Separating surface storage from hyporheic retention in natural streams using wavelet decomposition of acoustic Doppler current profiles

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1. Introduction

Channel morphology and hydrodynamics play an important role in controlling nutrient retention, reactive transport and biogeochemical transformations in stream ecosystems [Mulholland et al., 1999]. Transient storage (hereinafter TS) refers to multiple processes by which solutes move into and out of near-bed sediments (hyporheic exchange [Bencala and Walters, 1983]) or are temporarily retained by surface features such as eddies and pools, vegetation and woody debris [Ensign and Doyle, 2005; Runkel and Kimball, 2002]. Although both mechanisms result in a delay in the downstream transport of solute mass, the ability to distinguish between hyporheic exchange and surface storage [Runkel et al., 2003] is important in many transport considerations such as stream nitrogen cycles [Kasahara and Wondzell, 2003], redox zonation and the interpretation of biogeochemical activity [Gooseff et al., 2005].

The ability to separate surface storage (e.g., due to in-channel features such as eddies and pools) from retention due to hyporheic exchange is important in many solute transport considerations; however, current stream tracer approaches do not allow such separation. We examined transient storage processes in a fourth-order Michigan stream using tracer studies, numerical flow and transport models, and hydrodynamic data obtained from acoustic Doppler current profiler (ADCP) surveys. Since the high-resolution, three-dimensional velocity fields obtained from an ADCP relate to in-channel processes, we used wavelet decomposition to separate the flow into regions of slow and fast moving zones and to estimate the relative sizes of the main channel (A) and the storage zones (A_s).

Transport modeling based on the tracer data provided estimates of storage zone sizes that included contributions from both surface storage and hyporheic exchange. By coupling the estimates from tracer data with those obtained from an ADCP we were able to assess the relative importance of surface storage in different stream reaches. Estimated (A_s/A) values in three test reaches ranged from 0.12 to 0.22, and transient storage residence times varied from approximately 4 min in a run reach dominated by surface storage (reach A) to about 13 min in a reach with some potential for hyporheic exchange (reach C). In reach A the (A_s/A) values estimated from tracer and ADCP data were in good agreement, indicating that in-channel processes were the main mechanism responsible for storage in this reach. Reach C estimates, however, showed that surface storage (A_s/A = 0.05) accounted for only a fraction of the transient storage estimated using tracer data (A_s/A = 0.12), which indicated that hyporheic exchange contributed to transient storage in the reach. The wavelet decomposition approach based on the ADCP data provides a framework to better constrain transient storage models and to eliminate unrealistic parameters.

includes the combined effects of a number of processes such as turbulent mixing, molecular diffusion, and mixing due to vertical and transverse shear. Efforts to interpret model parameters and to relate them to physical stream characteristics are often confounded by the inability of current stream tracer techniques to separate TS processes [Goosseff et al., 2005] as well as mathematical difficulties associated with parameter estimation, especially in the presence of competing parameters and false/singular convergence [Runkel, 1998].

Wavelet analysis has been used extensively in the past decade to analyze both time series and spatial data in geophysical and engineering applications. Examples include the analysis of streamflows [Costilbaly and Burn, 2004], seismogram data [Lockwood and Kanamori, 2006], long-term trends in geomagnetic activity [De Artigas et al., 2006], and ocean waves [Pairaud and Auclair, 2005]. Wavelet analysis [Mallat, 1989] allows the original signal or image to be split into different components so that each component can be studied with a resolution that is suitable for its scale. In particular, the signal can be decomposed into slow changing (i.e., coarse or low frequency) features and rapidly varying (i.e., fine or high frequency) features using low-pass and high-pass wavelet filters at different levels (or scales) in a multilevel decomposition. This feature of wavelet analysis is particularly attractive for studying TS processes in streams. Previous applications of two-dimensional wavelets include characterization of permeability anisotropy [Neupauer et al., 2006], studies of spatiotemporal dynamics of turbulence [Guan et al., 2003] and analysis of spatial rainfall data [Kumar and Foufoula-Georgiou, 1993].

The aim of this study was to propose a method to separate surface storage from hyporheic exchange by coupling traditional TS modeling with wavelet decomposition of data obtained from acoustic Doppler current profiler (ADCP) surveys. Our hypothesis is that the size of the surface storage zones can be estimated using ADCP data. By comparing the parameters estimated using traditional TS modeling with those obtained independently from ADCP data, we will be able to assess the relative importance of surface storage in different stream reaches. Recent advances in ADCP technology allow the measurement of high-resolution three-dimensional velocity data in rivers. Although ADCPs are routinely used to measure stream discharge, the idea of using ADCP data to estimate storage zone sizes has not been explored earlier. In this paper, we report new analyses as well as tracer and ADCP data for a fourth-order Michigan stream and demonstrate how improved understanding of TS processes can be obtained by coupling the two types of data.

2. Site Description

The Red Cedar River (RCR) is a fourth-order stream in south central Michigan, United States, that drains a landscape dominated by agriculture and urbanization. It originates as an outflow from Cedar Lake, Michigan, and flows into East Lansing and Michigan State University (MSU). The river then connects with the Grand River in Lansing, Michigan. The total stream length is approximately 70 km. The river and its tributaries drain an area of about 1,230 km², one fourth of which is drained by Sycamore Creek (Figure 1). Limestone, shale, sandstone, coal, and other sedimentary rocks arranged in almost horizontal layers compose the bedrock surface of the Red Cedar Watershed. The river is characterized by a meandering channel and low stream gradient. A USGS gauging station (04112500) is located at the Farm Lane Bridge (Figure 1) and measures runoff from about 75% of the basin. The study reach was bounded by Hagadorn Bridge on the East and the Kalamazoo Street Bridge on the West (Figure 1). The RCR meanders through the MSU campus over a stretch of approximately 5 km (our study reach in this paper) and has an average slope of 0.413 m/km. The river width varies considerably, from 16.3 m to 40.4 m, with an average of 28.1 m (Figure 2). The general trend is that the width decreases for the initial 2400 meters, increases over the next 2100 meters and finally decreases for the remaining 600 meters.

3. Materials and Methods

Four tracer studies (slug additions) were conducted in summer 2002 on 17 May, 31 May, 6 June, and 21 June. Average discharge (Q) recorded on these four days at the USGS gage near Farm Lane Bridge was 16.82, 14.41, 19.06 and 2.49 m³/s respectively. Discharge and hydrodynamic data were collected from ADCP surveys during the period 2002–2006. ADCP data could not be collected during the tracer tests in 2002 but on two occasions the discharge values were similar to those during the tracer tests. ADCP data collected on 19 March 2006 (Q = 19.89 m³/s) and 29 September 2005 (Q = 2.0 m³/s) were used to compare with results from the tracer studies conducted on 6 and 21 June 2002 respectively. Because of low-flow conditions (long travel times of the dye cloud), the tracer study on 21 June 2002 could not be completed. For this discharge, we present results for only one sampling location. Tracer transport was described using the TS equations [Bencala and Walters, 1983]. A one-dimensional hydrodynamic model based on the St. Venant equations was used to examine two of the estimated parameters in the TS equations (mean velocity and dispersion coefficient). For all the four slug injections, the dye was released at the Hagadorn Bridge and samples were collected at the Farm Lane, Kellogg and Kalamazoo Bridges, respectively (Figure 1). The distances to the three sampling locations from the point of release are: 1400 m (Farm Lane), 3100 m (Kellogg) and 5079 m (Kalamazoo). Flourescein dye was released in the middle 75% of the channel to ensure conditions of instantaneous mixing. Sampling was done at the middle of the cross section for each dye release. Fluorescein solution with a concentration of 179.06 g/L was used for all the tracer tests. The volume of the dye used was based on a desired peak concentration range of 10–20 µg/L at the last sampling point. Samples were analyzed using a Turner10-AU field fluorometer (Turner Designs Inc., Sunnyvale, CA). Observed tracer data was checked for mass conservation. The mass of the tracer in the dye cloud as it passed a sampling point was calculated using the relation

\[ m(t) = Q \int C(t) \, dt \]
recovery values for all the flows (and at different sampling locations) varied from 0.91 to 1.5 with a mean value of 1.29 and a standard deviation of 0.16. To facilitate comparisons with mathematical models and to produce a data set that obeys dye mass balance, fractional recovery corrections were applied to the individual concentration values as reported by Atkinson and Davis [2000].

[8] Point velocities and river stages were measured on eight consecutive days in 2002 (4–11 April) to calibrate the hydrodynamic model. Price AA current meters and a 16-MHz

Figure 1. Red Cedar River Watershed showing the Red Cedar River and the sampling locations (bridges).

Figure 2. Variation of river width as a function of distance from the Hagadorn Bridge ($x = 0$). The dye was released at $(x = 0)$, and samples were collected at the three downstream bridges. The 5 km study reach was divided into three test reaches (A, B, and C) for the transient storage modeling as shown.
Sontek acoustic Doppler velocimeter (ADV) (Sontek/YSI Inc., San Diego, CA) were used to measure velocity profiles and discharge in the river. In addition, a vessel-mounted, down-looking 1200 kHz Rio Grande ADCP (Teledyne-RD Instruments, Poway, CA) was used to measure discharge and three-dimensional velocity fields in the river using an Oceanscience® Riverboat (a low-drag trimaran) equipped with radio modems for real-time deployments. By towing the vessel-mounted ADCP across a river transect (perpendicular to the flow direction), we obtained snapshots of instantaneous velocity fields in the river which were later used for wavelet analysis (described below). Typical boat speeds were in the 0.5–1 m/s range. Several steps were involved before analyzing the ADCP data. The speed at which the ADCP was towed across the river depended on the discharge and the ADCP operating conditions. The ADCP data was exported from WinRiver (the ADCP operating software) for further processing in MATLAB including smoothing and removal of bad ensemble values. Details of ADCP principles and processing are described by Simpson [2001] and Dimehart and Burau [2005]. The discharge values obtained from ADCP measurements at the Farm Lane Bridge were compared with those from the USGS gauging station on several occasions over a period of four years and an excellent agreement was obtained.

4. Analysis

4.1. Transient Storage Modeling

[9] Tracer transport was described using the TS equations (2 and 3) which describe transport in the main channel and the storage zones respectively [Runkel, 1998].

\[
\frac{∂C}{∂t} = - \frac{Q}{A} \frac{∂C}{∂x} + \frac{1}{A} \frac{∂}{∂x} \left( AD \frac{∂C}{∂x} \right) + \frac{q_s}{A} (C_L - C) + \alpha (C_S - C) \tag{2}
\]

\[
\frac{∂C_S}{∂t} = \frac{A}{A_S} (C - C_S) \tag{3}
\]

If equations (2) and (3) are applied on a reach basis, then the velocity (Q/A) in (2) is a reach-averaged value and, in general, is not equal to the local (point) velocity u. The TS model equations were solved using a fourth-order accurate compact numerical scheme [Demuren et al., 2001]. Briefly, the spatial derivatives were approximated using a fourth-order scheme with spectral-like resolution and a low-storage fourth-order Runge-Kutta scheme was used for temporal differencing. The resulting tridiagonal matrix system of equations was solved using the Thomas algorithm [Press et al., 2002]. A low Courant number of 0.15 and a uniform grid of 1001 points were used for all model runs.

[10] The boundary and initial conditions for the model were as follows. The river was assumed to be initially at zero tracer concentration. The upstream boundary was modeled to simulate slug release into the main channel and the storage zone was assumed to be initially solute-free. A no-flux boundary condition was specified at the downstream boundary for the transport equations.

[11] Parameters in the TS model (i.e., A, A_S, D, α) were estimated using a global optimization procedure, the Shuffled Complex Evolution (SCE) algorithm [Duan et al., 1992] by minimizing the root-mean-square error (RMSE) between the observed data and the model. This algorithm found wide applications in the hydrologic research literature and was shown to be robust and efficient in finding the global minimum [Duan et al., 1992]. For all the dye studies, optimal parameters were obtained in 3000–6000 iterations on a 512-core Western Scientific Opteron Cluster computing system at MSU. The RCR is generally a gaining stream; however, within the study reach the gain was not significant enough to change the TS parameters. After running two separate optimizations, with and without q_L, we decided to use the parameters obtained with q_L = 0.

4.2. Multiresolution Wavelet Analysis of ADCP Data

[12] To examine the estimated parameters and to relate them to the physical characteristics of the river, we used three-dimensional velocity data obtained from ADCP surveys. Given the 3-D velocity field in a river, the dispersion coefficient D can be computed by numerically integrating the velocities [Fischer et al., 1979]. However, we did not follow this approach as earlier studies found that dispersion estimates based on time-dependent velocity fields tend to be highly sensitive to velocity fluctuations [Palancar et al., 2003]. In this paper we focus on the parameters A_S and A. Repeated ADCP surveys in the study reach clearly showed regions of high and low velocities and acoustic backscatter (a well-known measure of suspended solids concentration, SSC) for different cross sections and led us to test the hypothesis that in-stream TS zones can be identified using the three-dimensional velocity fields obtained from ADCP surveys. Earlier studies [Sukhodolov et al., 2004; Tipping et al., 1993] showed that the concentration of suspended particulate matter reduces in the dead zones due to sedimentation of faster sinking fractions of suspended matter in the decelerating flow. Therefore, by identifying regions of decelerating flow or low SSC using ADCP data we may be able to identify the relative importance of dead zones in a river reach. [Engelhardt et al., 2004] noted a correspondence between SSC and mean velocity vectors and this correspondence was also evident in our ADCP data. To identify regions of relatively fast and slow moving water, we use multiresolution wavelet analysis of the two-dimensional (y, z) normalized mean velocity fields obtained from ADCP surveys. The continuous wavelet transform (CWT) of a function f(y, z) for a two-dimensional wavelet is defined as the convolution with a scaled and shifted version of the wavelet function Ψ.

\[
W(a, b) = \frac{1}{\sqrt{a_1 a_2}} \int f(y, z) \Psi \left( \frac{y - b_1}{a_1}, \frac{z - b_2}{a_2} \right) dydz
\]

\[
a = \begin{bmatrix} a_1 \\ a_2 \end{bmatrix}, \quad b = \begin{bmatrix} b_1 \\ b_2 \end{bmatrix}
\]

where a and b are the scale and translation vectors. We used the two-dimensional discrete wavelet transform (DWT) for our analysis. Similar to the 1D wavelet transform (Figure 3), the 2-D transform can decompose a given function into its slow changing (or coarse) features (called approximations) and fine (or rapidly changing) features (called details). For the two-dimensional case, the details can be further decomposed into horizontal, vertical and diagonal details.
Here the $W$ parameters are the wavelet coefficient matrices, $N$ denotes the number of levels in the decomposition, and the superscripts $A, V, H$ and $D$ denote the approximations (the first term in equation (6) and the vertical, horizontal and diagonal details respectively (the last three terms in equation (6)). The decomposition (6) allows us to examine the different components at multiple levels or scales. Our aim was to extract the coarse features (the first term in equation (6)) at different levels as they retain the essential features of the velocity field (the high-frequency content simply adds detail). Since our primary interest was in making a distinction between regions of slow moving or stagnant water (dead zones) and the main channel (two distinct scales), we used two-level decomposition ($N = 2$). After plotting the single-level reconstructions based on the approximations at levels 1 and 2 in the physical space, $(A_g/A)$ was computed from the area of all the pixels greater than a threshold value $T$ (corresponding to the background pixel value in the two images):

$$W^D_{m,ij} = \int \int \Psi_{m,ij}^{H}(y,z)f(y,z)dydz$$  \hspace{1cm} (10)

The Haar and Daubechies-12 wavelets [Daubechies, 1988] were used in our analysis; however, other wavelets produced essentially similar results. After performing the wavelet decomposition and extracting the terms $\sum_{ij} W_{N,ij}^a \Psi_{N,ij}^a(y, z)$ for $N = 1$ and $N = 2$ and plotting them in the physical space (i.e., as a function of $y$ and $z$ to identify the channel and the relative locations of the dead zones within the channel), functions in the MATLAB image processing toolbox were used to compute the areas and the ratio in equation (11).

### 4.3. Hydrodynamic Modeling

Hydrodynamic modeling was used mainly to understand differences in average stream velocities in the three reaches. The St. Venant (or shallow water) equations governing the one-dimensional flow in the river are given in (12) and (13). These equations are based on the assumption that Coriolis and other acceleration terms normal to the direction of flow are negligible and that bottom slopes are moderate. River stages were computed from the continuity equation (12) and velocities from the momentum equation (13).

$$\frac{\partial A}{\partial t} + \frac{\partial Q}{\partial x} = 0$$  \hspace{1cm} (12)

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + g \frac{\partial h}{\partial x} + \frac{g n^2}{R^2 T} |u| u = 0$$  \hspace{1cm} (13)
Bathymetry data for the hydrodynamic model were obtained from ADCP surveys and studies conducted by the MSU Physical Plant. Equations (12) and (13) were solved using an explicit scheme as implemented in DYNHYD [Ambrose, 1988]. The study reach was divided into 103 segments keeping the stability restriction in mind: \( \Delta x_i \geq (u_i + \sqrt{gH_i})\Delta t \). Measured discharges at the Hagadorn Bridge were specified as the boundary condition for the hydrodynamic model at the upstream boundary while a head versus time condition was specified at the downstream Kalamazoo Bridge. Both are well-known forms of boundary conditions used in river hydrodynamic modeling [Martin and McCutcheon, 1999]. The model was calibrated by adjusting the Manning’s roughness parameter, \( n \). Roughness values changed for different stream reaches and ranged from 0.021 to 0.036.

4.4. Need to Incorporate Transient Storage

While describing longitudinal solute transport in streams several authors found that their models did not need a separate term for longitudinal dispersion as long as TS was properly accounted for [Gupta and Cvetkovic, 2000; Worman, 1998]. In other cases, the classical advection-dispersion equation (ADE), which is a special case of the TS model, was found to describe tracer transport adequately [Schmid, 2004]. Before proceeding with a detailed analysis based on the TS model, we wanted to determine if TS was indeed important in our study reach. To answer this question, we examined our tracer data using an analysis originally described by [Chatwin, 1980] to ascertain the non-Gaussian behavior of tracer profiles. The solution of the ADE is given by:

\[
C_1(x,t) = \frac{M}{A\sqrt{4\pi Dt}} \exp\left( -\frac{(x-vt)^2}{4Dt} \right) \tag{14}
\]

where \( v = (Q/A) \) and \( C_1 \) is the solution of equation (2) with \( q_L = \alpha = 0 \). Chatwin rearranged equation (14) as shown below.

\[
\left\{ \log \left( \frac{M}{AC\sqrt{4\pi Dt}} \right) \right\}^{1/2} = \frac{x}{2\sqrt{D}} - \frac{v}{2\sqrt{D}t} \tag{15}
\]

The left-hand side of the above equation (called the Chatwin parameter, \( S \)) represents a straight line when plotted against time \( t \). Deviations from the straight line are taken as an indication of non-Gaussian behavior. As shown in Figure 4 for \( Q = 19.06 \text{ m}^3/\text{s} \) (similar plots were obtained for all other data sets), our tracer data clearly deviated from (15) at longer times, indicating the need for incorporating TS processes in the transport models.

5. Results

5.1. Comparison of the Numerical Model With Analytical Solutions

The fourth-order accurate compact scheme used to solve the TS equations in this paper was tested extensively and was used to solve similar sets of equations in the past [Phanikumar and McGuire, 2004]. To assess the accuracy of the compact scheme for solving the TS equations, we compared our numerical solutions with the analytical solutions reported by [De Smedt et al., 2005]:

\[
C(x,t) = \int_0^t \left[ \alpha + \left( \frac{x^2 - v^2 t^2}{4Dt^2} - \frac{1}{2\tau} - 0 \right) \right] J \left( \alpha t - \frac{\alpha(t - \tau)}{\beta} \right) \right)
\]

where \( \beta = A\gamma / A \) and \( C_1(x,t) \) is the classical solution to the advection-dispersion equation (equation (14)) with the same initial and boundary conditions [Chapra, 1997]. The \( J \) function in the above equation can be evaluated using the relations:

\[
J(a,b) = 1 - e^{-a} \int_0^b e^{-a} I_0(2\sqrt{b\lambda})d\lambda = 1 - e^{-a} \sum_{n=0}^{\infty} \sum_{m=0}^{\infty} \frac{I_0^n}{m! n!} \tag{17}
\]

where \( I_0 \) is the modified Bessel function of zero order. A different analytical solution to the TS equations was presented by [Hart, 1995] based on stochastic considerations. We found that both analytical solutions produce almost identical results; the difference was mainly at low concentrations. For some parameter values the solutions of
Hart, 1995] met with convergence problems (e.g., $\alpha > 0.005$). In equations (16) and (17) the computational effort to evaluate the $J$ functions can be significant, especially if the equations are used for parameter estimation since thousands of runs are often required to estimate the optimal parameters, therefore numerical models such as OTIS or the present model may be preferable. We selected the solution of [De Smedt et al., 2005] for our comparison as it was found to agree well with the numerical model OTIS [Runkel, 1998]. Comparisons with our numerical solutions obtained using 200 grid points are presented in Figure 5 for different values of $\alpha$. An excellent agreement is noted between the two solutions.

5.2. Evaluation of Channel Features and Potential for Hyporheic Exchange

[16] To estimate the TS parameters, the 5 km study reach was divided into three test reaches as shown in Figures 1 and 2. Surficial sediments in the study reach consisted of a thin (5 cm or less) layer of sand and gravel underlain by a heavily consolidated clay layer of depth 0.25 m or more [Uzarski et al., 2004]. Thus there was limited potential for hyporheic exchange in the study reach although there were differences between the test reaches. Reach A (between the Hagadorn and Farm Lane Bridges), a relatively straight section of the river, was free of alluvium and surface storage was the primary mechanism contributing to TS in the reach. This reach had extensive vegetation growing near the banks and dead trees within the channel, particularly near the Hagadorn Bridge. Reach B (between the Farm Lane and Kellogg Bridges) was characterized by the presence of a large meander and undifferentiated sand units (mostly fine sand and silt) near the left bank over much of its length. Previous studies showed that meandering contributes to surface storage by increasing eddies and pools in the reach. In addition, reach B has a weir located near the Library Bridge. The backwaters of this weir extended upstream and created an impoundment resulting in enhanced surface storage. The sand and silt unit hugs the left bank in Reach B and extends all the way into Reach C (between Kellogg and Kalamazoo Bridges) and provides the opportunity for enhanced exchange of solutes between the banks and the main channel compared to reach A [Puzio and Larson, 1982]. Thus reach B had the potential for both surface storage and hyporheic exchange. In reach C, the river comes down following a north-south course and runs into the Mason esker [Leverett and Taylor, 1915], a well-defined ridge of sand and gravel oriented in the south-north direction. The grain size of the sediments decreases radially outward from the centerline of the esker. Since the adjacent regions are mainly loam and other finer material, the gravel acts as a ledge and the river adjusts its gradient producing a meandering channel. Reach C therefore has wide flood-plains, exhibits extensive meandering and is marked by the presence of wetlands and swamps near the edge of the river. Meandering of the channel changed the floodplain alluvium and created high-porosity sand and gravel deposits that provide conditions suitable for hyporheic exchange.

[17] Variations in velocities and channel widths between the reaches can be used to gain insight into how dispersion changes along the river. Comparisons between observed and simulated flows, heads and velocities at the Library Bridge are shown in Figure 6. Reasonable agreement was obtained between model predictions and observations and similar agreement was noted at other stations as well. Although there was a significant variability in the velocities and heads.
as a function of distance, each subreach can be considered a fair approximation of a channel representing uniform hydraulic properties. Examination of the spatial variations in flows and velocities for the first three dye studies showed that there was a significant difference in the average velocities in reaches A and B but velocities in reaches B and C had similar values (Figure 7). For example, for $Q = 16.82 \, \text{m}^3/\text{s}$, the mean velocities for the three reaches were 0.44, 0.75 and 0.71 m/s respectively.

5.3. Evaluation of Transient Storage Model Parameters

While the global optimization algorithm successfully estimated parameter sets for reaches A and B, reach C

Figure 6. Comparison of observed and simulated flows, velocities, and heads at the Library Bridge (Figure 1) during 4–10 April 2002.

Figure 7. Simulated velocities and heads as a function of distance along the river for three different flow rates (corresponding to the first three dye studies).
parameters were found to suffer from singular convergence [Fernald et al., 2001; Runkel, 1998]. The initial set of parameters estimated for reach C had very high values for the dispersion coefficient $D$ and a storage zone size ($A_S$) that was smallest of all the three reaches. This result for reach C was not supported by our observations. The storage zone size for reach C was expected to be the highest in the study reach due to wider floodplains and alluvium storage in addition to the presence of wetlands and meander bends. Our calibrated hydrodynamic model gave similar estimates of mean velocities in reaches B and C; therefore we believe that the high dispersion coefficients and the low $A_S$ values initially estimated for reach C were unreasonable. To illustrate the singular behavior, we performed a Monte Carlo simulation by fixing the parameter $A$ at its optimal value (this parameter was estimated uniquely) and varying the other parameters simultaneously using uniform random sampling. Six thousand model runs were conducted and the results were shown in Figure 8, showing RMSE as a function of normalized values of the parameters. From the slopes, it is clear that the parameters $\alpha$ and $A_S$ were estimated uniquely; however it was not possible to estimate the dispersion coefficient $D$ uniquely for reach C. The slope of the RMSE versus $D$ curve was flat for reach C which was probably an indication that, near the optimal value, $D$ was a redundant parameter. Similar behavior was observed for other data sets as well. This singular convergence behavior of $D$ was also observed by other researchers [Fernald et al., 2001]. To resolve this problem, we set the dispersion coefficient for reach C equal to the value obtained for reach B and this allowed us to successfully estimate the remaining parameters for reach C. In fourth-order mountain streams, [D'Angelo et al., 1993] found that the dispersion coefficient $D$ correlated with velocity. Our calibrated hydrodynamic model gave us estimates of average velocities in the three reaches and showed that reaches B and C had similar velocities for all dye studies. This result seems to support our choice of $D$ for reach C. Estimated TS parameters for all the four dye studies and the three sampling stations are summarized in Table 1, while comparisons between observed and simulated tracer concentrations are shown in Figure 9. We find a good overall agreement between model predictions and observations (RMSE varied from 0.36 to 4.41).

Figure 8. Results of Monte Carlo simulations showing the RMSE as a function of three ($\alpha, D, A_S$) of the four parameters. The main channel cross-sectional area $A$ for this discharge was fixed as it was estimated uniquely. Note that parameters on the X axis were normalized to make visual comparisons easier.
channel features and surficial geology presented in the previous section. The size of the TS zones also increased with discharge, from 1.88 m² for a low flow of 2.49 m³/s to 7.28 m² for Q = 19.06 m³/s. This is in contrast to the results of [Morrice et al., 1997], who found that the size of the TS zone decreased with increasing discharge in a first-order mountain stream. In the RCR, stream cross-sectional area (A) increased with discharge, an observation also made by Morrice et al. [1997]; however, at higher discharges the adjacent low-lying areas near the banks were filled with relatively stagnant water which provided additional surface storage that was not available at low discharges. Results from our ADCP surveys (presented in section 5.4) support this explanation.

The rate of exchange α between the main channel and the storage zones increased with discharge Q. A positive relation between α and Q was also noted by [D’Angelo et al., 1993] who attributed it to an increased availability of solute per unit time. We calculated the TS zone residence times (t5 = (A+Δ)/α) [Harvey et al., 1996]. Residence times ranged from 255 seconds (t5 = 6.26 × 10⁻⁴ s⁻¹ for Q = 16.82 m³/s to 790 seconds (t5 = 1.5 × 10⁻⁴ s⁻¹) for Q = 2.49 m³/s for the RCR. For all dye studies, residence times increased with distance in the downstream direction and generally decreased with discharge Q. We attribute the increase in t5 with downstream distance to the presence of sand and gravel deposits in the last two reaches compared to reach A. Our estimated α values are in the same range as the values reported by Salehinn et al. [2003] for the vegetated reach of an agricultural stream in Sweden (α = 6.1 × 10⁻⁴ s⁻¹). Harvey et al. [2003] also reported similar range of values (α = 4.7 × 10⁻⁴ – 5.6 × 10⁻⁴ s⁻¹) for a heavily vegetated stream in Arizona. The relatively high t5 values determined for reach C are attributed to the alluvium storage and the sediment characteristics (gravel and coarse sand) in this reach. Results from our ADCP surveys (and wavelet decomposition) showed that surface storage in this reach was relatively small indicating that hyporheic exchange was the primary mechanism that contributed to TS in this reach.

The cumulative effect of TS on downstream transport and reach-scale retention of water depends on the parameters A5, α and the flow velocity in the main channel. It is well-known that efforts to interpret TS model parameters often lead to misleading conclusions about the relative importance of TS processes compared to other processes [Runkel, 2002] as existing metrics such as the storage zone residence time do not describe the overall effect of the TS parameters described above. Runkel [2002] proposed the use of a new metric (F_med) which is the fraction (expressed as percent) of the median reach traveltime that is due to TS.

\[ F_{med} = \left[ 1 - e^{-\frac{L_{med}}{Q}} \right] \times \frac{A_s}{A + A_s} \]  

Stream reaches that substantially influence the downstream transport of solute mass due to TS will have higher values of F_med and vice versa. Since reach lengths (L) vary significantly in different studies, a F_med value obtained using a standard reach length of 200 m (F_med^200) was proposed as a metric to facilitate direct comparison with other streams. F_med and F_med^200 values for the Red Cedar River indicate that the importance of TS increases in the downstream direction (Table 1). Comparison of the F_med values with estimates for other streams [Runkel, 2002] showed that values for the RCR were at the lower end of the range and were comparable to those for the Snake River, an acidic and metal-rich mountain stream in Colorado [Bencala et al., 1990]. Although the RCR is a bigger stream in comparison, the potential for hyporheic exchange at this site (due to the consolidated clay layer) was noted by other researchers [Uzarski et al., 2004]. The estimated (A5/Q) values and exchange coefficients (α) in several subreaches are comparable for the two streams.

Dispersion coefficient depends on the transverse velocity variation, which is a function of the flow and channel geometry. D tends to increase between the first and the second sampling stations. The width of the channel between the two locations decreases continuously resulting in a steeper gradient in the transverse velocity profile. However, the width between the second and third sampling stations exhibits an increasing trend resulting in a decrease in the rate of increase in dispersion coefficients. In the absence of TS, the magnitude of the dispersion coefficient is primarily a function of the transverse velocity profile. However, with a significant TS effect, the role of the dispersion coefficient is somewhat similar to the other parameters (i.e., A5/Q and α).

5.4. Results From ADCP Surveys and Wavelet Analyses

Figure 10 shows the observed bathymetry, channel cross sections and the mean velocity fields at the Hagadorn.
Figure 9. Comparison of observed and simulated tracer concentrations for four slug releases conducted during summer 2002. (a) $Q = 2.49 \text{ m}^3/\text{s}$. The comparison is shown for a sampling point at $x = 0.87 \text{ km}$ (Bogue Street Bridge, Figure 1) from the point of release. (b) $Q = 14.41 \text{ m}^3/\text{s}$. The three curves (in Figures 9b–9d) correspond to the three sampling locations, namely, Farm Lane Bridge ($x = 1.40 \text{ km}$), Kellogg Bridge ($x = 3.10 \text{ km}$), and the Kalamazoo Bridge ($x = 5.08 \text{ km}$), respectively. (c) $Q = 16.82 \text{ m}^3/\text{s}$. (d) $Q = 19.06 \text{ m}^3/\text{s}$. 

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and Farm Lane Bridges for two different flow rates, 

\[ \dot{Q} = 5.49 \, \text{m}^3/\text{s} \quad (8 \text{ November 2003}) \quad \text{and} \quad \dot{Q} = 19.89 \, \text{m}^3/\text{s} \quad (19 \text{ March 2006}). \]

On both days, the flow near the Hagadorn Bridge was highly nonuniform and became relatively uniform with distance in the downstream direction. In addition, channel cross section was W shaped at the Hagadorn Bridge as opposed to the U-shaped cross sections (which favor uniform conditions) at the other bridges. As discharge increased, the river became wider and was marked by the presence of relatively stagnant water near the banks. The relative extent of the low-velocity or stagnant zones decreased in the downstream direction as the flow became more uniform between Hagadorn and Farm Lane Bridges. Since the tracer data was used to estimate reach-averaged values for \( A_s/A \), ADCP data collected at multiple stations within a reach can be averaged to compare with the tracer results. If in-channel processes were primarily responsible for storage within a reach, then we expect \( A_s/A \) estimates from ADCP and tracer data to agree. On the other hand, stream reaches dominated by hyporheic exchange are expected to produce widely different estimates of \( A_s/A \) from ADCP and tracer data. The approximations to the original velocity fields based on two-level decomposition allowed us to identify TS locations with the channel (Figure 11). The images marked \( L_1 \) and \( L_2 \) show the first term in equation (6) for level 1 and level 2 decompositions respectively. The relative locations of the dead zones within the channel given by the wavelet decomposition (\( L_2 \) approximation in Figure 11) agreed with our observations of relatively stagnant water during our field work. The ratios of the areas \( (A_s/A) \) calculated based on this decomposition are shown in Table 2 for reach A for two discharge values (2.0 and 19.8 m\(^3\)/s). The average values obtained from the ADCP data were based on four transects and were found to be in good agreement with results from tracer data. Using more transects will likely improve the ADCP estimates but data from other transects (e.g., at the Bogue Street Bridge) showed that conditions were similar to those at the Farm Lane Bridge. Since our primary focus was on ascertaining whether the sizes of the TS zones independently estimated using the ADCP and tracer data were of a comparable magnitude, we believe that the estimates in Table 2 are adequate. In first-order tropical headwater streams differing in channel morphology and hydraulic characteristics, Giucker and Boečhat [2004] compared the sizes of TS zones for different stream morphotypes including straight run, meandering, step-pool, and swamp reaches. They concluded that their \( (A_s/A) \) estimates were lowest for straight run reaches and highest for swamp reaches. In our case, reach A, predominantly a run reach, had the lowest \( (A_s/A) \) for all the discharge values. Since parameter values estimated based on the TS model were comparable to those estimated based on the ADCP data for similar discharge values, we conclude that TS was primarily controlled by in-stream processes in reach A. We could not obtain \( (A_s/A) \) estimates from the ADCP data for reaches B and C as the stream was too shallow to operate our 1200 kHz instrument and obtain good transect data. Since \( (A_s/A) \) values were highest in reach C (which was consistent with our observations, e.g., the presence of swamps, a meandering channel, wide floodplain), we wanted to test the hypothesis that hyporheic exchange was important in this reach. If this was indeed true, then we expect the \( (A_s/A) \) estimates from ADCP data to be relatively small compared to the estimates from tracer data. We were successful in obtaining several good transects at the Kalamazoo Bridge (our last sampling point) on
15 October 2006 ($Q = 3.35$ m$^3$/s). Data from one such transect is shown in Figure 12. The channel was relatively narrow at this station (Figure 2) and showed cross-sectional uniformity in velocity which was an indication that surface storage may be relatively unimportant. Average value obtained from wavelet analysis based on three transects at this site gave ($A_S/A$) = 0.05. Although we do not have estimates from tracer data to compare with this value, our fourth dye study was conducted under similar low-flow conditions ($Q = 2.49$ m$^3$/s). We obtained ($A_S/A$) = 0.12 for reach D between the Hagadorn and Bogue Street Bridges in Table 1. All other dye studies showed that ($A_S/A$) increased in the downstream direction and with discharge. It is therefore highly probable that the ($A_S/A$) value for reach C on 15 October 2006 was significantly higher compared to the number 0.05 estimated from the ADCP data.

[24] An important parameter in assessing the role of storage zones is the Damköhler index ($Da$) calculated from observed data. The parameter $Da$ reflects the relative importance of downstream processes in relation to $TS$ and is computed as the ratio of the time needed for the downstream tracer transport for a certain reach length to the mean tracer residence time in the storage zones

Table 2. Comparison of the Relative Sizes of the Transient Storage Zones Estimated Using Tracer Data and Independently Using the ADCP Data for Reach A

<table>
<thead>
<tr>
<th>$Q$ Tracer, m$^3$/s</th>
<th>$Q$ ADCP, m$^3$/s</th>
<th>$A_S/A$ (Tracer Data)</th>
<th>$A_S/A$ (ADCP)</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>19.06</td>
<td>19.80</td>
<td>0.16</td>
<td>0.16</td>
<td>Farm Lane-1</td>
</tr>
<tr>
<td>19.06</td>
<td>19.80</td>
<td>0.16</td>
<td>0.12</td>
<td>Farm Lane-2</td>
</tr>
<tr>
<td>19.06</td>
<td>19.80</td>
<td>0.16</td>
<td>0.19</td>
<td>Hagadorn-1</td>
</tr>
<tr>
<td>19.06</td>
<td>19.80</td>
<td>0.16</td>
<td>0.21</td>
<td>Hagadorn-2</td>
</tr>
<tr>
<td>19.06</td>
<td>19.80</td>
<td>0.16</td>
<td>0.17</td>
<td>average value</td>
</tr>
<tr>
<td>2.49</td>
<td>2.00</td>
<td>0.12</td>
<td>0.11</td>
<td>Farm Lane-1</td>
</tr>
<tr>
<td>2.49</td>
<td>2.00</td>
<td>0.12</td>
<td>0.14</td>
<td>Farm Lane-2</td>
</tr>
<tr>
<td>2.49</td>
<td>2.00</td>
<td>0.12</td>
<td>0.11</td>
<td>Hagadorn-1</td>
</tr>
<tr>
<td>2.49</td>
<td>2.00</td>
<td>0.12</td>
<td>0.17</td>
<td>Hagadorn-2</td>
</tr>
<tr>
<td>2.49</td>
<td>2.00</td>
<td>0.12</td>
<td>0.12</td>
<td>Hagadorn-3</td>
</tr>
<tr>
<td>2.49</td>
<td>2.00</td>
<td>0.12</td>
<td>0.13</td>
<td>average value*</td>
</tr>
</tbody>
</table>

*Estimated for subreach (reach D).

Figure 11. Multiresolution wavelet approximations for the images shown in Figures 10c and 10d. The velocity data shown in Figure 10 were normalized and replotted. After completing the wavelet analysis, the low-frequency content at wavelet levels 1 and 2 (denoted by $L_1$ and $L_2$) was plotted in the physical space. (a and b) Hagadorn Bridge. (c and d) Farm Lane Bridge.

Figure 12. Observed mean velocity field near the Kalamazoo Street Bridge obtained using a 1200 kHz ADCP on 15 October 2006 ($Q = 3.35$ m$^3$/s).
and Wagner, 2000]. Schmid [2004] analyzed slug release data and concluded that very close or nearly identical results are obtained by the AD model and TS model if $Da < 0.6$ or $Da > 60.0$. In our case, the calculated $Da$ based on the estimated parameters of the TS model ranged between 1.5 and 7.6 for all cases, which confirmed the preferred use of TS model over the AD model.

6. Discussion

[25] By examining the low-frequency contributions (the coarse features) at successive levels in multilevel wavelet decomposition of ADCP data, we were able to identify the relatively stagnant regions in the flow field. The observed velocity fields contained both the mean and the highly oscillatory components of flow. By taking the average of several transects, we were able to quantify the relative magnitudes of surface storage and hyporheic exchange in different test reaches using wavelet analysis. The decomposed states are plotted in the physical space as shown in Figure 11 and image processing was used to estimate the ratio of the two cross-sectional areas in the images. In equation (11), the main channel cross-sectional area $A$ can be computed from either the level 0 (i.e., the original image) or the level 1 approximation as shown in Figure 11 (they produced identical values for our data sets). The high-frequency subbands in the wavelet decomposition (e.g., the horizontal, vertical and diagonal details) were not shown as they did not contain information useful for our present analysis. In a study of transient storage and hyporheic flow along the Willamette River in Oregon, Fernald et al. [2001] indicated that they used a boat-mounted ADCP to measure discharges and main channel cross-sectional areas ($A$) at their sampling locations, although they did not show any comparisons between observed (ADCP) and estimated cross-sectional areas. In the present work, we did not directly compare the cross-sectional areas measured using the ADCP to the $A$ estimated from our TS modeling for the following reason. Depending on the mode of operation and due to the time of delay required to transmit and receive acoustic signals, ADCP data usually have a “blank distance” close to the transducer in which velocity measurements are not available. In addition, there are difficulties in making measurements close to the banks. These limitations become more pronounced in shallow environments. Recent ADCP models specially designed for shallow environments may be more suitable for the type of applications described in this paper. Instead of directly using the cross-sectional areas ($A$) obtained from the ADCP, we decided to focus on the ratio of the areas ($A_s/A$) as errors involved in the approximations of the areas may cancel out when ratios are involved. The assumption here is that the ($A_s/A$) values estimated from ADCP measurements (and wavelet analysis) are representative of the entire cross section including the areas that could not be reached using an ADCP. For the data reported in this paper, we were able to make measurements close to the banks, therefore this assumption is unlikely to affect our results and conclusions but it is not hard to imagine situations where a significant fraction of storage zones are in shallow areas and remain inaccessible to an ADCP. The success of studies aimed at understanding functional relationships between nutrient uptake and storage area depends critically on our ability to separate surface storage from hyporheic exchange [Runkel et al., 2003; Salehin et al., 2003]. This paper presents one approach for achieving this separation. We demonstrated that in one of our test reaches (reach A), the tracer data and the ADCP estimates of ($A_s/A$) were in good agreement for both high and low discharge values. Additional data sets and analyses are required to test the strengths and weaknesses of this approach. As noted by [Shields et al., 2003] and [Dinehart and Burau, 2005], the study of river reaches using ADCPs is hampered by the lack of custom software for data analysis. For this study, we created software for extracting ADCP data, smoothing, correction, visualization, wavelet analysis, image processing etc. The availability of standardized software will make it easier to study river reaches using ADCPs on a routine basis.

7. Conclusions

[26] In this paper, we examined solute transport processes in a Michigan stream and presented results based on our field observations and numerical modeling. Analyses based on the Chatwin parameter and the Damkohler index indicated that TS processes need to be incorporated in the models. Wavelet decompositions of the mean velocity fields based on our ADCP surveys for different stream cross sections were used to separate fast and slow moving regions and were used to estimate the parameter ($A_s/A$) in the TS model. Good agreement was obtained between the observed (ADCP/wavelet-based) and simulated (TS model) ($A_s/A$) values for reach A in which surface storage was expected to be the primary mechanism contributing to TS. We believe that the proposed method of separating surface storage from hyporheic exchange has the potential to better constrain TS models by avoiding physically unrealistic parameters and solutions.

Notation

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$a_1$, $a_2$, $b_1$, $b_2$</td>
<td>scale and translation parameters in wavelet transform.</td>
</tr>
<tr>
<td>$A$</td>
<td>flow zone cross-sectional area ($m^2$).</td>
</tr>
<tr>
<td>$A_s$</td>
<td>cross-sectional area of the storage zone ($m^2$).</td>
</tr>
<tr>
<td>$C$</td>
<td>concentration of tracer in the main channel (ppb).</td>
</tr>
<tr>
<td>$C_s$</td>
<td>concentration of tracer in the storage zone (ppb).</td>
</tr>
<tr>
<td>$C_L$</td>
<td>lateral inflow solute concentration (ppb).</td>
</tr>
<tr>
<td>$Da$</td>
<td>Damkohler index, equal to $[L(1+(A_s/A))/vT]$.</td>
</tr>
<tr>
<td>$g$</td>
<td>acceleration due to gravity ($m^2/s$).</td>
</tr>
<tr>
<td>$h$</td>
<td>water surface elevation (m).</td>
</tr>
<tr>
<td>$H_i$</td>
<td>mean depth of the channel for segment $i$ (m).</td>
</tr>
<tr>
<td>$D$</td>
<td>longitudinal dispersion coefficient ($m^2/s$).</td>
</tr>
<tr>
<td>$i,j$</td>
<td>position indices ($y$ and $z$ directions) in equations (5)--(10).</td>
</tr>
<tr>
<td>$L$</td>
<td>reach length (m).</td>
</tr>
<tr>
<td>$m(t)$</td>
<td>mass of tracer in the dye cloud (kg), equation (1).</td>
</tr>
<tr>
<td>$m$</td>
<td>wavelet scale index.</td>
</tr>
<tr>
<td>$M$</td>
<td>total mass of tracer injected (kg).</td>
</tr>
<tr>
<td>$n$</td>
<td>Manning’s roughness parameter.</td>
</tr>
<tr>
<td>$N$</td>
<td>number of levels in the wavelet decomposition.</td>
</tr>
</tbody>
</table>
\[ Q \] discharge (m\(^3\)/s).
\[ R \] hydraulic radius (m).
\[ R \] rectangular region enclosing the river cross section in the \( y \) and \( z \) directions.
\[ t_s \] transient storage residence time (s).
\[ \Delta t \] time step (s).
\[ v \] velocity, \( Q/A \) (m/s).
\[ W \] wavelet coefficients.
\[ x, y, z \] longitudinal, cross-stream, and vertical coordinates.
\[ i, j \] indices denoting positions in \( y \) and \( z \) coordinate directions.
\[ \alpha \] storage zone exchange rate \((s^{-1})\).
\[ \beta \] the ratio \((A_S/A)\).
\[ \phi, \psi \] scaling and wavelet functions respectively (1D).
\[ \Psi \] wavelet bases for a two-dimensional wavelet.

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**References**


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