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Evaluating the reliability of the stream tracer approach to characterize stream-subsurface water exchange

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Abstract. Stream water was locally recharged into shallow groundwater flow paths that returned to the stream (hyporheic exchange) in St. Kevin Gulch, a Rocky Mountain stream in Colorado contaminated by acid mine drainage. Two approaches were used to characterize hyporheic exchange: sub-reach-scale measurement of hydraulic heads and hydraulic conductivity to compute streambed fluxes (hydrometric approach) and reach-scale modeling of in-stream solute tracer injections to determine characteristic length and timescales of exchange with storage zones (stream tracer approach). Subsurface data were the standard of comparison used to evaluate the reliability of the stream tracer approach to characterize hyporheic exchange. The reach-averaged hyporheic exchange flux (1.5 mL s^{-1} m^{-1}), determined by hydrometric methods, was largest when stream base flow was low (10 L s^{-1}); hyporheic exchange persisted when base flow was 10-fold higher, decreasing by approximately 30%. Reliability of the stream tracer approach to detect hyporheic exchange was assessed using first-order uncertainty analysis that considered model parameter sensitivity. The stream tracer approach did not reliably characterize hyporheic exchange at high base flow: the model was apparently more sensitive to exchange with surface water storage zones than with the hyporheic zone. At low base flow the stream tracer approach reliably characterized exchange between the stream and gravel streambed (timescale of hours) but was relatively insensitive to slower exchange with deeper alluvium (timescale of tens of hours) that was detected by subsurface measurements. The stream tracer approach was therefore not equally sensitive to all timescales of hyporheic exchange. We conclude that while the stream tracer approach is an efficient means to characterize surface-subsurface exchange, future studies will need to more routinely consider decreasing sensitivities of tracer methods at higher base flow and a potential bias toward characterizing only a fast component of hyporheic exchange. Stream tracer models with multiple rate constants to consider both fast exchange with streambed gravel and slower exchange with deeper alluvium appear to be warranted.

Introduction

In drainage basins with a shallow water table the flow of surface water in channels is usually closely connected with groundwater flow. The factors that affect hydrologic exchange between channels and groundwater include aquifer geometry, hydraulic properties, and water balance [*Freeze and Witherspoon*, 1967, 1968; *Winter*, 1995]; channel slope, width, sinuosity, and penetration in the aquifer [*Sharp*, 1977; *Larkin and Sharp*, 1992]; and temporal fluctuations in water table heights and channel stage [*Pinder and Sauer*, 1971]. The effect of groundwater and surface water mixing on transport of solutes is increasingly being studied, including research on dissolved salts [*Konikow and Bredehoeft*, 1974], nutrients [*Newbold et al.*, 1983; *Triska et al.*, 1989; *Valett et al.*, 1994], oxygen [*McMahon et al.*, 1995], metals [*Bencala et al.*, 1984; *Benner et al.*, 1995], radionuclides [*Cerling et al.*, 1990], and organic contaminants [*Squillace et al.*, 1993].

The scale of exchange flows between channels and subsurface flow systems can be large or small in extent. Individual flow paths of exchange range in scale from hundreds of meters, in which transport occurs on a timescale of years, to centimeter-long flow paths, in which transport occurs on a timescale of minutes. Interactions are driven at small scales by steady flow of surface water over roughness features such as sand waves or pools and riffles. The resulting uneven pressure distributions on the channel bed cause surface water to flow into and out of the bed [*Thibodeaux and Boyle*, 1987; *Harvey and Bencala*, 1993]. We refer to small-scale (centimeter to meter) exchanges of water between channels and the subsurface as "hyporheic exchange" (Figure 1a) in order to emphasize the relation to the hyporheic zone identified by stream ecologists [*Hynes*, 1974; *Triska et al.*, 1989].

The delineating characteristic of the hyporheic zone is the recharge of channel water to the subsurface and mixing with groundwater that has not yet reached the channel. Since flow paths are short, a molecule of channel water may be exchanged between the channel and hyporheic zone many times. Hyporheic exchange keeps channel water in close contact with sediment, which may enhance solute transformations that reduce downstream transport. The usual means to investigate hyporheic exchange is to measure indicators of exchange such as

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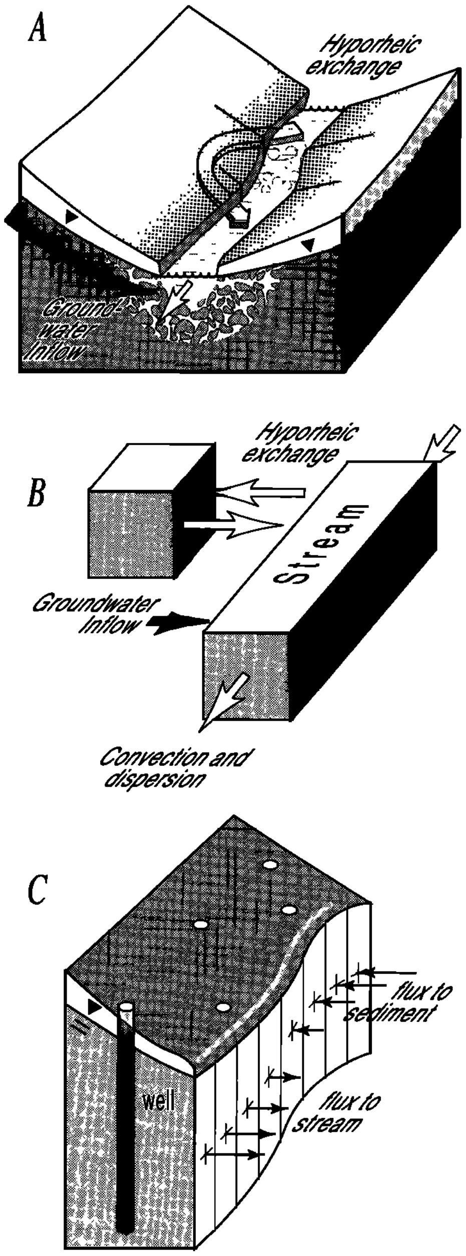

Figure 1. Stream-subsurface hydrologic exchange. (a) Schematic physical system indicating large-scale inflow from groundwater and small-scale exchange between stream and hyporheic zone. Two water balance analysis approaches were compared. (b) Reach-scale modeling of stream tracer injections to compute exchange fluxes. (c) Hydrometric approach using sub-reach-scale measurements of hydraulic heads and tracer movement to compute exchange fluxes.

temperature or hydraulic gradients [*White et al.*, 1987; *Valett et al.*, 1994] or to combine those measurements with observations of the movement of solute tracers in the subsurface [*Harvey and Bencala*, 1993; *Wondzell and Swanson*, 1996]. While those approaches are required to understand the fundamental processes, scaling-up those flux estimates is essential if the cumulative effects on drainage basin water quality are to be understood. A significant challenge is to link physical measurements of the hyporheic zone with characteristics determined using reach-scale injections of solute tracer.

Modeling of stream tracer experiments potentially provides a means to determine average characteristics of hyporheic exchange at scales of hundreds of meters. In such studies tracer-labeled stream water that enters hyporheic flow paths returns to the channel within the experimental reach but is delayed in downstream transport, producing a "signal" of hyporheic exchange in the tracer dynamics measured at downstream monitoring locations. Stream tracer experiments are usually simulated by adjusting reach-scale parameters of one-dimensional stream transport models (i.e., "dead zone" or "transient storage models" that include exchange with hydrologic storage zones) to achieve a "best fit" to measured tracer concentrations in the stream. Fitting may be done either by manual adjustment of parameters to match measured stream concentrations [*Bencala and Walters*, 1983; *Stream Solute Workshop*, 1990] or by using statistical approaches to select parameters [*Wagner and Gorelick*, 1986; *Young and Wallis*, 1993; *Hart*, 1995].

The question addressed in this paper was, To what degree do the hydrologic storage parameters determined by modeling actually represent exchange between the channel and hyporheic zones? The answer is uncertain because stream tracer experiments are also sensitive to exchange between the active channel and stagnant pools or recirculating eddies in surface flow [*Fischer et al.*, 1979]. Model storage parameters determined in a number of mountain stream settings vary widely, over several orders of magnitude [*Broshears et al.*, 1993]. Repeat investigations during periods of low and high base flow exhibit wide variability in parameter values [*Legrand-Marq and Laudelot*, 1985; *D'Angelo et al.*, 1993; *Morrice et al.*, 1996], demonstrating the importance of changes in flow conditions in affecting retention of solute tracers. Bivariate relationships indicate that model storage parameters show some relation to channel friction and other characteristics of flow in channels [*Thackston and Schnelle*, 1970; *Bencala and Walters*, 1983; *D'Angelo et al.*, 1993; *Lancaster and Hildrew*, 1993], but those statistical relationships cannot in themselves identify the relative importance of surface and subsurface storage mechanisms. Only recently has a concerted effort begun to directly compare stream tracer results with detailed observations of surface-subsurface water exchange [*Castro and Hornberger*, 1991; *Harvey and Bencala*, 1993; *Morrice et al.*, 1996]. The steps taken in the present study were as follows: (1) Conduct stream tracer experiments at St. Kevin Gulch during conditions of both low and high base flow and simultaneously measure hydraulic gradients and tracer movement in the subsurface (Figure 1a), (2) use the transient storage model and statistical methods to identify storage characteristics (Figure 1b), (3) compare detailed subsurface measurements with model storage characteristics (Figures 1b and 1c), and (4) assess the reliability of the stream tracer approach to characterize hyporheic exchange under variable flow conditions.

Study Site and Field Investigations

St. Kevin Gulch is a headwater catchment on the east side of the Continental Divide in the Rocky Mountains near Leadville, Colorado (Figure 2). The study area is in the lower part of the St. Kevin Gulch catchment. The experimental reach encompasses 50 m of third-order channel with an average slope of 0.07. Streamflow in this part of the catchment is sustained by inflow of groundwater from permanently saturated areas of the lower hillslope. About 300 m farther downstream the valley widens and the alluvium deepens; beyond that point the stream loses water in middle to late summer [*Zellweger*, 1994].

Streambed slope in the study reach varies on the scale of

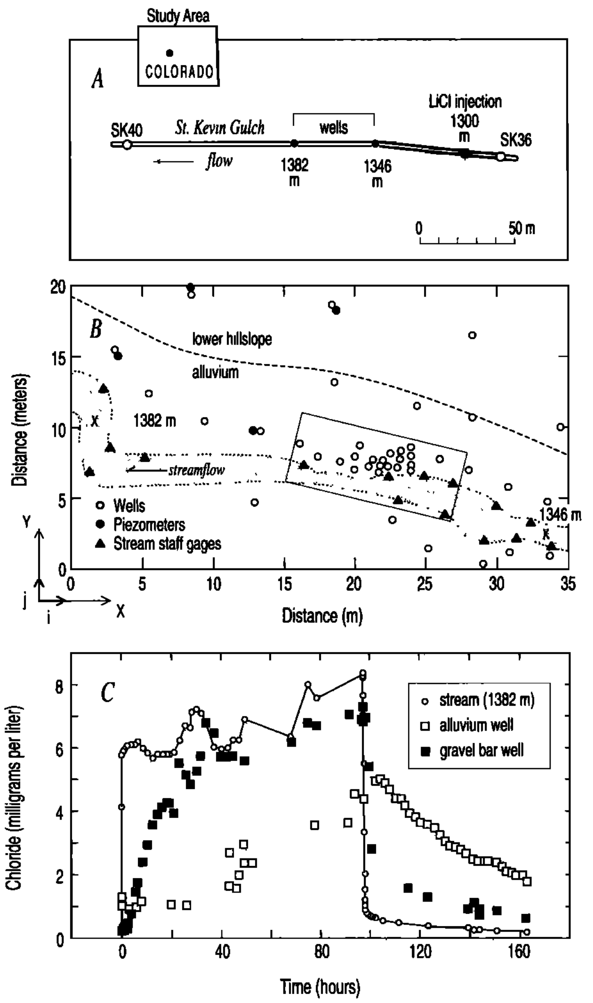

Figure 2. Experimental study reach, St. Kevin Gulch, Colorado. (a) Lithium chloride solute tracer was injected upstream of reach instrumented with wells. (b) Instrument location map. The rectangular inset box shows limits of the detailed measurements of stream tracer movement into hyporheic flow paths in a 12-m subreach. (c) Concentration of the stream water tracer in the stream at 1382 m and in one alluvium and one gravel bar well.

meters from much less than 1% to greater than 20%. Channel sediment is well-sorted sand and gravel, distributed in patches that range from fine sand to coarse sand to gravel; in the steeper channel units cobbles and small boulders are exposed. Alluvial sediment that surrounds the channel sediment is a poorly sorted sediment composed of mixed fine and coarse sand, gravel, and cobbles. Alluvial sediment is approximately 2 m thick, and it extends approximately 5 m on either side of the stream. Farther away from the stream, on the forested lower hillslope, the soil profile grades from organic horizon at the surface to a sandy loam and then to a clay loam at depth. Soils are underlain by a schistose and gneissic bedrock at a depth of several meters [*Singewald*, 1955].

Methods

Approximately 60 wells, piezometers, and staff gauges were emplaced along a 36-m study reach of stream (Figure 2). In August 1990 and in June 1991 lithium chloride (LiCl) was injected into the stream at a steady rate for a period of 4 days. Repeat experiments provided approximately a 10-fold contrast between low and high base flow in the stream (10 compared with 120 L s^{-1}). During tracer injections stream water samples were collected both at endpoints of the experimental reach and from a subset of wells along the reach (Figure 2b). Hydraulic heads were measured in all wells and at all staff gauges during the injections.

Subsurface Measurements of Hyporheic Exchange

Following the water balance approach outlined by *Harvey and Bencala* [1993], closely spaced hydraulic head measurements were used to compute the reach-averaged streambed flux and to partition that flux into its two components, stream-hyporheic exchange and stream-groundwater exchange. Reach-scale water balance calculations were supplemented by measuring chemistry in streamside wells along a 12-m subreach (rectangular inset in Figure 2b). Using standard mixing models adapted for use in stream and river studies [e.g., *Triska et al.*, 1989; *Bourg and Bertin*, 1993], the percent stream water at each well was computed (assuming steady state transport) using measurements of the distribution of nonreactive solute tracers in the stream and subsurface. Timescales of stream water movement into hyporheic flow paths were also determined using the procedures outlined by *Harvey and Bencala* [1993] and *Triska et al.* [1993]. The travel time needed for stream water to reach wells was determined by observing the arrival of the chloride tracer injected in the stream at wells. The 14 wells in the 12-m subreach were categorized as being either representative of well-sorted gravel bar deposits adjacent to the channel (the six gravel bar wells were 0.3 m or less from the channel) or representative of more poorly sorted alluvium to the sides and beneath the stream channel (the eight alluvium wells were 0.3 to 1.7 m from the channel).

Stream Tracer Experimentation and Modeling

One-dimensional models of advection and dispersion in natural channels are often extended to include a term for coupling the active channel with stagnant or slowly moving zones of flow [*Hays et al.*, 1966; *Thackston and Schnelle*, 1970; *Valentine and Wood*, 1977; *Bencala and Walters*, 1983]. The extended models (usually referred to either as "dead zone" or "transient storage models") were originally formulated in order to improve simulations of the "early time," nonequilibrium phase of dispersion that results from incomplete vertical or transverse mixing at stream sides or on the channel bottom [*Fischer et al.*, 1979]. Storage of solute is simulated as a mass transfer between the channel and a set of decoupled storage reservoirs (situated parallel to the stream) in which mixing is complete and instantaneous (Figure 1b). Although not originally envisioned as a model of stream-subsurface water exchange, the mass transfer formulation is a flexible approach which could represent an interaction between the channel and subsurface. In this paper we use the following familiar formulation of the stream tracer model equations:

$$\frac{\partial C}{\partial t} = -\frac{Q}{A}\frac{\partial C}{\partial x} + \frac{1}{A}\frac{\partial}{\partial x}\left(AD\frac{\partial C}{\partial x}\right) + \frac{q_L}{A}(C_L - C) + \alpha(C_S - C) \tag{1}$$

$$\frac{\partial C_S}{\partial t} = \alpha\frac{A}{A_S}(C - C_S) \tag{2}$$

where t and x are time and direction along the stream; C, C_S, and C_L are concentrations in the stream, storage zones, and

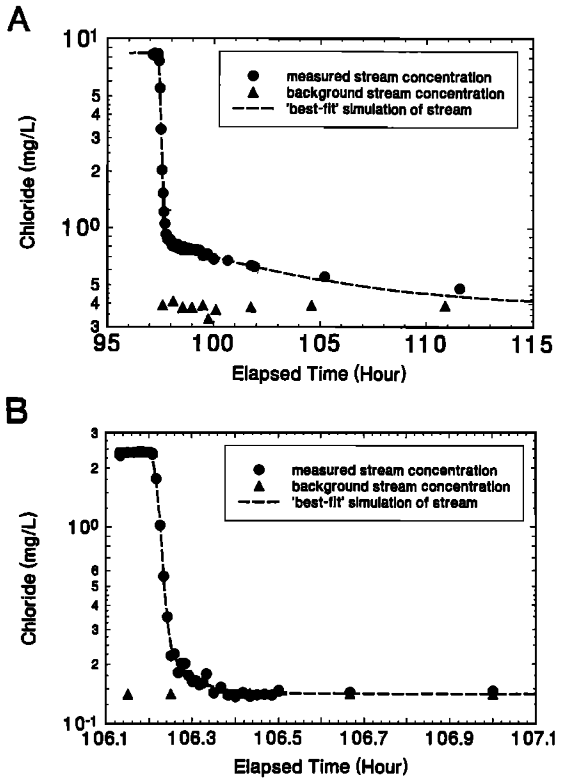

Figure 3. Stream tracer modeling results at St. Kevin Gulch: (a) low base flow and (b) high base flow.

groundwater, respectively $[M/L^3]$; Q is the in-stream volumetric flow rate $[L^3/T]$; q_L is groundwater inflow $[L^3/T/L]$; D is the longitudinal dispersion coefficient in the stream $[L^2/T]$; A and A_S are the stream and storage zone cross-sectional areas, respectively $[L^2]$; and α is the storage zone exchange coefficient $[1/T]$. We solved the model equations numerically by implementing the improved solution technique developed by *Runkel and Chapra* [1993].

Tracer experiments were conducted in August 1990 (period of low base flow) and in June 1991 (period of high base flow). During the 4-day injections of the chloride tracer, concentrations were measured at the upstream (1329 m) and downstream endpoints (1382 m) of a study reach that overlapped the 36-m experimental reach with wells. After cutoff of the tracer injection, tracer concentrations initially decreased rapidly in the stream, followed by a longer period in which tracer concentrations remained elevated above the background concentration (Figure 3). For each experiment streamflow discharge (Q) was determined by the dilution gauging method [*Kilpatrick and Cobb*, 1985]. Groundwater inflow (q_L) was estimated as the difference in streamflow at reach endpoints, divided by the reach length. The other parameters of the model (A, D, α, and A_S) were determined by inverse methods using the nonlinear, least squares regression approach described by *Wagner and Gorelick* [1986].

Assumptions Underlying the Use of the Stream Tracer Approach to Simulate Hyporheic Exchange

Use of the stream tracer approach to estimate hyporheic fluxes has the advantage of efficiency, because the hyporheic exchange flux can potentially be characterized at scales that are relevant to whole drainage basin studies, that is, at scales of hundreds of meters to kilometers where the expense of detailed subsurface measurements would be prohibitive. The reliability of the stream tracer approach to determine exchange with the subsurface is uncertain, because the method is also sensitive to exchange between the active channel and slowly moving surface water in pools and eddies. The purpose of this section is twofold: (1) to show the relation of familiar mass transfer parameters of the stream tracer model to the stream-hyporheic-groundwater mass balance presented by *Harvey and Bencala* [1993] and (2) to outline the assumptions that are needed to quantify hyporheic exchange using the stream tracer approach.

The downstream change in streamflow in a channel without tributaries that is closely connected with shallow groundwater flow is

$$dQ/dx = q_L^{\text{in}} + q_S^{\text{in}} - q_L^{\text{out}} - q_S^{\text{out}}, \tag{3}$$

where Q is streamflow discharge; x is downstream direction; and terms on the right-hand side of the equation are water exchange fluxes across the streambed; q_L^{in} is the reach-averaged groundwater flux into the stream; q_L^{out} is the reach-averaged stream water flux into groundwater; and q_S^{in} and q_S^{out} are the reach-averaged fluxes of stream water out of or into hyporheic flow paths, respectively [*Harvey and Bencala*, 1993].

Hyporheic flow paths are distinguished conceptually from groundwater flow paths by flow path length and by water source. Hyporheic flow paths are short, concentric-shaped flow paths that both enter and return from the subsurface within the stream reach of interest (Figure 1a). In contrast, groundwater flow paths are much longer flow paths that only leave or enter the channel once in the stream reach of interest. In practice, streambed fluxes are usually estimated by computing the difference in streamflow at upstream and downstream ends of channel reaches. Referred to as "seepage runs," these calculations estimate the net groundwater flux across the streambed ($q_L^{\text{in}} - q_L^{\text{out}}$). Hyporheic exchange fluxes cannot be estimated from seepage run data because the reach-averaged water flux into hyporheic flow paths is balanced (at hydrologic steady state) by return fluxes to the channel. Hyporheic exchange fluxes must therefore be estimated by a different means, either by a Darcian approach using hydraulic head and hydraulic conductivity estimates or by using the stream tracer approach. Darcian flux estimates suffer from the usual problems of highly uncertain estimates of hydraulic conductivity. Also, acquiring enough head measurements at the meter scale to compute reach-averaged fluxes at reach scales (hundreds of meters) will usually be prohibitive. Hence there is a growing interest in using the stream tracer approach to characterize hyporheic exchange.

The governing equations for solute transport in the stream and hyporheic zone include (3) and the following two equations,

$$A\frac{\partial C}{\partial t} = -\left(Q\frac{\partial C}{\partial x} + C\frac{\partial Q}{\partial x}\right) + \frac{\partial}{\partial x}\left(AD\frac{\partial C}{\partial x}\right) + C_L q_L^{\text{in}} - Cq_L^{\text{out}} + C_S q_S^{\text{in}} - Cq_S^{\text{out}} \tag{4}$$

$$A_S\frac{\partial C_S}{\partial t} = Cq_S^{\text{out}} - C_S q_S^{\text{in}} \tag{5}$$

where all variables are defined previously with (1)–(3). The new system of three equations ((3)–(5)) is simplified by substituting the right-hand side of (3) to replace the term dQ/dx in (4). Water fluxes q_S^{in} and q_S^{out} are replaced with the single variable q_S to reflect the equivalence of those fluxes at hydrologic steady state. Dividing (4) and (5) through by the respective cross-sectional areas yields the final governing equations for water and solute balance in a connected stream-hyporheic-shallow groundwater system:

$$\frac{\partial C}{\partial t} = -\frac{Q}{A}\frac{\partial C}{\partial x} + \frac{1}{A}\frac{\partial}{\partial x}\left(AD\frac{\partial C}{\partial x}\right) + \frac{q_L^{\text{in}}}{A}(C_L - C) + \frac{q_S}{A}(C_S - C) \qquad (6)$$

$$\frac{\partial C_S}{\partial t} = \frac{q_S}{A_S}(C - C_S) \qquad (7)$$

Comparison of the above equations with (1) and (2) shows the relation between the mass transfer and advective formulations of the exchange flux. The hyporheic exchange flux, q_S, is equal to the product of the mass transfer coefficient, α, and the stream cross-sectional area, A:

$$q_S = \alpha A. \qquad (8)$$

The fluid residence time in the hyporheic zone is

$$t_S = A_S/q_S. \qquad (9)$$

The average distance traveled in the channel by a water molecule before entering a storage zone is

$$L_S = u/\alpha. \qquad (10)$$

where L_S is the characteristic channel length for exchange with the hyporheic zone, u is the streamflow velocity, and α is the exchange coefficient.

There are three primary assumptions associated with using the stream tracer approach to characterize hyporheic exchange. First, the stream tracer approach assumes that stream parallel transport in hyporheic flow paths is negligible, an assumption that is valid if hyporheic flow paths are much shorter than the stream reach. Second, solute holding times in the hyporheic zone are assumed to be distributed exponentially, which is equivalent to assuming that the bulk response of all hyporheic flow paths can be modeled as a simple, first-order, mass transfer between channel and well-mixed reservoir [*Levenspiel*, 1972]. This assumption is violated if multiple storage reservoirs with distinctly different first-order rate contants are present. Third, the model assumes that exchange parameters uniquely characterize hyporheic exchange rather than mixing between the central channel and surface water storage zones. Surface water storage processes are assumed to be accounted for by the longitudinal dispersion coefficient. This third assumption is possibly the most critical assumption, because there is no way to guarantee that a storage zone model uniquely identifies subsurface storage processes.

Useful critiera do exist to estimate the channel length required for mixing in the stream to reach the equilibrium phase, after which surface water storage processes can adequately be accounted for by the longitudinal dispersion term of (1). *Rutherford* [1994] calculated the distance required to establish equilibrium mixing for a number of stream tracer studies. The distance to establish equilibrium mixing is related to the velocity of the stream and the timescale of transverse mixing in the channel:

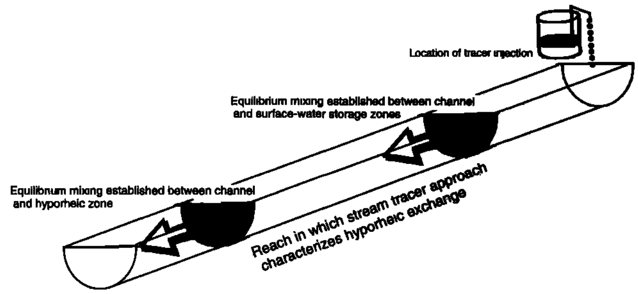

Figure 4. Schematic stream tracer experiment to characterize hyporheic exchange. Upstream and downstream tracer measurement points are located where equilibrium mixing has been established between channel and surface water storage zones and channel and hyporheic zones, respectively.

$$L_e = \beta(ub^2/k_z), \qquad (11)$$

where L_e is the characteristic distance for equilibrium mixing in the channel, u is streamflow velocity, b is channel width, k_z is the transverse dispersion coefficient, and β is a proportionality coefficient. Using relatively low values of the transverse dispersion coefficient (e.g., k_z often is estimated as $0.23du^*$ where d is channel depth and u^* is the shear velocity) *Rutherford* [1994] reports values of β in the range 1 to 10 for rough channels.

Figure 4 illustrates a tracer experiment that is intended to estimate hyporheic exchange parameters. Assuming that the timescale of hyporheic exchange is longer than that for surface water storage processes, stream tracer experiments are more likely to uniquely characterize hyporheic exchange if the upstream sampling point is located a distance from the injection where equilibrium mixing has been established with surface water storage zones. The location of the downstream sampling endpoint is also important; if located too far downstream, exchange with hyporheic zones will also reach the equilibrium phase, causing a shift in shape of the breakthrough curve that can be modeled equally well by adjusting the longitudinal dispersion coefficient or by adjusting the exchange parameters. This problem leads to difficulties with parameter identification because exchange parameters may become nonidentifiable; that is, after equilibrium mixing between channel and hyporheic zone is established it is possible that no unique combination of the exchange parameters and the dispersion coefficient will minimize the difference between measured and modeled tracer concentrations. Finally, sampling a reach that is as long as possible (up to a maximum length bounded by points where equilibrium mixing has been established with surface water storage zones and the hyporheic zone, respectively) is important because more tracer interchange occurs in longer reaches, which increases tracer concentration tailing and model sensitivity to hyporheic exchange and decreases uncertainty of exchange parameter estimates [*Wagner and Harvey*, 1994].

Results

Subsurface Results

Subsurface measurements indicated that hyporheic exchange accounted for between 40 and 80% of the total streambed water flux, depending on the magnitude of stream base flow. Hyporheic exchange was greatest at low base flow, 1.5 ×

Table 1. Surface-Subsurface Water Exchange Fluxes at St. Kevin Gulch, Colorado, Determined by the Hydrometric Approach

Flux	Base Flow: Low	Base Flow: High
Streambed Flux, mL s^{-1} m^{-1}		
Groundwater, q_L^{in}	0.3	1.6
Hyporheic, q_S^{in}	1.5	1.1
Total, $q_L^{in} + q_S^{in}$	1.8	2.7
Normalized Streambed Flux (1)		
Hyporheic, $q_S^{in}/(q_L^{in} + q_S^{in})$	0.8	0.4

10^{-6} m^3 s^{-1} m^{-1}, or fivefold larger than the net groundwater flux (0.3×10^{-6} m^3 s^{-1} m^{-1}) at low base flow (Table 1). Net groundwater inflow to the stream increased by a factor of 5 at high base flow, accompanied by a 30% decrease in hyporheic exchange relative to low base flow (Table 1). The slight decrease in hyporheic exchange at high base flow most likely resulted from increased groundwater flow toward the stream, which increased the resistance to stream water recharge into hyporheic flow paths.

Our previous work at St. Kevin Gulch showed that horizontal subsurface flow was perpendicular to water table contours [*Harvey and Bencala*, 1993]. In the present study we used water table contours to map individual hyporheic flow paths that ranged in length from centimeters to meters in the 12-m subreach. At high base flow the length and distance of penetration of individual hyporheic flow paths was reduced somewhat, and the percent stream water composition in hyporheic flow paths was less compared to low base flow (Figure 5), a finding consistent with the slight reduction of hyporheic exchange at higher base flow determined by the water balance (Table 1). Data on solute tracer arrival at wells indicated that timescales of hyporheic exchange ranged between minutes and tens of hours (Table 2). Hydrologic exchange between stream water and well-sorted gravel sediment directly adjacent to the stream was rapid (hours) relative to exchange with the more poorly sorted alluvial sediment that surrounds the active channel and gravel bar deposits (tens of hours). The average travel time for stream water to reach gravel bar wells was approximately 6 hours, compared to an average travel time of 84 hours for it to reach alluvium wells (Table 2).

Table 2. Nominal Travel Times of Stream Tracer to Reach Gravel Bar and Alluvium Wells, St. Kevin Gulch, Colorado

Well	Distance to Well, m	Travel Time, hours
Gravel Bar Well Series		
R1	0.10	0.3
R6	0.07	5.0
R12	0.07	0.1
R17	0.27	13.0
R16	0.13	3.5
R25	0.13	11.0
Average	···	6.0
Alluvium Well Series		
R4	1.33	69.9
R3	0.83	77.2
R2	0.43	89.9
R8	1.13	77.3
R15	1.00	80.1
R13	0.27	55.7
R23	1.67	111.
R22	0.73	111.
Average	···	84.1

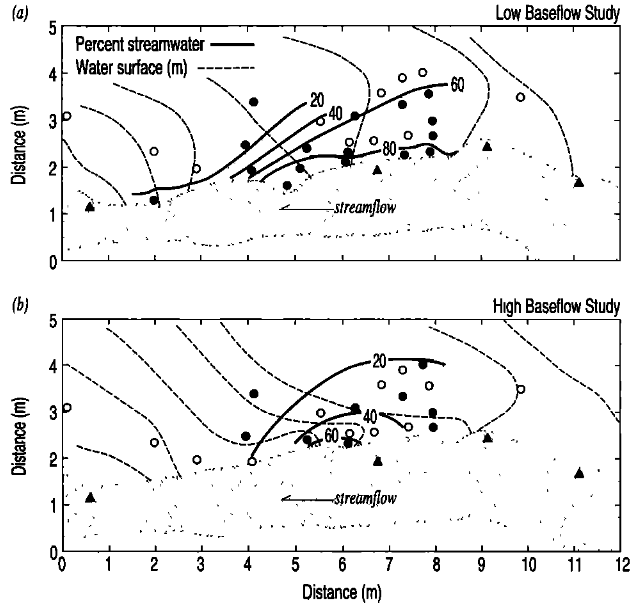

Figure 5. Map of hydraulic head equipotentials and percent stream water in hyporheic zone of a 12-m subreach at St. Kevin Gulch: (a) low base flow and (b) high base flow. Position of 12-m subreach within entire experimental reach is shown in Figure 2b. Hydraulic head was measured in all wells (circles); closed circles represent wells where tracer concentration was also measured.

Stream Tracer Results

Stream tracer modeling provided simulations that were a good match to measured tracer concentrations in the stream. Generally, good matches are indicated during experimental periods of both low base flow and high base flow by visual inspection of plots that compare model results and measurements (Figure 3) and by parameter uncertainty estimates that were mostly below 20% (Table 3). At low base flow the cross-

Table 3. Best-Fit Parameters and Uncertainties for Stream Tracer Model Simulations, St. Kevin Gulch, Colorado

Parameter	Base Flow: Low	Base Flow: High
α, s^{-1}	0.82E−04 (1.8)*	0.4E−03 (22.3)
A_S, m^2	0.21 (7.3)	0.025 (9.2)
D, m^2 s^{-1}	0.25 (6.4)	0.36 (17.4)
A, m^2	0.115 (0.8)	0.36 (1.3)
Q, m^3 s^{-1}	7.5E−03	9.6E−02
q_L, m^3 s^{-1} m^{-1}	0.5E−06	3.7E−06
t_S,† s	2.20E+04	1.73E+02
L_S,‡ m	800	670

Read, for example, 0.82E−04 as 0.82×10^{-4}.
*Percent uncertainty, that is, standard deviation of parameter estimate divided by best-fit value of parameter multiplied by 100.
†Storage zone residence time, $t_S = A_S/(\alpha A)$.
‡Characteristic channel length for exchange, $L_s = u/\alpha$.

sectional area of storage was twice the size of the stream cross-sectional area, and the residence time of storage was approximately 6 hours (Table 3). At high base flow results differed considerably. The cross-sectional area of the storage zone was reduced by an order of magnitude, the exchange coefficient increased by a factor of 5, and the fluid residence time in the storage zone decreased by more than 2 orders of magnitude (Table 3). At high base flow storage processes were apparently confined to a much smaller area with much shorter retention times (Table 3). Parameter uncertainties were higher at high base flow, with uncertainties for the longitudinal dispersion coefficient (17.4%) and the exchange coefficient (22.3%) approaching unacceptable levels.

Comparison of Stream Tracer Modeling Parameters With Subsurface Measurements

At low base flow the stream tracer approach was apparently sensitive to hyporheic exchange, as evidenced by a best-fit storage zone cross-sectional area that was twice as large as the best-fit stream cross-sectional area (Table 3). The best-fit storage zone residence time of approximately 6 hours at low base flow was consistent with measurements of travel time for the tracer to reach gravel bar wells (Table 2). At high base flow the stream tracer approach and hydrometric approaches were not consistent in their description of hyporheic exchange. Whereas the stream tracer approach indicated orders-of-magnitude decreases in storage zone area and fluid residence time at higher base flow (Table 3), subsurface data suggested only modest reductions in hyporheic zone dimensions (Figure 5) and a 30% reduction in the computed hyporheic flux (Table 1) at high base flow. Higher parameter uncertainties for the stream tracer approach at high base flow led us to investigate the possibility that the stream tracer approach was more sensitive to hyporheic exchange at low base flow than at high base flow.

Sensitivity of the Stream Tracer Model to Subsurface Storage Processes

Sensitivity analysis was used to assess the reliability of the stream tracer approach to characterize hyporheic exchange at low and high base flow. Sensitivity is the partial derivative of modeled stream tracer concentration with respect to a change in the value of a parameter,

$$S_{ij} = \partial c_i / \partial p_j \tag{12}$$

where S_{ij} is the sensitivity of stream tracer concentration at time I to the jth parameter, c_i is the concentration at time I, and p_j is the best-fit value of the jth parameter. The sensitivity analysis used here followed general procedures used previously in solute transport modeling [*Knopman and Voss*, 1987; *Sun and Yeh*, 1990]. For this study, sensitivities were determined with respect to the two exchange parameters of the stream tracer model, A_S; the cross-sectional area of the modeled storage zones; and α, the exchange coefficient. We calculated normalized sensitivities,

$$S^*_{ij} = \frac{p_j}{\sigma_i} S_{ij} \tag{13}$$

where S^*_{ij} is the normalized sensitivity, and σ_i is the estimated standard deviation for the concentration observation at time I.

A comparison of normalized sensitivities for the two exchange parameters indicated greater sensitivity of stream tracer concentrations to storage processes at low base flow (Figure 6). At their maximum the normalized sensitivities with respect to A_S and α were approximately twofold and fivefold larger for the low base flow simulation compared to high base flow. The lower sensitivity to hyporheic exchange at high base flow probably resulted from several factors.

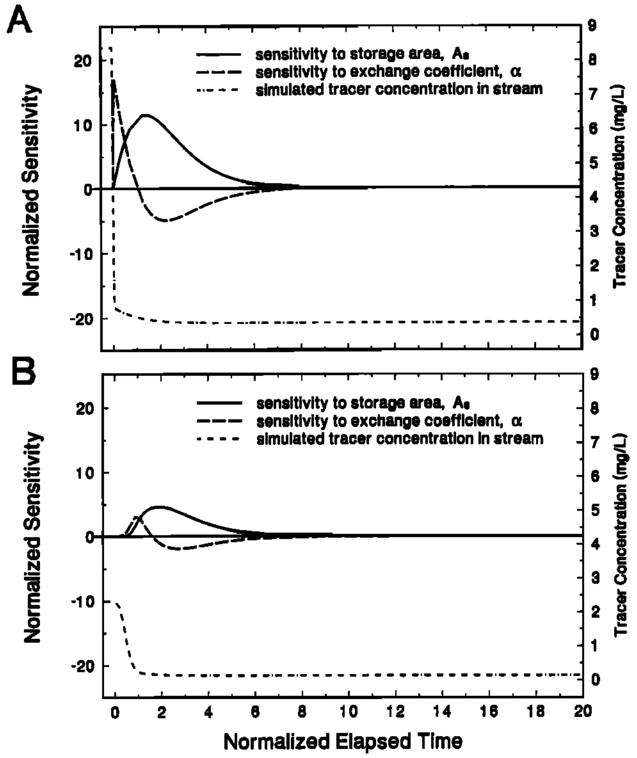

Figure 6. Sensitivity of modeled stream tracer concentrations at St. Kevin Gulch to storage area and exchange parameters: (a) low base flow and (b) high base flow. Time was normalized by subtracting cutoff time of stream tracer injection and dividing by the best-fit storage zone residence time.

One important factor affecting sensitivities was the larger cross-sectional area of the stream and higher streamflow velocities at high base flow, which meant that a much smaller proportion of the injected tracer was likely to enter the hyporheic zone in the experimental reach. Less interaction of tracer with the hyporheic zone at high base flow limited the effect that hyporheic exchange could have in producing the characteristic concentration "tail" that is indicative of storage (Figure 3). Another important factor affecting sensitivity was the lower plateau concentration of solute tracer in the stream at high base flow (approximately 2 mg L^{-1} compared to 8 mg L^{-1} at low base flow) which reduced the overall magnitude of the tracer signal and decreased sensitivity to storage processes.

We have no completely objective means to assess whether the stream tracer approach could uniquely distinguish surface and subsurface storage processes in our experiments, although simple calculations using (11) indicated that the equilibrium phase for mixing in the stream probably was not achieved in either of our experiments. Those calculations suggest that mixing in the stream probably could not be accounted for by adjusting the longitudinal dispersion coefficient and that some overlap of surface and subsurface storage was expected in the tracer signal. On the basis of measured timescales for storage at low base flow (hours from stream tracer modeling compared to minutes in surface water storage zones observed using fluoroscein dye), there is little doubt that subsurface storage processes dominated the stream tracer signal at low base flow. We suspect that the tracer experiment at high base flow was much more sensitive to surface water storage processes, because hyporheic flow paths were affected little by the change in stream stage (Figure 5), while surface water storage zones appeared to be larger at high base flow (i.e., stream stage increased significantly but not enough to swamp boulders and pool-riffle topography that create storage zones in surface flow). Considering all factors, we suspect that increased surface water storage processes and decreased sensitivity to subsurface exchange at higher base flow were the most important reasons that explain the inability to detect hyporheic exchange at high base flow at St. Kevin Gulch.

Even at low base flow the stream tracer approach could not detect all timescales of stream-hyporheic exchange fluxes at St. Kevin Gulch. The model exhibited maximum sensitivity to exchange processes that occurred quickly following the cutoff of the stream tracer injection, that is, within a window of time ranging between 0.2 and 2 best-fit storage zone residence times (Figure 6). The range of time where sensitivity was adequate to detect storage was similar to the range of travel times for stream water to reach gravel bar wells, that is, from 0.02 to 2 best-fit storage zone residence times (Figure 7). In contrast, travel times to deeper alluvium wells ranged between 9 and 18 best-fit storage zone residence times (Figure 7), which were much longer than the period when the stream tracer simulation was sensitive to exchange processes (Figure 6). Consequently, even under optimal conditions (minimum base flow) for using the stream tracer approach to detect hyporheic exchange at St. Kevin Gulch, the approach was not very sensitive to the interaction of the stream with deeper alluvium.

The main advantage of the stream tracer approach is its simplicity and efficiency at large scales of application. Yet clearly there are some fundamental limitations that result from assuming that exchange can be represented by a simple mass transfer between the channel and a well-mixed reservoir. The mathematical representation of storage in the stream tracer model has first-order dynamics, which is equivalent to assuming that the distribution of travel times in hyporheic flow paths is exponential [*Levenspiel*, 1972]. In contrast, observations of tracer movement to wells at St. Kevin Gulch suggested that there were two exponential distributions of travel times in hyporheic flow paths (Figure 7). The distribution of residence times in hyporheic flow paths at St. Kevin Gulch was determined by plotting well distance away from the stream-versus-tracer travel time to wells on a semilog scale. The linear trends on semilog plots (Figure 7) imply that travel times were approximately exponentially distributed, and different slopes for gravel bar and alluvium wells imply that there are two characteristic timescales of exchange between stream and hyporheic zone. The characteristic travel time to gravel bar wells was 6 hours, which closely agreed with the best-fit residence time of the storage zone specified by modeling (Table 3). The characteristic travel time to alluvium wells was 84 hours, more than an order of magnitude longer than for gravel bar wells. A stream tracer model with only one storage reservoir cannot characterize two very different residence time distributions of hyporheic exchange.

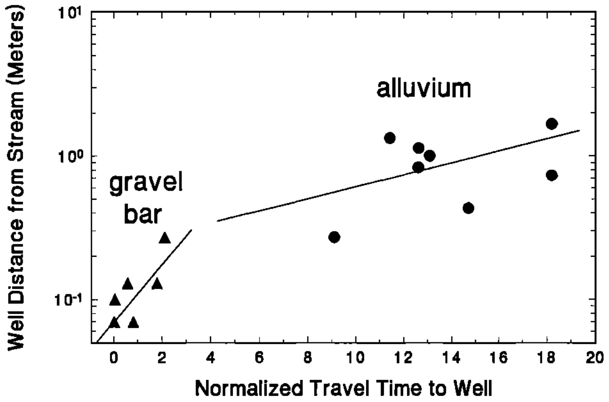

Figure 7. Well distances from stream versus normalized travel time for stream tracer to reach wells. Time is normalized to best-fit storage zone residence time. The linear trends on semilogarithmic plots imply that travel times in the hyporheic zone are exponentially distributed. Fast exchange with gravel bars and slower exchange with deeper alluvium is indicated.

Comparison of a Best-Fit Stream Tracer Simulation With a "Hydrometric-Based" Simulation

Another means to evaluate the reliability of the stream tracer approach to characterize hyporheic exchange was to compare the best-fit tracer transport simulations with a simulation based on detailed hydrometric measurements in the subsurface (referred to as the "hydrometric-based" simulation). The best-fit simulation used storage parameters obtained from the inverse analysis while the hydrometric-based simulation used storage parameters determined by subsurface measurements. The purpose of the comparison was to test the ability of the stream tracer approach to represent hyporheic zone dimensions and fluid residence times that were observed by independent observations in the subsurface. For the hydrometric-based simulation the exchange coefficient, α, was computed from (8) using the low–base flow values of q_S from Table 1 while holding constant the value of A for the low–base flow simulation (Table 3). The cross-sectional area of the storage zone, A_S, was roughly estimated from hydraulic head data from Figure 5 and *Harvey and Bencala* [1993], which indicated that the hyporheic zone was on the order of five times larger than the cross-sectional area of the stream. The value of A_S was therefore set equal to A (low–base flow value from Table 3) multiplied by 5. All other parameters of the hydrometric-based simulation were the same as reported in the low–base flow column of Table 3. Both simulations were compared with an independent data set, the solute tracer data collected in subsurface wells, to evaluate consistency between stream tracer and hydrometric approaches.

In addition to providing the best match to measured tracer concentrations in the stream (Figure 8), the best-fit simulation provided a good approximation of fast exchange with gravel bar sediment (Figure 9). However, the best-fit storage simulation was a poor simulator of slower exchange with alluvium, especially compared with the good match provided by the hydrometric-based simulation (Figure 9). These results illustrate that the simple mathematical representation of storage in the stream tracer model cannot simulate both rapid exchange with gravel bars and slower exchange with deep alluvium in the same simulation. In the procedure of fitting the model to stream tracer data, fast exchange with gravel bar sediment apparently had a greater effect on in-stream tracer concentrations than did slow exchange with alluvium. Slower exchange between stream and alluvium was simulated better using parameters estimated from detailed hydrometric measurements. Used alone, the stream tracer approach therefore provided an estimate of hyporheic exchange that was biased toward representing the faster exchange pathways between stream and subsurface.

Discussion

Previous studies indicated a relation between hyporheic zone dimensions and in-stream factors such as roughness features or streambed slope variation [*Thibodeaux and Boyle*, 1987; *Harvey and Bencala*, 1993]. The present study showed the importance of hyporheic exchange across a 10-fold range in stream base flow. At St. Kevin Gulch hyporheic zone characteristics were influenced by a balance between effects of in-stream bed topography and hillslope hydraulic potentials on hydraulic potentials adjacent to the stream. The magnitude of hyporheic exchange decreased by 30% at high base flow, owing to increased groundwater inflow, which resisted recharge of stream water to hyporheic flow paths. However, even when hyporheic exchange was reduced at high base flow, that component of the streambed flux was still nearly as large as the net groundwater flux across the streambed (Table 1), illustrating that hyporheic flow persists at all times whether base flow is high or low.

Our goal was to determine whether the stream tracer approach could efficiently characterize hyporheic zone dimensions and exchange timescales. We found that the stream tracer approach had only minimal sensitivity to surface-subsurface exchange at high base flow. Tracer mass flux in the channel was so large at high base flow that only a small proportion of the injected tracer could interact with the subsurface. At high base flow our stream tracer-hydrometric comparisons indicated that stream tracer methods are probably more sensitive to surface water storage processes than to hyporheic exchange. As a result only cautious conclusions about variability in hyporheic processes are possible based on stream tracer studies alone. Parameter differences between study sites [*Broshears et al.*, 1993] or changes in parameters from low to high base flow [*Legrand-Marq and Laudelout*, 1985; *D'Angelo et al.*, 1993; *Morrice et al.*, 1996; this study] therefore do not necessarily represent differences in hyporheic zone characteristics, and results need to be interpreted carefully. Our findings suggest that it may be possible to design stream tracer experiments for the specific purpose of identifying hyporheic zone characteristics, through careful consideration of assumptions reviewed in this paper and other network design principles [*Wagner and Harvey*, 1994]. Such an effort will need to consider changing sensitivities as a function of flow conditions in streams, as well as the possibility that multiple timescales of hyporheic exchange may need to be identified.

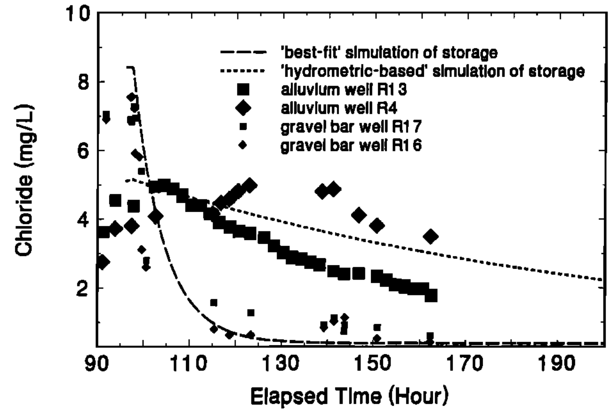

Figure 9. Comparison of best-fit and hydrometric-based simulations of exchange with storage zones for the low–base flow study at St. Kevin Gulch.

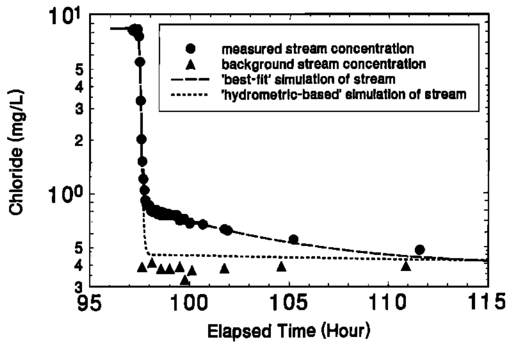

Figure 8. Comparison of best-fit and hydrometric-based simulations of stream tracer transport for the low–base flow study at St. Kevin Gulch.

Even under optimal conditions for identifying hyporheic zone characteristics (low base flow), the stream tracer approach still could not capture the two timescales of stream-hyporheic water exchange that were evident in subsurface data at St. Kevin Gulch. The best-fit simulation clearly had much greater sensitivity to the fast exchange between the stream and coarse gravel bar deposits compared to slower exchange with deeper alluvium. We suspect that a multirate exchange model formulation of the stream tracer model, with two classes of storage zone reservoirs with short and long time constants, respectively, would be necessary to develop a simulation that was consistent with all the field data collected in our study.

Most previous tracer modeling in streams has considered only a single timescale of storage. *Jackman et al.* [1984] examined the performance of several alternative formulations of storage zone submodels (including a linear reservoir mass transfer approach, a diffusion approach, and a highly dispersed plug flow approach). Jackman et al. found that the performance of all of the models was relatively equivalent, probably owing to the fact that the simple diffusive and dispersive formulations used each have first order dynamics and only one fundamental timescale for exchange. An exception to single-timescale exchange studies is the work of *Castro and Hornberger* [1991]; they used a times series model to identify two storage reservoirs in some stream reaches. Their two-timescale storage simulation appeared to be warranted, even with the added uncertainty of independently identifying additional parameters, which suggests that there may often be sufficient information in stream tracer experiments to determine multiple timescales of exchange.

Summary

The main advantage of the stream tracer approach to detect hyporheic exchange is the simplicity and efficiency at large scales of application compared to detailed subsurface observations. On the basis of a direct comparison of stream tracer and hydrometric methods we found the following: (1) Hyporheic exchange persisted across two seasons spanning conditions of low and high base flow, accounting for 40–80% of the total streambed flux; (2) differences in the magnitude of hyporheic exchange between periods of low and high base flow were accounted for by seasonal variation in groundwater inflow from the hillslope, a force that opposed localized recharge of stream water into hyporheic flow paths; (3) the stream tracer approach estimated hyporheic exchange with greater reliability at low base flow than at high base flow; (4) greater sensitivity at low base flow resulted from several factors, including more interaction of a larger proportion of the tracer with hyporheic flow paths at low flow and higher tracer plateau concentrations; and (5) even under more favorable low–base flow conditions, use of the stream tracer approach to determine hyporheic exchange parameters accurately characterized only the fastest exchange timescales (i.e., exchange between stream and gravel bars) identified by hydrometric analysis. Therefore, even under optimal conditions for application at minimum base flow, the stream tracer approach still may not be sensitive to longer timescale interactions with hyporheic flow paths in deeper alluvium. In some cases it may be possible to identify longer exchange timescales using an extended stream tracer model with multiple rate constants to identify characteristics both for the exchange with bed sediment and exchange with deeper alluvium.

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