ground water

Pumping Test Evaluation of Stream Depletion Parameters

by Hilary K. Lough[1] and Bruce Hunt[2]

Abstract

Descriptions are given of a pumping test and a corresponding analysis that permit calculation of all five hydrogeological parameters appearing in the Hunt (2003) solution for stream depletion caused by ground water abstraction from a well beside a stream. This solution assumes that flow in the pumped aquifer is horizontal, flow in the overlying aquitard or system of aquitards is vertical, and the free surface in the top aquitard is allowed to draw down. The definition of an aquitard in this paper is any layer with a vertical hydraulic conductivity much lower than the horizontal hydraulic conductivity of the pumped aquifer. These "aquitards" may be reasonably permeable layers but are distinguished from the pumped aquifer by their hydraulic conductivity contrast. The pumping test requires a complete set of drawdown measurements from at least one observation well. This well must be deep enough to penetrate the pumped aquifer, and pumping must continue for a sufficient time to ensure that depleted streamflow becomes a significant portion of the well abstraction rate. Furthermore, two of the five parameters characterize an aquitard that overlies the pumped aquifer, and values for these parameters are seen to be dependent upon the initial water table elevation in the aquitard. The field test analyzed herein used a total of eight observation wells screened in the pumped aquifer, and measurements from these wells gave eight sets of parameters that are used in a sensitivity analysis to determine the relative importance of each parameter in the stream depletion calculations.

Introduction

Stream depletion occurs when a well beside a stream either induces or increases seepage directly from the stream or intercepts seepage that would otherwise enter the stream. Analytical solutions for this problem have been obtained by Theis (1941), Glover and Balmer (1954), Grigoryev (1957), Hantush (1965), and Hunt (1999, 2003). The Hunt (2003) stream depletion solution is appropriate for the analysis of stream depletion in any system where the effective vertical hydraulic conductivity is lower than the horizontal hydraulic conductivity, provided that horizontal flow in any layers overlying the pumped aquifer is not significant. In the case where horizontal flow in overlying layers is significant, the solution may overestimate stream depletion in the transition phase between the point where vertical leakage decreases and the stream depletion rate increases.

All the analytical solutions, however, require values for parameters that characterize the pumped aquifer, the streambed, and, in some cases, aquitards that overlie the pumped aquifer. Field tests to evaluate stream depletion parameters have been carried out in numerous studies, including Sophocleous et al. (1988), Weir (1999) (later reported in Hunt et al. 2001), Chen (2000), Landon et al. (2001), Nyholm et al. (2002), Kelly and Murdoch (2003), Kollet and Zlotnik (2003), Cardenas and Zlotnik (2003), and Fox (2003, 2004). Some of these investigations have made direct measurements of streambed conductivities, while others have carried out pumping tests and used analytical or numerical solutions to analyze the resulting data. Pumping test measurements can be expected to give representative spatial averages for stream depletion parameters for the conceptual hydrogeological settings that

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are assumed when using analytical or numerical models for predictive purposes. Therefore, an ideal method for making parameter estimates would use a pumping test with drawdown measurements from at least one observation well to obtain values for all the required parameters. This paper explains and illustrates the use of such a method.

Test Site Description

Tests were conducted at the site used by Weir (1999) and described by Hunt et al. (2001). The pumped well was located 55 m from the edge of a small stream called the Doyleston Drain, which has a history of stream depletion problems caused by irrigation pumping. The Drain is 2.5 m wide with a silt- and gravel-lined streambed ~1.0 m below the ground surface. Underlying material consists of fluvial deposits of sand, silt, and gravel capped on top by 1 to 2 m of relatively fine-grained material. Well logs and a ground-penetrating radar survey carried out at the site by Lough (2004) showed that a significant amount of layering exists beneath the ground surface. A relatively shallow water table fluctuates seasonally at depths anywhere between the stream-free surface and a point 2 to 3 m below the ground surface.

The pumped well, which was screened between 8.87 and 10.87 m below the ground surface, and four observation wells, which were drilled to a depth of 6.2 m below the ground surface, were already in place from work carried out earlier by Weir (1999). Weir also used an irrigation well, screened from 7.39 to 9.39 m below the ground surface, as an observation well, but measurements from this well were inconsistent with measurements recorded in the 6.2-m-deep observation wells and were disregarded by Weir (1999) and Hunt et al. (2001).

Three new wells were drilled on the far side of the stream for these tests. Two of these wells were 6 m deep, and the third well was 10 m deep to allow for its possible use as a pumped well. However, the first pumping test carried out with these wells gave drawdowns that failed to decrease monotonically with distance from the pumped well, with the 10-m-deep well on the far side of the drain giving greater drawdowns than drawdowns recorded in the shallower wells that were closer to the pumped well. This result led to the conclusion that the 6-m-deep wells were too shallow to penetrate the pumped aquifer. Therefore, three of the wells were deepened to 10 m, and three new wells were installed to a depth of ~10 m, which gave a total of eight observation wells screened at the same depth as the pumped well. Drawdowns measured in these deeper wells during two additional pumping tests decreased monotonically with distance from the pumped well and were considered to give reliable measurements of drawdown in the pumped aquifer. The final site layout is shown in Figure 1, and a sketch of the geology is shown in Figure 2. More details of the well layout and logs for four of the wells are given by Lough (2004).

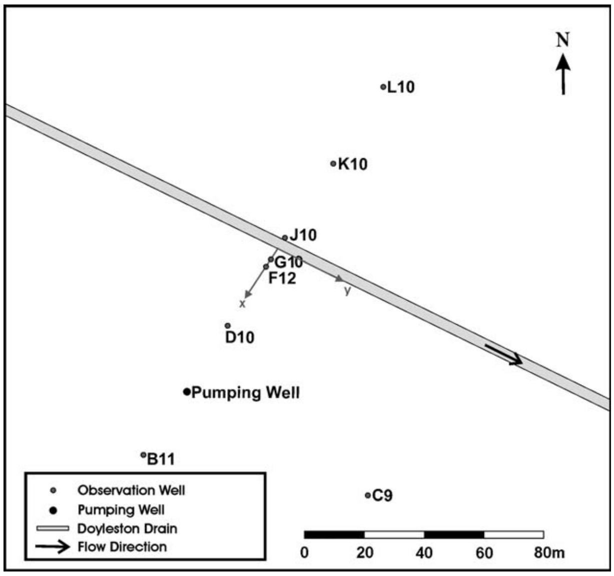

Figure 1. The test site layout.

Analytical Solution Behavior

The aquifer layering described previously and the observed responses of observation wells all suggested that drawdowns in the pumped aquifer could be described most appropriately with the analytical solution obtained by Hunt (2003). The geology assumed for this solution is shown in Figure 3, where Q = well abstraction rate; T and S are transmissivity and storativity of the semiconfined pumped aquifer, respectively; W = stream width; L = distance between the pumped well and the stream; B' = saturated thickness of the aquitard; K' and S_y are aquitard hydraulic conductivity and specific yield, respectively; B'' = aquitard thickness between the streambed and the pumped aquifer; and K'' = aquitard hydraulic conductivity beneath the stream. The solution models flow in the pumped aquifer as horizontal, flow in the overlying aquitard as vertical, and allows the water table in the aquitard to draw down as water is pumped from the underlying aquifer. The stream is modeled by setting the difference between horizontal flows on each side of the stream in the pumped aquifer equal to vertical flow entering the pumped aquifer from the overlying stream. When no stream is present, flow to a well in this type of semiconfined aquifer system is described by the Boulton delayed-yield solution, as shown by Boulton (1973) and Cooley and Case (1973).

Figure 4 shows a typical response in an observation well halfway between a stream and a pumped well when no stream is present (curve abcd) and when a stream is present (curve aefg). Segments abc and aef have similar shapes, although aquifer recharge provided by the stream causes smaller drawdowns along segment ef. The storativity, S, and transmissivity, T, of the pumped aquifer determine the horizontal position and slope, respectively, of segments ab and ae. The ratio of vertical aquitard hydraulic conductivity, K', to aquitard thickness, B', controls the vertical position of segment bc, so that decreasing K'/B' slows vertical recharge into the pumped aquifer and moves segment bc upward (the ratio K'/B' is referred to as the aquitard resistance term). Similar comments apply to segment ef for the case when a stream is present, except that decreasing recharge from the stream by

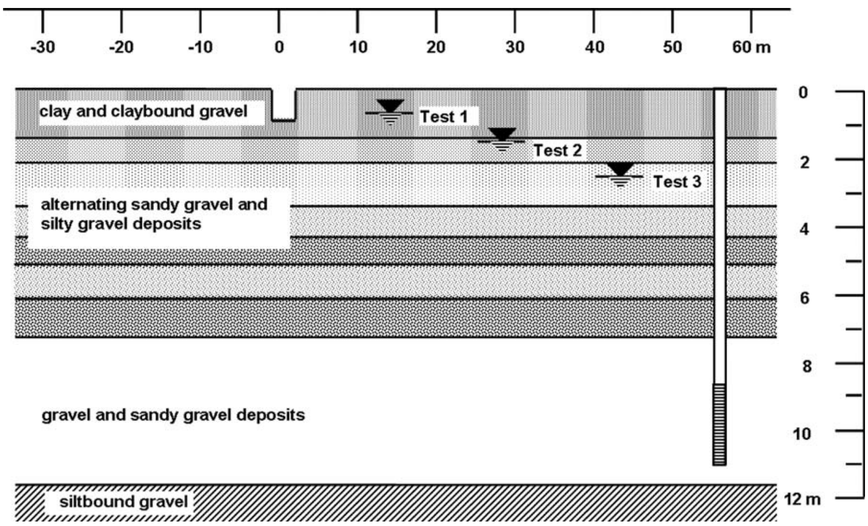

Figure 2. Conceptual hydrogeology for the test site and water levels for each of the three pumping tests.

decreasing the streambed conductance term, λ, also moves segment ef upward. (Hunt [2003] defines $\lambda = (K''/B'')W$, where K''/B'' = ratio of aquitard permeability to thickness beneath the stream and W = stream width. Curve aefg approaches curve abcd as λ approaches zero.) The specific yield, S_y, of the aquitard and transmissivity, T, of the pumped aquifer control the horizontal position and slope, respectively, of segment cd. Since the transmissivity of the pumped aquifer does not change with time, the asymptotic slopes of segments ab and cd are identical. (Sometimes, however, the lengths of segments ab and cd are not sufficient to allow these asymptotes to be approached very closely.) In some cases, the slope of cd will be flatter than the slope of ab in a plot of measured drawdowns. This indicates that horizontal flow in the overlying aquitards is significant, which leads to an increase in the transmissivity of the system. When a stream is present, segment fg eventually approaches a horizontal asymptote, indicating that a condition of steady flow has been approached. At this point, the stream depletion rate equals the well abstraction rate, and vertical recharge from the aquitard has ceased at all points away from the stream.

Since the vertical position of segment ef is controlled by values of both K'/B' and λ, values of these two parameters will not be determined uniquely by a drawdown curve that fails to extend beyond point f. On the other hand, since λ determines the horizontal asymptote of segment fg, a drawdown curve that includes all three of the segments ae, ef, and fg can be expected to give unique values for all five of the aquifer and streambed parameters T, S, K'/B', S_y, and λ.

Field Measurement Results

A total of three different pumping tests were carried out. The first test was run in August 2003, when the water table was ~0.97 m below the ground surface. Measurements in this test indicated that only one observation well was deep enough to penetrate the pumped aquifer. Therefore, three of the existing wells were deepened and three new wells were installed, giving a total of eight observation wells screened at about the same depth as the pumped well, before carrying out a second test in December 2003. The water table at the start of the second test was ~1.45 m below the ground surface. A third test was run in February 2004, when the steam was completely dry and no longer served as a recharge source for the pumped aquifer. The water table was ~2.70 m below the ground surface at the start of the third test. Water table elevations at the start of each test are shown in Figure 2.

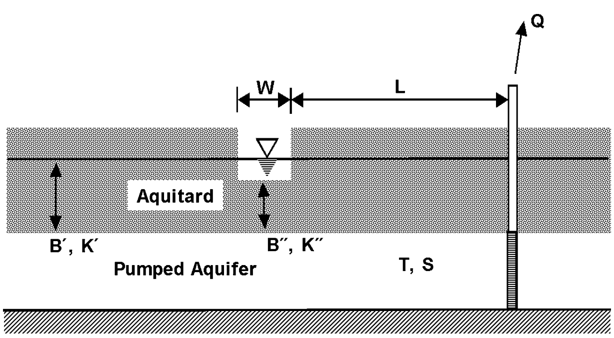

Figure 3. The conceptual model for the Hunt (2003) solution.

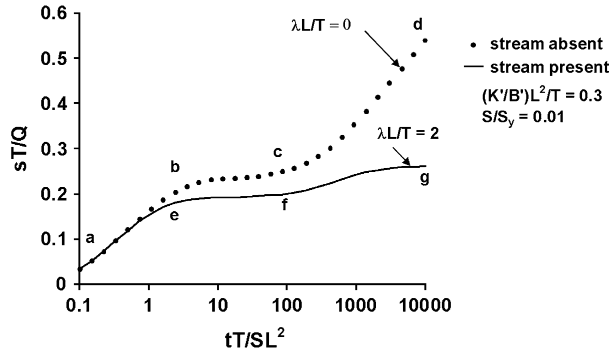

Figure 4. Drawdown curves for an observation well halfway between a stream and a pumped well.

The first pumping test abstracted 12 L/s from the pumped well over a period of 48 h. The second test abstracted 12 L/s from the pumped well over a period of 72 h, and the third test abstracted ~10 L/s over a period of 117 h. Automatic water level recorders were used to measure drawdowns both during the period of pumping and during the period of recovery after pumping stopped. Measurements in a nearby well were used to obtain relatively small corrections for piezometric-level background trends during the tests, and abstracted flows were discharged back into the Doyleston Drain at a point 300 m downstream from the pumped well.

Well abstraction rates were not identical for all three tests. Furthermore, abstraction rates for the third test were decreased during the test to avoid obtaining excessively large drawdowns. Therefore, since drawdowns for the Hunt (2003) and Boulton (1973) solutions are both directly proportional to the abstracted flow, all drawdowns were normalized by dividing their values by the abstracted flow to enable comparison between the tests and to correct for the pumping rate drop in test 3. This technique is only appropriate for correcting for small changes in the pumping rate and where the time scale used for the analysis is coarse in comparison to the expected recovery period caused by the drop in pumping rate. Normalized drawdown and recovery values for well L10 are plotted with solid lines in Figure 5 together with fitted curves, shown with filled circles, calculated from the corresponding analytical solutions. This particular well was ~60 m from the stream and was on the far side of the stream from the pumped well.

Analytical solutions were fitted to experimental measurements by using the analytical solution behavior described earlier to adjust aquifer parameters until satisfactory agreement was obtained between calculated and measured values. In particular, T and S were adjusted to achieve the correct slope and horizontal position, respectively, of segment ae in Figure 4, K'/B' and λ were decreased to raise the vertical elevation of segment ef, and λ was decreased to increase the vertical elevation of the steady-flow asymptote at point g. As stated subsequently, the steady-flow asymptote at point g was not well defined in these measurements, and the calculation of λ was aided considerably by modeling both drawdown and recovery phases. The writers prefer to think of this curve-fitting process as a process of successive approximation rather than trial and error, since the result from an adjustment in a parameter value is used in a systematic way to obtain an improved value for the parameter in the next step.

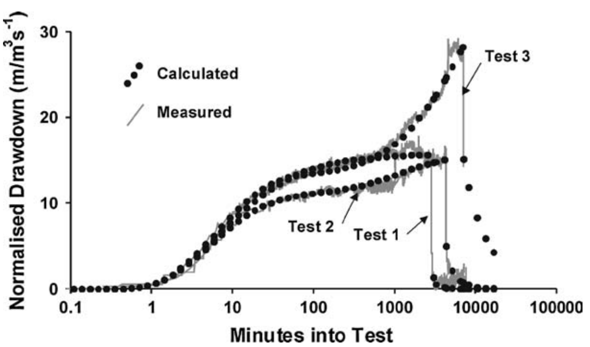

Figure 5. Drawdown and recovery plots for well L10.

There are at least four significant conclusions that can be drawn from studying Figure 5. The first conclusion concerns the appropriateness of the analytical solutions used for the analysis. Well logs show that significant layering occurs between the ground surface and the pumped aquifer. However, these well logs were made over a period of 44 years by three different drilling firms, and it is not possible to use them to estimate either relative conductivities or the lateral extent of different layers. Hunt and Scott (2005) have shown that the Boulton solution, which gives drawdowns for a pumped aquifer overlain by a single aquitard containing a water table, can be used to describe flow to a well in an aquifer overlain by N aquitard layers if the specific yield of the top aquitard, which contains the water table, is much greater than the storativity for any of the other layers and if the ratio K'/B' is replaced with the following equivalent value:

$$(K'/B')_{\text{eq}} = \frac{1}{\sum_{i=1}^{N} \frac{B_i'}{K_i'}} \tag{1}$$

where B_i' and K_i' are the thickness and hydraulic conductivity, respectively, of aquitard i. The implication is that the ratio K'/B' in the Hunt (2003) stream depletion solution should also be replaced by the equivalent value shown in Equation 1 for the case where there are multiple aquitard layers above the pumped aquifer. Comparisons by Hunt and Scott (2005) with a MODFLOW model for flow to a well in an infinite aquifer with no stream show exceptionally close agreement when $(K'/B')_{\text{eq}}$ is calculated from Equation 1, provided that the largest aquitard transmissivity does not exceed ~5% of the pumped aquifer transmissivity. If this condition is violated, then the transmissivity of the aquifer-aquitard system at large times, when the aquifer and aquitard system behave as a single unconfined aquifer, is given by the sum of transmissivities of the pumped aquifer and the aquitard layers. Since the asymptotic slope of segment cd in Figure 4 is inversely proportional to aquifer transmissivity, this causes segment cd to have a smaller asymptotic slope than segment ab, which has a slope determined by the transmissivity of the pumped aquifer alone. Inspection of the later part of the drawdown curve for test 3, which was carried out when the stream was dry and the Boulton solution was applicable, shows no evidence of having a smaller asymptotic slope than the asymptotic slope of the first segment, corresponding to segment ab in Figure 4. This suggests that the Boulton (1973) and Hunt (2003) solutions are appropriate models for analyzing the pumping test data for test 3 and for tests 1 and 2, respectively.

The second conclusion concerns the use of both drawdown and recovery data for fitting analytical solutions to the field data. It was pointed out earlier that the streambed conductance term, λ, can be determined uniquely only if drawdowns have been measured for a sufficiently

Table 1
Parameter Values for Each of the Three Tests

Test	Well	K'/B' (s^{-1})	S_y	S	T (m^2/s)	λ (m/s)
1	L10	1×10^{-9}	0.005	3×10^{-4}	1.0×10^{-2}	8×10^{-5}
2	L10	1×10^{-7}	0.004	3×10^{-4}	1.0×10^{-2}	8×10^{-5}
	K10	1×10^{-7}	0.012	3×10^{-4}	3.5×10^{-3}	8×10^{-5}
	J10	3×10^{-8}	0.002	5×10^{-5}	1.0×10^{-3}	1×10^{-4}
	G10	4×10^{-8}	0.003	9×10^{-5}	1.5×10^{-3}	9×10^{-5}
	F12	6×10^{-8}	0.002	8×10^{-5}	1.5×10^{-3}	1×10^{-4}
	D10	8×10^{-8}	0.007	1×10^{-4}	2.1×10^{-3}	2×10^{-4}
	B11	2×10^{-7}	0.020	1×10^{-4}	3.3×10^{-3}	2×10^{-4}
	C9	2×10^{-7}	0.004	2×10^{-4}	3.0×10^{-3}	8×10^{-5}
3	L10	2×10^{-7}	0.020	3×10^{-4}	1.0×10^{-2}	
	K10	4×10^{-7}	0.080	3×10^{-4}	3.5×10^{-3}	
	J10	3×10^{-7}	0.140	5×10^{-5}	1.0×10^{-3}	
	G10	3×10^{-7}	0.100	9×10^{-5}	1.5×10^{-3}	
	F12	3×10^{-7}	0.130	8×10^{-5}	1.5×10^{-3}	
	D10	3×10^{-7}	0.080	1×10^{-4}	2.1×10^{-3}	
	B11	3×10^{-7}	0.060	1×10^{-4}	3.3×10^{-3}	
	C9	3×10^{-7}	0.040	2×10^{-4}	3.0×10^{-3}	

long time to allow segment fg in Figure 4 to approach its horizontal asymptote. Although drawdown measurements shown in Figure 5 were taken until drawdowns appeared to have stopped changing with time, horizontal asymptotes for tests 1 and 2 are not clearly defined in Figure 5. However, since time translation and superposition allow recovery drawdowns to be computed from the difference of two solutions, and since the first of these two solutions calculates drawdowns at larger times than the time at which pumping stops, recovery measurements and calculations include a period of time during which λ has a significant and unique influence upon solution behavior. As a result, it was a simple process to obtain a unique value for λ when matching measured and calculated curves in Figure 5 for tests 1 and 2.

The third conclusion concerns the difference in drawdown behavior for tests 1 and 2. The stream was flowing for both tests, and both drawdown curves were measured in the same well with the same abstracted flow from the pumped well. Yet, significant differences exist between these two curves. In fact, the only difference between these two tests was that the water table was ~0.5 m lower at the start of test 2 than at the start of test 1. The calculated value for K'/B' for well L10 in test 2 was 2 orders of magnitude larger than for test 1, so from this it was concluded that the water table for test 1 was in a layer of very fine material but was below this layer for test 2. Since increasing K'/B' increases vertical recharge and decreases drawdowns in the pumped aquifer, and since drawdowns at intermediate times for test 2 lie below drawdowns measured for test 1, this explanation is consistent with the observed behavior. It is not possible to draw a similar conclusion about drawdowns measured during test 3, since recharge from the stream is seen from Figure 4 to have a significant effect on drawdown magnitudes, and stream recharge was absent during test 3 but was present during tests 1 and 2.

The fourth conclusion concerns values calculated for the specific yield, or effective porosity, from tests 1, 2, and 3. Specific yield values calculated from the data shown in Figure 5 were 0.005 and 0.004 for tests 1 and 2, respectively, but 0.02 for test 3. Since no other deeper wells existed when test 1 was carried out, it is not possible to compare specific yield values for tests 1 and 2 in any of the other wells. However, specific yield values for tests 2 and 3 can be compared for all eight of the deeper wells, and this comparison shows that specific yield values for test 3 are typically ~1 order of magnitude greater than specific yield values for test 2. Since the water table for test 3 was ~1.25 m lower than for test 2, and since the specific yield that appears in the analytical solution is the specific yield at the free surface, it appears that the water table location also has a significant effect upon the value of specific yield that is used in the analytical solution.

Values calculated for all the aquifer and streambed parameters, for all three tests, are given in Table 1, where K'/B' = hydraulic conductivity to thickness ratio for the overlying aquitard system (the aquitard resistance term), S_y = specific yield of the aquitard containing the water

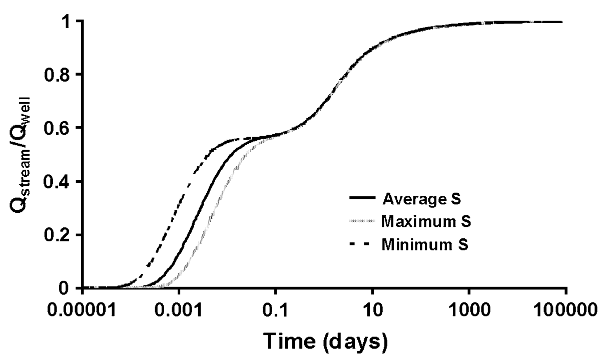

Figure 6. Sensitivity of stream depletion estimates to changes in the pumped aquifer storativity.

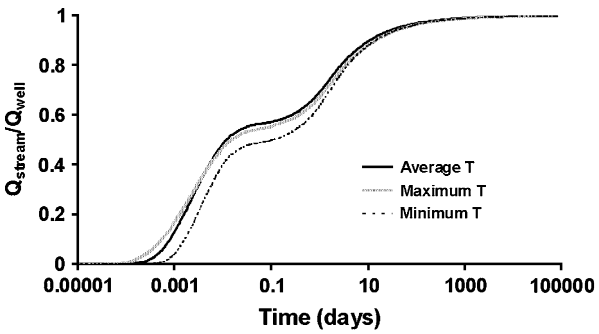

Figure 7. Sensitivity of stream depletion estimates to changes in the pumped aquifer transmissivity.

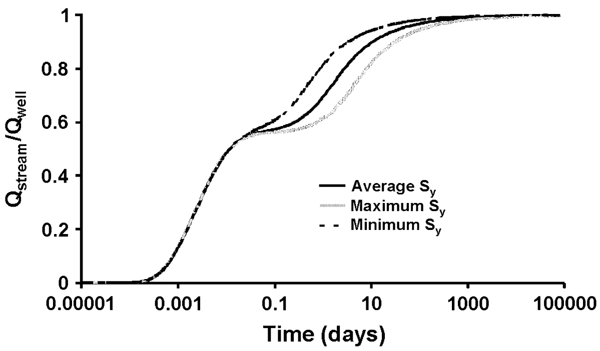

Figure 9. Sensitivity of stream depletion estimates to changes in the aquitard specific yield.

table, S = storativity of the pumped aquifer, T = transmissivity of the pumped aquifer, and λ = the streambed conductance term. All parameter values were allowed to vary from one well to the next. However, since S and T characterize the pumped aquifer, these parameters were kept constant at each well for all three tests. Since flow occurred in the stream for both tests 1 and 2, values of λ for well L10 were also set equal to the same value for tests 1 and 2. On the other hand, since changes in water table level can have an effect upon K'/B' and S_y, and since initial water table levels were significantly different for all three tests, values of K'/B' and S_y for each well were allowed to have different values for each test.

Aquifer parameter values obtained from this study were not compared with values obtained by Hunt et al. (2001). As stated previously, observation wells used in the earlier study were found in this study to be too shallow to penetrate the pumped aquifer. Therefore, the measured drawdowns were not representative of drawdowns in the pumped aquifer and could not be analyzed to give meaningful hydrogeological parameters.

Stream Depletion Estimates

Lough (2004) has carried out a sensitivity analysis in which stream depletion was calculated using the arithmetic mean of values from the eight observation wells for test 2 for four of the five parameters and setting the fifth parameter equal to its minimum, average, and maximum values. The results, which are shown in Figures 6 through 10, give an indication of the relative importance of obtaining accurate values for each of the five aquifer and streambed parameters.

Figure 6 shows that uncertainties in the aquifer storativity, S, create changes only at very early times as this parameter controls the initial release of water from elastic storage. Figure 7 shows that the predicted stream depletion is relatively insensitive to uncertainties in the aquifer transmissivity, T, which controls the flow rate through the aquifer. Uncertainties in the aquitard resistance term, K'/B', affect the vertical recharge rate, and Figure 8 shows that this leads to significant changes only at intermediate values of time as vertical recharge from overlying aquitards ceases at larger times. Figure 9 shows that uncertainties in the aquitard specific yield, S_y, affect the time of transition from vertical recharge from the aquitard system and the stream to complete stream depletion as this parameter controls storage in the overlying layers. The streambed conductance term, λ, controls the flow rate from the stream to the aquifer, and Figure 10 shows that this parameter influences stream depletion behavior at both intermediate and large values of time.

Conclusions

Measurements from three different pumping tests, all carried out at the same location, have been matched with analytical solutions obtained by Boulton (1973) and Hunt (2003) to obtain values for aquifer and streambed

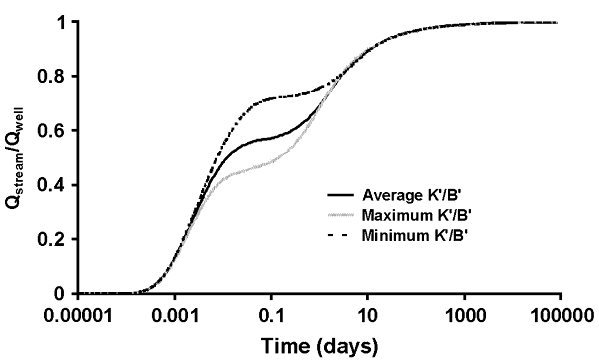

Figure 8. Sensitivity of stream depletion estimates to changes in the aquitard resistance term.

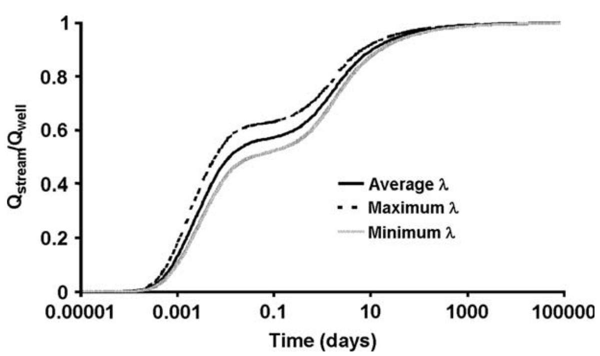

Figure 10. Sensitivity of stream depletion estimates to changes in the streambed conductance term, λ.

parameters. Each parameter influences a different part of a drawdown curve, which suggests that an accurate and complete set of drawdown measurements from a single observation well, provided it is screened at the same depth as the pumped well, is sufficient to obtain unique values for all five parameters. This hypothesis has been tested by analyzing data obtained by Lough (2004). The results show that a pumping test must be carried out for a sufficiently long period of time to allow leakage from the stream to have a significant influence upon the measured drawdown curve. Calculations also show that analyzing drawdown recovery measurements after a pump has been turned off can significantly shorten the duration of a pumping test required for this analysis. Each of the three pumping tests started with different water table elevations in an overlying aquitard, and results show that values for the aquitard resistance term, K'/B', and specific yield, S_y, can be influenced significantly by the initial level of the water table. Finally, parameter values calculated from the pumping tests have been used in a sensitivity analysis to determine the relative importance of each of the five parameters in stream depletion calculations.

Program Availability

Programs used for these calculations are contained in the spreadsheet software Function.xls, which has been described by Hunt (2005). This software and a manual describing its use may be obtained without charge from the Web site http://www.civil.canterbury.ac.nz/staff/bhunt.asp.

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References

Boulton, N.S. 1973. The influence of delayed drainage on data from pumping tests in unconfined aquifers. *Journal of Hydrology* 19, no. 2: 157–169.

Cardenas, M.B., and V.A. Zlotnik. 2003. Three-dimensional model of modern channel bend deposits. *Water Resources Research* 39, no. 6: 1–1 to 1–12.

Chen, X. 2000. Measurement of streambed hydraulic conductivity and its anisotropy. *Environmental Geology* 39, no. 12: 1317–1324.

Cooley, R.L., and C.M. Case. 1973. Effect of a water table aquitard on drawdown in an underlying pumped aquifer. *Water Resources Research* 9, no. 2: 434–447.

Fox, G.A. 2004. Evaluation of a stream aquifer analysis test using analytical solutions and field data. *Journal of the American Water Resources Association* 40, no. 3: 755–763.

Fox, G.A. 2003. Estimating streambed and aquifer parameters from a stream/aquifer analysis test. *Proceedings of the Twenty-Third Annual AGU Hydrology Days*, Colorado State University, Fort Collins, Colorado, 68–79.

Glover, R.E., and G.G. Balmer. 1954. River depletion from pumping a well near a river. *Transactions of the American Geophysical Union* 35, no. 3: 468–470.

Grigoryev, V.M. 1957. The effect of streambed siltation on well-field yield in alluvial aquifers [in Russian]. *Water Supply and Sanitation* 6: 110–118.

Hantush, M.S. 1965. Wells near streams with semipervious beds. *Journal of Geophysical Research* 70, no. 12: 2829–2838.

Hunt, B. 2005. Visual basic programs for spreadsheet analysis. Computer note. *Ground Water* 43, no. 1: 138–141.

Hunt, B. 2003. Unsteady stream depletion when pumping from semiconfined aquifer. *Journal of Hydrologic Engineering* 8, no. 1: 12–19.

Hunt, B. 1999. Unsteady stream depletion from ground water pumping. *Ground Water* 37, no. 1: 98–102.

Hunt, B., and D. Scott. 2005. An extension of the Hantush and Boulton solutions. *ASCE Journal of Hydrologic Engineering* 10, no. 3: 223–236.

Hunt, B., J. Weir, and B. Clausen. 2001. A stream depletion field experiment. *Ground Water* 39, no. 2: 283–289.

Kelly, S.E., and L.C. Murdoch. 2003. Measuring the hydraulic conductivity of submerged sediments. *Ground Water* 41, no. 4: 431–439.

Kollet, S.J., and V.A. Zlotnik. 2003. Stream depletion predictions using pumping test data from a heterogeneous stream-aquifer system (a case study from the Great Plains, USA). *Journal of Hydrology* 281, no. 1–2: 96–114.

Landon, M.K., D.L. Rus, and F.E. Harvey. 2001. Comparison of instream methods for measuring hydraulic conductivity in sandy streambeds. *Ground Water* 39, no. 6: 870–885.

Lough, H.K. 2004. The evaluation of hydrological parameters for stream depletion assessments. M.E. thesis, Department of Civil Engineering, University of Canterbury, Christchurch, New Zealand.

Nyholm, T., S. Christensen, and K.R. Rasmussen. 2002. Flow depletion in a small stream caused by ground water abstraction from wells. *Ground Water* 40, no. 4: 425–437.

Sophocleous, M., M.A. Townsend, L.D. Vogler, T.J. McClain, E.T. Marks, and G.R. Coble. 1988. Experimental studies in stream-aquifer interaction along the Arkansas River in Central Kansas—Field testing and analysis. *Journal of Hydrology* 98, no. 3–4: 249–273.

Theis, C.V. 1941. The effect of a well on a nearby stream. *Transactions of the American Geophysical Union* 22, no. 3: 734–738.

Weir, J. 1999. Stream depletion from groundwater pumping. M.E. thesis, Department of Civil Engineering, University of Canterbury, Christchurch, New Zealand.