J. N. Am. Benthol. Soc., 1996, 15(1):3–19 $\ensuremath{\mathbb{C}}$ 1996 by The North American Benthological Society

Seasonal and storm dynamics of the hyporheic zone of a 4th-order mountain stream. I: Hydrologic processes

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Abstract. The objective of this study was to quantify fluxes of ground water and advected channel water through the shallow aquifer adjacent to a 4th-order mountain stream. A network of wells was installed from 1989 to 1992. Water-table elevations were measured seasonally and during storms. These data were used to calibrate MODFLOW, a 2-dimensional groundwater flow model. The fluxes of water through the subsurface were estimated from the head distributions predicted by the model for 8 steady state model runs bracketing the observed range in baseflow conditions, and for 1 transient simulation of a large storm. The overall pattern of subsurface flow changed little over the course of the year, even though the relative flux of advected channel water and ground water changed among seasons and during storms. Apparently the longitudinal gradient of the main valley, the location of the stream, and the influence of secondary channels determined the pattern of subsurface flows. Subsurface fluxes through a gravel bar were dominated by advected channel water but fluxes through the floodplain were dominated by ground water. Flow rates were positively correlated to estimated stream discharge during base-flow periods, but decreased slightly during storms because of precipitation inputs to the aquifer. The mean residence time of water stored within the aquifer was approximately 10 d for the gravel bar and 30 d for the floodplain during baseflow periods. Even though precipitation during the simulated storm equaled 12% and 23% of the water stored in the gravel bar and the floodplain, respectively, the mean residence time of water remained long.

Key words: hyporheic zone, advected channel water, ground water, groundwater flow models, streams, aquifer, water budget, flow path, geomorphology.

Interest in the exchange of surface water between streams and the subsurface or hyporheic zone is increasing, in part because the chemistry of stream water is influenced by biogeochemical processes that occur in this zone (Grimm and Fisher 1984, Triska et al. 1989, Duff and Triska 1990, Bencala et al. 1993). However, the subsurface hydrology of shallow aquifers adjacent to streams is complex, with water from a variety of sources flowing within a complex flow net. Ground water enters the shallow unconfined aquifer within the floodplain as drainage from adjacent hillslopes or from larger regional aquifers. Exchange flows of stream water into the aquifer occur along substream flow paths. We follow the terminology of Triska et al. (1989) in this paper. Advected channel water is stream water found within the streamside aquifer. The exchange of surface and interstitial water is exchange flow. Ground water refers to subsurface water from other sources. We do not differentiate between soil water draining from adjacent hillslopes and that in deep aquifers. The zone

beneath, and to the side of the stream, where subsurface water is a mixture of at least 10% advected channel water and ground water is the hyporheic zone.

Several factors drive the exchange flow of advected channel water into the subsurface. Inertial effects resulting from streambed roughness drive exchange flows over distances of a few cm (Savant et al. 1987, Thibodeaux and Boyle 1987). Advective transport of channel water into the subsurface over distances of several m results from changes in channel slope in stepped-bed streams (Vaux 1962, 1968, Harvey and Bencala 1993). Preferential subsurface flows in secondary channels create potential gradients between the main stream and drive exchange flows over distances of 10s to 100s of m (Gibert et al. 1977, Stanford and Ward 1988, Ward 1989). Finally, the change between constrained and unconstrained stream reaches in alluvial river valleys may lead to the development of hyporheic zones that exceed 1 km in width (Stanford and Ward 1988 1993).

Several conceptual models describe the interactions between surface water flowing in stream channels and subsurface water flowing through streamside aquifers (White et al. 1987, Triska et al. 1989, Valett et al. 1990, Hendricks and White 1991, Bencala 1993). Tracer studies and transient storage models have been used to study the effect of subsurface flows of water and of biochemical processes occurring in the hyporheic zone on stream-water chemistry (Bencala et al. 1984, Triska et al. 1989, Castro and Hornberger 1991, Kim et al. 1992). These studies have usually focused on small streams during periods of constant discharge, because tracer experiments become logistically difficult at high discharge or during periods of changing stream discharge (Bencala et al. 1993). Consequently, little is known about the hyporheic zones of larger streams and rivers. Moreover, few studies have attempted to determine how changes in stream and groundwater discharge between dry and wet seasons, or between baseflow and stormflow periods, affect the flows of advected channel water through the hyporheic zone. Further, the role of groundwater fluxes through the streamside aquifer in determining the location and extent of the hyporheic zone has not been widely studied (Rouch 1992, White 1993).

The focus of our study was to quantify the fluxes of both ground water and advected channel water through the shallow aquifer adjacent to a 4th-order mountain stream under the range of stream discharges observed for each season of the year, as well as the changes that occur during storms. We investigated subsurface flow paths using the groundwater flow model MOD-FLOW (McDonald and Harbaugh 1988) and used the calibrated model to estimate subsurface fluxes.

Study Site

The study site was on McRae Creek, a 4thorder stream within the Lookout Creek catchment and the H.J. Andrews Experimental Forest in the western Cascade Mountains of Oregon, USA (44°10'N, 122°15'W). The drainage area above the study is 1400 ha, and most of the catchment is forested. Elevation within the catchment ranges from 600 m at the study site to 1600 m along the drainage divide. Average annual precipitation is approximately 2500 mm, falling mainly between November and March (Bierlmaier and McKee 1989).

McRae Creek was not gauged; therefore, to estimate McRae Creek discharge we used records from Mack Creek which is 4.5 km away. We assumed that unit area discharges would be similar for the two catchments (Gordon et al. 1992) and multiplied Mack Creek discharge by the ratio in size between the two catchments (1.6). Estimated stream discharge was highly variable over the study period, ranging from a low of 100 L/s during September and October, to 600 L/s during baseflow periods throughout the winter, and with peak storm flows during fall and winter exceeding 5000 L/s. These estimates may be inaccurate because stream discharge often does not increase linearly with watershed area, especially during storms (Dunne and Leopold 1978). However, stream stage at McRae Creek was highly correlated with stream discharge at Mack Creek ($n = 93, r^2 = 0.92$) over the ranges observed, suggesting that channel routing of water through these two watersheds was similar. Thus, the estimation error should be small, even during storms.

The study site was 100 m long and 80 m wide and lay along the eastern bank of an unconstrained stream reach (Fig. 1). A complex of landforms is present within the study site, including a recently formed gravel bar, older floodplain surfaces, and terraces. Sediment of both the gravel bar and the stream channel is a poorly sorted mix of sand, gravel, cobbles, and boulders more than 1.5 m in depth. A layer of rounded, stream-worked cobbles and boulders, 10-50 cm in diameter, is present at 1 to 3 m depth within the floodplain. The sediment overlying this layer varies in texture from loam to fine sand. A small seep is present along the boundary between the terrace and floodplain, but is not gauged. There is no surface flow from this seep during late summer. Flows increase during the winter rainy season, and peak during storms.

Methods

Wells and well network

A single transect of wells was established during late summer in 1989 as a pilot study. Additional transects of wells were installed during the summer of 1990 and an additional 18



FIG. 1. McRae Creek study site showing landforms and well locations.

wells were established on, and adjacent to, the gravel bar during 1991 and 1992. All wells were driven by hand because the study site had no road access. Large cobbles and boulders throughout the study site hindered well placement so that the deepest wells penetrated only 2.5 m below the ground surface. Wherever possible, wells were placed in holes driven at least 50 cm below the surface of the water table at summer baseflow.

Well casings were made from PVC pipe that was "screened" by drilling 0.32-cm diameter holes into the bottom 50 cm of each PVC pipe, at an approximate density of 1 hole/cm. The locations of all wells were mapped (Fig. 1) and the elevations of the well head and the ground level at each well were surveyed.

Water-table elevations were measured from the well network and from stage plates in both McRae Creek and in pools of water in the secondary channel at the back of the gravel bar (Fig. 1). The frequency and timing of storm observations were based on the intensity of precipitation and changes in stream stage. Measuring water depths in all wells took 1–2 h. Observations (4–5) were spaced irregularly over a single day during intense storm periods to capture the rising leg, crest, and falling leg of the stream hydrograph and the associated rise and fall of the water table.

Slug tests

Saturated hydraulic conductivities (K) were calculated from falling-head slug tests following Bouwer and Rice (1976), Bouwer (1989), and Dawson and Istok (1991). This method is appropriate for partially penetrating wells in unconfined, heterogenous, anisotropic aquifers. Most of the well casings were 2.54 cm in diameter, and while the test is valid for small-diameter wells, the estimated value of K applies only to a small region around the well (Bouwer 1989). The mean hydraulic conductivity was 2.0 \times 10⁻² cm/s in the secondary channels (range 2.1×10^{-3} to 2.5×10^{-2}), 9.0×10^{-3} cm/s for the gravel bar (range 1.0×10^{-3} to 2.2×10^{-2}), and 4.7×10^{-3} cm/s for the floodplain (range: 6.4×10^{-4} to 1.2×10^{-2}).

Tracer tests

A tracer test using a continuous injection of Rhodamine WT dye was conducted from 10 to 15 July 1992 to confirm flow paths of water predicted from the numerical simulation. An injection well was selected at the head of the gravel bar and sample wells were located along the expected flow path of tracer. Samples were collected at intervals throughout the test to measure arrival times and concentrations of Rhodamine WT within the tracer plume. Concentrations of dye never reached steady state during the 5-d period, even in the well closest to the injection point, probably because dye was adsorbed to sediments (Bencala et al. 1983). Consequently data could not be used to establish flow velocities and estimate the saturated hy-

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draulic conductivity. However, these data do show the flow path of tracer through the aquifer. Concentrations of Rhodamine WT measured on 15 July, 122 h after the start of the dye injection, were contoured by hand.

Temperature was also used as a tracer to monitor both the distance to which stream water penetrated into the aquifer and the mixing of stream water and ground water within the aquifer. Temperature is not a conservative tracer because of heat stored in the sediment and diel temperature changes. However, high air temperatures combined with multi-day residence times of water in the subsurface allow temperature to be used as a "label" for stream water after long hot and dry periods. Therefore, measurements were made during late summer when the shape of the piezometric surface and the flow net were relatively constant, and the difference between stream temperature and ground water exceeded 5°C. A simple mixing model was used to calculate the percentage of stream and ground water in each well. The inverse-distance-squared method (SURFER, Golden Software Inc., Golden, Colorado) was used to interpolate between wells, and the interpolated surface was contoured.

Groundwater modeling

Model assumptions.—We calibrated the numerical flow model MODFLOW (McDonald and Harbaugh 1988, Prudic 1989, McDonald et al. 1991) to reproduce the head observed within the well network. The calibrated model was then used to estimate fluxes of both advected channel water and ground water through the subsurface. We used a 2-dimensional model because wells were shallow and calibration data were available only to describe the upper 1-2 m of the aquifer. Observations of sediment layers in stream banks and soil pits at the study site showed that individual sediment lavers were generally <2 m in thickness and did not exceed 3 m. Hence, the model was parameterized so that the saturated thickness of the modeled aquifer was approximately 3 m during summer low flow. Although deeper alluvial layers are most likely present at the study site, we assumed that there was no leakage through the bottom of the modeled aquifer because alternating layers of fine and coarse sediment characteristic of alluvial deposits would tend to re-

strict vertical flow to or from deeper layers. We assumed that subsurface flows did not cross beneath the stream because subsurface flow from both sides of the valley converge (or diverge) along the channel. Hence, the model domain was defined by a no-flow hydraulic boundary following the center of the stream channel (Anderson and Woessner 1992). We assumed that evapotranspirational losses from the aquifer could be ignored because the water table was more than 1 m below the ground surface during the summer, and because Douglas-fir and western hemlock trees are shallow rooted (Waring and Schlesinger 1985). Fluxes of water into the study site from the seep along the terrace boundary are not known because the seep was not gauged; thus these fluxes were not simulated.

Model calibration .- MODFLOW was initially calibrated with head data recorded from the well network during a low flow period on 28 September 1992. For a complete description see Wondzell (1994). The model was further calibrated with data from 7 additional sample dates—3 summer and 4 winter—that bracketed the range of stream and water-table elevations observed over the study period. Observations for winter baseflow were made during interstorm periods several days in length. Winter baseflow was modeled with a steady-state simulation even though stream and water-table elevations were changing slowly. The slow rate of change in the observed water-table elevations suggested that the flow of ground water through the floodplain was in near equilibrium with lateral inputs to the floodplain. However, model calibration runs showed a progressively worse fit under increasingly wet conditions, especially along the terrace boundary. These results suggested that the boundary conditions specified for the terrace-floodplain boundary did not account for the magnitude of groundwater flux from either the small seep or the adjacent hillslope during wet, mid-winter conditions. Consequently, the ground water input to the floodplain from the adjacent hillslopes was increased to fit the data collected during the wettest winter baseflow period.

Model confirmation.—A transient simulation of a storm was used to test the calibrated model. Specific storage relates changes in water table elevations to the change in the volume of water stored within the aquifer sediment. We used es-

timates of specific storage for sediment of similar texture (Dawson and Istok 1991): 0.32 for the gravelly sediment of the gravel bar, and 0.20 for the loam textured sediment of the floodplain. Hourly totals of precipitation recorded at rain gauges 7 km from the study site were used as model inputs to simulate this storm. Assuming equivalent rainfall at the study site, precipitation inputs of water were 564 m3 to the floodplain and 113 m³ to the gravel bar. These inputs are large relative to the 250 m³ of subsurface flux expected during winter baseflow conditions over the same time period. If the model underestimated subsurface fluxes, predicted heads would be higher than observed heads, because the water table would rise as precipitation inputs were stored within the aquifer instead of draining to the stream. Also, water-table elevations would return to the steady state, or prestorm condition, more slowly than conditions actually observed, because the rate at which stored water drained from the aquifer would be slowed. Conversely, if the model overestimated groundwater fluxes, precipitation inputs would drain rapidly from the aquifer, predicted heads would be lower than observed heads, and water-table elevations would return to pre-storm conditions faster than field observations. Thus predictions from the transient simulation of this storm could be compared with recorded changes in water table elevations to test the calibrated model.

Results

Groundwater model fit

The steady-state model predictions fit the observed data during the summer. The mean of the residuals (MR) ranged within ±2.0 cm of the observed values, and the mean of the absolute value of the residuals (MAR) ranged from 12 to 14 cm (Fig. 2). However, the model did not simulate wet conditions during winter baseflow, especially along the terrace-floodplain boundary where heads were underestimated by 30 to 70 cm. After the groundwater inputs from the adjacent hillslope were increased, the predicted heads fit the observed data well for sample dates during the winter baseflow period. The MR ranged between ± 4.0 cm and the MAR ranged between 11 to 12 cm for these three model runs. However, the predicted fluxes of

ground water from the terrace showed an abrupt, or step-like increase between summer and winter that is an artifact of increasing hillslope inputs to fit the winter data better (Fig. 2). These results suggested that a more complex model should be used that includes either seep or a function that gradually increases the groundwater flows from the terrace in concert with increases in stream discharge. However, judging from the magnitude and distribution of the residuals, we believe that the magnitude of error introduced into the model predictions was small, relative to the sensitivity of the predictions to variations in either K or the saturated depth. Consequently, no further adjustments were made.

The transient model accurately predicted the overall gradient of the water table across the study site. The average of the MRs calculated for each observation data set was <1 cm and the average of all the MARs was 16 cm. The model also accurately predicted changes in water-table elevations for wells such as PA07 located on the gravel bar and PV31 located on the upper half of the floodplain (Fig. 3). However, changes in water-table elevations were poorly predicted in wells such as PA51 along the terrace–floodplain boundary or PE30 on the lower part of the floodplain (Fig. 3).

Flow net

Only a short length of the stream channel in the upper part of the study site was a zone of groundwater discharge. Most of the stream channel was dominated by the recharge of advected channel water into the floodplain aquifer during baseflow periods (Fig. 4A). The observed pattern in water temperatures clearly shows the penetration of stream water into the aquifer (Fig. 5A) in the location expected from the height of the piezometric surface. Stream water penetrated as far as 20 m into the floodplain, and both the location and extent appear related to the location of the secondary channel into which both advected channel water and ground water upwell. The movement of Rhodamine WT dye through the aquifer (Fig. 5B) also followed the path expected from the shape of the piezometric surface during summer low flow. Flows of advected channel water and ground water converged on the secondary channel in the lower part of the floodplain (Fig. 4A). The conver-

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FIG. 2. Relationship between subsurface flux from each source and estimated stream discharge, for the gravel bar and the floodplain, predicted from 8 model runs bracketing the range in baseflow conditions. The mean of the absolute value of the residuals (MAR) and the mean residuals (MR) for each model run are shown.

gent pattern of flow along the secondary channel appeared to limit the lateral diffusion of the tracer plume in the lower half of the gravel bar.

The overall shape of the piezometric surface changed little between summer baseflow, winter baseflow, and peak storm flow, although the height of the piezometric surface increased by more than 0.5 m over the same period (Fig. 4). Consequently, there was little change in the flow net among seasons or during storms. The most notable changes were the extension of the zone of groundwater discharge in the upper portion of the stream reach and the steepening of gradients in the piezometric surface along the floodplain-terrace boundary.

The entire stream reach along the floodplain was a site for groundwater discharge during storms. Apparently, increased inputs of ground water to the floodplain raised the water table above streamwater elevations. Head gradients



FIG. 3. Comparison of observed (symbols) versus predicted (lines) head in 5 wells over the course of the 6-d transient simulation.

along the floodplain-terrace boundary were gradual during summer baseflow, indicating that flux of ground water across this boundary was small during the summer when there was little rain and adjacent portions of the catchment were dry. Head gradients were much steeper along the floodplain-terrace boundary during winter baseflow and in storms than in summer, showing that groundwater flux across this boundary increases during wet seasons. However, these lateral inputs were insufficient to develop a strong, cross-valley flow pattern, even in storms. The portions of the floodplain adjacent to the terrace had the lowest saturated hydraulic conductivity of any locations within the study site. Consequently, the lateral vector of flow increased only slightly at other locations within the floodplain during storms.

Head gradients showed that subsurface fluxes converged on the secondary channel in the lower part of the floodplain, suggesting that this

channel was a primary location for the discharge of ground water from the floodplain and advected channel water from the gravel bar (Fig. 4). Surface flow in the secondary channel was continuous, except in summer. Water was present only in small pools along the secondary channel at the end of the dry season in late summer, and discharge to the McRae Creek was entirely subsurface. Flow in this channel increased during the winter rainy season and peaked in storms, matching the predicted changes in subsurface fluxes. The drainage of subsurface water through the secondary channel created a steep head gradient between the secondary channel and stream. This effect was persistent, even in storms, and advected channel water flowed through the gravel bar under all observed conditions (Fig. 4C). The floodplain disappears at the lower end of the study site, where the stream channel abuts against the terrace. Here, ground water, advected channel water, and sur-

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FIG. 5. Mixing of stream water and true ground water during mid-summer (A) and concentration of Rhodamine WT dye on 15 July 1992, 122 h following the start of continuous injection of dye (B). Outlines of landforms and secondary channels follow Fig. 1, and equipotential lines are from Fig. 4A.

face water flowing in the secondary channel are all discharged to the stream.

The interface between the advected channel water and the ground water may be sharp, and stationary through time where a secondary channel creates zones of convergent flow. This situation existed along the lower half of the secondary channel where subsurface water was discharged and the zone of mixing was narrow

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(Fig. 5A). The width of the mixing zone in other locations depended on the magnitude of advected channel water and groundwater fluxes. For example, near the head of the gravel bar, advected channel water mixed with ground water in a zone >20 m wide during summer low flow (Fig. 5A). The width of this mixing zone was narrower in other seasons of the year and during storms when groundwater fluxes

FIG. 4. Piezometric surface predicted from model simulations for A) summer baseflow, B) winter baseflow, and C) peak storm flow. Flow lines were hand drawn and are only approximate. Only wells for which data were available are shown.

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FIG. 6. Summary of predicted fluxes of advected channel water and ground water and estimated stream discharge between seasons and during a storm. Fluxes out of the model domain are a sum of fluxes into the model domain. All fluxes are in L/s, including precipitation (Ppt.). Note that flux due to precipitation is the difference between precipitation inputs and the amount of water stored in the aquifer during each time step. Storm fluxes are averaged over the 6-d transient simulation (Fig. 7) and do not represent an "average storm".

through the floodplain were greater, limiting the recharge of the aquifer with stream water (Figs. 4B, 4C).

Subsurface fluxes

Subsurface fluxes along this 110-m stream reach were always small in comparison to estimated stream discharge (Fig. 6). The absolute magnitude of subsurface fluxes during baseflow periods was greater in the winter rainy season than in the summer dry season, but the relative magnitude of these fluxes decreased as stream discharge increased. The total subsurface flux was 0.79% of stream discharge at summer low flow, 0.02% during winter baseflow, and 0.007% in storms. Interestingly, subsurface fluxes into the model domain decreased slightly in storms when precipitation inputs were large (Fig. 6).

Gravel Bar.—The predicted flows of subsurface water through the gravel bar and the floodplain varied linearly with estimated stream discharge over the observed range of baseflow (Fig. 2). Predicted flow rates of advected channel water through the gravel bar were 0.60 L/s during the dry season when estimated stream discharge

was only 100 L/s, and increased to 0.90 L/s under the wettest baseflow condition in winter, when estimated stream discharge exceeded 600 L/s. The flux of advected channel water did not continue to increase with stream discharge during storms. Predicted flows of advected channel water through the gravel bar were 0.76 L/s before the storm, and only 0.83 L/s during the period of peak stream discharge 100 h later, when estimated stream discharge exceeded 2400 L/s. The patterns in the predicted fluxes of advected channel water did not increase with stream discharge during the periods of greatest precipitation as would be expected from the baseflow simulations (Fig. 7). Two factors seemed to account for this difference. First, the potential gradient from the stream to the secondary channel did not increase much during storms (Fig. 4C). Water levels in the secondary channel changed less than 5 cm between summer low flow and peak storm discharge. During base-flow periods, stage increased linearly with stream discharge (observed stage -0.008 to +0.200 m). However, at stages higher than 0.200 m, the relationship between stage and discharge was exponential. Therefore, the stage during

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FIG. 7. Bottom—average hourly precipitation rate for each 6-h stress period of the 6-d transient simulation. Top—predicted flux of advected channel water through the gravel bar; predicted discharge of ground water from the floodplain into the gravel bar; and predicted subsurface flux from precipitation (change in storage minus precipitation inputs) for times at which head data were observed from the well network. Middle—predicted flux of advected channel water through the floodplain, predicted flux of ground water into the floodplain from the boundary along the head of the floodplain and along the terrace, and predicted subsurface flux from precipitation (change in storage minus precipitation inputs) for times at which head data were observed from the well network.

peak storm discharge was similar to winter baseflow. Second, precipitation appeared to replace some of the expected flows of advected channel water, given the relationship between stream stage and subsurface flow under baseflow conditions. When precipitation rates during the storm exceeded subsurface fluxes, precipitation inputs were stored within the aquifer, raising the water table and leading to a decrease in the flux of advected channel water through the gravel bar.

Precipitation inputs stored within the aquifer drained rapidly from the gravel bar so that water-table elevations quickly returned to steadystate levels after the storm. Consequently, flow through the gravel bar had nearly returned to pre-storm rates within 24 h following the last precipitation from this storm (Fig. 7). At this time the water table was only slightly elevated above steady-state levels and the drainage of stored water accounted for <5% of the total flow through the gravel bar.

The gravel bar had a total area of 825 m². Assuming an average aquifer thickness of 3 m and a specific yield of 0.32 for a gravelly sand (Dawson and Istok 1991), the gravel bar would store 740 m³ of water. Consequently, the estimated mean residence time of water in the gravel bar varied from 10 to 12 d during base-flow periods, given the range in the predicted flux of advected channel water among seasons. Precipitation inputs to the gravel bar during the 6-d storm equaled 12% of the volume of water stored in the gravel bar. The estimated mean residence time of water in the gravel bar was 9 d during this storm because subsurface flow rates increased as a result of precipitation.

Floodplain.—The flux of ground water through the floodplain was much less than the flow of advected channel water through the gravel bar under baseflow conditions for every season of the year (Fig. 2). Predicted fluxes through the aquifer underlying the floodplain ranged from 0.30 L/s during the dry season to 0.45 L/s during the winter baseflow. The predicted inputs of ground water along the terrace boundary were positively correlated to estimated stream discharge and ranged from 0.05 L/s to 0.18 L/s between summer and winter baseflow, respectively (Fig. 6). The predicted inputs of ground water along the floodplain boundary were relatively constant over the observed range in base-flow stream discharge, with a range of 0.17-0.20 L/s. Fluxes of advected channel water into the floodplain were constant at 0.07 L/s during baseflow periods.

Subsurface fluxes through the floodplain were dominated by precipitation inputs throughout the transient simulation. Precipitation exceeded subsurface flow rates, consequently the water-table elevations increased as water was stored within the aquifer. By the end of the storm, drainage of stored water accounted for 0.55 L/s of subsurface flow and represented more than 50% of the total subsurface flow through the floodplain (Fig. 7). Predicted flows of ground water from both the floodplain and the terrace were small relative to precipitation during the simulation. The predicted flow from the terrace reached 0.22 L/s at the peak of the storm (Fig. 7), which was only 15% greater than winter baseflow inputs. Surprisingly, the model predicted that groundwater inputs along the head of the floodplain would decrease by 50% at the peak of the storm (Fig. 7); these decreased inputs resulted from a drop in the head gradient from 0.078 to 0.052 in the region adjacent to this boundary (Fig. 4C), which is consistent with the observed head data which also show a decrease in the head gradient from prestorm to peak flow.

The hydraulic conductivities in the finer sediments of the floodplain were lower than those of the gravel bar. Consequently, ground water drained slowly from the floodplain after the end of the storm, and would take at least 6 d to return to steady state. Our estimate was conservative. We made a linear extrapolation from the initial drainage rates of ground water from the floodplain (the final 24 h of the transient simulation, or the period between 54 and 78 h when little precipitation fell). The actual response should be non-linear because drainage rates should slow as the system nears steady state.

The floodplain had an area of 4000 m². Assuming an average aquifer thickness of 3 m and a specific yield of 0.20 for silt or sandy clays (Dawson and Istok 1991), the floodplain could store 2500 m³ of water. The estimated mean residence time of water in the floodplain varied from 57 to 87 d, given the range in the predicted fluxes of ground water among seasons. Precipitation inputs to the floodplain during the 6-d storm equaled 23% of the volume of water stored in the floodplain. The estimated mean residence time of water in the floodplain was 31 d during storms because subsurface fluxes increased.

Discussion

Model limitations

The heads predicted by the calibrated model agreed with the observed data for conditions ranging from a steady state following months without significant precipitation to a several day storm. However, the subsurface fluxes predicted by the model may not be realistic because flux estimates are proportional to hydraulic conductivity (K), which is difficult to measure (Freeze and Cherry 1979). The values used for K fall within the range of published values for the respective sediment types, but these values range over several orders of magnitude (Freeze and Cherry 1979, Dawson and Istok 1991). Flux predictions can be confirmed if an independent estimate for a predicted flux exists, in this case, the measured precipitation falling during the simulated storm. The precipitation flux was large relative to the rates of groundwater flow predicted by the model for base-flow conditions. The close match between predicted and observed head in wells on the gravel bar during the transient simulation, and the rate at which water-table elevations returned to steady-state levels after the storm, suggested that the predicted fluxes through the gravel bar were real-

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istic. However, the poor fit for wells on the floodplain suggested that either the simulated groundwater inputs along the floodplain/terrace boundary or the predicted groundwater fluxes through this part of the model domain were poor.

Model predictions were also limited by the lack of detailed geohydrologic data for the study site. We assumed that horizontal flux should predominate because sediment lavers within the aquifer are horizontally continuous, and because the alternating layers of fine and coarse sediment characteristic of alluvial deposits would restrict flow between shallow and deep sediment layers. However, no data are available on the stratigraphy of the floodplain or the depths to impermeable layers. The assumed aquifer thickness (3 m) seems reasonable given that sediment layers exposed in stream banks, soil pits and auger holes within the Lookout Creek catchment seldom exceeded this thickness. Still, the predicted groundwater fluxes are sensitive to the aquifer thickness, and the long history of fluvial disturbance and sedimentation and width of the valley floor in this stream reach suggest that the total thickness of sediment deposits greatly exceed 3 m. Much additional work, including the drilling of deeper wells, would be necessary to determine the ways in which water in the upper 3 m of the aquifer interacts with water in deeper sediment layers.

Because it is impossible to validate predictions of numerical models (Konikow and Bredehoeft 1992, Oreskes et al. 1994), the results of this study should be interpreted cautiously. The predicted head distribution can be checked against field data from each well; consequently the flow net plotted from these distributions should be accurate at the scale of the model. The relative change in the predicted groundwater flux with changing stream discharge among seasons or during storms should be reliable. However, the absolute magnitudes of groundwater fluxes are uncertain because of uncertainties in K and the difficulties in specifying conditions for the floodplain-terrace boundary.

Driving factors

Fluxes of advected channel water appeared to result primarily from the effects of preferential drainage through secondary channels within

the floodplain. We did observe downwelling zones at the heads of riffles and upwelling zones at the heads of pools, which show that exchange flows of the type reported by Vaux (1962, 1968) or Harvey and Bencala (1993) did exist; however, there was no evidence that the pool-riffle-pool morphology of the stream produced lateral exchange flows. Either the well network was too coarse for these flows to be observed, or the steep potential gradient between the stream and secondary channel prevented the development of arcuate flow paths. There was no indication that the change from a constrained stream reach (upstream) to an unconstrained reach (our study site) led to the development of an extensive hyporheic zone as was reported by Stanford and Ward (1988, 1993). Rather, inputs of ground water from adjacent hillslopes, or from tributaries debouching onto the floodplain, were the source of water in the floodplain aquifer.

Dynamics of the flow net

The hyporheic zone is commonly considered to be a dynamic boundary between surface-water and true groundwater systems. Relative fluxes of both ground water and advected channel water through the hyporheic zone are expected to vary over time depending on catchment wetness and stream discharge, resulting in changes in both the location and extent of the hyporheic zone (Hynes 1983, Meyer et al. 1988, Gibert et al. 1990, Hakenkamp et al. 1993, Palmer 1993, White 1993, Williams 1993). However, we found that the overall pattern of subsurface flow changed little over the course of the year. Apparently the longitudinal gradient of the main valley floor and the influence of the secondary channel are critical determinants of the pattern of subsurface flow. Similarly, Harvey and Bencala (1993) found little change in the flow net at St. Kevin Gulch over a 10-fold change in stream discharge, although during extremely wet periods, the flux of ground water from adjacent slopes did limit the lateral exchange of advected channel water.

Fluxes of advected channel water and ground water

Even though the magnitudes of subsurface fluxes from each source were dynamic, they were relatively constant in comparison to estimated stream discharge which ranged from 117 L/s to 2425 L/s. We standardized our estimates of subsurface flux to account for the magnitude of stream discharge by expressing them as a spiralling length (Newbold et al. 1981, 1982, El-wood et al. 1983), defined as:

$$S = 2 \times Q_{\text{stream}} \times L_{\text{reach}} / Q_{\text{ACW}}$$

where S is the spiralling length (m); Q_{stream} is the stream discharge (L/s); L_{reach} is the length of the stream in the study reach (m); and Q_{ACW} is the flux of advected channel water into the hyporheic zone over the entire study reach (L/s). Our estimate is multiplied by 2 to account for fluxes of advected channel water occurring through both sides of the channel. We calculate that the spiralling length of stream water was 9.3 km of channel at the end of the summer. The spiralling length increased rapidly with stream discharge, reaching 33 km during winter baseflow and 150 km during peak storm flows. An analogous length can be calculated for the length of channel required for groundwater inputs to equal stream discharge, which equaled 28 km during late summer, 86 km during winter baseflow, and 133 km during peak storm flows. The relative magnitude of exchange flows through the hyporheic zone, or of groundwater discharge, was inversely proportional to stream discharge as shown by the rapid increase in the spiralling length.

The relative importance of the hyporheic zone may also be related to stream size as suggested by White (1993). However, few studies have attempted to quantify fluxes of advected channel water or ground water, and we know of no study that has quantified subsurface fluxes over a range of stream sizes within a single stream network. Harvey and Bencala (1993) used an approach similar to ours to estimate fluxes of advected channel water and ground water in a 32-m reach of St. Kevin Gulch, a 3rd-order mountain stream with 1/10 the discharge of McRae Creek. Using their values, we calculated a spiralling length of 107 km for stream water during spring snow melt and 9.6 km during summer baseflow. Similarly, 74 km of channel would be required for groundwater inputs to equal stream discharge during spring, and 6.6 km would be required in summer.

Groundwater discharge was relatively greater than the exchange of advected channel water in St. Kevin Gulch (the smaller stream) as predict-

ed by the conceptual model of the hyporheic zone presented by White (1993). However, the flux of advected channel water, relative to stream discharge, did not increase as stream order increased. We believe that the difference in geomorphic factors driving exchange flow at the 2 sites can account for this. Lateral exchange is driven primarily by head gradients produced by the stepped-bed morphology of the stream channel at St. Kevin Gulch, but produced by preferential drainage through a secondary channel at McRae Creek. The effect of the secondary channel creates steeper lateral gradients within the stream side aquifer than does the steppedbed morphology at St. Kevin Gulch, and the effect of the secondary channel extends over a longer portion of the studied reach. Further, the valley floor is wider and alluvial deposits are deeper at McRae Creek. Thus exchange flows occur through a larger cross section of the streamside aguifer than at St. Kevin Gulch. Results from transient storage simulations of tracer studies support these conclusions. The relative magnitude of exchange flows generally decreases with increased stream discharge (Legrand-Marcq and Laudelout 1985) or with increasing stream order (D'Angelo et al. 1993). Stream morphology is a major factor determining the magnitude of exchange flows, especially in larger streams where floodplain width strongly influences flow patterns (D'Angelo et al. 1993).

In summary, our study shows that groundwater flow models may be applied to investigate the dynamics in fluxes of both advected channel water and ground water under conditions of changing stream discharge and catchment wetness. Our results suggest that geomorphic features of the valley floor-valley floor width, the gradient of the main stream, the stepped-bed morphology of the channel, the presence of secondary channels and the way in which these channels are connected to the stream channel, and the thickness and hydraulic conductivity of the alluvial sediments-control the development of the flow net, and thereby determine the location and extent of the hyporheic zone as well as the flux of advected channel water.

Acknowledgements

We thank Dominique Bachelet, Stuart Fisher, Rosemary Mackay, Philip Sollins, Gilles Pinay and two anonymous reviewers for helpful reviews of this paper; John Selker for helping us understand groundwater hydrology; Cliff Dahm, Jud Harvey, Frank Triska, and Gary Zellweger for helpful discussions; and Craig Creel, Greg Downing, and Al Levno for helping to install wells. This research was supported by the National Science Foundation through Doctoral Dissertation Improvement Award BSR-90-16186 and (partly) by grants BSR-85-14325 and BSR-90-11663 to the H.J. Andrews Long-Term Ecological Research program, and grants BSR-84-19373 and BSR-85-08356. Support was also received from the US Forest Service, Pacific Northwest Research Station; and Oregon State University, Department of Forest Science. This paper is a contribution of the H. J. Andrews LTER program.

Meteorologic and stream discharge data sets were provided by the Forest Science Data Bank, a partnership between the Department of Forest Science and Pacific Northwest Research Station, Corvallis, Oregon. Funding of these data was provided by the Long-Term Ecological Research (LTER) program and other National Science Foundation programs (NSF grants: DEB-7611978, DEB-8012162, DEB-8508356, BSR-8514325, BSR-9011663), Oregon State University, and US Forest Service Pacific Northwest Research Station.

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> Received: 30 May 1995 Accepted: 29 November 1995



Journal









American Benthological Society

Volume 15 Number 1 March 1996