Transactions of the American Fisheries Society 132:493–502, 2003

Measuring Groundwater–Stream Water Exchange: New Techniques for Installing Minipiezometers and Estimating Hydraulic Conductivity

COLDEN BAXTER*[1] AND F. RICHARD HAUER

The Flathead Lake Biological Station,
The University of Montana,
311 Bio Station Lane, Polson, Montana 59860, USA

WILLIAM W. WOESSNER

Department of Geology,
The University of Montana,
Missoula, Montana 59812, USA

Abstract.—Measurements of groundwater–stream water interactions are increasingly recognized as important to understanding the ecology of fishes and other organisms in stream and riparian ecosystems. However, standard measurement techniques are often feasible only at small spatial scales, in areas with easy access, or in systems with relatively fine substrata. We developed simple new techniques for installing minipiezometers and obtaining estimates of vertical hydraulic gradient, hydraulic conductivity, and specific discharge in gravel and cobble streambeds that allowed for large numbers of measurements to be obtained in remote locations. Our approach yielded values comparable to those obtained through more traditional methods. Consequently, these techniques may provide a labor cost-efficient way for detecting groundwater−stream water interaction patterns that are critical labor-attributes of stream and riparian systems at multiple scales.

In recent years there has been increased research on groundwater–stream water interactions and heightened awareness of the importance of hyporheic processes to the ecology of fishes and other organisms in stream and riparian ecosystems (e.g., Stanford and Ward 1993; Brunke and Gonser 1997; Boulton et al. 1998). Consequently, measurements of groundwater–stream water interactions are needed in environments and at spatial scales not easily addressed using traditional hydrogeologic techniques. Standard measurement techniques are often feasible only at small spatial scales, in areas with easy access, or in streams with relatively fine bed sediments.

In a recently published study (Baxter and Hauer 2000), we quantified patterns of groundwater–stream water exchange in gravel and cobble beds of third- and fourth-order streams of northwestern Montana and found those patterns to be related to the selection of spawning habitat by endangered bull trout *Salvelinus confluentus*. Most studies of hyporheic processes have been done within short stream reaches. However, we demonstrated that quantifying groundwater–stream water exchange in a spatially extensive manner across a hierarchy of scales was both critical and feasible. We developed a simple design for the construction and installation minipiezometers and new techniques for obtaining estimates of vertical hydraulic gradient (VHG), hydraulic conductivity (K), and specific discharge (v) in gravel and cobble streambeds that allowed large numbers of measurements to be obtained in remote locations. Our purpose here is to describe in detail these methodological advances and to compare their results with those of more standard approaches in hopes of providing tools that will assist and further stimulate research on the ecological rules of groundwater–stream water interaction.

Construction and Installation of Minipiezometers

In our previous study (Baxter and Hauer 2000), we measured groundwater–stream water exchange through the use of minipiezometers inserted into the bed of the stream (Lee and Cherry 1978; Dahm and Valett 1996; Figure 1). We developed an installation technique that permitted us to obtain measures from over 500 minipiezometers. The development of this technique was essential because it allowed upwelling and downwelling patterns to be studied over large areas and at relatively remote

* Corresponding author: baxterco@cnr.colostate.edu
[1] Present address: Department of Fishery and Wildlife Biology, Colorado State University, Fort Collins, Colorado 80523, USA.

Received December 15, 2001; accepted October 15, 2002

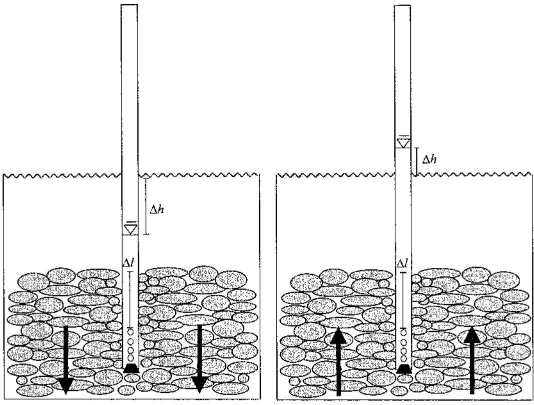

FIGURE 1.—Generalized features of minipiezometers and principles behind estimating vertical hydraulic gradient (VHG), as evident by comparing right and left panels showing how VHG is a function of Δh (the difference in head between the water level in the piezometer and the level of the stream surface) and Δl (the depth from the streambed surface to the first opening in the piezometer sidewall) under downwelling (black arrows down) and upwelling conditions (black arrows up).

stream sites that required equipment transport via backpack. We found that previously described piezometer and piezometer-installation designs were either too elaborate or heavy in their construction (e.g., Pollard 1955; Winter et al. 1988; Geist et al. 1998) or were not robust enough to penetrate the streambed of our sites (Valett et al. 1994; Dahm and Valett 1996). We went through numerous developmental stages in searching for the most effective installation mechanism. The apparatus we found to be most effective was fashioned on the concept of the dual-tube drilling system and allowed the use of many light (and relatively inexpensive), plastic minipiezometers.

The entire installation unit (Figure 2) included an outer sleeve or casing, a pointed driver rod that fit inside the casing, the minipiezometer itself, and a hammer cap that fit over the top of the driver; a 4.5-kg sledgehammer and vice grips completed the items needed. The outer sleeve or casing was 1.5 m in length and constructed from 1.9-cm (3/4-in) -diameter stainless steel pipe with a stainless steel collar ring welded 5 cm below its top. The driver rod was made of solid cold-roll steel that fit snugly inside the casing and had a machined point on one end. The outer casing was filed down where the point of the driver rod protruded so there was no lip that could get hung up during installation.

The minipiezometer consisted of 1.59-cm-diameter (5/8 in) chlorinated polyvinyl chloride (CPVC) pipe (inner diameter 1.11 cm [7/16 in]) that was perforated with approximately 30 evenly spaced holes (hole diameter, 0.238 cm [3/32 in]) over the bottom 15 cm of its length (150 cm) and plugged with a cork at the bottom. When the perforated length of the piezometer is more than eight times its radius ($L_p/R > 8$), available equations to estimate hydraulic conductivity by standard methods (as described below) become relatively straightforward (Freeze and Cherry 1979).

The procedure for installation was as follows (Figure 2): (1) the driver mechanism (the casing with the driver rod inserted) was placed on the stream bottom and a hammer cap was fitted on top of the collar, and the instrument was driven to the desired depth into the streambed by repeated blows with a 4.5-kg sledgehammer; (2) the steel driving rod was removed while the imbedded casing was held in place; (3) the CPVC minipiezometer was slipped inside the casing; and (4) while the minipiezometer was held in place (typically by pushing down on it from above with a short piece of CPVC), the steel casing was removed, leaving only the piezometer inserted in the streambed. We used a bright-colored tape around the outer steel casing to help judge when the appropriate depth had been reached. We typically installed piezometers to shallow depths (25–40 cm) in the stream substratum, though we also used piezometers nested at variable depths to check for vertical variation in hyporheic characteristics. We used vice grips when removing the driving rod from the casing and the casing from the streambed. We manually tamped the streambed sediment around the minipiezometer to ensure that river water would not directly flow along the casing to the perforated interval.

The piezometers were then developed to ensure the perforated interval was communicating with the hyporheic water. Water was extracted from the piezometer by inserting a short length of plastic tubing, applying a vacuum to the tube by mouth, then kinking off the tube, and then withdrawing it from the piezometer and emptying the tube. Repeated measures of the water level in the CPVC piezometer (see below) were used to document water level recovery and, thus, ensure good communication with subsurface water. An equilibration period preceded the recording of final field measurements. The time needed for equilibration depends on the design of the perforated interval and the hydraulic conductivity of the streambed material (Hvorslev 1951; Bouwer and Rice 1976; Freeze and Cherry 1979). We found that though most piezometers equilibrated within seconds or minutes, a few took up to several hours. Consequently we allowed 24 h for full water-level recovery.

Based on results in other stream systems (Pepin and Hauer 2002), it may be necessary to modify the driving mechanism of the instrument we described above to suit the nature of the streambed sediment being sampled. In fact, in a few stream reaches with very coarse cobble substratum, we found it necessary to use more ''heavy-duty'' piezometers to penetrate into the hyporheic zone. For this purpose, each piezometer consisted of a solid stainless steel tube (2.54-cm [1-in] inner diameter) with a machined point and a welded collar near the top where a hammer cap could be fitted. These piezometers were also driven into the streambed by repeated blows to the hammer cap with a sledgehammer. From the backpack-portable technique utilized in gravel and cobble streambeds (our study) to methods that have been developed for use in small, sandy desert streambeds of the Southwest (Dahm and Valett 1996) and in cobble streambeds of larger rivers, such as the Flathead (Stanford et al. 1994) and Columbia (Geist et al. 1998) rivers, it is apparent that basic piezometer design can be modified to obtain extensive hyporheic information in many settings.

Measuring Vertical Hydraulic Gradient

We quantified groundwater−stream water exchange by measuring vertical hydraulic gradient (VHG) and streambed hydraulic conductivity (K_h) in minipiezometers. Vertical hydraulic gradient is a unitless measure that is positive under upwelling conditions and negative under downwelling conditions. Specifically, VHG = $\Delta h/\Delta l$, where Δh is the difference in head between the water level in the piezometer and the level of the stream surface (cm) and Δl is the depth from the streambed surface to the first opening in the piezometer sidewall (the location of the middle of the perforated interval is also often used; Figure 1).

To obtain a measure of Δh, we needed to measure the water level height within the piezometer and the height of the stream surface and calculate the difference. To do this, we used a technique that involved the use of a pair of vice grips and a piece of 8-gauge wire approximately 1.6 m in length (Figure 3). The wire was abraded with a grinding stone along one edge of one-half so that it could be marked with yellow chalk (sidewalk artist chalk worked well). The wire was (1) gripped near the nonabraded end with the vice grips, (2) inserted into the piezometer until the vice grips came flush against the top of the piezometer tube, (3) withdrawn in a manner similar to an engine oil dipstick, and (4) laid alongside a meter stick to read the distance from the top of the tube (marked by the vice-grip) to the water mark. Alternative strategies include using a calibrated wooden dowel coated with chalk, a voltmeter with leads attached to the base of a calibrated wooden dowel, or a commercial water-level recorder (Dahm and Valett 1996).

To get accurate and precise estimates of the height of the stream surface relative to the water level in the piezometer, we found it was necessary to use a ''stilling well.'' The stilling well (Figure 3) was simply a hollow tube (same diameter as the piezometer) open at both ends that was attached to the side of the piezometer via a pair of plastic clips. The stilling well was always placed alongside of the piezometer in a line perpendicular to stream flow. The top of the stilling well extended above the stream's surface (but not above the level of the piezometer), and the bottom opened near the substratum but was not driven into the streambed. Once the stilling well was attached, the chalked wire and vice grips could be used to measure the distance from the top of the piezometer to the stream surface level inside the stilling well. In this manner, the distance to the stream surface and the water level in the piezometer were measured from the same location—the top of the piezometer. Attempting to determine the stream's surface elevation along the outside of the piezometer without the stilling well was not accurate or consistent because of frequent water run-up on the upstream side of the piezometer and an eddy on the downstream side.

Estimating Hydraulic Conductivity and Specific Discharge: A Comparison of Techniques

Hydraulic conductivity of streambed sediment was estimated using falling head slug tests, which relate the rate of water level change in a narrow well to the horizontal hydraulic conductivity (K_h) of the river substratum (Freeze and Cherry 1979; Fetter 1994; Butler 1998). The tests are initiated by filling the minipiezometer instantly with water. Data required are the design of the minipiezometer, the depth of the perforated interval below the streambed, and the measured change in water level over time until equilibrium is reestablished (Figure 4). Similarly, estimates of hydraulic conductivity may also be obtained using rising and constant head tests (Freeze and Cherry 1979; Fetter 1994; Butler 1998).

Standard slug-test methods for estimating hy-

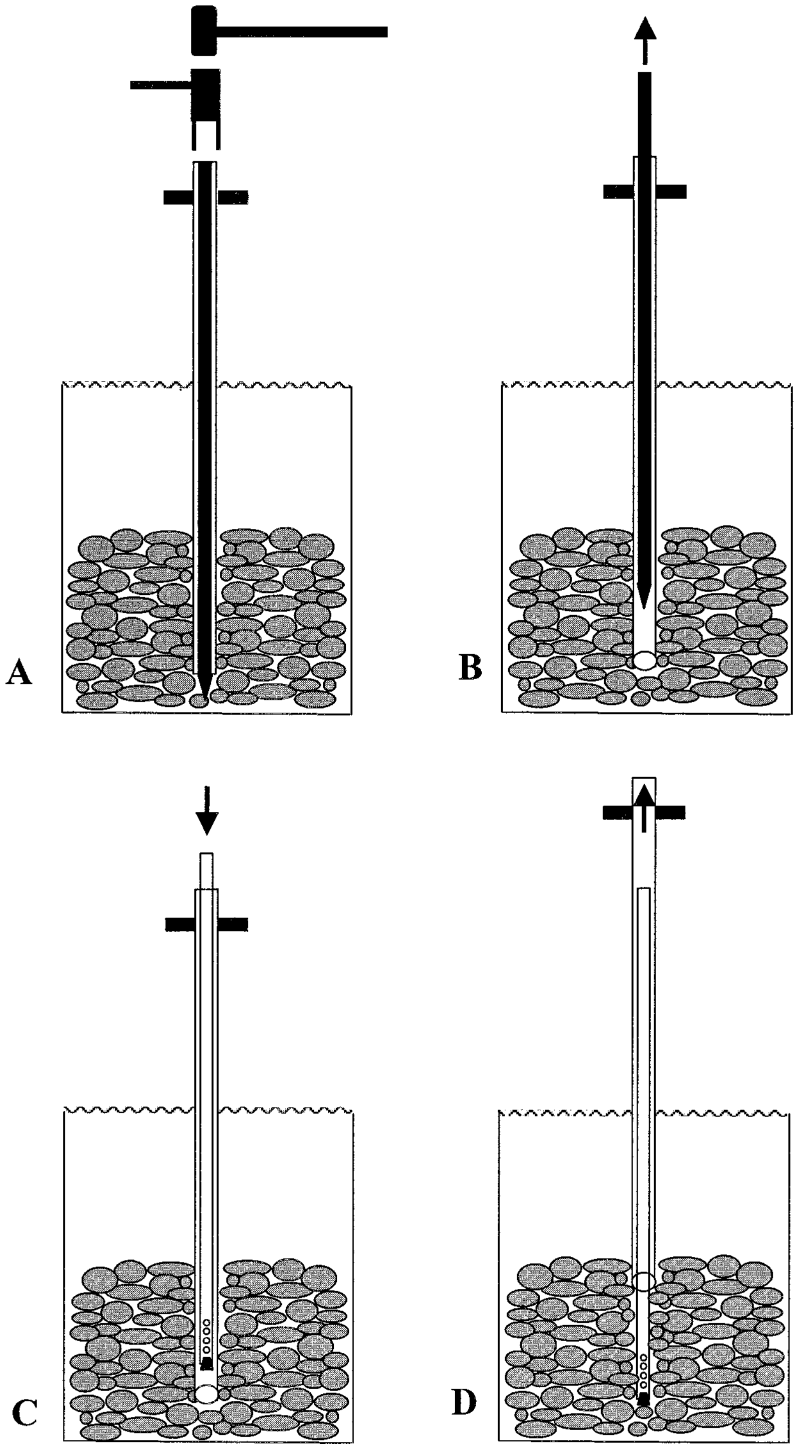

FIGURE 2.—Minipiezometer installation mechanism and sequential procedure. **(A)** The driver mechanism (the casing with the driver rod inserted) with a hammer cap fitted on top of the collar is hammered into the streambed. **(B)** The steel driving rod is removed, but the imbedded casing is held in place. **(C)** The minipiezometer is slipped inside the casing. **(D)** The minipiezometer is held in place while the steel casing is removed, leaving only the piezometer inserted in the streambed.

draulic conductivity require numerous measures of water level change during equilibration. However, in the majority of cases during the Baxter and Hauer (2000) study, minipiezometers equilibrated so rapidly (usually less than 10 s) that multiple measures could not be obtained using manual techniques. Under such circumstances, it is common to utilize pressure transducers and data loggers to monitor equilibration (Fetter 1994; Butler 1998). However, the use of such equipment may be precluded in some studies because of the small diameter of the minipiezometers, insufficient funds, study locations that are difficult to access, or the need for measurements from many locations. All of these were true for the Baxter and Hauer (2000) study. Consequently, we developed an alternative technique for estimating hydraulic conductivity based on the total time for equilibration. Below we briefly describe standard slug-test techniques, outline the derivation of our alternative equation for estimating hydraulic conductivity, and present a comparison of field estimates obtained via each of the approaches.

For some of our minipiezometers, equilibration occurred slowly enough that we were able to construct a curve of water level change with time. In these cases, we used two of the most widely applied techniques for estimating K_h: the Hvorslev (1951) and Bouwer and Rice (Bouwer and Rice 1976; Bouwer 1989) methods. The basic Hvorslev (1951) equation is

$$K_h = \frac{(r^2)\log_e(L_p/R)}{2L_pT_0},$$

where r = minipiezometer radius, L_p = length of the perforations, R = radius of the perforated interval, and T_0 is the basic time lag, a time value (t) derived from a plot of field data (Figure 4). This equation is valid for conditions such that (1) $L_p/R > 8$, (2) the perforated interval is located below the stream bottom, (3) unrestricted flow occurs between the perforated interval and the sediments (i.e., the size and number of perforations do not limit the movement of water between the sediments and piezometer), and (4) the groundwater movement in the sediment area is not influenced by the presence of an impermeable base or a limited lateral extent of the sediments. The water-level change is normalized for the maximum water-level change and plotted on a log scale versus time. The time value T_0 is associated with a corresponding normalized water-level change equal to 0.37 (see Freeze and Cherry 1979; Cedergren 1989; Fetter 1994). From this basic relationship, Hvorslev derived a number of equations that are used to estimate hydraulic conductivity under specific well designs and soil conditions (Hvorslev 1951; Cedergren 1989). The following equation was applicable to the design and conditions encountered in the Baxter and Hauer (2000) study and was used in our calculations:

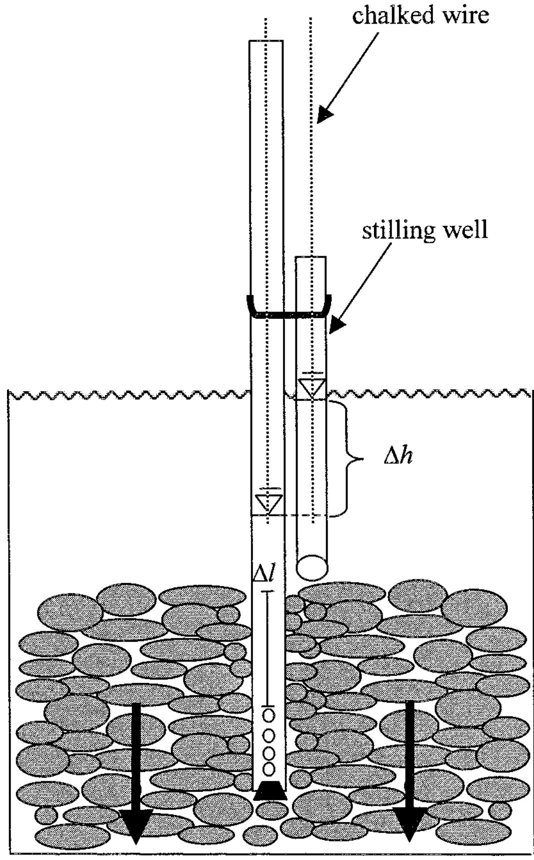

FIGURE 3.—Diagram of the stilling well and chalked wire setup used to measure the water level inside a minipiezometer and the stream water surface level at a site where downwelling occurs (see Figure 1 for definitions and additional explanation).

$$K_h = \frac{\pi(d_{\text{piezometer}})}{(11)(T_0)},$$

where $d_{\text{piezometer}}$ is the inside diameter of the piezometer.

The Bouwer and Rice (Bouwer and Rice 1976; Bouwer 1989) equation for estimating K_h is

$$K_h = \frac{(r^2)\log_e(R_e/r_w)}{2L_p}\left(t^{-1}\cdot\log_e\frac{h_0}{h}\right)$$

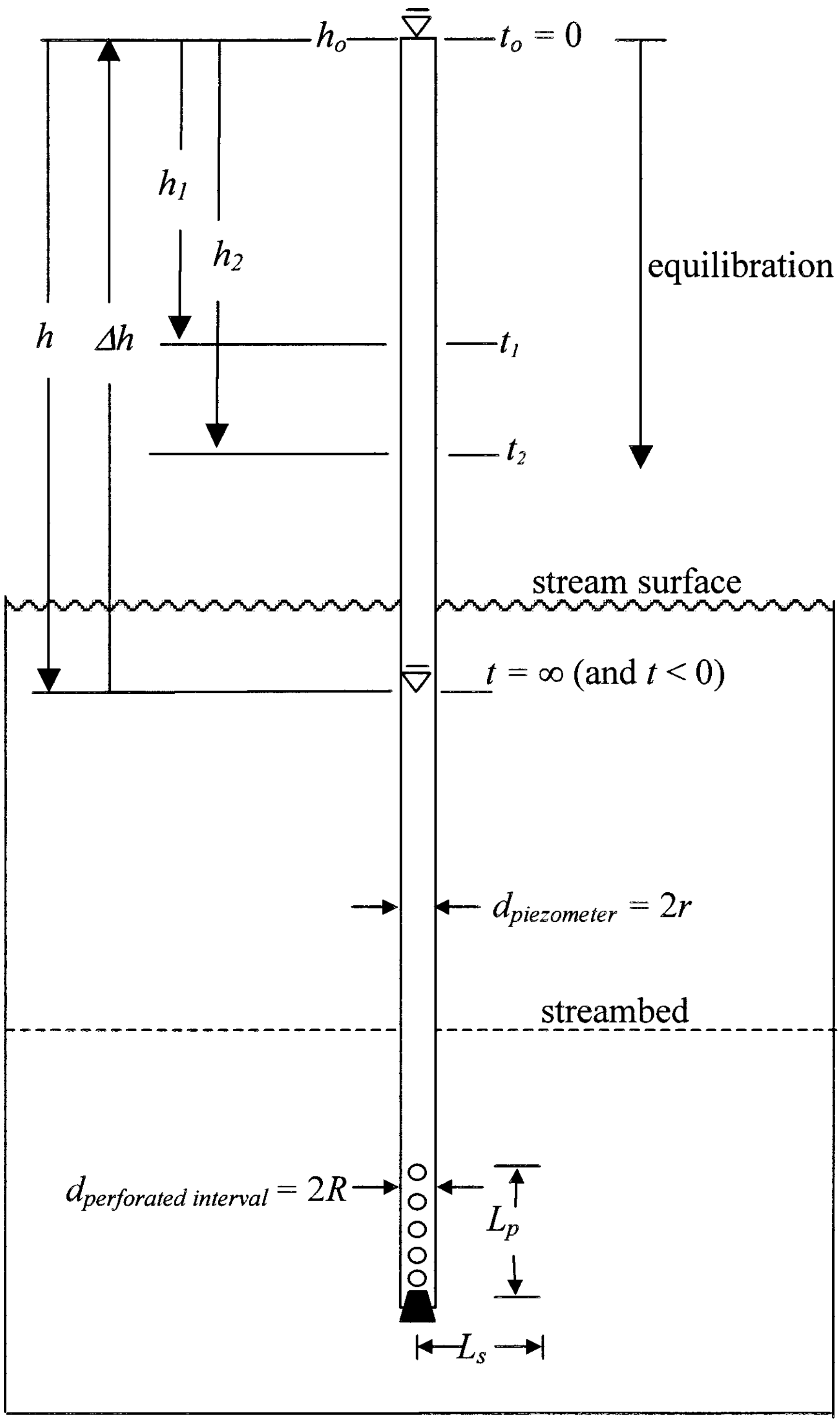

FIGURE 4.—Geometry and symbols for the falling head or slug test, where h = the head in the piezometer at time t, (note that Δh symbolizes the difference between the starting and finishing water levels—a different parameter than in Figures 1 and 3), L_s = the distance traveled by the slug of water into the sediments, L_p = the length of the perforations, R = the radius of the perforated interval, and $d_{\text{piezometer}}$ = the inside diameter of the piezometer.

where r = minipiezometer inside radius, h_0 = water level at time zero, h = water level at time t, R_e is the effective radial distance over which the water level drop h is dissipated into the surrounding sediments, and r_w is the radial distance between the undisturbed aquifer and the minipiezometer's center. In our case, with no sand or gravel pack around the well, this value is assumed to be the same as r. The value of $\log_e(R_e/r_w)$ is estimated via an additional equation based on the geometry of the piezometer system (Bouwer and Rice 1976; Fetter 1994). Similar to the Hvorslev (1951) method, a number of water-level measurements are taken during equilibration, and the water-level change is plotted on a logarithmic scale. The value of $\log_e(h_0/h)/t$ is then determined as the slope of the best-fitting line through the h versus t points and subsequently substituted into the equation for K_h.

To generate an approximate estimate of hydraulic conductivity when we could not obtain multiple measures of the equilibrating water level, we derived an alternative equation using the piezometer design data, an estimate of the time it took for the water level to reestablish equilibrium, and the relationship that the change in volume in the piezometer had to equal the change in volume exiting the perforated interval. The necessary equation was derived as follows (see Figure 4 for geometry and symbols):

$$\Delta V_{\text{piezometer}} = A_{\text{piezometer}} \cdot \Delta h$$

and

$$\Delta V_{\text{perforated interval}} = (K_h)\left(\frac{h}{L_s}\right)(A_{\text{perforated interval}})(\Delta t),$$

where V = volume, $A_{\text{piezometer}}$ = cross sectional area of the piezometer, $A_{\text{perforated interval}}$ = surface area of the perforated interval, L_s = distance traveled by the slug of water into the sediments, and h = the head in the piezometer at time t. Expanding the equations above gives the following (recall that the area of a circle = 0.7854·diameter^2 and the area of a cylinder = π·diameter·length):

$$Q_{\text{in}} = (0.7854)(d_{\text{piezometer}})^2\left(\frac{\Delta h}{\Delta t}\right) \quad \text{and}$$

$$Q_{\text{out}} = \frac{(K_h)(\pi)(d_{\text{perforated interval}})(L_p)(h)}{L_s},$$

where Q = flow, d = diameter, and L_p = the length of the perforated interval.

If $Q_{\text{in}} = Q_{\text{out}}$, then

$$[(d_{\text{piezometer}})^2(0.7854)][\Delta h] = (K_h)\left(\frac{h}{L_s}\right)(\pi)(d_{\text{perforated interval}})(L_p)(\Delta t).$$

By integrating both sides and rearranging,

$$\int [(d_{\text{piezometer}})^2(0.7854)][\Delta h] = \int (K_h)\left(\frac{h}{L_s}\right)(\pi)(d_{\text{perforated interval}})(L_p)(\Delta t),$$

which then yields the following, noting that at h_0, $t_0 = 0$ and at h, $t = t$:

$$K_h = \left[\frac{(L_s)(0.7854)(d_{\text{piezometer}})^2}{\pi\,(d_{\text{perforated interval}})(L_p)(\Delta t)}\right]\left[\log_e \frac{h_0}{h}\right]$$

If $L_s = L_p$ and $d_{\text{piezometer}} = d_{\text{perforated interval}}$, then this reduces to our equation

$$K_h = \left[\frac{(0.2501)(d_{\text{piezometer}})}{\Delta t}\right]\left[\log_e \frac{h_0}{h}\right].$$

In the Baxter and Hauer (2000) study, we performed each falling head test by adding water up to a set water level in the piezometer at start time ($t = 0$). We then monitored the elapsed time and the drop in head level by either (1) repeated measures using the chalked wire (in cases where equilibration was relatively slow), or (2) noting the elapsed time as the water level dropped to a set point in the piezometer (in situations where equilibration occurred too quickly for repeated measures). Estimated values of K_h in the Baxter and Hauer (2000) study ranged more than five orders of magnitude, from 2.32×10^{-6} to 3.37×10^{-1} cm/s.

For comparison purposes, we used water level and time measurements from 15 minipiezometers that were the only ones to equilibrate slowly (>20 s) in the Baxter and Hauer (2000) study. We computed hydraulic conductivity values via all three techniques. Values of K_h for these piezometers ranged from 2.32×10^{-6} to 4.72×10^{-2} cm/s. We found that values of K_h estimated via our alternative equation predicted estimates of K_h from the two standard approaches in a near 1:1 relationship ($r^2 = 0.99$ for both Hvorslev and Bouwer and Rice estimates; Figure 5). Of course, the standard approaches have been shown to yield the most accurate estimates of hydraulic conductivity (Fetter 1994; Butler et al. 1996; Landon et al. 2001) and should be used when such accuracy is required (e.g., studies of municipal water use, toxic waste

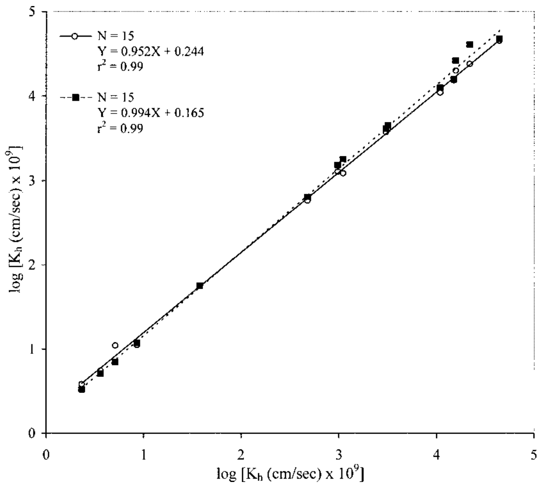

FIGURE 5.—Linear regression analysis of hydraulic conductivity (K_h) values for 15 piezometers in gravel and cobble-bed streams of northwestern Montana that were estimated via this study's alternative equation versus values obtained by the standard Hvorslev (unshaded circles) and Bouwer and Rice (black squares) techniques. Values were coded by being multiplied by 10^9 and then $\log_e$ transformed.

seepage, etc.). In addition, departures from the piezometer geometry we describe could yield different results or require changes to our alternative equation. However, our results suggest that our alternative approach to estimating K_h using the equilibration time may be applicable in coarse-grained, high-conductivity sediments and could be a promising alternative when standard methods are not feasible. Additional carefully controlled studies should be done to determine the accuracy of this approach in estimating streambed transmission characteristics and to evaluate the range of conditions over which it may be applicable.

Finally, after estimating K_h we wanted to estimate the actual flow of water through the streambed sediments. Because the hydraulic conductivity values estimated from our alternative equation yielded comparable values to the standard equations, we considered these values to be representative of horizontal properties of the bed sediments. Vertical hydraulic conductivity (K_v) values, needed to compute vertical fluxes, were assumed to be 0.10 of horizontal values (Anderson and Woessner 1992). After estimating K_v, one can estimate the vertical component of water flux through the streambed in the vicinity of each piezometer, the specific discharge (v; $cm^3 \cdot cm^{-2} \cdot s^{-1}$) being $v = K_v(\Delta h/\Delta l)$, where $\Delta h/\Delta l$ is the vertical hydraulic gradient (VHG; Figure 1) derived from the minipiezometers driven in the streambed (Freeze and Cherry 1979).

Applications in Studies of Stream and Fish Ecology

The minipiezometer design and installation method we described should provide a cost-effective, labor-saving means for quantifying hyporheic processes in remote settings. Such environments, frequently accessible only by backpack, are often the focus of stream and fish ecology research. In addition, the techniques we described may assist researchers working in systems with gravel and cobble streambed sediments because coarse streambed materials present challenges for installing minipiezometers and estimating hydraulic conductivity.

In addition to Baxter and Hauer (2000), numerous studies have demonstrated the importance of groundwater influence as a critical habitat attribute for stream fish (e.g., Benson 1953; Cunjak and Power 1986; Nielsen et al. 1994; Curry and Noakes 1995). Hyporheic exchange is known to occur across a hierarchy of spatial scales, including valley segment, reach, channel unit, and subunit scales (Stanford and Ward 1993; Brunke and Gonser 1997; Boulton et al. 1998; Woessner 2000). However, most studies of groundwater influences on stream fish ecology have been carried out at relatively small spatial scales, often focusing on channel-unit or subunit patterns and their effects on the distribution or success of spawning among salmonid fishes (e.g., Sowden and Power 1985; Curry and Noakes 1995; Garrett et al. 1998). The Baxter and Hauer (2000) study demonstrated how large-scale geomorphic and hyporheic patterns set the context for interpreting results of work at smaller scales. More studies of streams and stream fishes are needed that address groundwater–stream water interactions across a hierarchy of scales. The techniques described here should make this research more feasible.

Although these methods provide new tools for researchers, the ability to detect and quantify patterns in groundwater–stream water exchange at nested spatial scales may be enhanced through the use of techniques complementary to measurements from minipiezometers. In particular, accretion studies of stream flow (e.g., Riggs 1985; Kondolf et al. 1987; Stanford et al. 1994), thermal mapping (Silliman and Booth 1993; Ebersole et al. 2001; Torgersen et al. 2001), and winter ice observations (Benson 1953; Baxter and Hauer 2000) can complement piezometer use and yield a more complete perspective on valley segment to reach scale patterns of groundwater–stream water exchange. Quantifying groundwater–stream water interactions at smaller spatial scales (usually within a reach) may involve (1) the use of minipiezometers at a high sampling resolution (e.g., Valett et al. 1994; Baxter and Hauer 2000), (2) fine-scale measurements of streambed temperature (White et al. 1987), (3) use of seepage meters (Lee and Cherry 1978), (4) digging sampling pits and performing dye injections (Dahm and Valett 1996), or (5) injection of conservative tracers (e.g., Triska et al. 1989; Harvey and Bencala 1993). Any attempt to characterize patterns of groundwater–stream water interaction can benefit from a multiscale approach, as well as the use of multiple, complementary methods. It is our hope that the techniques we have presented here will broaden the range of tools available and will contribute to further research on the ecological roles of groundwater–stream water interaction.

Acknowledgments

We would like to acknowledge Lenny Weaver-Baxter and Joe Giersch for assistance in the field. We are also grateful to Mark Potter and Will Young for assistance with materials and construction of the piezometers and installation device. Additional experience, insight, and ideas came from Jack Stanford, Bob Beschta, Hiram Li, Judy Li, Kristopher Wright, and Christian Torgersen, as well as David Pepin, Jim Craft, and Bonnie Ellis.

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