



Analysis of pumping-induced stream–aquifer interactions for gaining streams

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Abstract

This paper presents analytical solutions that can be used to evaluate stream infiltration and baseflow reduction induced by groundwater pumping in nearby aquifers. Critical time, infiltration reach, and travel times can also be calculated to determine the hydraulic connectivity between the well and the stream. The critical time indicates the earliest time of reversal of hydraulic gradient occurring along the stream–aquifer interface, the infiltration reach is the stream segment where stream water recharges the aquifer, and the shortest travel time for the stream water particle to get into a pumping well is along the meridian line. The transient features of the two stream depletion components, baseflow reduction and stream infiltration, are evaluated separately. The rate of baseflow reduction can be greater than the rate of stream infiltration for a stronger gaining stream. However, for a given distance between the stream and well, a higher pumping rate or a weaker gaining stream results in higher rate of stream infiltration, although the total depletion rate is the same for different pumping rates or varied hydraulic gradient of the baseflow. When a steady-state condition is assumed for a transient flow, the rate and volume of stream infiltration can be overestimated; this overestimation can be very significant in the early stage of pumping.

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Keywords: Stream depletion; Baseflow reduction; Induced infiltration; Groundwater pumping

1. Introduction

Theis (1941) developed an analytical solution to calculate stream depletion from groundwater extraction in nearby aquifers. His solution was derived on the basis of simplified stream–aquifer systems, yet it has been a valuable tool in the analysis of stream–aquifer interactions. Glover and Balmer (1954) re-derived the Theis solution and presented it in another

form. Application examples of the Theis model in the evaluation of stream depletion include US Bureau of Reclamation (1960), Jenkins (1968), and Glover (1974). Wallace et al. (1999) extended the Theis model to the analysis of depletion process for cyclic pumping wells. Other analytical solutions for evaluation of stream–aquifer interactions include Hantush (1965), Hunt (1999), and Huang (2000). More recently, while numerical models are often used to evaluate the time-dependent depletion process for more realistic stream–aquifer systems, the Theis model continues to be used for verification of

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the appropriateness of numerical model design (Sophocleous et al., 1995; Chen and Yin, 1999).

The assumptions for the Theis model, as summarized by Jenkins (1968) and Sophocleous et al. (1995), include: (1) stream fully penetrates the aquifer and forms a linear boundary; (2) stream stage remains constant in space and time; (3) the stream and the aquifer are initially at hydraulic equilibrium; (4) the aquifer has a constant thickness and homogenous hydraulic conductivity in space and time, extends semi-infinitely, and rests on a horizontal, impervious base; and (5) the well fully penetrates the aquifer and is pumped at a constant rate. Fig. 1a shows a schematic stream–aquifer system for the Theis model. Under these assumptions, all the depletion water is from the stream.

For stream–aquifer systems where there is a baseflow toward the stream (Fig. 1b), the stream gains water from the discharge of the aquifer. Pumping of groundwater in the nearby aquifer can reduce the baseflow, which otherwise would discharge to the stream, and the pumping will then induce stream infiltration to the aquifer when a sufficiently long pumping generates a reversal of hydraulic gradient from the stream to the aquifer. Apparently, the depletion process, consisting of two components, baseflow reduction and induced stream infiltration, is more complex for this stream–aquifer system with an ambient flow. The summation of the two components leads to total stream depletion. The Theis model provides only the total depletion but does not describe the two components separately.

Glover (1974) noticed the existence of natural groundwater gradient near streams but did not analyze the percentage of baseflow reduction in the total depletion; he concluded that the impact of regional groundwater flow on the calculation of total depletion is very small. While Hantush (1964) evaluated the stream depletion processes for sloping unconfined aquifers where baseflow is present, his analytical models do not separate the baseflow reduction from stream infiltration in total stream depletion. Wilson (1993) demonstrated that a separate analysis of the two components is particularly important for areas where water-quality issues are concerned. Wilson's solution (1993), taking into account the natural hydraulic gradient, is, however, good only for steady-state groundwater flows. As pointed out by

Wilson, steady-state conditions rarely exist in nature and solutions for transient flows are needed for the separate analysis of the two depletion processes. In the river valleys where pumping of groundwater may be seasonal and cyclic (for example, irrigation wells in the High Plains region of the United States), a steady-state cone of depression can be hardly established by

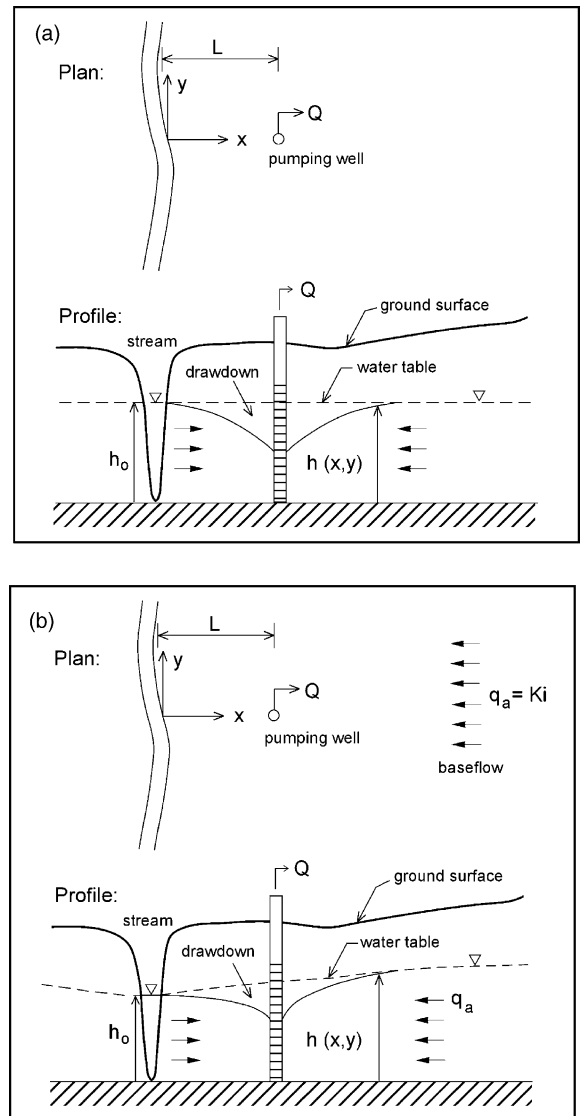


Fig. 1. Diagram showing pumping-induced stream infiltration for two hypothetical stream–aquifer systems: (a) no baseflow between stream and aquifer; (b) a regional baseflow toward the stream: h_0 , water table prior to pumping; h , hydraulic head at location x and y ; Q , pumping rate (modified from Wilson, 1993).

such short-term pumping, and the induced interactions between stream and aquifer are time-dependent. Indeed, the solutions developed by Wilson (1993) for steady state flows have limited applications. If utilized for a transient flow, they can overestimate stream depletion. To my knowledge, there is lack of analytical solutions that take into account the two depletion components separately for transient flow conditions.

Chen and Yin (2001) used the US Geological Survey's MODFLOW to design gaining stream–aquifer systems and analyzed the time-dependant rates of baseflow reduction and stream infiltration. An obvious advantage of a numerical model is that it offers options to represent more realistic hydrologic conditions of a stream–aquifer system. Although an analytical solution provides results only for much simplified systems, it is useful in helping to understand the relationship between the two depletion components and for verification of some numerical results. In addition, analytical solutions can provide some results that require a much complex procedure of a numerical simulation, for example, the determination of the critical time and the infiltration reach where the reversal of hydraulic gradient is formed along the stream–aquifer interface.

This paper will first present a solution for determining the time when reversal of hydraulic gradient begins, the section of stream where the stream infiltration has been induced for a given pumping period, and the shortest travel time for infiltrated stream water to arrive at the pumping well. Analytical solutions are presented for calculation of the rates and volumes of the baseflow reduction and the stream infiltration in pumping and post-pumping periods. The relationships between the two depletion components were analyzed. Comparison is made to determine the difference in the stream–aquifer interactions under steady state and transient conditions.

2. Critical infiltration time

The hydraulic head around the pumping well in the unconfined aquifer is

$$h = \sqrt{h_0^2 + 2h_0ix - \frac{Q}{2K\pi} [W(u_R) - W(u_I)]} \quad (1)$$

where h_0 is hydraulic head at the boundary between stream and aquifer ($x = 0$) and h_0 assumed to be constant for the entire pumping period, i the hydraulic gradient of the ambient flow perpendicular to the stream, Q the pumping rate at the well, $u_R = ((x - L)^2 + y^2)S/(4Tt)$ for the real well, $u_I = ((x + L)^2 + y^2)S/(4Tt)$ for the image well, and $W(u)$ is the well function. Here, S is specific yield for an unconfined aquifer, T aquifer transmissivity, L the distance between the stream and well, t the pumping time, and x and y are the coordinates. For a confined aquifer, Eq. (1) is written

$$h = h_0 + ix - \frac{Q}{4\pi T} [W(u_R) - W(u_I)].$$

The hydraulic gradients along the x direction was determined from Eq. (1), with

$$\frac{\partial h}{\partial x} = \frac{h_0}{h} i - \frac{Q}{2\pi Kh} \times \left[\frac{(L - x)}{r_R^2} e^{-u_R} + \frac{(x + L)}{r_I^2} e^{-u_I} \right] \quad (2)$$

where $r_R^2 = (x - L)^2 + y^2$ for the real well and $r_I^2 = (x + L)^2 + y^2$ for the image well. When the L is sufficiently large or the pumping rate Q is relatively small or the pumping duration is relatively short, the baseflow is able to provide sufficient water to the pumping well, and the cone of depression generated by the pumping well will never intercept the stream. According to Wilson (1993), as long as $Q < i\pi TL$, the pumping will not cause the infiltration of the stream water into the aquifer. However, when $Q > i\pi TL$ and the pumping is sufficiently long, the cone of depression will reach the stream to induce surface water infiltration.

For a well that has a potential to induce stream infiltration, it is interesting to know when the stream is about to recharge the aquifer. This information is important for wellhead protection. The shortest distance between the well and the stream is along the x -axis and is equal to L . For a given time, the hydraulic gradient along the x direction at the origin ($x = y = 0$) must be equal to zero. i.e.

$$\frac{Q}{\pi TL} e^{-u_0} - i = 0 \quad (3)$$

where $u_0 = L^2S/(4Tt)$.

Rewriting Eq. (3) gives

$$t_c = -\frac{L^2 S}{4T \ln(\pi i TL/Q)} \quad (4)$$

Eq. (4) provides the earliest time (critical time) for a reversal of hydraulic gradient to occur, initially, at $x = y = 0$ (Fig. 1b). Apparently, stream infiltration is to be induced if the pumping duration is longer than this critical time. When the pumping duration $> t_c$, the length of the infiltrated stream segment gradually increases toward both up- and downstream.

According to (4), a larger pumping rate shortens the critical time, while a larger value of i , L , or S increases the critical pumping times. An increase of the distance L is the most effective way to increase the critical time because of the square feature in (4). When $\pi i TL/Q > 1$, Eq. (4) yields a negative value. This unrealistic value indicates that the pumping will never intercept the stream. When $\pi i TL/Q = 1$, $t_c \rightarrow \infty$.

Fig. 2 shows the variations of the critical time for a number of baseflow gradients. Each of the curves was calculated for a different distance L that ranges from 150 to 575 m. The various symbols on each curve indicate five baseflow gradients: $i = 0.0001$, 0.0003, 0.0005, 0.0008, and 0.001 from left to right. The values of the following parameters remain constant in these computational examples: $K = 100$ m/d, $S = 0.2$, $h_0 = 15$ m, and

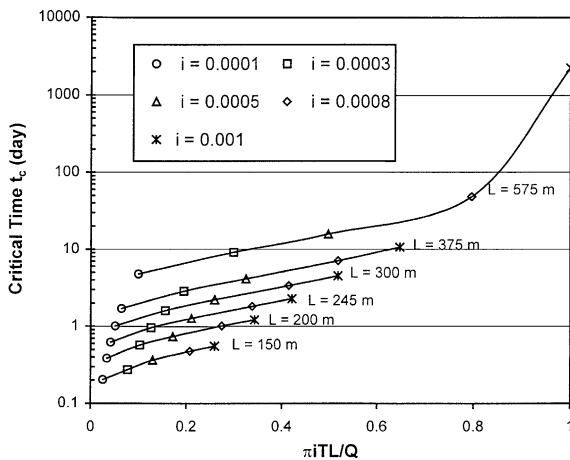


Fig. 2. Critical time of stream infiltration for varied baseflow gradients. A weaker gaining stream has a shorter critical time.

$Q = 2727$ m³/d. As shown in these curves, a larger i or L results in a longer critical time of pumping. When the well is farther from the stream, for example, $L = 575$ m, the effect of i on the critical time is stronger; a slight increase in i can lead to a significantly larger value of the critical time t_c . For some special cases ($\pi i TL/Q > 0.8$), the values of t_c can exceed 100 days. For $\pi i TL/Q = 0.995$, t_c can be as large as 2208 days (see Fig. 2 for $i = 0.001$).

Assume the stream infiltration has begun and the pumping continues. The infiltrated stream water will migrate toward the well. The water particles that move along the x -axis, the meridian line, need the shortest travel time to arrive at the well. The time can be computed using the particle tracking technique (Anderson and Woessner, 1992) such that

$$\Delta t = \frac{1}{v} \Delta x \quad (5)$$

where Δt is the time for a water particle to move a distance Δx ,

$$v = -\frac{K}{\theta} \frac{dh}{dx},$$

and here θ is the effective porosity. The migration process is generally slow. For example, for an aquifer where $K = 100$ m/d, $S = \theta = 0.2$, $h_0 = 15$ m, $Q = 2727$ m³/d, and $i = 0.001$, the migration time to the well along the meridian line will be 70.3 days for $L = 150$ m but the time increases to 393.5 days when L is doubled. Summation of the critical time and this travel time indicates the maximum duration of well pumping that brings the first stream water particles to the well. The percent of stream water increases gradually at the production well if the well continues to pump.

3. Infiltration reach

The cone of depression initially intercepts the stream–aquifer boundary at $x = y = 0$. At this time, the streamflow begins to be affected. The magnitude of the stream depletion is a function of the hydraulic gradient along the stream–aquifer boundary ($x = 0$)

$$\left. \frac{\partial h}{\partial x} \right|_{x=0} = \frac{QL}{\pi Tr^2} e^{-u} - i \quad (6)$$

where $r = (L^2 + y^2)^{0.5}$, $u = r^2S/(4Tt)$ and $T = Kh_0$. The first term in Eq. (6) represents the hydraulic gradient generated by the pumping well.

When

$$\frac{QL}{\pi Tr^2} e^{-u} < i$$

(Fig. 3a), the stream gains less water from the aquifer than before pumping began; this reduction in baseflow results in a decrease of streamflow.

$$\frac{QL}{\pi Tr^2} e^{-u} = i$$

indicates that a reversal of hydraulic gradient is to occur. When the pumping is longer, the hydraulic gradient along the stream–aquifer boundary for a given reach becomes

$$\frac{QL}{\pi Tr^2} e^{-u} > i$$

and the stream recharges the aquifer; this reach extends an equal distance up- and downstream (y' and $-y'$ in Fig. 3b). Wilson (1993) referred to y' as the stagnation point; y' is constant for a steady-state cone of depression as assumed by Wilson (1993). Stream water infiltrates into the aquifer between the two stagnation points, while the stream continues to gain water but at a reduced rate from the aquifer outside the infiltration zone ($y > |y'|$). The concept of stagnation point has also been used in the study of the capture

zone in areas where streams are not present (Javandel and Tsang, 1986; Franzetti and Guadagnini, 1996).

The stagnation point can be determined from (6) such that

$$f(y) = \frac{QL}{\pi Tr^2} e^{-u} - i = 0 \tag{7}$$

Newton’s method was used to determine y' through iterations (Carnahan et al., 1969). Generally, a longer pumping period generates a larger y' as long as the other parameters are given. When a pumping period is sufficiently long and near steady-state flow is established, e^{-u} approaches 1. The stagnation point can be simply calculated, as demonstrated by Wilson (1993), from

$$y' = L \sqrt{\frac{Q}{\pi i T L} - 1} \tag{8}$$

y' calculated using (8) represents the farthest stagnation point a pumping well can generate.

The total length of the infiltration reach at a given time is $2y'$. Fig. 4 shows the semi-ranges (y') of the infiltration zone vary in different stream–aquifer systems. As shown in Fig. 4, the y' values increase rapidly in the earlier time of pumping, and after about 60 days of continuous pumping, the growth of the y' toward up- and downstream becomes very slow. The maximum normalized semi-reach (y'/L) is smaller than 1.6. The length of the infiltration reaches reduces rapidly and is

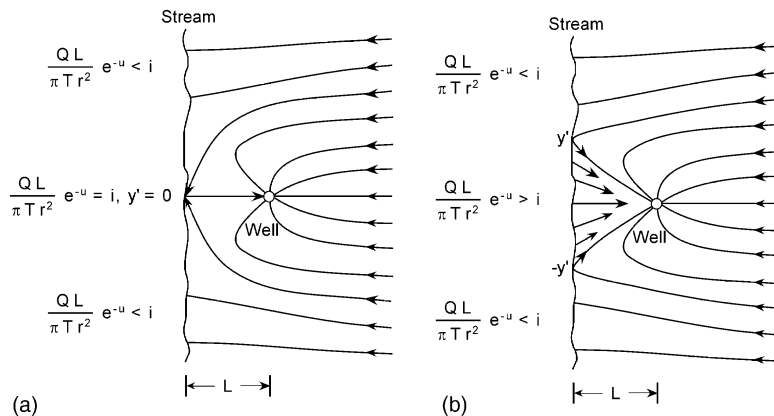


Fig. 3. Schematic showing: (a) baseflow has been reduced and a reversal of hydraulic gradient will occur at the stream–aquifer interface ($x = y = 0$), and (b) a pumping-generated infiltrating reach (between y' and $-y'$) where stream water recharges the aquifer due to a reversal of hydraulic gradient (modified from Wilson, 1993).

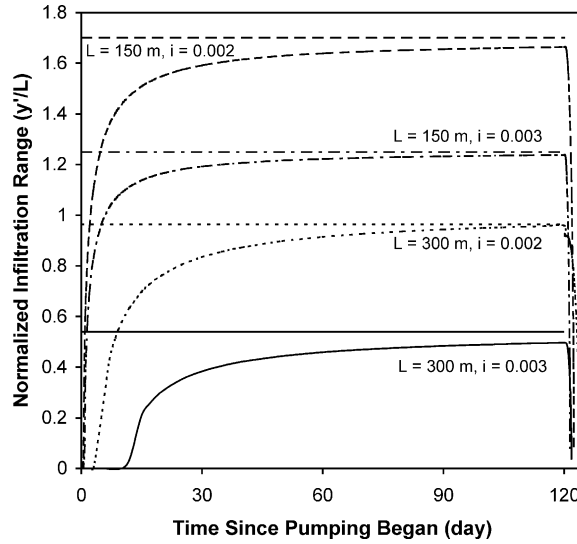


Fig. 4. Expansion of the semi-infiltration reach as the pumping continues; the length of the infiltration reach approaches a near-constant condition after 60 days of pumping. The four horizontal lines represent the constant semi-infiltration reaches for steady state conditions.

nearly zero only a few days after pumping stops. As shown in Fig. 4, y' remains zero in the first 10 days of pumping for $i = 0.003$ and $L = 300$ m.

The normalized semi-reaches would be from 0.54 to 1.7 for the four computational cases if a steady-state condition were assumed and Eq. (8) were used for the computations. They are also plotted in Fig. 4 for a comparison. As clearly shown in Fig. 4, an overestimation of infiltration reach would occur if steady-state conditions were assumed for a transient system and this overestimation can be large in the early time of the pumping. The hydraulic parameters for these examples are $K = 100$ m/d, $S = 0.2$, $h_0 = 15$ m, and $Q = 5455$ m³/d (1000 gallons/min). Two L values ($L = 150$ and 300 m) are used; for each L , two hydraulic gradients of ambient flow, $i = 0.002$ and 0.003 , are used in the computational examples.

4. Stream infiltration and baseflow reduction

Because the reversal of hydraulic gradient has occurred in the infiltration reach, the stream begins to recharge the aquifer. The stream infiltration (Q_s) is

calculated by integration between y' and $-y'$ such that

$$Q_s = \frac{2QL}{\pi} \int_0^{y'} \left(\frac{e^{-u}}{r^2} \right) dy - 2y'iT \quad (9)$$

The first term in Eq. (9) is evaluated using a numerical integration technique called Simpson's rule (Carnahan et al., 1969). For a steady-state flow condition (Wilson, 1993), $e^{-u} = 1$ and Eq. (9) becomes

$$Q_s = \frac{2Q}{\pi} \arctan\left(\frac{y'}{L}\right) - 2y'iT \quad (10)$$

In the infiltration reach, the stream no longer gains the discharge of the aquifer. This part of baseflow has been diverted to the pumping well. The rate of the reduced baseflow between y' and $-y'$ is calculated from

$$Q_{b1} = 2y'iT \quad (11)$$

For the stream reaches outside the infiltration zone, the rate of the reduced baseflow is calculated using

$$Q_{b2} = \frac{2QL}{\pi} \int_{y'}^{\infty} \left(\frac{e^{-u}}{r^2} \right) dy \quad (12)$$

The sum of Q_{b1} and Q_{b2} yields the total baseflow reduction; i.e.

$$Q_b = Q_{b1} + Q_{b2} \quad (13)$$

The total volume of the water infiltrated into the aquifer V_s is the product of Q_s and the pumping time

$$V_s = \sum Q_s \Delta t \quad (14)$$

and similarly, the total volume of baseflow reduction V_b is

$$V_b = \sum Q_b \Delta t \quad (15)$$

For water quality issues, Eqs. (7) and (9) provide critical information about the possible contamination zone and the volume of stream water leaked into the

aquifer if the river contains contaminants. For issues of stream depletion, Eqs. (11) and (12) offer the portion of total depletion that comes from the baseflow. The summation of Eqs. (9) and (13) is equal to the total rate of stream depletion described by the Theis model.

Fig. 5a shows the rates of baseflow reduction and stream infiltration (Q_b/Q and Q_s/Q) for three pumping rates. The well is located 300 m from the stream with the hydraulic gradient of the baseflow $i = 0.001$. For all the cases, the rate of baseflow reduction is greater than the rate of

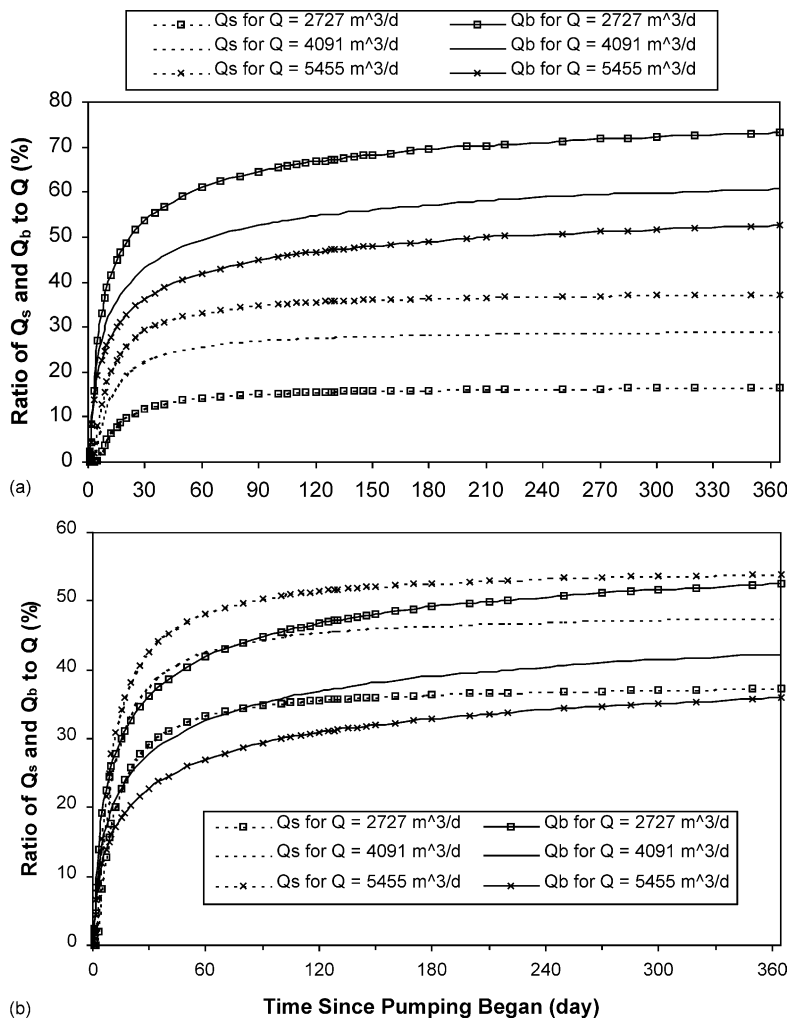


Fig. 5. Depletion curves for a constant stream–aquifer distance but three pumping rates: (a) for a stronger gaining stream ($i = 0.001$); (b) for a weaker gaining stream ($i = 0.0005$).

Table 1
Summary of the volumes of depleted baseflow and infiltrated stream water

	$i = 0.001, t = 365$ days			$i = 0.0005, t = 365$ days		
Q (m ³ /d)	2727	4091	5455	2727	4091	5455
V_s (m ³)	149,020	399,400	686,750	343,300	659,010	995,800
V_b (m ³)	657,800	810,850	926,920	463,459	551,240	617,870
V (m ³)	995,400	1,493,200	1,991,100	995,400	1,493,300	1,991,000
V_s/V_b	0.23	0.49	0.74	0.74	1.20	1.61
V_s/V (%)	15.0	26.7	34.5	34.5	44.1	50.0
V_b/V (%)	66.0	54.3	46.6	46.6	36.9	31.0
$(V_s + V_b)/V$ (%)	81.0	81.0	81.0	81.0	81.0	81.0

V , total pumped groundwater.

stream infiltration. A lower pumping rate seems to generate a higher rate of baseflow reduction but a much lower rate of stream infiltration (Fig. 5a). Note that the total depletion rate ($(Q_b + Q_s)/Q$) is the same for any Q at a given L . However, as indicated in Fig. 5, the rates of the two depletion components vary when Q changes.

For a comparison, Fig. 5b shows the rates of the baseflow reduction and stream infiltration for the three pumping rates but for a weaker gaining stream ($i = 0.0005$). The distance between the well and the stream remains the same ($L = 300$ m). Here, the rate of stream infiltration is greater than the rate of baseflow reduction for $Q = 4091$ and 5455 m³/d. Thus, a weaker gaining stream or higher pumping rates can result in more stream infiltration.

The depletion volumes of the baseflow reduction and stream infiltration were calculated using Eqs. (14) and (15). Table 1 summarizes the volumes of stream depletions for the two stream–aquifers (Fig. 5a and b) after 365 days of pumping; the ratios of the total volume of infiltrated stream water to the volume of reduced baseflow are also shown in this table. As shown in this table, at a given L and for the same pumping rate, a larger value of i leads to a higher rate of baseflow reduction and a lower rate of stream infiltration, although the total depletion is the same for varied i values.

If Eq. (10) is used for the calculation of Q_s for a transient flow condition, the total volume of infiltrated stream water V_s will be overestimated. Fig. 6 shows the overestimated stream water that infiltrates into aquifers for two baseflow gradients $i = 0.001$ and 0.0005 ; this overestimation is very significant for the first three months of pumping.

5. Residual effects

After pumping stops, the stream will continue to recharge the aquifer. This pumping-induced infiltration stops only as the hydraulic gradient at the origin ($x = y = 0$, see Fig. 1b) becomes zero; i.e.

$$\frac{Q}{\pi TL} (e^{-u} - e^{-u_a}) - i = 0 \quad (16)$$

where $u = L^2 S/(4Tt)$, $u_a = L^2 S/(4Tt_a)$, t the time since pumping begin and t_a is the time since the pumping stops.

In the post-pumping period, the aquifer will recover the storage loss during the pumping. There are two sources of water in the system: stream water

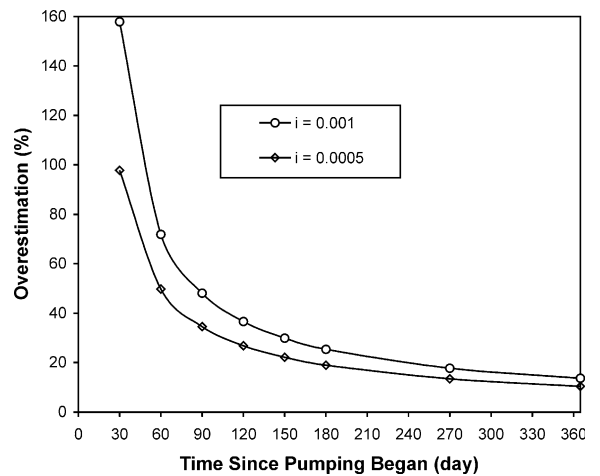


Fig. 6. Overestimation of the total volume of infiltrated stream water can be very significant if a steady-state condition is assumed for a transient flow ($Q = 2727$ m³/d, $L = 300$ m, $K = 100$ m/d, and $h_0 = 15$ m).

and baseflow. Both can replenish the aquifer. The rate of the replenishment at any time in the post-pumping period can be calculated using Eqs. (9)–(12). The stagnation point y_a in the post-pumping period must be determined from

$$f(y_a) = \frac{QL}{\pi Tr^2} (e^{-u} - e^{-u_a}) - i = 0 \quad (17)$$

where $u = (L + y_a)^2 S / (4Tt)$, and $u_a = (L + y_a)^2 S / (4Tt_a)$. Eq. (17) was derived using the superposition techniques.

The rates of baseflow reduction and stream infiltration were calculated for a period of 365 days with a pumping period of the first 120 days that is followed by a recovery period of 245 days. The rate of depletion occurring after the end of pumping is called the residual effects (Jenkins, 1968).

Two distances of stream-well, as well as two hydraulic gradients of baseflow, were considered.

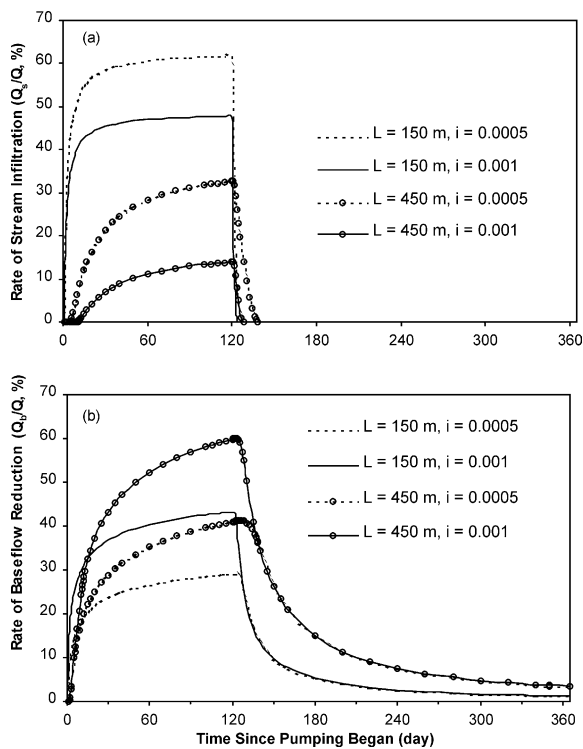


Fig. 7. Rates of baseflow reduction and stream infiltration during the pumping and post-pumping periods. Residual effects of stream infiltration often disappear shortly after the end of pumping (a), while the residual effects of baseflow reduction can last much longer (b).

The pumping rate was 4901 m³/d. Fig. 7a shows the rate of stream infiltration and Fig. 7b shows the rate of baseflow reduction. As shown in Fig. 7a, the rate of stream infiltration decreases rapidly after the cessation of pumping and it becomes zero in a relatively short time. When $L = 450$ m and $i = 0.0005$, the stream infiltration disappears at $t_a = 18.43$ days; when $L = 150$ m and $i = 0.001$, it took only 3.81 (t_a) days for the stream infiltration to vanish. Fig. 7a also indicates that stream infiltration may not occur until several days after the pumping begins. For example, when $L = 450$ m and $i = 0.001$, the stream begins to recharge the aquifer after the pumping has continued for 10.3 days.

In contrast, Fig. 7b indicates that the rate of baseflow decreases at a much slower pace in the post-pumping period. Fig. 7b seems to suggest that for a same L , a higher rate of baseflow reduction is likely to occur during the pumping period for a stronger gaining stream. However, as shown in Fig. 7b, the depletion curve for the same L merges shortly after the cessation of the pumping, and the rate of baseflow reduction becomes the same for streams of varied gaining rates. The aquifer parameters for these computations include $K = 100$ m/day, $h_0 = 15$ m, and $S = 0.2$.

The continuation of the effect of the baseflow reduction in the post-pumping period is because the aquifer storage was reduced during the pumping and the recovery of the aquifer storage consumes

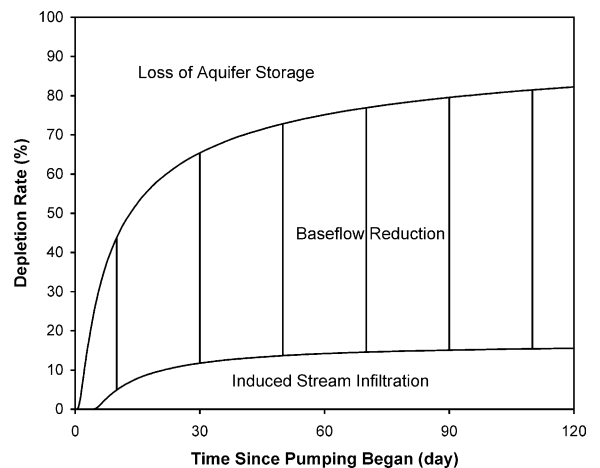


Fig. 8. The contribution of the loss of aquifer storage, baseflow reduction, and stream infiltration during the pumping of a well.

the baseflow. Fig. 8 shows the contributions of water from baseflow reduction, stream infiltration, and aquifer storage balance the pumping rate at the well. As shown in Fig. 8, the loss of aquifer storage occurs immediately after the pumping begins, the baseflow reduction takes place at a slightly later time (0.8 day after the beginning of the pumping in this case), and the infiltration of the stream water to the aquifer is not induced until about 4.6 days after the pumping has began. The hydraulic parameters for this example include $L = 300$ m and $i = 0.002$, and $Q = 5455$ m³/d.

6. Summary and conclusions

Analytical solutions are presented for the analyses of stream–aquifer interactions resulting from the pumping of groundwater in a nearby aquifer. The solutions calculate two depletion components: baseflow reduction and stream infiltration. Although the analytical solutions are for simplified gaining streams, they provide results, such as critical time and infiltration zone, that are important in understanding the migration of infiltrated stream water in the aquifer.

For a stronger gaining stream, the baseflow reduction is a major contributor in the total depletion, and its residual effect can last much longer than that arising from the induced stream infiltration. In contrast, stream infiltration becomes a major component in the total depletion when the baseflow gradient is smaller. While the total depletion is the same for different regional hydraulic gradient i or for varied pumping rates at a given location L , the depletion rates of the two components differ.

Reversal of hydraulic gradient is initially formed at $x = y = 0$ when the pumping duration approaches the critical time t_c , and an infiltration reach is formed for $t > t_c$. Stream water infiltration then occurs in the reach. When infiltrated, the stream water particles take a relatively long time to get to the pumping well compared to the critical time. For a well located several hundred meters from the stream, the infiltrated water may take many months or years to reach the well. For areas where wellhead protection is an issue, the critical

time and the travel time are both parameters to consider.

An overestimation in the length of infiltration reach and stream depletion will occur if a steady-state condition is assumed for a transient condition where the cone of depression continues to grow. This overestimation can be very significant for the early stage of pumping. For both water quality and quantity issues, transient conditions must be considered for the areas where pumping of groundwater is only seasonal.

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