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The Effect of Streambed Topography on Surface-Subsurface Water Exchange in Mountain Catchments

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A numerical hydrological simulation suggested that water exchange between stream channels and adjacent aquifers is enhanced by convexities and concavities in streambed topography. At St. Kevin Gulch, an effluent stream in the Rocky Mountains of Colorado, subsurface hydraulic gradients and movement of ionic tracers indicated that stream water was locally recharged into well-defined flow paths through the alluvium. Stream water–filled flow paths in the alluvium (referred to as substream flow paths) returned to the stream a short distance downstream (1 to 10 m). Recharge to the substream flow paths occurred where stream water slope increased, at the transition from pools (<1%) to steeper channel units (5–20%). Return of substream flow paths to the stream occurred where stream water slope decreased, at the transition from steeper channel units to pools. A net water flux calculation is typically used to characterize water and solute fluxes between surface and subsurface zones of catchments. Along our study reach at St. Kevin Gulch the net inflow of water from subsurface to stream (1.6 mL s^{-1} m^{-1}) underestimated the gross inflow (2.7 mL s^{-1} m^{-1}) by 40%. The influence of streambed topography is to enhance hydrological fluxes between stream water and subsurface zones and to prolong water-sediment contact times; these effects could have important consequences for solute transport, retention, and transformation in catchments.

1. INTRODUCTION

Topography has fundamental importance in controlling interactions between regional and local groundwater flow and water exchange between groundwater and surface water of lakes. Slope discontinuities in the land surface create numerous localized groundwater flow paths that are isolated from the regional groundwater flow [*Toth*, 1963]. The length and depth of local groundwater flow paths depend on the amplitude of topographic variations, as well as on geometric and hydraulic properties of groundwater aquifers [*Freeze and Witherspoon*, 1968]. The interaction between lakes and regional groundwater flow paths is strongly influenced by surface topography, aquifer geometry and hydraulic properties, and on depression-focused recharge across the landscape [*Winter*, 1983].

Interactions between streams and surrounding surface and subsurface drainage systems were reviewed and broadly classified by *Eagleson* [1970] and *Schumm* [1977]. The importance of hillslope topographic shape to flow paths and timing of storm flow generation from catchments has been demonstrated by field studies and modeling [*Dunne and Black*, 1970; *Freeze*, 1972; *Beven and Kirkby*, 1979]. In this paper it is our purpose to turn attention to the effect of streambed topography on stream-subsurface interactions. A similar topic has received previous attention; vertical exchange between flowing surface water and underlying gravel or sand was examined theoretically by *Vaux* [1968] and in laboratory flume studies by *Thibodeaux and Boyle* [1987], *Savant et al.* [1987], and *Elliot* [1990]. Our interest was in expanding knowledge of this hydrological link to catchments, where hydrological interaction with the unconfined aquifer of the valley bottom could occur. Our general prediction was that localized subsurface flow paths would exist for stream water, in isolation from the larger-scale system of water groundwater delivery to the stream, and that streambed topography would be a significant control.

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Streambed and stream water surface slopes vary due to local controls imposed by boulder, log, and gravel bar obstructions [*Leopold et al.*, 1964]. Mountain stream reaches are segregated into channel units with distinct streambed and stream water slopes which vary on the spatial scale of 1 to 10 channel widths along the stream [*Grant et al.*, 1990]. Variation in streambed topography, and the resulting variation in stream water slope, influences the potential energy distribution at the boundary between the stream and subsurface and could be a significant control on surface-subsurface interactions in mountain streams. Specifically, we hypothesized that (1) a topographic control on surface-subsurface water exchange is imposed by the slope discontinuity and spacing between stepped-bed units, (2) as a result, stream water travels temporarily in localized subsurface flow paths in the alluvium beneath and to the side of steeper channel units, and (3) localized subsurface flow paths are isolated within a larger system of subsurface water exchange between hillslope groundwater and streams.

In this paper we demonstrate the relation of localized subsurface water flux to streambed topography at a field site and by numerical simulation and show that localized stream-subsurface exchange fluxes, when integrated over a study reach, are comparable in magnitude to the net groundwater inflow to the stream. We suggest that localized stream-subsurface water exchange has significant potential to influence solute transport, biogeochemical cycling, and water quality in catchments.

2. NUMERICAL SIMULATION OF SURFACE-SUBSURFACE WATER EXCHANGE

Local bed slopes of mountain streams are grouped with distinct classes of channel morphology known as channel units [*Grant et al.*, 1990]. Bed slope and stream water slopes increase progressively through pool, riffle, rapid, cascade, and step unit classes. The potential effect of streambed

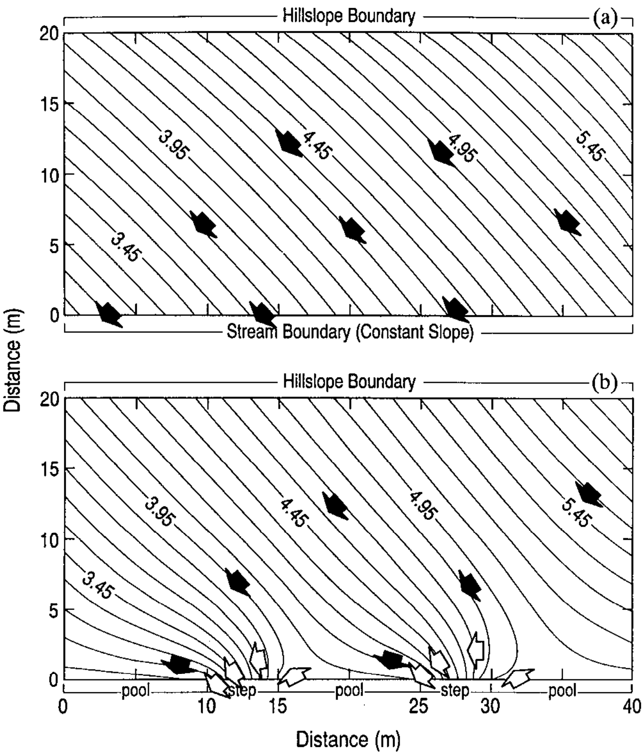

Fig. 1. Model prediction of phreatic surface contours near a stream boundary. The stream boundary is at the bottom of each panel and the direction of streamflow is to the left. (*a*) Constant stream water slope: groundwater inflow pathways (solid arrows) are uniformly distributed along stream. (*b*) Variable stream water slope: a gradual slope in pools (0.33%) alternates with a steep slope (19%) in stepped-channel units. Phreatic surface contours sweep in an arc around steps, indicating that stream water–filled flow paths exist in the subsurface (open arrows). Stream water enters the subsurface upstream of steps, flows around steps, and reenters the stream at the downstream end of steps. Groundwater flow (solid arrows) only enters the stream between steps.

topography was examined using a numerical hydrological model. Stream-subsurface water exchange was simulated by solving Laplace's equation for conditions of horizontal flow in an unconfined aquifer,

$$\frac{\partial^2 h^2}{\partial x^2} + \frac{\partial^2 h^2}{\partial y^2} = 0, \tag{1}$$

where h is the hydraulic head and x and y are horizontal direction coordinates.

A finite difference approximation to (1) was employed using Gauss-Seidel iteration with successive overrelaxation for the solution method. Heads on the boundaries were prescribed using a linear increase in slope that was equivalent in the directions upstream and away from the stream (positive x and y directions, respectively). The initial, simplified simulation ignored the stepped-bed morphology because a constant slope was used to define heads along the stream boundary. The result is that groundwater flow enters the stream without distortion at the stream boundary (Figure 1*a*). The second simulation was unchanged except for the specification of a more realistic stepped decrease in head along the stream boundary. The gradient in hydraulic head at the stream boundary was adjusted so that it varied between a gradual (0.33%) and steep (19%) slope, without changing the average gradient at the stream boundary. The calculated equipotentials for this case sweep in arcs around steps (Figure 1*b*), indicating that stream water flows into the subsurface upstream of steps and then returns to the stream at their downstream end. The simulated substream flow paths are embedded within the larger-scale flow system of groundwater transport toward the stream; hillslope groundwater inflows are segregated such that groundwater enters the stream in only in zones between steps.

3. STUDY SITE

St. Kevin Gulch is located in the upper Arkansas River drainage basin on the east side of the continental divide near Leadville in Colorado. Past research has been conducted along a 257-m reach of stream in the lower quarter of the catchment (Figure 2). Site names and downstream distances are referenced from a zero point established for several studies within the Upper Arkansas Toxic Substance Hydrology Project [*Kimball et al.*, 1988; *McKnight et al.*, 1988].

St. Kevin Gulch is a third-order, gravel bed, stream with an average slope of 6.7% within this study reach. Pools with gradual water surface slopes (<1%) alternate with steeper channel units (riffles, rapids, cascades, and steps) that have slopes of 20% or greater. A 36-m subreach was chosen without tributaries or seeps for intensive subsurface investigations. This subreach gains water perennially from subsurface drainage, as opposed to the situation further down the valley where the stream loses water to the subsurface in late summer [*Zellweger and Maura*, 1991].

Within the instrumented subreach an alluvial deposit of mixed cobble, gravel, and sand is present near the stream. The alluvium extends 5 m laterally on either side of the stream and is approximately 2 m in depth. The alluvium is composed of a large fraction of sand and fine gravel in the size range between 0.5 and 5 mm with pebbles and cobbles interspersed throughout. Soils on the lower hillslope are 0.5–3 m thick; they are composed of a surficial organic horizon above a sandy loam which grades to a clay loam at depth. Soils are underlain by a bedrock of schist and gneiss [*Singewald*, 1955].

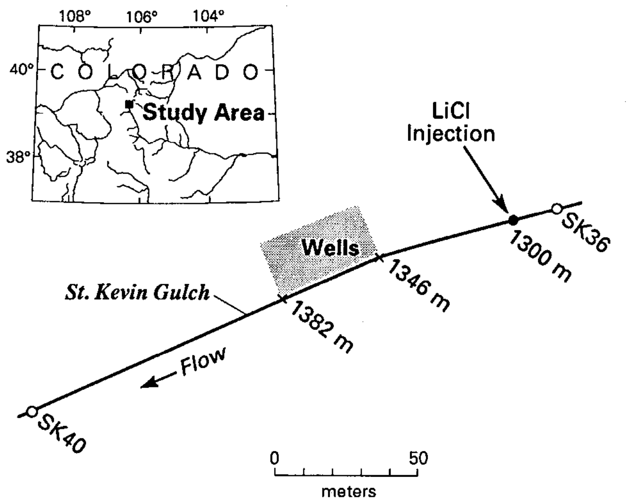

Fig. 2. Site location and study reach, St. Kevin Gulch, Lake County, Colorado. A subreach with perennial subsurface inflow from subsurface to stream was selected for intensive subsurface investigation (shaded here and shown in detail in Figure 4).

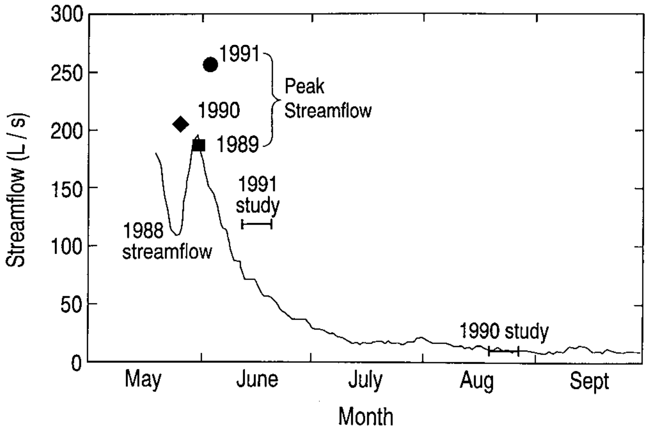

Fig. 3. Streamflow at location SK40, St. Kevin Gulch. Brackets indicate average streamflow during study periods in June 1991 and August 1990.

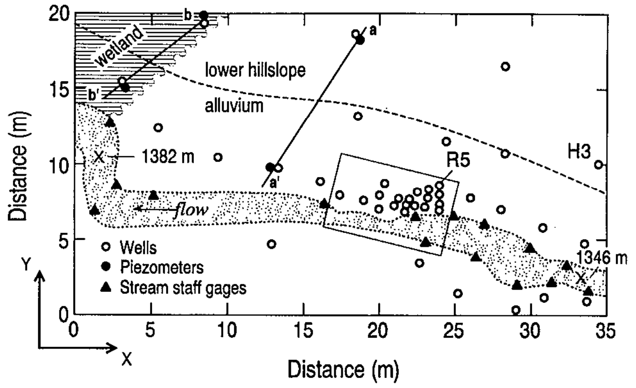

Fig. 4. Instrument location map and sites of subsurface tracer injections, St. Kevin Gulch. The stream channel is coarsely stippled and flow is to the left. Upstream and downstream limits for calculating surface-subsurface water fluxes are denoted with an X and labeled with the distance downstream from a common zero point. The rectangular inset box shows the limits of subsurface investigation in August 1990 (1356 to 1362 m). Potassium bromide (KBr) was injected into well R5 in August 1990 and well H3 in June 1991.

Our work was conducted in two periods, June and late August, which provided a contrast between high and low streamflow and wet and dry conditions on the lower hillslope. Most precipitation occurs as snowfall at high elevations in the Colorado Rocky Mountains. Peak streamflow occurs in May and is followed by a 3–4 month recession in streamflow with minor flow pulses caused by rainfall. Streamflow differed by tenfold between the experimental periods in which our work was conducted, spanning a large proportion of the typical annual range in streamflow at this site (Figure 3). No rainfall was recorded during either study.

4. Methods

4.1. *Field Instrumentation*

Investigations began in 1990 with installation of four nonpermanent stream staff gages, 19 cased wells, and six shallow sampling pits in exposed gravel bars. The 1990 array was compact, extending 6 by 3 m at streamside. In 1991 the network was expanded to cover an area of 35 m along the stream and up to 15 m away on the right bank. Sixteen additional wells, four piezometers, and ten permanent staff gages were added to the network in 1991. In 1992, eight additional wells and staff gages were installed along the left bank of the stream (Figure 4).

Well and piezometer emplacement was accomplished by drilling 4-inch (10 cm) diameter holes with a gas-powered auger or by excavating with hand tools to a depth below the phreatic surface. Holes were cased with 2-inch (5-cm) (nominal) polyvinyl chloride well screen and extension pipe. Well screens are 0.5–1.5 m in length, emplaced 30–70 cm below the water table. Holes were backfilled with coarse native sand. Piezometer screens were 15 cm long; they were emplaced 100–200 cm below the water table and were backfilled with coarse sand around the screen to a depth of 15 cm above the screen. A 30-cm-thick plug of bentonite pellets was installed on top of coarse sand backfill; the area above the bentonite plug was backfilled with native sediment. Staff gages were installed in quiescent zones of the stream channel next to the bank.

Horizontal positions and vertical elevations of well and staff tops were mapped using standard surveying techniques. A graduated rod, fitted for sensing a change in electrical resistance when water was contacted, was used to measure hydraulic head in wells and piezometers. Where staff gages were present, the hydraulic head at the stream boundary was estimated from staff readings. Combined accuracy and precision of hydraulic head measurements was estimated to be ±0.25 cm.

4.2. *Solute Tracer Experiments*

Two compounds were used to trace water flow: lithium chloride (LiCl) and potassium bromide (KBr). Transport experiments and sorption experiments conducted elsewhere indicated that Br is useful as a conservative tracer in soils and groundwater [*Bowman*, 1984; *Levy and Chambers*, 1987]. Both Cl and Li are useful conservative tracers for water flow in the Snake River, Colorado, an acidic mountain stream similar in chemistry to St. Kevin Gulch [*Bencala et al.*, 1990; *McKnight et al.*, 1988]. In our study, Br- and Cl-labeled solutions were injected at a constant rate and concentration into groundwater and stream water, respectively. Injection procedures and water sampling from streams and wells followed the general procedures outlined by *Triska et al.* [1989]. The locations and durations of injections are summarized in Table 1. Cl was selected over Li as the conservative ion to trace stream water movement

TABLE 1. Summary of Tracer Injections at St. Kevin Gulch, Colorado

Injection Type	Tracer	Input Location	Injection Duration, days	Start Day
		*1990**		
Stream	LiCl	1300 m	4	Aug. 20
Subsurface	KBr	R5	2.5	Aug. 21
		1991†		
Stream	LiCl	1300 m	4	June 13
Subsurface	KBr	H3	4	June 11

*Average streamflow at SK36 during injection was 10.6 L s^{-1}.
†Average streamflow during injection was 118.3 L s^{-1}.

in order to utilize available automated instrumentation for tracer quantification. Water samples were analyzed for Br and Cl using ion chromatography.

4.3. *Subsurface Water Velocity and Hydraulic Conductivity*

Water velocity and hydraulic conductivity in the valley bottom alluvium were estimated from bromide transport information. In the 1990 study, transport of Br-labeled water was measured from the point of injection (well R5) to the downstream well that received bromide at highest concentration (well R22). Average water velocity was estimated by dividing the horizontal flow path length by the travel time of bromide to well R22. By appearance, bromide concentrations at well R22 were distributed symmetrically in time around the time of the peak concentration. For this study an estimate of the travel time to well R22 was calculated as the difference between the time to peak bromide concentration and the midpoint of the period of injection. Although not generalizable, this calculation is reasonable for our study given the relatively symmetric distribution of bromide concentrations versus time and a period of injection into well R5 that was shorter than the transport time to well R22. Hydraulic conductivity was computed from the product of water velocity and porosity divided by the average hydraulic gradient along the flow path. A porosity of 0.3 was used for this calculation; this value was chosen from a range of 0.25 to 0.40 for gravel [*Freeze and Cherry*, 1979].

4.4. *Quantification of Surface-Subsurface Water Exchange*

A stream reach with steady flow is considered where water is exchanged between the stream and subsurface. The mass balance for water in the stream is

$$\frac{dQ}{dl} = q_i, \tag{2}$$

where Q is streamflow, l is the downstream space coordinate, and q_i is inflow, expressed as a water flux across the streambed per unit length of stream. To calculate inflow, we assumed that subsurface flow adjacent to the stream was nearly horizontal in a homogeneous and isotropic aquifer. We adopted the Dupuit assumption that the slope in the phreatic surface is a good approximation of the horizontal hydraulic gradient. The specific discharge of water in the subsurface is

$$\mathbf{q} = -K\left(\frac{\partial h}{\partial x}\mathbf{i} + \frac{\partial h}{\partial y}\mathbf{j}\right), \tag{3}$$

where h is the hydraulic head on the phreatic surface, K is the average hydraulic conductivity, and x and y are horizontal distance coordinates whose directions are shown on Figure 4. Assuming that subsurface fluxes are horizontal and symmetrical on either side of the stream, the inflow of water to the stream is

$$q_i = \frac{2}{L}\int\int \mathbf{q}\cdot d\mathbf{s}, \tag{4}$$

where ds is a differential area element, situated vertically beneath one side of the stream, with an outward normal vector pointed toward the stream. L is the total length of stream reach under consideration.

Phreatic surface elevations measured at St. Kevin Gulch on June 18, 1991, were interpolated onto a closely spaced grid (0.5 m) by kriging. Inflow was calculated using interpolated head data from

$$q_i = \frac{2}{L}\sum_{n=1}^{N}\left[-K\left(\frac{\delta h}{\delta x}\mathbf{i} + \frac{\delta h}{\delta y}\mathbf{j}\right)\cdot\Delta\mathbf{s}\right], \tag{5}$$

where each area element n is positioned vertically in the sediment at a node on the right bank of the stream, extending downward 2 m to the bottom of the alluvium and laterally in a direction that is locally parallel with the stream bank; 65 area elements resulted. The partial derivatives were estimated for each element by central difference approximation.

4.5. *Evaluation of Assumptions Used to Quantify Stream-Subsurface Water Exchange*

Two potential errors arose from assuming horizontal flow in the calculation of water fluxes. Horizontal gradients were computed from heads on the phreatic surface, which generated an error that depended on the phreatic surface slope [*Bear*, 1972]. That error was less than 1% for our phreatic slopes. Flux computations were also slightly in error because depth-averaged heads differed slightly from heads on the phreatic surface. Within the 36-m subreach at St. Kevin Gulch, vertical hydraulic gradients were slightly upward or downward (0.030 and −0.028, transect a–a′, Figure 4); errors of about 3% in discharge estimates could have resulted, according to an estimate made using *Bear*'s [1972] procedure.

Phreatic slopes and piezometer data supported the use of Dupuit assumptions to calculate subsurface flow across the stream boundary in the instrumented reach. Outside the 36-m subreach in a wetland just downstream, vertical hydraulic gradients were significant (0.21 and 0.38, transect b–b′); upward flow was occurring beneath the wetland. The significant effect of vertical gradients on subsurface flow calculations in the 36-m subreach was apparently confined only to the wetland vicinity at the downstream end of the study reach.

In addition to assuming that flow was horizontal at the stream boundary, flux calculations assumed that a no-flow boundary existed beneath the stream. Published data on hydraulic gradients beneath mountain streams typically show horizontal gradients on both sides of streams that converge into vertical gradients directly beneath the stream [e.g., *Gburek and Urban*, 1990]. The assumption that a no-flow boundary existed beneath the stream centerline at St. Kevin Gulch was supported by field data collected in May 1992. Additional wells and staffs installed on the left bank indicated symmetry between hydraulic heads on either side of the stream in the study reach.

5. RESULTS

5.1. *Convective Exchange of Water and Solute Between Surface and Subsurface*

Assuming horizontal flow in a homogeneous and isotropic alluvium, the phreatic surface suggests that stream water

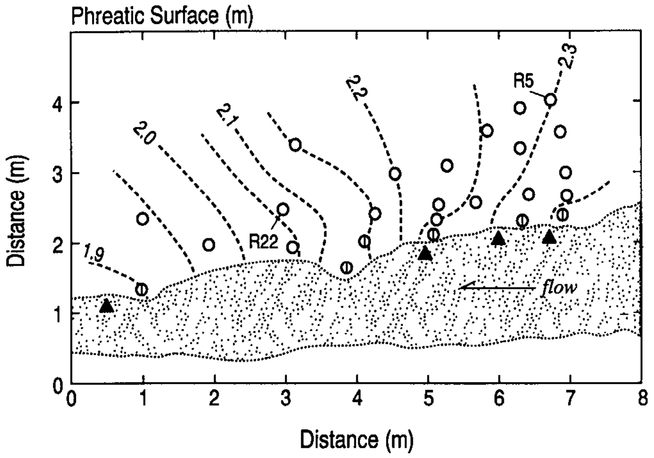

Fig. 5. Map of phreatic surface, St. Kevin Gulch, August 22, 1990. The network of wells pictured here is shown in the inset box of Figure 4. Dashed curves are contours of equal phreatic surface elevation in meters above an arbitrary elevation. Open circles show the position of cased wells. Circles filled with a vertical line locate shallow pits that were dug in gravel bars at the streams edge. Solid triangles are staff gages. In August 1990, KBr was injected into well R5 and monitored for travel time at well R22.

flowed into the subsurface at the upstream end of the 1990 array and that subsurface water flowed into the stream at the downstream end (Figure 5). Here we refer to these flow paths, where stream water flows temporarily through the alluvium and then returns to the stream, as "substream" flow paths.

Stream water and groundwater tracers moved along paths consistent with streamlines suggested by the phreatic surface. Over the 4-day period of stream water tracer (LiCl) injection in 1990, chloride moved a distance exceeding 2 m into the subsurface alluvium at the upstream end of the array but not at the downstream end (Figure 6*a*). The instream chloride injection was terminated on August 24. On August 27 (7 days after the beginning of the injection), chloride was still present at high concentration in the subsurface (>6 mg L^{-1}) but was detectable at only very low concentrations (0.4 mg L^{-1}) in the stream channel (Figure 6*b*). The plume of chloride-labeled water had moved farther into the subsurface zone at the upstream end of the well network by August 27 and had begun to curve back toward the stream at the downstream end of the network (Figure 6*b*).

The subsurface bromide injection in 1990 began on August 21. On August 24, bromide had moved parallel to the stream and had begun to curve back toward the stream. By August 27 the plume of bromide had reached the stream boundary at the downstream end of the well network (Figure 7). Hydraulic conductivity was estimated from bromide travel time in the subsurface and average hydraulic gradient along the flow path. The peak in tracer concentration occurred at well R22, located 4.4 m downstream of the injection well, after 4 days (Figure 8). The resulting calculation of hydraulic conductivity (1.1×10^{-2} cm s^{-1}) is in the range expected for a sand [*Freeze and Cherry*, 1979].

5.2. *Spatial Segregation of Substream Flow Paths From Hillslope Groundwater Inflow*

Stream water–filled flow paths in the subsurface were nested within a larger system of groundwater flow to the stream. Although groundwater on the lower hillslope moved uniformly toward the stream, phreatic contours (Figure 9*a*) and the inferred flow directions near the stream (Figure 9*b*) suggested that hillslope groundwater only entered the stream at locations between substream flow paths. Segregation of groundwater and substream flow paths is illustrated by the movement of the bromide tracer in 1991. Bromide-labeled water that was injected into well H3 appeared at streamside wells (Figure 9*a*) but turned away from the stream and moved around the outside margin of a substream zone before entering the stream (Figure 9*b*).

The direction and magnitude of water exchange between stream and subsurface varied on a spatial scale of 1–10 m along the 36-m subreach. Inflows for individual boundary elements specified in (5) are plotted versus distance downstream in Figure 10, along with the stream surface elevation. Local maximums and minimums in inflow corresponded respectively to concave and convex breakpoints in the slope of the stream water surface. Flow from the sediment to the stream occurred at transitions from steps to pools. Flow from the stream to the sediment occurred at transitions from pools to steps (Figure 10).

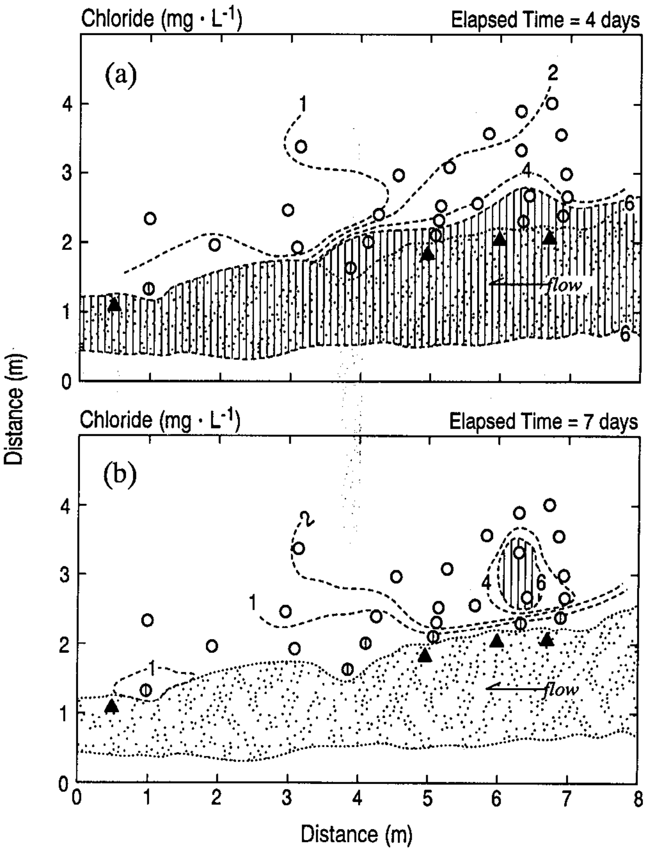

Fig. 6. Map of distribution of stream water tracer (chloride) in surface and subsurface water, St. Kevin Gulch, August 24 and 27, 1990. Chloride was injected 56 m upstream beginning August 20; elapsed times are times since injection began. (*a*) Chloride on day 4, just prior to cutoff of the instream injection and (*b*) chloride on day 7. Dashed curves are contours of equal Cl concentration in milligrams per liter; background chloride concentration was 0.6 mg L^{-1} in stream and subsurface. Hatched area indicates interval of highest concentration (>6 mg L^{-1}).

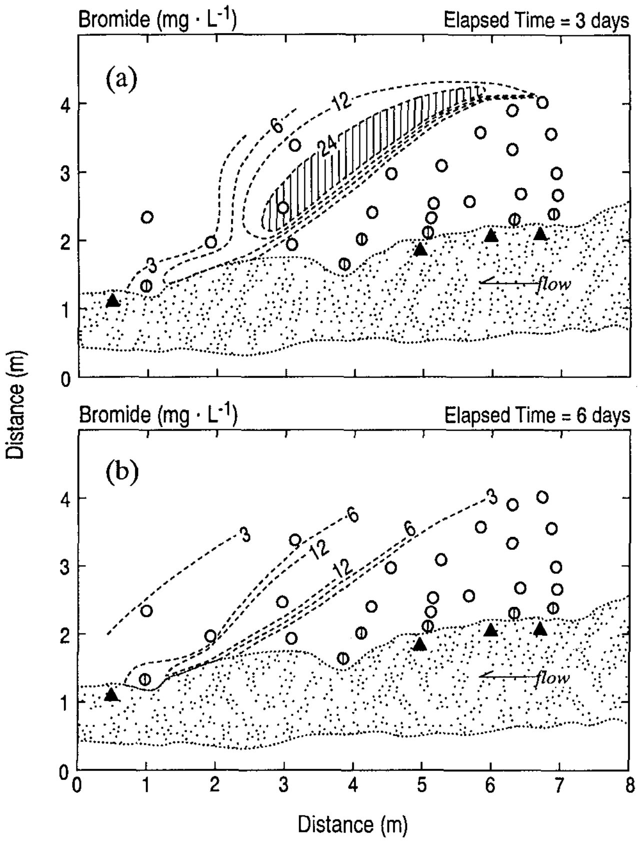

Fig. 7. Map of distribution of subsurface water tracer (bromide), St. Kevin Gulch, August 24 and 27, 1990. Bromide was injected into well R5 (Figure 4) beginning August 21; elapsed times are times since injection began. (*a*) Bromide on day 3, 10 hours after injection stopped and (*b*) bromide on day 6. Dashed curves are contours of equal Br concentration in milligrams per liter; background bromide concentration was less than 0.01 mg L^{-1} in stream and subsurface water prior to injection. Hatched area indicates interval of highest concentration (>24 mg L^{-1}).

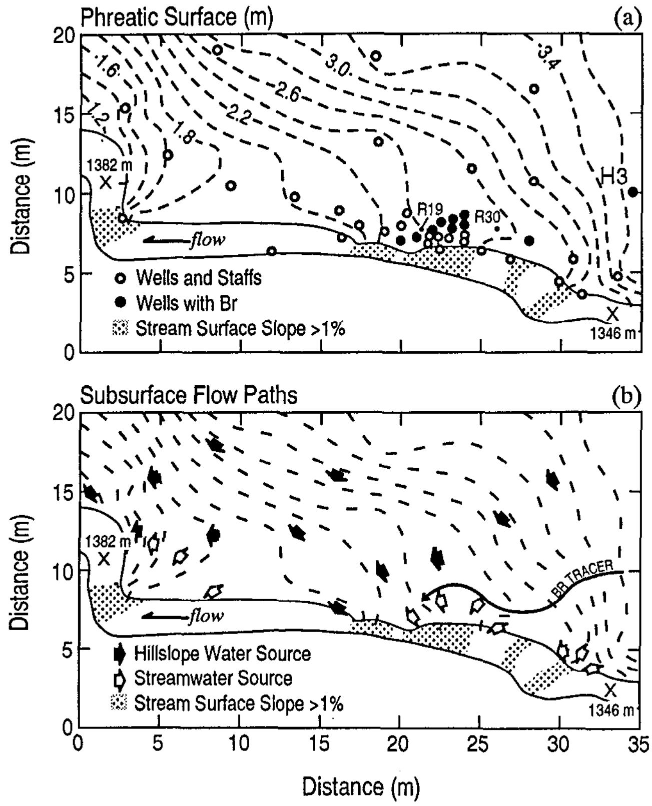

Fig. 9. Isolation of hillslope water and stream water–filled zones in the valley bottom alluvium, St. Kevin Gulch, June 18, 1991. (*a*) Phreatic surface contours and bromide tracer observations. Dashed curves are phreatic surface contours in meters above an arbitrary datum. Bromide injected into well H3 reached the indicated wells (solid circles) by June 29, 1991, at significant concentration (5% or more of highest measured concentration). Well R30 received the highest-bromide concentration (196.8 mg L^{-1}), and well R19 received the second highest concentration (123.3 mg L^{-1}). (*b*) Inferred subsurface flow paths. Solid arrows indicate flow paths for water originating as groundwater flow from the hillslope; open arrows indicate flow paths recharged with stream water. Flow directions assume a homogeneous and isotropic aquifer. The solid curve labeled BR TRACER connects the location of Br injection with wells that received the highest and second highest concentrations of tracer.

5.3. *Gross and Net Inflow Comparisons*

Phreatic surface contours provided evidence for the existence of two sources of water to the stream in our study reach, that is, hillslope groundwater and stream water that had been temporarily routed through substream flow paths in the alluvium (Figure 9*b*). In order to partition these two flow paths in the mass balance on water, (2) was expanded:

$$\frac{dQ}{dl} = q_i = (q_L^{\text{in}} + q_S^{\text{in}}) - q_S^{\text{out}}, \tag{6}$$

where subscript L refers to exchange with hillslope groundwater and subscript S refers to exchange with substream zones. The term $(q_L^{\text{in}} + q_S^{\text{in}})$ in (6) is the gross inflow to the stream and the term q_S^{out} is outflow from the stream into substream flow paths of the alluvium. If the reach under consideration is long relative to the average substream flow path, then substream flows that leave the channel will return to the channel in that reach, so that $q_S^{\text{in}} - q_S^{\text{out}} = 0$, and q_L^{in} equals the net inflow q_i.

Net and gross inflows were integrated for the 36-m subreach using (5); the gross inflow was calculated by summing fluxes only for boundary segments where inflow was positive. From (6) the difference between the gross inflow and the net inflow fluxes is the substream flux q_S^{out}. The subreach

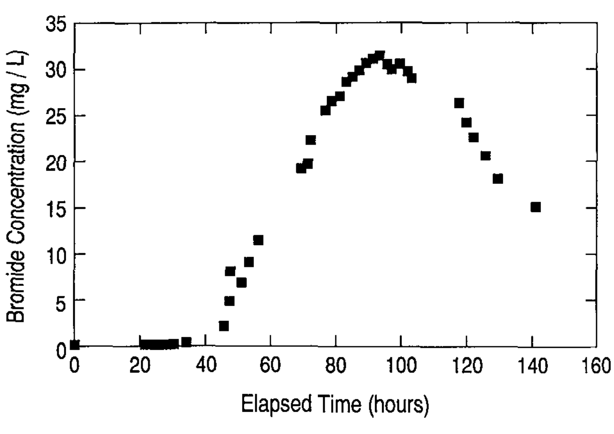

Fig. 8. Bromide concentration at well R22 versus elapsed time since injection began at well R5. The injection lasted 2.5 days.

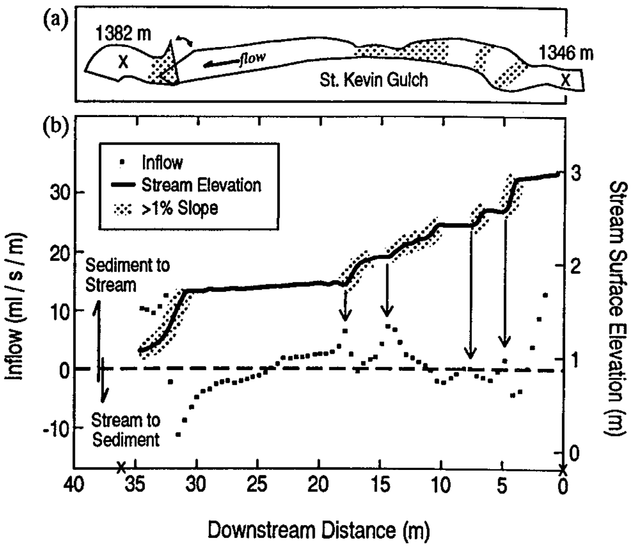

Fig. 10. Inflow from sediment to the stream versus location of stepped channel units, St. Kevin Gulch, June 18, 1991. (*a*) Overhead view of instrumented subreach; general locations of stepped-channel units with slope >1% are shaded. (*b*) Inflow and stream elevation versus downstream distance. Maximum inflows occurred locally where the stream water elevation profile is concave (shown with arrows); minimum inflows occurred at points where the profile is convex.

was gaining water on June 18, 1991, as indicated by a positive net inflow; however, gross inflow was approximately 2 times larger than net inflow in the subreach. The difference between gross and net inflows is the substream flux, which was nearly equal to the net inflow of hillslope groundwater to the stream (Table 2).

6. Discussion

6.1. *Routing of Stream Water Through Substream Flow Paths*

Our study showed that water and solute are exchanged in both directions between stream and subsurface in an effluent mountain stream. Stream water was locally recharged to the subsurface and routed in flow paths that returned to the stream a short distance downstream. Hillslope groundwater entered the stream at locations between substream flow paths. The principal evidence for this segregated system of flow paths were streamlines inferred from hydraulic gradients and paths of tracer movement that agreed with those streamlines (Figures 5–10).

TABLE 2. Reach-Integrated Surface-Subsurface Flux Calculations, St. Kevin Gulch, June 18, 1991

Parameter	Value
Reach, m	1346–1382
Inflow Fluxes (mL s^{-1} m^{-1})	
Gross ($q_L^{in} + q_s^{in}$)	2.7
Net (q_L^{in})	1.6
Substream (q_s^{in})	1.1
Normalized Inflow	
Substream ($q_s^{in}/(q_L^{in} + q_s^{in})$)	0.4

Inflow from substream calculated by difference between gross and net inflows.

Two-way convective exchange of water between the stream and subsurface was hypothesized by *Kennedy et al.* [1984] and *Jackman et al.* [1984]. Our detailed hydrological investigation adjacent to an effluent stream verified the existence of local zones in the valley bottom alluvium of mountain catchments that are recharged by stream water. The term used by *Kennedy et al.* [1984] and *Jackman et al.* [1984] to describe stream-subsurface water exchange was "underflow"; this term has also been used to describe subsurface flow that does not necessarily intersect nearby streams [e.g., *Savant et al.*, 1987]. *Triska et al.* [1989] referred to the subsurface zone in streambed gravel that received significant quantities of tracer-labeled stream water as the "hyporheic zone." Our desire was to quantify surface-subsurface water fluxes and, in particular, to partition the flux component that originated in the stream and returned to the stream. When integrated for a specific length of channel, we refer to this flux component as the "substream" flux.

6.2. *Identification of a Topographic Control on Surface-Subsurface Water Exchange*

We observed that alternating channel morphological units of different streambed slope influenced convective exchange between the surface and subsurface in our mountain stream. The primary pieces of evidence that supported our conclusion were the qualitative correlations between subsurface hydraulic gradients with tracer movements and the relation between calculations of positive and negative inflows with concavities and convexities in stream water slope. Conclusions based on field data were further supported by results of the numerical hydrological model.

Convective exchange of surface water with water in bed sediment has been investigated previously in laboratory flumes. Localized pressure variations over wavelike bed forms were found to cause significant vertical convection between surface and subsurface zones [*Savant et al.*, 1987]. Pressure variations of 100–1000 Pa were typically observed along sand waves. Convective velocities in the subsurface up to 10^{-3} cm s^{-1} were observed; subsurface flow path lengths were of the order of the wave height (about 5 cm). The sand wave bed form model used in the laboratory is probably an excellent model of bed roughness for many rivers [*Dingman*, 1984], yet wave bed forms represent only one type of bed roughness that is present in mountain streams. Other important roughness features could include gravel bars [*Prestegaard*, 1983], large boulders [*Wiberg and Smith*, 1991], alternating pools and boulder steps [*Whittaker*, 1987], and woody debris obstructions [*Keller and Swanson*, 1979].

Bank storage sometimes explains the appearance of stream water solutes in alluvial groundwater next to what normally are effluent streams; however, a sequential increase and decrease in stream stage is required to reverse the direction of water exchange from stream to subsurface [*Cooper and Rorabaugh*, 1963; *Pinder and Sauer*, 1971]. Relatively short, flow-through pathways for stream water in the subsurface at St. Kevin Gulch, that were seasonally persistent, cannot be explained by bank storage.

Stream curvature could influence surface-subsurface fluxes. This effect was considered at St. Kevin Gulch in flux calculations that account for curvature, but the effect was small compared to the influence of streambed topography. Variation in hydraulic properties is another possible control which was not considered in our study. An average hydraulic conductivity was used in our flux calculations, which was based on movement of a tracer along a 6-m flow path. As a result, larger-scale variation in hydraulic conductivity along our 36-m reach would not have been detected, and heterogeneities may have influenced water exchange at our site. However, hydraulic heterogeneities were not needed to produce substream flow around stepped channel units in the numerical simulation, in a pattern similar to what was observed in the field. We feel therefore that our general conclusions about the importance of streambed topography are correct. Nevertheless, the influence of heterogeneous hydraulic properties of the alluvium on surface-subsurface water exchange is a high priority to be considered in future research.

6.3. *Three-Dimensional Nature of Substream Flow Paths*

At St. Kevin Gulch we demonstrated horizontal exchange between the stream and the adjacent alluvium with detailed hydrometric analysis and tracer experiments. Conceptualizing the flow as predominantly horizontal outside the stream boundaries, and predominantly vertical beneath the stream, was a useful approximation for this shallow aquifer that simplified quantification of surface-subsurface water exchange. A fully three-dimensional conceptualization of substream zones in mountain streams may be important in future work at St. Kevin Gulch and may be critical at streams that are surrounded by deeper unconfined aquifers.

Vertical convective exchange was previously inferred in the streambed of a Michigan stream from the vertical displacement of temperature isotherms and chemical parameters beneath a riffle [*White et al.*, 1987; *Hendricks and White*, 1991]. Similar measurements of vertical hydraulic gradients and tracer movement beneath the streambed were impractical at St. Kevin Gulch. However, at another mountain stream, Little Lost Man Creek, California, we have measured both horizontal and vertical hydraulic gradients in a pool just upstream of a steep cascade (J. W. Harvey, unpublished data, 1992). Both horizontal and vertical flow were out of the stream and into the sediment, a finding consistent with a control imposed by the local increase in stream water slope at that location.

6.4. *General Importance of Topographic Control on Surface-Subsurface Water Exchange*

Spatially segregated zones of groundwater recharge and discharge were first recognized at the catchment scale [*Toth*, 1963; *Freeze and Witherspoon*, 1968; *Winter*, 1983]. *Dunne and Black* [1970], *Freeze* [1972], and *Beven and Kirkby* [1979] subsequently investigated the importance of topographic variation at the hillslope scale. In the present study we found that streambed topography influences groundwater–surface water interactions at a scale as small as channel bed slope units by creating localized pathways for streamflow in the subsurface that return to the surface a short distance downstream. Small-scale topographic controls on surface-subsurface water exchange can potentially affect solute transport processes at larger scales.

Stream-subsurface water exchange is generally calculated by streamflow differencing, using the steady state continuity equation for streams with groundwater inflow:

$$q_i = \frac{1}{L}(Q_d - Q_u), \tag{7}$$

where subscripts d and u represent downstream and upstream locations and L is the length of stream between them. This calculation estimates only the net water exchange and ignores fluxes of stream water into and out of substream flow paths. We showed that the return of water from substream flow paths to the channel can be a significant component of the gross inflow to streams. At St. Kevin Gulch the gross inflow of water to the stream exceeded the net inflow by nearly twofold (Table 2).

A comparison between (6) and (7) illustrates that the mass balance on water for a stream reach is not affected by considering gross exchange fluxes in addition to net inflow (because outflow and inflow from substream zones cancel in (6)). However, the unsteady mass balance on solute could be substantially affected by the choice between (6) and (7) for a water balance equation. For example, the Cl tracer in our study that was injected into stream water was retained in the substream flow paths and released slowly to the stream for days following the termination of the instream injection. Such behavior could result in the long tails of elevated tracer concentration versus time that are frequently observed in mountain stream tracer studies [*Kennedy et al.*, 1984; *Bencala et al.*, 1984; *Castro and Hornberger*, 1991]. Our work suggests that substream zones in the valley bottom alluvium of mountain catchments are a principal zone for hydrological storage of solutes. High specific surface areas of alluvial sediment in substream flow paths could further enhance the retention of reactive stream solutes; such surface-subsurface solute interactions would not be predicted by a hydrological model with only net inflow.

6.5. *Temporal Persistence of Substream Flow Paths*

Substream flow paths were persistent across late spring and summer, spanning a tenfold range in streamflow and a transition from high catchment wetness to low wetness. Qualitatively, it was observed that the high streamflows in June were not high enough to swamp the effects of bedslope changes on stream water slopes. However, extreme increases in hydraulic heads in the alluvium and on the lower hillslope affected the presence of substream flow paths.

On June 1, 1991, fresh snow fell on bare ground at the study site, followed immediately by rapidly warming temperatures; streamflow also peaked about that time (Figure 3). Streamflow on June 3, 1991, was 173 L s^{-1}, a flow 45% higher than the average streamflow during the 1991 study that began about 10 days later. Phreatic surface potentials on June 3, 1991 (Figure 11), were roughly perpendicular to equipotentials plotted for June 18, 1991 (Figure 9), and August 22, 1990 (Figure 5), which suggests that the local source of water to the alluvium on June 3 was groundwater from the hillslope, in contrast to the stream water source identified at the other times. This preliminary evidence

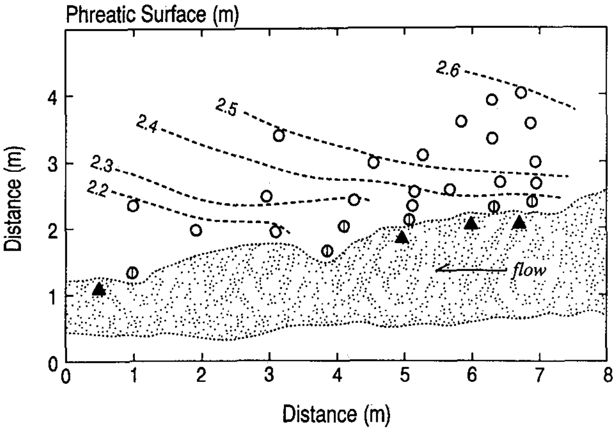

Fig. 11. Map of phreatic surface, St. Kevin Gulch, June 3, 1991. Phreatic surface contours indicated that hillslope groundwater flowed toward the stream at this location (1356–1362 m, see Figure 4), in contrast to the suggested stream water source for subsurface flow at this location on August 22, 1990 (Figure 5), and on June 18, 1991 (Figure 9).

suggests that substream flow paths may disappear temporarily at St. Kevin Gulch when the catchment is very wet.

The apparent effect of catchment wetness on water exchange with substream zones is summarized below. At times of low catchment wetness, variability in the stream water surface slope causes embedding of localized substream flow paths within the larger-scale groundwater flow system. During very wet periods the supply of groundwater from the lower hillslope overwhelms the influence of stepped-channel units on potential forces at the stream boundary. Substream zones disappear temporarily, and groundwater inflow occurs more uniformly along the channel.

7. SUMMARY

A numerical hydrological model demonstrated the potential for surface-subsurface water interactions to be controlled by streambed and stream water slope variation in catchments. Field evidence from St. Kevin Gulch supported the hypothesis that a stepped-bed morphology controls stream-subsurface water exchange. Hydrometric and tracer data documented that stream water and solute were locally recharged into well-defined "substream" flow paths in the alluvium that return to the stream at downstream locations. These stream water–filled flow paths in the subsurface began at the downstream end of pools, continued via a subsurface route around steeper units (e.g., cascades and steps), and then reentered the stream at the upstream end of the next pool. Groundwater from the lower hillslope was transmitted to the stream only in zones located between substream flow paths. Gross water exchange fluxes between stream and subsurface exceeded net inflows of groundwater from the lower hillslope by twofold. In addition, field results suggested that the effect of streambed topography on surface-subsurface exchange is lessened when the catchment is wet, due to the greater influence of hillslope groundwater heads on the head potential distribution near the stream. This study showed that streambed topography can influence solute transport in catchments, an effect which has potential consequences for biogeochemistry and aquatic ecology in catchments.

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