Groundwater

Depletion and Capture: Revisiting "The Source of Water Derived from Wells"

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Abstract

A natural consequence of groundwater withdrawals is the removal of water from subsurface storage, but the overall rates and magnitude of groundwater depletion and capture relative to groundwater withdrawals (extraction or pumpage) have not previously been well characterized. This study assesses the partitioning of long-term cumulative withdrawal volumes into fractions derived from storage depletion and capture, where capture includes both increases in recharge and decreases in discharge. Numerical simulation of a hypothetical groundwater basin is used to further illustrate some of Theis' (1940) principles, particularly when capture is constrained by insufficient available water. Most prior studies of depletion and capture have assumed that capture is unconstrained through boundary conditions that yield linear responses. Examination of real systems indicates that capture and depletion fractions are highly variable in time and space. For a large sample of long-developed groundwater systems, the depletion fraction averages about 0.15 and the capture fraction averages about 0.85 based on cumulative volumes. Higher depletion fractions tend to occur in more arid regions, but the variation is high and the correlation coefficient between average annual precipitation and depletion fraction for individual systems is only 0.40. Because 85% of long-term pumpage is derived from capture in these real systems, capture must be recognized as a critical factor in assessing water budgets, groundwater storage depletion, and sustainability of groundwater development. Most capture translates into streamflow depletion, so it can detrimentally impact ecosystems.

Introduction

In a classic and often-cited paper, Theis (1940) explains the sources of water derived from a pumping well. Among other things, Theis (1940) concludes that "All water discharged by wells is balanced by a loss of water somewhere. This loss is always to some extent and in many cases largely from storage in the aquifer. Some groundwater is always mined." He then notes that "After sufficient time has elapsed ... further discharge by wells will be made up at least in part by an increase in the recharge if previously there has been rejected recharge. ... further discharge by wells will be made up in part by a diminution in the natural discharge." The combination of increased recharge and decreased discharge is termed "capture" (Lohman et al. 1972; Bredehoeft and Durbin 2009; Leake 2011).

These generic relations show that at early times the principal source of water to a well is from depletion of storage in the aquifer (Figure 1). With increasing time, the fraction of pumpage derived from storage depletion (a nondimensional "depletion fraction") tends to decrease, and the fraction derived from capture increases. Eventually, provided that sufficient potential increases in recharge and decreases in discharge are available, a new equilibrium will be achieved when no more water is derived from storage and heads or water levels in the aquifer stabilize. The actual response time for an aquifer system to reach a new equilibrium is a function of the dimensions, hydraulic properties, and boundary conditions for the specific case. The response time will change as these conditions are varied. For example, the response time will decrease as the hydraulic diffusivity of the aquifer increases (see Theis 1940; Barlow and Leake 2012). The response time can range from days to millennia (Bredehoeft and Durbin 2009; Walton 2011). An important corollary to Theis' (1940) principles is that the average predevelopment rate of natural recharge itself is largely irrelevant to storage depletion and capture responses (Bredehoeft et al. 1982; Bredehoeft 1997; Barlow and Leake 2012). However, the natural recharge does serve as a constraint on capture—in the sense that it controls the natural predevelopment groundwater discharge, which is subject to capture by pumping wells.

Capture includes several factors and processes, but is often considered synonymous with (or dominated by) streamflow depletion (e.g., Alley et al. 1999; Barlow and Leake 2012). This includes increased recharge through

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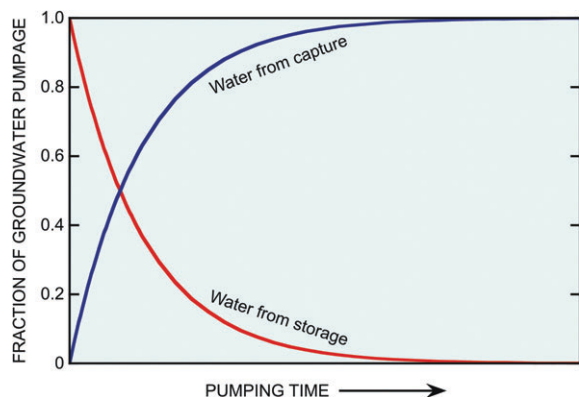
Figure 1. Sources of water to a well can change with time. Time scale for curves depends on the hydraulic characteristics of the aquifer and the distance of the well from recharge and discharge locations. Modified from Alley et al. (1999) and Barlow and Leake (2012).

induced infiltration from streams (and other surface water bodies), as well as decreases in groundwater discharge to springs, streams, and other surface water bodies (i.e., decreases in base flow). However, capture can also include (1) increased recharge facilitated by water-table declines in areas where potential recharge from precipitation under natural conditions is rejected and runs off the land surface because high water tables preclude infiltration, and (2) decreased evapotranspiration in areas where the water table is close to the land surface but declines due to pumpage-induced drawdown (Theis 1940, 1941; Bredehoeft et al. 1982; Walton and McLane 2013). If recharge were to increase or discharge were to decrease, either coincidentally or through intentional water management policies (e.g., artificial recharge, especially using imported water, or phreatophyte control), the effects of well pumpage would be additionally offset or balanced accordingly.

Theis (1940) notes that aquifers are bounded; Walton and McLane (2013) expand on this point and note that because of bounds, full capture of supply components may not be feasible. What are the consequences if sufficient capture is not available to meet the demands imposed by substantially increased pumpage? Then the response will be constrained and a new equilibrium may never be achieved (Bredehoeft and Durbin 2009; Barlow and Leake 2012). As explained by Theis (1940), if the amount of pumping in an area exceeds the amount available for capture, water levels will continue to decline and pumping therefore will continue to be derived from storage depletion. Pumping under these constraints is clearly unsustainable. However, most studies in the literature of the effects of pumping on storage depletion and streamflow capture have assumed the presence of at least some boundary conditions that would allow a new equilibrium to be achieved. For example, there is always some flow in a stream or river that bounds an aquifer, as assumed by Theis (1941).

Surface-water bounding conditions that place limits on capture potential are more likely to occur in arid

climates, but how common or large this constraint might be is uncertain. Lack of recognition of constraining boundary conditions might lead to erroneous estimates of storage depletion and sustainability. On the other hand, too much weight can be given to bounds on capture. For example, Wada et al. (2010) assume that groundwater storage depletion equals the excess of groundwater pumpage over natural recharge, except in humid climates. Pokhrel et al. (2012) estimate "unsustainable groundwater use" on the basis of estimated total water demand, and further assume that such groundwater use is equivalent to groundwater depletion. By essentially ignoring capture, they may substantially overestimate groundwater depletion (Konikow 2013a). Such analyses effectively assume that the storage depletion fraction of pumpage is 1.0 and the capture fraction is 0.0, and ignore the time dependency of the relations as shown in Figure 1.

The purpose of this study is to further characterize the partitioning of sources of well pumpage between capture and storage depletion. The study examines quantitatively the responses to pumpage under capture-constraining conditions, which have not been well elucidated in the literature, and compares responses under such conditions with those under unconstrained conditions. The study also examines storage depletion in real-world aquifer systems on the basis of long-term records and assessments in specific aquifer systems, in part to document the long-term depletion fractions in large-scale systems developed for long periods of time, and in part to assess the reasonableness of assumptions that estimate storage depletion on the basis of pumpage while ignoring capture.

Capture–Constrained Case

Aquifers are often bounded by surface-water features, such as streams or lakes. If such features have a limited availability of surface water, it is hypothesized that the balancing of well withdrawals by increases in recharge and/or decreases in discharge to or from that bounding feature would be constrained, and the general balance between storage depletion and capture (Figure 1) would be disrupted. This might occur, for example, if the stream or lake goes dry. If growth of capture is limited, then the abatement of storage depletion with time is thereby also diminished and storage continues to provide water to the well. In this case, the relative fraction of pumpage balanced by storage depletion is greater than would otherwise occur. If pumpage is so large that capture can never balance the withdrawals, then a new equilibrium cannot be attained, water levels would continue to decline, and the system will continue to be depleted until well yields are necessarily reduced or eliminated (Bredehoeft and Durbin 2009).

To test and evaluate this hypothesis in a quantitative framework, a hypothetical desert-basin aquifer was simulated. The model was based closely on the hypothetical desert-basin aquifer, as developed and documented by Barlow and Leake (2012), which includes a through-flowing river along the eastern edge of the basin. For
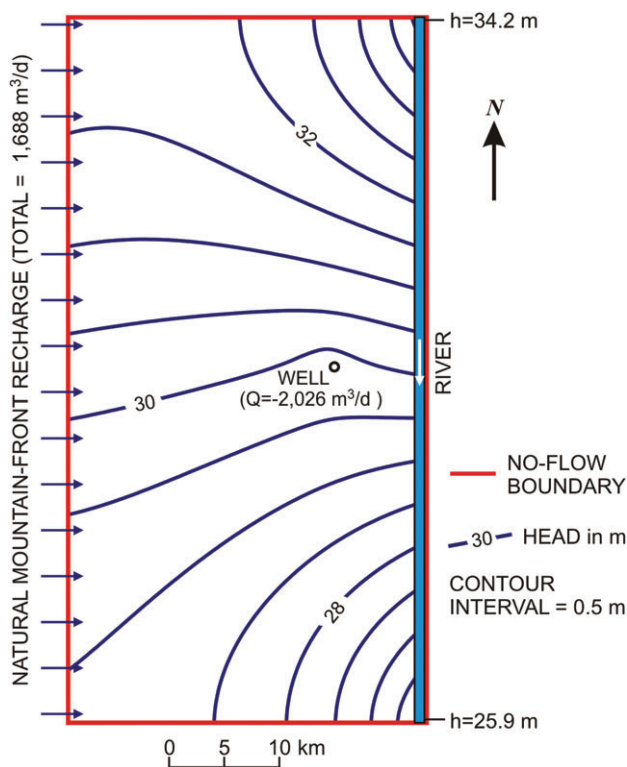
Figure 2. Map view of hypothetical desert-basin aquifer with a through-flowing river along the eastern edge of the basin, showing boundary conditions and calculated heads after 200 years of pumping from a single well. Modified from Barlow and Leake (2012).

Table 1
Characteristics of Hypothetical Aquifer System and Model

Property	Value
Basin dimensions	32.2×64.4 km
K_h	15.24 m/d
Aquifer thickness	157.4 m
Specific yield	0.2
Natural mountain-front recharge	1688 m^3/d
River width	10.0 m
River depth	0.001 m
Streambed K_z	3.05 m/d
Streambed thickness	0.305 m
Grid spacing	804.7 m
Number of rows	80
Number of columns	40

single-layer grid of 80 rows and 40 columns yielding square cells with a length of about 805 m on each side. The base case for development includes one pumping well located approximately 8.05 km west of the river and halfway between the northern and southern impermeable boundaries of the rectangular basin. The assumed pumping rate for the well (2026 m^3/d) represents the actual pumpage (and consumptive use), further assuming that none of the pumped water subsequently recharges the aquifer (e.g., see Bredehoeft 2011a). (If some of the pumped water subsequently recharged the aquifer, the net effect on the hydraulic responses in the aquifer would be the same as if the well discharge were reduced by the amount of return flow.) The river flows southward, and the streambed elevation varies linearly downstream from an elevation of 34.2 m to 25.9 m at the two ends of the stream reach. For a base case simulation, it is assumed that the flow rate entering the upstream end of the river is 20,000 m^3/d.

The aquifer system is simulated numerically using MODFLOW-2005 (Harbaugh 2005). The river is represented using the Streamflow Routing (SFR) Package, with a specified depth of water in the stream (Niswonger and Prudic 2005). The model computes a fluid flux between a stream and underlying aquifer at each relevant node of the grid based on head gradients between the stream and aquifer, and routes streamflow downstream after adjusting for the computed aquifer flux at a given location. It further assumes that as long as there is flow in the river, it remains connected to the aquifer (as opposed to disconnected, as described by Brunner et al. 2011). If the stream goes dry at a particular location, then no flow can be routed downstream and the boundary condition is automatically adjusted to preclude a stream-aquifer flux where a stream cell is dry. If the groundwater head at a downstream location is higher than the streambed elevation, then groundwater discharge will restart flow in the stream. Computational methods and assumptions are described in detail by Prudic et al. (2004) and Niswonger and Prudic (2005). A base-case simulation was run starting with an

simplification, it is assumed that the river lies on the easternmost edge of the alluvial aquifer, and that impermeable bedrock exists beyond the location of the river. Relative to the model developed by Barlow and Leake (2012), all length units were converted to the metric system, and a finer grid spacing was imposed in the *y*-direction. This is a two-dimensional model with no areally diffuse recharge from precipitation—noting that the natural predevelopment recharge rate would not affect the total streamflow depletion (e.g., see Bredehoeft et al. 1982; Barlow and Leake 2012). The only sources of recharge to this hypothetical aquifer are from natural mountain-front recharge at a fixed rate along the western boundary of the model and from head-dependent leakage from a river along the east side of the basin (Figure 2). Mountain-front recharge is a common and important phenomena in arid alluvial basins and typically represented as a boundary condition in groundwater models of a basin (Wilson and Guan 2004). As such, the specified recharge cannot be affected or directly captured by wells pumping from a basin aquifer. However, this type of recharge creates a downgradient discharge from the system that indeed can be captured. Also for simplification, it is assumed that there are no evapotranspirative losses from the water table that could potentially be captured.

Properties and characteristics of the aquifer and the model are listed in Table 1. The aquifer is approximately 32.2 km wide and 64.4 km long. It is discretized into a

Table 2
Hydrologic Budget for Base-Case Simulation (Flux Values Are in m³/d)

	Steady-State Predevelopment	50 Years	200 Years
Mountain-front recharge	1688	1688	1688
Recharge from river infiltration	5785	6585	6859
Groundwater discharge to river	7473	6855	6649
Well pumpage	0	2026	2026
Change in storage	0	608	128

initial steady-state simulation to represent predevelopment conditions. Then a 200-year transient simulation was run using annual time steps and a constant rate of pumping from a single well (location and pumping rate are shown in Figure 2). The resulting hydrologic budgets at three reference times are shown in Table 2. The flow field at the end of the simulation, as depicted by the head distribution (also shown in Figure 2), shows the effect of the pumping well, recharge and flow from the upper reaches of the river into the aquifer, lesser recharge from the western boundary of the model, and groundwater discharge to the lower reaches of the river.

After 47 years, most of the total cumulative pumpage is derived from capture and the amount derived from storage depletion has nearly stabilized by the end of the 200-year simulation period (Figure 3A). While the pumping rate remains constant in time, the rate of capture increases exponentially and the rate of storage depletion decreases exponentially (Figure 3B). The components (sources) of capture include increases in recharge (arising from increased stream leakage into the aquifer induced by declining groundwater levels) and decreases in groundwater discharge (base flow to the river) relative to predevelopment conditions. At early times, the latter is somewhat larger, but the two sources of capture stabilize in a few decades to where increases in recharge contribute about 56% of capture and decreases in discharge account for 44% of capture. For the conditions of this simulation, both factors combined (i.e., total capture) result in (and are equivalent to) streamflow depletion.

In terms of sources of water derived from the well, capture increases exponentially while storage depletion decreases exponentially. Although the generic relations shown in Figure 1 offer no specific time scale, in this test case the relative contributions still had not stabilized after 200 years (Figure 4). Storage depletion and capture can be compared to pumping on the basis of either cumulative volumes or instantaneous rates. Because the storage depletion rates decrease with time, the depletion curve based on cumulative volumes, which integrate system responses over time, will reflect higher fractional values at any particular time during the transient evolution of the response than that based on instantaneous rates. Conversely, capture (or streamflow depletion) fractions based on cumulative volumes will be smaller at any particular time than those based on instantaneous rates (also see Barlow and Leake 2012, 16 to 17). On the basis of cumulative volumes, in the base case simulation the results were depletion dominated for the first 47 years and then were capture dominated after that (Figure 4). When the fractions are computed on the basis of flow rates (for annual time steps in this case), the cross-over occurs earlier—after only about 17 years. A large difference is also present for the two calculations at any given time. For example, after 100 years, the depletion fraction based on cumulative fluxes was about 36% whereas the depletion fraction based on flow rates was about 18%.

The timing and relative magnitude of the response of the stream-aquifer system depends on the hydraulic properties of the aquifer, its boundary conditions, and the distance of the well from the recharge and discharge boundaries (Theis 1940). The sensitivity of the response in the hypothetical desert-basin aquifer to well location was evaluated by varying the well position in an east-west direction between the two lateral boundaries. The results (Figure 5) show that the time it takes for the system to reach a new equilibrium condition increases with distance of the well to the river (it was assumed that steady-state conditions are attained when 99.9% of the ultimate storage depletion has occurred). The storage depletion fraction was even more sensitive to well location and varied from 0.01 to 0.18 over the range of tested distances. The total storage depletion volume at steady state also increased by a factor of almost 20 (from 4.5×10^6 to 8.8×10^7 m³) as the distance to the river increased from 0.805 km to 30.6 km.

In the base case, the river never goes dry during the 200-year simulation, so the potential for increasing recharge in response to drawdown in the aquifer is never limited. To evaluate the affects of constraints on increases in recharge, the specified inflow to the upstream end of the river was reduced to 10,000 and 6400 m³/d. The downstream flow profiles for the three different specified stream inflows show that the river goes dry only for the lowest inflow case (Figure 6). The changes in flow over space and time are identical when $Q_{in} = 20,000$ or 10,000 m³/d. However, when Q_{in} is reduced to 6400 m³/d, the river starts to go dry during the 22nd year of the simulation when the increasing stream leakage into the aquifer equals the total flow in the river. As time progresses, the dry reach advances further upstream, as indicated by the difference between the 50-year and 200-year curves for the case of $Q_{in} = 6400$ m³/d. The differences also show that the change in flow during the first 50 years was much greater than the change during the next 150 years. Streamflow increases in the downstream part of the river because of groundwater discharge into the river. The difference between the predevelopment profiles and the curves at a given time after pumping began represents capture (and streamflow depletion).

The calculated flow rates for the streamflow-limited case (Figure 7) can be compared to the same elements in the base case (Figure 3B). The results show that
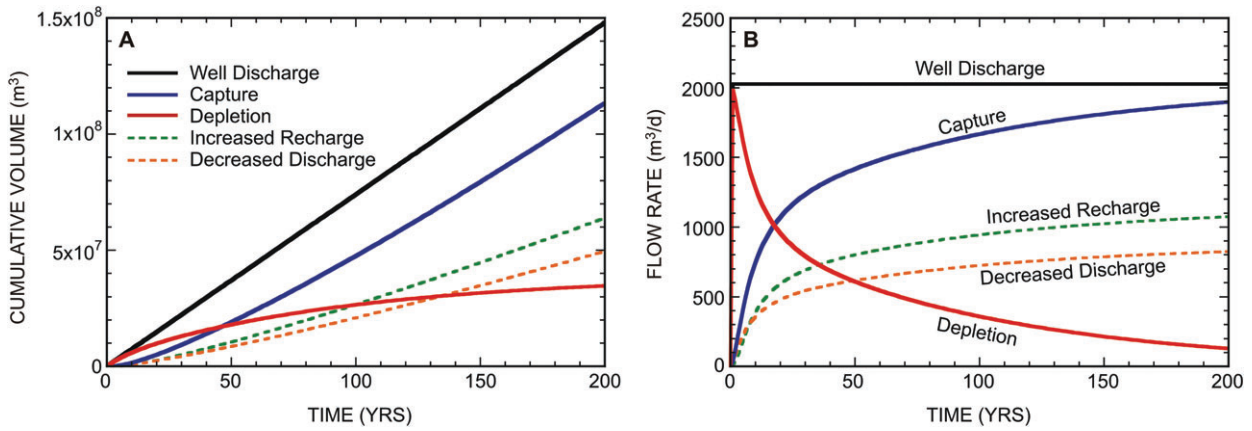
Figure 3. Simulated hydrologic budget elements for base case simulations of hypothetical aquifer expressed as cumulative volumes (A) and instantaneous flow rates (B).
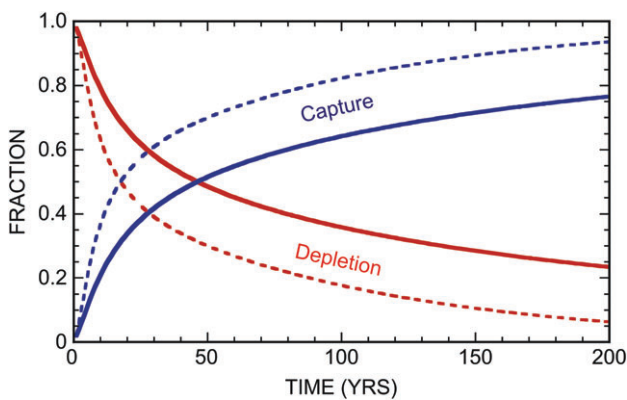

Figure 4. Simulated storage depletion (red curves) and capture (blue) fractions relative to pumping on both a cumulative (solid lines) and annual rate (dashed lines) basis.
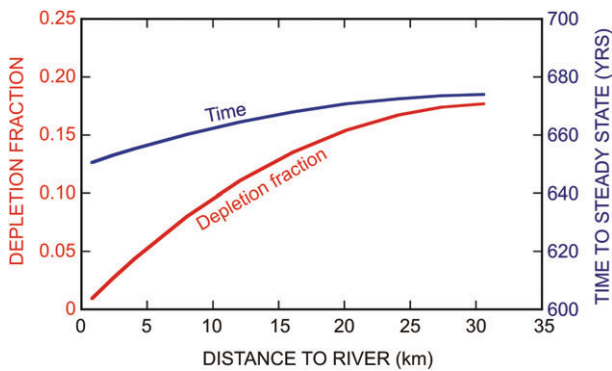

Figure 5. Sensitivity of time to achieve equilibrium, and of storage depletion fraction at equilibrium, to distance of pumping well to river. Depletion fraction is based on cumulative volumes at the time that a new steady-state condition is achieved.
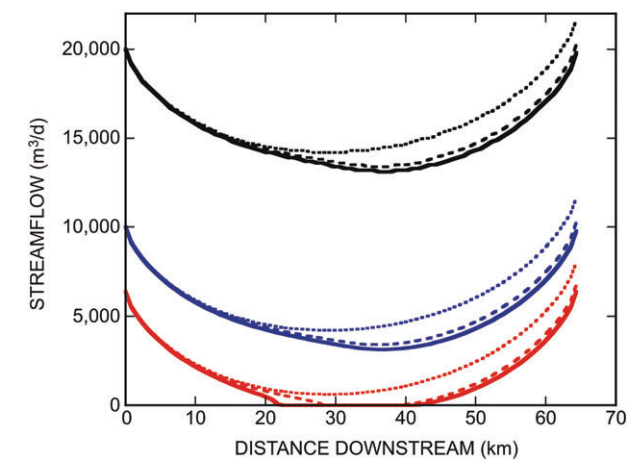

Figure 6. Change in flow along the length of the river when the specified upstream inflow is 20,000 m³/d (black curves), 10,000 m³/d (blue curves), and 6400 m³/d (red curves) for $t = 50$ years (dashed lines) and 200 years (solid lines). Streamflow profiles for steady-state predevelopment conditions are shown for comparison (dotted lines).

from stream leakage (induced infiltration), which reaches its maximum in year 22 and becomes constant after that because 100% of the upstream inflow to the river has been captured. Simultaneously, there is an increase (relative to the base case) in the amount of groundwater discharge to the downstream reaches of the river that is captured, though not enough to offset the constrained increase in recharge. Thus, after 22 years, the total capture is reduced and storage depletion is increased relative to the base case. The decreased capture relative to the base case is also reflected in a plot of storage depletion and capture fractions (Figure 8). After 200 years, the annual capture fraction is 0.84 for the streamflow-limited case, whereas it is 0.94 in the base case.

In the previous analysis, the total well pumpage is less than the total available capture and the rate of capture is still increasing after 200 years (Figure 7). But this should change if the total well pumpage were greater than the available capture. This was tested by adding nine
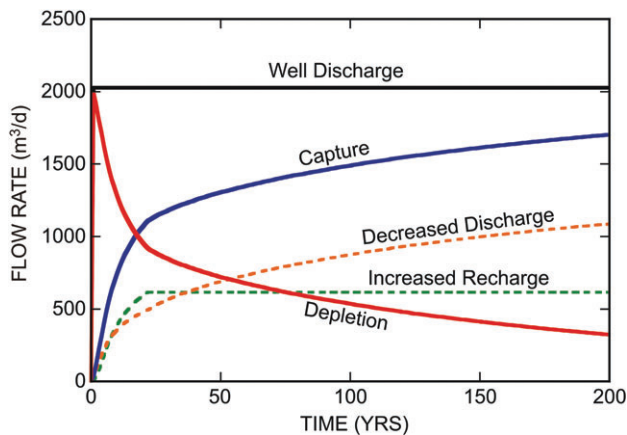
when there is insufficient water in the river to meet the drawdown-induced demand, the amount of pumpage derived from capture decreases and the amount derived from storage depletion increases. The effect is most noticeable on the increase in rate of recharge derived

Figure 7. Simulated hydrologic budgets for streamflow-limited case ($Q_{in} = 6400 \, \text{m}^3/\text{d}$) showing calculated flow rates for selected boundary conditions.
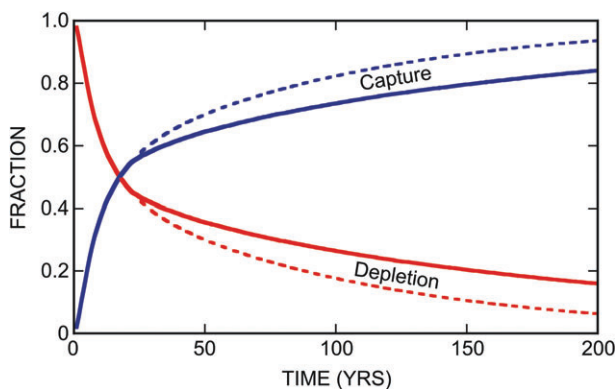

Figure 8. Simulated annual storage depletion (red curves) and capture (blue) fractions relative to pumping for the streamflow-limited case (solid lines) and the high streamflow base case (dashed lines).
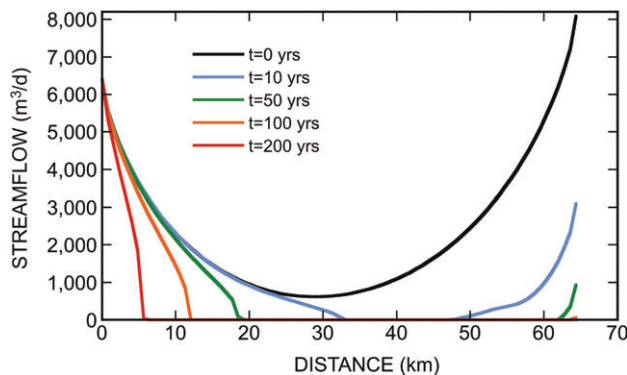

Figure 9. Change in flow along the length of the river for the streamflow-limited, high-pumpage case for selected times. Specified upstream inflow is $6400 \, \text{m}^3/\text{d}$. Total well pumpage is $20,260 \, \text{m}^3/\text{d}$.

more wells pumping at the same rate to the downstream (southern) half of the aquifer, so that the total well withdrawals are ten time greater than in the previous case ($Q_{tot} = -20,260 \, \text{m}^3/\text{d}$). With the increased pumpage, the stream first goes dry in the 6th year of the simulation, and captures all of the groundwater discharge in the 104th year (when the river outflow from the basin becomes zero). This is reflected in the changing downstream flow profiles at various times (Figure 9), which also illustrates the progressively longer length of the dry reach with time.

In terms of the hydrologic budget for the system under the higher pumping scenario, the rates of storage depletion and capture (streamflow depletion) become steady after 104 years (Figure 10A). Similarly, the fractions of annual pumpage derived from storage depletion (0.60) and capture (0.40) do not change after this time either (Figure 10B). This means that the cones of depression around the pumping wells will not stabilize and will continue to expand as long as the pumping continues and the boundary conditions remain the same. This is a classic groundwater mining situation, though slow recovery is possible if well pumpage is eliminated.

Analyses of real systems as well as of hypothetical desert-basin aquifers clearly demonstrate that streamflow depletion (capture) can continue long after pumping has ceased (Bredehoeft 2011b; Barlow and Leake 2012). They note that the rate of recovery depends on a number of factors, including hydraulic properties and boundary conditions. In the analysis by Barlow and Leake (2012) of the hypothetical desert-basin aquifer, following 50 years of pumping, it required an additional 100 years after pumping ceased to recover most (but not all) of the storage depletion. Under more severe streamflow-limited scenarios, such as reflected in the budgets of Figures 7 and 10, the recovery would take much longer.

Responses in Real Systems

Theis (1940) states that the source of water that balances well discharge is "always to some extent and in many cases largely from storage in the aquifer. Some groundwater is always mined." But he also points out that "in most artesian (i.e., confined) aquifers—excluding very extensive ones, such as the Dakota sandstone—little of the water is taken from storage." Were Theis' assessments basically correct? In real aquifer systems that have been developed (pumped) for decades, how much of the pumpage has been derived from storage? The previous analyses show that it can take many decades, if not centuries, for a stream-aquifer system—especially an areally extensive system in an arid climate—to reach a new equilibrium in response to long-term pumping stresses (also see Bredehoeft and Durbin 2009; Barlow and Leake 2012). During the transient response phase, the fraction of pumpage derived from storage depletion would tend to decrease with time, and the complementary capture fraction would correspondingly increase. The range of experiences in real aquifer systems that have been developed (pumped) for decades is examined, with analyses limited to aquifers, time periods, and areas for which adequate data are available for both estimates of storage depletion volume and estimates of total well withdrawals. Of course, at the scale of an aquifer system,
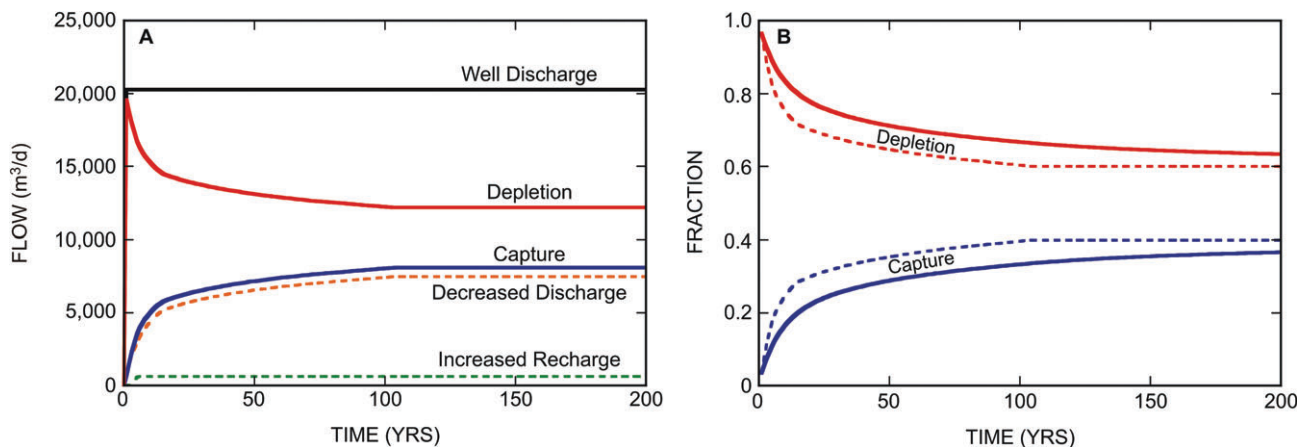
Figure 10. Simulated hydrologic budgets for streamflow-limited, high-pumpage case (A) and storage depletion (red curves) and capture (blue) fractions relative to pumping (solid lines show cumulative data and dashed lines show annual rates) (B).

the observed cumulative depletion is a complex response function of the interactions of multiple transient stresses, both natural and engineered, consistent with the principles of superposition.

Leakage from low-permeability confining units into pumped aquifers is a well-known and important process affecting the propagation of responses through an aquifer system. Typically, head declines will propagate slowly through confining layers, and the leakage will be derived largely from storage depletion in the confining unit until a new steady-state head distribution is eventually achieved (see Konikow and Neuzil 2007). Leakage also acts to slow the lateral propagation of head declines through an aquifer, thereby delaying the interaction with aquifer boundaries. Thus, streamflow depletion caused by pumping wells will take longer to occur and longer to reverse than in a nonleaky system.

Various estimates of long-term storage depletion in specific aquifers are available (e.g., see Konikow 2011, 2013b). There are 31 aquifers or areas in the United States and two outside the United States for which adequate data are available to estimate depletion and capture fractions (see Table S1, which shows that estimates for almost all areas represent cumulative volumes over periods of several decades). In many cases, the estimates of volumetric depletion include depletion in overlying and/or underlying confining units (methods and specific analyses are described by Konikow 2013b). Also, an estimate can be made for the United States as a whole based on cumulative withdrawals and depletion volumes over more than five decades. These aquifers and areas include a broad range of hydrogeologic settings and climates, so should be representative to some extent of global conditions. The areas for which data are available are mostly areas that have experienced relatively large-scale and long-term development of groundwater supplies. Because the estimates are generally based on long-term cumulative volumes, the depletion fractions for the most recent time would likely be smaller than the value computed on the basis of cumulative volumes and capture

fractions during the most recent time increments would likely be larger (see Figure 4). Note that these fractional values are not static. Rather, they would be changing slowly with time, although after several decades, the cumulative fractions are relatively stable and tend to change only very slowly.

In the United States, the distribution of depletion fractions shows a wide variance (Figure 11). The highest depletion fraction (0.97) is in the Death Valley regional flow system, which has an arid climate and few surface water resources. Outside the United States, the Nubian aquifer in North Africa has essentially zero recharge, no potential for increasing recharge, and an increasing magnitude of development. Even without the effects of development the system is undergoing a slow transient evolution of heads from a wetter period with recharge thousands to millions of years ago (Voss and Soliman 2013). Residual discharge is balanced by storage decreases. Yet a model study (CEDARE 2001) calibrated to 38 years of record (1960 to 1998) indicates that in 1998, the end of the study period, the storage depletion fraction was only 0.84 and the capture fraction was therefore 0.16, with the capture representing reductions in natural discharge (e.g., by a reduction in the discharge of springs at oases).

Theis' insight about confined aquifers was generally correct. For example, for 1901 through 1980 only about 30% of the pumpage in the areally extensive Cambrian-Ordovician aquifer in the Midwestern United States was derived from storage depletion. Theis' exception for the Dakota aquifer was also reliable, as about 78% of the withdrawals from the Dakota in South Dakota during 1881 through 1980 was balanced by a reduction in storage. However, as concluded by Konikow and Neuzil (2007), most of the storage depletion originated in the adjacent thick confining units—an aspect not noted by Theis. At the other end of the spectrum, intense groundwater development has occurred in the Floridan and adjacent aquifers in Florida and parts of Georgia and South Carolina. These areas have relatively high precipitation.
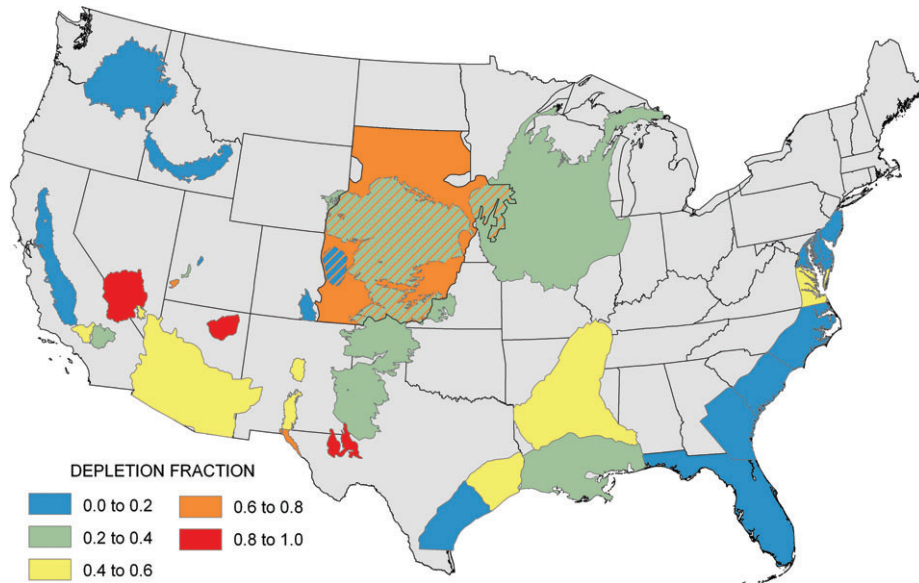
Figure 11. Estimated long-term cumulative storage depletion fraction in 31 areas and aquifers within the United States. Hatched patterns reflect areas where one aquifer overlies another. Also see data in Table S1.

The depletion fraction for this combined area is only about 0.01 for 1950 through 2005, so that about 99% of the pumping is derived from capture. For the United States as a whole for 1950 through 2005, the total net groundwater storage depletion volume is about 812 km^3 (Konikow 2013b) and the cumulative withdrawals are approximately 5340 km^3 (Kenny et al. 2009). Thus, the long-term depletion fraction is about 0.15 and the capture fraction is about 0.85.

Considering all 31 areas in the United States, the United States as a whole, and two aquifer systems outside the United States (Nubian aquifer [CEDARE 2001] and North China Plain [Cao et al. 2013]), an analysis of the frequency distributions (Figure 12) indicate that most systems have evolved to low cumulative depletion fractions (mean = 0.39) and high cumulative capture fractions (mean = 0.61). However, there can also be a wide variation within any particular areally extensive aquifer system. For example, the largest volume of storage depletion in the United States occurs in the High Plains Aquifer system. This large system underlies parts of eight states, and state by state data are also available (e.g., see McGuire et al., 2003; McGuire 2007). For cumulative volumes during 1950 through 2000, the depletion fraction for the entire High Plains Aquifer was about 0.27, but it ranged from 0.00 in the Nebraska portion (where there were slight water-table rises during this time period) to 0.42 in the Texas portion.

The storage depletion fractions also show some correlation with climate (Figure 13). The 33 data points in Figure 13 include separate values for the Texas and Nebraska parts of the High Plains Aquifer, but exclude averaged values for the United States as a whole. In general, where precipitation is higher and water tables are higher, one would expect a greater potential for pumping-induced drawdown to cause increases in recharge and/or
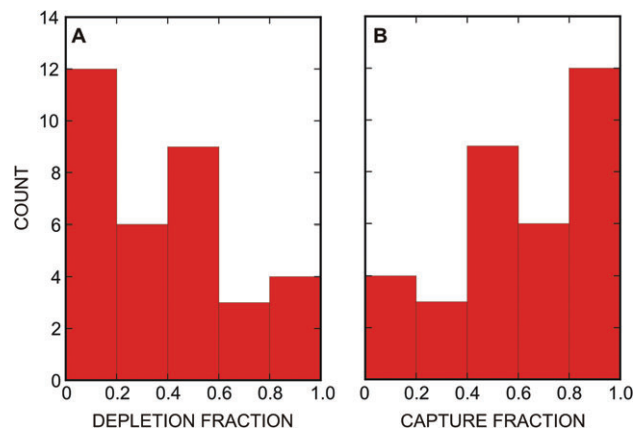

Figure 12. Histograms for (A) depletion fractions and (B) capture fractions in 34 areas and aquifers (from data in Table S1).

decreases in discharge. Also, in more humid climates, drainage densities tend to be higher, so that the effective distances from wells to surface water boundaries are generally shorter, especially in shallow aquifers; consequently, response times for inducing increased recharge or decreased discharge are shorter, which would tend to reduce relative storage depletion. The correlation coefficient (R) for this relation is 0.40, indicating a mild relation rather than a strong one, which can also be seen by the large spread of values about the regression line. It would be erroneous to assume that the cumulative depletion fraction can be accurately predicted on the basis of climate alone. Other factors that influence the cumulative depletion fractions include variability in the distances to influential boundary conditions, in hydraulic properties, and in time histories of well development and total aquifer withdrawals.
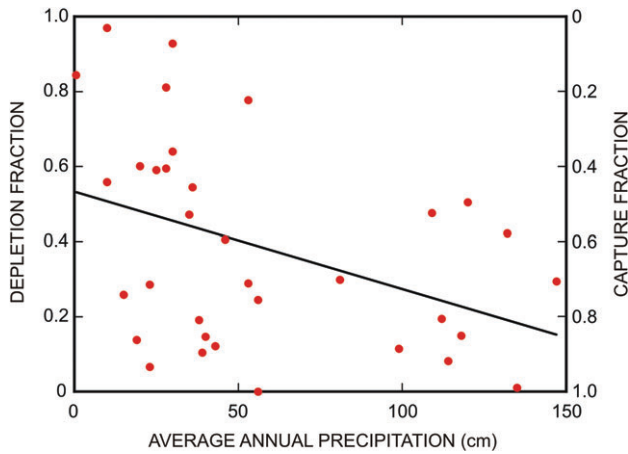
Figure 13. Relation between average annual precipitation and long-term cumulative depletion and capture fractions for 33 aquifer areas and subareas, showing a best-fit linear regression line (from data in Table S1).
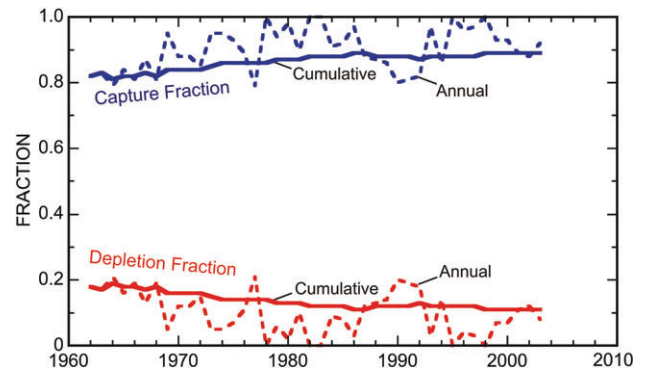

Figure 14. Results of water budget calculations of the Central Valley, California, calibrated groundwater-flow model (Faunt et al. 2009b), showing storage depletion (red) and capture (blue) fractions (solid lines for cumulative fractions; dashed lines for annual rates).

Well-calibrated and well-constructed simulation models of long-term responses in aquifer systems offer a means to analyze the sources of water derived from wells and how they vary with time. For this type of analysis, a well-constructed model would be free of artificial boundaries that would affect calculations of groundwater storage depletion and capture for a groundwater system. This will be illustrated briefly using two representative examples of such well documented model analyses.

The Central Valley of California is a major agricultural area in a large valley with an area of about 52,000 km^2 (Williamson et al. 1989; Bertoldi et al. 1991). The Central Valley has an arid to semiarid Mediterranean climate, where the average annual precipitation ranges from 13 to 66 cm (Bertoldi et al. 1991). Streamflow is an important factor in the water supply of the valley. Groundwater development began around 1880. By 1913, total well pumpage was about 0.44 km^3 annually (Bertoldi et al. 1991). During the 1940s and 1950s, the pumpage increased sharply, and by the 1960s and 1970s averaged about 14.2 km^3/yr. By the 1980s there were approximately 100,000 high-capacity wells in the Central Valley for either irrigation or municipal supply. During 1962 through 2003, withdrawals from irrigation wells averaged about 10.6 km^3/yr (Faunt et al. 2009a).

A transient groundwater-flow model of the Central Valley was developed for 1961 through 2003 (Faunt et al. 2009b). The model indicates that the decrease in groundwater storage from 1961 through 2003 was about 71.2 km^3. However, the total decrease in groundwater storage from predevelopment conditions until 1961 was about 58 km^3 (Williamson et al. 1989, 95), and this is not accounted for in the 1961 through 2003 model. As expected, the cumulative fractions are smoother than the annual fractions (Figure 14), and the year-to-year variability in annual fractions is largely controlled by variations in annual pumpage and precipitation. The depletion and capture fractions (both cumulative and rate based) for the first year of the simulation period are 0.18

and 0.82, respectively. But over the 42-year simulation period, the fractional rates did not change greatly, as reflected by the relatively small change in the cumulative storage depletion and capture fractions to 0.11 and 0.89, respectively, indicating that such long-term cumulative fractions (such as presented in Figure 4) are relatively stable and representative of conditions in the aquifer. Compared with the generic fractional curves (Figure 1), it is evident that this model of the Central Valley of California, which begins about 80 years after the start of pumpage, cannot and does not represent the expected early-time system responses of high depletion fractions and low capture fractions, so that the cumulative depletion fraction would be too small (and cumulative capture fractions too high) in the early years of these simulation results.

Antelope Valley, California, is a small (2400 km^2) topographically closed basin with an arid climate (average annual precipitation is less than 25 cm). The basin contains a thick (more than 1500 m in places) sequence of unconsolidated alluvial and lacustrine sediments. Surface water is limited, and the area includes several springs, playas, and intermittent streams that drain into the playas (Leighton and Phillips 2003). Delivery of some imported water began in 1986. Leighton and Phillips (2003) note that recharge to the groundwater system is primarily from the infiltration of precipitation runoff near the valley margins, and discharge from the aquifer system was primarily from evapotranspiration. Development of the groundwater system began around 1915 and increased rapidly into the 1950s. Pumpage peaked at more than 0.37 km^3/yr in the 1950s and 1960s, but by the mid-1980s had declined to about 0.12 km^3/yr (Galloway et al. 2003). Groundwater pumping has caused large water-level declines in the basin, resulting in a major decrease in evapotranspirative discharge (Leighton and Phillips 2003).

A 3D transient MODFLOW model was developed and calibrated to simulate groundwater-flow and aquifer-system compaction in the area (Leighton and Phillips 2003). The model was first calibrated to represent

predevelopment conditions prior to 1915. Then the transient model was developed using 81 1-year stress periods to simulate the period of 1915 through 1995 inclusive. More detailed descriptions of model parameters, boundary conditions, and the calibration process are presented by Leighton and Phillips (2003). The results of the transient simulation indicate that more than 10.5 km^3 of groundwater was removed from storage during 1915 through 1995, with most of the storage change occurring between about 1945 and 1975 (Leighton and Phillips 2003). The model-computed water budgets indicate that most of the pumpage was derived from storage depletion during the first few decades of development, but that the capture fraction generally increased with time—becoming dominant since the early 1970s (Figure 15). As would be expected in this type of basin, there appears to be a strong direct correlation between pumpage and storage depletion (Figure 15B). Similar to the Central Valley (Figure 14), in the Antelope Valley the fractions based on annual flow rates show greater variability than the cumulative fractions (Figure 15A), and the variability in annual fractions is largely controlled by variations in annual pumpage. During 1988 through 1990, the annual pumpage was the smallest since 1925, and during these 3 years there were small increases in net storage. Capture was comprised largely of increased recharge from irrigation return flows, but during the first 5 decades decreased evapotranspiration also contributed to capture. After 1985, increased recharge from imported water also provided a substantial offset of the effects of pumping.

The evolution of storage depletion and capture fractions in Antelope Valley indicate the transient nature of these factors, and the data (Figure 15A) show that the system is still continuing to evolve. It has not achieved a new permanent equilibrium state and storage depletion—though temporarily halted during 1988 through 1990—continues to increase even though annual pumpage has decreased substantially since its previous peak rates. The change in the fractions during the historical period of record covers the full range of values. The difference between the cumulative and rate-based fractions is much greater than seen in the results for the Central Valley. For example, at the end of this study period (1995) the annual storage depletion fraction was 0.18 while the cumulative depletion fraction was 0.59. Achieving a sustainable groundwater development practice would require that the annual depletion fraction approach and be maintained at values at or close to zero and that the environmental consequences of capture be acceptable.

Conclusions

Nearly 75 years have passed since Theis (1940, 1941) published his classic papers that clearly elucidated the sources of water derived from wells and the effect of pumping a well on flow in a nearby stream. His principles and guidance have stood the test of time, and are not only still relevant today, but should be required reading
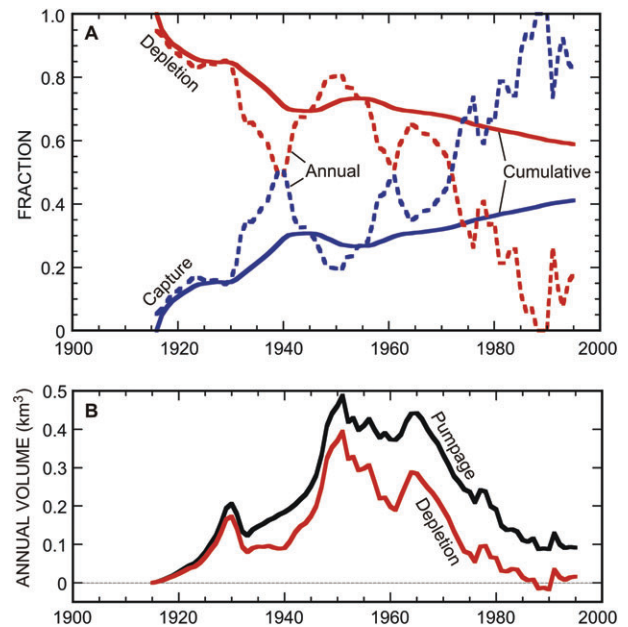

Figure 15. Results of water budget calculations of the Antelope Valley, California, calibrated groundwater-flow model (Leighton and Phillips 2003), showing (A) computed storage depletion fractions (red) and capture fractions (blue), with solid lines representing fractions based on cumulative data and dashed lines representing annual values, and (B) estimated annual pumpage (black) and calculated annual storage depletion volume (red).

for every groundwater analyst. His overriding principle is the simple message that all water discharged by a well must be balanced by a loss of water somewhere—either from storage or by capture. This study expands a little on Theis' work by examining two aspects that he did not focus on. First, we analyze how the balance is affected if capture is constrained by a limited availability of water. Theis (1941) had assumed "that the stream maintains a flow past the pumped area." Second, we analyze a number of real systems in which sufficient data are available to assess the partitioning of the balancing components into storage depletion and capture fractions after a long history of pumpage.

Groundwater storage depletion and capture can be measured in terms of nondimensional fractions relative to pumpage. These measures can be computed on the basis of either cumulative volumes or flow rates. The former will yield more moderated values that reflect long-term averaged responses (i.e., rates integrated over time), but may not accurately indicate system status at any particular time years after development started. These measures will tend to change exponentially with time, and the complementary fractional values of storage depletion and capture, based on flow rates, will effectively reach 0.0 and 1.0, respectively, if sufficient water for capture is available at aquifer boundaries. When this occurs, the aquifer system has attained a new equilibrium condition and continued development should be sustainable. However, if prior to equilibrium aquifer bounds are reached that preclude any further increases in recharge and decreases

in discharge, then a new equilibrium cannot be attained and storage depletion will continue to occur.

The potential for well withdrawals to be balanced by capture would be constrained if there is insufficient water available at aquifer boundaries to meet the increased demands imposed by drawdown-induced steepening of hydraulic gradients. Evidence of constraining conditions includes streams or springs going dry following an extended period of pumpage within the aquifer. When capture is constrained, the relative amount of pumpage balanced by (or derived from) capture decreases and the amount derived from storage depletion increases. In severely constrained cases, all sources of capture can reach their limits. Then, discounting natural fluctuations in recharge from precipitation, with continued steady pumpage the fractions of the pumping rate derived from capture and storage depletion will stabilize with time. This means that groundwater levels will continue to decline—a classic groundwater mining situation. This can then continue until drawdowns themselves start to limit the pumpage because of increased lifts and higher costs of pumping or because reduced saturated thicknesses decrease well yields. In this sense, groundwater storage depletion itself should eventually be self-limiting and unsustainable.

In an illustrative test problem representing pumping in a hypothetical desert-basin aquifer, the only source of capture was from the stream. In this case, rates of capture (streamflow depletion) exceeded storage depletion after 17 years. As long as the stream did not go dry at any point, the largest contributor to capture was increased recharge from induced infiltration. But if the stream did go dry and capture was thereby constrained, then the amount of pumpage derived from storage depletion increased relative to the nonconstrained condition, and the amount derived from capture correspondingly decreased. Also, under capture-constraining conditions, decreases in groundwater discharge to the stream became the larger contributor to total capture after 36 years because the central reach of the river went dry and induced infiltration could no longer increase.

There are 31 specific areas or aquifers within the United States and two outside the United States for which adequate data are available for both total withdrawals and cumulative storage depletion to allow estimates to be made of long-term storage depletion and capture fractions. The mean depletion fraction is 0.39 and the mean capture fraction is 0.61. For the United States as a whole during 1950 through 2005, about 15% of total pumpage was derived from a reduction of storage of groundwater—a depletion fraction of 0.15. But depletion fractions vary widely within the United States and even within any given large aquifer system. For example, the fraction of long-term (1950 to 2000) pumpage derived from storage depletion in the High Plains aquifer is about 0.27, but ranges from 0.0 in Nebraska (where there was a slight water-table rise) to 0.42 in Texas. In general, storage depletion fractions tend to be higher in arid regions, but the relation is not strong and depletion fractions cannot

be accurately predicted on the basis of climate alone. These fractions are time dependent, but analyses from the Central Valley and Antelope Valley, both in California, support the notion that cumulative fractions tend to be relatively stable at late times (typically a few decades after major development begins).

Well-calibrated simulation models offer a means to analyze the sources of water derived from wells and how the fractions vary with time—a modern tool not available to Theis. To reliably simulate the history of storage depletion and capture in a groundwater system, groundwater-flow models must start with initial conditions representative of predevelopment times and conditions. Such models also provide water managers with a tool to predict future changes in storage and streamflow depletion in response to possible changes (or no changes) in water management policies.

Groundwater storage depletion and capture problems must be confronted on local and regional scales, where water managers faced with unsustainable withdrawals will necessarily have to take actions to reduce demand and/or increase supply through managed aquifer recharge, desalination, and/or developing alternative sources. Otherwise, storage depletion of the aquifer system will itself ultimately limit withdrawals—in ways that are economically and environmentally less than optimal.

Acknowledgments

Supporting Information

Additional Supporting Information may be found in the online version of this article:

Table S1. Supporting Data and References for Estimates of Groundwater Depletion

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