)	co-ordinates of the deepest point of a streamline	(L, L)
Δy	compartment thickness in solute transport simulation	L
ε	porosity of aquifer	—
ϕ	potential	$L^2\ T^{-1}$
κ	streambed leakage coefficient	T^{-1}
λ	dispersion coefficient	L
Ψ	stream function	$L^2\ T^{-1}$
Ψ_0	stream function value for a certain streamline	$L^2\ T^{-1}$
Φ	complex potential	$L^2\ T^{-1}$

Introduction

Groundwater makes up 92% of the world's available freshwater resources. It plays a very important role in providing both drinking- and industrial water supplies, and also in sustaining natural water environments. It is often crucial in maintaining the wellbeing of surface water systems. Groundwater is vulnerable to contamination from a wide variety of sources. The use of fertilizers and pesticides spread over the land for long periods may cause diffuse pollution. This may build up over many years and the exact source of the problem can be difficult to identify.

Groundwater and surface water are not isolated components of the hydrologic system. The development or contamination of one usually affects the other. Understanding of the basic principles of interactions between groundwater and surface water is needed for effective management of water resources. In recent years studies of these interactions have expanded in scope to include studies of headwater streams, lakes, wetlands and estuaries (Winter 1995). The interaction between groundwater and lakes has been studied since the 1960s because of concerns related to eutrophication as well as acid rain. Recently, attention has been focused on exchanges between near-channel and in-channel water, which are key to evaluating the ecological structure of stream systems and are critical to stream-restoration and riparian-management efforts.

The quantification of response of surface water quality to groundwater flow and groundwater quantity requires hydrological models attuned to the contamination problem. Surface water-quality key parameters (e.g. residence times) are often ignored in regional water quantity-modelling studies. A comprehensive description of the relation between groundwater and surface water is required to address the impact of agronomic restrictions and intended measures. In regions with shallow groundwater tables and water discharge towards surface water, residence times are strongly influenced by the drain spacing and the depth of the local flow system. A sound description of the link between the local system and the regional system is of great importance for water quality simulations, because the greater part of the final discharge concentration depends on processes within the upper layer of the top system.

Beltman, Boesten and Van der Zee (1995) studied pesticide transport through the unsaturated zone described by an analytical solution of the convection-dispersion equation assuming steady water flow, a linear sorption isotherm and first-order transformation kinetics. Processes and parameters with the greatest impact on the fraction of applied pesticide reaching a drinking-water well were identified. Pesticide arrival in the well was only moderately sensitive to the characteristic travel time and transformation rate in the aquifer. For representative sandy soils under average Dutch rainfall conditions, processes in the unsaturated zone had a much larger impact on pesticide arrival in the wells than processes in the saturated zone. Surface water quality however, may be endangered by the solute migration through fast hydrological pathways such as flow to surface runoff, interflow, flow to drain tubes, flow through macro-pores (Groen 1997; Smelt et al. 2003; Jarvis et al. 2003).

The coherence between water quantity and water quality issues could be guaranteed by using two- or three-dimensional groundwater models. However, these type of models could not be applied at national or supra-regional scale for assessment of water quality parameters. Upscaling of groundwater flow descriptions and conceptualization of the relations involved has resulted to a feasible approach. The SWAP model (Feddes, Kowalik and Zaradny 1978; Belmans, Wesseling and Feddes

1983; Van Dam 2000; Kroes and Van Dam 2003) is a comprehensive one-dimensional physically-based mode for simulating the vertical transport of water, heat and solutes in the saturated and unsaturated top-soil layers. In the Dutch national application (Groenendijk and Boers 1999; Wolf et al. 2003) of the SWAP model the lower boundary is chosen at 13 m depth. The lateral boundary is used to simulate the interaction (discharge or infiltration) with surface-water systems.

Duffy, Kincaid and Huyakorn (1990) identified four general classes of modeling approach: the probabilistic approach, the input-output or systems approach, the multi-dimensional distributed-model approach and the coupled-transport and reaction-modeling approach. Both budget models and reservoir models interrelate with the probabilistic approach and linear-systems approach. Generally, the distributed-model approach and the technique of modeling transport and reactions coincide with the structural-model method. The linear-systems approach and the schematization of discharging zones in the groundwater, as implemented in the SWAP model, are examples of a structural model, which will be addressed in a subsequent section.

An example concerning the upscaling of a groundwater flow field as a method to predict drainage water quality by point models is presented as well as some results of experimental field work on the assessment of drainage water quality impacts of agricultural land management. Strengths, weaknesses, opportunities and threats of different modeling approaches are discussed. The unsaturated zone is an essential link in the chain between land management practices and the ecological status of freshwaters.

Concepts of groundwater – surface-water interactions

The relation between groundwater contamination from non-point sources and the pollutant load on surface waters is of major concern as regards phosphorus and nitrogen components. In the relation between groundwater and surface water pollution, the schematization of the hydrological system is of the utmost importance. A general review on concepts of groundwater – surface-water interactions is provided by Sophocleous (2002).

Groundwater moves along flow paths that are organized in space and form a flow system. In nature, the available subsurface flow domain of a region with irregular topography contains multiple flow systems of different orders of magnitude and relative, nested hierarchical order. Based on their relative position in space, Tóth (Tóth 1963) recognizes three distinct types of flow systems: local, intermediate and regional, which could be superimposed on one another within a groundwater basin. Water in a local flow system flows to a nearby discharge area, such as a field ditch or stream. Water in a regional flow system travels a greater distance than the local flow system, and often discharges to major rivers and lakes. An intermediate flow system is characterized by one or more topographic highs and lows located between its recharge and discharge areas. Areas of pronounced topographic relief tend to have dominant local flow systems, and areas of nearly flat relief tend to have dominant intermediate and regional flow systems (Freeze and Witherspoon 1967). The spatial distribution of flow systems also influences the intensity of natural groundwater discharge. Criteria for the occurrence of flow-systems nesting have been analysed by Zijl and Nawalany (1993) and Zijl (1999).

Hydrologic interactions between surface and subsurface waters occur by subsurface lateral flow through the unsaturated soil and by infiltration into or exfiltration from the saturated zones. Also, in the case of karst or fractured terrain, interactions occur through flow in fracture/solution channels. Water that enters a surface-water body promptly, in response to such individual water input events as rain or snowmelt, is known as event flow, to be distinguished from baseflow: water that enters a stream from persistent slowly varying sources. Beven (1989) defines interflow as the near-surface flow of water within the soil profile resulting in seepage to a stream channel within the time frame of a storm hydrograph. Interflow involves both unsaturated and saturated flows, the latter being in zones of limited vertical extent caused by soil horizons impeding vertical percolation. The mechanisms by which subsurface flow enters streams quickly enough to contribute to streamflow responses to individual rainstorms are summarized in various publications (Beven 1989). Rapid subsurface responses to storm inputs may be the result of fast flow through larger non-capillary soil pores, or macropores (Beven and Germann 1982).

Groundwater exfiltration occurs diffusely or at discrete locations. Perennial, intermittent or ephemeral stream-discharge conditions depend on the regularity of baseflow, which is determined by the groundwater level. For hydraulically connected stream-aquifer systems, the resulting exchange flow is a function of the difference between the river stage and aquifer head. A simple approach to estimate flow is to consider the flow between the river and the aquifer to be controlled by Darcy's law (Rushton and Tomlinson 1979). The assumption of a linear relationship between q and Δh is often too simplistic and takes no account of the decreased resistance to the passage of water as the stream volume increases. This phenomenon has been addressed in the agricultural drainage literature for more than forty years (Ernst 1962) and in the hydrogeological literature by Rushton and Tomlinson (1979). Most hydrological modeling approaches addressing the interaction between groundwater and surface water systems have used one-dimensional or two-dimensional models. Analysis and simulation of the three-dimensional nature of the problem is needed for a better understanding of these interactions. As the computation capacity increases every year, recent modeling efforts attempt to describe the comprehensive unsaturated/saturated soil-water dynamics, the overland flow and the surface-water flow in a fully integrated way. InHM is a physically based, spatially distributed, finite-element, integrated surface-water and groundwater model developed by VanderKwaak (1999), the MikeShe/Mike11 modeling system (Abbott et al. 1986a; 1986b; Refsgaard, Storm and Refsgaard 1995) is a physically based, spatially distributed, finite-difference, integrated surface-water and groundwater model, and WASH123 (WAterSHed systems of 1D Stream, 2D Overland and 3D subsurface media) is capable of simulating surface-water – groundwater interactions and overland flow in a finite-element framework (Yeh et al. 1998), based on FEMWATER (Lin et al. 1997). However, most models operating at a regional scale today are still not well equipped to deal with local phenomena related to flow near domain boundaries (MacDonald and Harbaugh 1988).

Water flows not only in the open stream channel but also through the interstices of stream-channel and bank sediments, thus creating a mixing zone with subsurface water. The region of mixing between subsurface water and surface water is a region of intensified biogeochemical activity (Triska, Duff and Avanzino 1993). The biogeochemical processes within the upper few centimeters of sediments beneath nearly all surface water bodies have a profound effect on the chemistry of groundwater entering surface water, as well as on the chemistry of surface water entering groundwater. Certain species of stream fauna do rely on the upwelling of groundwater for their survival. Many studies have been conducted on the biochemical and water-quality impacts of groundwater – surface-water interactions (Brunke and

Gonser 1997; Dahm et al. 1998).

Conceptualization of two-dimensional groundwater flow

One-dimensional

One-dimensional leaching models generally represent a vertical soil column. In the vertical soil column the ground level provides the upper boundary of the model and the lower boundary can be found at the hydrological basis of the system defined. The lateral boundary consists of one or more different drainage systems. The position of lower and lateral boundaries depends on the scale and type of model application.

Hydrological data, such as water fluxes and moisture contents of the distinct soil layers, are supplied by an external field-plot model (Kroes and Van Dam 2003) or a regional groundwater flow model (Querner and Van Bakel 1989; Van Walsum et al. 2004). The schematization of the soil profile and the main terms of the water balance for a particular drainage situation are depicted in Figure 1. Within the unsaturated zone, chemical substances are transported by vertical flows, whereas in the saturated zone the drainage discharge leaves the vertical column sideways. The drainage fluxes are

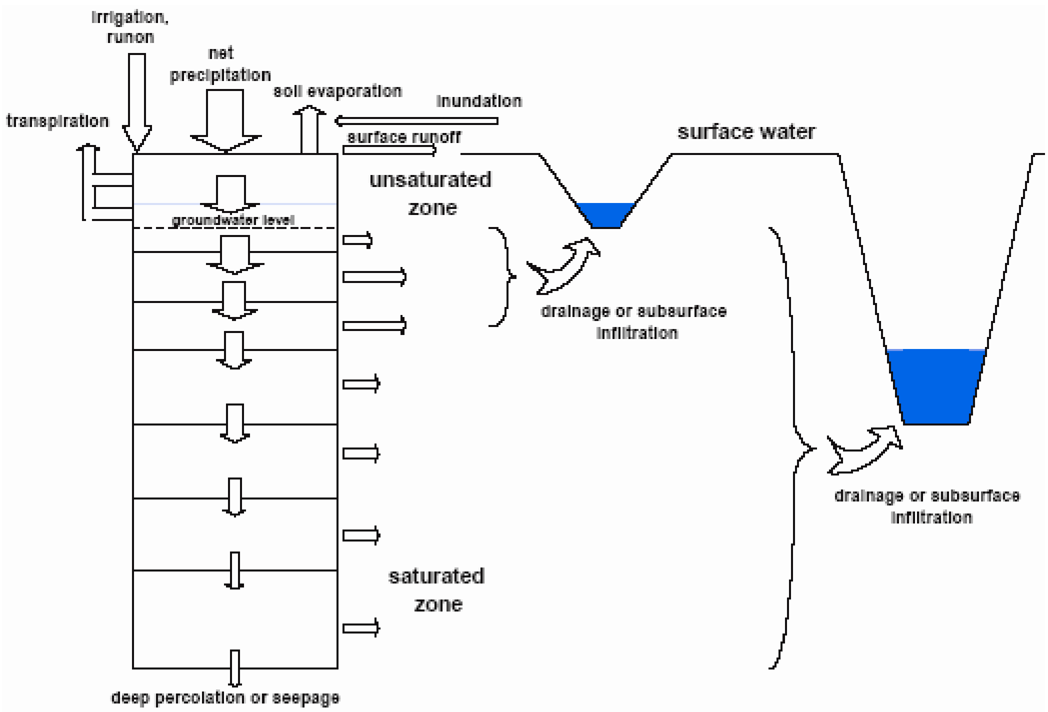

Figure 1. Schematization of water flows in a soil profile and the main terms of the water balance

considered lateral-sink terms $R_{dr}\,[T^{-1}]$ in the water balance. The distribution of lateral-drainage sink terms can be derived from the vertical flux (q_y) relation with depth according to:

$$\frac{dq_y}{dy} = -R_{dr} \qquad (1)$$

Two-dimensional

The main direction of the space co-ordinates (x, y) and the flux density (q_x, q_y) is defined positive from left to right and in upward direction (Figure 2). In our derivation of expressions for drainage-water travel times we consider only steady-flow conditions and homogeneous aquifer characteristics, and recharge between $x = \frac{B}{2}$ and $x = \frac{L}{2}$ is uniformly distributed at $y = 0$. Water flows to a drain with certain depth located between $x = 0$ and $x = \frac{B}{2}$, and additional point or line sources/sinks are disregarded. The relationship between the flux density and groundwater head is given by Darcy's law. For a homogeneous porous medium the following expressions hold:

$$q_x = -k_x \frac{\partial h}{dx} \text{ and } q_y = -k_y \frac{\partial h}{dy} \tag{2}$$

The potential ϕ is defined as the product of the groundwater head and the hydraulic conductivity vector: $\phi = k\, h$. The stream function has the property that the amount of water that passes a line between two points per unit of time is identical to the stream-function difference between those points (Van der Veer 1979):

$$q_x = -\frac{\partial \Psi}{\partial y} \text{ and } q_y = \frac{\partial \Psi}{\partial x} \tag{3}$$

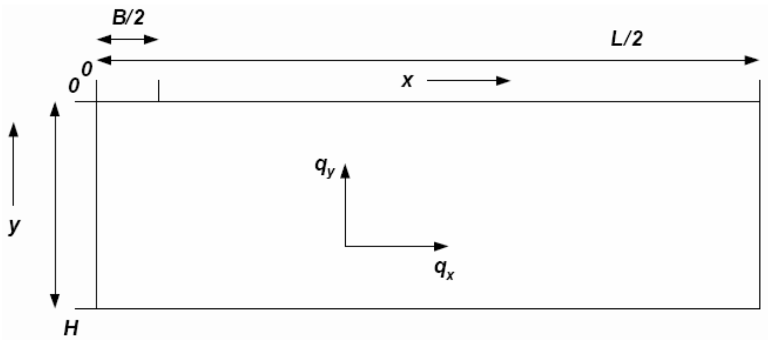

Figure 2. Definitions of place variables and flux densities

When a value for Ψ for a certain point (x, y) has been established, the complex potential function Φ can be composed of the potential ϕ and the stream function Ψ according to:

$$\Phi = \phi + i\Psi \tag{4}$$

Because ϕ and Ψ are conjugate harmonic functions of the space co-ordinates x and y, the complex potential Φ is an analytical function of the complex variable $z = x + iy$. In a two-dimensional flow field streamlines are mathematically described by the stream function $\Psi(x, y)\, [L^2T^{-1}]$. The stream function is discussed in several textbooks (Bear 1972; Strack 1984) and for a general discussion the reader is referred to these sources. Several extensions of the stream function and the associated stream-tube method have been published (Bear 1972; Zijl 1984). For a given value $\Psi(x, y) = \Psi_0$ streamlines can be constructed from the points (x, y) for which the stream function equals Ψ_0.

A groundwater flow field can be described by means of a vector function of the fluxes (Bear and Verruijt 1987). The equations describing the motion of a fluid particle in this flow field are:

$$q_x(x,y) = \varepsilon \frac{dx}{dt};\ q_y(x,y) = \varepsilon \frac{dy}{dt} \tag{5}$$

Rewriting yields:

$$dt = \frac{\varepsilon dx}{q_x(x,y)} = \frac{\varepsilon dy}{q_y(x,y)} \tag{6}$$

Assume that a fluid particle is at time t_0 in (x_0, y_0). The time it takes for this particle to reach a depth y_1 can be calculated by:

$$t - t_0 = \int_{y_0}^{y} \frac{\varepsilon dy}{q_y(x,y)} \tag{7}$$

where the integral should be taken over the streamline connecting the begin point with the end point. The relation of the vertical flux with depth determines implicitly the travel-time distribution of exfiltrating groundwater. A projection of the two-dimensional flow field on the vertical axis can be found by defining an *upscaled vertical flux* as a function of height $\overline{q}_y(y)$ for which the travel time is defined by:

$$t - t_0 = \int_{y_0}^{y} \frac{\varepsilon dy}{\overline{q}_y(y)} \tag{8}$$

The *upscaled vertical flux* $\overline{q}_y(y)$ is not dependent on the horizontal distance x. Let Q_d be the total discharge per unit length $[L^2\, T^{-1}]$ perpendicular to the plane of cross-section. A streamline for which $\Psi(x,y) = \Psi_0$ separates the flow field into two zones. Through the first zone a discharge equal to Ψ_0 is conveyed and the other zone drains the remaining part of the total discharge ($Q_d - \Psi_0$).For each streamline the deepest point can be found. The deepest point of each streamline for which holds that a certain part of the total discharge which will never pass that depth.

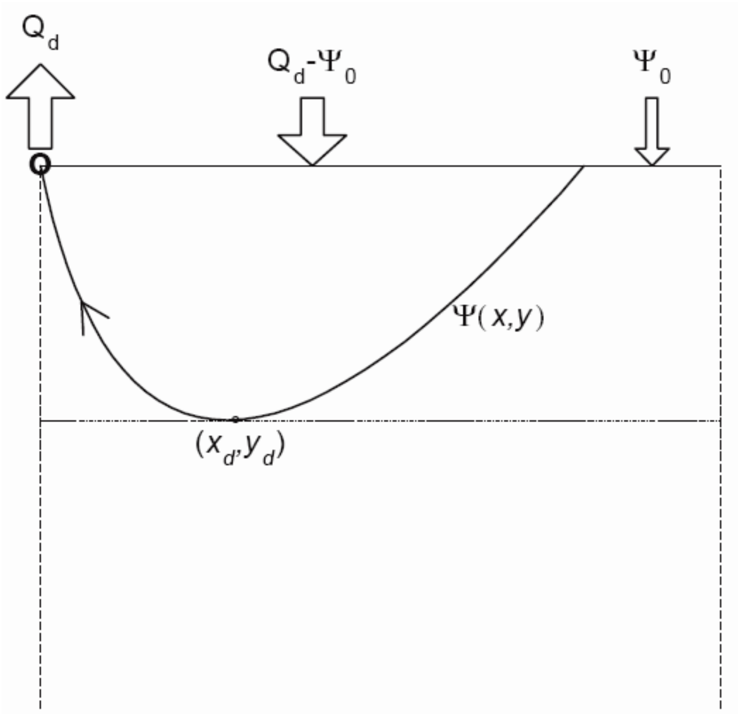

Figure 3. Separation of two-dimensional flow field by a streamline into two zones

This rule can be used to determine the x co-ordinate at a certain depth for which the vertical flux intensity on a streamline equals zero. For the deepest point of each streamline in our simplified flow field at depth y_d holds:

$$q_y(x_d, y_d) = 0 \tag{9}$$

This yields an expression for x_d as a function of y_d and Ψ for the flow domain considered. The total recharge entering the flow domain between $x = \frac{B}{2}$ and $x = \frac{L}{2}$ can be divided into a part passing through the plane at y_d and another part that has left the soil profile already before reaching the plane at y_d. Let the part of the drain discharge which will never pass through level y_d amount to $\Psi(x_d, y_d)$. In the schematized vertical soil column the remaining part $Q_d - \Psi(x_d, y_d)$ passes the plane at y_d as vertical flow. The average vertical flux at depth y_d is then found by substitution of x_d and y_d in the stream-function relation and division by half the drain spacing:

$$\bar{q}_y(y_d) = \frac{Q_d - \Psi(x_d, y_d)}{\frac{L}{2}} \tag{10}$$

Multilevel drain discharge

Upscaling of the relation between the residence-time distribution and the determining key parameters to assess the regional response of changed inputs on drainage water quality has been reported by several authors (Luther and Haitjema

1998; Van den Eertwegh 2002). The approach most widely applied is to consider a region a collection of independent fields and to calculate the drainage water quality as a flow weighted average. The interdependency between flow patterns which may arise when nested flow systems occur is often not accounted for in lumped models which describe the solute breakthrough of exfiltrating groundwater as a mixing process in a linear reservoir (Van der Molen and Van Ommen 1988). The hierarchical distribution of exfiltration lines or points as well as the influence of biochemical reactions on the concentration behavior necessitates distinguishing between the hydraulic and chemical properties of different soil layers. The SWAP model (Kroes and Van Dam 2003) has the ability to calculate lateral-sink terms for a multilevel drainage system. The flow pattern is schematized to a unilateral flow to perfect drains ignoring radial flow components in the vicinity of line drains.

In the drainage model describing the discharge to parallel equidistant water courses, the discharge flow $Q_{d,i}$ is defined as:

$$Q_{d,i} = q_{d,i} \frac{A}{\Sigma l_i} \tag{11}$$

where $q_{d,i}$ is the drainage flux density of system i; A is the area of the catchment and Σl_i is the total length of drainage system i. An essential assumption made in the SWAP model (Kroes and Van Dam 2003) is the proportionality of the ratio between the occupied flow volumes V_i and the ratio between the discharge rates $Q_{d,i}$:

$$\frac{V_i}{V_{i-1}} = \frac{Q_{d,i}}{Q_{d,i-1}} \tag{12}$$

The classification of the line drains and the compilation of the hierarchy allows for the superposition of drainage fluxes. First-order drains act also as second- and higher-order drains. Figure 4 depicts the schematization of the regional groundwater flow, including the occupied flow volumes for the nested drain systems. The volume V_i consists of summed rectangles $L_i H_i$ of superposed drains, where H_i is the thickness $[L]$ of discharge layer i. The method employed by the SWAP model has been based on the assumption that the groundwater volume occupied between the upper boundary and the zones indicated by H_1 and H_2 can be neglected. This assumption is rather questionable.

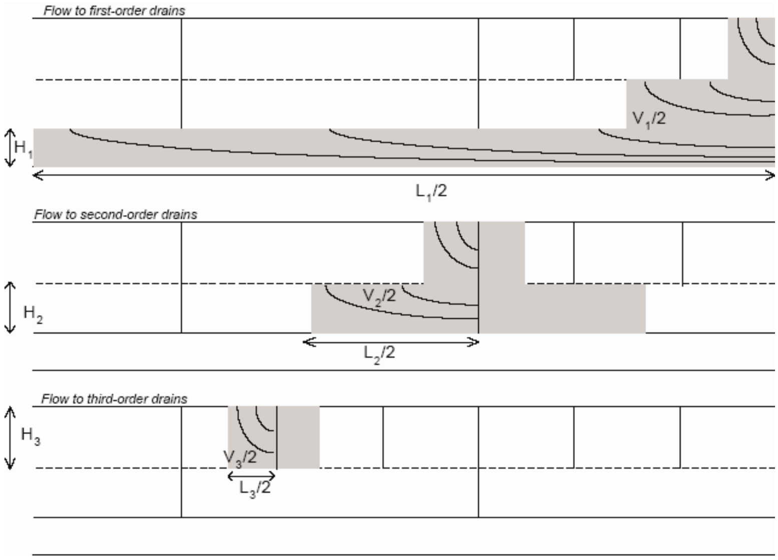

Figure 4. Schematization of groundwater flow to three different types of line drains

If n is the total number of drainage classes distinguished, the flow volume V_i assigned to a drainage system i is related to drain distances L_i and thickness H_i of discharge layers as follows:

$$V_i = \sum_{j=i}^{n} L_j H_j \tag{13}$$

Using the proportionality between occupied flow volumes and flow discharges and rearranging Eq. 13 yields:

$$L_1H_1 \ : \ L_2H_2 \ : \ L_3H_3 = (q_{d,1}\frac{A}{\Sigma l_1} - q_{d,2}\frac{A}{\Sigma l_2}) : (q_{d,2}\frac{A}{\Sigma l_2} - q_{d,3}\frac{A}{\Sigma l_3}) : \ q_{d,3}\frac{A}{\Sigma l_3} \tag{14}$$

If the horizontal conductivities exhibit a stratified constitution, the heterogeneity can be taken into account by substituting transmissivities T for layer thicknesses. The thickness of a certain layer can be derived by considering the vertical cumulative transmissivity relation with depth as depicted in Figure 5. In Figure 5 the average groundwater level is indicated by φ_{avg}. The lateral-flux relation per unit soil depth shows a uniform distribution. Lateral drainage fluxes $q_{d,i,k}$ to drainage system i for each nodal compartment k of the simulation model are calculated by the drainage flux $q_{d,i}$ multiplied by the ratio between the transmissivity of nodal compartment k and the total transmissivity of the *discharge layer* with depth H_i. The average concentration in the discharge water to a line drain results from flux weighted $(q_{d,i,k})$ averaging the nodal concentrations.

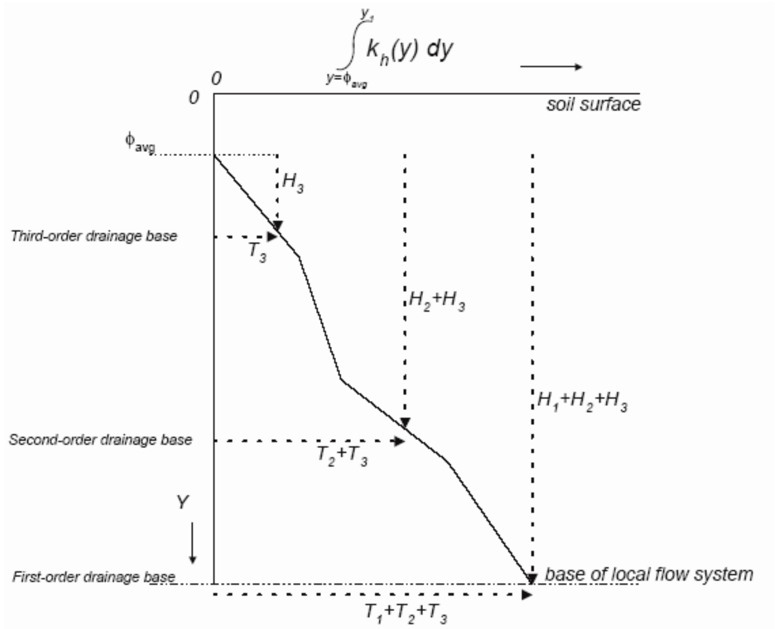

Figure 5. Discharge layer thickness H_i as function of cumulative transmissivity T_i in a heterogeneous soil profile

Analytical solutions for some selected cases

Analysis of flow patterns will allow predictions of both the groundwater quality and the composition of the outflow from the system. The classical two-dimensional flow patterns in vertical cross-sections of water flowing to equidistant and parallel drains may be applied for this purpose. Elementary analytical solutions for the travel time have been summarized by Ernst (1973) and De Vries (1975). The groundwater flow in an aquifer with perfect drains and the flow to parallel drains in an infinitely deep aquifer will be discussed.

Discharge to perfect drains

The unidirectional flow model based on the Dupuit assumption provides the simplest basis for estimating travel times. Many authors (Gelhar and Wilson 1974; Raats 1978; Van Ommen 1986; Meinardi 1994; Haitjema 1995; Rijtema, Groenendijk and Kroes 1999; Van den Eertwegh 2002) took this model as a starting point for the derivation of the travel time as a function of the distance to the drain. However, it can be demonstrated that for groundwater flow in an aquifer drained by fully penetrating drainage canals, the solution to Eq. 8 based on two-dimensional groundwater flow is identical to the solution based on the unidirectional Dupuit-assumption. Ernst (1962; 1973) utilized the Dupuit assumption but considered a two-dimensional flow field by using complex variables:

$$\Phi = \frac{R}{2H}(z - z_0)^2 \tag{15}$$

where R $[LT^{-1}]$ is the recharge of the aquifer and z_0 is given by: $z_0 = \frac{L}{2} - iH$. Expressions for the stream function and the vertical flux density can be derived:

$$\Psi(x,y)=\frac{R}{H}(x-\frac{L}{2})(y+H) \tag{16}$$

and

$$q_y(x,y)=\frac{R}{H}(y+H) \tag{17}$$

The deepest point on each streamline is found at the outflow boundary at $x=0$. Consequently $x_d=0$. For the total discharge for the area between the drain and midway between the drains holds: $Q_d=R\frac{L}{2}$. Substitution of $x_d=0$, the discharge and Eq. 16 into Eq. 10 yields the following relation for the one-dimensional vertical flux (Figure 6):

$$\overline{q}_y=R\frac{H+y}{H} \tag{18}$$

Accordingly, the time t_1 to reach a depth y_1 after infiltration at $t_0=0$ at depth $y=0$ holds (Eq. 8):

$$t_1=\frac{\varepsilon H}{R}\int_0^{y_1}\frac{dy}{H+y}=\frac{\varepsilon H}{R}\ln\left(\frac{H+y_1}{H}\right) \tag{19}$$

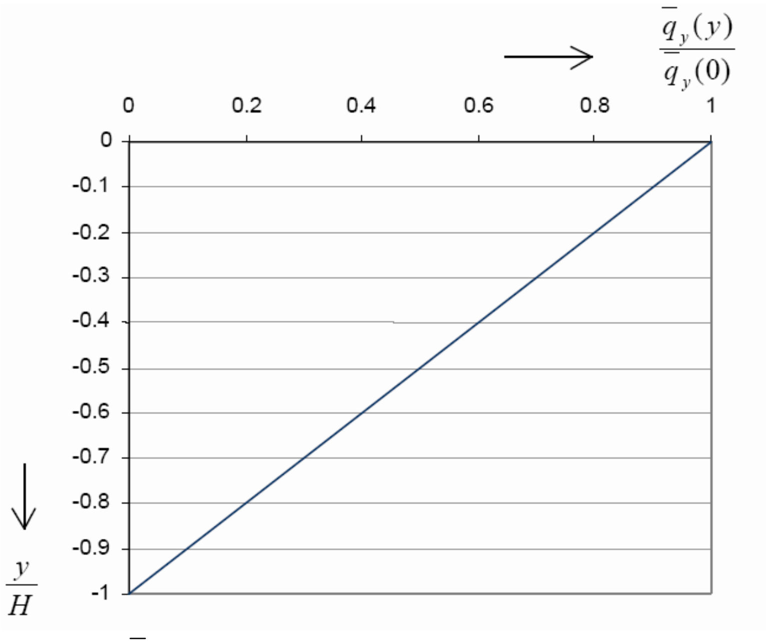

Figure 6. Vertical flux $\overline{q}_y$ in an aquifer with perfect drains

Discharge to line drains in aquifers of infinite thickness

For a drainage system with a small spacing relative to the aquifer thickness the Dupuit assumption does not hold. Relations for hydraulic head as a function of aquifer thickness, radius of drainage mean, drain spacing and steady-state recharge to be used in design practice have been derived by Hooghoudt (1940), Kirkham (1966) and Ernst (1962) without assuming *a priori* unidirectional flow. An aquifer with infinite thickness can be considered an extreme situation which will never occur. An analysis of the travel time of this extreme situation provides understanding of the maximum deviation from the ideal situation as expressed by Eq. 19. The complex potential function has been given by Ernst (1973):

$$\Phi = -\frac{R\,L}{\pi}\ln\left(\frac{\exp(\frac{2\pi\, i\, z^*}{L})-1}{2}\right) \tag{20}$$

with $z^* = x - iy$. Expressions for the stream function and the vertical flux are given by:

$$\Psi(x,y) = -\frac{RL}{\pi}\arctan\left(\frac{-\exp(\frac{2\pi y}{L})\sin(\frac{2\pi x}{L})}{\exp(\frac{2\pi y}{L})\cos(\frac{2\pi x}{L})-1}\right) \tag{21}$$

and

$$q_y(x,y) = R\frac{\cos\left(\frac{2\pi x}{L}\right)-\exp\left(\frac{2\pi y}{L}\right)}{\cos\left(\frac{2\pi x}{L}\right)-\cosh\left(\frac{2\pi y}{L}\right)} \tag{22}$$

At a certain height y_d , the deepest point of a streamline is found by stating $q_y(x_d, y_d) = 0$. This condition is met when the denominator of Eq. 22 equals 0:

$$x_d = \frac{L}{2\pi}\arccos\left(\exp(\frac{2\pi y_d}{L})\right) \tag{23}$$

Substitution of this expression and Eq.16 into Eq. 10 yields for the vertical flux $\bar{q}_y$:

$$\bar{q}_y(y) = R\frac{2}{\pi}\arctan\left(\frac{\exp(\frac{2\pi y}{L})}{\sqrt{1-\exp(\frac{4\pi y}{L})}}\right) \tag{24}$$

The relation is graphically presented in Figure 7. It appears from Figure 7 that in the upper layers of the saturated zone a considerable amount of total drainage volume is discharged. Nearly half of the drain discharge is conveyed above the plane at $y/L = 0.1$. For the travel time t_1 after infiltration at $t_0 = 0$ at depth $y = 0$ holds:

$$t_1 = \int_0^{y_1}\frac{\varepsilon dy}{R\frac{2}{\pi}\arctan\left(\frac{\exp(\frac{2\pi y}{L})}{\sqrt{1-\exp(\frac{4\pi y}{L})}}\right)} \tag{25}$$

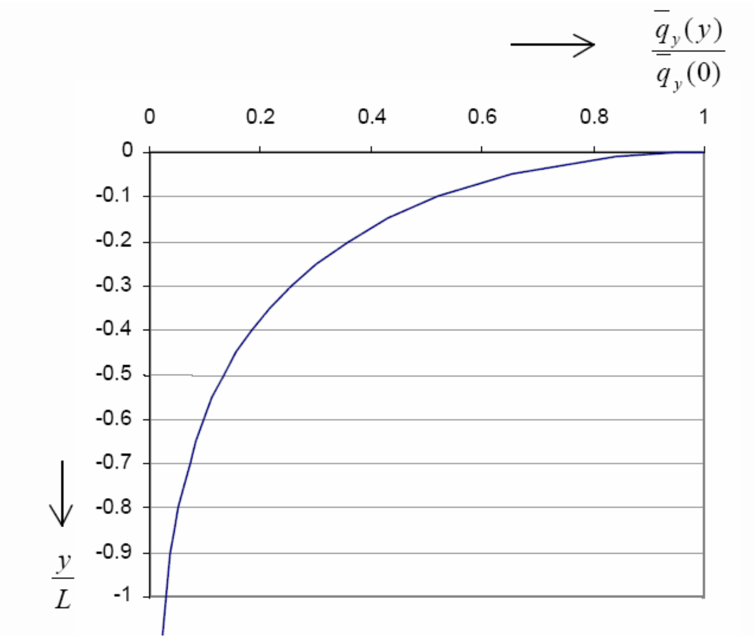

Figure 7. Vertical flux $\overline{q}_y$ in an aquifer with infinite thickness and line drains

Direct solutions to this integral equation for verification purposes are hard to find. An indirect proof by applying the expression for $\overline{q}_y$ in a solute breakthrough simulation will be given in a subsequent section. Ernst (1973), Raats (1978), Van der Molen (1987) and Van der Molen and Van Ommen (1988) present a solution for the transit time to be distributed according to:

$$t = \left(\frac{\varepsilon L}{2R}\right)\left(\frac{\pi x}{L}\right)\tan\left(\frac{\pi x}{L}\right) \qquad (26)$$

Elaboration provides an impulse–response relation for solute breakthrough to be used for validation of Eq. 25.

Application to solute transport

In an aquifer with perfect drains construction of isochrones for solute displacement after uniform infiltration at the phreatic level yields horizontal lines, because the vertical fluxes do not depend on the horizontal distance relative to the origin. The isochrones are regarded as imaginary boundaries between soil layers and each of the soil layers may be regarded as a perfectly mixed reservoir (Groenendijk and Roest 1996). Part of the inflow is conveyed to underlying soil layers, the remainder flows horizontally to the water course or drainage tube.

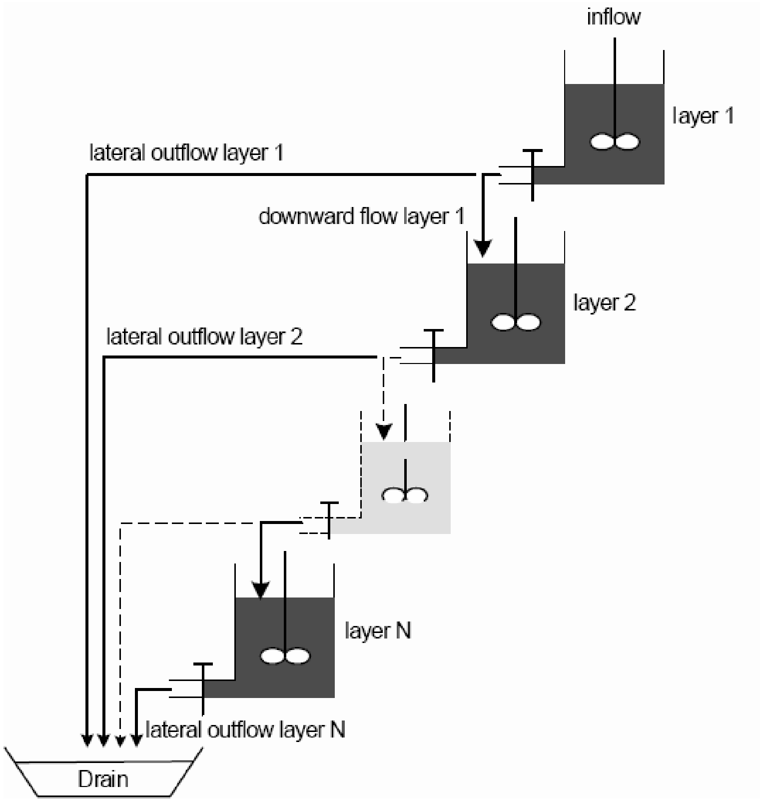

Figure 8. Solute migration in an aquifer represented by the flow through a series of perfect mixed reservoirs

The upscaled one-dimensional solute flux J_s of an inert substance through the aquifer is the sum of an advective flux and a dispersive flux attributed to the combination of molecular diffusion and dispersional mixing:

$$J_s = \bar{q}_y \, c - \varepsilon \, D \frac{\partial c}{\partial y} \tag{27}$$

where D is the dispersion coefficient [$L^2 \, T^{-1}$]. The dispersion coefficient is assumed linearly proportional to the water velocity according to $D = \lambda \frac{\bar{q}_y}{\varepsilon}$, where λ is the dispersion length [L]. Relating solute flux to mass conservation gives:

$$-\frac{\partial J_s}{\partial y} = -\frac{\partial}{\partial y}(\bar{q}_y c - \varepsilon D \frac{\partial c}{\partial y}) - R_{dr} c = -\frac{\partial}{\partial y} \bar{q}_y (c - \lambda \frac{\partial c}{\partial y}) - R_{dr} c \tag{28}$$

Space is discretized in compartments and Taylor expansion is used to translate the differential equation into a finite differenced expression where second- and higher-order terms are ignored. The concentrations are defined at the center of each compartment and have the subscript i. The inflowing and outflowing water flux are considered at the top and the bottom of compartment i, respectively, and are defined as $\bar{q}_{i-\frac{1}{2}}$, and $\bar{q}_{i+\frac{1}{2}}$, respectively. The concentration at the interface between compartments $i-1$ and i is calculated as the weighted average of c_{i-1} and c_i. For steady-state flow conditions the partial derivative for transport can be developed as follows:

$$-\frac{\partial J_s}{\partial y} \approx -\frac{\bar{q}_{i+\frac{1}{2}}}{\Delta y_i}\left(\frac{\Delta y_i - 2\lambda}{\Delta y_i + \Delta y_{i+1}}\right)c_{i+1} + \left\{\frac{\bar{q}_{i+\frac{1}{2}}}{\Delta y_i}\left(\frac{\Delta y_{i+1} - 2\lambda}{\Delta y_i + \Delta y_{i+1}}\right) - \frac{\bar{q}_{i-\frac{1}{2}}}{\Delta y_i}\left(\frac{\Delta y_{i-1} + 2\lambda}{\Delta y_i + \Delta y_{i-1}}\right) - R_{dr}\right\}c_i + \frac{\bar{q}_{i-\frac{1}{2}}}{\Delta y_i}\left(\frac{\Delta y_i + 2\lambda}{\Delta y_{i-1} + \Delta y_i}\right)c_{i-1} \tag{29}$$

Diffusion/dispersion can be described by utilizing the numerical dispersion which results from this calculation scheme. Taking identical thicknesses and defining the dispersion length λ as half of the compartment thickness reduces the expression into:

$$-\frac{\partial J_s}{\partial y} \approx -\left(\frac{\bar{q}_{i-\frac{1}{2}}}{\Delta y_i} + R_{dr}\right)c_i + \frac{\bar{q}_{i-\frac{1}{2}}}{\Delta y_i}c_{i-1} \tag{30}$$

Utilization of this spatial discretization and preserving the conservation term as differential quotient yields a set of first-order linear differential equations which can be solved analytically.

$$\begin{pmatrix} \frac{dc_1}{dt} \\ \frac{dc_2}{dt} \\ \vdots \\ \frac{dc_{N-1}}{dt} \\ \frac{dc_N}{dt} \end{pmatrix} = \begin{pmatrix} -\frac{R}{\varepsilon_1 \Delta y_1} & 0 & \cdots & 0 & 0 \\ \frac{\bar{q}_y(y_1)}{\varepsilon_2 \Delta y_2} & -\frac{\bar{q}_y(y_1)}{\varepsilon_2 \Delta y_2} & \cdots & 0 & 0 \\ \vdots & \vdots & \ddots & \vdots & \vdots \\ 0 & 0 & \cdots & -\frac{\bar{q}_y(y_{N-2})}{\varepsilon_{N-1}\Delta y_1} & \frac{\bar{q}_y(y_{N-2})}{\varepsilon_{N-1}\Delta y_{N-1}} \\ 0 & 0 & \cdots & \frac{\bar{q}_y(y_{N-1})}{\varepsilon_N \Delta y_N} & -\frac{\bar{q}_y(y_{N-1})}{\varepsilon_N \Delta y_N} \end{pmatrix} \begin{pmatrix} c_1 \\ c_2 \\ \vdots \\ c_{N-1} \\ c_N \end{pmatrix} + \begin{pmatrix} \frac{R}{\varepsilon_1 \Delta y_1} c_{inp} \\ 0 \\ \vdots \\ 0 \\ 0 \end{pmatrix} \tag{31}$$

The solution to this set of differential equations is found by applying linear-algebra theory (Cullen 1979). The flux-weighted average concentration in groundwater exfiltration water reads:

$$\bar{c}(t) = \frac{\sum_{i=1}^{n} c_i(t)\left\{\bar{q}_y(y_{i-1}) - \bar{q}_y(y_i)\right\}}{q_{dr}} \tag{32}$$

This simple model is applied to an aquifer with perfect drains, assuming n reservoirs with identical thickness and an initial concentration c_0 in each reservoir. The solution to the differential equations yields the concentration course over time in reservoir j (Groenendijk and Roest 1996):

$$\frac{c_j(t) - c_{inp}}{c_0 - c_{inp}} = \sum_{i=1}^{j}\binom{n}{i-1}\binom{n-i}{j-i}(-1)^{i+1}\exp\left(-\frac{(n-i+1)R}{\varepsilon H}t\right) \tag{33}$$

In the aquifer with perfect drains, the vertical flux density is a linear relation with depth (see Eq. 18). Thus the lateral outflow is uniformly distributed with depth. Since the thicknesses of the reservoir are assumed to be identical, the resulting outflow and consequently the resulting output concentration can be found as the arithmetic mean of all reservoir concentrations. Lengthy but straightforward summation of the binomial series in Eq. 33 yields a simple relation for the drainage-water concentration (Figure 9):

$$\frac{\bar{c}(t)}{c_{inp}} = 1 - \exp(-\frac{R}{\varepsilon H} t) \tag{34}$$

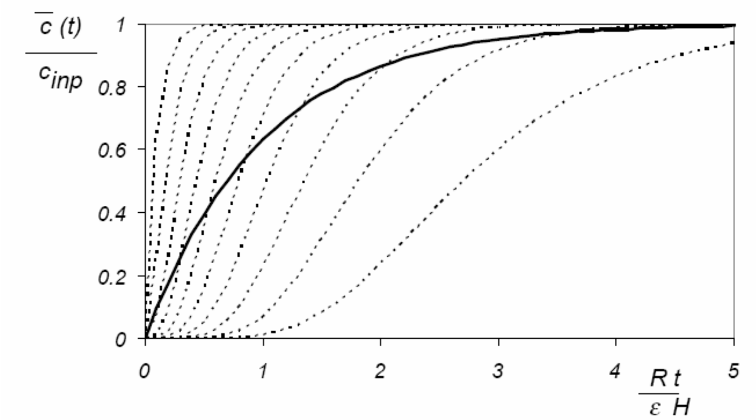

Figure 9. Concentration in groundwater compartments (dashed lines) and break through curve of drainage water concentration (solid line)

Figure 9 results from a simulation of solute displacement in an aquifer subdivided into 10 layers with an identical thickness and porosity. The time is scaled to a dimensionless variable. The residence time in the first compartment equals one tenth of the overall residence time ($T_1 = \frac{\varepsilon \frac{H}{10}}{R}$), whereas the residence time of the last reservoir T_{10} equals ($\frac{\varepsilon \frac{H}{10}}{\frac{R}{10}} = \frac{\varepsilon H}{R}$). The breakthrough curves show the effect of numerical dispersion which was introduced by assuming perfect mixing in the groundwater layers. However, since the resulting average drainage water concentration is identical to the concentration derived by a piston flow model, the total effect of vertical dispersion that has been introduced by defining distinct soils layers can be neglected. This result agrees with findings of Ernst (1973), Gelhar and Wilson (1974) and Raats (1978), who showed that the breakthrough curve of a field with fully penetrating drainage canals is identical to the breakthrough curve of a completely mixed reservoir. According to Van Ommen (1986) it can be demonstrated that Eq. 34 is a valid approximation for any flow for which the Dupuit assumption holds.

The underlying processes are completely different. Perfect mixing in a reservoir is completely absent in aquifers where piston flow solute migration is assumed. The analogy should not be transferred to relations involving a solute front (Van der Molen and Van Ommen 1988). Also when heterogeneously distributed soil properties as nonlinear sorption behavior influence the groundwater concentration the use of Eq. 34 is dissuaded.

Considering the solute migration to a line drain in an aquifer of infinite thickness, the total drain discharge is apportioned into 10 identical fractions. Thus, the outflow of each reservoir is assumed identical and the resulting drain concentration may be calculated by taking the arithmetic mean of the reservoir concentrations. The thickness of the reservoirs can be derived from Eq. 24. The relation between cumulative thickness and cumulative drain discharge of the array of reservoirs should be attuned to the $\overline{q}_y(y)$ relation. Table 1 denotes the characteristics of a simulation with the 'cascade of reservoirs' model.

Since the aquifer has an infinite thickness, the bottom boundary of the last layer is approximated by the depth at which $\overline{q}_y(y)/\overline{q}_y(0)$ equals 1%.

Table 1. Cascade reservoir characteristics for simulation of solute breakthrough in an infinite deep aquifer drained by a line drain

No.	$\frac{\overline{q}_y(y=y_{i-1})}{\overline{q}_y(0)}$	Δy	Diagonal coefficient
1	1	$0.004\ \frac{L}{2}$	$253.598\ \frac{2R}{\varepsilon L}$
2	0.9	$0.012\ \frac{L}{2}$	$74.812\ \frac{2R}{\varepsilon L}$
3	0.8	$0.021\ \frac{L}{2}$	$38.534\ \frac{2R}{\varepsilon L}$
4	0.7	$0.031\ \frac{L}{2}$	$22.781\ \frac{2R}{\varepsilon L}$
5	0.6	$0.043\ \frac{L}{2}$	$14.000\ \frac{2R}{\varepsilon L}$
6	0.5	$0.059\ \frac{L}{2}$	$8.499\ \frac{2R}{\varepsilon L}$
7	0.4	$0.082\ \frac{L}{2}$	$4.865\ \frac{2R}{\varepsilon L}$
8	0.3	$0.122\ \frac{L}{2}$	$2.450\ \frac{2R}{\varepsilon L}$
9	0.2	$0.217\ \frac{L}{2}$	$0.923\ \frac{2R}{\varepsilon L}$
10	0.1	$\approx\ 0.732\ \frac{L}{2}$	$\approx\ 0.137\ \frac{2R}{\varepsilon L}$

Results of the simulation are presented in Figure 10. Due to the large difference in response between the top layers and the deeper layers, the time variable at the horizontal axis is given on a logarithmic scale.

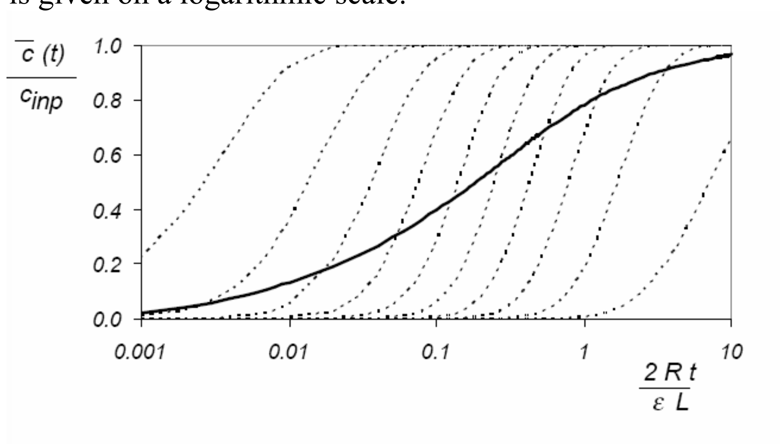

Figure 10. Concentration respons to input cinp in aquifer of infinite thickness (dashed lines represent the concentration in groundwater compartments and solid line represent the break through curve of drainage water)

The result can be validated with the analytical expression given by Ernst (1973) and Van der Molen and Van Ommen (1988):

$$\frac{2Rt}{\varepsilon L} = \frac{1}{2}\frac{\bar{c}(t)}{c_{inp}}\tan\left(\frac{\pi}{2}\frac{\bar{c}(t)}{c_{inp}}\right) \quad (35)$$

It appears from Figure 11 that the results of the reservoir model fit reasonably well to the analytical solution. The maximum deviation occurs at $\frac{2Rt}{\varepsilon L} \approx 0.1$ and amounts to approximately 5% of the input concentration, which is sufficient for practical applications. The choice of vertical flux values and thickness of groundwater layers implies a linearization of the vertical flux relation as presented in Figure 7. The concentration range which shows the maximum deviation corresponds to the region where the linearized vertical flux deviates most from the analytical solution (Eq. 24). The results of the reservoir model show a faster breakthrough. The faster breakthrough can be attributed to the composition of reservoir volumes. The very thin layers near the water table are approximated by completed stirred reservoirs and allow the solute to pass too quickly**.** More accurate results can be obtained by extending the cascade with some more compartments and by choosing the mutual volume ratios according to expected dispersivity. The result can be validated with the analytical solution given by Ernst (1973) and Van der Molen and Van Ommen (1988):

$$\frac{2Rt}{\varepsilon L} = \frac{1}{2}\frac{\bar{c}(t)}{c_{inp}}\tan\left(\frac{\pi}{2}\frac{\bar{c}(t)}{c_{inp}}\right) \quad (36)$$

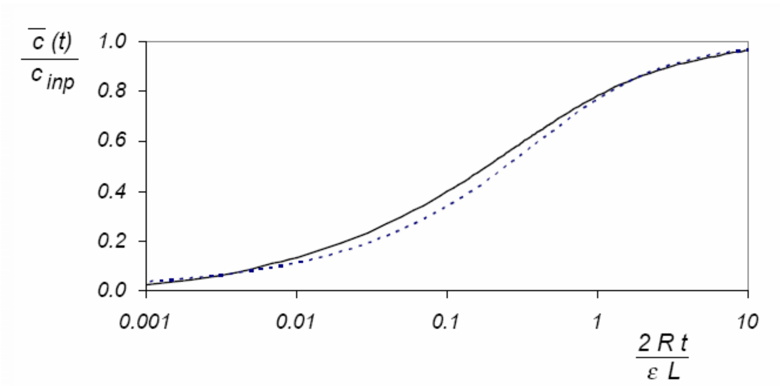

Figure 11. Concentration in drainage water of an aquifer with inifinite thickness simulated by the cascade of reservoirs model (solid line) and calculated with the analytical solution (dashed line)

Age-class approach

Theory

Systems analyses, based on the impulse–response relationships, have been used to predict surface water loads and drainage concentrations (Jury 1975). The method presupposes linearity of the relation between groundwater quality and discharge concentrations. An example of the response of the system to a unit step function is given in Figure 12. The relation can be regarded as the breakthrough curve of drainage water after continuously contaminating an initially clean groundwater system. The reaction of a linear system to a time-dependent input can be obtained by separation of a discrete form of the block response into unit-salinographs, which are determined by Jury (1975) and Van Ommen (1986):

$$\mu(\tau,t) = S(t) - S(t-\tau) \quad \text{for}: t > \tau \tag{37}$$

These time salinographs represent the influence of the input concentration on the output between the times t and $t-\tau$. Assuming the time axis to be divided into n time intervals with identical length, the drainage water concentration at time t equals:

$$\bar{c}(t) = \sum_{i=1}^{i=n} \mu(\tau, i\cdot\tau) c_{inp}(n-i+1)\,, \tag{38}$$

where $c_{inp}(i)$ is the input concentration to the groundwater system.

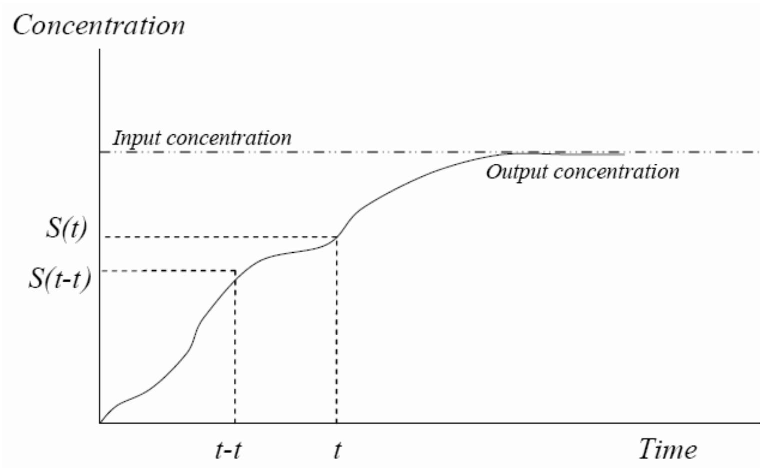

Figure 12. Example of a step response function for concentration breakthrough

Once the function $\mu(\tau,t)$ has been deduced, estimates of the surface water load can easily be made. The function can be derived for solute migration which obeys mathematical linearity rules. This implies the description of sorption according to a linear sorption isotherm and transformation according to a first-order decay relation. If the dynamics of groundwater flow greatly influence the resulting discharge concentrations, the method will be invalid. The method presupposes the availability of input concentrations, resulting from the unsaturated zone. For phosphate and nitrogen, this variable depends to a great extent on processes within the root zone and the unsaturated subsurface soil. A large proportion of the phosphorus load on surface water originates from the zone in the vicinity of the mean highest water table. Linearization of transport processes in such a dynamic situation will lead to predictions with low reliability.

The age-class approach has been proposed by Meinardi (1994) for analysis of tritium concentrations in groundwater of sandy regions and has been further elaborated by Van den Eertwegh (2002) for the analysis of drainwater concentrations as a function of meteorologic variation and land use. Both authors refer to Ernst (1973) and Bruggeman (1999). Meinardi (1994) and Van den Eertwegh (2002) based their approach on the logarithmic travel-time distribution relation as a function of depth, which results from a schematization not considering the flow domain above drain level. Such a relation has been used before to describe flow velocities as a function of depth (Hoeks 1981; Strack 1984). For sandy soils in lowland regions this approach may be adequate, but special hydrological situations may occur for drained loamy and clay soils and for sloping fields where the flow volume above the drain should be taken into account.

A relation for the travel time as function of distance relative to the drain can be derived considering the Dupuit assumption and steady-state conditions (Figure 13).

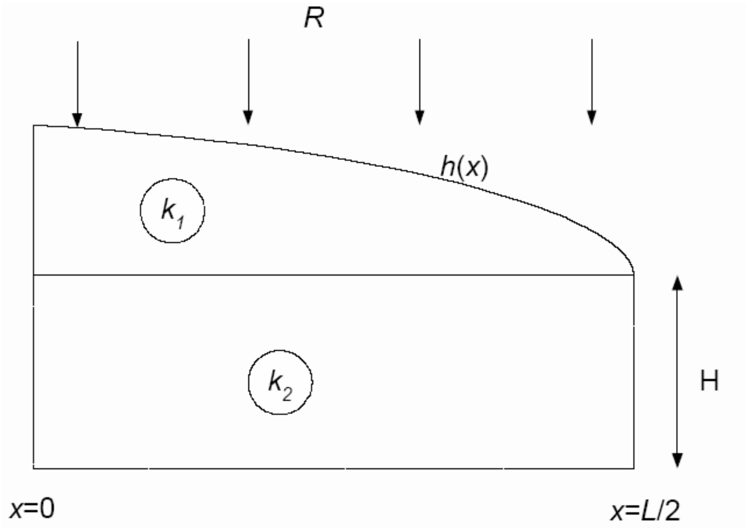

Figure 13. Schematisation of soil profile into a zone above drainlevel with thickness $h(x) - H$ and a zone below drainlevel with thickness H

Two zones are distinguished: the top boundary of the first zone is determined by the groundwater elevation $h(x)$ and the bottom coincides with the drain level. The second zone has a permanent thickness H and its conductivity equals k_2. The groundwater elevation $h(x)$ relative to the hydrologic basis can be found from:

$$\frac{d}{dx}\left(-(h(x)-H)k_1\frac{dh(x)}{dx}-k_2H\frac{dh(x)}{dx}\right)=R \tag{39}$$

When a zero-flow condition applies at $x=0$ and $h(x)$ takes a value H at $x=\frac{L}{2}$, the following expression is obtained for the groundwater elevation as a function of the distance relative to the drain:

$$h(x)=H-H\frac{k_2}{k_1}+\sqrt{H^2\left(\frac{k_2}{k_1}\right)^2+\frac{R}{k_1}\left(\left(\frac{L}{2}\right)^2-x^2\right)} \tag{40}$$

The travel time $t(x)$ in the saturated zone follows from:

$$t(x)=\varepsilon\int_x^{L/2}\frac{h(x)}{R\,x}dx \tag{41}$$

The solution reads:

$$t(x)=\frac{\varepsilon H}{R}\ln\left(\frac{L^*}{2x^*}\right)+\frac{\varepsilon H}{R}\frac{k_2}{k_1}\left\{1-\ln\left(\frac{L^*}{2x^*}\right)-\sqrt{1+\frac{R}{k_1}\left(\frac{L^{*2}}{4}-x^{*2}\right)}\right\}+ \\ \frac{\varepsilon H}{R}\frac{k_2}{k_1}\sqrt{1+\frac{R}{k_1}\frac{L^{*2}}{4}}\ln\left(\frac{L^*}{2x^*}\frac{\sqrt{1+\frac{R}{k_1}\frac{L^{*2}}{4}}+\sqrt{1+\frac{R}{k_1}\left(\frac{L^{*2}}{4}-x^{*2}\right)}}{\sqrt{1+\frac{R}{k_1}\frac{L^{*2}}{4}}+1}\right), \tag{42}$$

where $x^*=\frac{x}{H}\frac{k_1}{k_2}$ and $L^*=\frac{L}{H}\frac{k_1}{k_2}$. Some special situations are considered:

- When $k_1 \gg k_2$ the flow volume in the upper region can often be ignored. This case expresses the assumption made in many analyses: $h(x)-H\approx 0$. This assumption applies mainly to sandy aquifers with relatively high conductivities. If the assumption $h(x)\approx H_s$ is valid, the travel time $t(x)$ is calculated by:

$$t(x)=\frac{\varepsilon H}{R}\ln\left(\frac{\frac{L}{2}}{x}\right) \tag{43}$$

- In heavy clay soils most of the drainage systems are designed to allow a certain maximum height difference relative to the drains. Soil-ripening processes have resulted in higher conductivities above the drain level where the permanent saturated layers exhibit a relatively low permeability. As a consequence, most of the precipitation excess is conveyed through the upper flow domain to surface waters. For these types of soils the assumption of constant stream-zone thickness coinciding with the elevation line does not apply. This case applies often to the situation where drain tubes are located on top of an impermeable basis: $H_s=0$. An expression of the travel time is found from:

$$t(x)=\frac{\varepsilon}{\sqrt{R\,k_1}}\int_x^{\frac{L}{2}}\frac{\sqrt{(\frac{L^2}{4}-x^2)}}{x}dx=\frac{\varepsilon\frac{L}{2}}{\sqrt{R\,k_1}}\ln\left(\frac{\frac{L}{2}+\sqrt{\frac{L^2}{4}-x^2}}{x}\right)-\frac{\varepsilon\sqrt{\frac{L^2}{4}-x^2}}{\sqrt{R\,k_1}} \tag{44}$$

For age assessment of the discharge water the total discharge volume infiltrating between $x=0$ and $x=\frac{L}{2}$ is subdivided into N identical fractions. The fractions determine the time intervals of N age classes. The water arriving at the drain attributed to age class j had its starting point between $x=\frac{N-j}{N}\frac{L}{2}$ and $x=\frac{N-(j-1)}{N}\frac{L}{2}$. Let us define $\tau(j)$ as the upper time boundary of age class j. For the situation where $h(x)\approx H$ the class boundaries follow from:

$$\tau(j)=\frac{\varepsilon H}{R}\ln\left(\frac{N}{N-j}\right) \tag{45}$$

and for drain tubes located on top of an impermeable basis:

$$\tau(j) = \frac{\varepsilon \frac{L}{2}}{\sqrt{R\,k}} \left\{ \ln\left(\frac{N}{N-j} + \sqrt{\left(\frac{N}{N-j}\right)^2 - 1} \right) - \sqrt{1 - \left(\frac{N-j}{N}\right)^2} \right\} \qquad (46)$$

Table 2 presents five age classes related to identical discharge fractions for three discerned hydrological situations: an aquifer of infinite thickness ($H \to \infty$), an aquifer where the upper flow domain can be ignored ($h(x) \approx H$) and a flow domain completely above drainage level ($H = 0$).

Table 2. Age classes of arrived drainage water for three discerned hydrological situations

Age class	$H \to \infty$	$h(x) \approx H$	$H = 0$
1	$0 \le \tau < 0.03\frac{\varepsilon}{R}\frac{L}{2}$	$0 \le \tau < 0.23\frac{\varepsilon H}{R}$	$0 \le \tau < 0.09\frac{\varepsilon L}{2\sqrt{Rk}}$
2	$0.03\frac{\varepsilon}{R}\frac{L}{2} \le \tau < 0.15\frac{\varepsilon}{R}\frac{L}{2}$	$0.23\frac{\varepsilon H}{R} \le \tau < 0.51\frac{\varepsilon H}{R}$	$0.09\frac{\varepsilon L}{2\sqrt{R\,k}} \le \tau < 0.30\frac{\varepsilon L}{2\sqrt{R\,k}}$
3	$0.15\frac{\varepsilon}{R}\frac{L}{2} \le \tau < 0.41\frac{\varepsilon}{R}\frac{L}{2}$	$0.51\frac{\varepsilon H}{R} \le \tau < 0.92\frac{\varepsilon H}{R}$	$0.30\frac{\varepsilon L}{2\sqrt{R\,k}} \le \tau < 0.65\frac{\varepsilon L}{2\sqrt{R\,k}}$
4	$0.41\frac{\varepsilon}{R}\frac{L}{2} \le \tau < 1.23\frac{\varepsilon}{R}\frac{L}{2}$	$0.92\frac{\varepsilon H}{R} \le \tau < 1.61\frac{\varepsilon H}{R}$	$0.65\frac{\varepsilon L}{2\sqrt{R\,k}} \le \tau < 1.31\frac{\varepsilon L}{2\sqrt{R\,k}}$
5	$\ge 1.23\frac{\varepsilon}{R}\frac{L}{2}$	$\ge 1.61\frac{\varepsilon H}{R}$	$\ge 1.31\frac{\varepsilon L}{2\sqrt{R\,k}}$

Van den Eertwegh (2002) elaborated the travel-time distribution approach as presented by Eq. 45. Meinardi and Van den Eertwegh (1995a; 1995b) expanded the approach by accounting for the seasonal variation by introducing a factor linear with time. Van den Eertwegh and Meinardi (1999) used a conceptual model to explain the daily variation of travel-time distribution of arriving drainage water as could be deduced from solute concentrations. Van den Eertwegh (2002) used also a conceptual model to identify three discharge components: surface runoff, shallow groundwater discharge and deep groundwater discharge. Application of Eq. 45 with H = 2 m, R = 0.325 m a^{-1} and $\varepsilon = 0.35$ and assuming steady-state conditions, leads to annual travel-time fractions f_τ of the total drain water as shown in Table 3. Results of a further expansion of the approach by considering the flow domain above drainage level are also presented in Table 3. Both k_1 and k_2 were set 0.01 m d^{-1} and the drain spacing amounts to 16 m. The additional volume of the saturated zone above drain level results in a slight increase of travel times and consequently in diminished fractions f_1, f_2 and f_3.

Table 3. Example of travel-time distribution of tile drainage water

Fraction	Age class of travel time (−)	f_τ (−) (Van den Eertwegh 2002)	f_τ (−) (Eq. 42)
f_1	$t \leq 1$	0.37	0.32
f_2	$1 < t \leq 2$	0.23	0.19
f_3	$2 < t \leq 3$	0.15	0.13
f_4	$3 < t \leq 4$	0.09	0.09
f_5	$t > 4$	0.16	0.26

Analysis of field observations

Meinardi and Van den Eertwegh (1995a; 1995b) performed field experiments and analysed water and material budgets for a number of tile-drained fields in The Netherlands. Summarized results of the experimental sites in Flevoland and Hupsel-Assink will be presented in the subsequent sections.

Results of the Flevoland site

The experimental site Flevoland was located on clay soils cropped by arable crops. Field observations were carried out for one year at two sequentially monitored plots. The hydrological conditions were found to be different at each plot. Surface runoff was not observed but ponding did occur. The distance between the tile drains amounted to 48 m and the depth of the discharge layer beyond drainage level was estimated at 3 and 5 m for Plot1 and Plot 2, respectively. The porosity of the saturated zone was estimated at 0.30 as an effective average of both the soil crack volume of the subsoil and the matrix porosity (Van den Eertwegh 2002). The precipitation excess was estimated at 0.425 ($m\ a^{-1}$). Discharge of the precipitation surplus did not only consist of tile drainage but also direct flow to field ditches and to the main canals of the surface water system was observed.

The age-class approach assuming steady-state conditions as described in a preceding section was initially used to estimate the travel time of drainage water, which yielded reasonable results (Table 4). Improved results were obtained by applying a numerical model that employs a dual-porosity approach for water flow and solute transport, taking detailed soil properties, accurate boundary conditions and seasonal variation into account. Numerical transient modeling showed better results as compared to the steady-state approach, but the results of the steady-state approach

Table 4. Arrival fractions of drainage water calculated by the steady-state age-class approach for the Flevoland field sites

Fraction	Age class (a)	f_τ Plot 1	Plot 2
f_1	$t \leq 1$	0.25	0.38
f_2	$1 < t \leq 2$	0.19	0.23
f_3	$2 < t \leq 3$	0.14	0.15
Total	$t \leq 3$	0.58	0.76

were still reasonable. The volume fractions pertaining to the age classes indicate that any significant change in land management practices will yield 70-75% of its response to non-reactive solutes within three years. It was concluded from the numerical-modeling results that the steady-state age-class approach is less suitable to estimate the field-scale travel-time distribution under dry circumstances and that the results as presented in Table 4 could be improved by taking into account the soil conditions of each individual site concerning the location and the depth of sand, ripened clay and unripened clay layers.

Results of the Hupsel-Assink site

The Hupsel-Assink site was located in the eastern sand district of the Netherlands within the Hupsel brook basin. Results of these experiments have been reported by Van den Eertwegh and Meinardi (1999) and Van den Eertwegh (2002). The land use was permanent grassland and the soil at the experimental site was classified as loamy sand. The sandy phreatic aquifer is about 3 m deep and overlies a 30 m thick low-permeable Miocene clay layer. The distance between the tile drains amounted to 14 m and the depth amounted to 0.8 - 0.9 m below soil surface. The effective porosity of the permanently saturated zone was estimated at 0.35. Field measurements were carried out on one plot from November 1992 through March 1994, which appeared a relatively wet period compared to recorded long-term rainfall excesses. Surface runoff was not observed although ponding occurred occasionally. The precipitation surplus was discharged through tile drains as well as by direct drainage to a collection ditch. The field water balance yielded an unmeasured drainage term of 31% of total drainage. Both the steady-state *age-class approach* and a two-dimensional transient numerical simulation were performed to produce travel-time estimates. After attuning the boundary conditions for drainage and deep percolation, the numerical model was able to reproduce the water-balance terms. The travel-time distribution fractions of drainage, as calculated by the steady-state age-class approach, were confirmed by the transient two-dimensional model (Table 5). By combining the tile-drainage water balance and the field water balance a subdivison into discharge components was deduced. The reduced flow to tile drains can be accounted for by reducing both the recharge and the thickness of the saturated layer occupied by tile drainage flow. The travel distribution of drainage water indicates that any significant change in

Table 5. Arrival fractions of tile drainage water calculated for two time spans for the Hupsel-Assink site

Fraction	Age class (a)		f_τ 1985-1993	1993-1994	1993-1994
		depth	2.5 (m)	2.5 (m)	1.5 (m)
		recharge	0.330 ($m\,a^{-1}$)	0.550 ($m\,a^{-1}$)	0.385 ($m\,a^{-1}$)
f_1	$t \leq 1$		0.31	0.47	0.52
f_2	$1 < t \leq 2$		0.22	0.25	0.25
f_3	$2 < t \leq 3$		0.15	0.13	0.12
Total	$t \leq 3$		0.68	0.85	0.89

management practices will yield a major part of its response on non-reactive solutes within three years. It was concluded from numerical-simulation results that for the

Hupsel-Assink site, both the two dimensional model and the age-class approach can be used to describe the travel times of drainage water. Application of a transient 2D model yields more information on the dynamics of flow and transport processes (Van den Eertwegh 2002). Especially when plant transpiration and crop uptake of nutrient are to be considered, the results of the age-class approach will have a limited value.

Discussion and conclusions

General aspects

The importance of the unsaturated zone as an inextricable part of the hydrologic cycle has long been recognized. Theoretical and experimental studies on both water flow and solute transport in this zone have been further motivated by attempts to manage the root zone of agricultural soils optimally as well as concerns about soil and groundwater pollution. The interrelationships between soil, subsoil and surface waters make it unrealistic to treat the saturated and unsaturated zones and the discharge to surface waters separately. Agricultural practices over the last decades have resulted in higher emission of nutrients to surface water systems than would be allowed from an environmental point of view. The European Water Framework Directive sets targets concerning the chemical and ecological quality of groundwater and surface water. Insight into hydrological processes at field scale as well as at catchment scale and an associated understanding of the relation between drainage-water travel times and the loading of the surface water is of the utmost importance for an assessment of environmental impacts of agricultural management restrictions. An integrated approach that takes into account the subsoil, the upper groundwater zone and the surface water system is needed both for developing realistic water quality targets and for determining permissible losses at different spatial scales.

During the design phase of scientific research that focuses on surface water pollution as influenced by groundwater exfiltration a number of questions should be answered:

- What are the spatial and the temporal scales of input data and key numbers to be predicted?
- Could a simple one-dimensional approach suffice or should a two- or three-dimensional model be set up?
- Do soil heterogeneity and irregular surface water geometry play a role?
- Should the dynamic behavior of unsaturated/saturated soil moisture flow and drain discharge as results from climatically conditions be accounted for or could a steady-state approach be used?
- What types of eventual water management issues and load reduction measures have to be described?

Some simple approaches with a relatively low data requirement have been presented in the preceding sections. They contribute to the understanding of the relation between land use and surface water pollution, but when specific problems must be treated in detail more complex numerical models have to be applied.

Strengths

As was shown by Van den Eertwegh (2002) the interdependency of hydrological and hydrochemical information can be used to analyse the systems' behavior in more detail than for a situation of monodisciplinary-data availability. In many cases it appears impossible to measure all the terms of a field water balance independently. Information on solute migration in soil and breakthrough curves in drainage water of experimentally applied tracers can fill the gap in understanding the behavior of hydrological systems.

Different model approaches are available for the quantification of the pollutant load on surface water systems by groundwater discharge. An integrated-model approach is preferable to a segregated approach that addresses the unsaturated zone, the groundwater system and the discharge to surface water separately. The linear-systems approach and the associated age-class approach provide knowledge on the systems' behavior and can be applied to relatively simple problems. Complete physically based models that integrate three-dimensional water flow in the saturated and unsaturated zone, the two-dimensional surface flow and the flow within surface water systems have been successfully formulated and implemented. The validation of detailed process-oriented models at regional scale seems to present an insurmountable obstacle. For the purpose of prediction of environmental impacts of land and water management the application of a mixed type of models that comprise independent modules for certain knowledge domains still seems to be preferable.

This paper addresses the so-called *discharge-layer approach* as has been implemented in the SWAP model (Kroes and Van Dam 2003) and the age-class approach as was dealt with by Van den Eertwegh (2002). Both methods are easy to apply and require only a few input data. Although some shortcomings have been recognized, the methods can be applied when no detailed predictions are required. These types of models are useful to analyse the stochasticity of predicted drainage water concentrations.

Beside the approaches presented in this paper a large number of conceptual models can be found in literature. For reasons of linearity, both the lumped-parameter model (Van Ommen 1986) and the cascade of reservoirs are suitable to define easily solvable stochastic differential equations. These types of models can be extended by linear adsorption behavior and first-order transformation processes. A prerequisite for formulating a conceptual model is to define the most relevant key parameters as model variables. Upscaling and application on the regional scale of these models is relatively simple as compared to detailed mechanistic models.

Weaknesses

The limited validity of conceptual models should be kept in mind. The age-class approach as elaborated by Van den Eertwegh (2002) disregards the flow domain above drain level and assumes steady-state flow conditions. The thickness of the saturated flow domain is hard to assess for complex flow systems and the residence time in the unsaturated zone has been neglected. The discharge-layer approach (Kroes and Van Dam 2003) has been based on crude assumptions by presupposing perfect drains and ignores the flow volume occupied before a deeper unidirectional flow zone is reached. For diffusely scattered, low-adsorptive substances the relation used is considered valid. However, for distributed inputs at field level or high-adsorptive components in the presence of tile drains at a certain depth, the schematization may produce results of insufficient accuracy. Combined with the impact of solute transport via soil cracks and preferential flow, approximation of these problems in regional load models is considered to be an important challenge for further research.

Spatial scales of process-oriented regional models (Groenendijk and Roest 1996; Wolf et al. 2003) require more attention. For example, consistent data are required over a whole country for planning purposes and, on the other hand, farmers may require data on a very local basis in connection with a Nitrate-Vulnerable Zone. These types of scale issues have not been solved completely in the current models.

The coherence between quantity and quality key parameters emanated from the two- or three-dimensional approach is still not guaranteed. In a number of water quantity models the drainage resistance is used as a lumped parameter for describing the relation between groundwater head and discharge to surface water (Ernst 1978). The conceptualization of residence-time relations is not always tailored to the flow resistances. Also the recognition of the hierarchy and the classification of drainage systems as they can be identified on the basis of field data are biased by the model. Often the classification is chosen on the basis of water quantity-modeling purposes whereas a water quality model would have a different classification. This demonstrates the limited understanding of the systems' behavior of transient regional flow systems when both water quantity and quality have to be addressed.

The SWAP model (Kroes and Van Dam 2003) is used as a supporting model to generate hydrological information for regional prediction of nitrogen and phosphorus leaching (Groenendijk and Boers 1999; Groenendijk and Kroes 1999; Wolf et al. 2003). The choice of appropriate boundary conditions is still a hardly superable task. For example, the thickness of the soil profile for which the SWAP model simulates water flow, and associated with it the thickness of the discharge layers determine the response time of land management practices. The position of the bottom boundary should be chosen at the upper boundary of the regional flow system, but flow systems exhibit a dynamic behavior and field data for well-based model schematization are still missing.

Opportunities

It can be expected that the major concern with respect to the ecological status and ecological potentials of surface water bodies will lead to an intensification and optimization of the monitoring efforts in terms of both quantity and quality. More detailed information on land management, soil, groundwater and surface waters will be collected, which can be used for model validation purposes. On the other hand, the monitoring issues will stimulate further research on the critical links within these fields.

Many models of varying degrees of complexity and dimensionality have been developed during the past decades to quantify the basic physical and chemical processes affecting pollutant transport in the unsaturated zone. Models for variably unsaturated–saturated flow, solute transport, aqueous chemistry and cation exchange were initially developed independently of each other, and only recently there has been significant effort to couple the different models. Comprehensive models addressing the unsaturated–saturated zone, overland flow and surface water dynamics are available nowadays but require detailed input data and high skills to apply. Mixed approaches combining detailed and conceptual modules for different sub-domains of the hydrological cycle will be the most successful for practical applications. Such an approach implicitly comprises scale contrasts at the sub-domains' interfaces. The comprehensive models mentioned can be used to validate solutions attempting to manage scale discrepancies.

The discharge-layer approach presented in this paper as a method to conceptualize groundwater flow in a hierarchical flow system should be extended with the *averaged vertical-flux concept*, based on streamline patterns resulting from two- and three-dimensional groundwater models. The age-class approach coincides with the linear-systems approach. The age-class approach can be elaborated by accounting for the influence of radial flow in the vicinity of line drains and the flow domain above drainage level. Linearized conceptual models have proven to be useful for stochastic analysis.

The work of Van den Eertwegh (2002) demonstrates the importance of collecting accurate field data to compare and validate model approaches. Different types of approaches were compared: analysis of field data, application of a simple conceptual age-class method and setting up a complex numerical dual-porosity model. The combination of water-balance data as well as a solute balance at different spatial scales proved to be an indispensable source of information for validation. Combined analysis of water, tracers and reactive nutrients contributes to the understanding of groundwater-flow-driven contamination of surface water bodies.

Threats

The most detailed models that incorporate the latest technologies and concepts are generally the most technically defensible and have, in theory, the widest range of applicability. Practically, however, these may not be the most useful models for application in farm management or regional land-use decisions, given the required knowledge of the system and the data requirements.

To account for the influence of spatial variability and the atypical behavior of structured soils on water flow and solute transport a wide range of methods is required. The method to be chosen must be attuned to scale of data and scale of problem to be addressed. During the last decade, models based on the Richard equation were often a point of reference. Some extensions were made to treat macropore flow (Groen 1997; Hendriks, Oostindie and Hamminga 1999), preferential flow and soil variability (Kroes and Van Dam 2003), but the question concerning the validity and the usefulness of the Richard equation for these soils got less attention.

Understanding the relation between land management and the nutrient load on surface water is important to assess the ecological status and the ecological potentials of surface waters. However, transformation and accumulation processes within the upper few centimeters of the sediment often reduce the influx to many European surface water systems by more than 50%. For decision making and evaluation of environmental impacts the main focus of attention will be on these retention zones. Movement of solutes in soils is reasonably well understood. The added value of further refinement of describing solute transport in the unsaturated zone will be limited as the uncertainties in other knowledge fields determine the accuracy of final surface water-quality estimates.

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