



Channel water balance and exchange with subsurface flow along a mountain headwater stream in Montana, United States

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[1] Channel water balances of contiguous reaches along streams represent a poorly understood scale of stream-subsurface interaction. We measured reach water balances along a headwater stream in Montana, United States, during summer base flow recessions. Reach water balances were estimated from series of tracer tests in 13 consecutive reaches delineated evenly along a 2.6 km valley segment. For each reach, we estimated net change in discharge, gross hydrologic loss, and gross hydrologic gain from tracer dilution and mass recovery. Four series of tracer tests were performed during relatively high, intermediate, and low base flow conditions. The relative distribution of channel water along the stream was strongly related to a transition in valley structure, with a general increase in gross losses through the recession. During tracer tests at intermediate and low flows, there were frequent substantial losses of tracer mass (>10%) that could not be explained by net loss in flow over the reach, indicating that many of the study reaches were concurrently losing and gaining water. For example, one reach with little net change in discharge exchanged nearly 20% of upstream flow with gains and losses along the reach. These substantial bidirectional exchanges suggest that some channel interactions with subsurface flow paths were not measurable by net change in flow or transient storage of recovered tracer. Understanding bidirectional channel water balances in stream reaches along valleys is critical to an accurate assessment of stream solute fate and transport and to a full assessment of exchanges between the stream channel and surrounding subsurface.

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1. Introduction

[2] Exchanges between stream channel and subsurface flows are driven by variability in hydraulic gradients that are induced by structural variability in channels and valley floors [Harvey and Bencala, 1993; Woessner, 2000; Kasahara and Wondzell, 2003]. Channel and valley structures that influence stream-subsurface exchange occur at multiple scales [Dent et al., 2001; Dahl et al., 2007; Cardenas, 2008], and a few examples include: substrate bed forms (e.g., sand ripples and dunes [Worman et al., 2002]), channel units (e.g., step sequences [Wondzell, 2006]), channel meanders [e.g., Boano et al., 2006], and convergent/divergent valley floors [e.g.,

Stanford and Ward, 1993]. These structures can induce subsurface flow paths that both recharge and discharge in the stream channel (hyporheic flow), resulting in multiscaled hyporheic flow paths that span a broad range of transport times and distances [Cardenas, 2008]. Therefore, hyporheic flow paths can have multiple concurrent effects on channel flow within a given reach, including any combination of: (1) an increase in channel flow along the reach (gross gain, e.g., Figure 1 flow path B); (2) a decrease in channel flow along the reach (gross loss, e.g., Figure 1 flow path D); and (3) retention of channel water for a period of time before arrival at the base of the reach (transient storage, e.g., Figure 1 flow path F). Stream hydrologic studies infrequently consider the potential for concurrent gross gain and gross loss over a reach. However, acknowledging the potential for bidirectional channel water balance is critical to understanding the full influence of hyporheic flow on channel water and associated solute loads [Zellweger, 1994; Harvey and Wagner, 2000; Ruehl et al., 2006].

[3] Net change in channel flow over a stream reach is seldom separated into its constituent gross gains and gross losses. Water balance is typically represented only by net gain or net loss, as indicated by the difference between downstream and upstream channel flow. For example, gross loss to the subsurface is commonly assumed to be negligible in net

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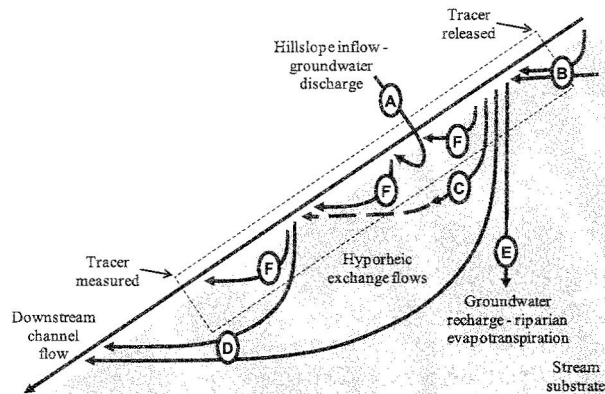


Figure 1. Conceptual, schematic profile of stream channel exchange with an underlying nested subsurface flow network. The dashed box represents the region influenced by an individual reach-scale tracer test. Thick arrows represent flow paths in the stream channel and in the underlying substrate. The thick dashed arrow (tail end of flow path C) represents a subsurface flow path with residence time longer than the duration of tracer concentration measurement. Flow paths A, B, and C (tail end) are unlabeled with tracer and dilute the measurable tracer concentrations in the channel. These flow paths are considered gross hydrologic gain in the context of the tracer test. Flow paths C (head end), D, and E remove tracer mass during the tracer test and reduce the amount of tracer mass recovered. These flow paths are considered gross hydrologic loss in the context of the tracer test. Flow paths labeled F return tracer to the stream in measurable concentrations: a temporary period of storage relative to channel flow. Tracer flow paths considered transient storage during the tracer and do not influence the stream water balance.

gaining stream reaches. This simplification not only underestimates stream-subsurface exchange over a reach where gross loss is substantial, but also may result in a more general underestimate of hyporheic interactions aggregated over consecutive reaches of a valley segment [Ruehl *et al.*, 2006]. A few field studies have suggested that concurrent gross gain and gross loss can be substantial in streams and rivers [Zellweger *et al.*, 1989; Ruehl *et al.*, 2006; Covino and McGlynn, 2007]. However, we lack the extensive field data necessary to understand where and when channel water balance may be important to stream water quantity and quality, and we lack the systematic approaches needed to gather those data. We suggest that substantial gross gain and gross loss (>10%) should be measurable along small stream channels with substantial channel and valley structural variability, and we provide supporting water balance analyses from a mountain headwater stream.

(4) Multiple tracer tests were used to estimate bidirectional channel water balances in reaches along a first- to second-order mountain stream, draining a 5.5 km² watershed in the Rocky Mountains of west-central Montana, United States. Tracer masses were released instantaneously near the end points of 13 contiguous reaches, delineated evenly along 2.6 km of the stream valley. Dilution and mass recovery of tracer were used to estimate net change in discharge, gross gain, and gross loss for each reach. Four series of tracer tests

were used to obtain "snapshots" of water balance along the study stream at different summer base flows during seasonal discharge recessions. Our immediate objectives for analyses of water balance data are: (1) to examine the magnitude and distribution net change in flow, gross gains, and losses that are likely influenced by stream-subsurface exchange along a segment and (2) to examine how bidirectional channel water balances respond to changing base flow conditions.

2. Methods

2.1. Study Site Description

[5] This study was performed at the Tenderfoot Creek Experimental Forest, a research watershed located in the Little Belt Range of the Rocky Mountains in Montana, United States (lat. 46°55'N, long. 110°52'W) and managed by the United States Department of Agriculture Forest Service. Our study focused on Stringer Creek, a headwater draining a 5.5 km² subcatchment of the experimental watershed (Figure 2). We performed stream tracer experiments along 2.6 km of the valley, from near the confluence with Tenderfoot Creek to near initiation of flow during the lowest base flows (early autumn). A perennial tributary flows into Stringer Creek about 2.2 km upstream from the confluence with Tenderfoot Creek, dividing the first- and second-order segments of the study stream. Water balance was estimated in 13 contiguous study reaches, selected by evenly dividing the 2.6 km of valley into 200 m segments. Each reach is denoted by valley distance from the downstream end of the reach to the gauge at the base of the study

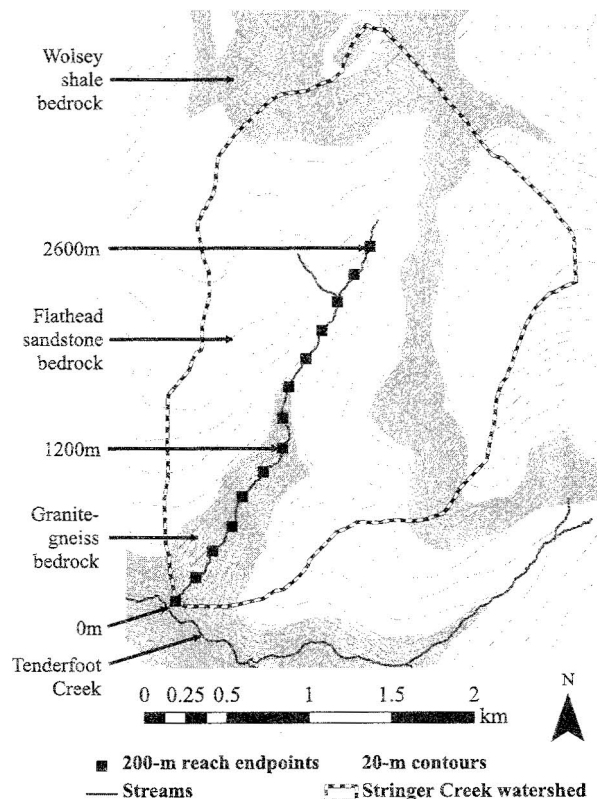


Figure 2. Plan view of the Stringer Creek watershed (latitude 46°55'N, longitude 110°52'W).

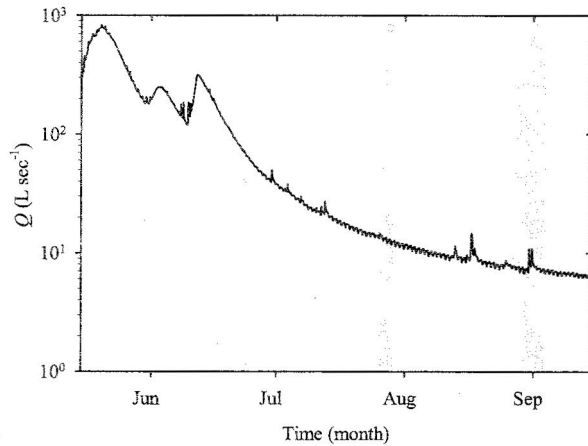


Figure 3. Spring-summer 2006 hydrograph measured at the gauge near 1200 m in Stringer Creek. Shaded areas represent days when three consecutive slug series were performed in 2006.

stream (every 200 m from the 0 m reach at the base to the 2400 m reach at the head, Figure 2).

[6] An abrupt change in valley structure along Stringer Creek corresponds to a bedrock transition between sandstone upstream and granite-gneiss downstream (Figure 2) [Reynolds, 1995]. The structural transition in the valley occurs approximately 1.2 km upstream from the confluence with Tenderfoot Creek (near the base of the 1200 m reach). The valley upstream from the transition has a wide floor and shallow downvalley slope (5.7%) relative to the constrained floor and steep slope (9.0%) downstream. Riparian vegetation also differs with the transition in valley structure. Upstream, the valley floor is nearly free of trees, and flora is composed primarily of grasses and forbs [Mincemoyer and Birdsall, 2006]. Grassy meadows on the upstream valley floor abruptly change to lodgepole pine forest (*Pinus contorta*) at the toe of adjacent hillslopes. Meadows more frequently extend to hillslopes and become wider with distance upstream, culminating in large "parks" between the tributary confluence (~2200 m) and initiation of lower base flows (~2600 m). In sharp contrast, there are relatively few riparian meadows downstream of the structural transition and trees are common near the active channel. In general, hillslope soil depths are approximately 0.5–1.0 m, and depths of unconsolidated materials in the riparian zone are approximately 1.0–2.0 m, based on depths to refusal during shallow well installations [Jencso et al., 2009].

[7] There are two stream gauges along Stringer Creek. A 4 foot H flume is located just upstream of the confluence with Tenderfoot Creek (0 m), and a 3.5 foot H flume is located just downstream of the structural transition and the 1200 m reach. Design and stage discharge rating equations for these gauges were developed by the U.S. Department of Agriculture [Brakensiek et al., 1979; Farnes et al., 1999]. Stages in both flumes were measured in stilling wells and logged at 15 min intervals by capacitance rods accurate to ± 0.5 mm (TruTrack, New Zealand, note that the use of trade or firm names in this publication is for reader information and does not imply endorsement by the U.S. Department of Agriculture or U.S. Geological Survey of any product or service). Snowmelt and spring rains dominate annual run-

off, and the typical convective summer storm has little sustained influence on streamflow (Figure 3).

2.2. Interpretation of Reach Water Balance From Tracer Tests

[8] A common field method in stream hydrology is to release a conservative tracer to the channel at the head of a study reach, then analyze the tracer concentration over time (breakthrough curve) in the channel at the base of that study reach. Breakthrough curve analyses have typically focused on residence time of recovered tracer mass (tracer mass that is measurable in concentrations at the base of the reach [e.g., Payn et al., 2008]) and have typically neglected any gross losses that might be indicated by loss of tracer mass [Ruehl et al. [2006] is a recent exception). Tracer mass loss is caused by: (1) hyporheic flow paths that retain tracer longer than the duration of the tracer test (Figure 1 flow path C); (2) hyporheic flow paths that transport tracer beyond the downstream end of the reach (Figure 1 flow path D); and (3) flow paths do not return to the channel (Figure 1 flow path E). Neglect of substantial gross loss also results in underestimates of tracer dilution due to gross gain. Tracer mass in the channel will be diluted by: (1) flow paths that contribute groundwater (or water that has never been in the stream, Figure 1 flow path A); (2) hyporheic flow paths that return water that was lost from the channel upstream from the tracer release (Figure 1 flow path B); and (3) hyporheic flow paths that return water that was lost from the channel downstream from the tracer release, but was lost before being labeled with tracer (Figure 1 flow path C). Specific subsurface flow paths that cause dilution (Figure 1 among flow paths A, B, or C) or mass loss (Figure 1 among flow paths C, D, or E) cannot be individually identified using tracer data from the channel. However, the influence of gross losses and gross gains can be distinguished from net change in flow using dilution gauging and mass recovery analyses.

[9] Independent tracer releases, dilution gauging, and mass recovery analyses were used to estimate discharge, gross losses, and gross gains in each of the 13 study reaches, working consecutively from downstream to upstream. We used dilution gauging over short "mixing length" reaches to estimate discharge at the downstream (Q_D) and upstream (Q_U) ends of each reach. Net change in discharge for each reach was calculated by difference ($\Delta Q = Q_D - Q_U$). Tracer mass that was lost in transport over the reach (M_{LOSS}) was used to estimate gross hydrologic loss (Q_{LOSS}). Gross hydrologic gain (Q_{GAIN}) was estimated by water balance ($Q_{GAIN} = \Delta Q - Q_{LOSS}$ where gain is positive and loss is negative). Details of tracer test design and analyses are in the following section.

[10] A substantial M_{LOSS} indicates influence of subsurface flow paths at larger space and time scales than flow paths indicated by transient storage [Ruehl et al., 2006]. This scale dependence is related to the "window of detection" of transient storage analyses, which suggests that estimates of transient storage are only sensitive to channel storage and hyporheic flow paths at relatively small spatial scales and short temporal scales [Harvey et al., 1996; Wagner and Harvey, 1997]. Therefore, M_{LOSS} may be due to larger spatial scale subsurface flow paths that do not return to the stream channel within the study reach (Figure 1 flow path D) or to longer temporal scale hyporheic flow

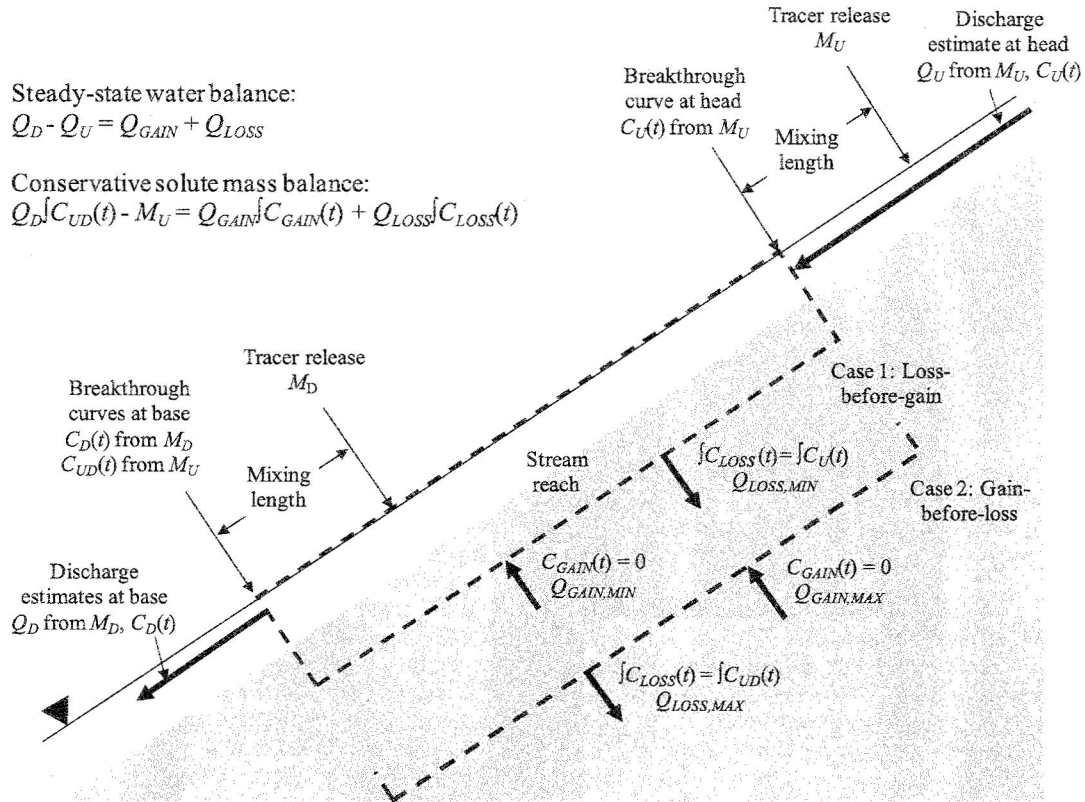


Figure 4. Summary of slug releases, breakthrough curve measurements, and discharge estimates necessary for estimate of water balance for one reach. A slug mass (M_D) is released a mixing length upstream from the base of the reach, and a breakthrough curve is measured at the downstream end of the reach ($C_D(t)$). Then, a second slug mass (M_U) is released a mixing length upstream from the head of the 200 m reach, and breakthrough curves are measured at the upstream ($C_U(t)$) and downstream ends of the reach ($C_{UD}(t)$). Discharges are estimated from releases at the base and head of each reach (Q_D, Q_U), and mass recovery of M_U indicated by $C_{UD}(t)$ is used to estimate gross hydrologic gain and loss (Q_{GAIN}, Q_{LOSS}). Two distributions of gross gain and loss are represented, where loss occurs upstream of gain (case 1) and gain occurs upstream of loss (case 2). Illustration is not to scale; that is, mixing lengths are exaggerated relative to the reach lengths.

paths that do return within the study reach, but not until after termination of the tracer test (Figure 1 flow path C). References to "small or large" spatial scales and "short or long" temporal scales here are relative to the length of the study reach and duration of the tracer test under consideration.

[11] In this study, durations of tracer tests were determined by an apparent return to background conditions after an instantaneous tracer release. Return time to background was typically around 2 to 3 channel transport times after the release of tracer (approximately 0.7 to 3 h depending on discharge), where channel transport time is defined as the time from tracer release to peak in tracer concentration. Thus, the functional definition of Q_{LOSS} for this study is: water that started in the channel at the head of the study reach and did not reach the channel at the base of the study reach within 2 to 3 channel transport times.

2.3. Tracer Test Design and Analyses

[12] Sodium chloride (NaCl) was used as a conservative tracer, and concentrations were estimated by calibrating temperature corrected electrical conductivity measurements (EC) to known tracer concentrations in stream water

[Gooseff and McGlynn, 2005; Wondzell, 2006]. EC measurements were made with Campbell CR510 or CR10X data loggers and CS-547 A-L temperature/conductivity probes (Campbell Scientific, Inc., Logan, Utah, United States). Each probe was independently calibrated in 2006, and a calibration curve from a single probe was used for all probes in 2005. Calibrations were performed using standards made from known masses of NaCl dissolved in known volumes of stream water. Water for standards was collected from the study stream on the day of calibration or the day before. Calibrations were performed one or two days before each series of tracer tests, or on the day after. All stream EC measurements were corrected for background EC before applying the calibration slope to estimate NaCl concentration.

[13] Selections of "mixing length" reaches for dilution gauging (Figure 4) were based on structure of the wetted stream channel (e.g., pools, riffles, runs, drops, etc.). The goal was to maximize the likelihood of complete mixing within a representative volume of moving stream water, but minimize discharge overestimates due to tracer mass loss over mixing lengths. Mixing distances are not reliably predicted by theoretical functions of channel morphometries in the complex channels of mountain headwater streams [Day, 1977].

Therefore, the locations for EC measurements and tracer releases were determined by on-site observation, with the understanding that wetted channel structure and reach lengths may change with flow conditions. First, we selected an EC measurement location by avoiding locally unmixed regions directly within pools or downstream of identifiable inflows. Then, we selected an upstream tracer release location such that the mixing length consisted of at least three transitions between divergent and convergent streamflow (e.g., pool-riffle sequences), in order to maximize repeated, turbulent self-contact within a representative volume of moving water. In a few cases (~4-5), tracer experiments were repeated immediately, sometimes over a different mixing distance, when initial breakthrough curves showed obvious signs of equipment failure or incomplete mixing. Further experiments to ensure complete tracer mixing at every location and flow condition in this study were not practical, so here we focus on more pronounced variability in results and repeatability of general patterns. Mixing lengths in this study were between 5 and 30 m in valley distance, depending on channel structure and discharge, where mixing lengths were generally longer at higher discharge.

[14] Tracer tests for a given reach were initiated with an instantaneous release of predissolved tracer mass (M_U) a mixing length above the downstream end of the reach (Figure 4). Tracer concentration was measured at the base of the reach ($C_D(t)$) after M_D was released (Figure 4). Dilution gauging [Day, 1977] was used to estimate Q_D , assuming constant discharge and complete mass recovery:

$$Q_D = \frac{M_D}{\int_0^t C_D(\tau) d\tau} \quad (1)$$

where τ is the time variable of integration, and t is the time of the experiment between the release time ($T = 0$) and the return to background EC ($T = t$). We used trapezoidal numerical approximation for all integrations of discrete concentration breakthrough curves, which were logged at 2 to 5 s intervals. After a return to background EC, indicating $C_D(T = t) \sim 0$, a second tracer mass (M_U) was released a mixing length from the upstream end of the reach (Figure 4). Tracer concentrations resulting from the M_U release were then measured at both the upstream ($C_U(t)$) and downstream ($C_{UD}(t)$) ends of the reach. Dilution gauging was used to estimate Q_U from $C_U(t)$ and M_U (similar to equation (1)), and net change in discharge over the reach was calculated by difference ($\Delta Q = Q_D - Q_U$)

where ΔQ is a net gain if positive and a net loss if negative.

[15] We used a method similar to Rieckermann *et al.* [2005, 2007] to estimate M_{LOSS} or the loss of M_U in transport along the reach. This M_{LOSS} is tracer that was not recovered in measurement of $C_{UD}(t)$ before an apparent return to background EC, indicating $C_{UD}(T = t) \sim 0$. The recovered tracer mass (M_{REC}) is the integration of downstream tracer load ($M_{REC} = Q_D \int_0^t C_{UD}(\tau) d\tau$), assuming constant Q_D , which requires the independent estimate of Q_D from the first tracer test. We calculated the corresponding M_{LOSS} by conservation of mass:

$$M_{LOSS} = M_{REC} - M_U = Q_D \int_0^t C_{UD}(\tau) d\tau - M_U \quad (2)$$

where M_{LOSS} has negative value when a portion of M_U is lost from the reach.

[16] The amount of water associated with transport of M_{LOSS} was used to infer Q_{LOSS} and Q_{GAIN} during the second tracer test (Figure 1). The method was similar to that of Zellweger *et al.* [1989] and followed the suggestions of Harvey and Wagner [2000], with two notable distinctions. First, we elected to use dilution gauging rather than velocity gauging to estimate downstream discharge Q_D , because dilution gauging is likely to be more accurate than velocity gauging in small, tortuous stream channels [Day, 1977; Zellweger *et al.*, 1989], such as Stringer Creek. Second, we estimated a range of gross hydrologic gain and loss dependent on the order in which they occur, rather than the single estimate suggested by Harvey and Wagner [2000]. The M_{LOSS} from a reach is the integration of tracer load defined by concentration of tracer in lost water ($M_{LOSS} = Q_{LOSS} \int_0^t fC_{LOSS}(\tau) d\tau$, assuming constant Q_{LOSS}). However, $C_{LOSS}(t)$ is determined by tracer concentration in the channel at the location of Q_{LOSS} , and tracer concentration in the channel may be influenced by upstream Q_{GAIN} . Hence, estimate of Q_{LOSS} from a given M_{LOSS} depends on the magnitude and location of tracer dilution by Q_{GAIN} to the channel. The minimum estimate $Q_{LOSS,MIN}$ assumes minimum dilution before loss, or all loss occurring upstream of all gain (i.e., $fC_{LOSS}(t) = fC_U(t)$). The maximum estimate $Q_{LOSS,MAX}$ assumes maximum dilution before loss, or all gain occurring upstream of all loss ($fC_{LOSS}(t) = fC_{UD}(t)$, Figure 4). Therefore, $Q_{LOSS,MIN}$ and $Q_{LOSS,MAX}$ were calculated by

$$Q_{LOSS,MIN} = \frac{M_{LOSS}}{\int_0^t C_U(\tau) d\tau} \quad (3a)$$

$$Q_{LOSS,MAX} = \frac{M_{LOSS}}{\int_0^t C_{UD}(\tau) d\tau} \quad (3b)$$

[17] Finally, we calculated the corresponding range of gross gain ($Q_{GAIN,MIN}$ and $Q_{GAIN,MAX}$) by mass balance (Figure 4), assuming no change in the volume of storage in each reach ($Q_{GAIN} = \Delta Q - Q_{LOSS}$). $Q_{GAIN,MIN}$ and $Q_{GAIN,MAX}$ were thus calculated from $Q_{LOSS,MIN}$ and $Q_{LOSS,MAX}$, respectively.

[18] Discharge from the head of each reach was used as the discharge at the base of the next reach upstream (downstream reach Q_U = upstream reach Q_D). Progressing upstream in this fashion conserves time and tracer because each reach can be completed with a single tracer release, with the exception of the first reach during each day of tracer tests. Each morning of a given series, a mixing length tracer test was repeated at the last location from the previous day, to account for any difference in discharge when calculating channel water balances for the adjacent reaches.

[19] Tracer test series were conducted in Stringer Creek on four occasions over two summers: 4-6 August 2005, 22-24 June 2006, 25-28 July 2006, and 26 August to 4 September 2006 (Figure 3). The four series of tracer tests spanned base flow discharges ranging from 15 L s⁻¹ to 101 L s⁻¹ at 0 m ($Q(0)$ m), Figure 5). Thus, results allow comparisons of

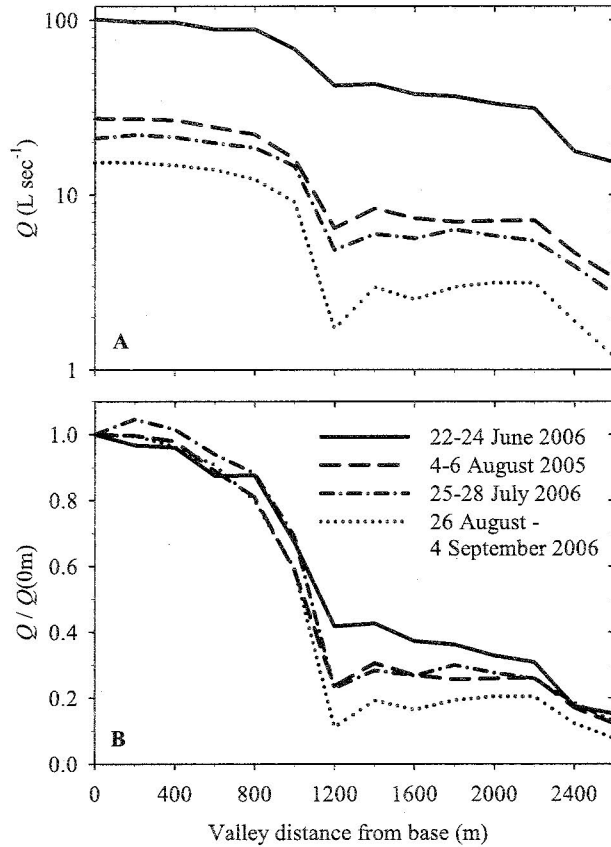


Figure 5. (a) Stream channel discharge (Q) estimated every 200 m from dilution gauging slugs along Stringer Creek and (b) discharge along Stringer Creek as a fraction of channel flow at 0 m ($Q/Q(0\text{ m})$).

longitudinal distributions of stream water balance across a wide range of base flow conditions in this stream.

3. Results

[20] Repeated series of tracer tests revealed variability in channel water balance both through time and across space. Net variability in discharge provides a context of stream-flow generation during the base flow recession study period. Variability in gross exchanges of channel water balance reveals when and where further stream-subsurface exchanges were likely to influence channel flow.

3.1. Net Changes in Channel Flow Over Time and Along the Valley

[21] The highest flows and net gains among the four tracer test series occurred during June 2006, when stream discharge longitudinally increased from $Q(2600\text{ m}) = 15\text{ L s}^{-1}$ to $Q(0\text{ m}) = 101\text{ L s}^{-1}$ (Figure 5a and Table 1). Series performed during August 2005 and July 2006 reflect intermediate flows among the experiments, when $Q(2600\text{ m}) = 3\text{ L s}^{-1}$ in both years and $Q(0\text{ m}) = 27\text{ L s}^{-1}$ in 2005 and $Q(0\text{ m}) = 21\text{ L s}^{-1}$ in 2006 (Tables 2 and 3). In general, 2005 was a wetter year than 2006, resulting in generally higher discharges later in the summer. The lowest flows among the four tracer test series occurred in late August of 2006, when $Q(2600\text{ m}) = 1\text{ L s}^{-1}$ and $Q(0\text{ m}) = 15\text{ L s}^{-1}$ (Table 4).

Table 1. Tracer Masses, Areas Under Breakthrough Curves, Mass Losses, Discharge Estimates From Tracer Tests When $Q(0\text{ m}) = 101\text{ L s}^{-1}$ at the Base of Stringer Creek

Reach (m)	M_D (g)	$\int_0^t C_D(\tau)d\tau$ (mg L ⁻¹ min)	Q_D (L s ⁻¹)	M_U (g)	$\int_0^t C_U(\tau)d\tau$ (mg L ⁻¹ min)	Q_U (L s ⁻¹)	ΔQ (L s ⁻¹)	M_{LOSS} (%)	$Q_{LOSS,MIN}$ (L s ⁻¹)	$Q_{LOSS,MAX}$ (L s ⁻¹)	$Q_{GAIN,MIN}^a$ (L s ⁻¹)	$Q_{GAIN,MAX}^a$ (L s ⁻¹)
0	600	98.8	101	1200	204	97.8	3	0.3	0	0	3	3
200	1200	204	97.8	1100	189	97.2	0.7	-1.3	-1.3	-1.3	1.9	2.0
400	1100	189	97.2	1000	188	88.5	8.7	0.5	0.4	0.5	8.7	8.7
600	1000	188	88.5	1000	175	95.4	-6.4	-6.4	-6.1	-6.0	-0.9	-0.9
800	400	81.5	81.8	900	221	68.0	13.8	-3.4	-2.3	-2.9	16.1	16.7
1000	900	221	68.0	800	316	42.2	25.8	1.5	0.6	1.0	25.8	25.8
1200	800	316	42.2	700	271	43.1	-0.9	-2.7	-1.2	-1.2	0.3	0.3
1400	700	271	43.1	600	265	37.7	5.4	1.1	0.4	0.5	5.4	5.4
1600	600	265	37.7	600	266	37.6	0.1	-2.1	-0.8	-0.8	0.9	0.9
1800	200	93.5	35.6	500	251	33.3	2.4	4.2	1.4	1.4	2.4	2.4
2000	500	251	33.3	500	267	31.3	2.0	1.1	0.4	0.4	2.0	2.0
2200	500	267	31.3	400	376	17.7	13.5	1.0	0.2	0.3	13.5	13.5
2400	400	376	17.7	300	325	15.4	2.4	0.2	0.0	0.0	2.4	2.4

^aPositive gross losses were considered as errors and were not considered in calculation of gross gain, so gross gain is equal to net gain when gross loss is positive.

Table 2. Tracer Masses, Areas Under Breakthrough Curves, Mass Losses, Discharge Estimates, and Water Balance Estimates From Tracer Tests When $Q(0 \text{ m}) = 27 \text{ L s}^{-1}$ at the Base of Stringer Creek

Reach (m)	M_D (g)	$\int_0^t C_D(\tau) d\tau$ ($\text{mg L}^{-1} \text{ min}$)	Q_D (L s^{-1})	M_U (g)	$\int_0^t C_U(\tau) d\tau$ ($\text{mg L}^{-1} \text{ min}$)	Q_U (L s^{-1})	$\int_0^t C_{UD}(\tau) d\tau$ ($\text{mg L}^{-1} \text{ min}$)	$\frac{M_{\text{LOSS}}}{M_U}$ (%)	ΔQ (L s^{-1})	$Q_{\text{LOSS},\text{MIN}}$ (L s^{-1})	$Q_{\text{LOSS},\text{MAX}}$ (L s^{-1})	$Q_{\text{GAIN},\text{MIN}}^a$ (L s^{-1})	$Q_{\text{GAIN},\text{MAX}}^a$ (L s^{-1})
0	803	488	27.4	1603	979	27.3	872	-10.6	0.1	-2.9	-3.2	3.0	3.4
200	1603	979	27.3	1403	872	26.8	788	-8.0	0.5	-2.1	-2.4	2.6	2.8
400	1403	872	26.8	1304	892	24.4	755	-6.8	2.5	-1.7	-2.0	4.1	4.4
600	1304	892	24.4	1201	902	22.2	737	-10.4	2.2	-2.3	-2.8	4.5	5.0
800	1201	902	22.2	1001	1040	16.1	633	-15.8	6.1	-2.5	-4.2	8.7	10.3
1000	1001	1040	16.1	902	2420	6.2	646	-30.9	9.9	-1.9	-7.2	11.8	17.0
1200	400	989	6.7	802	1590	8.4	1230	-38.1	-1.6	-3.2	-4.1	1.6	2.5
1400	802	1590	8.4	801	1810	7.4	1340	-15.6	1.0	-1.2	-1.6	2.2	2.6
1600	801	1810	7.4	800	1900	7.0	1590	-12.4	0.3	-0.9	-1.0	1.2	1.4
1800	800	1900	7.0	700	1640	7.1	1410	-15.2	-0.1	-1.1	-1.3	1.0	1.2
2000	700	1640	7.1	600	1440	6.9	1190	-15.7	0.2	-1.1	-1.3	1.3	1.5
2200	300	676	7.4	501	1790	4.7	1060	-6.1	2.7	-0.3	-0.5	3.0	3.2
2400	501	1790	4.7	200	997	3.3	654	-8.5	1.3	-0.3	-0.4	1.6	1.8

^aPositive gross losses were considered as errors and were not considered in calculation of gross gain, so gross gain is equal to net gain when gross loss is positive.

Table 3. Tracer Masses, Areas Under Breakthrough Curves, Mass Losses, Discharge Estimates, and Water Balance Estimates From Tracer Tests When $Q(0 \text{ m}) = 21 \text{ L s}^{-1}$ at the Base of Stringer Creek

Reach (m)	M_D (g)	$\int_0^t C_D(\tau) d\tau$ ($\text{mg L}^{-1} \text{ min}$)	Q_D (L s^{-1})	M_U (g)	$\int_0^t C_U(\tau) d\tau$ ($\text{mg L}^{-1} \text{ min}$)	Q_U (L s^{-1})	$\int_0^t C_{UD}(\tau) d\tau$ ($\text{mg L}^{-1} \text{ min}$)	$\frac{M_{\text{LOSS}}}{M_U}$ (%)	ΔQ (L s^{-1})	$Q_{\text{LOSS},\text{MIN}}$ (L s^{-1})	$Q_{\text{LOSS},\text{MAX}}$ (L s^{-1})	$Q_{\text{GAIN},\text{MIN}}^a$ (L s^{-1})	$Q_{\text{GAIN},\text{MAX}}^a$ (L s^{-1})
0	500	394	21.1	1400	1060	22.1	967	-12.4	-1.0	-2.7	-3.0	1.8	2.0
200	1400	1060	22.1	1000	777	21.4	710	-5.9	0.7	-1.3	-1.4	1.9	2.1
400	1000	777	21.4	1000	840	19.8	734	-5.6	1.6	-1.1	-1.3	2.7	2.9
600	1000	840	19.8	1000	906	18.4	788	-6.2	1.5	-1.1	-1.3	2.6	2.8
800	500	442	18.9	1000	1140	14.6	769	-12.9	4.3	-1.9	-2.8	6.2	7.1
1000	1000	1140	14.6	1000	3430	4.9	937	-17.9	9.7	-0.9	-3.2	10.6	12.9
1200	1000	3430	4.9	800	2230	6.0	1820	-33.7	-1.1	-2.0	-2.5	0.9	1.3
1400	800	2230	6.0	700	2150	5.4	1630	-16.5	0.6	-0.9	-1.2	1.5	1.7
1600	300	853	5.9	700	1840	6.3	1670	-16.0	-0.5	-1.0	-1.1	0.5	0.7
1800	700	1840	6.3	600	1710	5.8	1760	11.6	0.5	0.7	0.7	0.5	0.5
2000	600	1710	5.8	600	1830	5.5	1600	-6.8	0.4	-0.4	-0.4	0.7	0.8
2200	600	1830	5.5	600	2370	4.2	1530	-16.1	1.3	-0.7	-1.0	1.9	2.3
2400	100	463	3.6	400	2470	2.7	1790	-3.2	0.9	-0.1	-0.1	1.0	1.0

^aPositive gross losses were considered as errors and were not considered in calculation of gross gain, so gross gain is equal to net gain when gross loss is positive.

Table 4. Tracer Masses, Areas Under Breakthrough Curves, Mass Losses, Discharge Estimates, and Water Balance Estimates From Tracer Tests When $Q(0 \text{ m}) = 15 \text{ L s}^{-1}$ at the Base of Stringer Creek

Reach (m)	M_D (g)	$\int_0^t C_D(\tau) d\tau$ ($\text{mg L}^{-1} \text{ min}$)	Q_D (L s^{-1})	M_U (g)	$\int_0^t C_U(\tau) d\tau$ ($\text{mg L}^{-1} \text{ min}$)	Q_U (L s^{-1})	$\int_0^t C_{UD}(\tau) d\tau$ ($\text{mg L}^{-1} \text{ min}$)	$M_{D/U}^{\text{MAX}}$ (%)	ΔQ (L s^{-1})	$Q_{\text{LOSS}, \text{MIN}}$ (L s^{-1})	$Q_{\text{LOSS}, \text{MAX}}$ (L s^{-1})	$Q_{\text{GAIN}, \text{MIN}}$ (L s^{-1})	$Q_{\text{GAIN}, \text{MAX}}$ (L s^{-1})
0	300	325	15.4	700	761	15.3	724	-4.5	0.1	-0.7	-0.7	0.8	0.8
200	700	761	15.3	700	789	14.8	748	-1.8	0.5	-0.3	-0.3	0.8	0.8
400	300	337	14.8	700	836	14.0	749	-4.7	0.9	-0.7	-0.7	1.5	1.6
600	700	836	14.0	700	951	12.3	830	-0.6	1.7	-0.1	-0.1	1.8	1.8
800	300	405	12.3	700 (300) ^b	549	9.1	754	-20.2	3.2	-1.8	-3.1	5.1	6.4
1000	300	549	9.1	700	8150	1.4	1000	-21.7	7.7	-0.3	-2.5	8.0	10.2
1200	200	1670	2.0	600	3370	3.0	2670	-46.5	-1.0	-1.4	-1.7	0.4	0.8
1400	600	3370	3.0	600	4490	2.2	4140	22.7	0.7	0.5	0.5	0.7	0.7
1600	200	1180	2.8	400	2240	3.0	1980	-16.2	-0.1	-0.5	-0.5	0.3	0.4
1800	400	2240	3.0	400	2130	3.1	2260	0.5	-0.2	0.0	0.0	-0.2	-0.2
2000	100	527	3.2	200	1060	3.1	1040	-1.2	0.0	0.0	0.0	0.1	0.1
2200	200	1060	3.1	100	880	1.9	-	-	1.2	-	-	-	-
2400	100	880	1.9	100	1400	1.2	-	-	0.7	-	-	-	-

^aPositive gross losses were considered as errors and were not considered in calculation of gross gain, so gross gain is equal to net gain when gross loss is positive.

^bThe 700 g release was applied to estimate $Q_{D/U}$, and a second 300 g slug was applied to estimate Q_D due to an equipment failure in the upstream probe during the 700 g experiment.

[22] Streamflow gradually decreased with time during a given series of tracer tests due to the base flow recession (Figure 3). Decrease in discharge during a given series of tracer tests (working downstream to upstream) causes overestimation of net gain and underestimation of gross loss, depending on the period of time under consideration. The greatest decreases occurred over the three days of testing in June 2006, when the 3 day decrease at the upstream gauge (near 1200 m) was 24% and the maximum decrease over a single day of tracer tests was 6%. Therefore, the net gain over the entire study stream in June 2006 ($\Delta Q = Q(0 \text{ m}) - Q(2600 \text{ m}) = 86 \text{ L s}^{-1}$) was likely overestimated by about 24%. However, the largest decrease over a given day of tracer tests was only 6% and individual tracer tests lasted from 0.7 to 3 h. Therefore, even the most rapid decreases in discharge during a tracer test series influenced reach water balance estimates by much less than 6%.

[23] Discharge estimates along the stream showed two strong spatial discontinuities during all four series of tracer tests (Figure 5). There was a consistent increase in discharge due to the tributary near 2200 m and a substantial net gain downstream from the valley structure transition near 1200 m. The gain downstream of 1200 m is likely explained by multiple visible springs and seeps in this region, and occurred in the absence of a surface tributary.

[24] Channel flows and net gains along the study stream generally decreased through the summer recession (Figures 5a and 6). Reaches from 0 m to 1200 m were predominantly net gaining during all series, despite the general decrease in flow. In reaches from 0 m to 1000 m, discharges relative to the outlet flow were similar across all series of tracer tests (Figure 5b), indicating that the rates of recession along this segment were similar to the rate of recession at the outlet. The 1000 m reach consistently had the highest net gain, and its relative contribution to channel flow increased from 75% at higher flows to 540% at lower flows (Figure 7). In contrast, the streamflow between 1200 m and 2000 m changed from net gaining to net losing over the recession, and the rates of recession along this segment were generally greater than the rate of recession at the outlet (Figure 5b). The 1200 m reach was consistently net losing, and the stream had more losing than gaining reaches from 1200 m to 2000 m during experiments at lower base flows (Figure 7). Overall, net losing reaches occurred much less frequently than net gaining reaches, and net losing reaches were only marginally more frequent at lower flows relative to higher flows (2 when $Q(0 \text{ m}) = 101 \text{ L s}^{-1}$, 2 when $Q(0 \text{ m}) = 27 \text{ L s}^{-1}$, 3 when $Q(0 \text{ m}) = 21 \text{ L s}^{-1}$ and 3 when $Q(0 \text{ m}) = 15 \text{ L s}^{-1}$, Figures 6 and 7).

3.2. Bidirectional Channel Water Balance Over Time and Along the Valley

[25] Many stream reaches showed substantial tracer mass and water loss (some well over 10%) during intermediate and lower base flows (Figures 7 and 8). Tracer mass loss in these reaches was infrequently explained by net loss in discharge, indicating concurrent influence of gross gain and gross loss over individual reaches (Figures 8 and 9). In contrast, the largest estimate of tracer mass loss over a reach at higher base flows was only 6% (Figure 7a).

[26] A physical explanation for reaches with positive mass loss (i.e., mass influx, Figure 7) is improbable, because substantial increases in background EC were un-

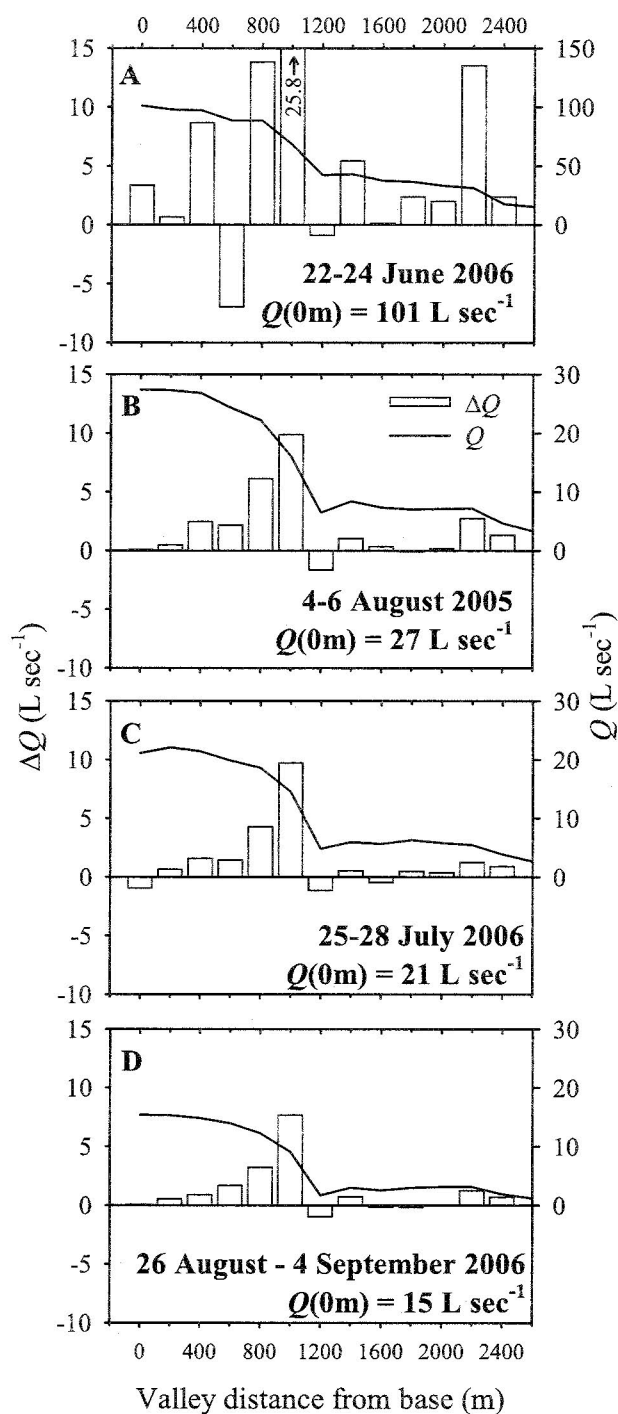


Figure 6. (a–d) Net change in discharge (ΔQ , positive is net gain, negative is net loss) and discharge measured at the end points of each 200 m reach of Stringer Creek (Q). Bar values correspond to the reach immediately upstream from the designated distance (e.g., 0 m datum is for the reach between 0 m and 200 m). Plots are sorted by discharge at 0 m, $Q(0 \text{ m})$ from high to low.

likely during an individual tracer test. Therefore, positive mass loss estimates likely represent other experimental errors. However, there is little evidence of frequent or systematic error of this nature, and the magnitudes of most

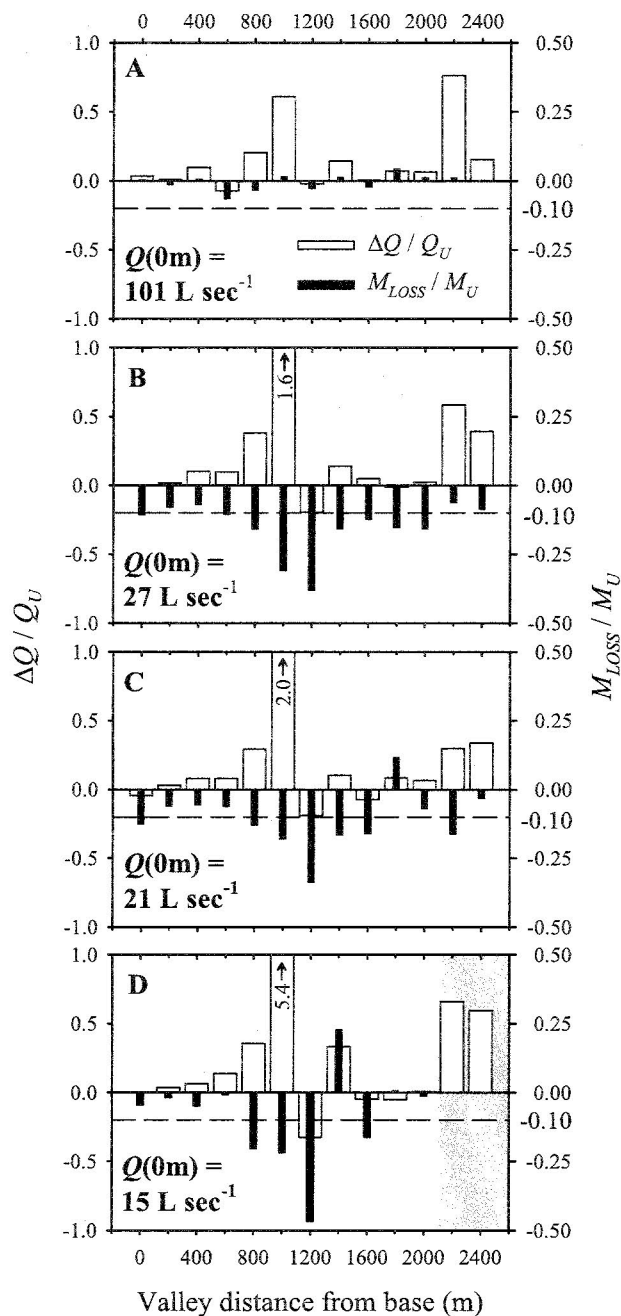


Figure 7. (a–d) Net change in discharge as a fraction of discharge at the top of each reach ($\Delta Q/Q_U$) and the fraction of the upstream tracer release that was not recovered at the base of the reach (M_{LOSS}/M_U , negative). Values correspond to the reach immediately upstream from the designated distance (e.g., 0 m data are for the reach between 0 m and 200 m). Plots are sorted by discharge at 0 m, $Q(0 \text{ m})$ from high to low, and shaded area of Figure 7d indicates where no mass loss data is available due to very low channel flow.

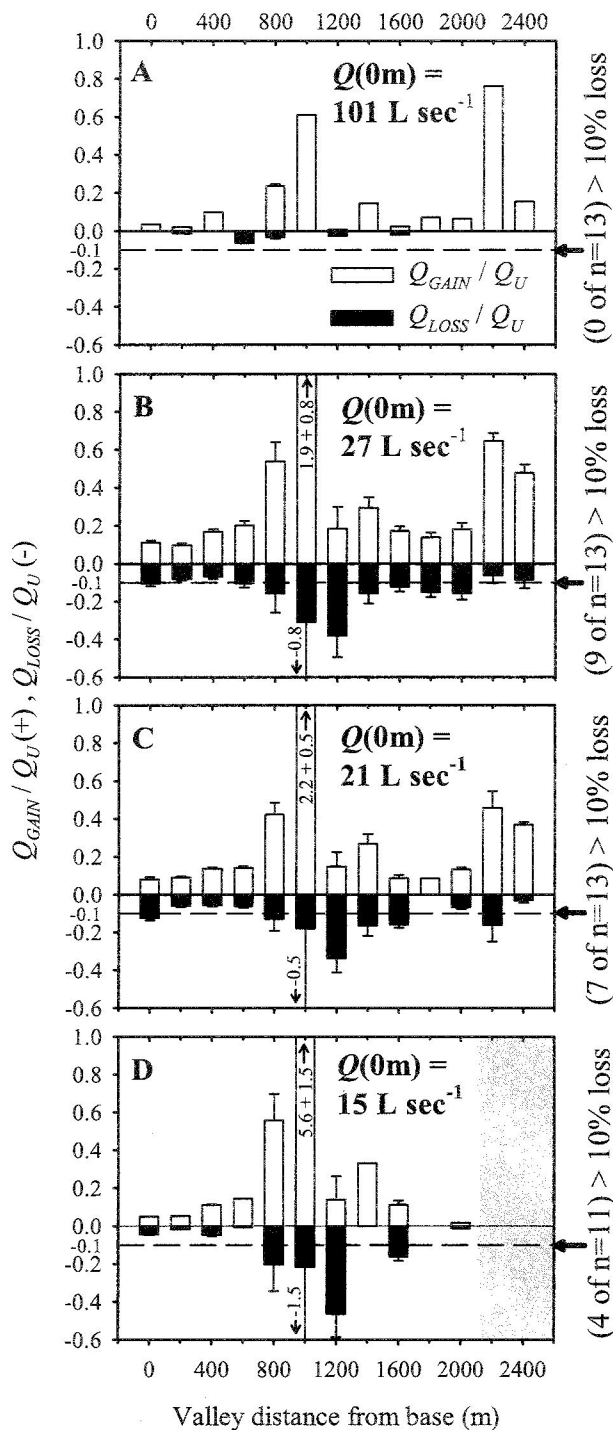


Figure 8. (a–d) Apparent gross hydrologic gain (Q_{GAIN}) and loss (Q_{LOSS}) over 200 m reaches as a fraction of the discharge measured at the top of each reach (Q_U). Plots are sorted by discharge at 0 m, $Q(0m)$ from high to low, and shaded area of Figure 8d indicates where no water balance data are available due to very low channel flow. Bar heights are the minimum estimate of gross exchanges ($Q_{GAIN,MIN} / Q_U$, $Q_{LOSS,MIN} / Q_U$) and single-sided error bars extend to the maximum estimate of gross exchanges (relative to $Q_{GAIN,MAX} / Q_U$, $Q_{LOSS,MAX} / Q_U$).

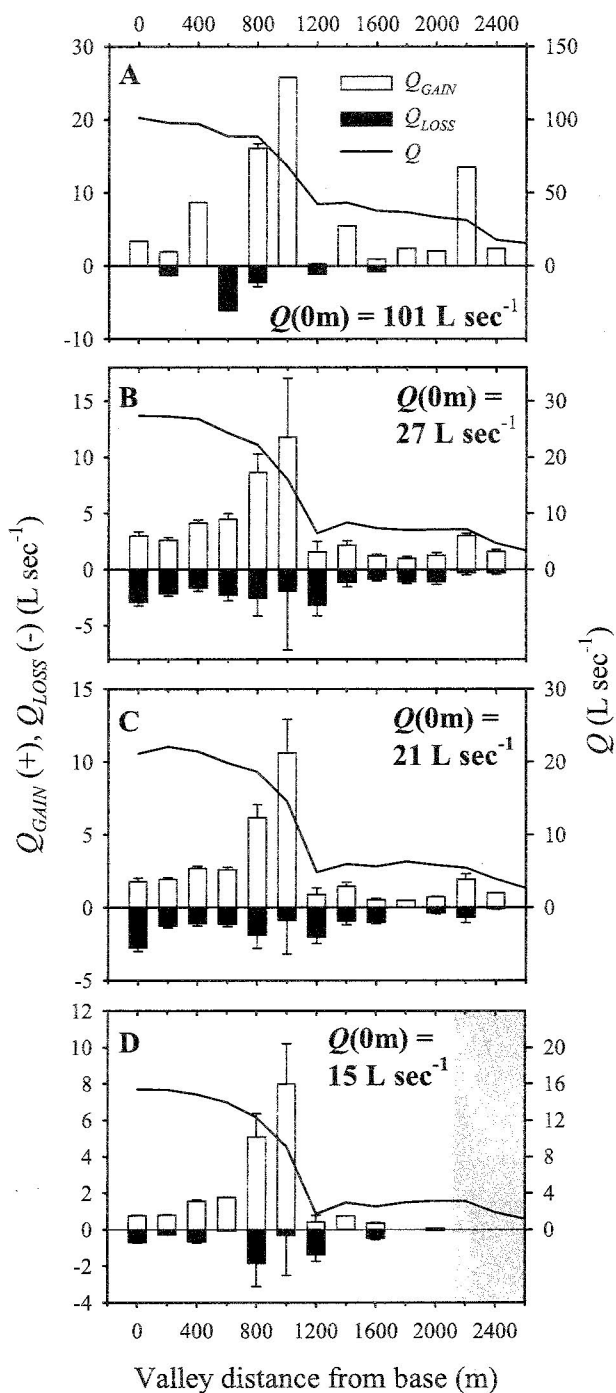


Figure 9. (a–d) Apparent gross hydrologic gain (Q_{GAIN}) and loss (Q_{LOSS}) for each 200 m reach and discharge measured at the end points of each reach (Q). Plots are sorted by discharge at 0 m, $Q(0m)$ from high to low, and shaded area of Figure 9d indicates where no water balance data are available due to very low channel flow. Bar heights are the minimum estimate of gross exchanges ($Q_{GAIN,MIN}$, $Q_{LOSS,MIN}$) and single-sided error bars extend to the maximum estimate of gross exchanges ($Q_{GAIN,MAX}$, $Q_{LOSS,MAX}$).

observed positive mass losses were likely less than general uncertainty in the dilution gauging method (e.g., positive mass losses at higher base flow, Figure 7a), with the exception of two estimates that were likely due to aberrant discharge estimates (1800 m reach when $Q(0\text{ m}) = 21\text{ L s}^{-1}$ and 1400 m reach when $Q(0\text{ m}) = 15\text{ L s}^{-1}$). For the purpose of this study, positive mass loss estimates are assumed to represent negligible gross hydrologic loss ($Q_{\text{Loss}} = 0$), to provide a clear illustration of gross gains and losses in water balance figures (Figures 8 and 9). This simplification substantially affects estimates for only two reaches, and thus has little effect on the general spatial patterns of observed channel water balance.

[27] The frequency and magnitude of both relative (Figure 8) and absolute (Figure 9) gross hydrologic loss increased between tracer tests at higher base flows (Figures 8a and 9a) and those at intermediate and lower base flows (Figures 8b, 8c, and 8d and 9b, 9c, and 9d). The segment from 1200 m to 2000 m demonstrated a seasonal increase in both gross loss and net loss, but gross loss also increased elsewhere, including in many reaches that were net gaining. Gross loss was greater than 10% of upstream discharge in a majority of reaches during tracer tests at intermediate flow (Figures 8b and 8c). Curiously, this pattern did not persist through lower base flows (Figure 8d), though this could reflect less confidence in smaller differences between smaller discharge estimates.

[28] Repeatability of tracer test results was demonstrated by strong similarities between patterns in water balance during similar flow conditions in 2005 and 2006. Different EC probes were used each year and different personnel made on-site decisions regarding specific tracer measurement and release locations, according to the protocol above. Therefore, the similar patterns between the two years were not likely caused by either systematic error or frequent random errors.

4. Discussion

[29] Spatiotemporal variability in channel water balance along Stringer Creek was apparently driven by exchanges with multiple scales of subsurface flow paths. Large scale variability in valley and watershed structure appeared to control the net influence of stream-subsurface exchanges, as indicated by net change in discharge along the valley. Bidirectional channel water balances along the stream indicated further subsurface exchanges at intermediate and low base flows. Gross gain and loss indicated subsurface or hyporheic flow at scales that were not indicated by net change in discharge or transient storage of recovered tracer. Subsurface flow paths associated with gross gains and losses have strong implications for interpreting stream solute fate and transport.

4.1. Net Changes in Channel Flow and Connection With the Catchment

[30] Channel water distribution along Stringer Creek was strongly related to valley and catchment geologic structure and varied during the base flow recession. Channel discharge in most reaches downstream of the transition in valley structure (0 m to 1000 m) tended to decrease at the same rate as the seasonal base flow recession at the outlet (Figure 5b). In contrast, discharge in many reaches upstream

of the structural transition decreased faster than discharge at the outlet, and the dominant direction of stream-subsurface exchange in this segment changed from gain to loss over the recession. Relative contributions from gains in the 1000 m reach increased dramatically with time, such that channel losses and more rapid recession in the upstream segment (1200 m to 2000 m) did not influence the rate of recession downstream from 1000 m. The dynamics in longitudinal discharge distributions reflect a shift in the relative distribution of channel water from upstream to downstream through the recession, where the shift occurred around a structural discontinuity in the valley and watershed systems.

[31] Increasing relative contributions from gains over the 1000 m reach may be explained by multiple interactions with adjacent stream and catchment systems. One explanation is an increasing relative contribution of return flow from large spatial scale hyporheic flow paths recharged by losses along the upstream valley (i.e., underflow *Larkin and Sharp*, 1992]). This explanation is consistent with the hyporheic corridor concept [*Stanford and Ward*, 1993] and with field observations by *Baxter and Hauer* [2000], who found a general pattern of enhanced subsurface inflows to streams where wider valleys are pinched into more constrained valleys. An alternative explanation is a top-down draining effect through the recession, where sources of streamflow from the watershed upstream of 1200 m were depleted before sources of streamflow downstream of 1200 m. Further explanations might be intersection of the downstream valley with larger-scale groundwater systems and differences in hydrologic storage and transmissivity among the bedrock units. These explanations are not mutually exclusive.

[32] Further source water separation studies [e.g., *Covino and McGlynn*, 2007] and water aging studies [e.g., *McGlynn and McDonnell*, 2003] of gains in channel flow would be useful for distinguishing between hyporheic return flows and streamflow generation from hillslopes or groundwater (Figure 1 flow paths A, B, and C). Channel water balance data do not indicate connections of the subsurface flow network among gross losses, gross gains, and watershed sources or sinks. However, bidirectional channel water balances suggest where the end points of large spatial scale or long temporal scale subsurface flow paths are located along the channel of a stream segment. Multiple series of reach water balances show how the various influences of these flow paths change over a base flow recession. Understanding the locations and dynamics of these flow paths would be invaluable to the design of source water separation and water aging studies in parallel with further water balance and transient storage studies. As a whole, this integrative approach may reveal the full distribution of water residence time in the stream system (channel plus hyporheic flow, Figure 1, flow paths B, C, D, and F) as well as a spatial distribution of hydrologic connections between the stream system and watershed (Figure 1, flow paths A and E).

[33] To date, most conventional reach-scale studies of stream tracer dynamics have neglected the presence and distribution of the bidirectional water balance in consecutive reaches, yet an accurate, reach-aggregate model of a valley segment depends on full characterization of reach interaction with adjacent systems and other reaches. Streams are

open systems with respect to the surrounding subsurface and are a relatively easy access point for measuring the spatial distribution of watershed processes, at least those that influence measurable characteristics of the stream. Continued large-scale, longitudinal stream studies will improve the breadth and physical accuracy with which we connect the stream to the adjacent watershed.

4.2. Concurrent Gross Gains and Losses in Stream Reaches

[34] The magnitude, consistency, and frequency of gross losses along Stringer Creek suggest that a substantial amount of channel water was lost to subsurface flow paths during reach-scale tracer tests. As a result, net gains over these reaches frequently underestimated the influence of gross gains, which suggest additional exchanges from the subsurface to the channel (Figure 1). Note that gross loss and gross gain are likely to covary where gross losses are driven by hyporheic flow paths with relatively long transport times (Figure 1, flow path C). Gross loss and gross gain are rarely reported explicitly in stream hydrologic studies (exceptions being Zellweger *et al.* [1989], Ruehl *et al.* [2006], and Covino and McGlynn [2007]), despite their inclusion in popular conceptual [Harvey and Wagner, 2000] and quantitative [Runkel, 1998] models of stream-subsurface interaction. Here, we have demonstrated substantial, and often concurrent, gross loss and gain of channel water in 200 m stream reaches. This distance is of particular relevance given that solute transport experiments are commonly performed in reaches of approximately 200 m in length [Runkel, 2002].

[35] Results from Wondzell [2006] provide a direct example of tracer mass loss from a mountain headwater channel to subsurface flow paths. During a constant-rate release (constant tracer load imposed at the head of the reach), tracer concentrations in some near-stream wells continued to increase after the tracer concentration had come to an apparent steady state (constant load) at the base of the reach. Increasing tracer concentration in these wells suggests unaccounted tracer mass loss from the channel, which might have been measurable as a difference between upstream and downstream tracer loads. The study did not include a mass recovery analysis to determine how much tracer mass was lost to subsurface flow. However, it provides direct physical evidence of subsurface flow paths receiving tracer mass, where the mass loss was not evident in tracer recovery at the base of the reach.

[36] The influence of gross losses along Stringer Creek appeared to increase with the decrease in discharge. This interpretation is consistent with a shift in stream-subsurface hydraulic gradients with the lowering of the surrounding water table [e.g., Harvey and Bencala, 1993; Wroblicky *et al.*, 1998]. However, the observed water balance dynamics may also be the result of limitations in tracer techniques when applied across variable discharge conditions. At high discharge, relatively high flow velocities would provide less time for tracer mass to be removed by a flow path that causes gross loss. Also, a constant gross loss that is measurable at low discharges might be indistinguishable from error at higher discharges due to its relative decrease in influence on channel flow. These limitations are similar to those described for transient storage analyses of channel breakthrough curves [Wagner and Harvey, 1997; Harvey

and Wagner, 2000], and they apply both to temporally variable discharges during the base flow recession and to spatially variable discharges along the stream.

[37] Estimates of gross gain and loss are likely to be important to conclusions about hyporheic flow across many scales of analysis. Ruehl *et al.* [2006] and Covino and McGlynn [2007] reported gross gains and losses of water over net losing river and stream segments. In a river in California (~12 km of river length with discharge on the order of 100s of $L\ s^{-1}$), Ruehl *et al.* [2006] found that dilution during tracer tests was likely explained by gross gains of water from returning hyporheic flow, despite a net loss of water along the valley. Covino and McGlynn [2007] found gross hydrologic exchanges in a net losing stream segment that flowed from an alpine to piedmont topography in Montana (hundreds of meters in length). Between two gauges, they estimated a total gross gain of 69,509 m^3 using chemical source water separation over a spring-summer hydrograph, nearly equaling a total net loss of 72,541 m^3 from the volumetric flow differences between the gauges. Results from Stringer Creek show that bidirectional channel water balance is also quantifiable in smaller streams and over much shorter stream reaches and time scales than those indicated by Ruehl *et al.* [2006] or Covino and McGlynn [2007]. Together, these studies indicate bidirectional water balances are evident across a broad range scales, relative to the mechanisms of water balance, stream sizes, and analytical methods. Furthermore, results from Stringer Creek confirm speculation from Ruehl *et al.* [2006] that gross losses and gains may occur in the presence of net gains, as well as in the presence of net losses as indicated by their study.

4.3. Implications of Channel Water Balance on Interpreting Solute Fate and Transport

[38] Covino and McGlynn [2007] used the term "turnover" to differentiate between the effects of gross exchanges of channel water and the effects of net changes in channel flow. Turnover is the replacement of some portion of a moving volume of channel water as it flows past exchanges with gross losses and gross gains along the stream (i.e., the Lagrangian perspective of a water volume passing through a Eulerian reference frame of multiple discrete reach water balances, in the sense of Doyle and Ensign [2009]). Stringer Creek water balance data emphasize that turnover of tracer-labeled channel water may not be evident in the net change in discharge or the transient storage of tracer mass. Thus, turnover and reach water balance have profound implications on interpreting stream solute fate and transport from solute dynamics in the channel.

[39] Quantifying turnover of channel water due to gross gain and gross loss is particularly important to closing a solute mass balance between upstream and downstream solute loads. Kimball *et al.* [2002] applied dilution gauging and synoptic sampling to characterize net water and solute mass balances in stream reaches of tens to hundreds of meters in length, in an effort to evaluate contaminant mass fluxes in a stream influenced by mine drainage. They acknowledged the potential for hyporheic transport beneath the dilution gauging points [e.g., Zellweger *et al.*, 1989], and subsequently specified that their measurements did not differentiate between channel and subsurface flow in total

downvalley discharge. Therefore, they consider net changes in the sum of stream channel and hyporheic flow paths between paired locations of discharge measurements made by dilution gauging. Stringer Creek data quantitatively demonstrate this conceptual framework, where dilution gauging actually measured an aggregate of channel and subsurface flow paths moving at a broad range of transport distances and velocities. Stringer Creek data also further encourage consideration of bidirectional water balance when making assumptions of mass conservation in dilution gauging, even in relatively short net gaining reaches. The frequency and magnitude of gross losses in Stringer Creek show that use of the net change in discharge alone can underestimate solute mass both leaving and entering along the reach. As a result, consideration of channel water turnover is important to analyses of solute dynamics that depend on mass balances, e.g., geochemical fate and transport of contaminants [Kimball et al., 2002] or biological assessment of whole-stream metabolism [McCutchan et al., 2002; Hall and Tank, 2005].

[40] The effect of a bidirectional reach water balance on channel solutes is a function of the spatial distribution of gain and loss within a given reach. In this study, the distribution of gain and loss within reaches is unknown, and many different distributions of gain and loss can be conceptualized by random selections of reaches overlying multiscaled subsurface flow paths (e.g., Figure 1). Unknown spatial distributions of gain and loss over a reach mean that the degree of tracer dilution by gross gain before gross loss is also unknown. Hence, substantial gross gain over a reach introduces uncertainty in the estimate of hydrologic loss based on a given tracer mass loss. The effect of gross gain on uncertainty in gross loss is evident in data from Stringer Creek, where the reaches with the largest gross gains had the largest ranges of potential gross loss (Figures 8 and 9). As a corollary to these effects of dilution, the frequent assumption that downstream tracer concentrations are insensitive to channel water loss is invalid when gross losses occur upstream of gross gains. In this case, gross loss will effectively decrease tracer concentrations by removing water and tracer mass from the channel before dilution by gross gains.

[41] Both hydrologic gain and loss in streams have potential influence on the incremental quality and quantity of water along stream channels (Figure 9). Stream channel gains and losses, and the consequent addition and removal of channel solute loads, are critical elements to understanding the influence of hyporheic exchange over the contiguous reaches of a stream segment.

5. Conclusions

[41] Channel water balance along Stringer Creek was related to a large-scale transition in valley structure, and changed dramatically through the summer hydrologic recession in flow. Repeated series of water balance data demonstrate that the general decrease in flow at the watershed outlet was due to a dynamic interaction of gains and losses along the stream, where downstream and upstream segment dynamics differed strongly. At intermediate and lower flows, the frequency of reaches that demonstrated concurrent gross gain and gross loss of water suggest that stream-subsurface exchanges would not have been fully

explained by transient storage of recovered tracer and net change in discharge. Gross exchanges of a channel water balance are required to fully couple a stream channel to the surrounding hyporheic zone and hydrologic sources/sinks in the surrounding catchment.

[43] Results from Stringer Creek illustrate how extensive channel water balance data provide useful information to stream hydrology. First, characterization of gross gain and loss along streams would likely improve understanding of multiscaled hyporheic flow. Second, frequent concurrence of gross gain and loss reiterates the importance of mass recovery in interpretation of dilution gauging and solute loading, even in net gaining streams. Third, bidirectional channel water balances along consecutive reaches provide a spatially explicit context for more extensive study of stream-subsurface exchange along stream segments. Finally, we suggest that accurate, independent estimates of downstream discharge be added to whole-reach tracer experiments when it is important to understand the influence of bidirectional water balance on experimental results.

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