Environ Fluid Mech (2010) 10:121–135
DOI 10.1007/s10652-009-9148-1

ORIGINAL ARTICLE

Estimating travel time of recharge water through a deep vadose zone using a transfer function model

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Received: 30 March 2009 / Accepted: 13 August 2009 / Published online: 28 August 2009

Abstract We estimate the travel time of percolating water through a deep vadose zone at the regional scale using a transfer function model and a physical based conceptual flow model (Hydrus-1D), thereby exploiting the time series of precipitation, actual evapotranspiration and groundwater piezometry and generic vadose zone data. With the transfer function model we observe a high variability of estimated travel time varying from 0.9 to 3.1 years, corresponding to estimated vertical water flux velocities varying from 6.6 to 28.0 m/year. These results were compared with the travel time estimated from the physical based conceptual model. With the flow model, estimated travel time varies between 4.7 and 15.5 years, corresponding to water flux velocities varying between 1.7 and 4.1 m/year. The estimated travel time calculated with the flow model were therefore about five times larger than those estimated with the transfer function model. This could be explained by the fact that the transfer function model considers heterogeneous recharge from the vadose zone as well as from the vicinity of the piezometer through the so called "pushing effect". In addition, the flow model requires various hydrogeological and hydrodynamic parameters which were estimated using generic parametrisation approaches, that are largely affected by uncertainty and may not reflect the local conditions. In contrast, the transfer function model only exploits available measurable time series and has the advantage of being site-specific.

Keywords Transfer function model · Travel time · Soil · Groundwater level fluctuations · Brusselian sands

1 Introduction

The vadose zone is a key component within terrestrial systems and plays an essential role in the hydrological cycle. The vadose zone encompasses the unsaturated soil root zone and controls thereby the fluxes of water, matter and energy between the atmosphere, land surface

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and subsurface water bodies. Knowing the travel time of water in the vadose zone is therefore a prerequisite for quantifying the recharge and load of groundwater bodies in terms of variable boundary conditions at the land surface interface [12].

Characterizing the travel time of water and associated matter in the vadose zone is complicated and this because of many reasons [2]. First, a wide set of physical, chemical and biological processes affect flow in the vadose zone. Second, the size and geometry of the vadose zone in natural conditions is often not well known. Third, the physical, chemical and biological properties of the vadose zone are variable in space and in time which makes the experimental assessment tedious. Fourth, as compared to technologies and tools available for characterizing surface hydrological and hydrogeological properties, much less technologies have been designed to characterize the partially saturated vadose zone. Fifth, water flow in the vadose zone is a highly non-linear process and depends very much on the saturation degree of water which on its turn is controlled by the variable water input at the land surface. Given this complexity, assessing travel time of water through the vadose zone has received far less attention as compared to travel time assessments in surface and groundwater bodies. The vadose zone is therefore still often a large gray box in our current hydrological knowledge.

Travel times of water through the vadose zone depend very much on the thickness of the vadose zone. In flat alluvial regions, where only a thin vadose zone is present, travel time may be small such that pressures exerted at the land surface will propagate fast to the groundwater body. In undulating landscapes, however, where the size of the vadose zone will generally be much larger, higher travel time is expected. In arid environments, for instance, the vadose zone can be hundreds of meters thick, and infiltration fluxes very low [10], resulting in the residence time of water ranging from several hundreds to thousands of years [5]. However, also in such a deep vadose zone, flow can be extremely fast. For instance Levitt et al. [29] showed that a surface applied tracer could reach a 78 m deep groundwater tables in 8–14 days, proving the small travel time in such large formations.

Different methods exist to assess vadose zone travel times. Analysing at a given depth in the vadose zone the breakthrough of a land surface applied tracer allows determining the travel time from the moments of the time dependent breakthrough curve. Use have been made of single ionic inert tracers such as chloride [1,9,20,34,35], stable and radiogenic isotopes [3,15], nitrate [37], or a combination of different ionic tracers [41]. Yet, given the complexity to carry out large scale tracer breakthrough experiments and the long time frames involved, an alternative approach relies on the inversion of conceptual flow and transport models. With such approaches use is made of some measured vadose zone property which is modelled using a transport model in which the travel time is considered, directly or indirectly, as the fitting model parameter. With the inverse approach, measured moisture content or resident tracer profiles have been used as fitting data by Wang et al. [49], McElroy and Hubbell [32], Hubbell et al. [17], Flint and Ellett [13] and Wu et al. [50]. Robinson et al. [39] combines measurements of moisture content with the analysis of a tracer experiment. Constantz et al. [7], proposes the logging of temperature profiles in combination with a coupled flow and heat transport model.

When using the inverse approach, physically based conceptual flow and transport models can be used that are based on the physical laws of mass, momentum and energy conservation and flow. The use of such flow and transport models is often prohibitive given the lack of data that are available for parametrising such models. Where nothing but only few time series of vadose zone properties are available, time series models are often preferred to the conceptual flow and transport models [27]. One of these, the transfer function model, was proposed by Jury [25] to characterize field scale solute transport through an unsaturated soil. Using a transfer function model, it is considered that the internal transport mechanisms are unknown. Jury [25] derived a simple theory for estimating the average and extreme behaviour of solutes moving through the vadose zone as a function of a simple field-measurable travel time probability density function (pdf). Inverting the transfer function for the parameters of the travel time pdf allows estimating the travel time through the vadose zone. This approach was applied for example by Vanderborght et al. [46] and Javaux and Vanclooster [20] to determine the travel time through large undisturbed soil columns from resident concentration data; by Gasser et al. [14] to predict nitrate leaching under potato crops of an agricultural field; by Stewart and Loague [43] to predict pesticide load of ground water at the regional scale; by van der Velde et al. [47] to estimate the travel time through a volcanic soil on a small atoll in the Pacific Ocean.

Notwithstanding the availability of these methods, uncertainties in the estimation of the travel time through the vadose zone remain very high [11,16]. For instance, Cook et al. [8] compared ionic tracer studies with radiogenic isotope tracer studies and concluded on the inconsistency between the estimates of the travel time with both methods. Given these uncertainties and inconsistencies, quite some scope exist to improve our knowledge on the travel time of water through the vadose zone. Upon our knowledge, while the Jury transfer function has been used to simulate solute transfer through the vadose zone, no study has been presented so far which exploits the easily measurable temporal dynamics of the vadose zone lower boundary condition, i.e. the groundwater position, as fitting data in the inversion process.

The objective of this study is to evaluate the travel time of percolating water through a deep unsaturated zone using transfer function theory, thereby exploiting the measured time series of precipitation, actual evapotranspiration and groundwater piezometry. The underlying hypothesis is that the water flux wave at the soil surface will propagate through the vadose zone and, when it hits the ground water body, result in a ground water depth perturbation. Modeling this traveling wave by means of the transfer function model allows inferring the large scale travel time through the vadose zone. The results obtained by means of the transfer function model will further be compared with the travel time estimated by means of a physical based conceptual model (Hydrus-1D).

2 Material and methods

2.1 Site and data description

This study focuses on the vadose zone of the Brusselian aquifer situated in the center of Belgium. The aquifer has a surface area of 965 km^2 and encompasses an unconfined ground water body which is of primary importance for drinking water supply in the region. The aquifer is composed of Tertiary sedimentary sands and is overlain by a Quaternary loess layer of variable thickness (0–15 m). The Brusselian sands outcrop mainly in the valleys where sandy and sandy loam soils develop. Transmissivity of the aquifer varies from 2.9×10^{-5} to 1.2×10^{-2} m^2/s and its permeability varies from 1.4×10^{-6} to 6×10^{-3} m/s [19].

Daily precipitation measured at Uccle (Belgium) from 1961 to 2007 was provided by the European Climate Assessment & Dataset [44], and was aggregated to monthly totals. A time series of monthly actual evapotranspiration was calculated by Oger [36], using daily weather data collected at Gembloux (situated in the study region) from 1950 to 1989. In order to extrapolate these data to the period for which precipitation series were available (1961–2007), we calculated the mean actual evapotranspiration for each month and completed the time series of monthly evapotranspiration rates with the calculated mean per month. Completing the

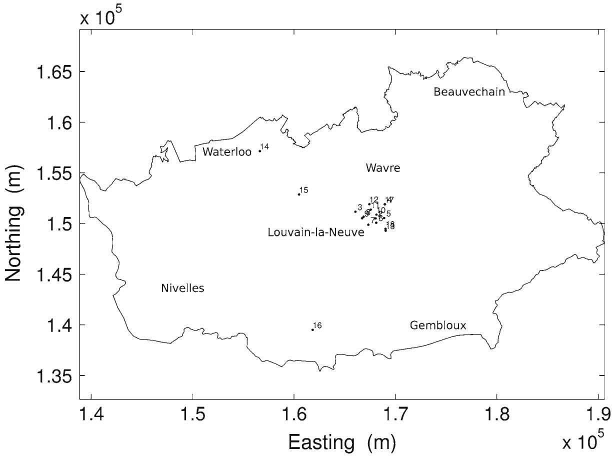

Fig. 1 Location of the piezometers in the Brusselian sands ground water body

missing values as such is justified by the high reproducibility of the actual evapotranspiration cycle from 1 year to another. The time series of the piezometric height were provided by the regional administration [33] and from the Hydrological Service of the Université catholique de Louvain. In the study area, ground water levels are measured on 241 piezometers. However, the vast majority of these piezometers cannot be used in the context of this study since either the length of the time series is insufficient, or the frequency with which the measurements are performed does not allow to capture the dynamics of the system. From the 241 available piezometers, only a subset of 18 piezometers were selected for which water levels are measured for a long period (1967–2007) and at a high measurement frequency (Fig. 1).

2.2 The transfer function model

We use a transfer function model to describe water transport through the vadose zone. The underlying hypothesis is that a water flux wave at the soil surface will propagate through the heterogeneous vadose zone and perturbate the water level when it hits the groundwater body. The vadose zone in this case study includes the biologically active soil root zone, a loamy substratum, and the unsaturated sandy substratum. The water flux wave at the soil surface is determined by infiltration of rain water in the soil, from which evapotranspiration rates are subtracted. The model further assumes that no lateral run-off occurs and that no dispersion of percolating water takes place other than that which is implicitly represented by the travel time variations of the system [25]. Water transport is assumed to be convective with a stochastic component expressing the variability of water travel time, due to heterogeneous pathways in the soil and subsoil [14]. We consider only the vertical flux since the vertical travel time in the vadose zone is generally 1–3 orders of magnitude smaller than the horizontal one [26].

The transfer function model uses a simple travel time probability density function (pdf) to relate an output time series to an input time series. The approach of Jury [25] derives a distribution function based on the distribution of the travel time of solutes from the soil surface to a reference depth. Since the vadose zone between the surface and the aquifer smooth out the variations of water flux, a transfer function model can be used to describe the relation between the variation in rainfall and evapotranspiration and the variations of the level of the groundwater table. We use the transfer function model to describe the convolution of the variation in the rainfall minus the actual evapotranspiration with the transfer function to predict variation in the level of the ground water table:

$$P(t) - \mu_P = \alpha \int_0^t f(t)(I(t-\tau) - \mu_I)dt \tag{1}$$

where $I(t) = R(t) - E(t)$ is the input time series (percolating water, mm/month); $R(t)$ is the rainfall (mm/month); $E(t)$ is the actual evapotranspiration (mm/month); μ_I is the mean of I (mm/month); $I(t) - \mu_I$ is the variation of input around μ_I (mm/month); $f(t)$ is the transfer function equal to the travel time probability density function (pdf) (months^{-1}); t is time (months); τ is the time lag (months); α is the scaling factor (−); $P(t)$ is the piezometric level (mm/month); μ_P is the mean of P (mm/month); $P(t) - \mu_P$ is the variation of P around μ_P (mm/month).

The integral in Eq. 1 expresses the convolution between the variations around the mean percolating water ($I(t) - \mu_I$) with the travel time probability density function to predict the variations around the mean of the piezometric level ($P(t) - \mu_P$). Here t is time and τ the lag time between inputs and piezometric variations, α is a scaling factor, $f(t; \mu, \sigma)$ is the travel time probability density function. A log-normal probability density function was chosen for the transfer function $f(t; \mu, \sigma)$ to reflect the transport and mixing of the percolating water as it moves through a porous medium exhibiting a log-normal distribution of pore-water velocities [38]. Hence:

$$f(t; \mu, \sigma) = \frac{1}{\sigma t \sqrt{2\pi}} \exp\left(-\frac{(\ln(t) - \mu)^2}{2\sigma^2}\right). \tag{2}$$

The model $f(t; \mu, \sigma)$ corresponds to a log-normal probability density function where μ and σ respectively correspond to the mean and standard deviation of the logarithm of the travel time through the vadose zone.

The transfer function was used to determine the modal transfer time, defined as the time where the probability density function reaches its maximum value, and the 5th and 95th percentiles, defined as the time needed by 5 and 95% of the percolating water to reach groundwater table. The parameters α, μ, and σ were obtained by minimizing the sum of squared errors for all observations within each piezometer [47]:

$$\Phi(\alpha, \mu, \sigma) = \frac{1}{n} \sum_{i=1}^{n} \left[(P - \mu_P)_{i,\text{simulated}} - (P - \mu_P)_{i,\text{observed}}\right]^2 \tag{3}$$

where $\Phi(\alpha, \mu, \sigma)$ is the objective function to be minimized; $(P - \mu_P)_{i,\text{simulated}}$ is the simulated variation of P around μ_P (mm); $(P - \mu_P)_{i,\text{observed}}$ is the observed variation of P around μ_P (mm); n is the number of observations in each piezometer.

To avoid ill-posedness, an a-priori parameter domain was divided into a regular grid, covering a range from 0 to 5.5 for μ (Eq. 2) (corresponding to mean travel time from 1 month to 20 years), from 0 to 1 for σ (Eq. 2) and from 0 to 150 for the scaling factor α (Eq. 1). The objective function (Eq. 3) was calculated for each combination of the grid selected parameters. Afterward, the parameters minimizing the objective function were used as an input of the Nelder–Mead multidimensional unconstrained nonlinear minimization algorithm available in Matlab™ for a more accurate estimation of the optimal parameters.

We identified first the travel time pdf for each piezometer separately. Subsequently we identified an effective travel time by inverting all data simultaneously and compared these to the mean travel time pdf obtained from the local estimates. In case the traveling wave process through the vadose zone is a linear process, than the local and effective travel time should converge.

2.3 The physical based conceptual flow model

We compared the estimated travel time by means of transfer function theory with predictions made by means of a physically based flow and transport model. We used the Hydrus-1D model [42] which solves numerically the 1-D governing flow model (Richards equation) for the vadose zone:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left[K(h)\left(\frac{\partial h}{\partial z} + 1\right)\right] - S \tag{4}$$

where z is the vertical coordinate (positive upwards), t is time (day), S is a sink term (day^{-1}), h is the pressure head (m), θ is the volumetric moisture content and K is the unsaturated hydraulic conductivity function (m/day); and the convection dispersion solute transport model:

$$\frac{\partial C}{\partial t} = D\frac{\partial^2 C}{\partial z^2} - \nu\frac{\partial C}{\partial z} \tag{5}$$

where C is the solute concentration (m^3 m^{-3}), D is the dispersion coefficient (m^2 day^{-1}) and ν is the water flux (m day^{-1}). Hydrus-1D solves Eqs. 4 and 5 for vertical heterogeneous porous media and transient boundary conditions by means of the Galerkin finite element method. The soil hydraulic functions used in Hydrus-1D are modelled according to the van Genuchten–Mualem model [48].

A variable flux condition was imposed on the soil surface, considering the same inputs $I(t)$ than those used in the transfer function model (Eq. 1). At the bottom of the flow domain a constant pressure head equal to zero was imposed, representing the position of the water table. The initial conditions were defined through a warming up procedure. In a first loop, the model was run from 1961 to 2007. The obtained moisture and pressure head profile at the end of this warming up loop was subsequently used as initial moisture and pressure head profile for the tracer simulation loop. In this second loop, a tracer pulse was applied on the top of the soil column during one time unit and followed through the profile allowing to infer convective solute travel time.

In contrast to the transfer function approach, we need information about the lithology of the sub-soil (i.e. a sandy layer overlain by a sandy loam layer) and the physical properties of each sub-soil layer to parametrise the formation profile in Hydrus-1D. The thickness of the different layers could be inferred from lithological logs that were collected when installing the piezometers. However, these logs were only available for 12 out of the 18 piezometers. No data were available to parametrise the soil physical properties, and therefore use was made of three generic approaches to estimate these parameters (Table 1). The first set comes from the Hydrus catalog [6], the second one from Rosetta [40] and the third one from literature of measured hydraulic properties of media in the study area [18,22,23]. Three typical different parameter sets describing the soil properties were used since we didn't know the real soil properties at the location of the piezometers.

Except for the soil parameters, the generic parameters were used (e.g. single porosity, no hysteresis), because of the absence of data needed for a better parametrisation of the physically-base model.

Table 1 Soil properties used in the Hydrus-1D simulations

Source	Soil layer	Depth (m)	θ_r (−)	θ_s (−)	α (m^{-1})	n (−)	λ (−)	K_{sat} (m/year)	Dispersivity (m^{-1})
Hydrus catalog	Sandy loam	0–4	0.065	0.41	7.5	1.89	0.5	387.265	0.1
	Sand	4–14.3	0.045	0.43	14.5	2.68	0.5	2601.72	1
Rosetta	Sandy loam	0–4	0.0387	0.387	2.67	1.4484	0.5	139.612	0.1
	Sand	4–14.3	0.053	0.3747	3.53	3.1798	0.5	2346.88	1
Literature	Sandy loam	0–4	0	0.3894	0.4084	1.204	0.5	108.6	0.1
	Sand	4–14.3	0.0818	0.33	2.87	5.64	−0.555	189.8	1

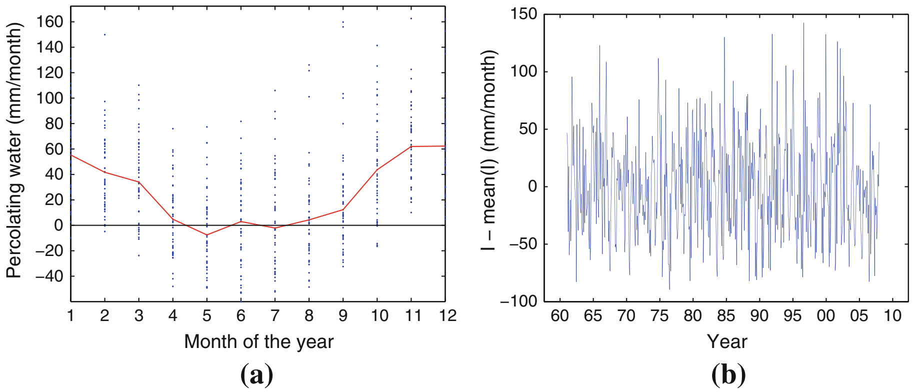

Fig. 2 **a** Monthly total (*dots*) and monthly median of the monthly total (*line*) percolating water flux at the soil–atmosphere interface (mm/month), calculated from 1965 to 2007 and **b** variation of the monthly total percolation water flux around its mean value (mm/month) used as input time series of the transfer function model

3 Results

3.1 Input of the transfer function model

The monthly median of the monthly total percolating water flux at the soil–atmosphere interface, $I(t)$, calculated from 1965 to 2007, varies from nearly null from April to August, to 62 mm in December (Fig. 2a). The variability of the monthly total percolating water flux for each month is large. The input time series of the transfer function model was obtained by subtracting the average of the percolating water flux calculated for the complete time series from the monthly total percolating water flux time series $I(t)$ (Fig. 2b).

3.2 Travel time estimated from the transfer function model

Figure 3 presents, for two piezometres randomly sampled in the study area, the results of simulated versus measured variation of piezometry around the mean (Fig. 3a, b) together with its corresponding optimal log-normal transfer functions (Fig. 3c, d). We observe that the transfer function model allows to describe the phase of the dynamic signal propagating through the deep vadose zone, but has more difficulty to predict correctly its amplitude.

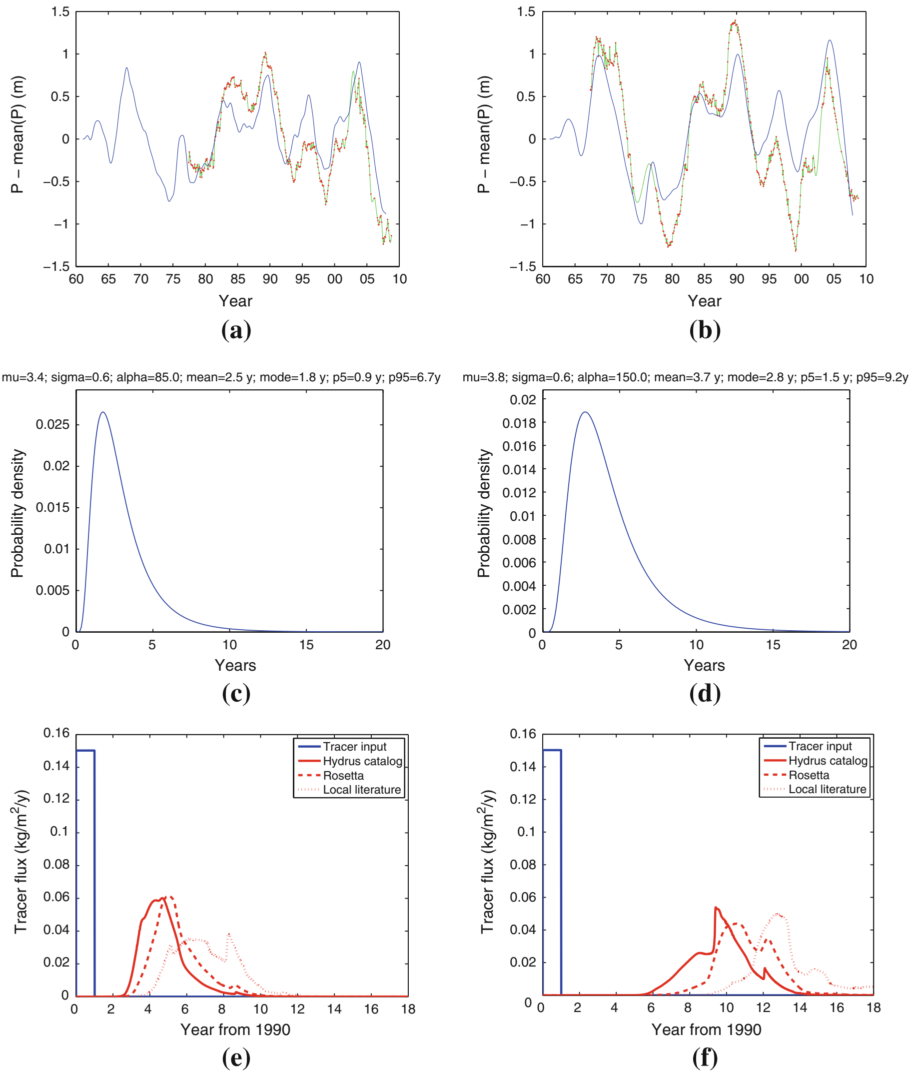

Fig. 3 Measured (*red dots*) and simulated with the transfer function model (*line*) piezometric time series (**a, b**), their optimal log-normal transfer functions (**c, d**) and the breakthrough curves calculate with Hydrus-1D (**e, f**) for two different piezometers

We also observe a systematic overestimation of the simulation between 1993 and 1998 which can be explained by the exploitation regime of ground water in this area, and which is not considered in the transfer function model. However, only the phase is important for estimating the travel time.

The modal value of the travel time of recharge water through the unsaturated zone, calculated from the transfer function model for 18 individual piezometers varies from 0.9 to 3.1 years (Table 2). The average of the modal transfer time of these 18 piezometers was 2.3 years, with a standard deviation of 0.6 years. Furthermore, the fifth percentile of the travel time for the 18 piezometers varies from 0.5 to 1.8 years and the 95th percentile from 5.2 to

Table 2 Depth to the water table, travel time and water flux velocities calculated with the transfer function model and Hydrus-1D (with the three sets of soil properties) for the 18 piezometers

Piezometer	Depth	Transfer function		Hydrus (Hydrus set)		Hydrus (Rosetta set)		Hydrus (Local literature set)	
	(m)	Travel time (year)	Velocity (m/year)	Travel time (year)	Velocity (m/year)	Travel time (year)	Velocity (m/year)	Travel time (year)	Velocity (m/year)
1	37.7	2.92	12.9	12.4	3.04	13.4	2.82	15.5	2.43
2	36.7	2.25	16.3	10.4	3.53	12.5	2.94	13.3	2.76
3	41.4	2.92	14.2	10.5	3.94	10.0	4.14	12.0	3.45
4	14.3	1.75	8.2	4.7	3.04	5.0	2.86	8.3	1.72
5	33.8	2.25	15.0	10.4	3.25	12.6	2.69	13.4	2.53
6	26.9	2.17	12.4	9.2	2.92	9.1	2.96	9.3	2.89
7	39.5	1.92	20.6	10.1	3.91	9.9	3.99	11.7	3.37
8	16.9	2.58	6.6	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
9	17.7	2.25	7.8	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
10	36.5	2.33	15.7	10.2	3.58	10.7	3.41	12.6	2.9
11	33.7	2.92	11.6	9.5	3.55	9.6	3.51	11.6	2.91
12	49.1	3.08	15.9	12.6	3.9	12.9	3.81	13.5	3.64
13	29.7	2.75	10.8	9.4	3.16	10.6	2.8	12.8	2.32
14	25.6	0.92	28.0	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
15	39.3	2.25	17.4	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
16	22.9	1.33	17.2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
17	14.5	2.17	6.7	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
18	29.8	2.50	11.9	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.

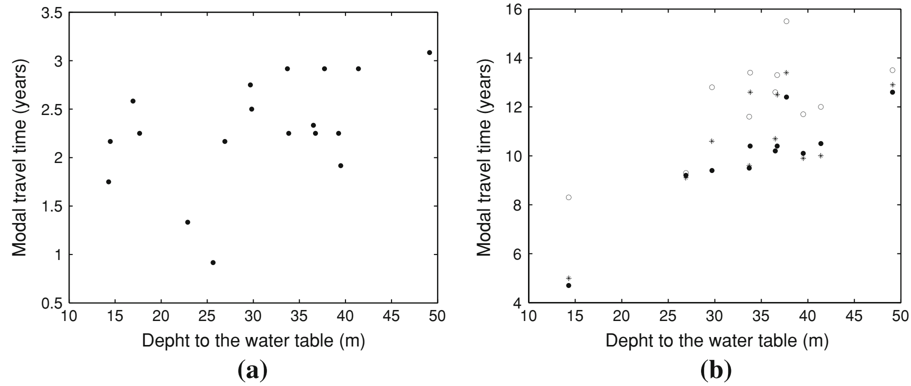

Fig. 4 Relation between the depth to the water table and the modal travel time calculated with **a** the transfer function model and **b** Hydrus-1D (*black circles*, *stars* and *white circles* correspond to travel time calculated with the soil parameters of the Hydrus catalog, Rosetta and local literature, respectively)

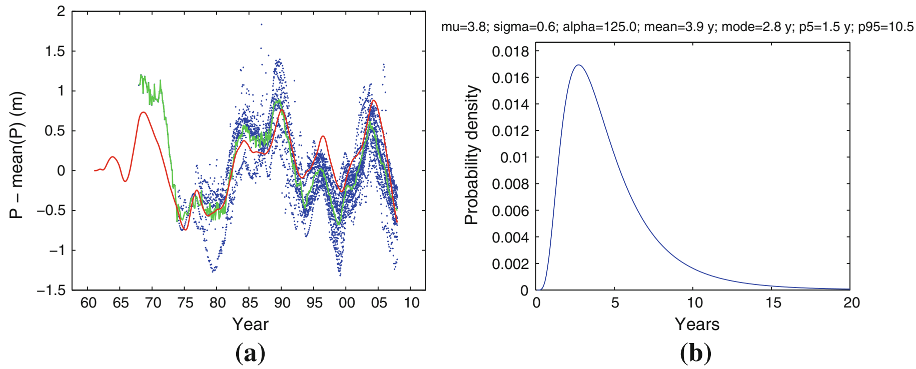

Fig. 5 Measured data of all the piezometers (*blue dots*), their mean (*green line*) and the simulated piezometry (*red line*) (**a**) and corresponding optimal log-normal transfer function (**b**)

15.1 years, which shows that, for a same location, the water wave propagates very quickly through the unsaturated zone while at other places wave propagations is much slower. The variability of the travel time between the piezometers can be explained by the differences in the thickness of the unsaturated zone, variability of hydrogeological properties, and flow phenomena that are not considered within the transfer function model. For instance, we observe that travel time increases with depth to the water table (Fig. 4a). By dividing the depth to the water table (Table 2) by the modal travel time (Table 2) for each piezometer, we can estimate that the water travel velocity through the vadose zone varies from 6.6 to 28.0 m/year (Table 2).

We also attempted to derive a regional transfer function model by fitting one log-normal transfer function to the mean of all the data of the 18 piezometers (Fig. 5a). The resulting optimal log-normal pdf is characterized by a modal travel time of 2.8 years, which is close to the mean of the modal travel time of the 18 individual piezometers (2.3 years) (Fig. 6). The 5th percentile of the travel time estimated on the mean of all piezometers was 1.5 year and the 95th was 10.5 years (Fig. 5b).

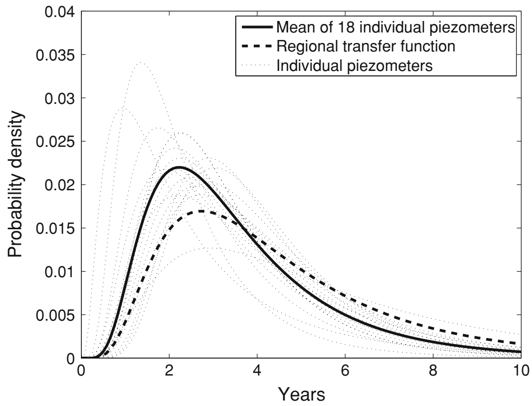

Fig. 6 Transfer function models of the 18 individual piezometers (*dotted lines*), their mean (*line*) and regional transfer function model (*broken line*)

3.3 Travel time estimated from the conceptual flow and transport model

Travel time was also calculated from the breakthrough of a conservative tracer, simulated for 12 piezometers using the Hydrus-1D model. All 18 piezometers of the previous section could not be used since information about the thickness of the sandy and sandy loam layer was not available for 6 out of them. The breakthrough curve of the inert tracer at the bottom of the unsaturated zone column had a log-normal-like shape with a modal transfer time going from 4.7, 5.0 and 8.3 years to 12.6, 13.4 and 15.5 years for the first, second and third set of soil properties, respectively (Table 2). Figure 3e, f presents the simulated breakthrough curves for two selected piezometers. Obviously, we observe also that travel time increases with depth to the water table (Fig. 4b).

4 Discussion

The values of the travel time estimated with the physical based conceptual model are about five times larger than those estimated with the transfer function model. Differences can be explained by conceptual differences between both modeling approaches and the uncertainty associated with the parametrisation of the physically based conceptual flow and transport model.

Indeed, the travel time estimated from the simulated breakthrough curve of the inert tracer by means of the physical based conceptual model, considers 1D convective mass transport of an inert tracer with the moving water body through an unsaturated column on top of the piezometer. In this case, it has been hypothesized that the full water body within the vadose zone contributes to the conservative mass transport, and that the estimated travel time of the recharging water corresponds to the simulated travel time of the conservative tracer. In reality however, the measured piezometric position will not only be influenced by the local vertical recharge through the 1D vadose zone column on top of the piezometer, but also by the lateral recharge through the vadose zone and the groundwater body itself.

The transfer function model considers the geologic formation between the observation point (i.e. the local piezometer) and the surface input as a black box with no well defined geometry. With this modeling approach the flow domain does not necessarily pertain to a 1D unsaturated soil column on top of the observation point, but may be equivalent to a geometrically irregular flow domain, encompassing stream tubes through the vadose zone and the groundwater body itself, where water is transported to the observation point in different directions. Hence, the transfer function model may conceptually encompass the recharge of the water coming from the vadose zone on top of the observation point as well as water that laterally feeds the observation point. The transfer function model will also consider the water that feeds the observation point through the so-called "pushing effect". Indeed, when pressure waves propagate through the vadose zone, resident water in the vicinity of the piezometer may be mobilized and contribute to piezometric variation. Pressure waves generally propagates through natural porous media at velocities which are much faster than mass transport waves [4], and piezometric variation does not necessarily reflect mass transport through the vadose zone. In case the lateral recharge is more important than the vertical recharge through the vadose zone, or in case the "pushing effect" is significant, then travel time inferred from the transfer function modeling approach may easily be smaller than estimated from a physically based 1D conceptual mass transport model through the vadose zone.

Further, vertical flow rate in the vadose zone may be faster than estimated from the conceptual mass transport model. Many experimental evidence show that only part of the available water in the vadose zone contributes to effective mass transport. Using small scale sandy columns collected from the same region as in the present case study, Vanclooster et al. [45] showed that mass transport could much better be described with a two porosity flow model rather than a single porosity flow model, considering mobile water fractions ranging between 0.2 and 0.8 of total resident volumetric soil moisture. In case the mobile water fraction would indeed be 0.2, then effective velocities would be five times larger which is consistent with the estimates from the transfer function model in the present study. Based on detailed solute transport experiments at variable flow rates on an undisturbed 0.5 m^3 size sandy soil monolith, collected from the sandy substratum of the Brusselian aquifer, Javaux and Vanclooster [21] showed that effective pore water velocity contributing to the mass transport of the inert tracer ranges between 0.8 and 1.6 times the theoretical pore water velocity, suggesting mobile water fractions between 0.6 and 1.2 times the total volumetric moisture content. Also in the majority of the experiments of this last study, larger effective velocities than pore velocities estimated from a single region mass transfer model were obtained.

The unavailability of pore water to contribute to effective mass transport will further be exacerbated when heterogeneous flow develops in the flow domain feeding the piezometer. In-situ heterogeneous flow may first be induced by the local variability of the vadose zone transport parameters. The local scale variability of the vadose zone flow properties is indeed very large. Mallants et al. [30], for instance, showed that the local scale saturated hydraulic conductivity of the loamy soil specific for the present study area could vary with six orders of magnitude, while saturated porosity could vary between 0.38 and 0.51. Together with the local scale variability of the unsaturated hydraulic properties, these induce variability of fluxes of a factor 12. Similar results were obtained when analyzing the local scale variability of the sandy substratum of the present case study area [23]. Heterogeneous flow may further be promoted at the formation scale by the presence of distinguished structural features, such as clay lenses and discontinuous clay layers. Indeed, the sandy substratum of the study area is intersected with different discontinuous clay layers which reflect the genesis of these natural formations. Flow across different textured layers will be unstable and induce heterogeneous flow. Within the unsaturated sandy soil core analyzed by Javaux et al. [24], a distinguished structural feature was present which induced clearly heterogeneous flow at the macroscopic core scale. Similarly, when analyzing inert tracer transport through a 6 m deep observation well below an artificial lake, different clay layers were observed. The measured local flow velocity across the different layers varied between 2.5 and 13.1 m/year which is in the same order of magnitude of flow velocities that are calculated in the present study.

The difference in estimated travel time with both methods could also partially be explained by the uncertainty on the hydrogeological and hydrodynamic data feeding the 1D conceptual flow model. The 1D flow model requires a good knowledge of the lithology of the vadose zone and the unsaturated flow hydraulic properties. In case the hydraulic properties are not know, they can be estimated based on the knowledge of the granulometry and bulk density using so-called pedo-transfer functions. For the considered piezometers in this case study, the lithology of the vadose zone could rather correctly be reconstructed using the lithological logs that were collected when installing the piezometers. However, the hydraulic properties are far more uncertain. To consider the uncertainty in vadose zone hydraulic properties, three parametrisation strategies were considered in the present study. The different parametrisation schemes yield estimated travel time variations ranging between 4.7 and 15.5 years, corresponding to velocity variations from 1.7 to 4.1 m/year. Although largely significant, this suggests that uncertainty in hydraulic parametrisation only marginally explains the difference in estimated travel between both methods.

As a final remark, the important travel time through the vadose zone may have considerable consequences for water management in the study area. Indeed, the unconfined groundwater body of the Brusselian aquifer is exploited for drinking water purposes and is subjected to many pressures. Mattern et al. [31] for instance describe the current pollution of the groundwater body by nitrates, while Leterme et al. [28] discuss pollution by pesticides. Within the context of the implementation of the European Water Framework Directive and the European Nitrate Directive, the regional administration currently implements a set of land use management measures, envisaging to reduce emissions of pollutants from the land surface and the root active zone to the groundwater body. The important travel time through the vadose zone suggests that the impact of these measures on ground water quality can not be expected at the short term.

5 Conclusions

In this study, we used a transfer function model to estimate the travel time of percolating water through a deep vadose zone (Brusselian sands, Belgium) and we showed a high variability across the study area, with modal travel time varying from 0.9 to 3.1 years and transfer velocities varying from 6.6 to 28.0 m/year. These water flux velocities were compared with estimates made by means of a physically based conceptual flow model (Hydrus-1D). The transfer velocity calculated with the transfer function model was about five times higher than those simulated with the flow model. This could be explained by the fact that the flow model requires various hydrogeological inputs which are affected by considerable uncertainty and that it doesn't take the local soil characteristics and heterogeneities into account. In contrast, the transfer function only needs easily measurable time series and has the advantage to be site-specific. However, in case the "pushing effect", which intrinsically is modeled with the transfer function model, would be important, then the travel time estimated from the transfer function would not reflect real mass transport from the surface to the groundwater body. The importance of the "pushing effect" could not be quantified with the present set-up. It is therefore suggested to complete this study with more advanced mass tracer experiments.

Acknowledgements S. Mattern is a Research Fellow supported by the "Fonds pour la formation à la Recherche dans l'Industrie et dans l'Agriculture" (F.R.I.A. Belgium). The authors are grateful to the "Direction Générale des Ressources Naturelles et de l'Environnement" (Ministère de la Région Wallonne) for delivering the "10-Sous" database and to Gérard Peeters (Université catholique de Louvain) for delivering hydrogeological and piezometric data.

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