Parameter Estimation of the Transient Storage Model for Stream–Subsurface Exchange

Andrea Marion\(^1\); Mattia Zaramella\(^2\); and Aaron I. Packman\(^3\)

Abstract: The transient storage model (TSM) is the most commonly used model for stream–subsurface exchange of solutes. The TSM provides a convenient, simplified representation of hyporheic exchange, but its lack of a true physical basis causes its parameters to be difficult to predict. However, the simple formulation makes the model a useful practical tool for many applications. This work compares the TSM with a physically based pumping model. This comparison is advantageous for two reasons: Advection pumping is known to be an important hyporheic exchange process in many streams, and the pumping model can be used to derive dimensionless transient storage parameters that are properly scaled with important physical stream parameters. Transient storage model parameters are shown to be dependent on both the timescale of observation and the shape of the breakthrough curve, i.e., on the temporal evolution of the solute concentration in the surface water. This indicates that the transient storage model can, in practice, lead to incorrect predictions when model parameters are obtained without consideration of the stream flow dynamics, the properties of the stream bed, or the process timescale. This work emphasizes the limitations of simplified models for hyporheic transport, and indicates that such models need to be carefully applied.

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Introduction

Stream–subsurface exchange has received significant attention in recent years, primarily for its role in controlling the fate of contaminants and ecologically relevant substances. Stream–subsurface exchange can control the net transport of reactive substances because this exchange exposes solutes and suspended matter to subsurface biogeochemical processes. The attention given to the exchange of solutes between the surface water of a river and the hyporheic zones has led to the development of various types of mathematical models for exchange (Bencala and Walters 1983; Castro and Hornberger 1991; Elliott and Brooks 1997a; Runkel 1998; Wörman et al. 1998; Wörman et al. 2002). These models have been applied extensively to solute transport studies in many streams (Bencala 1984; Castro and Hornberger 1991; Rutherford et al. 1993, 1995; Vallet et al. 1996; Mulholland et al. 1997). Concerns about the generality of existing models has led to several recent attempts to reconcile existing models (Packman and Bencala 2000; Lees et al. 2000). This work will compare a simplified model with a fundamentally based model for advective hyporheic exchange in order to further define the limitations of the commonly used models.

The compared models are mostly used to analyze hyporheic exchange: They are commonly known as the pumping model (PM) and the transient storage model (TSM). The TSM provides a way to synthesize the net mass transfer from a stream to its bed. Its parameters are a storage zone of constant depth identified in the upper part of the sediments and a coefficient commonly known as “mass transfer coefficient.” The exchange is assumed to be proportional to the difference of concentration between water and the storage zone through the mass transfer coefficient. The TSM is usually formulated as follows [Bencala and Walters (1983)].

\[
d\frac{C_b}{dt} = \alpha \frac{A}{A_s} (C_w - C_b) \quad (1)
\]

where \(A\)=cross-sectional area of the stream; \(A_s\)=cross-sectional area of the storage zone; \(\alpha\)=mass transfer coefficient; \(C_w\)=concentration of solute in the stream; \(C_b\)=concentration of solute in the storage zone; and \(t\)=time.

The exchange parameters are evaluated on the basis of curve fitting the results of solute transport experiments conducted in the stream of interest. The TSM has been applied to a large number of streams, principally to evaluate metals transport in mining areas and the influence of hyporheic exchange on nutrient transport (Triska et al. 1989, 1990, 1993; Winter et al. 1998; Jones and Mulholland 2000). Transient storage parameters can be evaluated directly from tracer measurements (Hart 1995; Wagner and Harvey 1997; Packman and Bencala 2000). Several other models exist that are similar to the TSM in that they represent hyporheic exchange as mass transfer from the stream to an idealized subsurface reservoir using an empirical exchange coefficient (e.g., O’Connor 1988).

Recent work has probed the underlying processes which drive hyporheic exchange. Hyporheic exchange is now known to be

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\(^{1}\)Assistant Professor, Dept. of Hydraulic, Maritime, Environmental, and Geotechnical Engineering, Univ. of Padua, via Loredon 20, 35100 Padova, Italy (corresponding author). E-mail: marion@idra.unipd.it

\(^{2}\)PhD Candidate, Dept. of Hydraulic, Maritime, Environmental, and Geotechnical Engineering, Univ. of Padua, via Loredon 20, 35100 Padova, Italy.

\(^{3}\)Assistant Professor, Dept. of Civil and Environmental Engineering, Northwestern Univ., 2145 Sheridan Rd., Evanston, IL, 60208-3109.

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driven primarily by advective processes. Advective exchange flows develop at several spatial scales due to separate mechanisms such as flow over bedforms, around obstacles, and around meanders (Thibodeaux and Boyle 1987; Savant et al. 1987; Harvey and Bencala 1993; Elliott and Brooks 1997a,b; Hutchinson and Webster 1998; Packman and Brooks 2001). The bedform-induced PM developed by Elliott and Brooks calculates hyporheic exchange within the stream bed based on a fundamental analysis of hydrodynamic interactions between the stream and subsurface.

The case of uniform stream flow over an infinitely deep stream bed covered by regular, two-dimensional bedforms, is approximated by a sinusoidal pressure distribution acting on a flat bed. Considering a coordinate system (x,y) parallel and perpendicular to the bed, respectively, the induced pore water flow has Darcy velocity components

\[ u = -kKh_m \cos(kx) \exp(-ky) \]
\[ v = -kKh_m \sin(kx) \exp(-ky) \]

in the x and y directions, respectively. Exchange is driven by the sinusoidal variation of head at the bed surface with half-amplitude \( h_m \), the bedforms are characterized by their wavelength \( \lambda \) or wave number \( k = 2\pi/\lambda \), and the sediments have hydraulic conductivity \( K \). The half-amplitude \( h_m \) is a function of the stream velocity and bed form geometry (Feltham 1985). The stream–subsurface exchange flux can be found by evaluating Eq. (2) at the bed surface. Further, solute mixing in the subsurface can be calculated explicitly because Eq. (2) provides the entire subsurface velocity field. The PM has been successfully applied to several different sets of experimental results (Elliott and Brooks 1997b; Packman et al. 2000a; Marion et al. 2002).

The most important feature of the PM is that physical quantities like the stream velocity, permeability of the sediments, and channel geometry are used to evaluate hyporheic exchange. Because of this, it has predictive capability and can also be modified to be applicable to a slightly different physical system or extended to include chemical reactions in a detailed way. The effects of layered subsurface heterogeneity, an impermeable subsurface layer, and active bed sediment transport have been included in various pumping model formulations (Packman et al. 2000a; Packman and Brooks 2001). The PM has also been extended to predict the transport of reactive solutes and colloidal particles (Eylers et al. 1995; Rutherford et al. 1995; Packman et al. 2000a,b). However, because it requires detailed knowledge of streambed conditions, the PM has been directly applied to few natural streams (Rutherford et al. 1993, 1995). Wörman et al. (2002) have recently presented a methodology that allows application of the PM to stream solute breakthrough data using moment methods, and demonstrated the utility of this approach in assessing geomorphic controls on hyporheic exchange in an agricultural stream.

In this work, the TSM will be compared to the fundamentally based PM in order to examine if their simplified formulations can adequately represent the pumping exchange process. Model simulations will be used to examine the behavior of the idealized model parameters and to determine how the semiempirical exchange parameters can be estimated from stream properties.

**Exchange Models: Theory**

**Model Parameter Normalization**

The PM is normally applied in a dimensionless form with all parameters scaled appropriately with characteristic physical dimensions and hydrogeologic properties of the stream system (Elliott and Brooks 1997a). Exchange is normally represented with the equivalent depth of penetration, \( M \), which is the volume of water exchanged per unit bed surface area. Pumping exchange can then be given in terms of a dimensionless time and dimensionless exchange

\[ t^* = \frac{k^2 K h_m}{\theta} t \]
\[ m^* = \frac{2 \pi k}{\theta} M \]

where \( t^* = \text{dimensionless time}; m^* = \text{dimensionless penetrated mass}; \) and \( \theta = \text{porosity of the bed} \).

Cultures such as flow over bedforms, around obstacles, and around meanders, which is the probability that a particle that entered in the sediments at time \( \tau^* = 0 \) is still in the bed at a general time \( \tau^* \). If the tracer concentration \( C(t) \) in the water column is known, then the dimensionless penetrated mass can be evaluated as

\[ m^*(t^*) = 2 \pi q \int_0^{t^*} R(\tau^*) C^* (t^* - \tau^*) d\tau^* \]

where \( C^* = C/C_0 = \text{concentration normalized with the initial concentration} \), and \( q^* = -1/(2\pi K h_m) \int_{x=0}^{\lambda} v(x,y=0) dx = \text{dimensionless average flow through the bed surface} \).

A dimensionless formulation of the TSM can be developed. For a wide channel, or one with constant width, Eq. (1) can be written as follows:

\[ \frac{dC_b}{dt} = \frac{\alpha^*}{\delta} (C_w - C_b) \]

where, if \( H = \text{stream depth} \), the coefficient

\[ \alpha^* = \alpha \theta H \]

is the exchange rate with units of velocity and \( \delta = \text{depth of the storage zone} \). By introducing the PM scaling, the dimensionless exchange rate and storage zone depth are found

\[ \alpha^* = \frac{\alpha \theta H}{kK h_m} \]
\[ \delta^* = k \delta \]

According to these expressions, Eq. (5) becomes

\[ \frac{dC_b^* (t^*)}{dt^*} = \frac{\alpha^* \delta^*}{\delta^*} [C_w^* (t^*) - C_b^* (t^*)] \]

Eqs. (4) and (9) can be applied to any desired boundary and initial conditions. Three different stream concentration curves will be considered in the following analysis. These curves represent forcing functions that drive hyporheic exchange when the in-stream transport is uncoupled from hyporheic transport. The three functions have been selected for analysis here because they represent useful practical cases and have analytical solutions. Analytical solutions are needed in order to proceed with the numerical model.
evaluation. It is worth noting that field cases may have a variety of boundary conditions, so the models should be solved for more complex boundaries on a case-by-case basis.

**Transient Storage Model: Solution for Closed System, Exponential Decay, and Sinusoidal Concentration**

The simple formulation of the TSM makes it easily solvable for complex cases, so that it is particularly suitable for practical applications. Due to the simplification inherent in the TSM, values for the storage zone and the mass transfer coefficient that work for a particular case may not work for a different one, even for the same stream and the same bed configuration. This is the reason for evaluating three different boundary conditions representative of the different kinds of processes that occur in rivers.

**Closed Systems**

Consider laboratory tests where the volume of surface water (including the water column and sumps) and the volume of the water stored in the porous bed are fixed, and water is recirculated. This system is here termed "closed". It is here assumed that the porous bed is thick enough for the exchange process not to be affected by the bed thickness. The initial conditions $C_w=C_0$ and $C_b=0$ reflect the experimental conditions discussed in this work. Closed systems require a relationship linking the transferred mass to the concentration in the water. A mass balance provides the required expression

$$\frac{dm(t)}{dt} = -d' \frac{dC_w^*(t)}{dt}$$  \hspace{1cm} (10)

where $d'$ = effective stream depth defined as the total volume of stream water, including off-channel components (sumps), divided by the exchange area, i.e., the area of the interface between the surface flow and the bed. Eq. (10) is valid for both the TSM and PM.

Defining

$$d^* = kd'$$  \hspace{1cm} (11)

= normalized stream depth, integration of Eq. (10) leads to the following:

$$C_w^*(t) = 1 - \frac{\theta}{2\pi} \frac{m^*(t)}{d^*}$$  \hspace{1cm} (12)

The effective depth of penetration into the bed can also be written as a function of the concentration in the storage zone

$$m(t) = \theta \delta C_b^*(t)$$  \hspace{1cm} (13)

which can be normalized according to Eqs. (8) and (3), resulting in

$$m^*(t) = 2\pi \delta^* C_b^*(t)$$  \hspace{1cm} (14)

The dimensionless expression relating the concentrations in the bed and stream, found by substituting Eq. (14) into Eq. (12), is

$$C_b^*(t) = [1 - C_w^*(t)] \frac{d^*}{\theta \delta^*}$$  \hspace{1cm} (15)

By combining Eqs. (15) and (9), the TSM solution for the mass transfer from the stream to the bed is found

$$C_b^*(t^*) = \frac{1 - \exp\left(-\frac{\theta \delta^*}{d^*} + 1\frac{\alpha^*}{\delta^*} t^*ight)}{\theta \delta^*}$$  \hspace{1cm} (16)

It can be shown from Eqs. (16) and (17) that the solute concentration in the pore water increases with an initial rate equal to $\alpha^*$, and the solute concentration everywhere eventually approaches an asymptotic value given by complete mixing through the entire transient storage zone

$$C_f^* = \frac{1}{\theta \delta^*} \frac{1}{d^* + 1}$$  \hspace{1cm} (18)

The case of a sudden change of the solute concentration in a river followed by steady conditions (step change) is now examined. The solution can be derived readily from Eq. (17), as the limit for $d^* \to \infty$

$$m^*(t^*) = 2\pi \delta^* \left(1 - \exp\left(-\frac{\alpha^*}{\delta^*} t^*ight)\right)$$  \hspace{1cm} (19)

**Sinusoidal Variation**

Another important case is when the concentration in the water changes periodically, which may occur due to diurnal or seasonal variations of the tracer concentration in the water. This case is also applicable to variations produced by tides or other periodic events such as industrial discharge cycles. To represent a periodic variation of the concentration in the water, the following expression is introduced

$$C_w(t^*) = C_m \sin\left(\frac{2\pi}{\Pi \tau^*} t^*\right)$$  \hspace{1cm} (20)

where $\Pi^* = k^2 Kh_m / \theta \Pi =$ normalized period of oscillation. Eq. (20) should be taken as a periodic deviation from some mean background concentration. The normalized depth of penetration that occurs due to this periodic forcing is

$$m^*(t^*) = \frac{2\pi \delta^*}{\Pi^* \alpha^*} \left[\sin\left(\frac{2\pi}{\Pi^* \alpha^*} t^*\right) + 1\right]$$

$$- \frac{2\pi \delta^*}{\Pi^* \alpha^*} \left[\cos\left(\frac{2\pi}{\Pi^* \alpha^*} t^*\right) - \exp\left(-\frac{\delta^*}{\alpha^*} t^*\right)\right]$$  \hspace{1cm} (21)

**Exponential Decay**

A fourth interesting case is the exponential decay of the concentration in the water. This is representative of an accidental discharge of a contaminant that degrades exponentially over time. It may also represent a substance that is consumed due to a reaction that occurs only in the stream and not in the subsurface, such as photodegradation. In this case, the concentration in the stream is assumed to be

$$C_w(t^*) = C_0 \exp\left(-\frac{t^*}{\Gamma}\right)$$  \hspace{1cm} (22)

where $\Gamma^* = k^2 Kh_m / \theta \Gamma =$ characteristic decay time. The analytical solution of the TSM can easily be found for this boundary condition.
The TSM parameters will be evaluated for each of these cases.

**Basis for Model Comparison**

Models can be evaluated or compared in several different ways. The approach that will be taken here is to fit the TSM to the PM and then to evaluate whether the resulting idealized exchange parameters, $\alpha^*$ and $\delta^*$ are reasonable. If the idealized model works well, then its parameters should behave in a manner that is realistic and consistent with model assumptions. The model-fitting approach is appropriate because the PM is a theoretically based description for curve fitting without a firm physical basis. It may indicate that it may merely provide a convenient mathematical description for curve fitting without a firm physical basis.

The parameter $\Delta$ is defined as

$$\Delta(\alpha^*, \delta^*, t^*) = m^*_p(t^*) - m^*_{\text{TSM}}(t^*)$$

then the mean-square error of the function $\Delta$ is given by

$$S(\alpha^*, \delta^*, T^*) = \frac{1}{T^*} \int_0^{T^*} [\Delta(\alpha^*, \delta^*, t^*)]^2 dt$$

where $T^*$ = time interval over which $S$ is evaluated and thus is equivalent to a “fitting timescale”. The nondimensional relative mean-square error

$$S^*(T^*) = \frac{1}{T^*} \int_0^{T^*} [\Delta^2 dt^*]$$

is similarly used to evaluate the performance of the model for different dimensionless timescales $T^*$.

The model comparison is made by finding the minimum value of $S$ as a function $\alpha^*$ and $\delta^*$. The best-fit values for the exchange parameters correspond to the points where the gradient of $S$ is 0, and where the Jacobian is strictly positive

$$\text{grad}(S) = \begin{bmatrix} \frac{\partial S}{\partial \alpha^*} & \frac{\partial S}{\partial \delta^*} \end{bmatrix}$$

$$= \begin{bmatrix} \int_0^{T^*} \frac{\partial }{\partial \alpha^*} [\Delta(t^*)]^2 dt^* & \frac{1}{T^*} \int_0^{T^*} \frac{\partial }{\partial \delta^*} [\Delta(t^*)]^2 dt^* \end{bmatrix}$$

$$= (0, 0)$$

$$J(S) = \begin{bmatrix} \frac{\partial^2 S}{\partial \alpha^*^2} & \frac{\partial^2 S}{\partial \alpha^* \delta^*} \\ \frac{\partial^2 S}{\partial \delta^* \alpha^*} & \frac{\partial^2 S}{\partial \delta^*^2} \end{bmatrix} > 0$$

**Fig. 1. Step change of the stream concentration. Best-fit values of (a) the dimensionless exchange coefficient $\alpha^*$ and (b) the dimensionless depth of the exchange layer $\delta^*$. On the right-hand side vertical axis the dimensionless mean square deviation $S^*$ is reported.**

Results will be presented in the next section as the best-fit values of the exchange parameters for various time scales of comparison.

**Results and Discussion**

**Model Comparison and Parameter Estimation**

Fig. 1 presents the best-fit values for TSM parameters for the case of a constant tracer concentration in the stream. For $T^* \rightarrow 0$, the normalized depth of the contaminated layer tends to infinity, while the normalized exchange rate assumes a finite value. This can be explained by a direct analytical comparison of the models. The PM initially has a linear increase of the exchanged mass. This behavior can only be modeled with the TSM using an infinite depth of the storage zone. Near the origin, the residence time function has an approximate value of 1 (Elliott and Brooks 1997a). By substituting $R(t^*)=1$ in Eq. (5) and remembering that $C^*(t^*)=1$, then

$$m^*_p(t^*) \approx 2\pi q^*(t^*)$$

as an approximation for the pumping exchange at short times. This is equivalent to assuming that the normalized TSM exchange rate is equal to the normalized flux through the bed surface, $\alpha^* = q^*$. In this case, $\alpha^*(0)=1/\pi$, as shown in Fig. 1.

As the fitting timescale increases, the pumping model loses its linear behavior, and a finite best-fit value of the storage zone
depth can be derived through optimization. If the optimization is made for long timescales, the TSM storage zone depth increases according to the net mass exchange, resulting in a logarithmic growth of $d^*$ ($T^*$). This reflects the fact that pumping exchange is characterized by mixing to a greater depth over time. However, as $d^*$ increases, the exchange rate $\alpha^*$ decreases over time. The TSM represents pumping exchange with various degrees of approximation, as indicated by the value of the normalized mean-square deviation $S^*$.

The simulations in Fig. 2 show the application of the transient storage model for two values of $T^*$. For a timescale $T^* = 10$ [Fig. 2(a)], which falls within the region of minimum $\delta^*$ in Fig. 1, the TSM provides a good approximation of the PM. The TSM does not perform nearly as well for a comparison timescale $T^* = 500$ [Fig. 2(b)]. The TSM still provides a fair representation of the PM curve, but it underpredicts the initial exchange, asymptotic behavior and intermediate exchange. This behavior results because the TSM does not provide a good approximation for the pumping process over large timescales. The idealization of a defined transient storage zone characterized by a given area or depth is not representative of hyporheic exchange in an unconstrained alluvial stream. The TSM exponentially approaches a constant tracer concentration in the storage zone, while advective pumping continues to mix tracer deeper into the bed. Pumping exchange slows considerably at long times due to the fact that pore water velocities decrease exponentially with depth, but it never actually stops. Figs. 2(a and b) indicate that this process cannot be represented with constant values of $\alpha^*$ and $\delta^*$, and can only be reproduced in the TSM if $\delta^*$ is allowed to increase over time and $\alpha^*$ is allowed to decrease.

Optimization over the two different timescales also yields very different values of the TSM parameters even though the stream conditions are the same. The optimal values $\alpha^* = 0.32$ and $\delta^* = 2.05$ were found for $T^* = 10$, while $\alpha^* = 0.08$ and $\delta^* = 4.97$ were found for $T^* = 500$. This result has serious implications for the use of the TSM in conjunction with tracer experiments to characterize hyporheic exchange in streams.

The same optimization procedure has been used to compare the TSM and the PM for the cases where the tracer concentration in the stream varies sinusoidally or decays exponentially. Results are shown in Fig. 3 for the sinusoidal case and in Fig. 4 for the exponential case. The different curves in Fig. 3 correspond to different oscillation periods. All of the curves have similar values for short timescales because the PM behaves similarly for different $\Pi^*$ in the initial linear exchange. The periodic variation of the tracer concentration in the stream water causes the TSM parameters to approach a steady state at long times. The concentration in the stream and the exchanged mass oscillates with the same frequency but with different phases. It is important to note that different oscillation periods yield different asymptotic values of the TSM parameters even though the magnitude of the concentration variation and all physical stream parameters are the same. This again shows that the best-fit TSM parameters obtained from solute measurements are dependent on the nature of the imposed solute breakthrough curve in the stream.

Fig. 5 shows the TSM behavior for an oscillation period $\Pi^* = 10$ and a timescale $T^* = 50$. The TSM provides good results in this case because the relatively short oscillation period results

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**Fig. 2.** Step change of the stream concentration. Comparison of the transient storage model and the pumping model for (a) small dimensionless timescales and (b) large dimensionless timescales.

**Fig. 3.** Sinusoidal change of the stream concentration. Best-fitting values of (a) the dimensionless exchange coefficient $\alpha^*$ and (b) the dimensionless depth of the exchange layer $\delta^*$. 

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in only a superficial contamination of the bed. Longer oscillation periods cause deeper bed layers to become mixed, resulting in poor model performance similar to that shown in Fig. 2 for a high comparison timescale.

The results for the exponential decay case are shown in Fig. 4. Asymptotic constant values of $\alpha^*$ and $\delta^*$ are also found in this case, and these depend on the magnitude of the decay timescale $\Gamma_*$. In all cases, the optimal TSM parameters follow the curve for the steady concentration case until the in-stream degradation causes them to deviate and approach an asymptotic constant value. Substances that degrade more quickly have a higher apparent exchange rate $\alpha^*$ and a lower $\delta^*$. Fig. 6 shows the comparison between the models for $\Gamma^*=100$ and $T^*=600$. The relatively long decay time of the tracer concentration in the water column (corresponding to a persistent contaminant) causes the tracer to penetrate deeply into the bed, and this deep penetration is not well modeled by the TSM. This is analogous to the behavior of the steady concentration case at long comparison times.

**Comparison with Experimental Results**

Experimental data obtained by Marion et al. (2002) will be used to demonstrate a few key principles. These experiments involved the observation of stream–subsurface exchange in a relatively large flume with a coarse sandbed. Four different bed geometries were used, including both naturally formed and artificially shaped bedforms. The fitting procedure has been applied to analyze the optimal values of the TSM parameters $\alpha^*$ and $\delta^*$ to reproduce the experimental data. These curves of $\alpha^*$ and $\delta^*$ versus the fitting timescale $T^*$ are shown in Fig. 7. The results obtained from the experimental data follow the form of the theoretical curves, especially for the natural bedforms. This example confirms that the best-fit TSM parameters are dependent on the timescale of the experiment. This approach shows also that the normalization presented in Eqs. (3), (7), and (8) is capable of simulating bedform-driven hyporheic fluxes. This normalization is a useful way to compare experimental results and model predictions for different stream conditions.

**Conclusions**

The TSM is a very useful tool for the practical analysis of stream–subsurface exchange. However, it does not fully represent the physics of the pumping process, wherein advective flows are produced by the pressure gradients at the bed surface associated with the presence of bedforms. The results shown here indicate that the two calibration parameters employed by TSM are not constant for a given stream, but instead change depending on the method used to obtain them. Comparison of the TSM with the fundamentally based PM shows that the TSM parameters change with the timescale of the theoretical evaluation and, correspondingly, vary with the duration of experiments used to probe the hyporheic exchange process. The quality of the TSM fit to the pumping model also depends on the timescale of comparison. Furthermore, the model evaluation has been carried out for three different concentration curves, representing common contamination cases. TSM parameters are shown to also be dependent on
The following symbols are used in this paper:

- \( A \): cross-sectional area of the stream;
- \( A_s \): cross-sectional area of the storage zone;
- \( C_b \): concentration of tracer in the storage zone;
- \( C_w \): concentration of tracer in the stream;
- \( C_0 \): initial tracer concentration in the stream, equal to \( C_w/C_0 \);
- \( C^* \): dimensionless tracer concentration;
- \( C^*_f \): dimensionless final dilution concentration;
- \( d' \): effective stream depth;
- \( d^* \): dimensionless water depth above sediments, equal to \( k d' \);
- \( H \): stream depth;
- \( h_m \): half-amplitude of the sinusoidal distribution of pressure on the bed surface;
- \( K \): hydraulic conductivity;
- \( k \): bedform wavenumber, equal to \( 2\pi/\lambda \);
- \( M \): net exchanged mass per unit bed surface area;
- \( m \): equivalent penetration depth, equal to \( M/C_0 \);
- \( m^* \): dimensionless equivalent penetration depth, equal to \( 2\pi k m/\theta \);
- \( \bar{R} \): flux weighted, spatially averaged residence time function;
- \( S \): mean square deviation;
- \( T^* \): fitting timescale;
- \( t^* \): dimensionless time, equal to \( k^2 k h_m t/\theta \);
- \( u \): Darcy horizontal velocity;
- \( u^* \): dimensionless Darcy horizontal velocity, equal to \( u/(k k h_m) \);
- \( v \): Darcy vertical velocity;
- \( v^* \): dimensionless Darcy vertical velocity, equal to \( v/(k k h_m) \);
- \( x \): horizontal coordinate;
- \( x^* \): dimensionless longitudinal coordinate, equal to \( k x \);
- \( y \): vertical coordinate;
- \( y^* \): dimensionless vertical coordinate, equal to \( k y \);
- \( \alpha \): mass transfer coefficient;
- \( \alpha' \): exchange rate, equal to \( \alpha \theta H \);
- \( \alpha^* \): dimensionless exchange rate, equal to \( \alpha \theta H/(k k h_m) \);
- \( \Delta \): disagreement between pumping model and transient storage model;
- \( \delta \): storage zone depth;
- \( \delta^* \): dimensionless storage zone depth, equal to \( k \delta \); and
- \( \theta \): porosity of bed sediment.

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- \( C_w \): concentration of tracer in the stream;
- \( C_0 \): initial tracer concentration in the stream, equal to \( C_w/C_0 \);
- \( C^* \): dimensionless tracer concentration;
- \( C^*_f \): dimensionless final dilution concentration;
- \( d' \): effective stream depth;

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