Domestic Well Capture Zone and Influence of the Gravel Pack Length

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Abstract

Domestic wells in North America and elsewhere are typically constructed at relatively shallow depths and with the sand or gravel pack extending far above the intake screen of the well (shallow well seal). The source areas of these domestic wells and the effect of an extended gravel pack on the source area are typically unknown and few resources exist for estimating these. In this paper, we use detailed, high-resolution groundwater modeling to estimate the capture zone (source area) of a typical domestic well located in an alluvial aquifer. Results for a wide range of aquifer and gravel pack hydraulic conductivities are compared to a simple analytical model. Correction factors for the analytical model are computed based on statistical regression of the numerical results against the analytical model. This tool can be applied to estimate the source area of a domestic well for a wide range of conditions. We show that an extended gravel pack above the well screen may contribute significantly to the overall inflow to a domestic well, especially in less permeable aquifers, where that contribution may range from 20% to 50%; and that an extended gravel pack may lead to a significantly elongated capture zone, in some instances nearly doubling the length of the capture

zone. Extending the gravel pack much above the intake screen therefore significantly increases the vulnerability of the water source.

Introduction

Most households in rural areas of the United States, outside the service area of incorporated cities, rely on domestic wells for their water supply (McCray 2005, U.S. EPA 1997). And many of these domestic wells are constructed with a well-screen at depth and a sand or gravel pack that extends upward to the mandatory minimum depth of the well seal, which is dictated by local and state guidelines. A question commonly asked by homeowners is: Where does our water come from? The capture zones (also referred to as the source area or recharge area) of domestic wells are rarely determined. Attention has instead focused on public supply wells and their capture zones as these are regulated through U.S. EPA's source water protection program. Domestic wells, typically serving a single family, are often constructed to relatively shallow depths when compared to public or municipal water supply wells (Burow et al. 2004).

Methods for delineating well capture zones range from very simple to very complex. In general, the various approaches fall into four categories (Harter, 2008):

1. Geometric or graphical methods involve the use of a pre-determined fixed radius without any special consideration of the flow system, or the use of simplified shapes that have been pre-calculated for a range of pumping and aquifer conditions.
2. Analytical methods allow calculation of distances for protection zones using equations that can be solved using a hand calculator or microcomputer spreadsheet program.

3. Hydrogeologic mapping involves identifying the recharge zone and the source zone based on geomorphic, geologic, hydrologic, and hydrochemical characteristics of an aquifer.

4. Computer modeling methods involve devising, calibrating, and applying complex analytical or numerical models that simulate groundwater flow and contaminant transport processes.

The long-term average pumping rate of domestic wells typically ranges from less than 4 L/min [1 gallon/min] to 20 L/min [5 gallon/min]. Using the graphical method employed by California's Drinking Water Source Assessment and Protection (DWSAP) Program (California DHS, 1999), for example, the default source area of a domestic well pumping 1,233.5 m^3/year (1 acre-foot per year, the typical annual consumption of a U.S. single family household) is a circle with a radius of 15 m (~50 ft) for areal recharge of 450 mm/year (typical for very humid areas or rural residences in semi-arid areas surrounded by irrigated lawn and fields) or with a radius of 31 m (~100ft) at a recharge rate of 100 mm/year (typical of many semi-arid regions). This simple geometric approach neglects the effects of the regional groundwater flow on the capture zone of a domestic well.

On the other hand, where regional groundwater flow is dominant and local recharge is negligible, the capture zone of a domestic well can also be easily computed if the well fully penetrates the aquifer system or does not strongly affect regional groundwater flow. The width, w, of the capture zone of a domestic well is then obtained by simple mass balance (Todd 1980):

$$w = Q / (T * i) \qquad (1)$$

where Q is the pumping rate, T is the aquifer transmissivity, and i is the regional hydraulic gradient. For example, at a relatively low transmissivity, T, of 10 m^2/d, a regional groundwater gradient of 0.5% and a pumping rate of 1,233 m^3/year, the width of the capture zone is approximately 60 m

(~200 feet). At values of T typical for productive aquifers, the width of the capture zone is often on the order of 1 m - 10 m (~ 3 feet - ~30 feet) or even less.

Both, the geometric approach and equation (1) above provide simple approximations for extremely idealized conditions. Here, our objective is to determine the capture zone of a domestic well with a sand or gravel pack, completed in an unconfined aquifer, where both, recharge and regional groundwater flow are significant. We use high-resolution computer simulations to determine the source area and to explicitly determine the influence of the gravel pack on the well capture zone. For reference, we compare those to a simple analytical model of the capture zone for a low-producing well in an unconfined aquifer with recharge. Our study's focus is on rural domestic wells in irrigated agricultural regions, e.g., of the Southwestern United States, where significant recharge is due to irrigation return flows and much of the groundwater production is for irrigation purposes. Our findings have general implications that are independent of this particular climate scenario.

Conceptual Framework

Domestic wells in rural areas are assumed to be completed near the uppermost portion of a regional aquifer system. Furthermore, we assume that a significant downward gradient exists in the regional aquifer system due to recharge at the water table and due to significant groundwater production (mostly for irrigation) from the deeper portions of the aquifer system (e.g., Belitz and Phillips, 1995). Burow et al. (2004), for example, report typical recharge rates in irrigated areas in the San Joaquin Valley, California, to be on the order of 550 – 750 mm/a with the majority of recharge originating from irrigation return flows. For simplicity, regional groundwater flow is considered to be uniform around the source area of the domestic well and at steady-state. The superposition of regional groundwater flow with the downward gradient induced by water table recharge and deeper groundwater production yields a groundwater flow field that is vertically inclined relative to the slope of the water table (**Figure 1**).

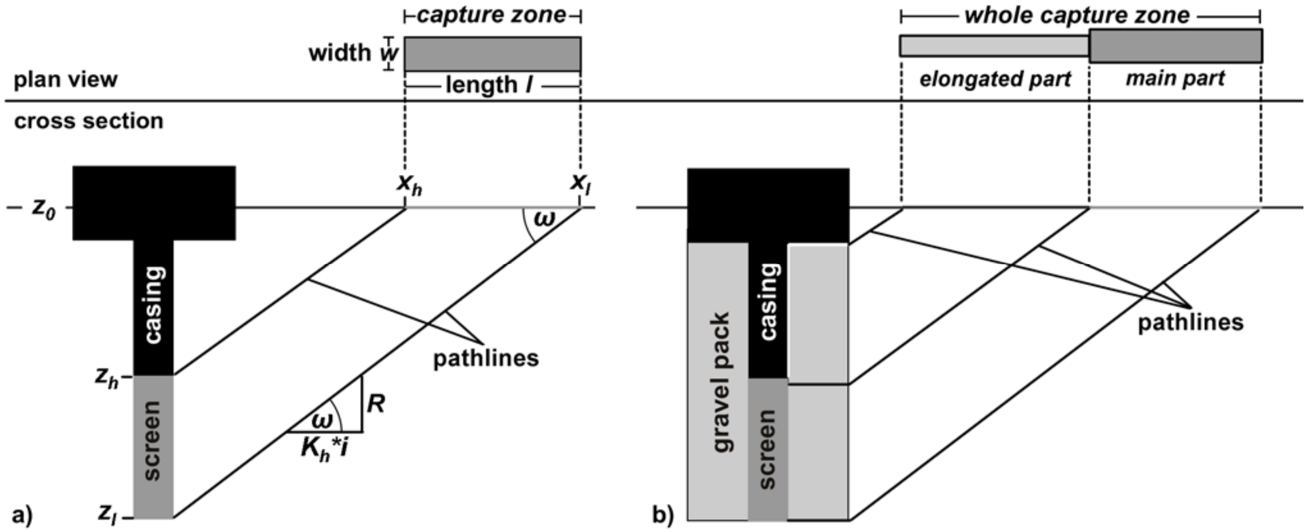

Figure 1: Conceptual framework of groundwater flow towards a partially penetrating domestic well (a) without and (b) with a gravel pack that extends for several meters to several tens of meters above the well screen. Top: plan view, bottom: cross-sectional view. Regional groundwater flow is from right to left with a vertical flow component controlled by uniform recharge at the top and aquifer pumping from large production wells dispersed in the deep part of the aquifer below. The aquifer bottom is assumed to be much deeper than the typical depth of the (relatively shallow) domestic well; *l*: length, *w*: width. All other symbols: see text for details.

A simple method to compute the approximate source area size and location is available, if we neglect the effects of the gravel pack and the effects of domestic well pumping, *Q*, on the local groundwater flow field. Then, the source area location is obtained from the length and depth of the domestic well screen, and from the angle, ω, of groundwater flow relative to the slope of the water table (**Figure 1a**):

$$x_h = x_w + \frac{z_0 - z_h}{\tan\omega} \quad (2)$$

$$x_l = x_w + \frac{z_0 - z_l}{\tan\omega} \quad (3)$$

$$l_{theo} = x_l - x_h \quad (4)$$

$$A_{theo} = Q/R \quad (5)$$

$$w_{theo} = \frac{A_{theo}}{l_{theo}} \quad (6)$$

where x_w is the location of the well (along the regional groundwater gradient), x_h is the location of the downgradient edge of the recharge (source) area, x_l is the location of the updgradient edge of the recharge area, z_0 is the elevation of the water table, z_h is the elevation of the top the well screen, z_l is the elevation of the bottom of the well screen (**Figure 1a**), l_{theo}, w_{theo} , A_{theo} are the theoretical length, width, and area of the recharge zone, and:

$$\tan \omega = R / (K_h * i) \quad (7)$$

where R is the uniform recharge rate, K_h is aquifer hydraulic conductivity, and i is the regional hydraulic gradient. Equations (2) – (7) provide a simple analytical model to determine the capture zone of a domestic well in an unconfined aquifer with uniform flow, recharge, and deep production.

To account for the influence of domestic well pumping on the local groundwater flow system around the well and to account specifically for the influence of the gravel pack on the recharge area (**Figure 1b**), we constructed a numerical model, described in the next section.

Modeling Methods

The capture zone of a domestic well with a gravel pack is computed for a fully three-dimensional steady-state groundwater flow field. The steady-state head and flux distribution are computed using the MODFLOW groundwater flow model (McDonald and Harbaugh 1988). The capture zone corresponding to a particular groundwater flow solution is delineated using the backward particle tracking model MODPATH (Pollock 1994).

Briefly, MODFLOW solves the steady-state groundwater flow equation

$$\nabla \underline{\underline{K}} \nabla h = 0 \qquad (8)$$

where h is the hydraulic head, by using a fully three-dimensional block-centered finite difference scheme for the user-specified boundary conditions, $\boldsymbol{K}$ is the hydraulic conductivity tensor. In the following simulations pumping induces only a small drawdown of the piezometric surface, so the linear flow model (8) is sufficiently accurate for our purposes. We effectively invoke the Dupuit assumption equivalent to the MODFLOW "unconfined layer" algorithm. There, the unconfined layer thicknesses are set constant and only updated iteratively. From the hydraulic head solution, MODFLOW also computes the flux, q, across each of the six faces of each finite difference cell in the modeling domain. The flux solution becomes input to MODPATH, which computes backward particle travel paths given the linear groundwater velocity, $v = q/n$, where n is the effective porosity, across each finite difference cell face. Starting locations for backward particle paths are user-defined. MODPATH uses a semi-analytical linear interpolation scheme to compute a spatially continuous particle path (Pollock 1994).

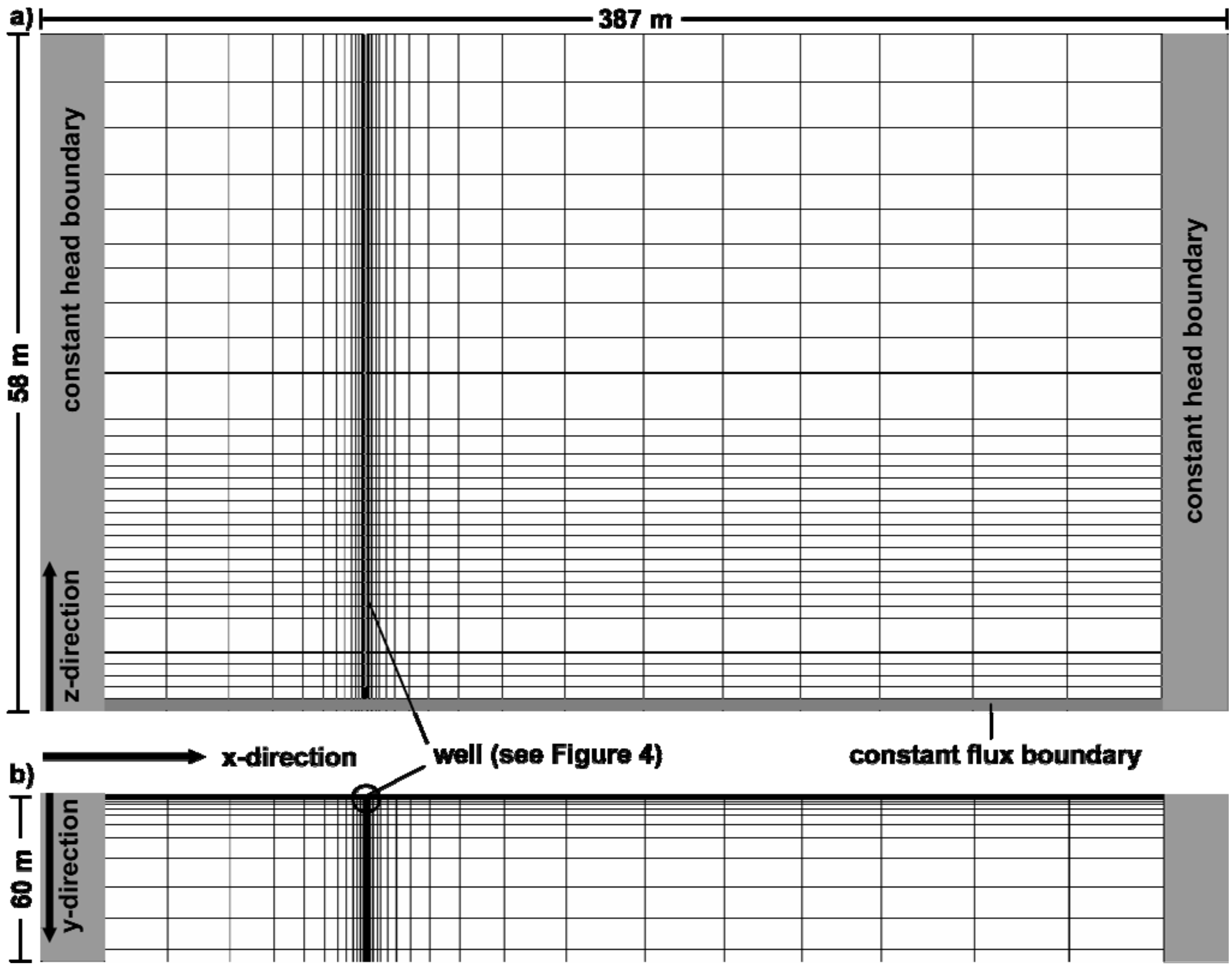

Figure 2: Model grid in (a) cross-sectional view at $y = 0$ (Vertical exaggeration = 4.2x) and (b) in plan view. Due to the symmetry of the flow field, the model domain simulates only half of a well and half of the capture zone. The well and gravel pack are very finely discretized. A close-up view of the model around the well screen is shown in Figure 4.

Our modeling domain is a finite difference grid with 141,750 cells of which 137,937 are active. The modeling domain is 58 m high, 387.23 m long and 59.695 m wide and consists of 45 rows, 90 columns, and 35 layers. The modeling domain takes advantage of the symmetry in the well flow field, which is symmetric across the x-axis (mean flow direction) centered on the domestic well (y = 0, see below). The model is therefore designed to model only one-half of the well capture zone (**Figure 2**). The second half of the well-capture zone mirrors the first half. Grid spacing is nonuniform in both the vertical and horizontal direction. Vertical grid spacing varies from 1 m at the elevation of the well screen to 4 m elsewhere (**Figures 2, 3**). Horizontal grid-spacing varies from 0.01 m near the well and in the gravel pack to nearly 20 m near the model boundaries. The horizontal increase in cell-size between adjacent rows or columns of the finite-difference grid is set to not exceed 50 % of its width.

The hydraulic gradient along the x-axis is produced by defining a constant head boundary of 58.00 m to the exterior block of cells at the upgradient vertical side of the model and a constant head of 57.61 m at the downgradient vertical side of the model (**Figure 2**). This is equivalent to a hydraulic gradient of 0.0018, which is typical for the study area. The other two vertical planes of the model are assigned no-flow boundary condition: the vertical plane adjacent to the well half is a symmetry plane. The vertical plane opposite of the half well is at sufficient distance to the well that the local effect of pumping on the groundwater flow field can be neglected and flow is parallel to regional groundwater flow. The average (steady-state) recharge rate is set to 0.669 m/year, a value typical for semi-arid, irrigated agricultural regions such as the Modesto Area, San Joaquin Valley, California. The bottom of the model domain is considered permeable and open to the regional aquifer system below. It is assigned a uniform constant (downward) flux boundary condition, with total outflow across the bottom boundary set equal to the difference between the total recharge inflow at the top and the well outflow rate. In this way we implicitly enhance our model to greater aquifer depths. In the Modesto Area large irrigation wells up to a depth of almost 370 m below land surface pump large amounts of water and produce a vertical flow component, even through a confining clay unit above the irrigation wells (Burow et al. 2004).

The well construction was chosen to be representative of domestic well construction in the San Joaquin Valley, California (e.g., Burow et al., 2004). The model well has a total depth of 56 m below the water table. A seal to 18 m below the water table overlies a 30 m long gravel pack around a blank well casing. The casing has a diameter of 0.2 m. The perforated well screen is located at 48 m to 55 m below the water table, followed by a conceptual well sump from 55 m – 56 m. Casing and screen are surrounded by a 0.09 m thick gravel pack. The total borehole diameter is 0.38 m. Due to the relatively low pumping rate, the well-loss and skin effect are assumed to be negligible. Inflow along the screen is computed by the model and non-uniformly distributed.

The grouted well seal above the gravel pack and the well casing are modeled as "no-flow" cells (black cells in **Figure 3**). The pump is simulated by 74 "well" cells inside the casing. They are located significantly above the top of the screen, opposite of the well seal bottom, which creates an upward flow inside the screen and casing. The MODFLOW "well" package is used to simulate the pump cells (light-grey cells in **Figure 3**). The total pumping rate of the domestic well is 3.5 m^3/d, half of which is uniformly distributed across the individual "well" cells at the top of the casing. Flow inside the model well casing was modeled by approximating the flow with eq. (8) using very high hydraulic conductivity. The gravel pack (grey cells in **Figure 3)** is modeled by choosing a separate hydraulic conductivity that is higher than that of the surrounding aquifer and ranges between 50 and 1000 [m/d] (**Table 1**). Modeling the pump inside the well allows the model to properly distribute the flow across the well screen, with screen inflow highest near the top of the screen and lowest at the bottom of the screen.

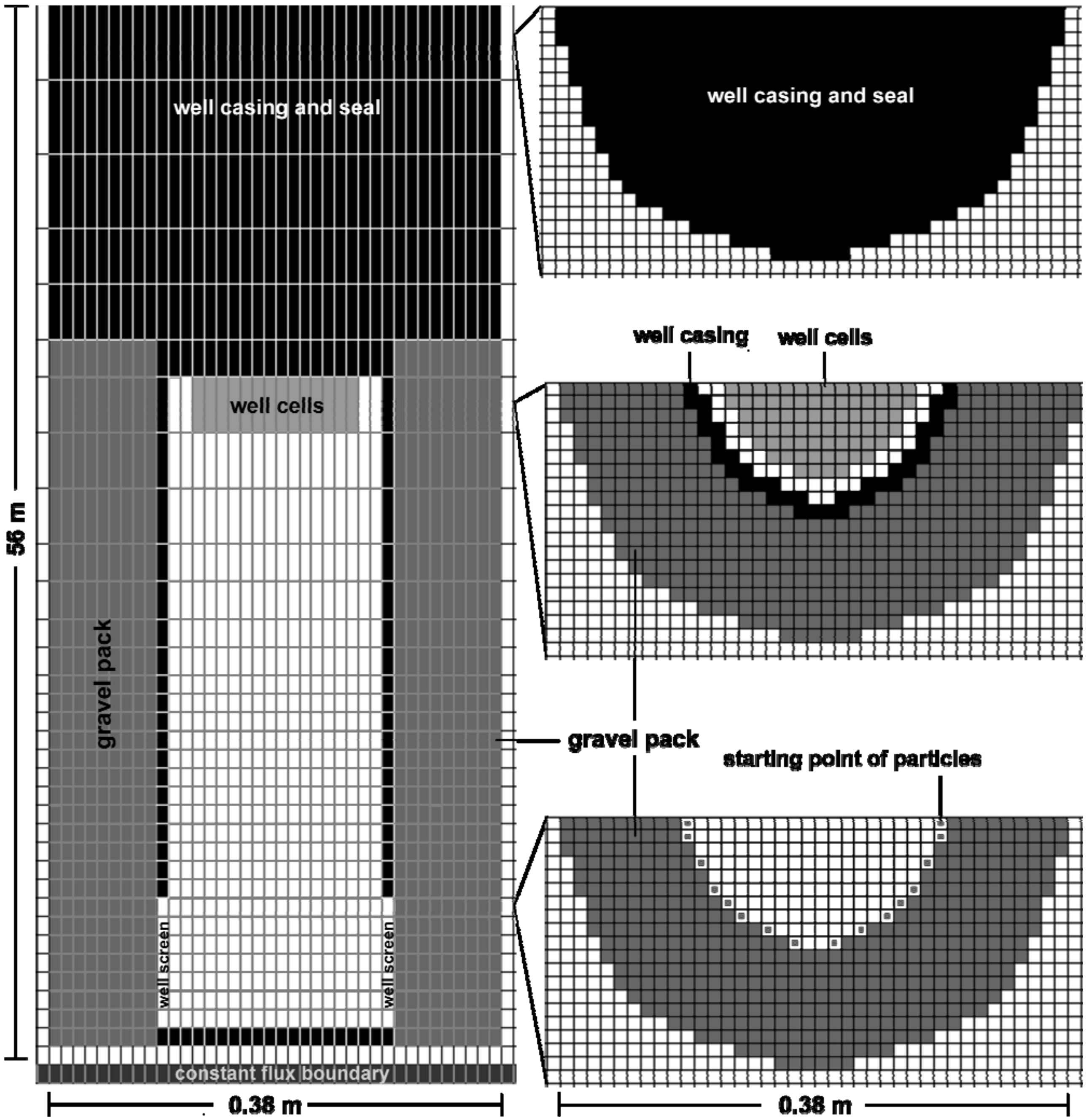

Figure 3: Model well configuration and grid discretization around the well. Left: Cross-section at the model boundary ($y = 0$). Right: Plan view at the land surface (right top), at the top layer of the casing containing the well cells (right center), and at the screen elevation (right bottom). Black cells: casing and well seal (impermeable). Grey cells: gravel pack. Dark grey cells: constant flux boundary cells at the model bottom. Grey dots in the lower left panel indicate the starting location for backward particle tracking.

The hydraulic conductivity, K_h, is assumed to be isotropic in the horizontal plane, while the vertical aquifer hydraulic conductivity, K_v, is lower, as typically observed in alluvial aquifers (e.g., Phillips et al., 2007). Two representative anisotropy ratios, K_h/K_v = 5 and 2, were chosen to bracket a representative range typically found in alluvial aquifers (*ibid.*). The gravel pack itself is assumed to have a completely isotropic hydraulic conductivity, K_g, that is larger than K_h. For illustration and application purposes, we modeled well capture zones for a wide range of representative values for the horizontal hydraulic conductivity, K_h, and the gravel pack hydraulic conductivity, K_g, and for two anisotropy ratios (**Table 1**).

K_h	K_v	K_g
1	0.2	50, 125, 250, 500, 750, 1000
1	0.5	50, 125, 250, 500, 750, 1000
3	0.6	50, 125, 250, 500, 750, 1000
3	1.5	50, 125, 250, 500, 750, 1000
5	1	50, 125, 250, 500, 750, 1000
5	2.5	50, 125, 250, 500, 750, 1000
10	2	50, 125, 250, 500, 750, 1000
10	5	50, 125, 250, 500, 750, 1000
30	6	50, 125, 250, 500, 750, 1000
30	15	50, 125, 250, 500, 750, 1000
100	20	125, 250, 500, 750, 1000
100	50	125, 250, 500, 750, 1000
300	60	500, 750, 1000
300	150	500, 750, 1000

Table 1: Model configurations with various combinations of the horizontal hydraulic conductivity, K_h, the vertical hydraulic conductivity, K_v, and the gravel pack hydraulic conductivity, K_g. All values are in units of [m/d].

Results

Head contour configurations in the aquifer around the domestic well are highly dependent on the aquifer and gravel pack hydraulic conductivities. Cross-sectional head contour lines along the regional flowpath are vertical under strictly regional flow with no recharge and no pumping. As expected from the analytical model above, the modeled contours deviate from the vertical due to the vertical flow component imposed by the recharge at the top of the model area and the regional pumping below the modeled zone. Contour lines increasingly deviate from the vertical alignment with smaller and smaller ratios of K_h / R (**Figure 4**). In addition, in aquifers with relatively low hydraulic conductivity, the domestic well creates a distinct zone of local influence in the aquifer around the well screen, whereas the influence is minimal in the highly permeable aquifer. The anisotropy of the aquifer hydraulic conductivity creates significant flow zonation: much of the impact of domestic well pumping on the pressure field is seen at the elevation of the well screens, especially for those cases with the higher aquifer anisotropy. Another distinct horizontal zone is created by the top of the gravel pack. The higher the gravel pack hydraulic conductivity (relative to K_h), and the higher the aquifer anisotropy ratio, K_h/K_v, the more pronounced is the effect that the transition between the top of the gravel pack and the annular seal has on the head contour lines (e.g., **Figure 4**). Inflow to the well varies non-uniformly along the screen. It is highest near the top of the screen, which is nearest to the pump intake inside the well-casing. The difference between the screen inflow at the top (layer 26) and the screen inflow near the bottom (usually in layer 31 just above the bottom of the layer) varies from approximately 45% for highly permeable aquifers to more than 100% for very low permeable aquifers with very high gravel pack K_g. This is consistent with analytical models (Nahrgang, 1954; Garg and Lal, 1971) and with field observations on large production wells (VonHofe and Helweg, 1998).

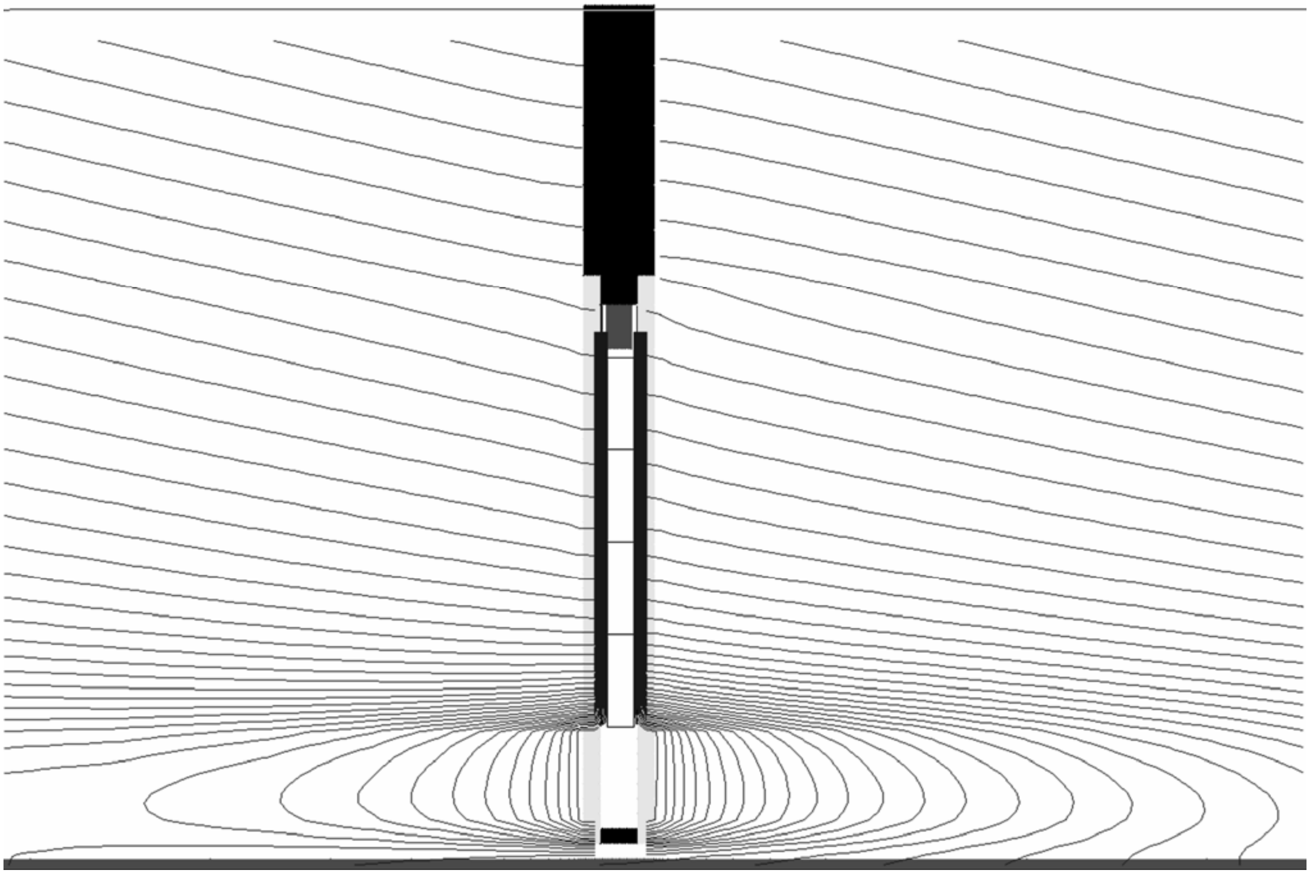

Figure 4: Head contour lines around the well for a conductivity of 10 m/d, an anisotropy ratio of 2, and hydraulic gravel pack conductivity of 750 m/d. The heads depend on the conductivity of the aquifer, the anisotropy and the relative difference in the conductivities between gravel pack and aquifer. Horizontal dimension is 7 m, vertical dimension is 58 m. Due to the horizontal exaggeration (12.4x) the inclination of the head contours in the regional flow field (near top of the cross-section appears nearly horizontal although it is actually nearly vertical.

Corresponding to the head field, pathlines in low hydraulic conductivity aquifers are significantly steeper and the capture zone is much closer to the well-head than in an aquifer with high hydraulic conductivity (**Figure 5**). For $K_h \geq 10$ m/day, the modeled pathlines are in fact sufficiently flat that the source area is outside the model area. In those simulations, we computed the pathlines outside the numerical modeling area by analytically calculating the extension of the pathlines to the water table using equation (7). Also, for model scenarios with hydraulic conductivities of 1, 3, and 5 m/d, the pathlines in the top aquifer layer were computed from eq. (7), because MODPATH computations in the top layer were subject to numerical error.

The source area of the domestic well has a distinct shape composed of two features: the main capture zone, a relatively large and wide oval area, corresponding to pathlines that enter the annulus of the well below the top of the well-screen for horizontal delivery into the well. At the downgradient (well-facing) side of this main capture zone, we observe a narrow elongated capture subzone that represents those pathlines that enter the gravel-pack of the well at some distance above the well screen. These pathlines capture domestic water through the high permeability field of the gravel pack above the well screen (**Figure 1b, Figure 5**). The greater the hydraulic conductivity difference between gravel pack and aquifer, the higher is the relative downward flow in the upper gravel pack, and the more MODPATH virtual water particles enter the well flowing through the upper gravel pack. Moreover, the steeper the particle path gradient, the higher is the highest point of entry into the gravel pack of pathlines that ultimately will be captured by the well. Thus, the gravel pack, where it extends to elevations much higher than the well-screen, significantly extends the length of the source area towards the well, albeit within a very narrow transverse range (**Figure 5**).

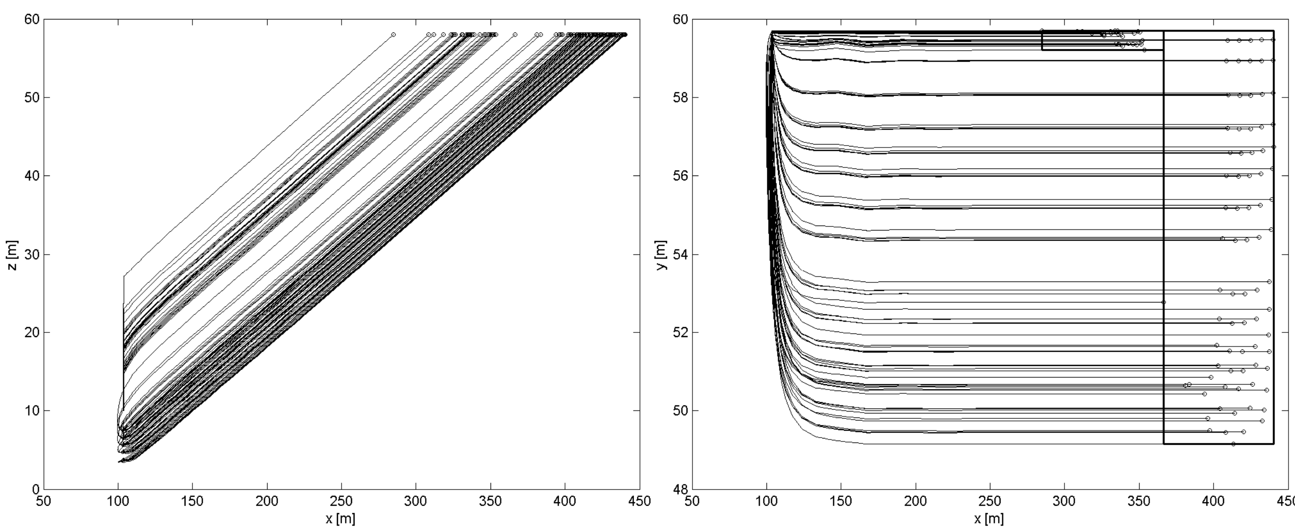

Figure 5: Pathlines in cross section (left) and plan view (half the well, right) with the elongated and main capture zone parts for an aquifer conductivity of 10 m/d, an anisotropy ratio of 2, and gravel pack conductivity of 750 m/d. Corresponding heads are shown in Figure 4.

For further analysis of the capture zone location and size, we separately refer to the width and length of the narrow, "elongated" part of the capture zone nearer to the well and of the "main" part of the capture zone (**Figure 6**) from where the majority of the water originates. The simulations show that the length of the elongated part increases faster than the length of the main part as horizontal aquifer conductivity increases, but the gravel pack conductivity has a significant influence only on the length of the elongated part (**Figure 6c, d**). The same is true for the width of the two capture zone parts: The gravel pack conductivity has a significant influence only on the width of the elongated part but little, yet discernable influence on the main part. The width of the elongated part increases several-fold with gravel pack hydraulic conductivity, K_g, especially in less productive (low K) aquifers. By the same token, the widths of the main and elongated parts (**Figure 6a, b**) decrease with higher aquifer conductivities (more narrow, but longer source area). For low gravel pack conductivities, the width of the elongated part of the capture zone remains nearly constant, regardless of aquifer conductivity

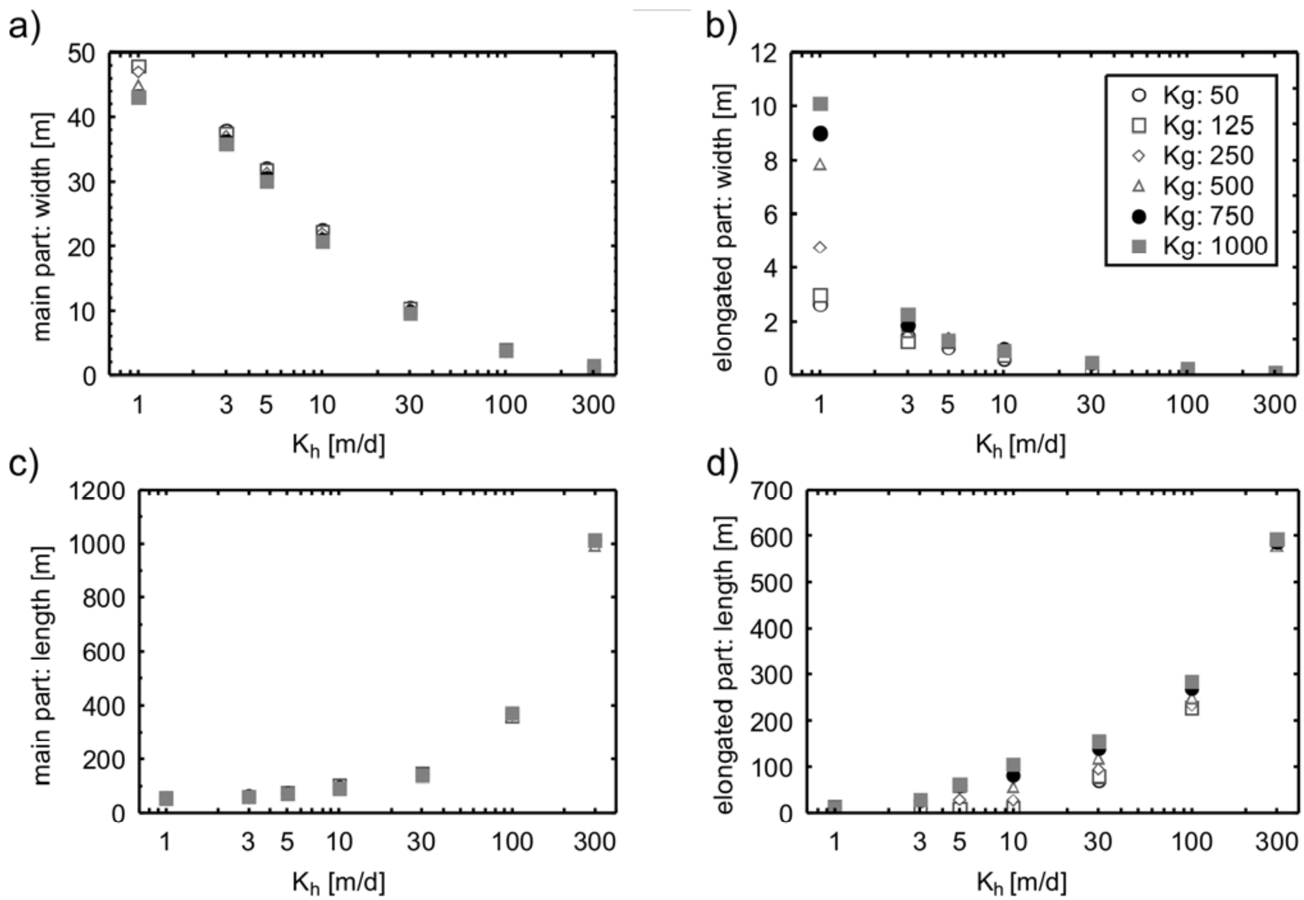

Figure 6: Widths (top panels) and lengths (bottom panels) of the elongated part (right panels) and the main part (left panels) of the capture zone for an anisotropy ratio of K_v: K_h = 1 : 2. Behavior of the models with an anisotropy ratio K_v : K_h = 1 : 5 is similar.

The analytical model (eqs. 2,3) of the source area location provides good approximations of the source area only in highly permeable aquifers. For aquifers with intermediate and low conductivity, the gravel pack has significant influence on the distance of the downgradient edge of the capture zone from the well (**Figure 7a**), where the source area can be as much as 90% closer to the well than estimated from eq. 2. The analytical approximation of the distance from the well to the upgradient edge of the source area (eq. 3) is relatively close to the numerical simulations if aquifer hydraulic conductivities are above 5 m/d. In those cases, the relative difference between analytical and numerical model is on the order of 10% or less (**Figure 7b**), regardless of anisotropy ratio and gravel pack hydraulic conductivity.

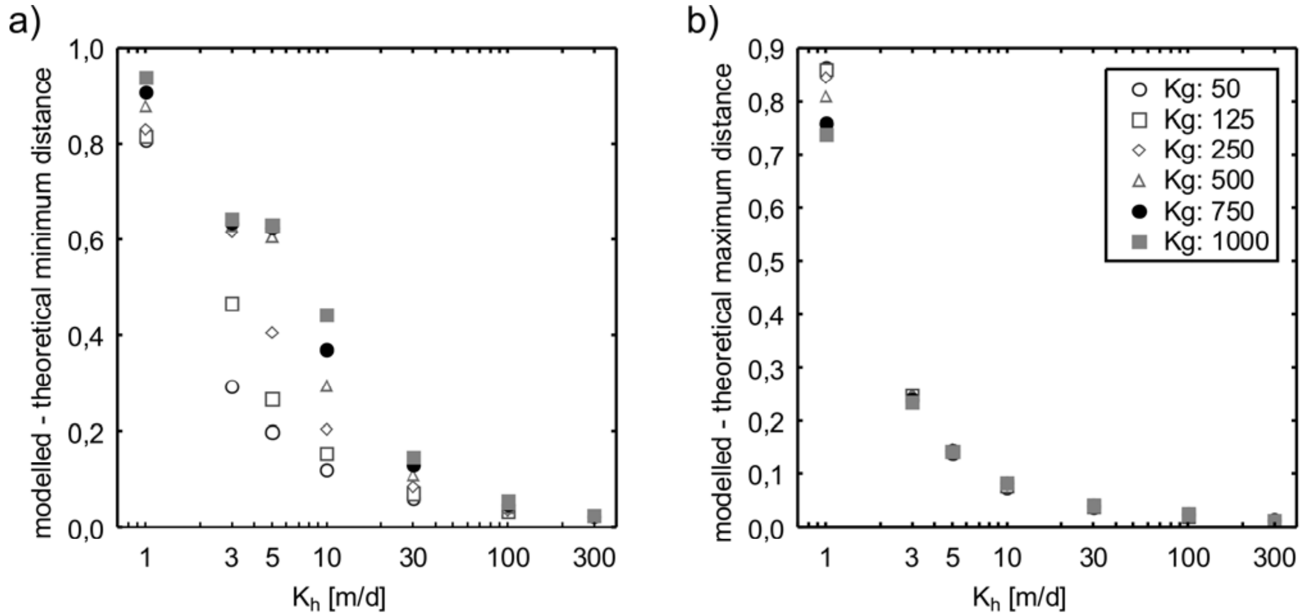

Figure 7: Comparison of the distances of the source areas to the well provided by the numerical and by the analytical model exemplarily for an anisotropy ratio of K_v: K_h = 1 : 2. (a) Normalized differences between the modeled and analytically calculated distances of the downgradient edges of the source areas to the well. (b) Normalized differences of the distances of the upgradient edges of the source areas to the well.

The simulation results show that water moves downward inside the gravel pack above the well-screen from considerable distances: For K_h less than 10 m/d and high gravel pack hydraulic conductivities, water travels downward from as far as the top of the gravel pack, 30 m above the well-screen (**Figure 8**). Again, the more permeable the gravel pack in the annulus, the larger the above-screen capture of source water. The fraction of well pumpage that originates from capture in the gravel pack above the well-screen increases as the aquifer hydraulic conductivity decreases (**Figure 9**). In intermediate and low permeable aquifers, domestic wells with highly permeable gravel packs receive from 20% to 50% of the total well flow from the extended gravel pack above the screened aquifer horizon. This model result is qualitatively consistent with the field data of Houben (2006), who found iron oxide incrustations in the gravel pack significantly above the top of the well screen, where the incrustations were due to a significant amount of water flowing through the upper part of the gravel pack. At high aquifer conductivities (K_h > 10 m/d), less than 12 % of the total domestic well flow originate from the gravel pack above the well screen. Aquifer anisotropy has little influence on the height of the capture zone within the gravel pack.

The height of the gravel pack participating in flow to the well and the percent fraction of the pumpage originating from the gravel pack above the screen can be expressed quantitatively: **Table 2** provides the regression coefficients obtained by fitting data in **Figures 8a and 8b** to nonlinear exponential regression equations of the form:

$$y = a*\exp(-\log(K_h)/b) \qquad (9)$$

using the Levenberg-Marquardt algorithm for optimization. For application to a specific site, linear interpolation of the values for *a* and *b* in **Table 2** may be used to compute the height of capture in the gravel pack and the proportion of flow originating from the gravel pack above the well screen for values of the anisotropy ratio and of K_g other than those given in the Table. This modified analytical tool provides a much more realistic source area than the much simpler graphical method employed in many states as part of their source water assessment programs (e.g., California DHS 1999).

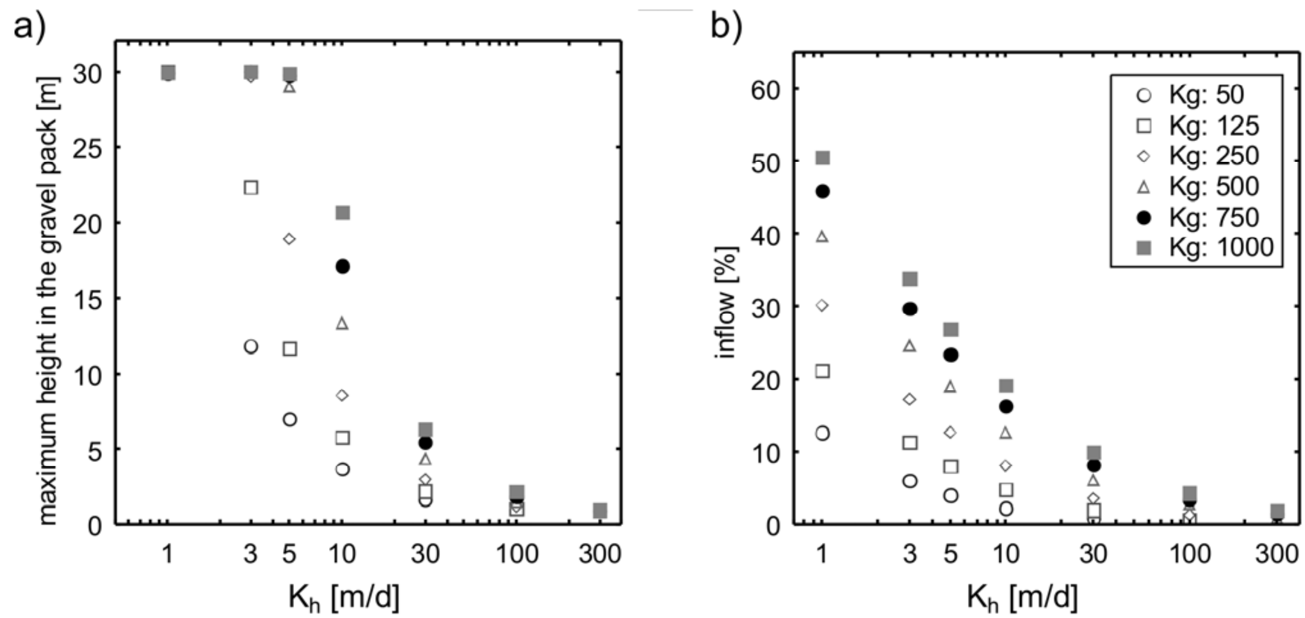

Figure 8: (a) Maximum virtual water particle heights in the gravel pack above the well screen serving to capture water (b) Percentage of inflow into the well screen flowing through the gravel pack from above the screen. Both for an anisotropy ratio of 2.

Anisotropy	K_g	Parameter for maximum heights:		Adjusted r^2	Parameter for inflow from above:		Adjusted r^2
		a	b		A	b	
2	50	33.31	1.05	1.00	12.69	0.60	0.99
2	125	77.01	0.88	0.99	21.44	0.69	0.99
2	250	87.95	1.02	1.00	30.59	0.76	0.99
2	500	163.35	0.93	1.00	40.57	0.86	0.99
2	750	172.04	1.00	1.00	47.16	0.92	0.99
2	1000	226.70	0.96	0.94	52.10	0.96	0.98
5	50	79.01	0.78	0.95	14.94	0.61	1.00
5	125	98.28	0.93	0.97	25.14	0.70	0.99
5	250	216.33	0.80	0.93	35.57	0.77	0.99
5	500	276.67	0.87	0.99	47.59	0.85	0.99
5	750	396.12	0.85	0.99	55.54	0.91	0.99
5	1000	401.91	0.88	1.00	61.39	0.96	0.98

Table 2: Coefficients and adjusted coefficients of determination (r^2) for the equations describing the maximum heights of the capture zone in the gravel pack, and the inflow of water entering the well from the gravel pack above the screen.

Discussion

For application to specific sites, Figure 7 provides a tool to estimate the additional source area due to the gravel pack, when compared to the simple approximation (eq. 2). These results can also be applied for conditions with smaller or larger recharge rates, *R'*, than the rate $R = 0.669$ m/a used in our computations. For *R'* not equal to *R*, results shown in Figures 7-10 and expressed in the above equation are looked up for a scaled hydraulic conductivity *K'* rather than for the actual hydraulic conductivity *K*, where $K' = K \cdot R'/R$. This scaling procedure is approximate because it does not simultaneously scale other parameters controlling the observed results, e.g., screen length and pumping rate. However, for applications in unconsolidated sedimentary aquifers, this scaling approach works well as the drawdown created by domestic wells is relatively small. For depths to the top of the screen different from that used here, the simple geometric conceptual model outlined in Figure 1 and expressed in eq. 2 provides a framework for adjusting the distance of the source area from the well head. Equation 9 (with Table 2) can be used to estimate the fraction of flow originating from the elongated part of the source area.

The numerical modeling shows the significant influence of the gravel pack on the source area of a domestic well, particularly for lower permeable aquifers (horizontal hydraulic conductivities of less than 10 m/d). In highly permeable aquifers (relative to the recharge rate of 0.669 m/year used in this study), the analytical model (eqs. 2, 3) provides a relatively good approximation of the upgradient and downgradient edge of the source area. Lower hydraulic conductivities lead to significantly longer capture zones than predicted by the analytical model (eqs. 2-3). In our configuration of screen length and gravel pack length, which represents an average domestic well construction for Central California, the elongation due to the presence of a gravel pack constitutes up to 70 % of the total length of the capture zone. The elongation is relatively narrow but higher gravel pack conductivities lead to significant increases in that width. The width of the main capture zone, in turn, slightly decreases at higher gravel pack conductivities. The greater the difference between hydraulic conductivity of the aquifer and that of the gravel pack, the greater is the elongated part relative to the total length of the capture zone.

For many contaminants, chemical or microbial, aquifer attenuation is a dynamic, time-dependent process. Travel times for potential contaminants decrease approximately linearly with increased gravel pack length above the well screen. This is due to the strong influence of recharge on vertical downward displacement of water (and contaminants) and the relatively small influence that the domestic well pumping exerts on the overall groundwater flow field. A linear decrease in travel time from the time of recharge until arrival at the gravel pack is associated with exponentially increased contaminant concentrations. The gravel pack itself typically provides much less attenuation capacity than the aquifer material. Hence, a short seal and vertically extended gravel pack constitute a potential short-circuit for contaminants.

We also note that the fraction of flow captured by the gravel pack above the well screen may be relatively small in a productive (high K) aquifer. But for some contaminants the resulting dilution with (good) groundwater collected by the well at the depth of the screen may not be sufficient. This includes contaminants that reach the water table at concentrations that are several orders of magnitude above regulatory drinking water limits including solvents, pesticides, other organic chemicals, and pathogens. A possibly common source of such contamination are septic tank leach fields, which - in rural and semi-rural housing developments - are often located in the vicinity of domestic wells.

Conclusions

Our work provides a tool to quickly estimate the size and location of the source area of domestic wells in regions with significant recharge (for example, due to irrigation). The influence of the gravel (or sand) pack in the well annulus above the well screen is explicitly accounted for. Results allow for estimation of source area and gravel pack impact for a wide range of scenarios. Importantly, we show that the gravel pack above the well screen poses a significantly increased risk for domestic well contamination. A gravel pack that extends significantly above the well screen (due to short seal length), may significantly enhance the length of the source area, thus exposing the well to a larger cross-section of potential contaminant sources. The extended gravel pack also decreases travel time and distance for contaminants from the source area to the well allowing for contaminants to partially circumvent natural aquifer attenuation. This is especially true in aquifers with low to intermediate hydraulic conductivity ($K \leq 10$ m/d). We therefore strongly recommend that the gravel (or sand) pack not be extended more than a few meters above the well screen of a domestic well.

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