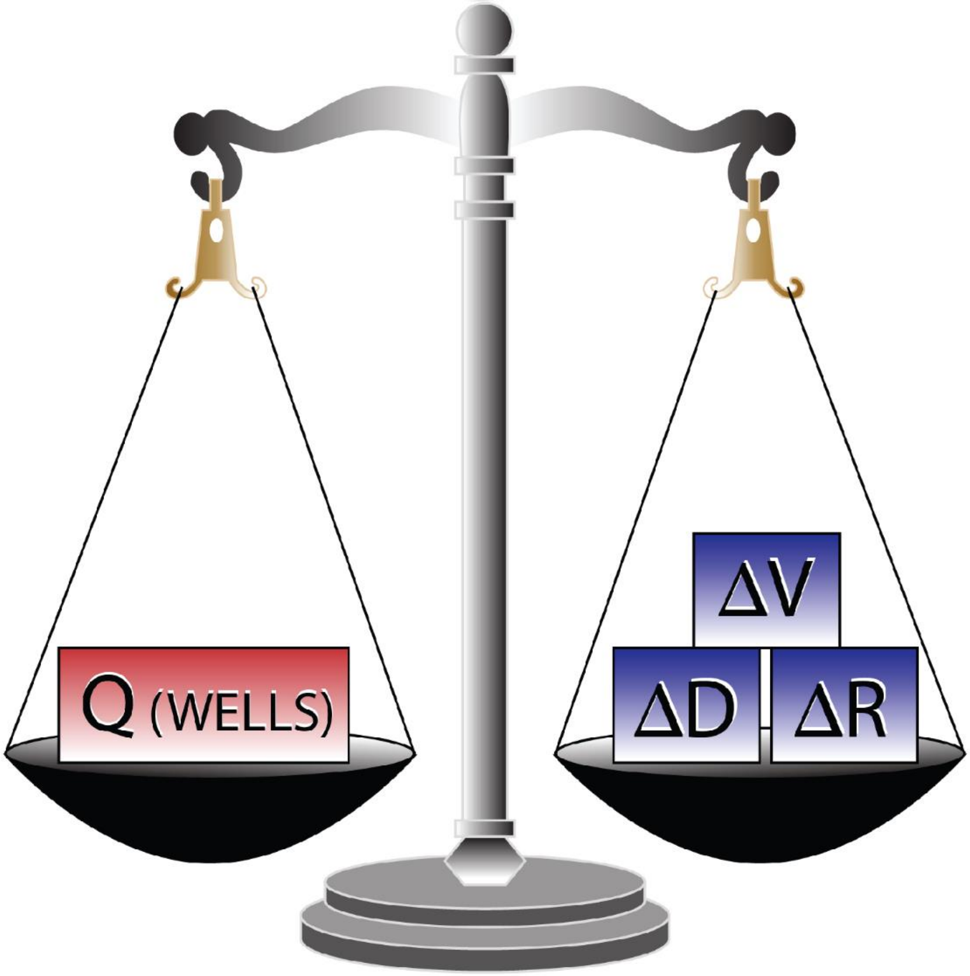

Groundwater Resource Development

Effects and Sustainability

Leonard F. Konikow and John D. Bredehoeft

THE GROUNDWATER PROJECT

Groundwater Resource Development: Effects and Sustainability

The Groundwater Project

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Thank you.

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Groundwater Resource Development: Effects and Sustainability

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Dedication

Dedicated to the generous sharing of groundwater knowledge.

The Groundwater Project Foreword

The United Nations Water Members and Partners establish their annual theme a few years in advance. The theme for World Water Day of March 22, 2022, is "Groundwater: making the invisible visible." This is most appropriate for the debut of the first Groundwater Project (GW-Project) books in 2020, which have the goal of making groundwater visible.

The GW-Project, a non-profit organization registered in Canada in 2019, is committed to contribute to advancement in education and brings a new approach to the creation and dissemination of knowledge for understanding and problem solving. The GW-Project operates the website https://gw-project.org/↗ as a global platform for the democratization of groundwater knowledge and is founded on the principle that:

"Knowledge should be free and the best knowledge should be free knowledge." Anonymous

The mission of the GW-Project is to provide accessible, engaging, high-quality, educational materials, free-of-charge online in many languages, to all who want to learn about groundwater and understand how groundwater relates to and sustains ecological systems and humanity. This is a new type of global educational endeavor in that it is based on volunteerism of professionals from different disciplines and includes academics, consultants and retirees. The GW-Project involves many hundreds of volunteers associated with more than 200 hundred organizations from over 14 countries and six continents, with growing participation.

The GW-Project is an on-going endeavor and will continue with hundreds of books being published online over the coming years, first in English and then in other languages, for downloading wherever the Internet is available. The GW-Project publications also include supporting materials such as videos, lectures, laboratory demonstrations, and learning tools in addition to providing, or linking to, public domain software for various groundwater applications supporting the educational process.

The GW-Project is a living entity, so subsequent editions of the books will be published from time to time. Users are invited to propose revisions.

We thank you for being part of the GW-Project Community. We hope to hear from you about your experience with using the books and related material. We welcome ideas and volunteers!

The GW-Project Steering Committee

September 2020

Foreword

For centuries humans have been pumping aquifers but it wasn't until 1935, when C.V. Theis at the United States Geological Survey (USGS) introduced the unsteady-state solution to the groundwater flow equation, that the effect of aquifer storage in confined aquifers was taken into account based on the physics of the system. This was the start of the modern era of assessment of aquifer exploitation and sustainability in which all of the potential sources of extracted groundwater are conceptualized for the aquifer water budget. However, surprisingly for many aquifers around the world where there is reliance on the extracted water, the water budget is not well understood, or is insufficiently quantified to serve for effective water management.

Depletion of aquifers is the norm in many countries. When an aquifer is pumped for a substantial period of time, such as years or decades, the answers to the questions: 'Where is the pumped water coming from? and What is changing with time?' often are not known. Hence, many aquifers enter into depletion without understanding of what is happening.

This book (Groundwater Resource Development: Effects and Sustainability) provides the concepts and principles that underpin understanding of aquifer depletion and sustainability. It explains the role that groundwater recharge plays in sustainability of groundwater exploitation. Although simple in theory, the role of groundwater recharge has remained elusive in the quest for science-based water management.

The authors of this book, Lenny Konikow and John Bredehoeft (both emeritus of the USGS), have played key roles in the development and applications of the most advanced computer models for simulating aquifer exploitation and have long been leaders in the quest to integrate the appropriate models and concepts into sustainable groundwater development.

John Cherry, The Groundwater Project Leader

Guelph, Ontario, Canada, September 2020

Preface

This book discusses ideas associated with development of groundwater at the macro scale, including how well pumpage affects surface water, groundwater storage, and the long-term sustainability of groundwater development. We do not focus on well hydraulics, well interference, or design of wells or well fields. Instead, we begin with where water comes from when one pumps a single well and proceed to the scale at which an entire aquifer system is developed. These ideas date back to a paper by Theis (1940) on the source of water derived from wells. Theis (1940) noted that all pumpage is balanced by a loss of water somewhere, with the loss during early time coming largely from storage. As time advances, pumpage tends to be balanced increasingly by some combination of increases in recharge and decreases in discharge.

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Leonard Konikow
John Bredehoeft

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1 Introduction

When one installs a supply well, it is usually with the hope that the well will reliably provide water for a long time (that is, its use will be sustainable for future generations). The science of hydrogeology has developed principles that can potentially (1) guide development projects that are created to last for a long time, and/or (2) assess the longevity of an existing development.

In this Book it is our intent to discuss the ideas associated with development of groundwater at the macro scale. We do not focus on well hydraulics, well interference, or design of wells or well fields. Instead, we begin with where the water comes from when one pumps a single well and proceed to the scale in which we develop an entire aquifer system. These ideas are not new; they date back to a paper by Theis (1940) on the source of water derived from wells -- many investigators think this is Theis' best paper. Theis (1940) noted that all pumpage is balanced by a loss of water somewhere, with the loss during early times coming largely from groundwater storage. Theis (1940) concludes 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 had been rejected recharge. … further discharge by wells will be made up in part by a diminution in the natural discharge.*" (Rejected recharge is water available to potentially enter the aquifer but cannot because of aquifer storage capacity or conductance limitations.)

There are a number of processes and environmental consequences involved in groundwater development. Many relevant ideas are discussed in other Groundwater Project books, including: the basic theory of groundwater flow and transport, multiphase flow, unsaturated zone flow, the physics of recharge, the geologic occurrence of groundwater, groundwater-surface water interactions; and the related topics of land subsidence/consolidation and coastal hydrogeology/seawater intrusion.

2 Groundwater Use

Groundwater withdrawals in the United States and globally expanded dramatically during the last half of the 20th century. Estimates of groundwater withdrawals in the United States are available from the United States Geological Survey for every five-year period since the 1940s. These data (Figure 1) show a steady growth through the 1970s, followed by a more or less stabilization of groundwater use. In 2015, total groundwater withdrawals were about 114 km^3/yr (82,300 million gallons per day [MGD]), and 69 percent of the groundwater withdrawals was used for irrigation and 18 percent was used for public supply (data from Dieter et al. 2018). Approximately half the population of the world depends upon groundwater for its drinking water supply, and groundwater supplies about 43 percent of the global irrigation water supply (Siebert et al., 2010; WWAP, 2015). Total global groundwater use in 2010 was approximately 982 km^3/yr, and about 70 percent of that was used for irrigation (Margat and van der Gun, 2013). As of 2010, the country with the largest groundwater use was India (about 250 km³/yr) -- more than the second and third largest users combined (China and the United States, with about 110 km³/yr each). This large and expanding use of groundwater is the primary driving force for concerns about groundwater depletion and sustainability of groundwater pumping.

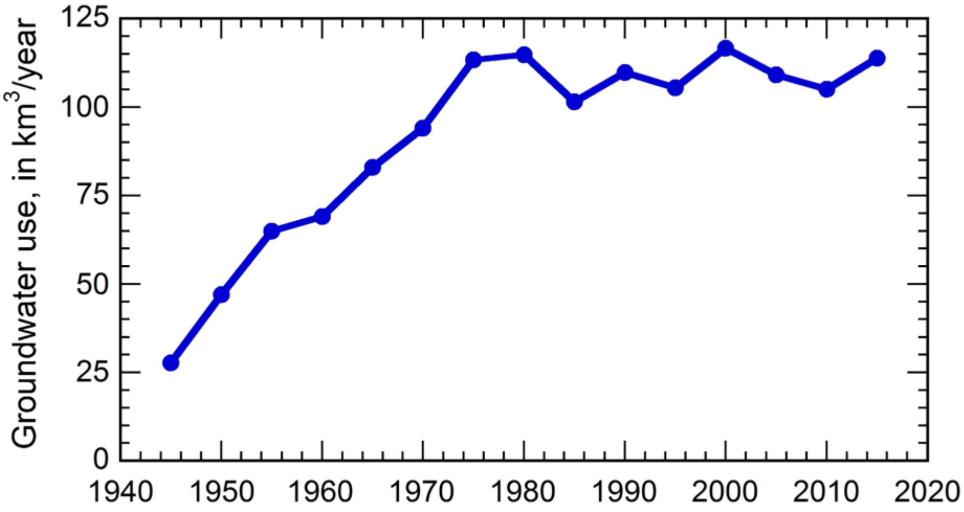

Figure 1 - Total fresh groundwater withdrawals in the United States from 1945 to 2015 (data source: Guyton, 1950; https://waterdata.usgs.gov/nwis/wu↗).

3 Sustainability of Groundwater Development

Groundwater is often characterized as a "renewable resource." Yet data now accumulating indicate that much of the current development of groundwater is depleting the resource at rates that cannot be sustained -- in many places groundwater is being "mined" at high rates -- contradicting its renewability over human timeframes. This poses a challenge to groundwater scientists and managers -- *can the resource be developed in a sustainable manner, and if so, how can that goal be accomplished?* This is the premier task for the groundwater community in the 21st Century.

"Groundwater sustainability" has been defined in various ways. Alley et al. (1999) define groundwater sustainability as "development and use of groundwater in a manner that can be maintained for an indefinite time without causing unacceptable environmental, economic, or social consequences." However, they note that "the definition of 'unacceptable consequences' is largely subjective and may involve a large number of criteria." It might, for example, include streamflow depletion, drying of springs or wetlands, loss of vegetation, and/or water-level declines in wells. Price (2002) argues that sustainability should be related to a specific time period over which it is to be evaluated.

The concept of *safe yield* of a groundwater system is often used by water managers to place limits on the total number of wells and/or total pumping from a given aquifer. This need arises because groundwater is a common pool resource in which extraordinarily high usage by one or more parties may be highly beneficial to those parties (and their self-interests) but harmful to the long-term viability of the resource (through excessive depletion) and to everyone else's continued future use of the common resource. Overdevelopment of an aquifer is a classic example of "The Tragedy of the Commons" (Hardin, 1968).

Meinzer (1923) defined *safe yield* as "the rate at which water can be withdrawn from an aquifer for human use without depleting the supply to such an extent that withdrawal at this rate is no longer economically feasible" (Alley and Leake, 2004). Freeze and Cherry (1979) state, "Todd (1959) defines the *safe yield* of a groundwater basin as the amount of water that can be withdrawn from it annually without producing an undesired result. Any withdrawal in excess of safe yield is an *overdraft*." Freeze and Cherry (1979) indicate that there is widespread dissatisfaction with the term among hydrologists. Alley and Leake (2004) indicate that the dissatisfaction arises in large part because the term is vague. Misinterpretation implies a fixed underground water supply. Groundwater supply depends on the particular locations of wells and is only fixed when the locations of wells are specified. A yield that is safe from one perspective, such as depletion of groundwater storage, might not be so safe from the standpoint of discharge areas of aquifers, such as lakes, springs, and wetlands (Alley and Leake, 2004). Pumping rates that are considered safe by well owners may yield streamflow depletion that is an undesired result for surface-water users. Freeze and Cherry (1979), among others, suggest that an optimization approach within a socioeconomic framework would be a better way to assess an *optimal* (rather than *safe*) level of development in a groundwater basin.

The desirability and value of sustainable development of groundwater, and management approaches to help achieve that, have been discussed by Gleeson et al. (2012). Sustainable groundwater development seeks to preserve the resource for use by future generations. But "sustainability" should be assessed in a larger perspective that includes impacts on surface-water flows, other environmental consequences (e.g., land subsidence and water-quality changes), as well as other linkages, such as socio-economics (Alley and Leake, 2004; Hiskock et al., 2002; Kendy, 2003; MacEwan et al., 2017; National Research Council, 2013; and Van der Gun and Lipponen, 2010).

In contrast, "groundwater mining" is the removal of water from storage in the aquifer that cannot be renewed (or replaced) within a human timeframe (Thomas, 1955; and Bredehoeft and Alley, 2014). By definition, such rates of groundwater development cannot be maintained indefinitely. However, the length of time that such mining can continue depends on the stock of groundwater in storage (i.e., the volume of recoverable groundwater in the aquifer) and the rate of withdrawal through wells. Clearly there are cases in which substantial rates of groundwater development and mining can continue for decades and even centuries. It is possible that the economic, social, and political benefits of the water provided by such groundwater mining may be very large, and that some mining is acceptable within a socioeconomic framework.

If sustainable development of a groundwater resource is not practically achievable, then the question might be whether to manage the resource in order to extend its life or just let development eventually lead to functional limitations on groundwater withdrawals. Should non-sustainable groundwater development (sometimes called "overdraft" or "overexploitation") ever be considered acceptable? Should society weigh the shorter-term economic and societal benefits of a time-limited use of the groundwater resource in an aquifer against the "costs" and environmental effects of the development, and how would that be done? Price (2002) points out that many developments in human history were non-sustainable, but contributed substantially to human progress. Perhaps non-sustainable groundwater development (groundwater mining) is acceptable to society if it is done with a full knowledge and understanding that such groundwater use can only continue for a limited (but reliably estimated) time. Depletion of groundwater resources without an understanding of its existence, timing, and consequences should be considered unacceptable policy. Hydrogeologists can provide that predictive understanding, which offers a long-view scientific basis for policy makers and water managers to make sound and defensible policy decisions.

3.1 Basic Assumptions

In considering the impacts of groundwater development, we make a number of basic assumptions that simplify the discussion; they are:

1. All water molecules under consideration have the same composition -- in other words we are not considering groundwater transport of differing quality water.
2. The groundwater flow equation describes the material (mass) balance -- and Darcy's Law can be applied to obtain the direction and rate of groundwater flow at any point within the system. The dependent variable within the groundwater system being considered is hydraulic head.
3. Groundwater systems exist over long periods of time prior to development. Before development they are in a kind of steady state or long-term dynamic equilibrium in which short-term and seasonal fluctuations in precipitation and recharge average out over the longer-term period.
4. Conservation of the mass of water within the system is preserved.

All of these assumptions can be relaxed; needless to say, one must take care in relaxing these ideas.

3.2 Water Balance

Conservation of mass is a basic principle for understanding groundwater flow. Combining this principle (as a continuity equation) with Darcy's Law yields a partial differential equation describing changes in head and flux through a groundwater system in response to stresses and boundary conditions. This is discussed in another Groundwater Project book↗ that describes the principles of groundwater flow, including the groundwater flow equations. We can also present a simpler but useful algebraic equation that describes a water balance in an aquifer. A quantitative global water balance statement for a groundwater system can be expressed as Equation 1:

$$\Delta V/\Delta t = (R_0 + \Delta R_t) - (D_0 + \Delta D_t) - Q_t \tag{1}$$

where:

ΔV = change in the volume of water in storage in the aquifer (L[3])

Δt = length of a time increment of interest (T)

$\Delta V/\Delta t$ = global (aquifer-wide) rate of change in storage (L[3]/T)

R_0 = total recharge rate into the undisturbed natural system (the virgin recharge) (L[3]/T)

ΔR_t = global change in recharge caused by the pumping (L[3]/T)

D_0 = total discharge rate from the undisturbed system (prior to development) (L[3]/T)

ΔD_t = global change in discharge caused by the pumping (L[3]/T)

Q_t = global rate of pumping during the time period, t (L[3]/T)

3.3 The System Prior to Development

Prior to development (Q=0) one assumes that on average the recharge to a groundwater system is balanced by the discharge, and there is no long-term change in storage (ΔV =0). This is especially true when considering that groundwater systems evolve over geologic time. Admittedly there are wet and dry years, as well as seasonal and even daily variability in precipitation and recharge, but over the long-term it is reasonable to assume that the annual fluctuations balance out. Under this assumption:

$$R_0 = D_0 \tag{2}$$

for conditions prior to development (Figure 2). In many investigations considerable effort is spent in trying to estimate the undisturbed (or "natural") recharge, R_0. Usually it is best to estimate both the undisturbed recharge and discharge simultaneously, as they must be equal and constrain one another. As explained in the "A New Perspective" section of this book, the undisturbed recharge is not as important in analyzing groundwater development as many people think.

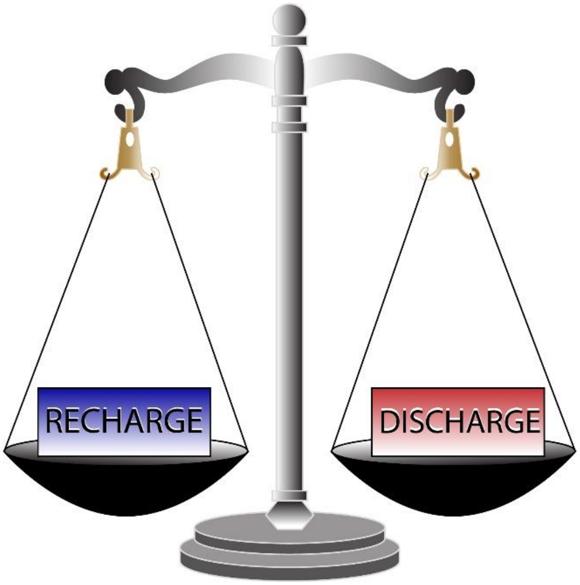

Figure 2 - Schematic illustration of the natural groundwater balance, the long-term average balance between recharge and discharge in an undisturbed (natural) groundwater system (Konikow and Bredehoeft, 2020).

This is not to say that some aquifer systems cannot continue to undergo slow transient changes in their water balance over tens of thousands of years or longer if they are relatively large and somewhat slow to respond to major changes in climate and/or sea level during "recent" geologic times. If climate change near the end of the Pleistocene has caused a substantial reduction of average recharge rates to the aquifer, the new recharge regime cannot balance or maintain the discharge from the system. One implication of this imbalance is that the system will continue to drain (and water in storage will be reduced over time) as the system tries to reach a new equilibrium condition and discharge is reduced accordingly. If the properties and boundary conditions of the aquifer are such that reaching this new equilibrium would take centuries to millennia, then we can today see groundwater systems that are still responding to the vast changes in climate, glacial extent, and sea level 10,000 to 20,000 years ago. Groundwater occurring in such systems is sometimes referred to as "fossil water" because it is essentially nonrenewable under present climatic conditions. One example is the Nubian aquifer of North Africa, where modern natural predevelopment discharge (and storage depletion) was estimated to be about 86 m^3/s (2.7 km^3/yr) prior to development (about 1960 and earlier), while natural recharge was close to zero (Voss and Soliman, 2014). But such examples are the exception rather than the rule -- most aquifer systems were in balance (between recharge and discharge) prior to human development.

3.4 Pumping

When a well is pumped, it introduces a new discharge from the system. As elucidated by Theis (1940), water to supply the well comes from (or is balanced by) three potential sources: (1) an increase in the recharge to the aquifer caused by the pumping, (2) a decrease in groundwater discharge from the aquifer caused by the pumping, and (3) a reduction in groundwater storage in the aquifer system, or some combination of the three. This principle simply means that water mass is conserved, and that additional (new) discharge by pumping must be balanced or compensated by changes in other elements of the aquifer's water budget (as graphically illustrated in Figure 3). Based on the equivalency expressed in Equation 2, one can simplify the quantitative global water balance statement for the system (Equation 1) to:

$$\Delta R_t - \Delta D_t - \Delta V/\Delta t \ = Q_t \tag{3}$$

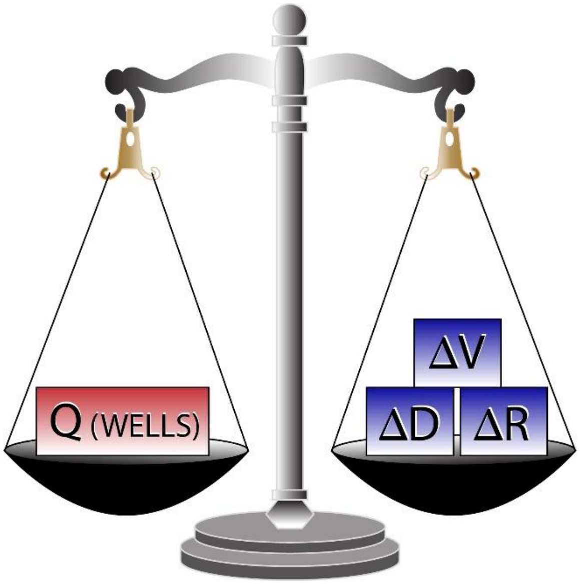

Figure 3 - Schematic illustration of the developed groundwater balance required to offset a new pumping stress in an aquifer; ΔV is the change in storage, ΔD is the change in groundwater discharge, and ΔR is the change in recharge (Konikow and Bredehoeft, 2020).

We define a new quantity ($\Delta R_t - \Delta D_t$) as the *capture* (Lohman et al., 1972)**.** That is, *capture* encompasses the changes in recharge and discharge caused by drawdown and changes in hydraulic gradients resulting from pumping a well, and represents water "captured" by the well that otherwise would not have entered the groundwater system or otherwise would have discharged from the groundwater system naturally. When the change in the volume of water in storage, ΔV, is negative, the change represents a depletion of the volume (or mass) of groundwater stored in the aquifer. The balance depicted in Figure 3 holds regardless of whether the individual terms are all represented as rates or as cumulative volumes. It is important to distinguish *capture* from *capture zone*, which represents the three-dimensional, volumetric portion of a groundwater flow field that discharges to a well (Anderson, et al., 2015; Barlow et al., 2018). The capture zone may or may not include the area where capture is occurring.

3.5 Long-Term Pumping Equilibrium (Development)

In every case of groundwater development some water must be removed from storage in order to create a cone of depression that will create a local head gradient adjacent to the well that causes groundwater to flow into the well. However, commonly the intent is to end up with a system that will persist through time (i.e., where the pumping will be sustainable indefinitely, or at least can be maintained for some acceptably long period of time). Development that can persist through time is one in which eventually storage is no longer being depleted and pumping is balanced by capture:

$$\Delta V / \Delta t = 0 \tag{4}$$

In this new equilibrium (or dynamic equilibrium) pumping is balanced entirely by capture. From this perspective the questions for an investigator or water manager are:

1. What are the changes in recharge and/or discharge brought on by the pumping -- what is the capture?
2. Can the system that now includes the pumping reach a new equilibrium in which storage is no longer being depleted?
3. If a new equilibrium can be reached, how long will that take to occur?
4. If a new equilibrium cannot be achieved, for how long can the pumping be maintained?
5. Is the capture acceptable from a water (and environmental) policy and management perspective?

The answers to these questions depend upon the particulars of the system being considered. For example, where pumping is situated relative to discharge often makes a difference -- both in the feasibility of reaching a new equilibrium and in how long it takes. It makes a big difference where wells are located in a groundwater system.

At early times after a well starts to pump and the cone of depression begins to develop, the pumped water is derived exclusively from a reduction (or depletion) in the volume of water stored in the aquifer. As the cone of depression spreads out and begins to affect aquifer boundaries, more and more of the water pumped from the well will be balanced by capture (Figure 4). This time dependence is a key to understanding and predicting the effects of groundwater development. The time scale for the response curves depends on the hydraulic properties of the aquifer and the distance of the well from recharge and discharge locations. But as noted by Bredehoeft and Alley (2014), this transition period can be very long -- perhaps years to decades to centuries -- especially if the system under development is dominated by water-table conditions with large storativity. The pumping rate itself does not affect the relative (or fractional) responses shown in Figure 4, but will be proportional to the actual volumetric magnitude of the effects. A quantitative method to estimate the timing of capture is presented in detail in the section of this book titled "Estimating the Magnitude and Timing of Streamflow Depletion."

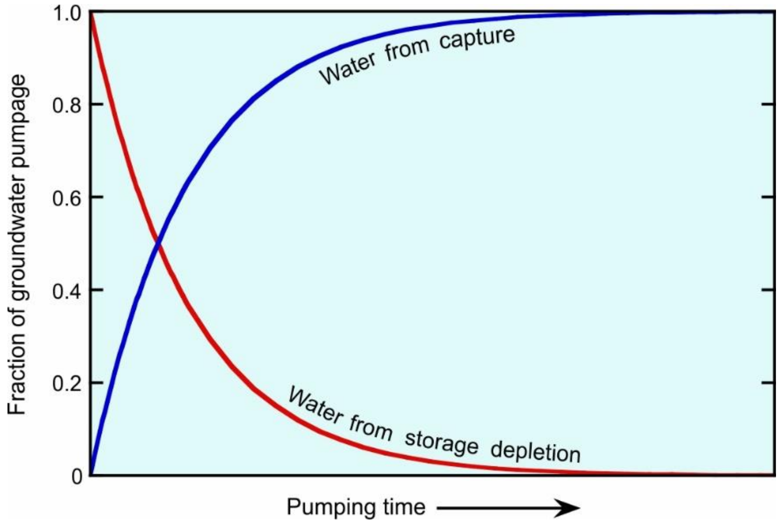

Figure 4 - Sources of water to a well can change with time (*from* Konikow and Leake, 2014, as modified from Alley et al., 1999, and Barlow and Leake, 2012).

3.6 A New Perspective

For many these ideas represent a new perspective. One must understand the dynamics of the groundwater system to assess development impacts. The static information on what was the undisturbed recharge is not nearly as important as the dynamics of capture. This comes as a shock to many who consider undisturbed ("natural") recharge to be all important in evaluating sustainable groundwater development (Bredehoeft, 2007; Bredehoeft et al., 1982). It is a common misconception that groundwater development is "safe" if the average annual rate of groundwater withdrawal from an aquifer does not exceed the average annual rate of natural recharge (Alley and Leake, 2004). Bredehoeft et al. (1982) call this the "water-budget myth" and clarify that hydraulically maintainable development depends on pumping becoming balanced by capture.

Describing the dynamics of how a particular groundwater system functions is best done today with a well-calibrated numerical groundwater model, as is illustrated later in this book with two case studies. Such a model solves the groundwater flow equation for the specified boundary and initial conditions and provides the global mass balances (water budgets) described by Equations 1 through 4.

4 Storage Depletion

As discussed, one factor that contributes to balancing of groundwater pumping is the removal of water from storage in the pore spaces of the subsurface saturated zone. Coincident with storage depletion are water-level declines, either from a declining water table in an unconfined aquifer or from lowering of the potentiometric head in a confined aquifer. Groundwater storage depletion and accompanying water level declines have a number of consequences.

The volume of water that an aquifer releases from storage per unit surface area per unit decline in head in response to pumping is characterized by the dimensionless storage coefficient (S) (Lohman et al., 1972). In a confined aquifer, the water released from storage is derived from the expansion of water and compression of the aquifer as the head declines. In an unconfined (or "water table") aquifer, the primary source of water from storage comes from gravity drainage of the pore spaces as the water table declines, and S is essentially equivalent to the dimensionless specific yield (S_y) (Lohman et al., 1972). In confined aquifers, values of S typically range from 5×10^{-5} to 5×10^{-3}, whereas for unconfined aquifers the values of S_y typically range from 0.01 to 0.30 (Freeze and Cherry, 1979). Consequently, a given (or unit) withdrawal rate in a confined aquifer will result in greater drawdown near the well and a cone of depression that spreads faster and further than would result from an identical withdrawal rate in an unconfined aquifer.

4.1 Some Effects of Storage Depletion

Economic and Societal Benefits

Where groundwater storage depletion is large, even larger volumes of groundwater have been withdrawn and used. These uses provide water for drinking, for factories, and for agriculture, and therefore have great value. Such beneficial use and value must be carefully weighed when considering the economic and environmental effects and costs of the storage depletion. For example, substantial and continuing groundwater depletion might adversely impact the availability of water for irrigated agricultural productivity (e.g., CAST, 2019). Water managers and policy makers, of course, must also make their decisions in light of the legal framework applicable to their areas -- a framework that is highly variable from one political entity to another (even within a single nation).

Water-Level Declines

Water levels always decline in response to pumping. This is normal and not always a problem. However, if water-level declines in an aquifer or in an individual well are substantial, it can have detrimental physical and economic effects. Large water-level declines lead to reduced well yields and increased energy costs because of the greater lift required to move water from the well to the discharge point at or above the land surface. One reason that a greater lift results in a reduced well yield is because most pumps with a fixed power rating and capacity will produce a discharge of water in an inverse relation to the magnitude of the lift. For a flowing artesian well, the reduced water levels (or heads) will reduce the gradient driving the flow to the land surface, and hence reduce or eliminate the flow discharging from the well. In an unconfined aquifer, lowering the water table will also reduce the saturated thickness of the aquifer adjacent to the open or screened interval of the wellbore, and that will consequently reduce the effective transmissivity of the aquifer at the well, which in turn will reduce the well yield. In a confined aquifer, even as the potentiometric head drops, the aquifer will remain fully saturated, except in extreme circumstances. It is possible that the head can decline to a level below the elevation of the top of the aquifer, in which case the confined aquifer will start to drain and dewater. However, this is a rare occurrence.

If the water level in a well drops below the pump intake, the well will "go dry." This will have economic impacts on the well owner, who then must choose among several costly alternatives, including lowering the pump (if possible), deepening the existing well, drilling a new deeper well, abandoning the well and the former use of its produced water, or purchasing water (or water rights) and conveyance infrastructure for an alternate source of water supply if one is available and such transfers are legally allowable.

Land Subsidence

Groundwater storage depletion and the concomitant water-level declines can also have some poroelastic effects and consequences. The most common, widespread, and consequential is land subsidence, which can damage infrastructure and is widely recognized to be associated with groundwater use and storage depletion. Examples of affected areas include the Houston, Texas, area, the Central Valley of California, Mexico City, Bangkok, Tokyo, Jakarta, Venice, and other areas including those discussed by Galloway et al. (1999) and Poland (1984). The mechanisms whereby groundwater withdrawals can cause land subsidence are discussed in more detail in other Groundwater Project books. In summary, if the aquifers contain clayey lenses or layers, then lowering the heads can reduce the pore pressure in these materials, which in turn causes them to compress in an inelastic manner as the clay mineral structure itself realigns in a more compact (and irreversible) manner.

Sea-Level Rise

Most depleted groundwater ultimately finds its way into the oceans -- the ultimate sink. In a sense, groundwater depletion can be viewed as a large-scale, long-term transfer process of water from the continents to the oceans. If the long-term cumulative volume of depleted groundwater is large enough, it can contribute to sea-level rise, and there is good evidence that it has (e.g., Sahagian et al., 1994; Konikow, 2011; Church et al., 2011; Döll et al., 2014). The studies indicate that in the first decade of the 21st century, global groundwater depletion may have contributed 0.3 to 0.4 mm/yr to sea-level rise -- about 10 percent of the observed sea-level rise.

4.2 Methods to Estimate Depletion

Measuring or estimating the change in the volume of groundwater in storage over a time period is not straightforward as it cannot be measured directly. Estimations can rely on several alternative approaches, but all require the use of some unmeasured and uncertain properties and/or fluxes.

Head Changes

Perhaps the most direct approach to estimating depletion is to map the changes in head over the affected area of the aquifer and integrate those with estimates of the storage coefficient (or storativity; specific yield for unconfined aquifers). An example of this approach is presented for the High Plains aquifer as provided in Box 1↴. The storativity can vary horizontally and vertically, and in some cases, one may need to account for that variation to estimate reliable regional average values.

Models

This approach is inherent to the way that a numerical groundwater flow model simulates the hydrodynamic responses in an aquifer. As part of its numerical solution of the governing groundwater flow equation, a model that is calibrated to observed changes in head (and/or fluxes) in the aquifer will provide calculations of changes in groundwater storage during each model time step, as well as cumulatively for the time period from the start of the model simulation, in its water budget output. Examples of simulation models applied to aquifers having substantial depletion include Faunt et al. (2009) for the Central Valley of California and Clark and Hart (2009) for the Mississippi Embayment aquifer.

Water Budgets

One can also use a water budget approach for estimating depletion. For example, Kjelstrom (1995) used pumpage data in conjunction with other water budget estimates for the Snake River Plain aquifer in Idaho and eastern Oregon to estimate the changes in groundwater storage. But this approach has limited applicability because of the large uncertainties in estimates for all of the water budget elements. The difficulty arises because the change in storage may be small relative to the fluxes in and out of the aquifer so that errors in the estimated fluxes may exceed the magnitude of the rate of storage change. Nevertheless, a fair number of studies have been published using large scale (or even global) models of atmospheric and land-surface processes to estimate groundwater depletion, which is calculated as a residual in the budget equation (e.g., Wada et al., 2010). In its simplest form, depletion equals the difference between recharge and pumpage. Recharge is assumed to equal precipitation minus runoff and evapotranspiration, and pumpage estimates are typically based on land-use characterization and expected water use rather than on direct measurements of withdrawals. A primary difficulty here is that these approaches generally cannot simulate or predict the very effects and processes discussed in this book -- namely that pumpage will be balanced by a combination of increased recharge and decreased discharge -- because the water budget approaches do not simulate the hydrodynamic changes within a groundwater flow system nor its relation to surface water. Therefore, these global water-balance approaches tend to substantially overestimate the magnitude of groundwater storage depletion because they erroneously ignore capture and/or its hydrodynamic processes.

Gravity Measurements

Another alternative approach to estimating groundwater depletion is through geophysical methods. Where groundwater depletion occurs, the mass of material in the Earth's subsurface is reduced, and this affects the Earth's gravity field. Very sensitive gravity measurements repeated over time can detect relatively small changes in the gravity field. If these gravity changes are derived solely or primarily from changes in total water storage (TWS), then the observed gravity changes can be used to estimate groundwater depletion if other possible contributors to changes in TWS are negligible or can be reliably estimated (such as changes in surface-water storage, snow and ice, and vadose zone/soil moisture). Land-based surface gravity measurements were repeated annually over several years in Arizona to estimate changes in groundwater storage in alluvial aquifers (Pool and Anderson, 2008). However, they reported that correlations between gravity-based estimates of storage change and water-level changes in observation wells were sometimes only poor to moderate.

GRACE Remote Sensing

A much larger scale of gravity measurement is provided by the GRACE satellites (Gravity Recovery and Climate Experiment, a mission jointly sponsored by the United States and Germany), which are a pair of coupled satellites to measure spatial and temporal changes in the Earth's gravity field (e.g., Tapley et al., 2004; Famiglietti and Rodell, 2013; and Famiglietti et al., 2015). GRACE has provided useful information about global changes in TWS, as well as global groundwater depletion for many large aquifer systems (e.g., Famiglietti et al., 2011; Tiwari et al., 2009; and Rodell et al., 2009). However, GRACE data provide estimates of total change in water storage over a relatively large footprint -- a resolution on the order of 100,000 km^2 as discussed by Scanlon et al. (2016). This scale is much larger than many aquifers. Although the accuracy is on the order of 1.5 cm of equivalent water height, the low spatial resolution of GRACE limits its ability to provide groundwater depletion data at a scale amenable for water managers to use effectively as discussed by Alley and Konikow (2015). Furthermore, the analysis of GRACE data still faces the challenge of separating all the components contributing to total water storage change as explained by Scanlon et al. (2015). Hence, using this approach to estimate groundwater depletion is most applicable to large aquifers in arid to semi-arid climatic areas and requires some caution in interpretation.

Subsidence

An estimate of the minimum value for groundwater storage depletion in an area undergoing land subsidence due to groundwater withdrawals can be made by calculating the volume of subsidence. The groundwater storage depletion volume must equal or exceed the subsidence volume because the removal of pore water and the subsequent compaction of the sediments are the drivers for the subsidence. For example, in the Gulf Coastal Plain near Houston, Texas, the volume of land subsidence was estimated by

Konikow (2013) using maps of historical subsidence (1906–2000). The calculated cumulative subsidence volume was 10.5 km^3. For comparison, the cumulative water budget from a numerical model calibrated to field observations made from 1891 to 2000 indicates that 10.8 km^3 of groundwater was removed from storage in the unconsolidated clay units as the clays compacted and subsidence progressed -- essentially all of it during the 20th century according to Kasmarek and Robinson (2004). The small difference of less than 3 percent provides good support for the use of the subsidence approach, as well as for the quality of the model calibration and resulting reliability of the model calculations. For comparison, the volume of groundwater depletion derived from storage losses in interbedded clays (and associated with land subsidence) represents about 36 percent of the total groundwater storage depletion of 28.9 km^3, with the remaining depletion derived from storage losses in the sand layers (Kasmarek and Robinson, 2004).

Confining Layers

It is well established that confining layer storage is a significant source of water when confined aquifers are developed (e.g., Theis, 1940; Jacob, 1946; Hantush, 1960; Bredehoeft et al., 1983). In a regionally extensive confined aquifer, direct recharge may be limited to outcrop areas at the margins of the aquifer's extent. So a stress on the aquifer in the form of groundwater withdrawal from wells distant from the outcrop area cannot be readily balanced by an increase in recharge. Thus, drawdown propagates large distances laterally and changes in vertical hydraulic gradients induce leakage from adjacent confining beds. Because the hydraulic conductivity of confining beds is low relative to that of the aquifer, precluding timely propagation of the head change to the other side of confining bed, that leakage will be derived primarily from storage depletion in the confining beds over timeframes of decades to centuries. In fact, the magnitude of storage depletion in the confining units can be much greater than the storage depletion in the confined aquifer itself. An example of this occurred in the Dakota Aquifer system as explained in Box 2. Konikow and Neuzil (2007) summarize a number of approaches for estimating the volume of depletion from confining units. Most methods require measurements or estimates of the hydraulic conductivity and specific storage properties of the confining layer, as well as observations of head changes within the confining layer. However, such data are rarely available because water supply wells rarely have open intervals in confining units. Konikow and Neuzil (2007) offer a simplified method that is based on head changes in the confined aquifer at the boundary with low-permeability confining units.

4.3 Magnitude of Storage Depletion

Groundwater storage depletion is becoming recognized as an increasingly serious global problem that threatens sustainability of water supplies (e.g., Schwartz and Ibaraki, 2011). Long-term cumulative groundwater storage depletion, both in the United States and globally, was estimated by Konikow (2011; 2013) using calibrated groundwater models, analytical approaches, gravity-based analyses, and/or volumetric budget analyses for multiple aquifer systems. Estimates were derived from bringing together information from the literature and from new analyses.

Long-term cumulative depletion volumes in the United States were assessed for 40 separate aquifers or areas as shown in Figure 5 and one broader land-use category (agricultural and land drainage where the water table has been permanently lowered).

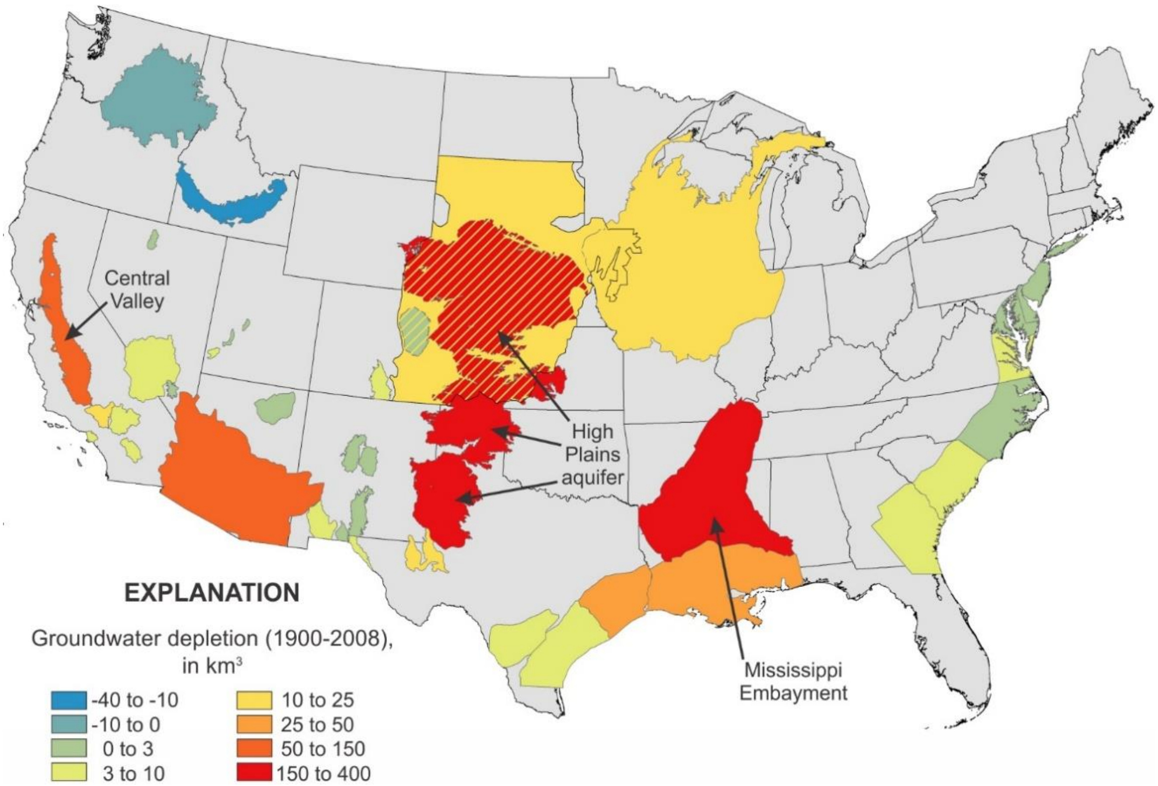

Figure 5 - Cumulative long-term volumetric groundwater storage depletion in the United States during 1900-2008, in km^3 (modified from Konikow, 2013). The three aquifers with the highest depletion volumes are identified. Hachured areas are where a shallow aquifer overlies a deeper aquifer.

Estimated total groundwater depletion in the United States during the 20th century was approximately 800 km^3, increasing by about 25 percent during the next 8 years for a total of 1,000 km^3 during 1900-2008. The rate of groundwater depletion has increased markedly since about 1950 (Figure 6), with maximum rates occurring during the recent period (2001–2008) when the depletion rate averaged almost 25 km^3 per year (compared to 9.2 km^3 per year average for the 1900-2008 timeframe). Two large aquifer systems in the northwestern United States showed long-term negative depletions, in other words, water-table rises thus increased groundwater storage. This is attributable primarily to the diversion and application of surface water for irrigation purposes, which increases recharge above rates that occur naturally.

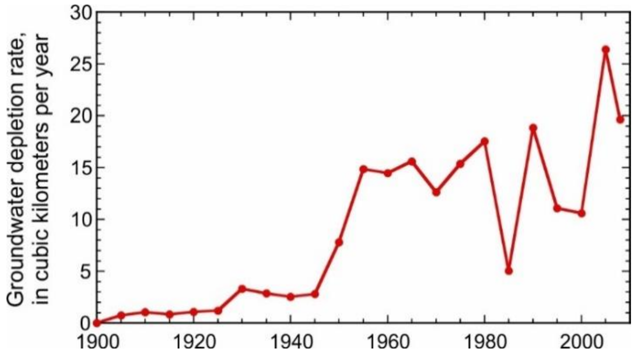

Figure 6 – Five-year averaged rates of groundwater depletion in the United States, 1900 through 2008. Final value represents average rate during 3-year period, 2006 through 2008 (from Konikow, 2015).

Depletion volumes for individual aquifer systems vary substantially across the United States (Figure 5). The three aquifer systems having the largest depletion volumes are the High Plains aquifer (341 km^3), the Mississippi Embayment aquifer system (182 km^3), and the Central Valley of California (145 km^3). However, this does not tell the whole story. For example, because the High Plains aquifer encompasses a very large area (~450,000 km^2), the average depletion over the entire area is less than in some other systems, and depletion is nonuniform - being much greater in its southern part. Another way to assess the magnitude of aquifer depletion is to normalize the depletion volume by the area of the aquifer, yielding a measure of *depletion intensity* (Konikow, 2015). During the 20th century, the highest depletion intensities occurred in three relatively small basins in southern California. However, during the beginning of the 21st century, the highest depletion intensity was in the Central Valley of California, which has an area of about 52,000 km^2. The aquifer-wide depletion intensity there averaged 0.075m/yr during 2001-2008 (Konikow, 2015). Consistent with this measure, the Central Valley has been experiencing increasing water shortages, additional water-level declines, and accelerated land subsidence in the most affected parts of the valley since 2000 (Faunt et al., 2016).

Groundwater storage depletion and capture can be measured in terms of nondimensional fractions relative to pumpage (Konikow and Leake, 2014). Reliable estimates of cumulative pumpage in the United States are available for the United States as a whole for 1950-2005, during which the cumulative withdrawals were approximately 5,340 km^3 (Kenny et al., 2009). During that same time period, the total net groundwater storage depletion was about 812 km^3 (Konikow, 2013). Thus, about 15 percent of the total pumpage was derived from a reduction in the volume of groundwater in storage; that is, the long-term depletion fraction is about 0.15 and the capture fraction is about 0.85. But depletion fractions vary widely throughout the United States. There are adequate withdrawal and depletion data available for 31 specific areas or aquifers within the United States. The mean depletion fraction for these areas is 0.39 and the mean capture fraction is 0.61 (Konikow and Leake, 2014). Overall, even though groundwater storage depletion is a serious problem in many places, it is evident that over periods of years to decades that capture is generally larger than depletion and constitutes an even more serious concern.

To assess the potential contribution of groundwater depletion to sea-level rise, one can make a bounding calculation by assuming that the oceans represent an ultimate sink for essentially all depleted groundwater. Then the contribution of groundwater storage depletion to sea-level rise can be estimated by spreading the volume of depletion across the surface area of the oceans (a total area of about 3.61×10^8 km^2). On this basis, groundwater depletion in the United States alone would account for (or balance) as much as 2.2 mm of sea-level rise during the 20th century. During this 100-year period, the observed rate of sea-level rise averaged about 1.7 mm/yr. Thus, groundwater depletion in the United States alone can explain about 1.3 percent of the observed global sea-level rise during the 20th century.

Groundwater storage depletion is a global problem. Konikow and Kendy (2005) note that excessive groundwater depletion affects major regions of North Africa, the Middle East, South and Central Asia, North China, North America, and Australia, as well as localized areas throughout the world. Konikow (2011) estimated the cumulative global groundwater depletion, as well as its equivalent sea-level rise for 1900-2008 (Figure 7). Obtaining data on water-level changes from many parts of the world is extremely difficult, so the resulting depletion estimates have a greater uncertainty than those for just the United States. The total depletion volume is on the order of 4,500 km^3, which could explain about 12.6 mm of sea-level rise. However, during the most recent part of this evaluated time period (2001-2008), the rate of global groundwater depletion had increased from an average of 33.7 km^3/yr during the 20th century to approximately 145 km^3/yr (equivalent to a sea-level rise contribution of 0.40 mm/yr). For these same reference time periods, the rate of sea-level rise had increased from about 1.7 mm/yr to about 3.1 mm/yr. Therefore, during the first part of the 21st century, global groundwater depletion can explain almost 13 percent of the observed rate of sea-level rise.

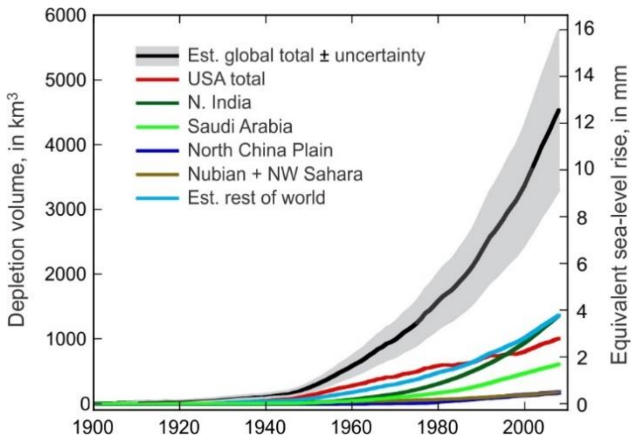

Figure 7 - Estimated cumulative global groundwater storage depletion, 1900-2008, and equivalent contribution to sea-level rise (from Konikow, 2011).

5 CAPTURE

The second mechanism that contributes to balancing of groundwater pumpage is capture - a combination of increased recharge and decreased discharge of groundwater (Lohman et al., 1972; Bredehoeft and Durbin, 2009; Leake, 2011; Barlow and Leake, 2012; and Barlow et al., 2018).

The decrease in groundwater discharge typically represents a reduction in the discharge to surface water, such as streams, lakes, wetlands, springs, drains, estuaries, and/or along coastlines. However, evapotranspiration is also a type of discharge from a groundwater system that also can be captured. In fact, in areas where surface water is sparse or absent, as in closed desert basins, the reduction in groundwater discharge (and capture) can primarily encompass a reduction in evapotranspirative losses from the groundwater system as the water table declines over time.

Capture also includes increases in recharge in response to pumping and water-level declines. For example, if a lowland area has a water table that lies at or immediately below the land surface, precipitation falling onto that surface cannot infiltrate to recharge the aquifer because no pore space is available to absorb it -- all the pore spaces are already fully saturated. This potential recharge is “rejected.” However, if the water table drops because of pumping, then future precipitation onto this same land surface area will be able to infiltrate the soil and recharge the groundwater system. Also, increases in recharge in response to pumping can occur where surface water features intersect aquifers. If drawdown due to pumping is so large as to reverse the hydraulic gradient, then where groundwater flow was formerly directed from the aquifer to the stream it now becomes directed from the stream into the aquifer. This seepage loss from the stream is usually termed “induced infiltration.”

But in general, capture typically is composed mostly of streamflow (or other surface water) depletion. In the United States, groundwater discharge represents from 15 to 90 percent of total annual streamflow -- about 50 percent on average (Winter et al., 1998). Thus, any reduction in groundwater discharge supporting streamflow can have serious detrimental consequences. Streamflow depletion is most often observed as a reduction in the base flow (or low flows) of streams. In the extreme, streams can go dry (Figure 8). Such streamflow depletion is of great concern for water-supply managers, for those with senior rights to use surface water, and because of potential environmental impacts. In fact, streamflow depletion due to groundwater pumping has been the subject of several United States Supreme Court cases in recent years, whose decisions have clearly recognized the relation between groundwater and surface water (Alley and Alley, 2017).

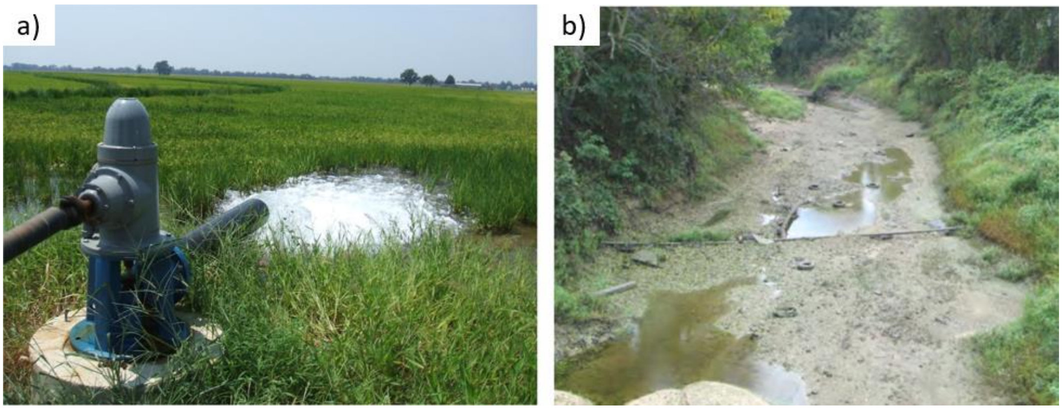

Figure 8 - There is a relation between groundwater pumping and streamflow depletion. a) Groundwater pumped from the Mississippi River alluvial aquifer for flood irrigation of a rice field in the Mississippi Delta, USA (Photograph by David E. Burt, Jr., United States Geological Survey; *source:* Barlow and Leake, 2012). b) A Delta stream (Big Sunflower River) that is nearly dry during the summer because of loss of base flow (Photograph by Matt Hicks, United States Geological Survey; *source:* Barlow and Clark, 2011).

5.1 Streamflow Depletion

Streamflow depletion arises from both a reduction in the discharge of groundwater into a stream and an increase in recharge from the stream to the aquifer as gradients are reduced or reversed between the aquifer and the stream. Both constitute capture, and both result in a depletion of streamflow downstream from the capture area.

An idealized representation of the sequence of changes for stream-aquifer interaction and its response to pumping is depicted in Figure 9. The simplified schematic illustration shows that under natural (predevelopment) conditions (Figure 9a) the average recharge to the saturated zone equals the average discharge to the stream (assuming no evapotranspiration from the water table and ignoring any short-term variations in precipitation that may affect recharge rates or stream stage). After a well is drilled and begins pumping (Figure 9b), a cone of depression develops and water is removed from storage. The drawdown reduces the gradient towards the stream and flow from the aquifer to the stream (i.e., groundwater discharge) is decreased. After more time, in some cases the hydraulic gradient at the stream-aquifer boundary may be reversed, which locally reverses the flow direction and induces water in the stream to flow into the aquifer, thereby increasing recharge (Figure 9c). When pumping becomes fully balanced by capture, the heads will have stabilized and there will be no additional drawdown or removal of water from storage. The time it takes for this to occur is called the "time to full capture" (Bredehoeft and Durbin, 2009). At this time, the system will have attained a new equilibrium condition, and the pumping rate will be sustainable (or maintainable) from a hydraulic perspective. However, the streamflow depletion may have detrimental effects on downstream users and ecosystems, and may not be acceptable. If the pumping is turned off (Figure 9d), this sequence is reversed and groundwater levels begin to recover, and groundwater discharge to the stream increases. Given sufficient time, groundwater heads may return to their original levels, and recharge and discharge in the system will again achieve a long-term equilibrium condition (Figure 9e).

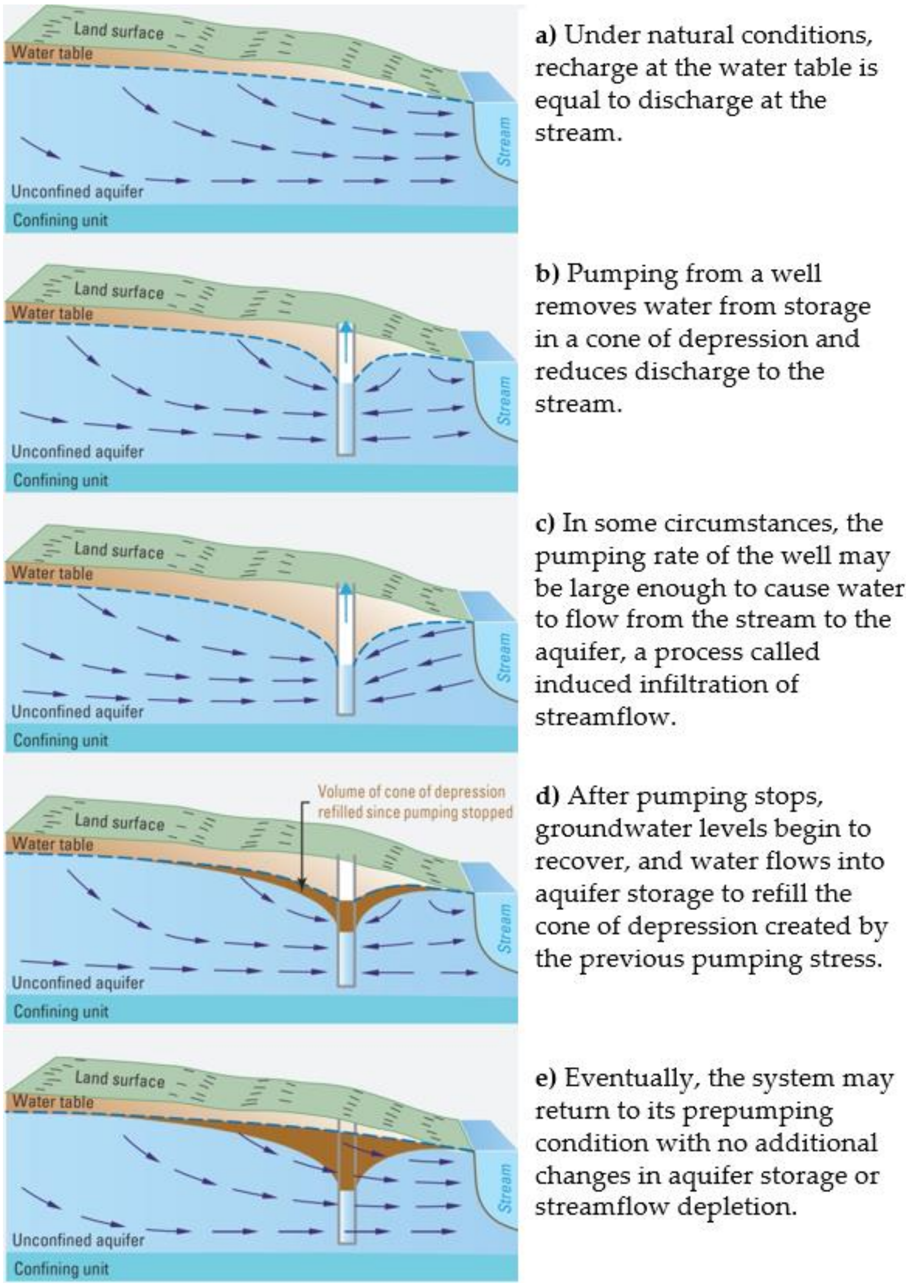

Figure 9. Effects of pumping groundwater from a hypothetical water-table aquifer that discharges to a stream. Sequence shows progressive changes to groundwater flow and streamflow before, during, and after pumping at a hypothetical well site (from Leake and Barlow, 2013; as modified from Heath, 1983; Alley et al., 1999).

The capture and streamflow depletion can be manifested in several ways. During low-flow periods in a stream, the groundwater discharge constitutes a larger fraction of the total streamflow than during high-flow periods, so a given reduction in groundwater discharge would be more easily detectable. So, where groundwater pumping and storage depletion is affecting streamflow, we would expect to see the clearest signal in the low-flow records for the stream. For example, this is indeed evident in the records of a stream gage on the Sunflower River, Mississippi, USA, located within the area of the heavily pumped Mississippi Embayment regional aquifer system (Figure 10). The data show a significant decline in the minimum daily mean flow for each year shortly after groundwater use and storage depletion noticeably accelerated.

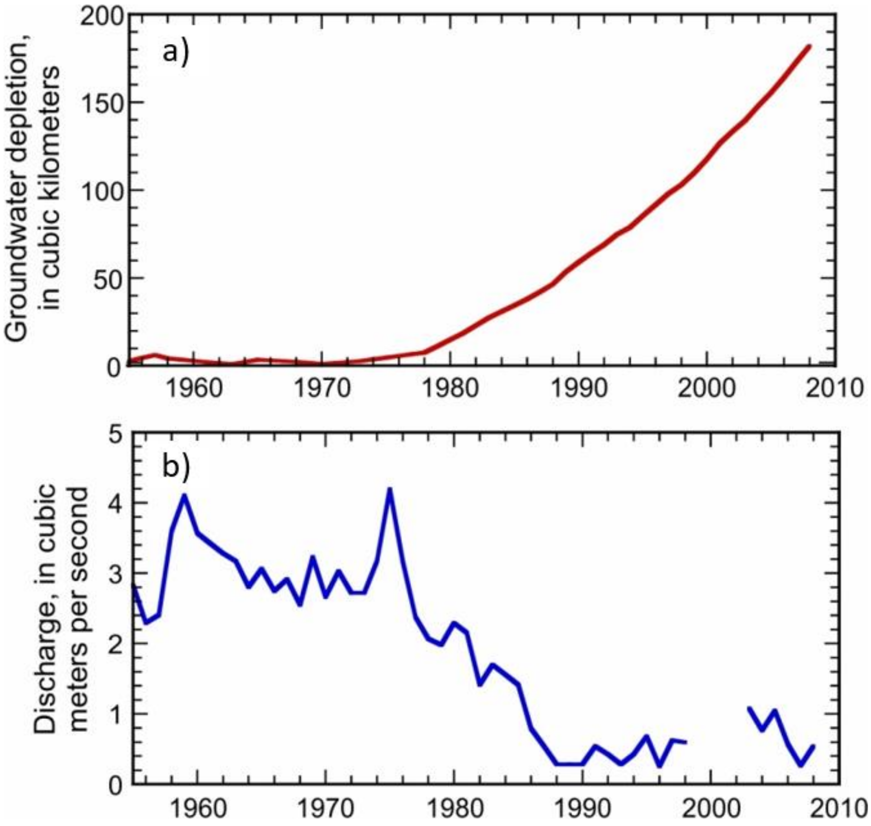

Figure 10 - Relation between groundwater depletion and stream discharge. a) Cumulative groundwater depletion in the Mississippi Embayment regional aquifer system, USA, 1955 through 2008 (Konikow, 2013). b) Annual minimum mean daily streamflow for the Big Sunflower River, Mississippi, USA, showing effects of withdrawals from the aquifer on base flow of the stream and indicating streamflow depletion after the late 1970s; data are missing for 1998-2002 (modified from Welch et al., 2010).

Another type of low-flow characteristic is how often (or for how long) a stream goes dry. If streamflow depletion is affecting the flow of the stream, then the frequency of days during which there is no flow in the stream might increase or the lengths of dry stream reaches might increase. The Cache River in northeastern Arkansas, USA, also lies within the boundaries of the Mississippi Embayment aquifer, and the number of zero-flow days first increased after 1980, shortly after groundwater depletion had increased (Figure 11). Another aspect of this phenomenon is that stream reaches that were perennially flowing prior to groundwater development can go intermittently dry so that those stream reaches are no longer considered to be perennial, as seen in western Kansas (Figure 12).

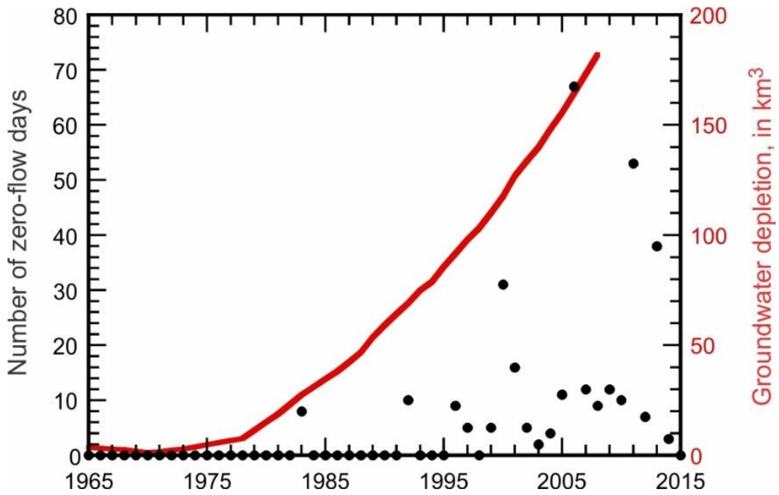

Figure 11 - The number of days during a year (1965-2015) that the Cache River at Egypt, Arkansas, USA, is dry (shown as black dots) increased markedly after cumulative groundwater depletion (red curve) in the Mississippi Embayment regional aquifer system became significant.

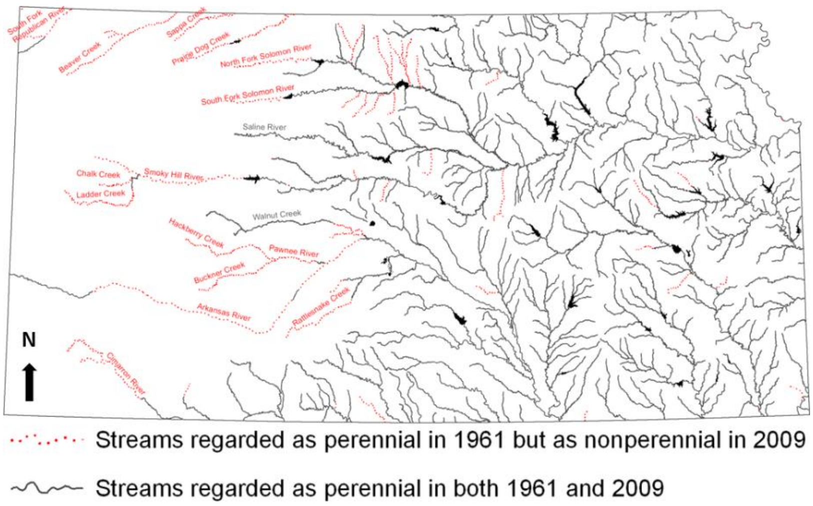

Figure 12 - Map showing major perennial streams in Kansas as of 1961 and 2009. In western Kansas, which is underlain by areas of the High Plains aquifer that have undergone substantial groundwater-level declines and storage depletion since the 1950s, many streams or stream reaches that had been considered perennial in 1961 were no longer so in 2009 (modified from Kansas Department of Agriculture, 2010).

5.2 Capture of Evapotranspiration

Groundwater evapotranspiration includes both evaporation from the water table and transpiration through plant roots that tap the uppermost part of the saturated zone. Evaporative losses from the water table involves a flux through the unsaturated zone to the atmosphere. This flux will be largest if the distance from the water table to the soil surface is shortest (that is, the water table is at or immediately below the land surface). It is also expected that there exists a depth below which evaporation from the water table becomes negligible. Similarly, transpiration from the saturated zone by phreatophytes depends on root penetration, which is most extensive when the water table is shallowest. Phreatophytes are plants that depend for their water supply upon groundwater that lies within reach of their roots (Robinson, 1958). However, there is a maximum depth that plant roots can penetrate. The limiting depth of the water table -- below which no evapotranspiration can occur -- is called the extinction depth or cutoff depth (McDonald and Harbaugh, 1988).

Overall, the groundwater evapotranspiration flux is inversely proportional to the depth of the water table below the land surface. Thus, as the water table declines in response to pumping and storage depletion, the potential evapotranspiration will decrease (i.e., it is captured). An extreme example of this effect is illustrated by photos of a riparian zone of a stream in an arid climate, where a long-term loss of vegetation due to large water-table declines reflects a capture of evapotranspiration (Figure 13).

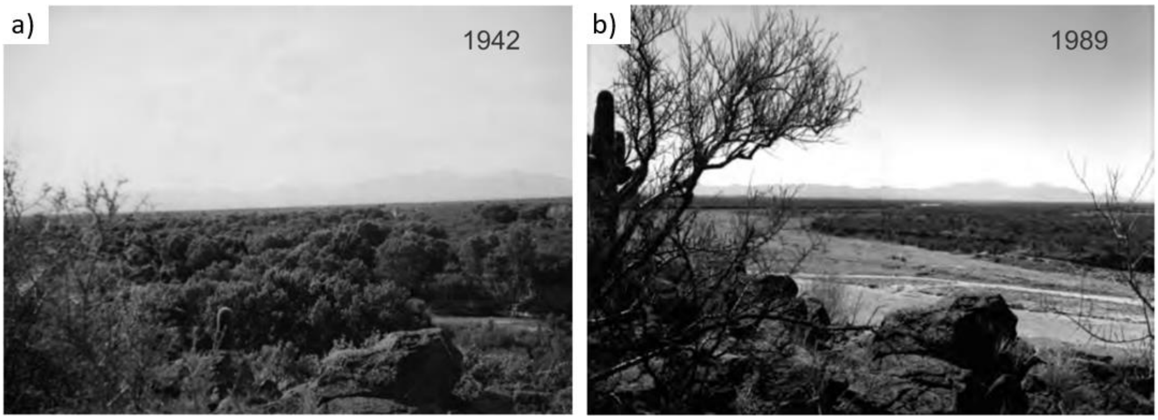

Figure 13 - Loss of riparian vegetation where the water table declined: a) 1942 photograph of a reach of the Santa Cruz River south of Tucson, Arizona, showing stands of mesquite and cottonwood trees growing in the riparian zone of the river (left photograph, Arizona Game and Fish Department); b) photograph of the same site in 1989 showing that the riparian vegetation has largely disappeared (right photograph, R.H. Webb, USGS). Data from two nearby wells indicate that the water table has declined more than 30 m because of pumping; this pumping (and its consequences) appears to be the principal reason for the vegetation loss (Healy et al., 2007).

The potential for salvaging groundwater evapotranspiration to help balance groundwater pumping was recognized by Theis (1940). With the goal of optimizing well locations and minimizing drawdown and streamflow depletion, Theis (1940) stated "pumps should be placed as close as economically possible to areas of … natural discharge where ground water is being lost by evaporation or transpiration by non-productive vegetation." A small decrease in the rate of groundwater evapotranspiration over a large area can yield a large volume of water.

5.3 Capture of Spring Discharge

Spring discharge is in part a function of local and regional hydraulic gradients. The spring elevation is fixed, but the distal heads can decline due to pumping, which reduces the head gradient towards the spring. This will reduce the flow towards the spring, which reduces its discharge -- exactly the same mechanism that causes streamflow depletion. This effect has been observed in long-term discharge measurements at regional springs in southern Nevada, USA (Figure 14), for example, as well as in springs in North Africa (Margat et al., 2006). It is not uncommon for springs to go dry because of groundwater development in the spring's source area. But dry springs can recover. Manse Spring (shown in Figure 14), having gone essentially dry in 1977 (although small intermittent winter flows were reported after that), started to flow again in the late 1990s (San Juan et al., 2010). The recovery of spring flow was in response to reduced pumping and rising groundwater levels since 1980, and discharge was reported to be about 1.9×10^6 m^3/yr in 2011 (Halford and Jackson, 2020).

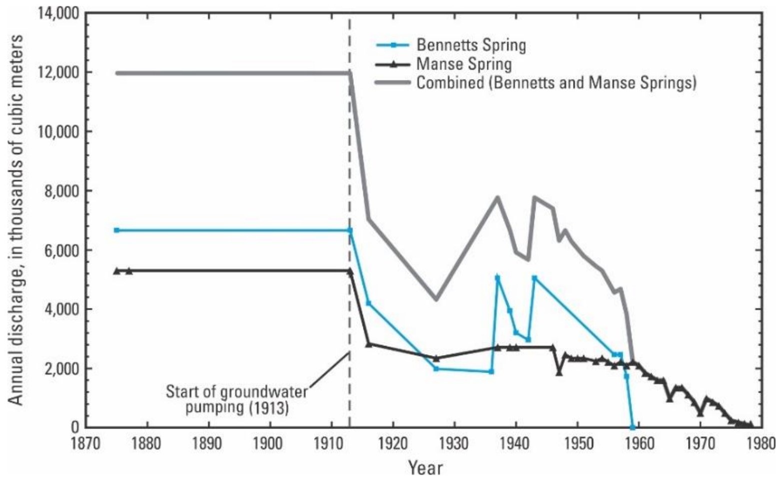

Figure 14 - Annual discharge from regional springs in Pahrump Valley, Nevada, 1875-1978, showing trend of decreasing spring flow after onset of pumping (modified from San Juan et al., 2010).

5.4 Estimating the Magnitude and Timing of Streamflow Depletion

The factors controlling the timing of streamflow depletion by groundwater capture are essentially the same factors that control the response of an aquifer to a pumping well. These include the geometry, dimensions, and hydraulic properties of the aquifer; the locations and hydraulic characteristics of aquifer boundaries, including streams; and the horizontal and vertical distances between wells and streams (Barlow and Leake, 2012). Barlow and Leake (2012) note that the two most important factors are the distance between a pumping well and a nearby stream and the hydraulic diffusivity of the aquifer (which is defined as the ratio of the transmissivity to storage coefficient). The magnitude of the streamflow depletion will be proportional to the magnitude of the rate of pumping, although the timing is not affected by the pumping rate.

Perhaps the best way to predict the magnitude and timing of streamflow depletion in response to pumping is to develop and calibrate a reliable simulation model that accurately (or adequately) represents all of the governing factors in a stream-aquifer system. Numerical groundwater modeling software, such as the public-domain MODFLOW software↗, offer a framework for accomplishing this. Numerical models can readily account for heterogeneities and irregular or nonlinear boundaries. However, if certain simplifying assumptions about the system can reasonably be made, then it is possible to apply an analytical solution, which can offer a faster way to make an estimate (e.g., Glover and Balmer, 1954; Theis and Conover, 1963). Commonly required assumptions include (but are not limited to): (1) a semi-infinite, homogeneous, isotropic aquifer, (2) aquifer transmissivity remains constant with time, and (3) the bounding stream is a straight line and fully penetrates the aquifer. As summarized by Barlow and Leake (2012), Glover's analytical solution allows one to compute the total rate of streamflow depletion with time, $Q_s(t)$, as shown in Equation 5.

$$Q_s(t) = Q_w\, erfc\,(z) \tag{5}$$

where:

$Q_s(t)$ = total rate of streamflow depletion with time (L³/T)

Q_w = pumping rate of the well (L³/T)

$erfc$ = complementary error function (dimensionless)

z = $\sqrt{(a^2S)/(4Tt)}$ (dimensionless)

a = distance from the well to the stream (L)

S = storage coefficient (dimensionless)

T = transmissivity (L²/T)

t = time (T)

To simplify the mathematical complexities of the analytical solutions, Jenkins (1968) used a semi-analytical approach in which he introduced the concept of a stream depletion factor (sdf), which had units of time and was defined as Equation 6.

$$sdf = a^2/D \tag{6}$$

where:

D = hydraulic diffusivity, $D = T/S$ (L²/T)

As described by Barlow and Leake (2012), the value of sdf for a given pumping location is a relative measure of how fast streamflow depletion will occur in response to new pumping. A high value of D will result in a relatively low value of sdf and a relatively fast response of streamflow depletion to pumping. Values of sdf at every location in an aquifer can be calculated using a numerical model and then mapped for use by water managers to readily assess and compare potential impacts of new wells on streamflow for various well locations.

As an illustrative example of streamflow depletion dynamics, we can look at northeastern Arizona, USA, where base flow of streams is maintained in part by groundwater discharge from the C aquifer. Leake et al. (2005) developed a numerical model of this aquifer to help assess the possible effects of the proposed withdrawals from the C aquifer. They calculated potential streamflow depletion for two possible pumping scenarios (Figure 15). The two withdrawal scenarios were simulated for a 51-year period of withdrawals followed by 50 years during which there were no withdrawals. Scenario A included a nearly constant withdrawal rate of about 9.0 ft³/s (0.25 m³/s), and scenario B simulated a more variable pumping rate with a maximum withdrawal rate of about 15.9

ft³/s (0.45 m³/s). The results show that streamflow depletion was undetectable for the first few years after withdrawals began, but then increased steadily during the rest of the withdrawal period. However, the depletion rate at the end of the withdrawal period was between 0.3 and 0.4 ft³/s (0.008 and 0.011 m³/s) for both scenarios -- substantially less than the withdrawal rates, largely because of the large distances of 20 miles or more between the wells and the streams.

An important lesson in these results is that the streamflow depletion continued to grow for decades after the pumping had stopped (Figure 15). These lag times complicate the management of water resources in stream-aquifer systems.

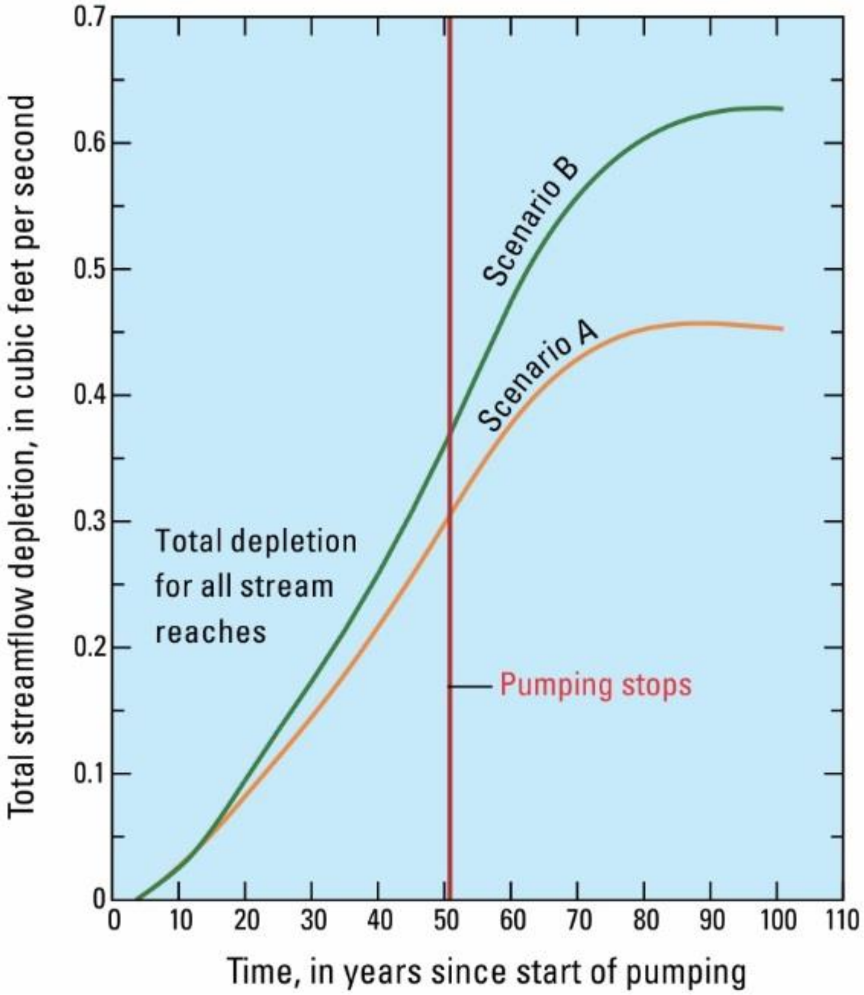

Figure 15 - Streamflow depletion as a function of time for two scenarios of groundwater pumping from the C aquifer, northeastern Arizona, USA (from Barlow and Leake, 2012, as modified from Leake et al., 2005).

5.5 Methods to Estimate Capture

For real-world problems in typically complex hydrogeologic environments, numerical models "… are the only approach to compute capture from different features" (Barlow and Leake, 2012). Models are widely used tools in groundwater analysis. The beauty of the model is that it can be used to project how a particular system might respond to different stresses in the future. There are too many unknowns and uncertainties in the problem to predict accurately and uniquely a future result. On the other hand, one can project how a system might respond, or a range of responses, and at the same time, given potential errors in input data, place some confidence interval about a future projection. This is useful in attempting to understand and manage the system. In other words, one can ask: if we do this, what is the resulting projected future state of the system.

Barlow and Leake (2012) provide an example of such an analysis for the Upper San Pedro Basin aquifer system in southern Arizona, USA, which was studied by Leake et al. (2008). Using this model, which included a representation of the evapotranspiration process, they assessed the response of the system to pumping a hypothetical well at various locations. The results for one such well are plotted in Figure 16 and show the shifting tradeoff over time between groundwater storage change and capture as sources of water to balance the pumpage. Furthermore, it illustrates that salvaged evapotranspiration can be a substantial component of the total capture. The streamflow depletion includes both induced infiltration (increased recharge to the aquifer) and decreased discharge of groundwater to the stream, though these are not shown separately in the plot (though the typical model output will include sufficient information to allow the user to do this).

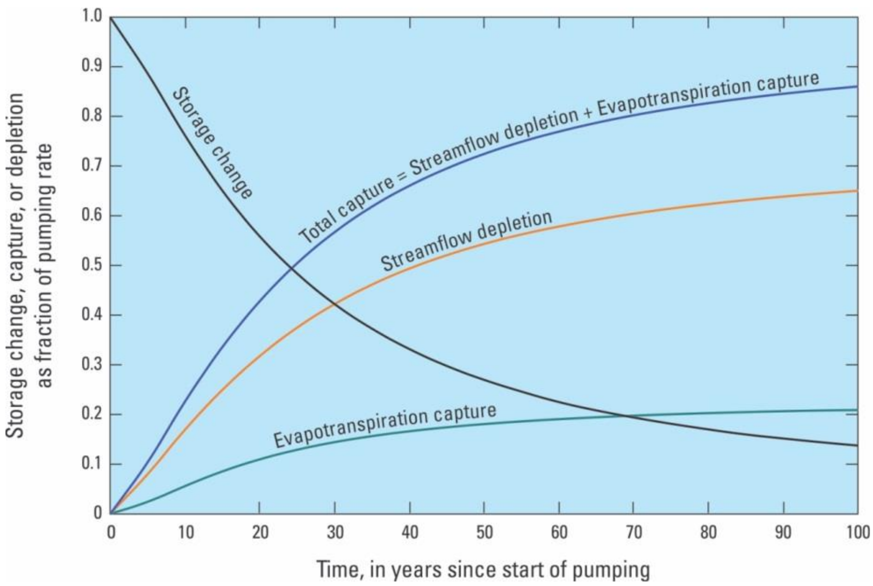

Figure 16 – Model-computed streamflow depletion, evapotranspiration capture, storage change, and total capture for the location of one hypothetical well that pumps for 100 years in the Upper San Pedro Basin, Arizona (from Barlow and Leake, 2012; after Leake et al., 2008).

6 CASE STUDIES ILLUSTRATING GROUNDWATER DEVELOPMENT DYNAMICS

In an effort to illustrate these principles it is useful to examine both a simple hypothetical aquifer system as well as a groundwater development case. Development of a simulation model for a simplified hypothetical problem offers the advantage of having control over the system and complete knowledge and definition of hydraulic properties and boundary conditions. For an illustrative example of a well-documented field problem, we look to the United States Geological Survey (USGS) Regional Aquifer System Analyses (RASA) studies. As part of one RASA study, an agricultural development in Paradise Valley, Nevada, USA, was studied in detail. We review the highlights and lessons of this study, which illustrate how the principles of development are applied to a complex system.

6.1 Case Study 1: Hypothetical Stream-Aquifer System

Description of Problem

When a hypothetical system is stressed in a simulation model, the responses can be analyzed and interpreted without the uncertainty normally associated with complex field systems. Furthermore, additional factors or complexities can be added in individual increments, thereby simplifying the analysis of cause and effect. Thus, the links and relations between stresses and responses (causes and effects) can be more clearly and definitively identified.

For this purpose, we identify a hypothetical desert basin with a perennial stream along one edge of the valley, based on one designed and used by Barlow and Leake (2012) to illustrate the effects of pumping on streamflow. We modify their example problem, similarly to the modifications of Konikow and Leake (2014), to illustrate additional effects of pumping a well on the hydrology of such a stream-aquifer system (Figure 17). Among other changes, we add more realistic and variable land-surface elevations that will induce spatially varying evapotranspiration (ET) losses, which vary as a function of the depth to the water table. ET consists of both direct evaporation from the water table surface and transpiration by plants. Substantial groundwater use by phreatophytes is evidenced by diurnal fluctuations of the water table (e.g., Butler et al., 2007). Groundwater evaporation is greatest when the water table is at or very close to the land surface. As the water table becomes deeper, the transport distance for water vapor between the water table and the atmosphere increases and dissipation of humidity in the soil is hindered, hence the evaporative flux from the saturated zone decreases. Transpiration is generally the larger component of ET. As the depth to the water table increases, the percentage of plant roots that are sufficiently long and deep to penetrate the water table tends to decrease, and the greater the energy requirement needed to lift water over the larger distance to the land surface. At some point, the water table may be so deep that no roots can reach it.

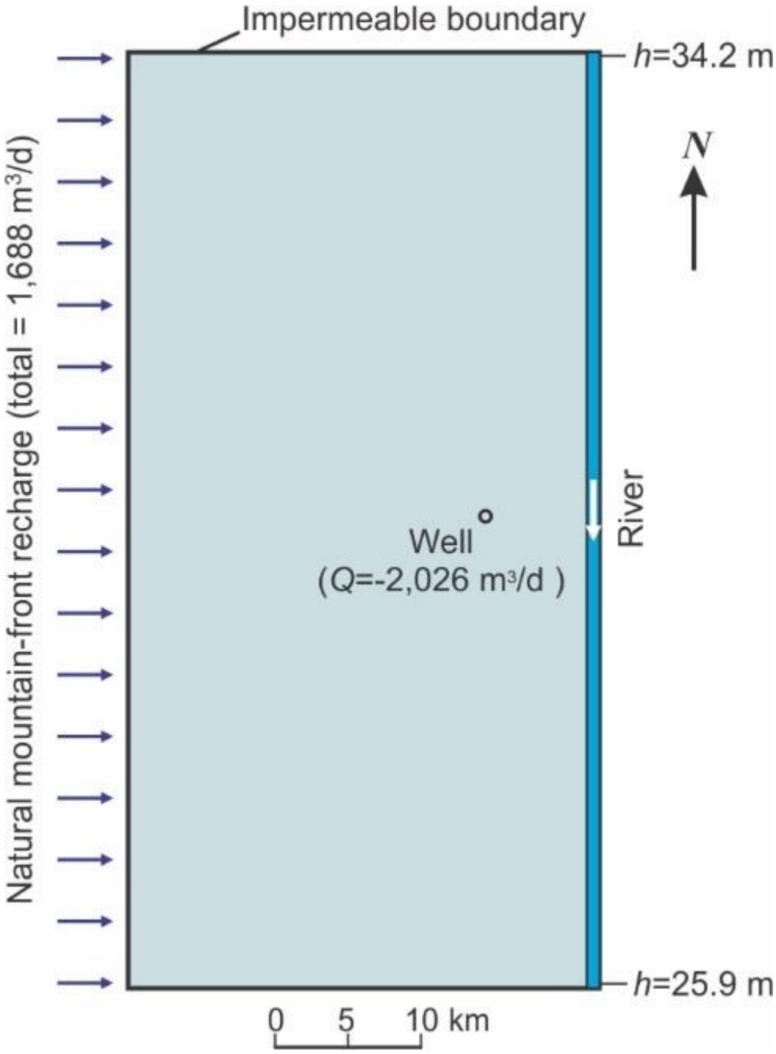

Figure 17 - Hypothetical and idealized desert basin aquifer with a perennial river along its eastern side and mountain front recharge along its western side (modified from Barlow and Leake, 2012).

The hypothetical aquifer is 32.2 km (20 mi) wide by 64.4 km (40 mi) long, with a total area of 2,072 km^2 (800 mi^2). The river is in good hydraulic connection with the alluvial aquifer; the specified flow into the upstream end of the river is 20,000 m^3/d. It is assumed, for simplicity, that there is neither direct precipitation on the river nor evaporation from the river, and that the inflow to the upstream end of the river is constant. Mountain front recharge totaling 1,688 m^3/d is uniformly distributed along the western boundary of the basin. The hydraulic conductivity of the aquifer is 15.2 m/d and its thickness averages about 150 m, which is much smaller than its lateral extent. The specific yield is 0.20. The rectangular aquifer is surrounded by impermeable boundaries, including along the distal side of the river. There is one fully penetrating well located near the center of the system, at a distance of 8.05 km (5 mi) from the river. It pumps at a rate of 2,026 m^3/d (0.83 ft^3/s).

The land surface is represented as having three terrace levels above the river (Figure 18). The elevation differences between adjacent terraces is about 1 m, with the highest elevations being the furthest from the river. Note that the land-surface elevation is not assumed to coincide with either the streambed elevation or the water surface in the river. Rather, the head in the river is assumed to vary linearly between the upstream and downstream heads shown in Figure 17 and to remain constant in time.

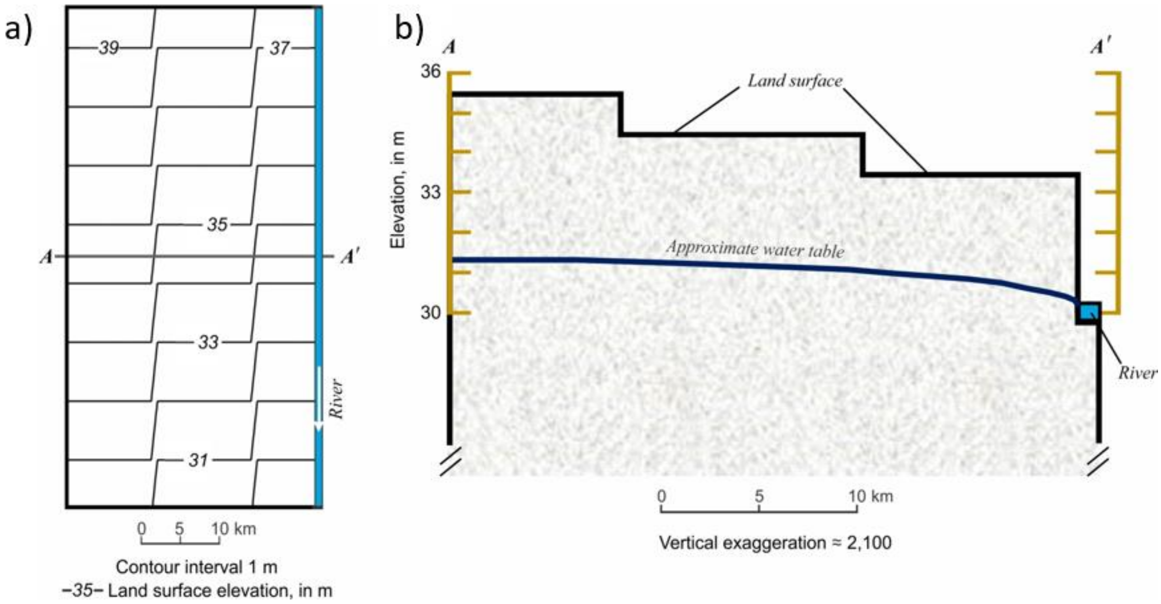

Figure 18 - Topography of the hypothetical stream-aquifer system; a) map view, b) cross-sectional view along line *A-A′*, showing approximate predevelopment water table. Water depth in the river is not shown to scale (Konikow and Bredehoeft, 2020).

Simulation Model

A two-dimensional numerical model was developed to simulate transient groundwater flow in this system. The model was developed using the MODFLOW-NWT code↗, (Niswonger et al., 2011). Streamflow was represented using the Streamflow Routing Package (SFR2) (Niswonger and Prudic, 2005). The model domain was discretized into 80 rows and 40 columns of square cells within a single model layer. The grid spacing was 805 m in each direction. Layer thickness varied depending on the water-table position. For simplification in the model, it was assumed that the streambed elevation can be specified as the linearly varying river heads shown in Figure 17 and that the water depth would remain constant and uniform at 0.001 m.

In the model, evapotranspiration (ET) is assumed to vary as a linear function of the depth to the water table (Figure 19). The ET Surface is the water-table elevation at which the maximum value of evapotranspiration loss occurs (assumed to coincide with the land surface). The extinction depth (or cutoff depth) is that depth of water table at which ET no longer occurs. The ET rate varies linearly between those two limits.

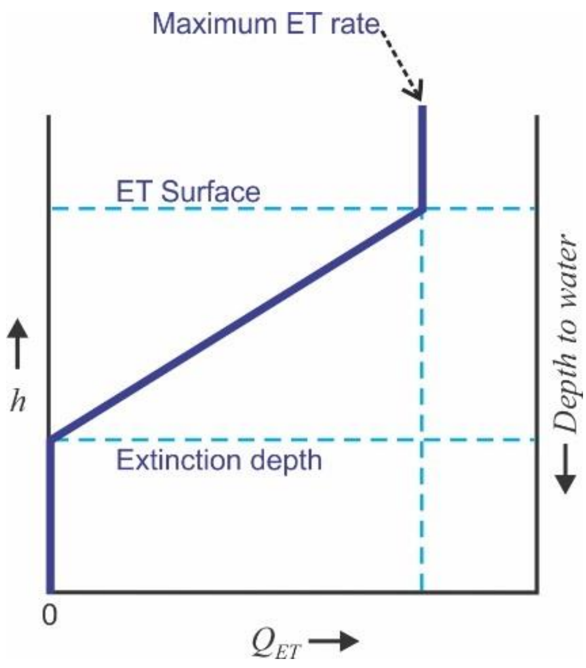

Figure 19 - The linear function relating ET rate (Q_{ET}) to depth of the water table in the MODFLOW-NWT model (*h* is the head [water level]) (Konikow and Bredehoeft, 2020).

Base Case: No Recharge and No ET (No Phreatophytes)

To develop a "base case" for assessing the effects of pumping, a scenario was first developed for a case with no areally diffuse recharge (from precipitation) and no ET losses (no phreatophytes), but including mountain front recharge as a specified flux in all cells along the western edge of the aquifer (represented as injection wells in the model). To develop initial conditions for a 200-year transient simulation, a steady-state run was made first to calculate self-consistent heads and fluxes for predevelopment (initial) conditions with no pumpage. Then it was assumed that the single well would pump for 200 years at a rate of 2,026 m^3/d and a transient simulation was run to simulate changes in heads and fluxes resulting from the imposition of this new pumping stress.

The calculated heads for predevelopment conditions (Figure 20a) show that recharge into the aquifer is dominated by inflow (stream infiltration) from the river along its northern reaches, and discharge of groundwater back out to the river along its southern reaches. The influence of mountain front recharge on the water table distribution is notably smaller than that of the river. Comparison of predevelopment heads with those computed after 200 years of pumping (Figure 20b) indicates that the drawdown from the well has only a small effect on the flow pattern.

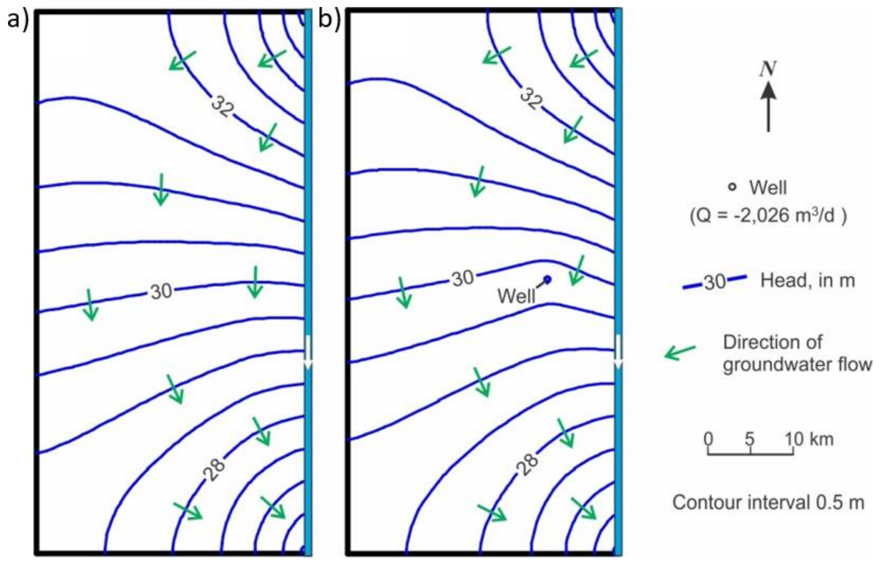

Figure 20 - Map showing calculated water-table elevations in the hypothetical stream-aquifer system for the base case in which there is no areal recharge and no ET: a) for predevelopment conditions, and b) after 200 years of pumping. The head values are relative to an arbitrary datum (Konikow and Bredehoeft, 2020).

The water budget for the predevelopment case is listed in Table 1. It shows that during predevelopment conditions, most of the inflow and all of the outflow were between the river and the aquifer. The streamflow also changed during the simulation (Table 2). The stream discharge at the downstream end of the model was 21,688 m^3/d during predevelopment conditions and decreased to 19,790 m^3/d after 200 years of pumping. The streamflow reduction of 1,898 m^3/d is balanced by the sum of the increase in stream infiltration to the aquifer (1,074 m^3/d) plus the reduction in groundwater discharge to the river (824 m^3/d).

Table 1 - Groundwater budgets for model base case for predevelopment conditions and after 200 years of pumping one well. All flux values are in m^3/d.

		Predevelopment	t = 200 Years
IN	Mountain Front Recharge	1,688	1,688
	Change in Storage	0	128
	Stream infiltration	5,785	6,859
	Total	**7,473**	**8,675**
OUT	Wells	0	2,026
	Outflow to stream	7,473	6,649
	Total	**7,473**	**8,675**

Table 2 - Streamflow budgets for model base case for predevelopment conditions and after 200 years of pumping one well. All flux values are in m^3/d.

	Predevelopment	t = 200 Years
River Inflow	20,000	20,000
River Outflow	21,688	19,790

The components of the water budget changed substantially during the 200-year transient simulation period (Figure 21). After the pumping begins, well discharge is initially balanced completely by groundwater storage depletion. However, over time, more and more well discharge becomes balanced by capture and less and less by storage depletion. After about 17.5 years of pumping, the amount of capture exceeds the amount of storage depletion. After 200 years, only 6.3 percent of pumping is being balanced by groundwater storage depletion while 93.7 percent of well pumping is being balanced by capture. In this case, capture is made up entirely of streamflow depletion. Streamflow depletion is composed of both induced infiltration in the upstream reaches of the river and reduced groundwater discharge to the river in its downstream reaches. As seen in Figure 21, the quantity of induced infiltration is always somewhat greater than the reduction in groundwater discharge in this particular aquifer system. The relative contributions in any particular system will always depend on the hydraulic properties and boundary conditions governing that system.

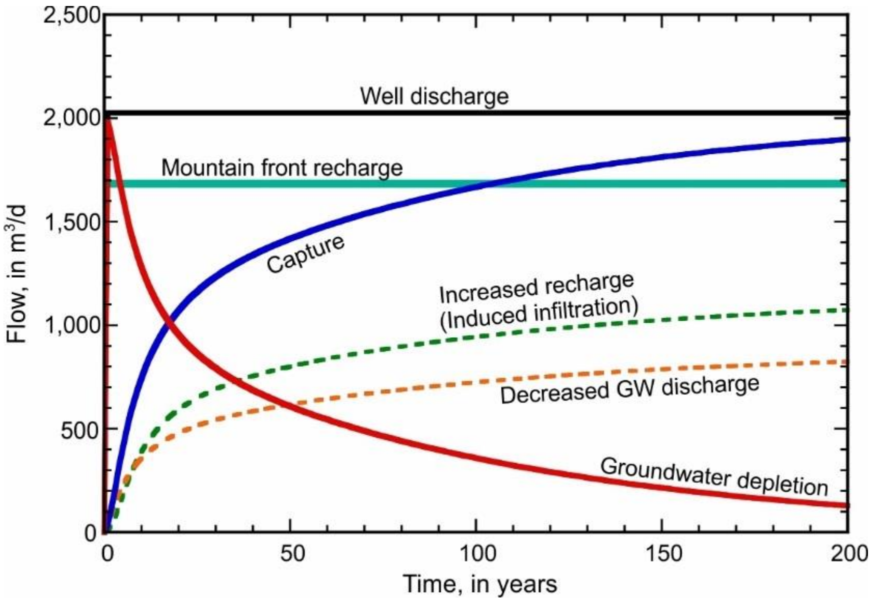

Figure 21 - Calculated changes in the water budget of the hypothetical desert basin aquifer during the 200-year simulation period for the base case model. Capture is the sum of its components represented by the dashed lines (Konikow and Bredehoeft, 2020).

The changes in the water budget and in the sources of water balancing the pumping can also be viewed in a nondimensional way (Figure 22). These fractions can be based on either annual rates or on cumulative volumes. As with the flow rates shown in Figure 21, the fractional sources based on annual rates cross at about 17.5 years, after which capture provides most of the water being pumped by the well. However, when viewed from the perspective of cumulative volumes of water pumped, the crossover does not occur until much later in time -- at about 46 years after pumping started.

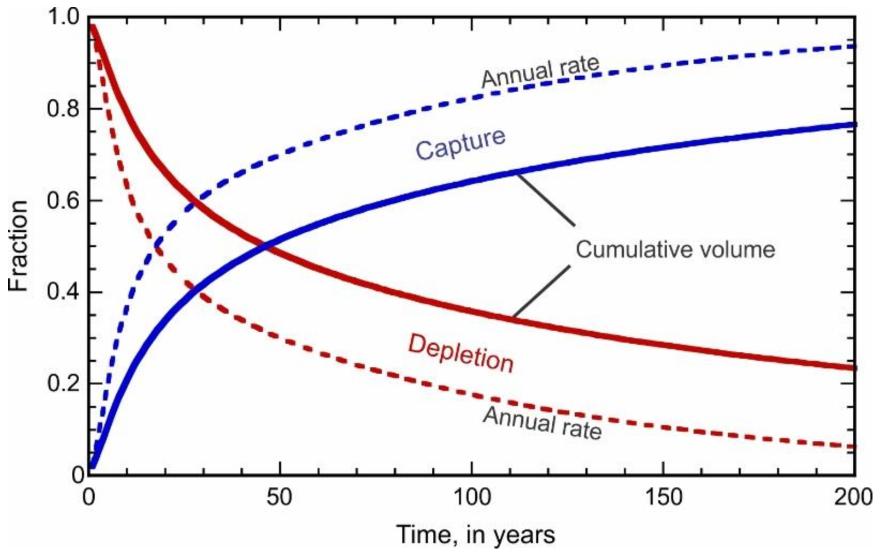

Figure 22 - Nondimensional sources of water being pumped based on both annual rates (dashed lines) and cumulative volumes (solid lines) for the base case model (Konikow and Bredehoeft, 2020).

Low ET Case (Phreatophytes)

To illustrate the effects of evapotranspiration (ET) on the flow system, the base case was modified by allowing the ET process to be represented in the simulation model. The extinction depth was set to 6.0 m, and the maximum ET rate was limited to a relatively low rate of 1.72×10^{-5} m/d (6.28 mm/yr), yielding a linear change in ET between the extinction depth and a water table depth at the land surface (Figure 23).

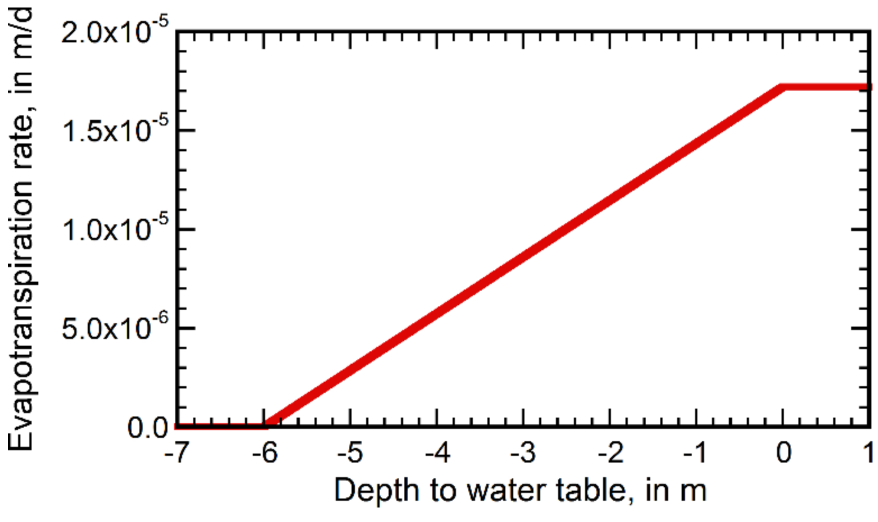

Figure 23 - Relation between evapotranspiration rate and depth to the water table as represented in the MODFLOW model for the Low-ET Case (Konikow and Bredehoeft, 2020).

The steady-state predevelopment scenario was simulated first with the addition of ET, followed by 200 years of pumping identical to the pumping rate in the base case. The calculated water budget for this case (Table 3) shows that the relatively low ET rate per unit area becomes a substantial stress for the aquifer when spread out over the large extent of the aquifer (about four times that of the pumping stress). This presence of ET in the predevelopment model results in groundwater loss and therefore lowers the groundwater level. This additional ET stress under predevelopment conditions (8,163 m^3/d) lowers heads enough to induce almost twice the infiltration from the river into the aquifer (compare 5,785 m^3/day in Table 1 to 9,702 m^3/day in Table 3). The lower heads also cause a 57 percent reduction in groundwater discharge back out to the river in the downstream reaches of the river (compare 7,473 m^3/d in Table 1 to 3,231 m^3/d in Table 3). The actual ET rate varies spatially because the depth to the water table varies, and over the area of the aquifer averages 1.5 mm/yr. Because of this additional ET consumptive loss relative to the base case, under predevelopment conditions the streamflow out of the last river reach is reduced to 13,530 m^3/d (compared to 21,688 m^3/d in the base case with no ET). That is, the ET loss from the aquifer is completely balanced by a reduction in streamflow out of the study area.

Table 3 - Groundwater budgets for model case with low ET rate for predevelopment conditions and after 200 years of pumping one well. All flux values are in m^3/d.

		Predevelopment	t = 200 Years
IN	Mountain Front Recharge	1,688	1,688
	Change in Storage	0	63
	Stream Infiltration	9,702	11,199
	Total	**11,390**	**12,950**
OUT	Wells	0	2,026
	Evapotranspiration	8,163	7,844
	Discharge to Stream	3,231	3,084
	Total	**11,394**	**12,954**

During the 200-year pumping period, the water table declines somewhat, and the depth-dependent ET loss is reduced by just 4 percent (relative to predevelopment rates). This represents captured ET, which helps offset (or balance) the pumping (Figure 24). Compared to the base case with no ET losses, after 200 years of pumping the induced infiltration rate into the aquifer is almost twice as high, and the groundwater discharge rate to the river is almost half as much as in the base case. Streamflow depletion is smaller than in the base case and total capture now includes a component of salvaged ET. Because total capture has increased, the groundwater storage depletion has decreased relative to the base case. The net effect on streamflow is that after 200 years, the streamflow out of the downstream end of the river is reduced from a predevelopment value of 13,530 m^3/d to 11,885 m^3/d (compared to a reduction from 21,688 m^3/d to 19,790 m^3/d in the base case).

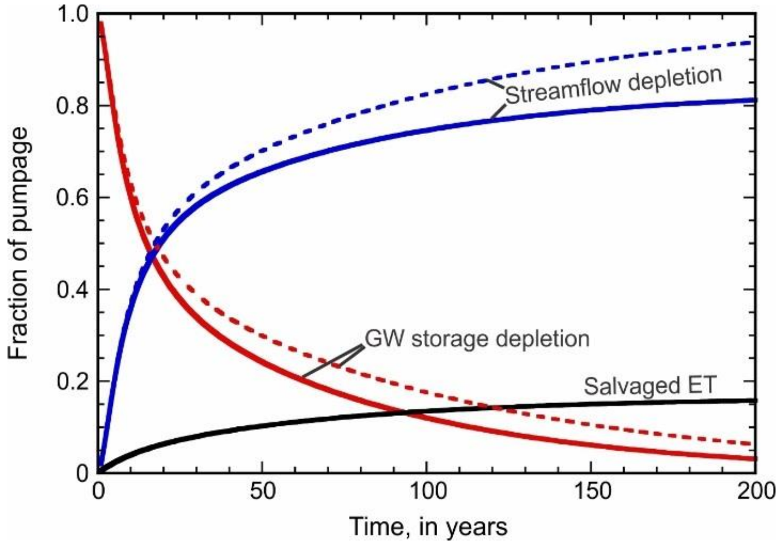

Figure 24 - Nondimensional sources of water being pumped based on annual rates for the base case model (dashed lines) and for the case with low ET (solid lines) (Konikow and Bredehoeft, 2020).

ET and Recharge Case (Phreatophytes and Rainfall)

The final scenario modeled applied a diffuse recharge rate at a uniform average long-term rate of 9.0×10^{-5} m/d (32.8 mm/yr) to represent recharge from precipitation and irrigation. The maximum ET rate was increased from the previous scenario to 2.8×10^{-4} m/d (103 mm/yr); the extinction depth was kept at 6.0 m. The calculated water budget for the simulation results in this case (Table 4) shows that the ET and diffuse recharge are now the dominant components of the water budget for both predevelopment conditions and the transient pumping conditions. However, considering all diffuse surface fluxes together, the difference between the total diffuse recharge and the total ET is much smaller -- a net discharge by ET flux of just 8,157 m³/d, which is not very different from the ET flux in the previous case and represents about four times that of the pumping stress.

Table 4 - Groundwater budgets for predevelopment conditions and after 200 years of pumping for case with both diffuse recharge and ET. All flux values are in m³/d.

		Predevelopment	t = 200 Years
IN	Mountain Front Recharge	1,688	1,688
	Diffuse Recharge	186,479	186,479
	Change in Storage	0	0
	Stream Infiltration	7,439	8,089
	Total	**195,606**	**195,256**
OUT	Wells	0	2,026
	Evapotranspiration	194,636	193,260
	Discharge to Stream	975	974
	Total	**195,611**	**196,260**

This volumetric rate averaged over the surface area of the aquifer is equivalent to 1.5 mm/yr. But it does vary spatially (Figure 25). Because diffuse recharge is applied uniformly over the area of the aquifer, whereas ET varies as a function of depth to the water table, the net surface flux is greatest where the depth to water is the greatest and ET is the lowest, such as the western edge of the aquifer furthest from the river (Figure 25a). The net surface flux is the most negative (ET loss greater than recharge) closer to the river, where the depth to water is minimal and ET is highest. Note that the north-south banding shown in Figure 25a derives from the stepped terraces of the land surface topography (Figure 18). Contours of change in surface flux after 200 years (Figure 25b) show that the change in ET is small, except very close to the pumping well where depth to the water table increases the most after pumping starts; of course, the contours in Figure 25b parallel the drawdown around the well.

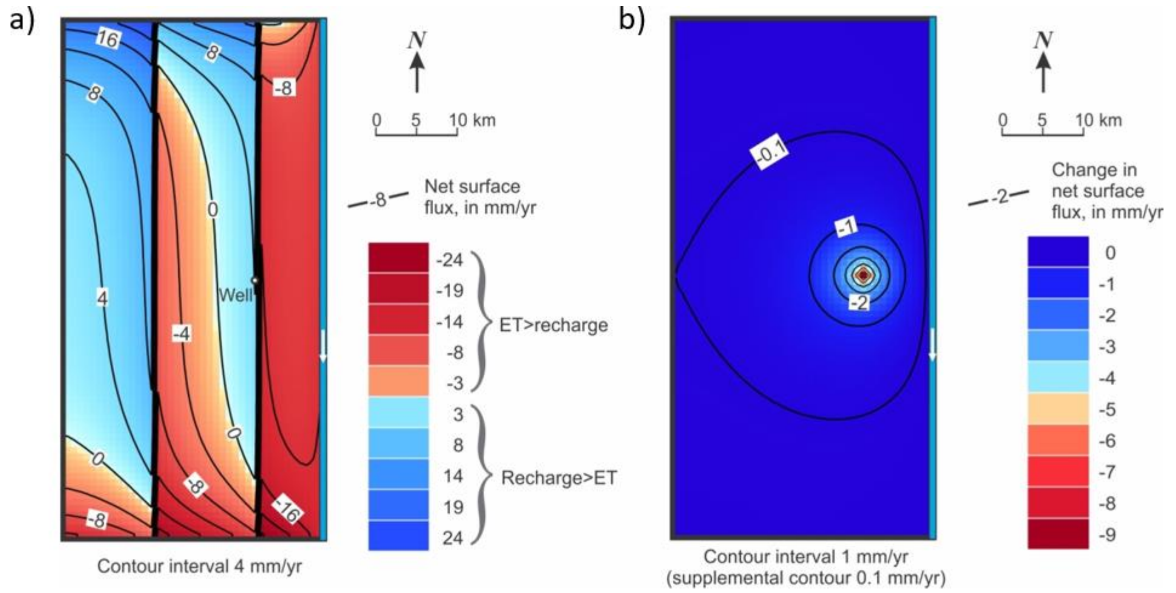

Figure 25 – Net surface flux is the difference between recharge and evapotranspiration: a) Net surface flux (recharge minus ET) for predevelopment conditions, and b) change in net surface flux after 200 years of pumping (Konikow and Bredehoeft, 2020).

The large fluxes of recharge and ET combined with the nonuniformity of the ET, visibly affect the head distribution in the stream-aquifer system (Figure 25). Comparing Figure 25 with Figure 20b, it is evident that heads near the western edge of the aquifer are generally higher than in the base case (no recharge or ET) because recharge generally exceeds ET in most of that band, whereas heads near the eastern edge (close to the river) are generally lower in Figure 26 because ET exceeds recharge in that band. This also induces infiltration along a greater length of the river in the ET/Recharge case than in the base case, as reflected by the angle that the head contours intersect the river. The predevelopment stream infiltration into the aquifer is almost 30 percent greater than in the base case, though smaller than in the previous low-ET case. The predevelopment aquifer discharge to the stream is much smaller than in base case, as reflected by the angle that the head contours intersect the river. The predevelopment stream infiltration into the aquifer is almost 30 percent greater than in the base case, though smaller than in the previous low-ET case. The predevelopment aquifer discharge to the stream is much smaller than in the base case, and also smaller than in the previous low-ET case, in large part because the head distribution only allows groundwater discharge to the river along a short downstream reach of the river.

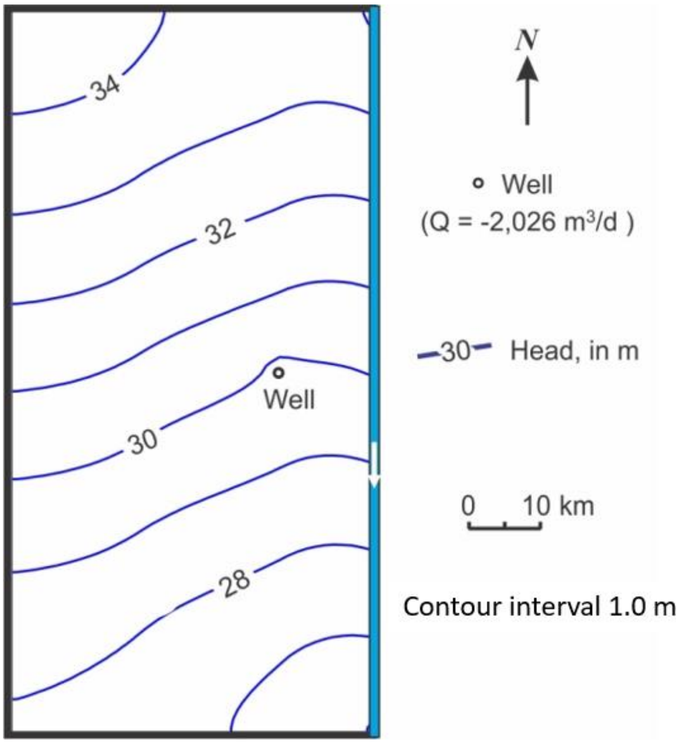

Figure 26 - Calculated water-table elevations after 200 years of pumping for the case in which there is both diffuse areal recharge and ET. Head is relative to an arbitrary datum. (Konikow and Bredehoeft, 2020).

After 200 years of pumping, the rate of change in storage is zero. This indicates that the system has reached a new equilibrium state by this time, and as seen in Figure 27, the new equilibrium was attained after about 75 years of pumping. In contrast, a new equilibrium had not been reached in 200 years with identical pumping for both the base case and the low-ET case. The primary reason for the change in relations here is that the much higher absolute magnitude of ET in this case permits much more of this ET discharge to be captured (or salvaged) to offset (or balance) the pumping, and it can happen faster than streamflow depletion. It is faster because streamflow capture requires time for the drawdown effects to propagate to the stream boundary, whereas ET capture occurs immediately and locally as the water table declines. Furthermore, recharge in this model scenario is a specified flux condition, and it is not affected by pumping or by drawdown. Because ET salvage is so fast and so large, the impact of pumping on streamflow is much less than in either previous case. Similarly, less groundwater storage depletion is needed to balance pumping. When storage depletion reaches zero, it means that heads have stabilized and no additional drawdown is occurring. This is the very definition of an equilibrium condition in an aquifer.

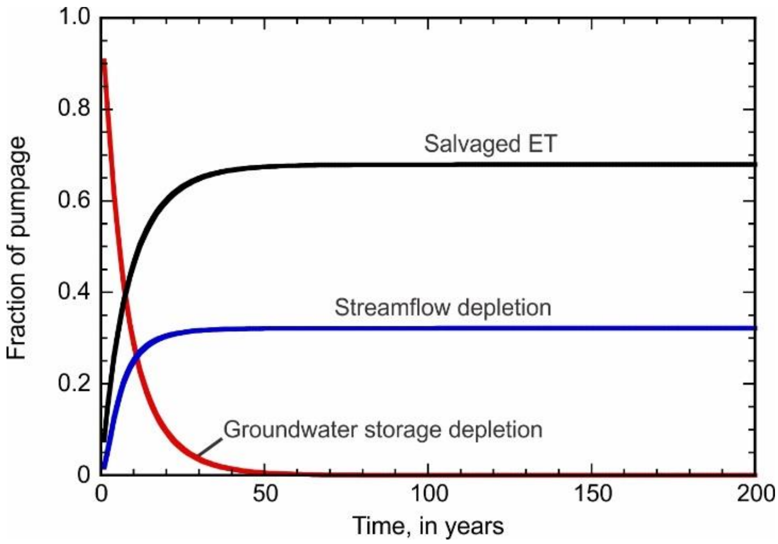

Figure 27 - Simulated nondimensional sources of water being pumped based on annual rates for the case with ET and recharge (Konikow and Bredehoeft, 2020).

After 200 years of pumping, the stream infiltration into the aquifer has decreased by 9 percent, but the groundwater discharge to the river has stayed at about the same low rate. The streamflow leaving the downstream end of the river is 13,536 m^3/d during predevelopment conditions, and that is reduced to 12,884 m^3/d after 200 years of pumping. This decrease in streamflow of about 652 m^3/d over time is less than half that in the low-ET case, and the fractional (nondimensional) proportion of pumping balanced by streamflow depletion is smaller than in the low-ET case (compare Figure 27 with Figure 24). Again, this is because a much greater fraction of pumping is more readily balanced by salvage (or capture) of ET losses from the aquifer.

Running the Model

The procedure for obtaining the model files, running the three scenarios presented for Case Study 1, post-processing the model outputs, and viewing the model results are explained in Box 3↴. The post-processing includes: creating contour maps of head and drawdown, plots of the water budget through time, calculating changes in streamflow with time, and plotting well hydrographs (i.e., water level as a function of time in a well).

Summary

Idealized hypothetical aquifers can be simulated to illustrate cause and effect relations in stream-aquifer systems. The simulations presented here have shown that groundwater withdrawn by wells must be balanced by a combination of increased recharge, decreased discharge, and depletion of groundwater in storage in the aquifer. An archive of the models, including input and output files, used to simulate the three primary scenarios described in this section are available in the file "CaseStudy1--Models.zip" as part of the online Supplemental Materials for this book↗. ET losses from the water table constitute one form of groundwater discharge, and lowering the water table can reduce ET losses. The reduction in ET can be viewed as salvaged ET, which helps balance pumping from wells. This might have environmental implications because of impacts on surface vegetation and changes in climatic conditions in areas where the evapotranspired water was contributing to precipitation. Additional changes in the fluxes between the aquifer and the bounding river also help to balance pumping, but constitute streamflow depletion. This might have environmental and legal implications because of impacts on aquatic ecosystems and existing surface water rights.

Salvaged ET can be a large component of capture. A small decrease in the ET rate over a large area can yield a large volume of water. Most salvaged ET offsets and reduces streamflow depletion. In a field setting, confirmation of ET salvage may be very difficult.

If the flow in the river were small enough that the effects of pumping caused reaches of the river to go dry, then pumping would have to be balanced by more storage depletion. The rate of head decline would be larger and the system could not reach a new equilibrium if there were no additional sources of capture.

6.2 CASE STUDY 2: PARADISE VALLEY, NEVADA

Unlike hypothetical examples, real-world groundwater systems are complex and heterogeneous, and there are never enough data available to define their properties and boundary conditions uniquely, accurately, and precisely. Hence, analyses of such systems are always made in light and consideration of this uncertainty. Any model developed for field aquifer systems is always an approximation and always subject to improvement as additional data become available. But hydrogeologists are typically tasked with analyzing field systems -- that is our job and that is our challenge -- and the complexity and uncertainty do not preclude us from developing reliable and useful models for the task at hand. Therefore, we next provide an illustrative example of an analysis of an aquifer system that has been developed in historical times.

Description of Study Area

Paradise Valley, located in north-central Nevada, USA, is a typical valley in the Basin and Range topographic province of Nevada and western Utah in the western United States (Figure 28). It is a north-south trending valley, approximately 40 miles long by 10 miles wide (64 km by 16 km), extending north from the valley of the Humboldt River. The town of Winnemucca is situated along the Humboldt River to the southwest of Paradise Valley. Paradise Valley is surrounded by mountains to the east, north, and west, and open to the valley of the Humboldt River to the south.

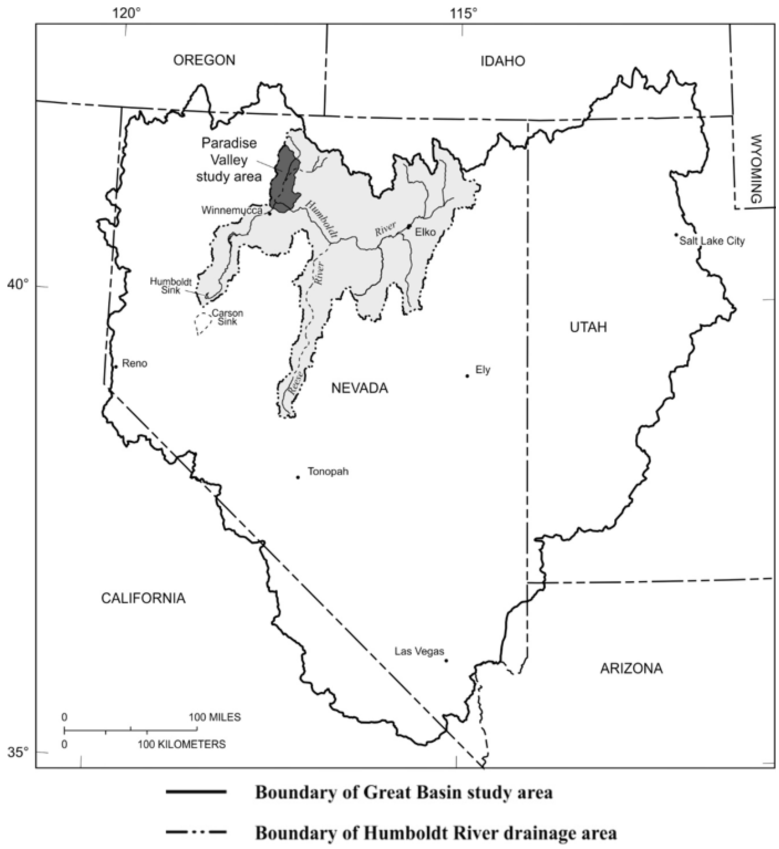

Figure 28 - Location of Paradise Valley study area and Humboldt River drainage basin in Nevada, western United States (modified from Prudic and Herman, 1996).

The region is arid. A 100-year rainfall record in Paradise Valley indicates that the average precipitation is approximately 8 inches per year (200 mm/yr) (Prudic and Herman, 1996). Direct recharge from precipitation on the valley floor is assumed to be negligible. Two streams enter the valley from the northeast: Martin Creek and the Little Humboldt River. Martin Creek has been continuously monitored since the early 1920s (Figure 29). The locations of these two streams are shown in Figure 30.

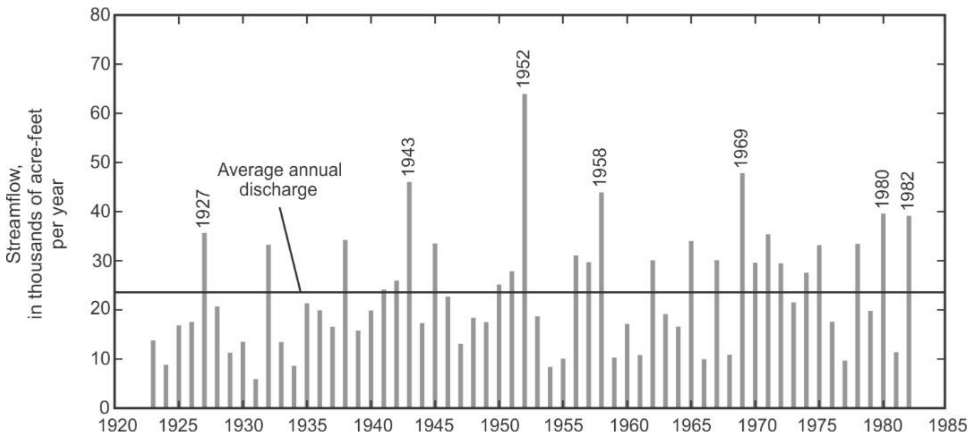

Figure 29 - Annual discharge of Martin Creek where it enters Paradise Valley, Nevada, USA. [1,000 ac-ft = 0.000123 km^3] (from Prudic and Herman, 1996).

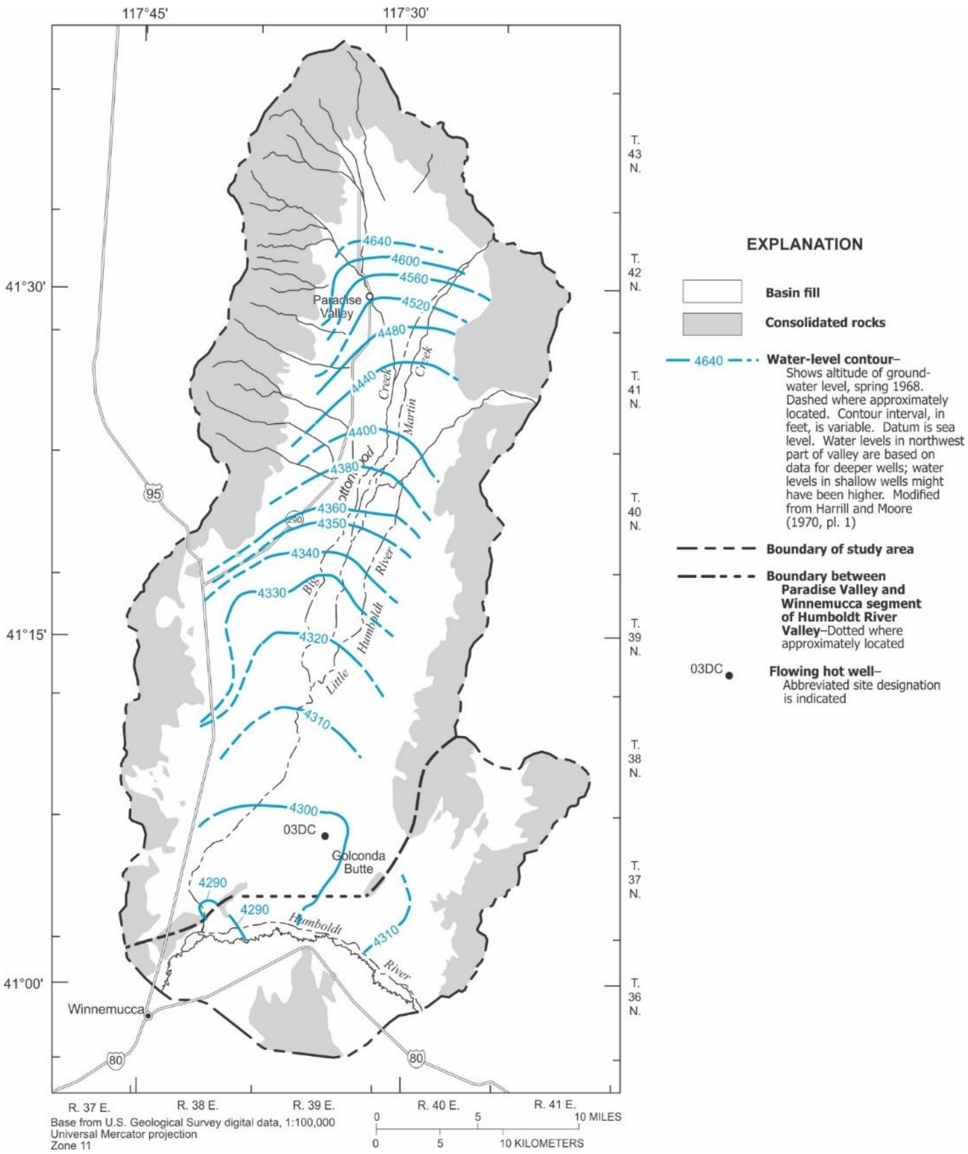

Figure 30 - Groundwater levels in Paradise Valley, Nevada, spring 1968 (from Prudic and Herman, 1996).

The annual discharge of Martin Creek is typical of a desert stream with wet and dry years. The Little Humboldt River is a similar stream; it has a small reservoir in the mountains to the east of Paradise Valley. There are only occasional measurements of the flow of the Little Humboldt River during the period shown on Figure 29, 1923 to 1982. With rare exceptions, all of the water from both Martin Creek and the Little Humboldt River infiltrates and recharges the alluvial aquifer in Paradise Valley near the northern end of the valley.

Western Nevada had a much wetter climate during the periods of Pleistocene ice advances. Lake Lahontan occupied much of western Nevada during the advance. Paradise Valley was on the fringe of the lake; two maximum rises of the lake extended into the lower reach of Paradise Valley. During the glacial period there was a through-going stream in Paradise Valley. This stream created a highly permeable alluvial deposit down the center of the valley (Bredehoeft, 1963). Paradise Valley is filled with alluvial sediments to a depth of 2,000 to 3,000 feet (600 to 900 m), but the thickness may exceed 8,000 feet (2,400 m) in the center of the valley (Prudic and Herman, 1996). The alluvium is underlain by igneous, metamorphic, and sedimentary consolidated rocks generally having low porosity and low permeability. The valley bottom slopes gently from north to south toward the Humboldt Valley. The water table generally mimics the topography of the valley floor (Figure 30).

The depth to water map (Figure 31) indicates a large area in the center of the valley where the water table is less than 10 feet (3 m) below the land surface. This is an area where phreatophytes flourished prior to groundwater development.

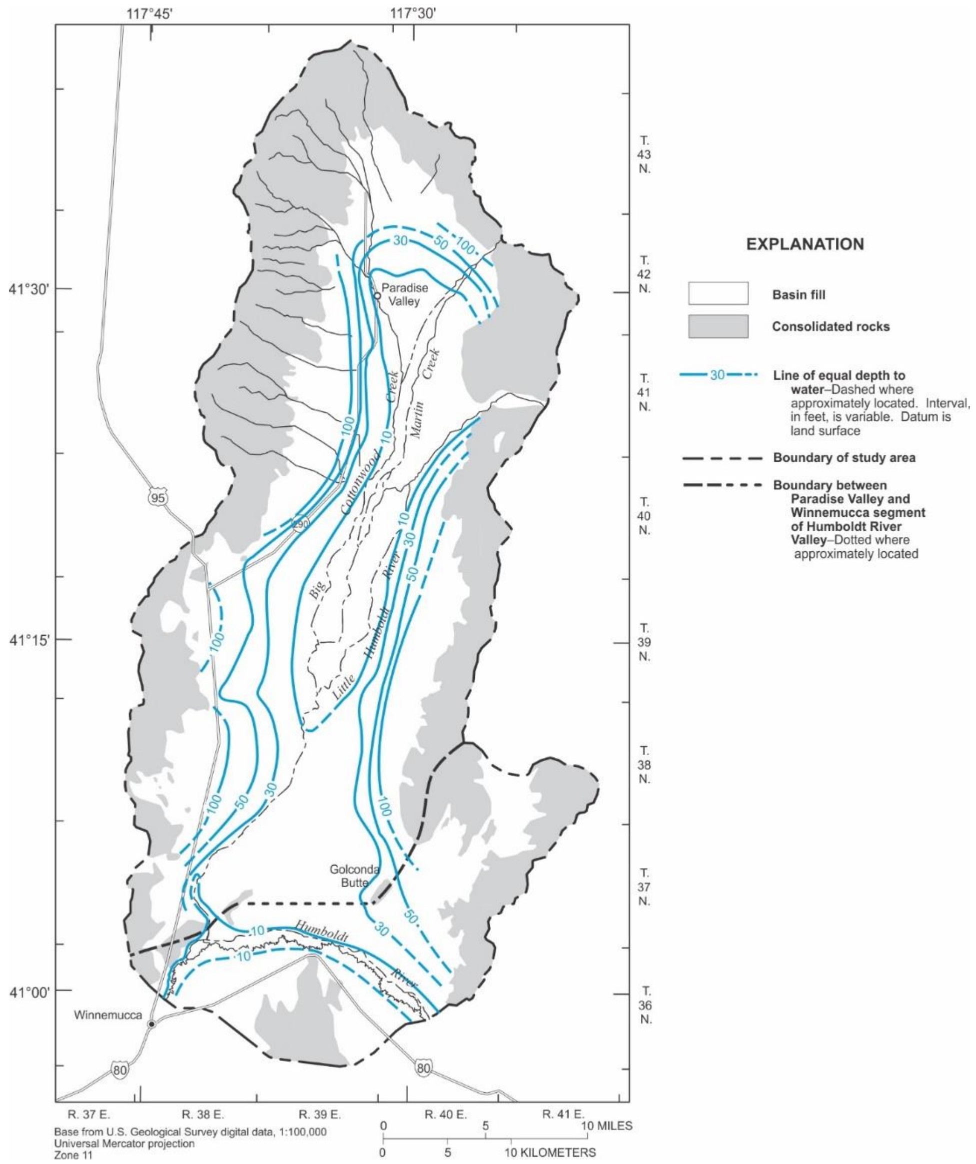

Figure 31 - Depth to groundwater in Paradise Valley, Nevada, prior to major development (pre-1969) (from Prudic and Herman, 1996).

A reconnaissance investigation of Paradise Valley prior to significant development indicated a valley with alluvium near fully saturated with groundwater (Loeltz et al., 1949). Water from Martin Creek and the Little Humboldt River infiltrates into the alluvial aquifer within a short reach of entering the valley, as usually do the flows in all other smaller tributaries draining off the adjacent mountains. The reconnaissance studies (e.g., Loeltz et al., 1949) suggested that the average annual recharge available from the two streams, Martin Creek and the Little Humboldt River, was nearly 40,000 ac-ft/yr (0.05 km^3/yr or 1.6 m^3/s), which was balanced by discharge through evapotranspiration from a shallow water table over much of the valley. As shown in Figure 31, the depth to the water table is less than 30 ft (9.1 m) throughout the large central part of the valley. Groundwater discharges as evapotranspiration; much of the discharge is transpiration by phreatophytes. The thick, permeable, alluvial aquifer makes the area ideal for groundwater development. Groundwater pumping in the area increased gradually during the 1950s and 1960s, but accelerated markedly during the 1970s when it was discovered that Paradise Valley was a good place for groundwater development (Figure 32).

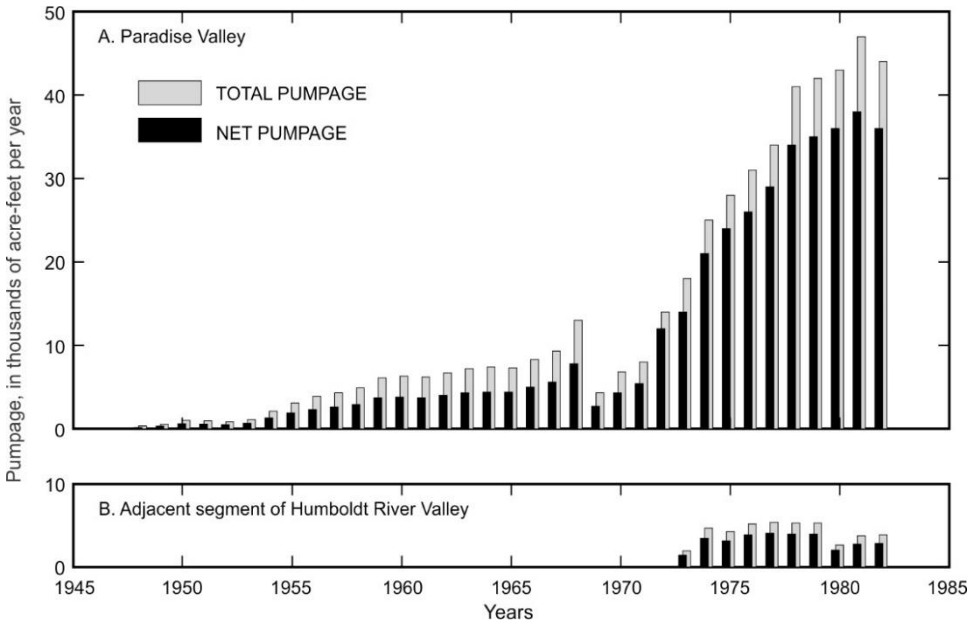

Figure 32 - Pumping in Paradise Valley and the associated part of the Humboldt River Valley (from Prudic and Herman, 1996). Net pumpage equals total pumpage minus the amount estimated to return to basin-fill aquifer through infiltration of irrigation water. [1,000 ac-ft = 1.23 million m^3 = 1.23×10^{-3} km^3].

When irrigating, some of the water applied to a field is not used by the crop; rather, it infiltrates into the soil as groundwater recharge from irrigation. This is especially true where the water table is close to the land surface. The net pumping is the amount of pumping used by the plants or evaporated -- the net pumping is the consumptive use.

Groundwater is used as the principal source for irrigation in the Paradise Valley area, and wells generally irrigate fields located close to the wells. Most of the irrigation wells in the area were drilled after 1969 and are located in the southern part of Paradise Valley (Figure 33). As the pumpage increased rapidly between 1969 and 1981, a substantial cone of depression developed in the area of heaviest pumping and irrigation in the southern part of the valley (Figure 34). Water-table declines exceeded 80 ft (24 m) in the center of the affected area. There was a decline of more than 10 ft (3 m) over a large area in the lower part of Paradise Valley.

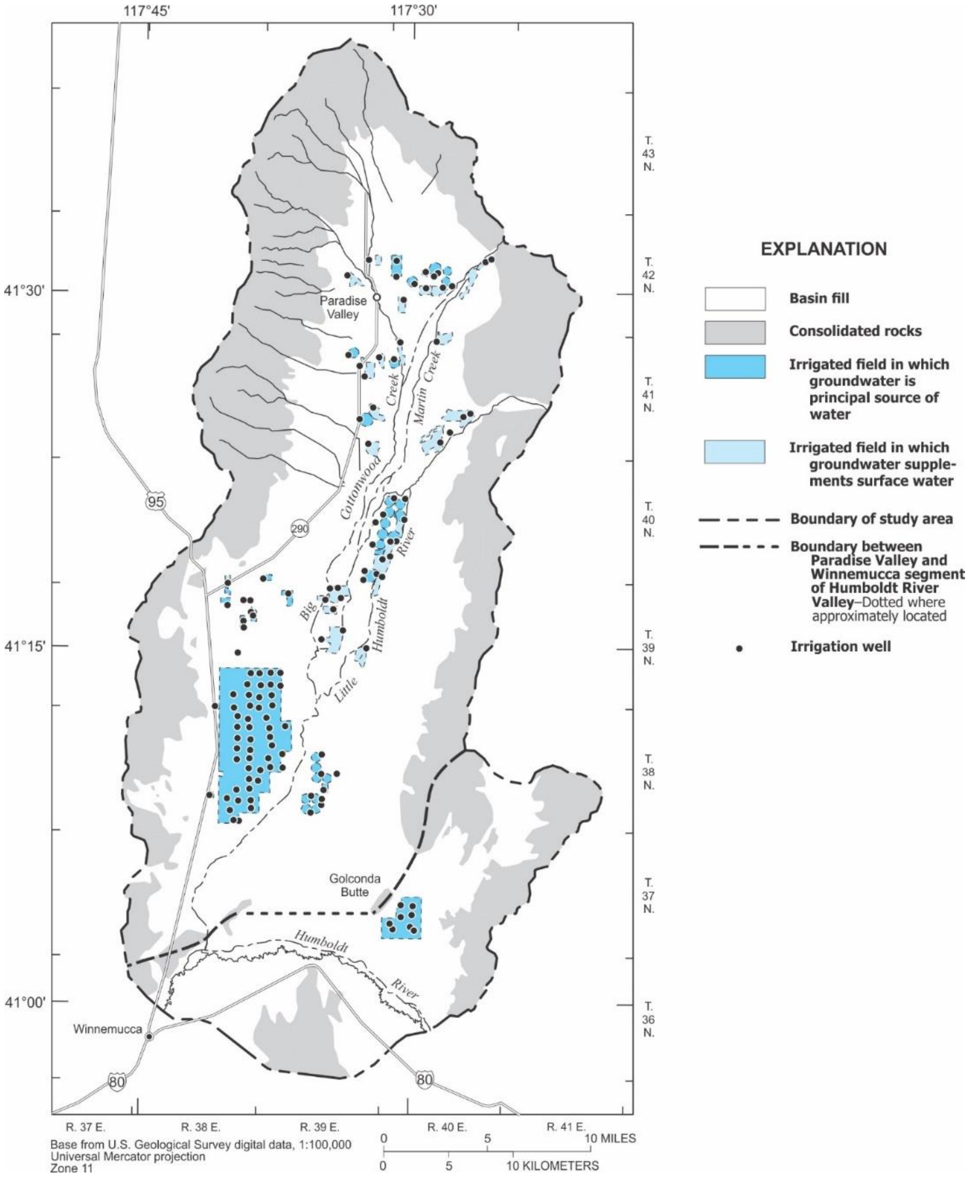

Figure 33 - Distribution of irrigated lands and irrigation wells in 1981 (from Prudic and Herman, 1996).

As groundwater development increased during the 1960s and 1970s, there were management concerns about what happens to groundwater in the valley with continued pumping. Is the pumping sustainable? Would drawdown from pumping lower the water table and capture much of the evapotranspiration -- leading to development that might be sustained for a long period? But a quantitative basis for assessing these questions was lacking. One approach to analyzing this question is to develop and calibrate a numerical model of the groundwater system, and then use the model to simulate the impact of continued development. In an effort to better understand the system, the USGS undertook a comprehensive numerical groundwater model study, described by Prudic and Herman (1996).

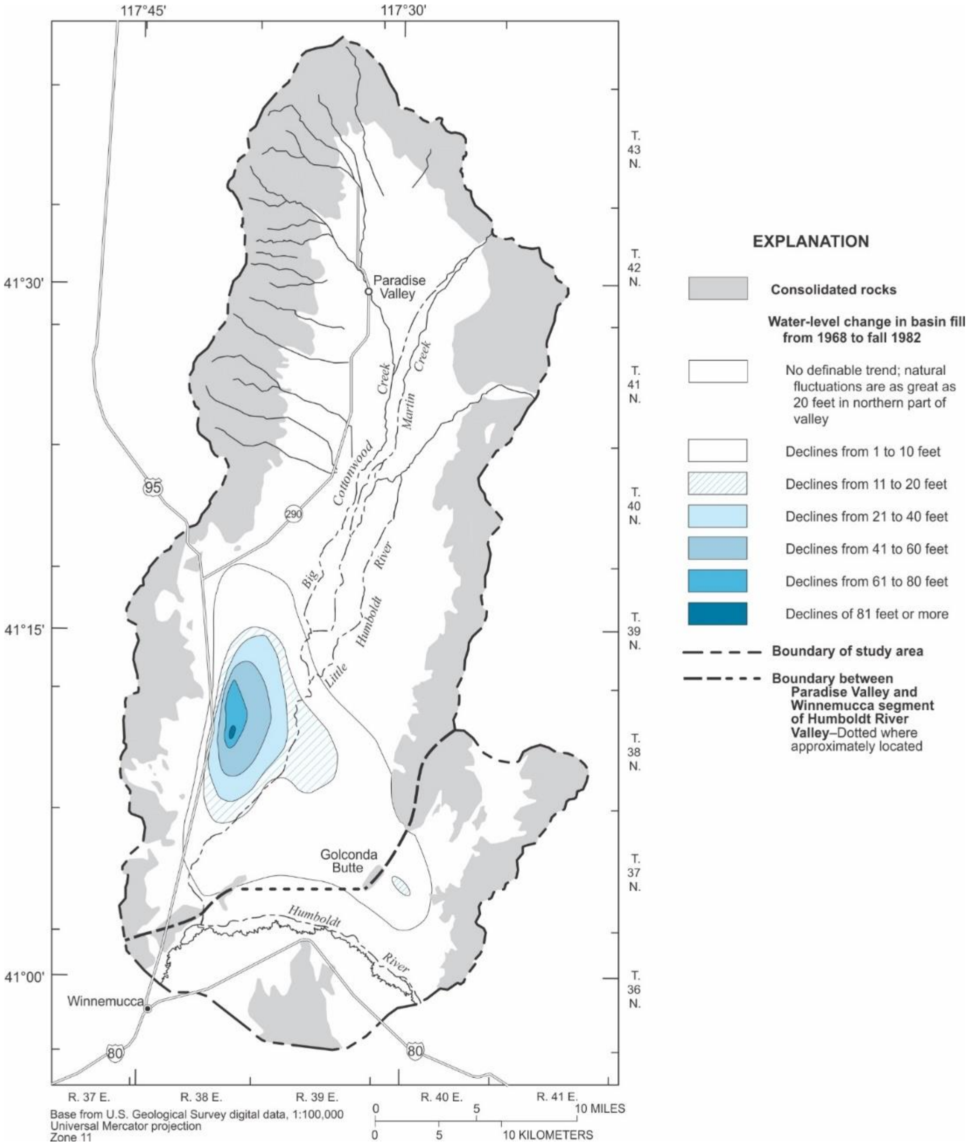

Figure 34 - Water-table decline, 1968 to 1982 (from Prudic and Herman, 1996).

Paradise Valley Simulation Model

A three-dimensional finite-difference model was developed for Paradise Valley using the generic MODFLOW-88 model (McDonald and Harbaugh, 1988). It was conceptualized to represent varying saturated thickness, heterogeneous hydraulic properties, recharge from stream leakage, inflow from adjacent bedrock, discharge from evapotranspiration, and withdrawal by pumping wells (Figure 35). The grid used three layers, 33 columns, and 89 rows of cells. Layer thicknesses were 600 to 1,200 ft (183-366 m). Horizontal grid dimensions were 2,500 ft (762 m) on a side, and the center of each cell is designated as a node. The Humboldt River was used to simulate the southern boundary of flow in Paradise Valley, and it was simulated as a head-dependent flux boundary condition. All other boundaries were represented as no-flow boundaries, although specified fluxes are applied to some bounding cells to represent inflow from adjacent bedrock (Prudic and Herman, 1996, for details).

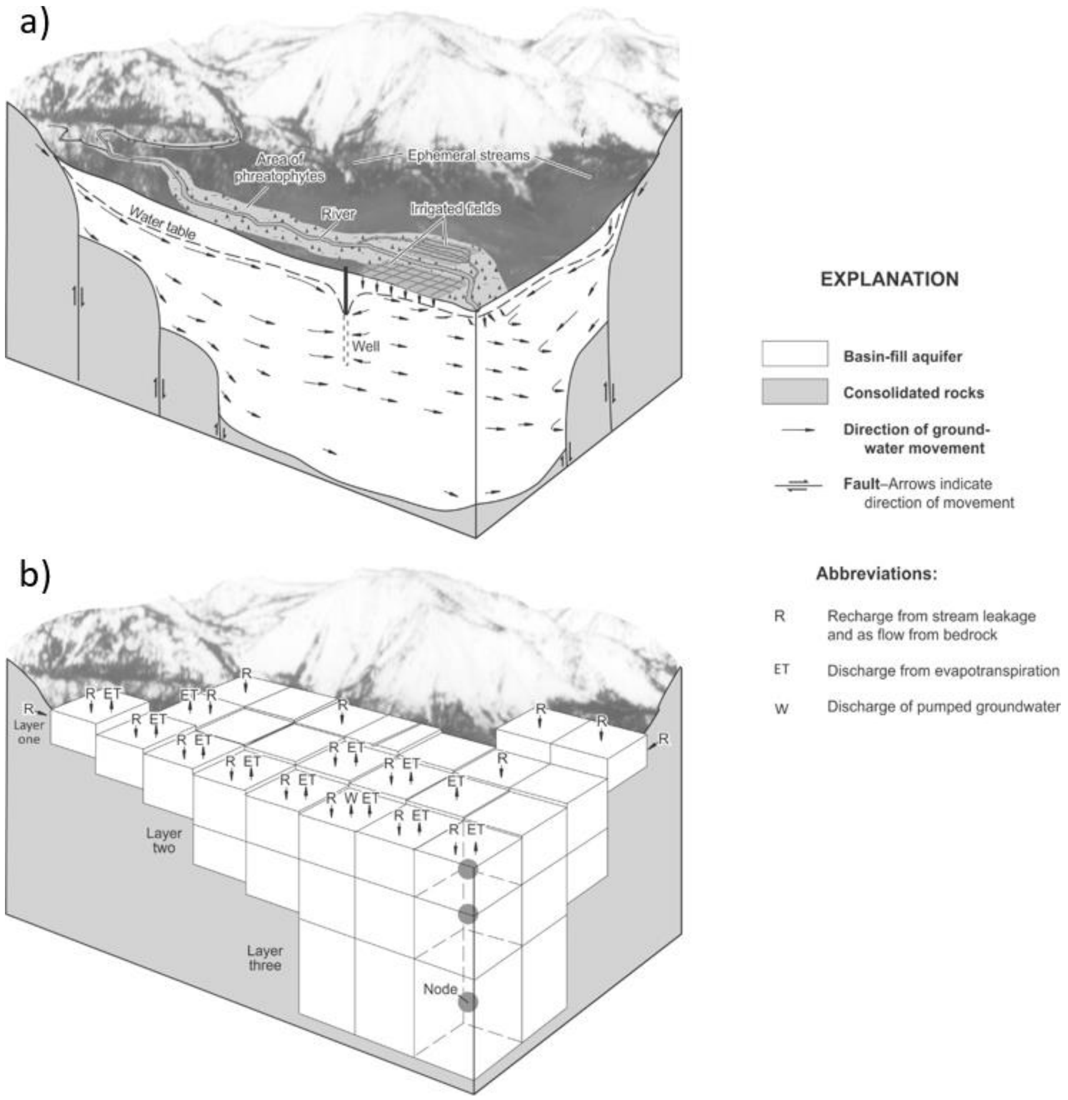

Figure 35 - Schematic three-dimensional diagram that illustrates how the groundwater system in Paradise Valley was subdivided into a finite system of rectangular cells for the purpose of numerical analysis (modified from Prudic and Herman, 1996).

An Initial Steady State

The usual procedure in modeling a groundwater system is to first model the system prior to development -- its undisturbed state. This procedure provides the best estimates of natural inflow (recharge) to and outflow (discharge) from the system. One uses the model to fit the observed undisturbed head distribution (Figures 30 and 31) and observed fluxes (if known). In calibrating the model, one adjusts the inflows and outflows, including their distribution, and the hydraulic parameters (such as hydraulic conductivity of the aquifer) until an adequate (or best) fit is achieved. Prudic and Herman (1996) developed a steady-state model to represent likely average conditions at the start of their transient simulations (1948).

Once one has adequately modeled the initial state of the system, the model yields a calculated water budget representative of the undisturbed system. Consistent with Equation 2, the total inflows (recharge) and outflows (discharge) under steady-state conditions are balanced (at a rate of about 74,000 ac-ft/yr [2.89 m^3/s]). The Paradise Valley groundwater system is thought to be very nearly undisturbed prior to 1948 and only slightly developed during 1948-1969.

Transient Historical Simulation

Transient flow simulations were made for 1948-1982 to match the historical period of record and help assess the reliability of the model. Although the model results show substantial annual variability, Prudic and Herman (1996) report that about 60 percent of the net pumpage was derived from a reduction in storage. Therefore, about 40 percent is balanced by capture, consisting of reduced evapotranspiration, increased recharge from streams, and decreased discharge to streams.

Simulated Future Development

It was considered that future development could potentially include a prolonged total pumpage equal to about 72,000 ac-ft/yr (2.82 m^3/s), an amount close to the long-term average annual streamflow into Paradise Valley. To help clarify the consequences of doing this, Prudic and Herman (1996) simulated the system under this scenario (among others) for a 300-year period of pumping. Figure 36 shows how critical inputs and outputs respond to the long 300-year period of pumping under the stresses of this proposed but hypothetical scenario.

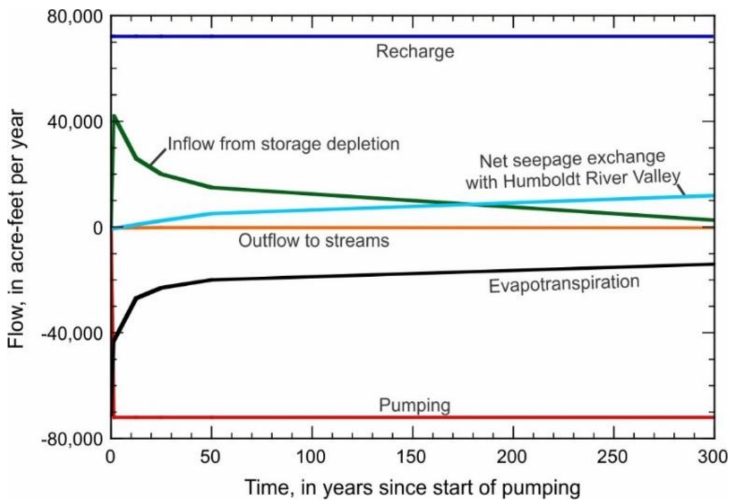

Figure 36. Model-calculated fluxes versus time for a scenario in which pumping of 72,000 ac-ft/yr (2.82 m^3/s) is distributed widely in Paradise Valley (data from Prudic and Herman, 1996, scenario 5). Recharge includes both mountain front recharge and infiltration from streams.

These results indicate that not all of the phreatophyte use is captured even after 300 years of pumping; about 14,000 ac-ft/yr (0.55 m^3/s) of use remains. After 300 years, pumping has induced an increase of approximately 12,400 ac-ft/yr (0.47 m^3/s) in the net underflow between the Humboldt River Valley and the Paradise Valley groundwater system (by increasing underflow into Paradise Valley from the Humboldt River Valley by 10,700 ac-ft/yr and decreasing underflow out of Paradise Valley into the Humboldt River Valley by 1,700 ac-ft/yr). There is still approximately 2,600 ac-ft/yr (0.10 m^3/s) continuing to be removed from storage. This means that after 300 years of pumping, the system still has not quite reached a new equilibrium state.

The drawdown after 300 years resulting from the pumping associated with this scenario exceeded 100 ft (30 m) in a large central part of the valley (Figure 37). In this particular pumping scenario, the pumping was distributed throughout the valley in the areas in which the phreatophytes grew. Drawdown in the area of riparian vegetation maximizes the capture of evapotranspiration by the pumping.

The Lesson of the Paradise Valley Example

Paradise Valley is ideally suited to groundwater development. The valley is filled with an alluvial aquifer that is both highly permeable and more than 2,000 feet (610 m) thick in most of the valley. One can drill highly productive wells in the valley. The climate, although arid, is well-suited for irrigated agriculture.

The model analysis indicates that the groundwater system in Paradise Valley was in a state of long-term equilibrium prior to development of its groundwater resources, with approximately 70,000 ac-ft/yr (2.7 m^3/s) of recharge and discharge and shallow groundwater levels throughout. The principal recharge was infiltration of water from both Martin Creek and the Little Humboldt River, and the principal discharge was by riparian phreatophytes.

In creating a groundwater development system that could be maintained for a long period in Paradise Valley, the strategy was to install a pumping system that would lower the water table the most in the area of the phreatophytes and "capture" their discharge. Various alternative pumping arrangements are feasible. The scenario shown in figure 37 spreads the pumping widely up and down the valley and minimizes the drawdown for that magnitude of pumping.

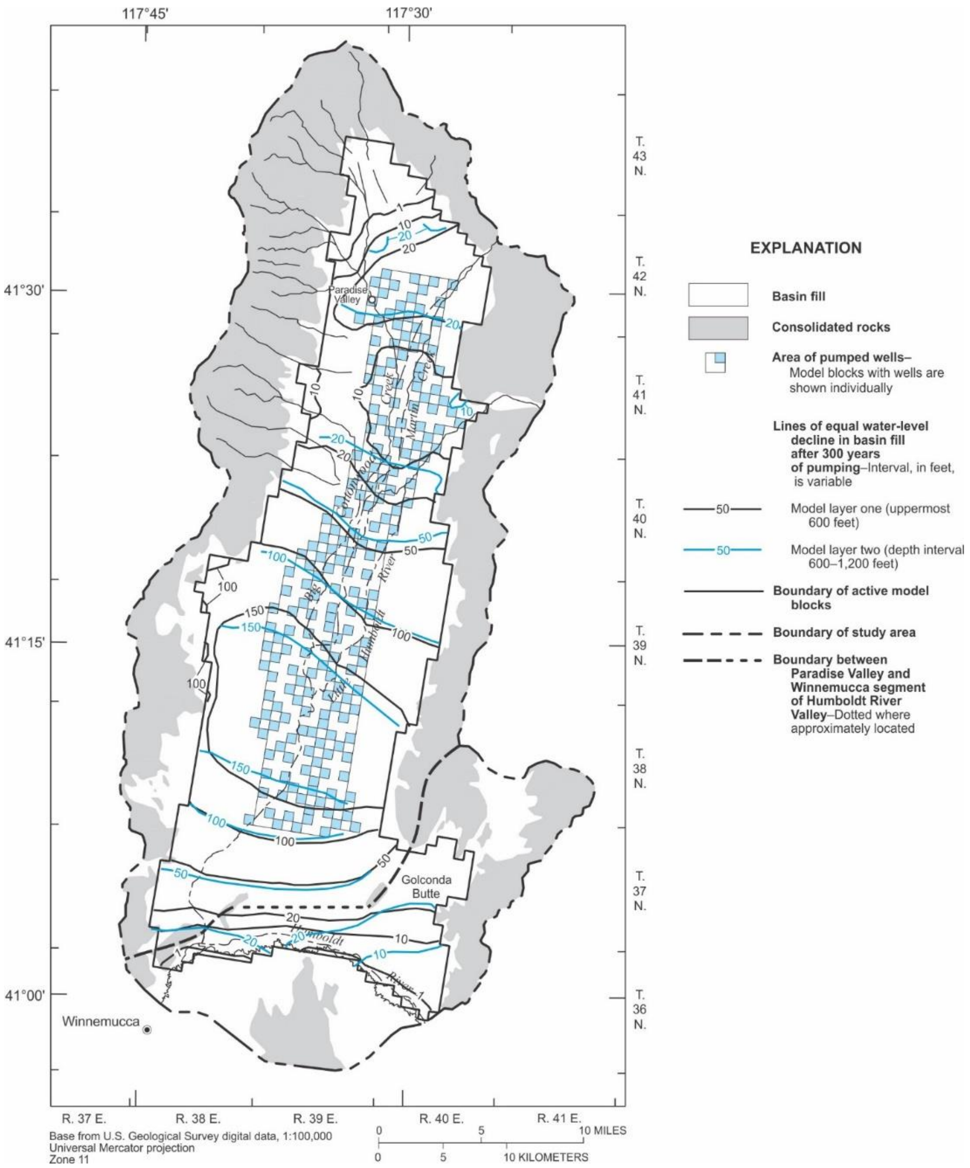

Figure 37 - Drawdown after 300 years resulting from scenario with pumping 72,000 ac-ft/yr (2.82 m^3/s) from the locations distributed widely in the area of evapotranspiration near the center of valley (from Prudic and Herman, 1996). Individual pumping wells are located in model cells shaded blue.

In the example presented, a simulated pumping period of 300 years does not quite produce a new steady-state (equilibrium) condition. At 300 years there is still approximately 3 percent of water being removed from the system coming out of storage. This hypothetical pumping in Paradise Valley also induces a net loss of approximately 12,400 ac-ft/yr (0.48 m^3/s) from the alluvial aquifer of the adjacent Humboldt River Valley. This is likely to be problematic because the additional loss of groundwater from the Humboldt River Valley alluvium will likely affect the flow of the Humboldt River, which is already fully appropriated. This is supported by observed trends in streamflow changes in the Humboldt River between the gage upstream of Paradise Valley (at Comus, Nevada, USA) and the next one downstream from Paradise Valley (near Imlay, Nevada, USA). During 1946-1975, this reach of the river consistently gained an average of 31 ft^3/s (0.88 m^3/s) in September, a month characterized by base flow dominance, but during 1987-2013, when groundwater pumping was substantially larger in Paradise Valley and nearby areas, the average September change in flow had been reduced to zero and the predevelopment gains in flow had been lost (D. Prudic, written communication., 2018 based on data from the USGS National Water Information System↗).

Even with these changes and consequences, from a hydraulic perspective the pumping scenario depicted in Figures 36 and 37 can be maintained indefinitely, as long as external factors (including climate change) do not change the average flows and capture potential from incoming and adjacent streams and additional pumping wells are not developed in the valley.

7 SUMMARY AND CONCLUDING REMARKS

Groundwater is a critically valuable resource for water supply around the world. But the development of groundwater resources has consequences -- its stock or reserves (i.e., the volume in storage in the aquifer) can be reduced and groundwater withdrawal can deplete surface-water flows and resources and have other environmental impacts. Specifically, new groundwater withdrawals through wells will have to be balanced by some combination of (1) a reduction in the volume of groundwater in storage in an aquifer, (2) an increase in recharge to the aquifer, and (3) a reduction in the discharge from the aquifer. The latter two constitute “capture,” and capture often manifests itself largely as a reduction in streamflow. The balancing between pumping and the three listed factors is dynamic, and it changes over time. If the groundwater storage depletion over time becomes negligible, then groundwater withdrawals are maintainable indefinitely (as long as other factors do not affect the aquifer’s water balance). However, even if groundwater withdrawals are maintainable from a hydrologic perspective, one must also consider its impacts on surface-water resources (and the timing of those impacts) because the changes in groundwater recharge and discharge (capture) may diminish those surface-water resources, and thereby potentially have economic, legal, political, and environmental consequences.

Historically, large-scale groundwater development and its subsequent effects (including storage depletion, capture, and land subsidence) have occurred in many areas prior to any recognition of its impacts or any consideration of its acceptability. Today, hydrogeologists have the knowledge and tools to understand and predict the magnitude and timing of these effects. This understanding and quantitative assessments can provide reliable scientific input to water managers and policy makers.

8 EXERCISES

The following exercises are based on the "Case Study 1: Hypothetical Stream-Aquifer System" described in this book. Some will require the application of MODFLOW to simulate the groundwater flow system, and the MODFLOW software is included in the online Supplemental Materials for this book↗; links to additional software are also provided.

Exercise 1) Effects of Well Location (Distance from Well to River)

Consider the hypothetical desert-basin, stream-aquifer system, as illustrated in Figure 17. Does the position of the pumping well relative to the stream affect the response of the system? Specifically, how does it affect (1) the magnitude and timing of the effect of pumping on surface water, and (2) the relative sources of water to the well? What is the nature of these effects? Consider two alternative well locations -- one further from the river and one closer to the river. Are any or all of these pumping scenarios sustainable? If you were a water manager for this stream-aquifer system, which well location would you consider preferable, and why? And are there any tradeoffs, such as the amount of drawdown in the pumping well?

Suggested approach: Download the files for the Case Study 1 model as described in Box 3. To determine the effect and importance of the distance between the well and a stream, if any, generate and simulate two additional variants of the base case in which the well position is changed. Move the location of the pumping well a distance halfway closer to the river (to the east) for one variant, and then also halfway closer to the distal "mountain" boundary (to the west). For these two new cases, the distances from the well to the center of the river are 4 km and 20 km, respectively. In the original "base case," that distance was 8 km (Figure 17). Plot the depletion and capture fractions over time for 200 years. The analyses can be completed using public domain United States Geological Survey software.

Click for solution to exercise 1↴

Exercise 2) Lower Ratio of Streamflow to Pumping

The Base Case analysis of Exercise 1 assumed that the streamflow was much greater than the well pumpage. What if the pumping rate was higher than in Exercise 1 and the ratio of streamflow to pumping (withdrawal) rate was much lower? In the Base Case, river inflow is about ten times greater than the well pumpage. Consider a case in which the well pumpage is increased by a factor of three (two more wells are drilled close to the original well to form a well field, and each well has the same pumping capacity, so the total Q in that cell of the model grid is -6,078 m^3/d) and the river inflow is reduced by a factor of three (Q_{in} = 6,667 m^3/d instead of 20,000 m^3/d). The ratio of streamflow entering the system to pumping out of the aquifer would then be about 1.1. How would that affect (1) the streamflow in space and time, (2) the drawdown in the aquifer, (3) the head distribution in the aquifer, and (4) the hydrograph for the pumping well? How does that affect (5) the water budget of the aquifer and (6) the fractional sources of water to the well? Is this pumping scenario sustainable?

Suggested approach: To determine the effect and importance of the relative strength of the pumping stress to the magnitude of streamflow, copy the Base Case input files into a new folder for Exercise 2 and modify the input parameters to match the assumptions of the above exercise. Specifically, you have to reduce the river's inflow (this requires changing the values "2.000000000000E+004" to "0.666700000000E+004" in the "Base.Case.sfr" input file and increase the pumping rate (by changing the value "-2.026000000000E+003" to "-6.078000000000E+003" in the last line of the "Base.Case.wel" input file). Also, consider adding another stream gage to monitor changes in streamflow (by (1) changing the number on the first line of the "Base.Case.gag" file to 2, (2) adding a third line to the Base.Case.gag" file "1 46 20206 1", and (3) adding a line to the "Base.Case.nam" file that says "DATA 20206 ..\Output.Files\Base.Case.sfrg2 REPLACE" after the similar line for the first gage).

Click for solution to exercise 2↴

Exercise 3) Analytical Solution for Streamflow Depletion

Estimate streamflow depletion using Glover's analytical solution (rather than a numerical model). How do these results compare with those from the numerical model used in the Base Case and in Exercise 1? Explain any differences.

Suggested Approach: Solve Equation 5 for each year of the 200-year simulation. There are a number of ways to do this, but one reasonable approach is to use formulas in an Excel spreadsheet. This exercise can be completed using information available in this book and the results for Exercise 1.

Click for solution to exercise 3⤵

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10 Boxes

Box 1 - Regional Unconfined Aquifer System: The High Plains Aquifer

The High Plains aquifer in the central United States provides a good example of applying the approach of estimating depletion volume by integrating head changes and storage coefficient estimates. Water-level changes in this regionally extensive unconfined aquifer have been estimated from many water-level measurements in a large number of observation wells over many years, producing maps of water-level changes since predevelopment times (Figure Box 1-1).

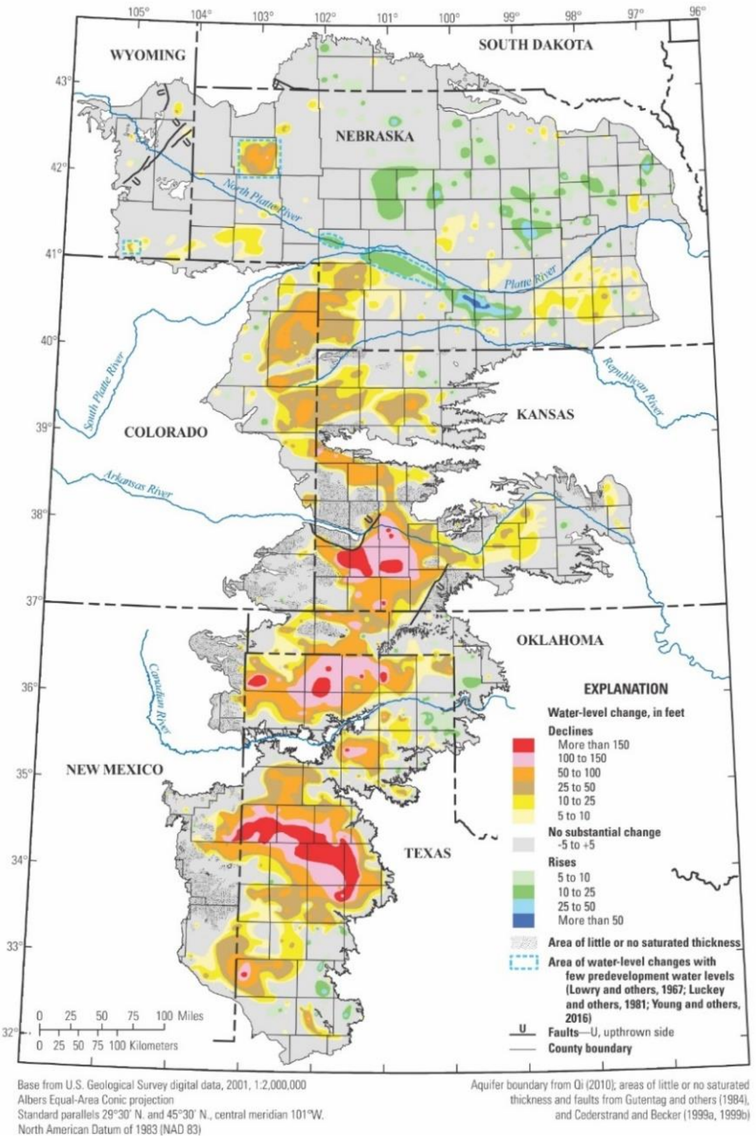

Figure Box 1-1 - Changes in groundwater levels in the High Plains aquifer, from predevelopment conditions (about 1950) to 2015 (from McGuire, 2017).

The area-weighted average specific yield varies by State from 0.081 in Wyoming to 0.185 in Oklahoma and is 0.151 overall for the aquifer (McGuire, 2017; Gutentag et al., 1984;

McGuire et al., 2012). McGuire (2017) accounts for this spatial variation and estimates that the total cumulative groundwater storage depletion in the High Plains aquifer since about 1950 is approximately 273.2 million acre-feet (337 km^3).

Return to where text links to Box 1↥

Box 2 - Storage Depletion in a Thick Confining Layer: Dakota Aquifer System

Konikow and Neuzil (2007) calculate storage depletion in the low-permeability beds that confine the regionally extensive Dakota Aquifer in South Dakota, USA (Figure Box 2-1), and that example is summarized here. The Dakota Sandstone and related sandstones in western-central North America form what is often considered a classic example of an artesian aquifer system. The Dakota Aquifer system is extensively developed and has played a particularly important role in the settlement and economic development of South Dakota. Study of the aquifer system began with Darton (1896; 1909) and helped shape current ideas about artesian aquifers (Bredehoeft et al., 1983).

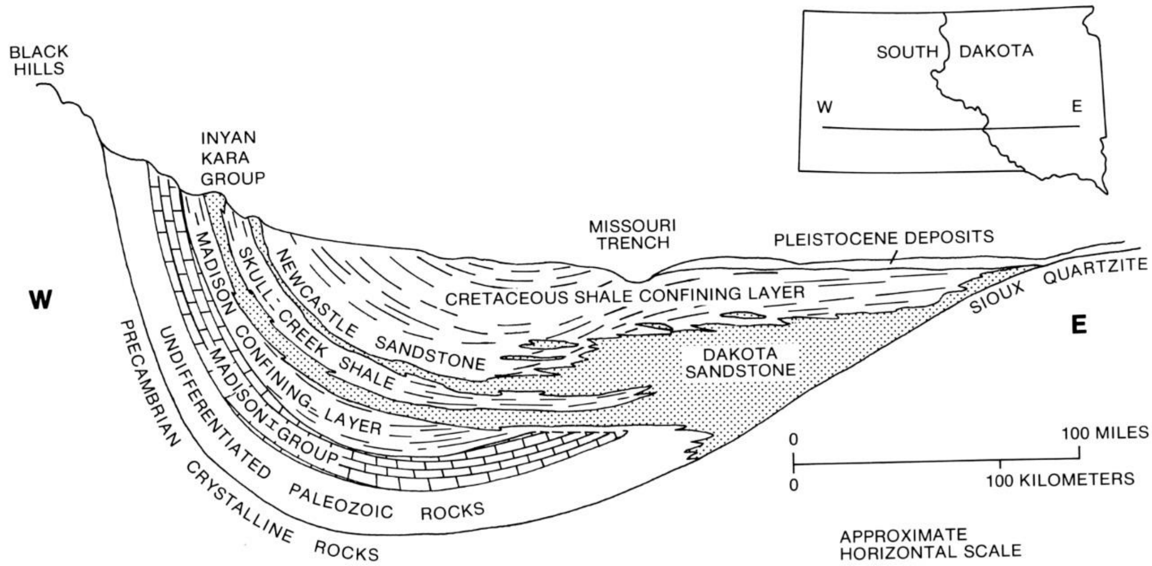

Figure Box 2-1 - Schematic east-west cross section of major aquifers and confining layers in South Dakota (*not to scale*) (modified from Bredehoeft et al., 1983). Vertical scale is greatly exaggerated.

The Dakota Aquifer underlies more than 171,000 km^2 of South Dakota as described by LeRoux and Hamilton (1985), though it also extends in several adjacent states. Significant recharge to the Dakota and the deeper Madison Aquifers occurs where they crop out on the flanks of the Black Hills (Figure Box 2-1). The Dakota aquifer discharges naturally at low elevations in the eastern part of the state. Discharge from pumped and flowing wells also has become an important source of discharge from the aquifer system (Case, 1984).

Substantial development of the aquifer system began by the early 1880s (Bredehoeft et al., 1983). By 1905, over 1,000 wells were producing water in the portion of South Dakota east of the Missouri River, supplying an estimated 1.2×10^6 m^3/d of water for irrigation and livestock (Bredehoeft et al., 1983). High rates of head decline in the Dakota Aquifer occurred before 1915. For example, eastern South Dakota experienced head declines averaging about 7 m/yr between 1909 and 1915. The rate of decline decreased to less than 0.5 m/yr by 1953 (Schoon, 1971). Estimated withdrawals stabilized at about 150,000 m^3/d by 1960 (Helgesen et al., 1984). Pumpage data presented by Bredehoeft et al. (1983), Helgesen et al. (1984), and Case (1984) indicate that the cumulative well discharge from the Dakota Aquifer system in South Dakota from the time before development until 1981 totaled about 19.7 km^3 of water. The history of development is incompletely documented, but Bredehoeft et al. (1983) estimate well discharge in 1912 as approximately 1.4 million m^3/d and then it declined dramatically to about 300,000 m^3/d in 1922; subsequently, it remained at rates less than half of the peak rate into the 1980s.

Groundwater storage depletion from the Dakota Aquifer system in South Dakota was evaluated using potentiometric maps showing predevelopment (Darton, 1909) and 1980 conditions (Case, 1984). The Inyan Kara, Newcastle, and Dakota Sandstones were treated as a single continuous unit forming the Dakota Aquifer. The spatial distribution of differences in head between the predevelopment and 1980 potentiometric surfaces indicated that the maximum decline in head was about 190 m and the average decline was 47 m. Storage coefficient values for the Dakota Aquifer ranged from 1.0×10^{-5} to 1.0×10^{-4} (Bredehoeft et al., 1983, table 3); a central value, 5.0×10^{-5}, was used to estimate depletion in this confined aquifer system. These data indicate that a total of about 0.4 km^3 of groundwater was derived from storage in the aquifer for the period from predevelopment through 1980, which represents about 2 percent of the estimated cumulative discharge of 19.7 km^3. The remaining 98 percent consists mostly of storage depletion in adjacent confining beds as water leaked from the confining beds into the aquifer.

Bredehoeft et al. (1983) used numerical models to analyze flow in the Dakota Aquifer system. They concluded that prior to development, most of the recharge and discharge occurred as steady-state leakage through the thick confining layers. Furthermore, their analyses indicate that since development, most of the water released from storage originated from the confining layers. Using Bredehoeft et al. (1983) model-calibrated estimate of specific storage for confining layers of 1.6×10^{-4} m^{-1}, Konikow and Neuzil (2007) estimated that the volume of water removed from storage in the confining units in South Dakota by 1980 was 14.9 +/-2.2 km^3, which represents approximately 76 percent of the estimated cumulative well discharge. This further implies that about 22 percent of the withdrawals are balanced by capture because only 2 percent was derived from storage in the aquifer. The capture includes both increased inflow (recharge) in or near upgradient outcrop areas and reduced outflow in downgradient discharge areas.

Return to where text links to Box 2↥

Box 3 – Running and Post-Processing the Model for Case Study 1

Supplemental material for running the Case Study 1 model is provided in CaseStudy1--Models.zip can be downloaded at the Groundwater Project↗. The material includes computer files for running the model described in the Case Study 1: Hypothetical Stream-Aquifer System. The files are designed to allow the reader to reproduce all the simulations and results related to the case study as presented in the Groundwater Project book "Groundwater Resource Development: Effects and Sustainability."

The zip file contains a folder named "Case Study 1 - Models" with subfolders that include all the input and output files for the three scenarios presented in this book. The subfolder "MODFLOW-NWT.Model" contains a copy of the executable software for MODFLOW-NWT (version 1.1.4), which was used in the analyses of this problem, as described in this book. It also includes a copy of the model documentation report. The computer source code for MODFLOW_NWT and additional documentation can be obtained by clicking here↗.

The Case Study includes three different scenarios, and the files for each are contained in separate subfolders, labeled (1) Base Case (No Recharge and No ET), (2) Low ET Case, and (3) ET and Recharge Case. Instructions for running a simulation for each case are provided below. The Base Case represents a scenario in which there is no areal recharge from precipitation and no evapotranspiration (ET) losses. The Low ET Case includes mountain front recharge but no areally diffuse recharge. The third scenario (ET.and.Recharge Case) includes both ET discharge (at a higher rate than in the previous case) and areal recharge from precipitation.

The folder for each of the three scenarios includes two subfolders. One subfolder contains all the input files needed to run that simulation and the other subfolder contains all the output files. The input file folder also includes the ModelMuse project file used to generate the input files (you do not need to use ModelMuse to run the simulation, although it is possible to do that). ModelMuse is a USGS public domain model pre- and post-processor↗. The input folder also contains a batch file that can be used to run the simulation using those input files.

Running the Model:

There are a number of alternative ways that the input files for each scenario can be run with MODFLOW-NWT. We offer one straightforward way that is consistent for the three scenarios. Specifically, we have placed a batch file ("*name*.bat") in each input folder (where "*name*" is the name of the scenario). Double-clicking on this batch file will cause it to execute a script contained within it. The scripts are written to link to the executable version of MODFLOW.NWT contained in the "Model" folder, start executing it, and provide it the name and location of the input files for each scenario. It will route all output files to the "Output" folder. Note that if you run (or re-run) the model with this folder and file configuration, the original output files will be overwritten and lost. If you want to save them for future comparisons, then you will first need to either rename the previous output files or move them to a separate new folder before running the simulation.

<u>Creating head (water level) and drawdown contour maps using ModelMuse:</u>

To start, if you have not already installed ModelMuse, download and install it by visiting the United States Geological Survey, ModelMuse web site: https://www.usgs.gov/software/modelmuse-a-graphical-user-interface-groundwater-models↗.

Once it is installed, go to the Case Study 1--Models folder, then down to the Base Case folder, then into the Input.Files folder and double click on the file “Base.Case.gpt” to open it in ModelMuse. You may want to stretch the ModelMuse window so you can see the entire model grid as shown in Figure Box 3-1. Model rows extend from left to right starting with row 1 at the top; model columns extend from top to bottom starting with column 1 on the left. The main window is a plan view of the grid, the lower window is the front-view (i.e., the cross section of the row highlighted in green on the plan view), and the window on the right is the side-view (i.e., the cross section of the column highlighted in blue on the plan view). The blue lines on the plan view indicate the model column shown on the right and the green lines indicate the model row shown at the bottom.

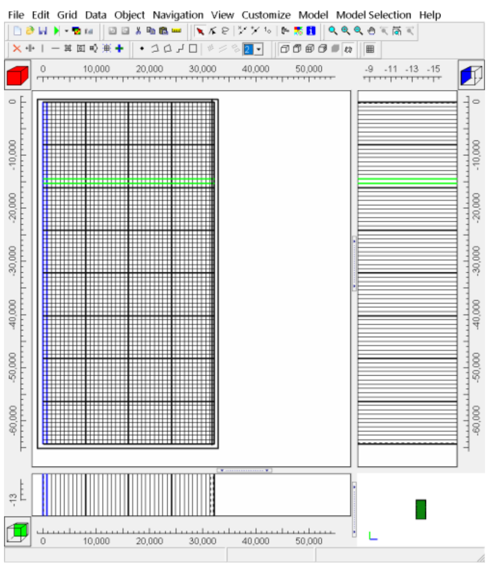

Figure Box 3-1 - Case Study 1 model grid displayed in ModelMuse (Konikow and Bredehoeft, 2020).

If you have not run the model directly from ModelMuse, then you need to import the data set you wish to contour. Under “File,” select “Import >Model Results” and navigate to the Case Study 1--Models folder, the Base Case folder and the Output.Files folder and select the file Base.Case.fhd and choose Open. A list of all the model simulation times for which the file contains head data will be shown and by default the last time (73051 days) will be selected. You can set “Display choice” to “Contour grid” and select OK and click on “Update the existing data sets with new values” to view a contour map of those heads, or you could scroll down to uncheck that and choose any other time. Figure Box 3-2 shows the head distribution for "Head: Period 1; Step: 1; Total Time: 1" which is a contour map of head for the undeveloped conditions. To change the default contours, open the Data Visualization dialog box and select Contour Data from the list on the left, then in the upper right enter a different contour interval and click apply and close. The drawdown distribution at a time that is 73,051 days after pumping began can be viewed by importing data from that time from the Base.Case.fdn file (Figure Box 3-3). Drawdown is the difference between the heads at two different times, typically predevelopment and a given time after pumping began.

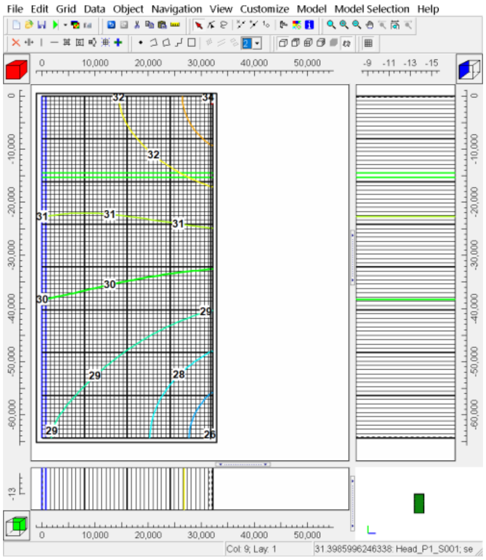

Figure Box 3-2 - Head contours for undeveloped conditions (Konikow and Bredehoeft, 2020).

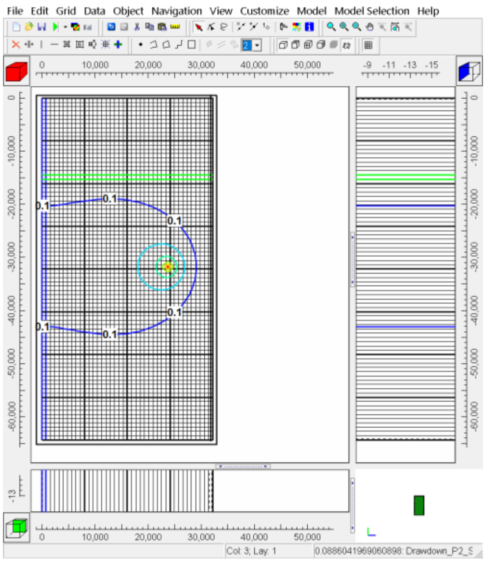

Figure Box 3-3 - Drawdown contours after 200 years of pumping (Konikow and Bredehoeft, 2020).

To view smaller magnitudes of drawdown further from the well, we suggest adding two new contours by opening the Data Visualization dialog box and selecting Contour Data from the list on the left, then click Specify Contours in the upper right. Increase the number of rows from 6 to 8. Then scroll down and set the contour values for the two new rows to 0.01 and 0.05. Next click "OK." and "Apply" and then "Close" in the Data Visualization dialog box. There are many options available for coloring and labeling the contours; feel free to experiment. Zooming in on the area around the pumping well shows that the drawdown at the well is on the order of 0.6 m and along the river is close to (but less than) 0.01 m.

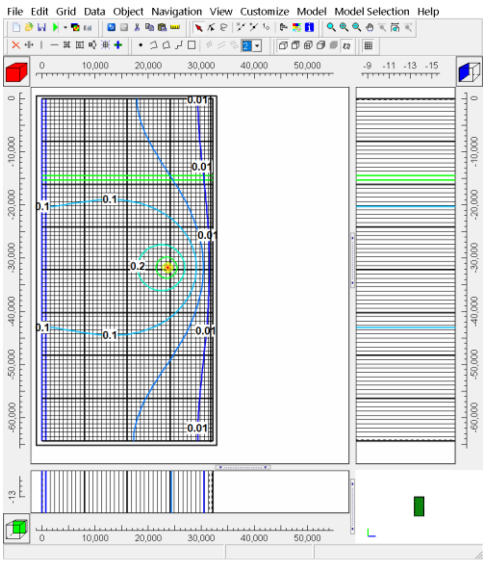

Figure Box 3-4 - Drawdown contours after 200 years of pumping including two contour lines for smaller drawdown (Konikow and Bredehoeft, 2020).

To remove grid lines, under View, select "Show or hide 2-D Grid>Show Exterior." To improve clarity of drawdown values and contours you can also select "Hide all objects" under the "Objects" pull-down menu. To export and save the image of the contour map, go to File>Export>Image (or click the camera icon). In the resulting dialog box, you have several options, but just click "Save image" to generate a file (select format or type) with the contour map. Click "Close." If you prefer that the contours be a single color, you can change the color scheme to "Blue only" or "Black only" in the Data Visualization dialog box. The resulting image is shown with black contours in Figure Box 3-5.

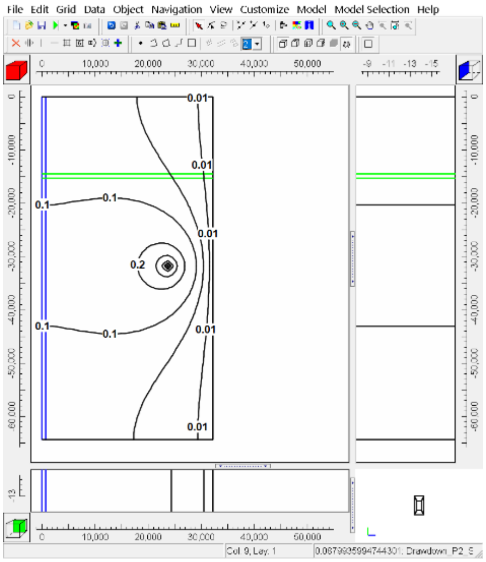

Figure Box 3-5 - Drawdown contours with the grid and object removed from view (Konikow and Bredehoeft, 2020).

Extracting and plotting MODFLOW budget data using GW_Chart software:

GW_Chart can be used to conveniently extract the budget data from the main MODFLOW or MODFLOW-NWT output file. If you have not yet installed GW_Chart, download and install it by visiting the United States Geological Survey, GW_Chart web site:

https://www.usgs.gov/software/gwchart-program-creating-specialized-graphs-used-groundwater-studies↗.

Once it is installed, open "GW_Chart." Under pull-down menu for "Chart Type/Convert," select "Water Budgets." Then in the lower right area, select "MODFLOW" under "Read Data From." Next, under "File," select "Open," and then navigate to the output folder for the Base Case and select "Base.Case.lst" to open that file and allow GW_Chart to read all the water budget data. This will generate plots for all selected variables for either cumulative or rate budgets. However, these plots are not at a high resolution and do not provide the numbers.

Therefore, we want to Save/export data using the lower-right middle button (); name the output text file (e.g., "Base.Case.Budget.txt") and select a destination folder. This will generate a text file that contains all the saved budget data for the model (as specified in the "output control" [".oc"] input file).

Open the "Base.Case.Budget.txt" file with Notepad or other word/text processing software. The file lists "cumulative" data first, followed by a listing of "rate" data. For this exercise, we will work with annual rates. Select all lines in the "RATES" category (bottom half of the listing (lines 206 to 408), and then select "COPY." Then open a blank Excel workbook, select the upper left cell (A1) and paste using the "Text Import Wizard." Click "next" and then select delimiters that clearly and properly separate the data columns (specifically, select "tab" and deselect "space"; the latter step will assure that the headers in row 2 align correctly with the proper data columns below). Then click "next" and "finish" to complete the import process. You can give the worksheet a name (e.g., by changing default name of "Sheet1" to "Base.Case" in the lower left tab). Save the spreadsheet file to the Output.Files folder, giving it an appropriate name (e.g., "Rate.Budgets").

Examine the headers in Row 2. If these column labels did not line up accurately with the data columns, adjust them manually for improved clarity in the spreadsheet. For improved clarity, select (highlight) all of Row 2 and then click on "Wrap Text" to see complete labels. You can also adjust column widths and number formats for data as desired. After completing these several steps, the first 11 rows (out of 203 Rows) and 17 columns (of 19) should look something like Figure Box 3-6.

	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	P	Q	R	S
1	RATES	units =	m^3/d																
2	Stress Period	Time Step	Time	In: STORAGE	In: CONSTANT HEAD	In: WELLS	In: ET	In: RECHARGE	In: STREAM LEAKAGE	In: TOTAL IN	Out: STORAGE	Out: CONSTANT HEAD	Out: WELLS	Out: ET	Out: RECHARGE	Out: STREAM LEAKAGE	Out: TOTAL OUT	IN - OUT	Percent Discrepancy
3	1	1	1.0	0.00E+00	0	1687.88	0.0	0.00E+00	5.842E+03	7.530E+03	0	0	0	0	0	7.534E+03	7.534E+03	-3.537	-0.05
4	2	1	366.3	1.98E+03	0	1687.88	0.0	0.00E+00	5.852E+03	9.524E+03	0.1836	0	2026	0	0	7.501E+03	9.527E+03	-3.572	-0.04
5	2	2	731.5	1.90E+03	0	1687.88	0.0	0.00E+00	5.878E+03	9.470E+03	0.1348	0	2026	0	0	7.447E+03	9.473E+03	-3.552	-0.04
6	2	3	1096.8	1.81E+03	0	1687.88	0.0	0.00E+00	5.921E+03	9.414E+03	8.69E-02	0	2026	0	0	7.392E+03	9.418E+03	-3.577	-0.04
7	2	4	1462.0	1.71E+03	0	1687.88	0.0	0.00E+00	5.972E+03	9.366E+03	3.81E-02	0	2026	0	0	7.344E+03	9.370E+03	-3.526	-0.04
8	2	5	1827.2	1.61E+03	0	1687.88	0.0	0.00E+00	6.026E+03	9.327E+03	2.15E-02	0	2026	0	0	7.304E+03	9.330E+03	-3.519	-0.04
9	2	6	2192.5	1.53E+03	0	1687.88	0.0	0.00E+00	6.076E+03	9.293E+03	1.27E-02	0	2026	0	0	7.270E+03	9.297E+03	-3.541	-0.04
10	2	7	2557.8	1.45E+03	0	1687.88	0.0	0.00E+00	6.123E+03	9.264E+03	2.93E-03	0	2026	0	0	7.241E+03	9.267E+03	-3.507	-0.04
11	2	8	2923.0	1.39E+03	0	1687.88	0.0	0.00E+00	6.164E+03	9.239E+03	2.93E-03	0	2026	0	0	7.216E+03	9.242E+03	-3.522	-0.04

Figure Box 3-6 - Excel spreadsheet with rates from budget for the Base Case of Case Study 1 (Konikow and Bredehoeft, 2020).

To assess the sources of water to the well, and how these change with time, we need to complete a few more calculations, which are aided by the use of formulas in Excel. We will examine these changes over the 200-year simulation period, so it will be convenient to create a column for time in years (because the model units of time are days) so that variables can be plotted in terms of years. Therefore, in Column T, add a label in Row 2 (something like "Time, in years"). Then in Row 3 of Column T, add a formula to convert time in days (Column C) to time in years (assume one year equals 365.25 days). Copy and paste that formula ("=(C3/365.25)-0.00273785") into the remaining cells of Column T. (Recall that the length of the initial steady-state stress period is arbitrary, and was set at 1 day, thus the subtraction of approximately 0.00273785 years) to account for the steady state portion of the simulation.

One source of water to the well is from a change in storage in the aquifer. In the MODFLOW budget terminology, "In: STORAGE" refers to the water that enters (flows into) the groundwater system by <u>coming out of storage</u> in the aquifer. We will use Column U to compute the net change in aquifer storage. So, in Row 2 of Column U, add a label "Net Change in Storage" or something similar. In <u>Row 3 of Column U</u>, insert a formula to compute net change in storage as the difference between the value in Column D and that in Column K ("=D3-K3"). Then copy that formula and paste it into every remaining cell in

Column U. With this order of subtraction, the results with a positive sign will represent a reduction (or depletion) of storage.

Next, we want to calculate capture, which in this case can only include the capture of streamflow (1) by increasing the seepage losses from the stream ("In: STREAM LEAKAGE"), which equals recharge to the aquifer, and/or (2) by decreasing the discharge from the aquifer to the stream ("Out: STREAM LEAKAGE"). So, we need to calculate how these change with time. First, we compute the increase in recharge from the stream in Column V. Add a label in Row 2 of Column V (something like "Increased Seepage Loss from River"). The increase in seepage loss in any time step during the transient stress period equals the difference between the "In: STREAM LEAKAGE" during that time step and the respective value during the initial steady-state stress period when there was no pumpage from the well. Set up a formula for that column to compute these values (the formula should look something like: "=I4-I$3" in Row 4 of Column V). Follow a similar procedure for "Decreased Groundwater Discharge to River" by entering "=P4-P$3" in Row 4 of Column W). Then in Column X, calculate the Capture by adding the absolute values of Columns V and W, that is "=ABS(V4)+ABS(W4)" in Row 4 Column X. Finally, we want to compute the nondimensional fractions of the sources of water to the well. That is, we want to compute the storage depletion and capture fractions. Use Columns Y and Z to calculate the storage depletion fractions and capture fractions, respectively, by dividing Columns U and X by the well pumpage (Column M). Note that the sum of these two columns should always equal 1. Plot the change in the storage depletion and capture fractions over the 200-year simulation period (using Excel or your favorite graphic/plotting software package). The graph should look like Figure Box 3-7.

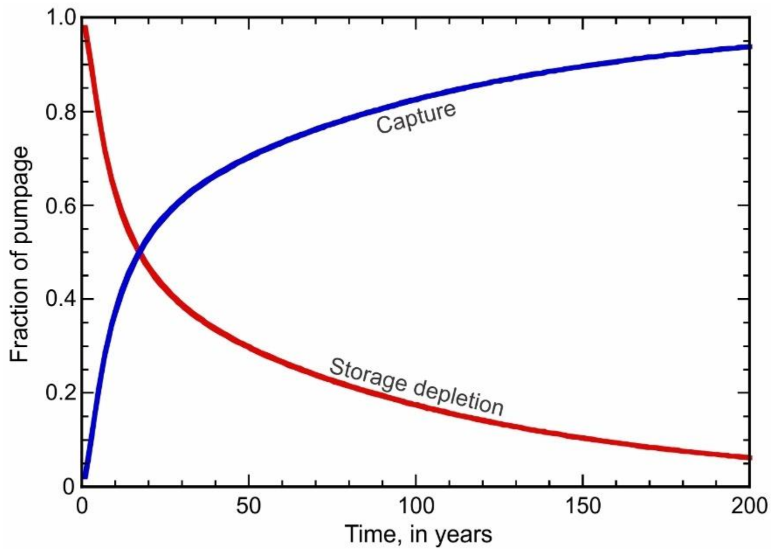

Figure Box 3-7 - Fractional sources of water entering the pumping well for the Base Case of Case Study 1 (Konikow and Bredehoeft, 2020).

Computing streamflow changes in Excel

To determine how streamflow varies in a downstream direction, first use a text editor to open the main listing (output) file "Base.Case.lst" from the Base Case simulation (folder: "Case Study 1--Models\Base Case\Output.Files", file "Base.Case.lst"). Find the stream listing data at 200 years (time step 200 in stress period 2) by searching for "STREAM LISTING PERIOD 2 STEP 200", which will be just above the last budget print out at the bottom of the file. Then copy the label lines and data (a total of 83 lines) and then paste it into cell A1 of an Excel spreadsheet. Next click on Data>Text to Columns, and choose "fixed width", then "next", then "finished." Delete all data except the stream reach number and the flow out of the reach (these are Columns E and H of the pasted data). Now, the reach number should be in column A in sequentially increasing order. Insert two new columns after column A to compute the distance downstream in meters and kilometers, respectively. Use formulas based on the knowledge that each reach covers a distance of 804.67 m. Column D should then include the streamflow out of each reach (in m^3/d) after 200 years of pumping. Next use Excel or other plotting or graphic software to plot the results, which should look like Figure Box 3-8.

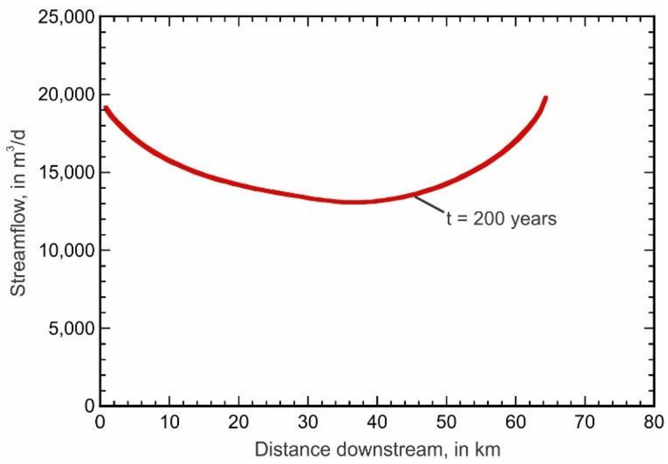

Figure Box 3-8 - Streamflow with distance downstream for the Base Case of Case Study 1 (Konikow and Bredehoeft, 2020).

To examine the stream losses and gains, add Column E with the label "Stream Loss(-) Gain(+)". In cell E4 enter the formula "=D4-20000" because there is a stream inflow of 20,000 m^3/d at the upstream end of the stream. In cell E5 calculate the difference between outflow in the reach and outflow in the upstream reach "=D5-D4" and copy the formula down the column. Because we are subtracting the flow at the upstream end of the reach from the downstream end, a negative value indicates stream loss in the reach and a positive value indicates gain. The result is shown in Figure Box 3-9.

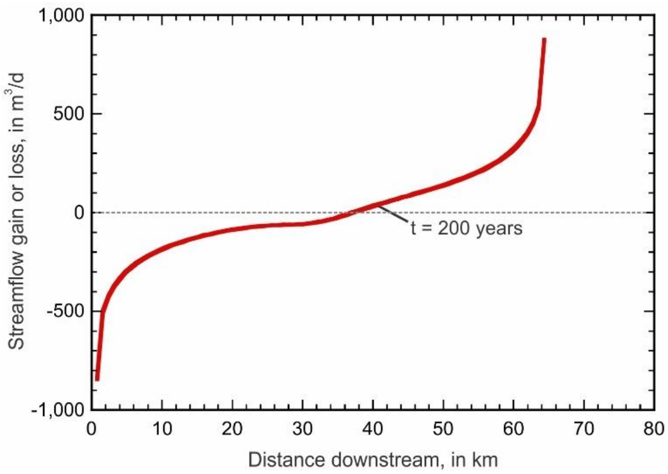

Figure Box 3-9 - Streamflow Gain(+) and Loss(-) with distance downstream for the Base Case of Case Study 1 (Konikow and Bredehoeft, 2020).

To determine the impact that 200 years of pumping had on the streamflow as well as the gains and losses, we want to add the predevelopment data to the previous plots. Follow the above steps for the predevelopment conditions by searching in the same "Base.Case.lst" file for "STREAM LISTING PERIOD 1 STEP 1", which will be close to the top of the file (for us it is line 1675 of the Base.Case.lst file.) As done before for 200 years, copy the data into the spreadsheet. Make similar calculations as done for the 200-year time and add the data to the graphs. They should look like Figure Box 3-10.

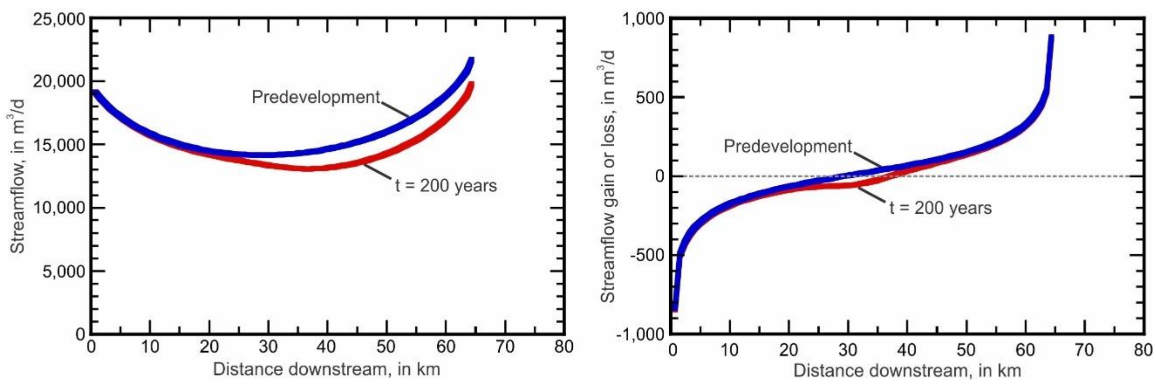

Figure Box 3-10 - Streamflow and stream Gain(+) and Loss(-) with distance downstream for the Base Case of Case Study 1 (Konikow and Bredehoeft, 2020).

<u>Plotting well hydrographs from MODFLOW output using GW_Chart:</u>

GW_Chart can be used to extract and plot water-level changes over time at individual nodes (cells) of the MODFLOW grid. This produces a well hydrograph showing either water levels (heads) or drawdown (changes in water level from an initial value), as selected by the user.

To do this, first if you have not yet installed GW_Chart, download and install it by visiting the United States Geological Survey, GW_Chart web site:

https://www.usgs.gov/software/gwchart-program-creating-specialized-graphs-used-groundwater-studies↗.

Once it is installed, open "GW_Chart." Then, under the pull-down menu for "Chart Type/Convert," select "Hydrographs." In the Data box, make sure that "MODFLOW head or drawdown file" is selected. Next, set Column-Row-Layer to the cell location for the pumping well (30,40,1 for the Base Case). Then under "File," select "Read Heads or Drawdown," and then navigate to the output folder for the Base Case and select either the formatted head file or the formatted drawdown file. We chose to view drawdown.

To get a better-quality graph with a time scale in years, we need to place the data into a spreadsheet or plotting package. You can select (highlight) all the data in the two columns of data listed in the upper middle box of GW_Chart, and then copy it (Ctrl-c). Next, paste it into cell A1 of a blank Excel worksheet. Alternatively, under the File pull-down menu, you can select "Save Heads or Drawdowns" to save the two columns of data to a separate new file. The first time shown is "1", which represents the arbitrary length of the initial steady-state stress period. There is no drawdown for this stress period, but MODFLOW uses the first period as the reference time and the final converged head is not known at the start of the calculations, so it calculates the change in head from the initial conditions to the converged steady state conditions. Therefore, change the drawdown value for the first time "1" to a value of 0 in Column B. Add a header (or label) to column C of the spreadsheet "Time, in years" and then define the values in that new column by a formula to convert time in days (Column A) to time in years (for example, the formula for C3 would be "=(A3-1)/365.25"). Copy and paste that formula into rows 2 through 202 of Column C. Then plot Column C on the horizontal axis and Column B on the vertical axis to produce the hydrograph as shown in Figure Box 3-11.

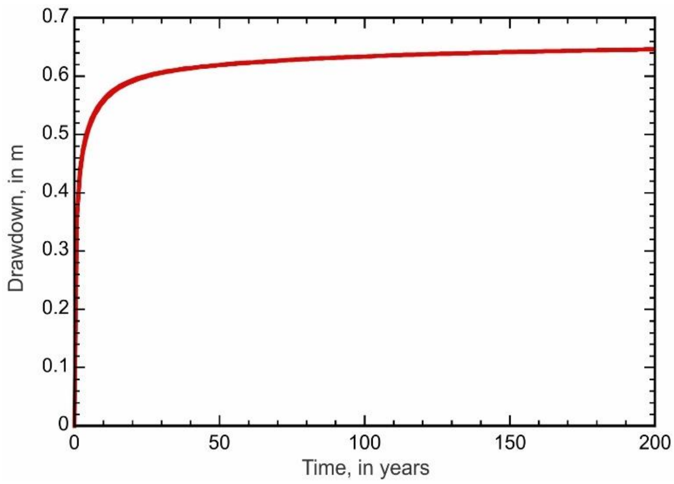

Figure Box 3-11 - Drawdown at the pumping well location for the Base Case of Case Study 1 (Konikow and Bredehoeft, 2020).

Streamflow changes with time are recorded in the Gage output file "Base.Case.sfrg1". In this case, we defined a stream gage as being located at the most downstream cell of the river (reach

80), representing surface water outflow from the system being simulated. Streamflow at the gage for the Base Case is illustrated in Figure 3 Box 3-12.

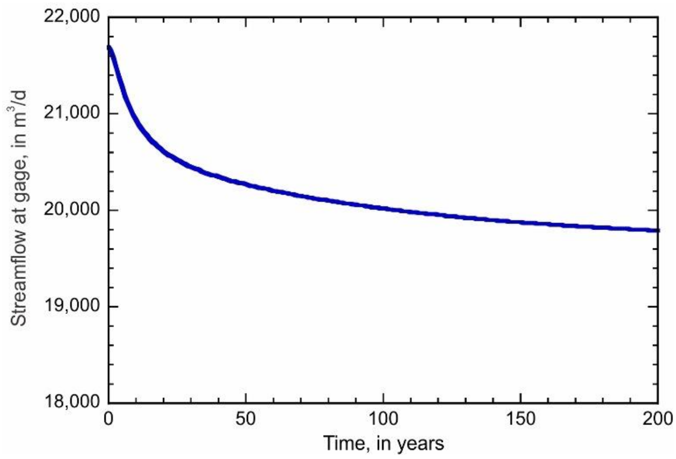

Figure Box 3-12 - Streamflow at model outlet for the Base Case of Case Study 1 (Konikow and Bredehoeft, 2020).

Return to where text links to Box 3

11 Exercise Solutions

Exercise 1 Solution: Effects of Well Location (Distance from Well to River)

The problem:

Exercise 1 requests that you consider the hypothetical desert-basin, stream-aquifer system, as illustrated in Figure 17 and answer the following questions. Does the position of the pumping well relative to the stream affect the response of the system? Specifically, how does it affect (1) the magnitude and timing of the effect of pumping on surface water, and (2) the relative sources of water to the well? What is the nature of these effects? Consider two alternative well locations—one further from the river and one closer to the river. Are any of these pumping scenarios sustainable? If you were a water manager for this stream-aquifer system, which well location would you consider preferable, and why? And are there any tradeoffs, such as the amount of drawdown in the pumping well?

How to run and analyze the model results:

If you have not already done so, it is useful to read Box 3, then run and post-process the results of the Base Case model of Case Study 1 before undertaking Exercise 1. To do this, first put the input files, MODFLOW-NWT executable code, and ModelMuse files for the Base Case of Case Study 1 on a Microsoft-OS computer by downloading the zip file "CaseStudy1--Models.zip" from the online Supplementary Information for this book↗. Extract the "Case Study 1" folders and subfolders onto your personal computer. Then go through the steps described in Box 3.

Acquiring a file folder for Exercise 1:

Next download the zip file "Exercise1.zip" from the online Supplementary Information for this book↗. To get you started, we have already copied the Base.Case input files into two new folders under the folder "Exercise 1" (Well.Closer.to.River.Case and Well.Further.from.River.Case); you can use these to simulate the cases with the well at two alternative locations -- one closer to the river and the other further from the river. However, the files have not been modified yet, so if you execute either of these simulations without changes, you will get the base case result. We suggest you work from these two folders to analyze and develop solutions to Exercise 1. For convenience, we have also installed a copy of the executable code for MODFLOW-NWT in a location that will work with the batch files in these two folders.

Modifying Input Files to move the well:

Next, modify the input files in the Input.Files subfolders of the two folders "Well.Closer.to.River.Case" and "Well.Further.from.River.Case" to move the position of the well closer to and further from the river, respectively. For each variant, it might be easiest to modify the input file for the WEL Package manually using a text editor, such as Notepad. In this approach, you only have to modify the column coordinate in the pumping well data, which is the very last line of the file "Base.Case.wel", which shows the well in row = 40 and column = 30. The stream is in column 40 so the new well will be half way between the base case well and the stream if it is placed in column 35. The mountains are to the left of column 1 so the new well will be half way between the base case well and the mountains if it is placed in column 15.

If you change the name of this file [or of any other input file], then you also have to make a corresponding change to the name of that file within file “Base.Case.nam”. If you change the name of the “name” file, then you have to make a corresponding change to its name in the file “Base.Case.bat”. You can run the modified problem by double-clicking on the batch file (named “Base.Case.bat”), assuming the folder “MODFLOW-NWT.Model” is at the same level as we setup in the zip file “Exercise1.zip”, that is, at the same level as the folders “Well.Closer.to.River.Case” and “Well.Further.from.River.Case”.

Alternatively, if you have the ModelMuse software on your Microsoft-OS computer, or download it as described in Box 3, you can modify the position of the well by opening the ModelMuse project file “Base.Case.gpt” and adjusting the well position therein, exporting the input files, and running the program from within ModelMuse. You will also be able to use ModelMuse to visualize the groundwater heads and drawdowns using the instructions from Box 3 to contour data from the *name*.fhd and *name*.fdn files.

Post-processing the new model results:

Box 3 provides step-by-step instructions for analyzing and plotting the water budgets and hydrographs, preparing contour maps of hydraulic head and drawdown, and computing streamflow gains and losses for the Base Case. For Exercise 1 repeat that process for the models with the well closer to and further from the stream. Once you complete the analysis for the other model, it is useful to include the results for all three cases in the same graph to facilitate comparison. Data from the model output of this exercise are included in spreadsheets located in the “DataSpreadsheets” subfolder of the “Exercise 1” folder.

Analysis of impact of well position on streamflow:

In assessing the effects of pumping on streamflow, it is important to remember that the river is interacting with the aquifer even under natural predevelopment conditions. Specifically, the river is losing water to the aquifer in the upstream reaches of the river and is gaining water from the aquifer in the lower downstream reaches of the river. This is clearly indicated by the predevelopment head distribution (Figure 20a). However, pumping the well disturbs these “natural” interactions.

After post-processing the streamflow result for the base case and the two cases with the well closer to and further from the stream, the results are compared at the end of the simulation (t = 200 years). The methods and steps used to calculate streamflow gains and losses are detailed in Box 3.

The results shown in Figure ExSol 1-1 indicate that the location of the well relative to the river has only a small effect (relative to the magnitude of the streamflow) on the streamflow profile along the river -- certainly a smaller effect than that of the pumping itself. At the most downstream point, where the river leaves the area of the model, the flow is essentially the same for all three well locations. The same is true at the halfway point down the river, which is the north-south position of the well in all cases. However, upstream from the halfway point, streamflow is highest for the case of the closer well and lowest for the case of the further well. The area upstream of the halfway point is where the river is losing water because of natural seepage losses as well as induced infiltration because of the drawdown caused by pumping. Downstream of the halfway point, the opposite is true—with higher streamflow when the well is located further from the river. In the downstream reach of the river, the streamflow is increasing because of groundwater discharge into the river. At a distance of about 40 km downstream (just past the midpoint), the difference in streamflow between the case with the well closer to the river and the case with the well further from the river after 200 years of pumping is the largest, at about 535 m^3/d, which is large relative to the magnitude of the well pumpage (Q = 2,026 m^3/d).

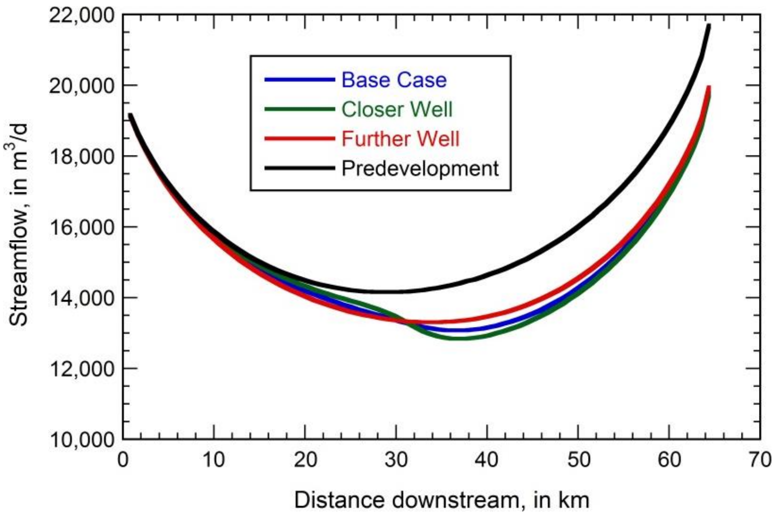

Figure ExSol 1-1 - Effect of well location on streamflow after 200 years. For comparison, streamflow is shown under steady-state predevelopment conditions (Konikow and Bredehoeft, 2020).

The stream gage at the outlet of the model represents surface water outflow from the system being simulated. Comparison of these data for the three well locations (Figure ExSol 1-2) show that most of the capture (and stream depletion) occurs during the early times. As might be expected, the fastest effect occurs when the well is located closest to the river. In this case, half the total depletion occurs in the first 15 years of pumping. But in the long run, the total depletion of streamflow is the same regardless of well location—it is a function of the overall water budget.

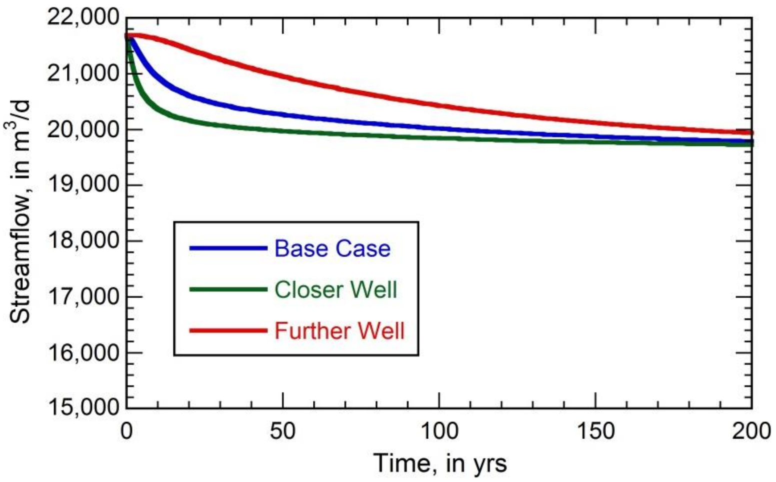

Figure ExSol 1-2 - Effect of pumping on streamflow as recorded at the stream gage at the outlet of the model during 200 years of pumping for three different well locations (Konikow and Bredehoeft, 2020).

It is also worth looking at the gains and losses in streamflow along the river, and how they change due to pumping from wells at different locations. The changes in streamflow gains and losses are determined by the differences between streamflow gains and losses during the steady-state predevelopment period and those during the transient pumping period. We evaluate those after 200 years of pumping; detailed steps are described in Box 3.

The streamflow loss is greatest in the first cell of the river (i.e., the most upstream reach) (Figure ExSol 1-3). For the predevelopment case, streamflow is diminished in the first reach of the river by 836 m^3/d out of an inflow of 20,000 m^3/d; the losses then decrease consistently in a downstream direction until they reach a minimum near the middle reach of the river. Then the streamflow starts to increase in reach 37 (about 30 km downstream), increasing exponentially to a maximum gain of 883 m^3/d in the most downstream reach of the river. Differences between predevelopment streamflow gains and losses and those after 200 years of pumping appear to be relatively small and only evident in the middle third of the river—closest to the pumping well. This indicates that most of the streamflow gains and losses are related to the natural predevelopment boundary conditions of the problem and that the changes caused by sustained pumping are small relative to that. Even though these changes are small, they still need to be considered as a real effect by water managers.

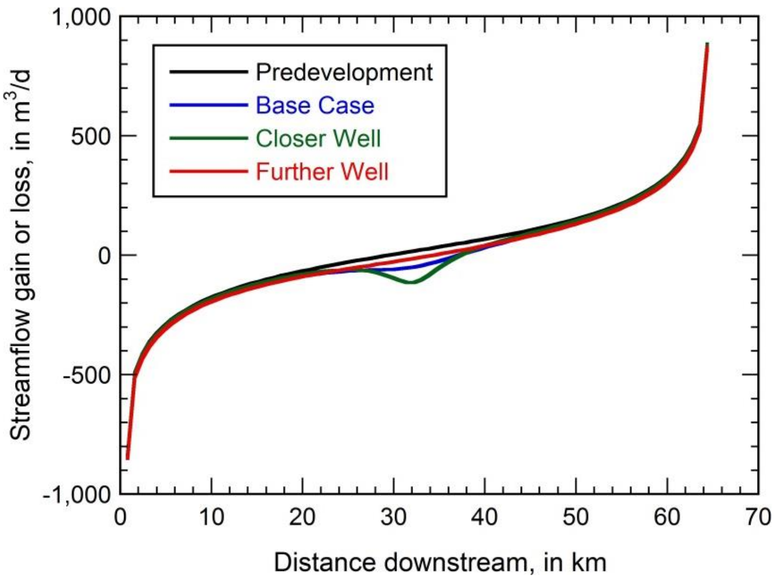

Figure ExSol 1-3 - Changes in streamflow along the length of the river for predevelopment conditions and after 200 years of pumping at three different well locations (Konikow and Bredehoeft, 2020).

The difference between the predevelopment values and those after 200 years for the three well locations represent the effects of pumping on streamflow (Figure ExSol 1-4). These results show that when the well is closest to the river, it will have the greatest magnitude of reducing streamflow per unit length of river (in this case, at a river location 32 km downstream from where it enters the valley). When the well is furthest from the river, its effects are smoothed out (or dampened) over a longer reach of the river.

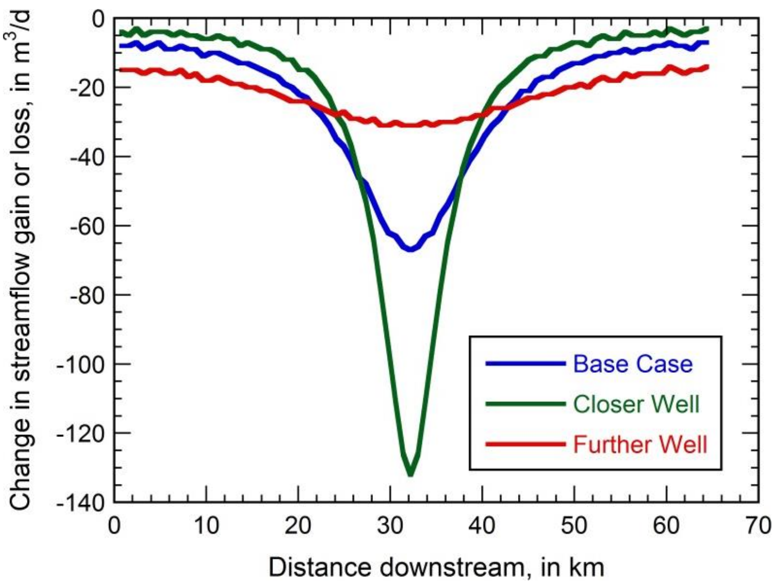

Figure ExSol 1-4 - Changes in streamflow gains and losses attributable to pumping for 200 years at three different well locations (Konikow and Bredehoeft, 2020).

Even for the well location that is closest to the river, the reduction in streamflow in the central reaches of the river never reach 1 percent of the streamflow. Thus, one can conclude that the effects of pumping on streamflow are small and tolerable given the benefits of the groundwater supply. However, one must also be cognizant that streamflow is variable in time and that droughts may cause low-flow periods in which streamflow is substantially less than the long-term steady rate assumed in this exercise. At those times, the impact of the well on streamflow would be proportionately greater.

Sources of Water to the Well:

As should be expected, the closer the well is to the river, the sooner the system becomes capture dominated and the greater are the near-term impacts on streamflow (and streamflow depletion). The aquifer's transition from storage-depletion dominated to capture-dominated, reflected in Figure ExSol 1-5 by the time location of the crossover of the two curves for each distance, increased from about 5 years for a distance of 4 km, to 18 years for a distance of 8 km, to 74 years for a distance of 20 km. Thus, where the well is positioned makes a substantial difference in the response of the system (and on the timing of its effect on surface water).

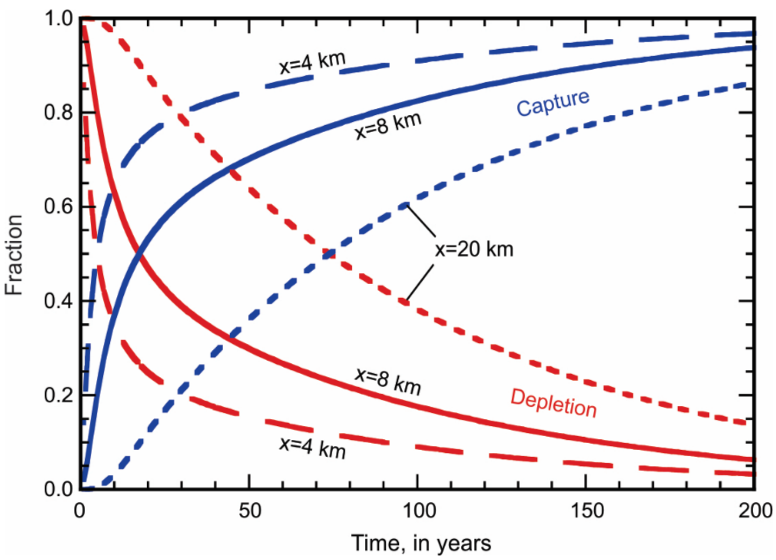

Figure ExSol 1-5 - Annual storage depletion and capture fractions for simulations of variants of the base case (Figure 17) in which the distance from the well to the stream is varied by placing it at a distance (x) of 4 km, 8 km (the base case), and 20 km from the river. Storage depletion fractions are shown in red and capture fractions are shown in blue (Konikow and Bredehoeft, 2020).

Sustainability:

The simulation model for all three well locations indicates that the pumping is sustainable for at least 200 years (at least from a strictly hydraulic perspective; in general, there may be other environmental considerations that render this too damaging to the environment to be considered "sustainable"). That is, as pumping continues over time, stream capture continues to increase. This means reduced streamflow, which may have unacceptable consequences from the perspective of downstream water users or ecological considerations. However, in this particular idealized case, that would not appear to be a major concern. However, these results also demonstrate that even "sustainable" groundwater development can cause streamflow depletion. Declining water tables also may impact the extent and duration of wetlands. During droughts or low-flow periods, the central reach of the river may dry up. Thus, "sustainability" must be evaluated from a broader perspective than just the physical ability to withdraw groundwater from the aquifer. These factors were not considered or evaluated in this simplified example.

If the pumping rate can be maintained until the system reaches a new equilibrium, then no further system changes will occur and the pumping can be sustained indefinitely based on hydraulic factors and processes. We can infer from the plots in Figure ExSol 1-5 that the system has not yet reached equilibrium in any of the three cases. That is, groundwater storage is still being depleted after 200 years of pumping, which means that water levels are still declining, which in turn means that the system has not reached a steady-state condition. To assess whether the system can eventually reach equilibrium (steady-state) conditions, additional simulations were run for longer periods of time. The results of these extended simulations yield water budget data indicating that a new hydraulic equilibrium can be reached in all three cases, although the time to equilibrium varies with the distance of the well to the stream (Figure ExSol 1-6). Equilibrium is reached soonest (in about 884 years) for the case where the pumping well is closest to the river (4 km), and the longest (in about 1,140 years) when the well is furthest from the river (20 km).

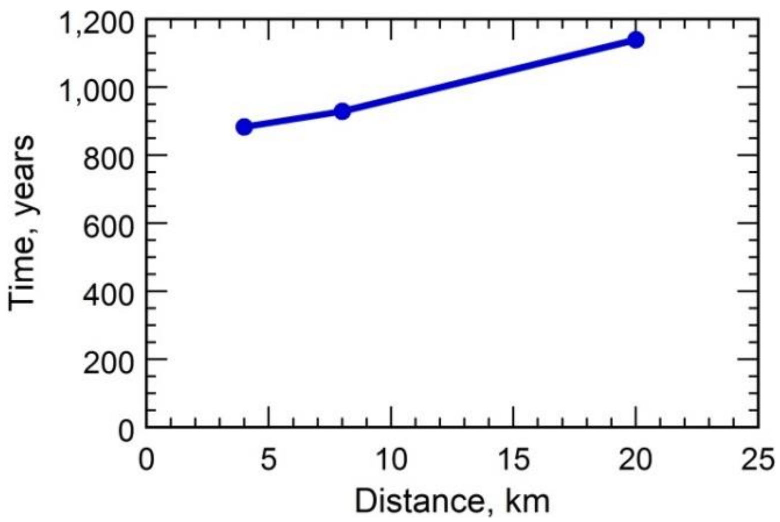

Figure ExSol 1-6 - Relation between time for model to reach a new steady-state condition and the distance from the pumping well to the river (Konikow and Bredehoeft, 2020).

Drawdown:

The drawdown in the pumping wells would be expected to vary for the three different well locations because each has a different distance to nearby barrier boundaries and recharge boundaries. These differences are seen in Figure ExSol 1-7, which compares calculated drawdowns in the pumping wells at the three different locations. Even though a new equilibrium condition has not strictly been achieved in 200 years for any of the three locations, the rates of change in head are very small at that time for all cases. The annual additional drawdown is less than 0.001 m/yr after just 16 years in the closer well, after 30 years in the base case, and after 94 years for the well location furthest from the river. These rates are very small compared to the average saturated thickness of the aquifer (150 m), so additional long-term drawdown would not be a major concern to water managers. The maximum difference in drawdown among the three cases is less than 0.3 m. Although pumping costs might be somewhat greater for the furthest well, which has the largest drawdown, the magnitude of this difference is quite small and would not likely be a major concern to water managers.

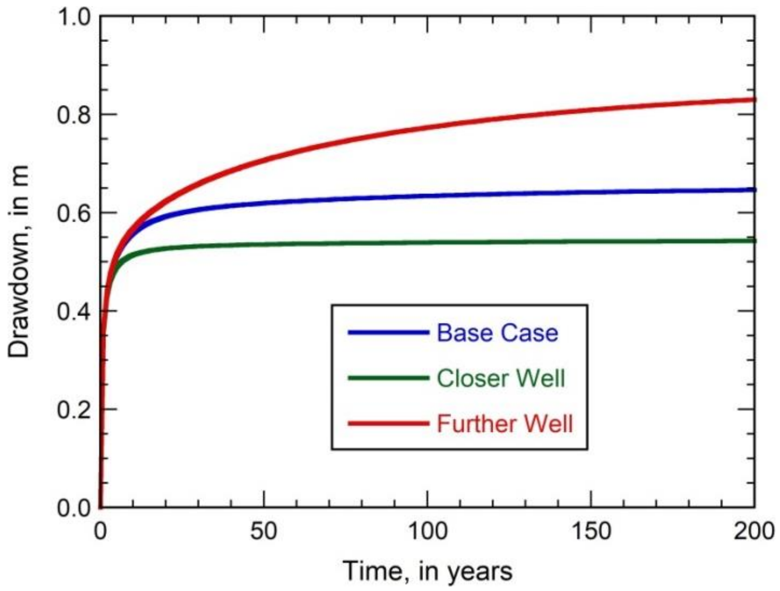

Figure ExSol 1-7 - Calculated drawdown in the pumping well for three different well locations (Konikow and Bredehoeft, 2020).

The drawdown in the aquifer can be mapped at various times to show how water levels change in time and space, or as a function of well location. The drawdown map for the base case after 200 years of pumping (Figure ExSol 1-8) indicates that the drawdown near the stream is only about 0.01 m or less, with the greatest drawdown in the middle reach of the river (which is also the closest point on the river to the well) and smaller drawdowns in either direction along the river away from the well location. However, even these small drawdowns are enough to affect the exchange of water between the river and the aquifer (as seen in Figure ExSol 1-3).

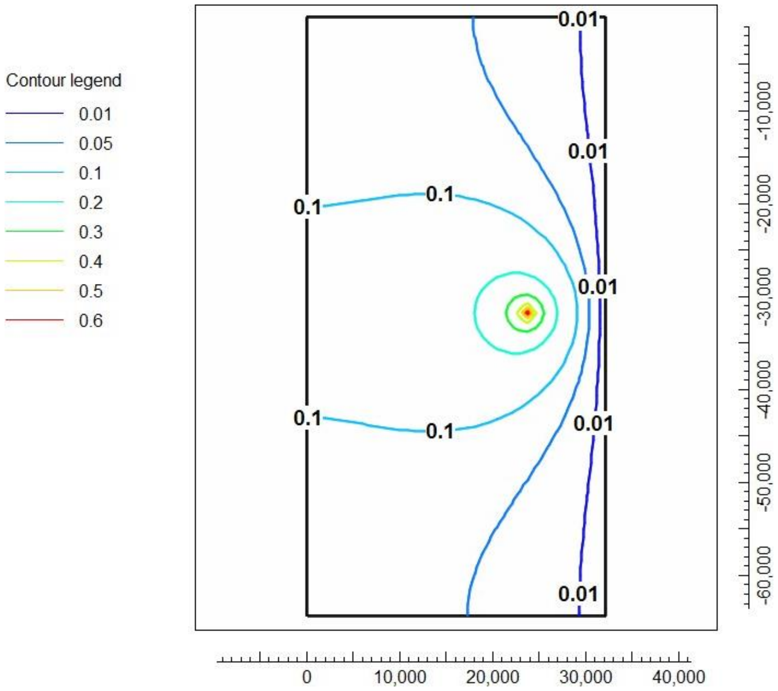

Figure ExSol 1-8. Contour map of drawdown (in m) in the aquifer after 200 years of pumping for the Base Case (Konikow and Bredehoeft, 2020).

Water Management Perspective:

The results indicate that the further the well is from the river, the greater the drawdown in the well. A lower water level increases the lift and energy costs to deliver the water. On the other hand, the further the well is from the river, the more the effect on streamflow is delayed and dampened. In this particular simplified hypothetical case, the streamflow is large relative to the well pumpage, and the pumpage is small relative to the transmissivity of the aquifer. Thus, the tradeoffs between well efficiency and streamflow capture are not dramatic and a water manager's decision is not clear or obvious. That is, in this particular case, it would not make a big difference whichever well location was selected. However, if all else were the same, and if seasonal streamflow variability were high, then preserving downstream water rights and protecting ecosystem water needs would indicate a preference for selecting the further well location. On the other hand, if hydraulic properties of the aquifer were less favorable and resulted in much greater drawdown for the same pumping rate, then the closer well location would seem preferable.

Return to exercise 1⮥

Exercise 2 Solution: Lower Ratio of Streamflow to Pumping

The problem:

The Base Case analysis of Exercise 1 assumed that the streamflow was much greater than the well pumpage. What if the pumping rate was higher than in Exercise 1 and the ratio of streamflow to pumping (withdrawal) rate was much lower? In the Base Case, river inflow is about ten times greater than the well pumpage. Consider a case in which the well pumpage is increased by a factor of three (two more wells are drilled close to the original well to form a well field, and each well has the same pumping capacity, so the total Q in that cell of the model grid is -6,078 m^3/d) and the river inflow is reduced by a factor of three (Q_{in} = 6,667 m^3/d instead of 20,000 m^3/d). The ratio of streamflow entering the system to pumping out of the aquifer would then be about 1.1. How would that affect (1) the streamflow in space and time, (2) the drawdown in the aquifer, (3) the head distribution in the aquifer, and (4) the hydrograph for the pumping well? How does that affect (5) the water budget of the aquifer and (6) the fractional sources of water to the well? Is this pumping scenario sustainable?

How to run and analyze the model results:

If you have not already done so, it is useful to read Box 3, then run and post-process the results of the Base Case model of Case Study 1 before undertaking Exercise 2. To do this, first put the input files, MODFLOW-NWT executable code, and ModelMuse files for the Base Case of Case Study 1 on a Microsoft-OS computer by downloading the zip file "CaseStudy1--Models.zip" from the online Supplementary Information for this book↗. Extract the "Case Study 1" folders and subfolders onto your personal computer. Then go through the steps described in Box 3.

Acquiring a file folder for Exercise 2:

Next download the zip file "Exercise2.zip" from the online Supplementary Information for this book↗. To get you started, we have already copied the Base.Case input files into a new folder under the folder "Exercise 2" (Revised Base Case--Change.Qs); you can use these to simulate the case with more pumping and less streamflow. However, the files have not been modified yet, so if you execute these files without changes, you will get the base case result. We suggest you work from this folder to analyze and develop solutions to Exercise 2. For convenience, we have also installed a copy of the executable code for MODFLOW-NWT in a location that will work with the batch files in these two folders.

Modifying Input Files

To determine the effect and importance of the relative strength of the pumping stress to the magnitude of streamflow, copy the Base Case input files into a new folder for Exercise 2 and modify the input parameters to match the assumptions of the above exercise. Also, to more clearly assess the impact on streamflow, it is suggested that you add an additional stream gage to a location in the river close to the well. The results of Exercise 1 indicated a minimum streamflow in reach 46, so place the gage there. These modifications can be accomplished by:

1. Increase the pumping rate for the well on the last line of the WEL Package input file "Base.Case.wel" by changing the value "-2.026000000000E+003" to "-6.078000000000E+003".
2. Reduce the streamflow entering the river for both stress periods in the file "Base.Case.sfr" by changing both occurrences of the value "2.000000000000E+004" to "0.666700000000E+004".
3. Add a stream gage by (1) changing the number on the first line of the "Base.Case.gag" file to "2", (2) adding a third line to the "Base.Case.gag" file "1 46 20206 1", and (3) adding a line to the "Base.Case.nam" file after the similar line for the first gage that says "DATA 20206 ..\Output.Files\Base.Case.sfrg2 REPLACE").

Next, run the model by double clicking on the batch file "Base.Case.bat" in the Input.Files folder.

Assessing Streamflow:

In assessing the effects of pumping on streamflow for the case with increased pumping and less incoming streamflow, we see that after 200 years, the streamflow along the river (from the upstream to downstream ends) is much lower than for the previous Base Case (Figure ExSol 2-1). This difference appears to be explainable primarily by the difference in the specified inflow at the upstream end of the river. To more fully understand the changes in streamflow, we can look at the records of the two stream gaging stations in the GAG output files (Figure ExSol 2-2). These results show that flow at the gage located near the middle reach of the river (Gage no. 2) declines rapidly for 5 years until the river goes dry at that location; it stays dry (no flow) for the remainder of the simulation period. The gage at the downstream end of the river (Gage no. 1) similarly declines rapidly at first, but then the rate of decline decreases after 5 years -- at the same time that the upstream reach has ceased flowing. At that time -- starting in year 6 -- the streamflow depletion is influenced only by the decreased groundwater discharge and is no longer affected by decreased infiltration in the upper reaches.

The nature of the cessation of streamflow in selected parts of the river can be analyzed by plotting the streamflow profiles along the river at various times (Figure ExSol 2-3). These results illustrate that the river first goes dry in a short 7-km central reach of the river at 6 years. The length of the dry section subsequently expands in both an upstream and downstream direction. After 200 years of pumping, 40 kilometers of the river are dry, and the expansion trend seems to be continuing. Therefore, with the smaller ratio of streamflow to pumping, the effects on the flow in the river are much more severe, noticeable, and environmentally damaging.

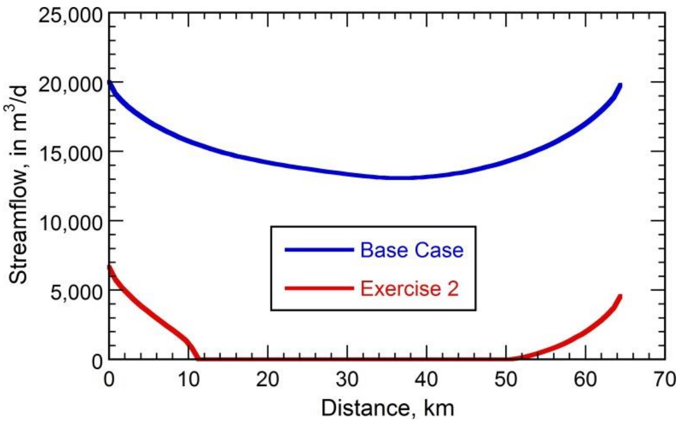

Figure ExSol 2-1 - Streamflow variations with distance after 200 years (Konikow and Bredehoeft, 2020).

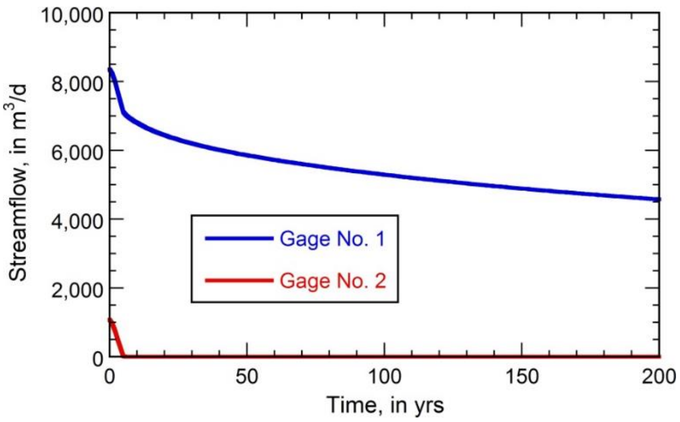

Figure ExSol 2-2 - Streamflow changes for 200 years at two gaging stations. Gage No. 1 is located at the outflow from the modeled area (reach 80) and Gage No. 2 is located at reach 46 (37km) downstream from where the river enters the modeled area) (Konikow and Bredehoeft, 2020).

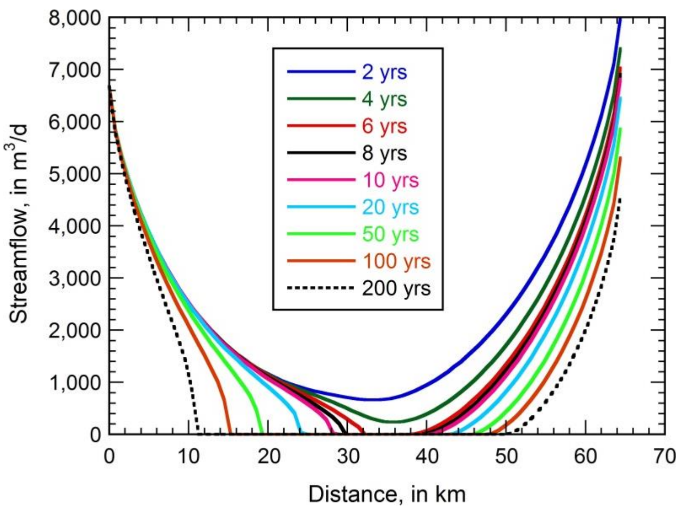

Figure ExSol 2-3 - Changes in the streamflow profile during the 200-year simulation period showing growth in the length of the dry central reaches after year 5 (Konikow and Bredehoeft, 2020).

<u>**Assessing Drawdown:**</u>

The drawdown in the pumping well would be expected to be greater than in the Base Case. As seen in Figure ExSol 2-4, that is indeed the case, although the differences are relatively small. Whereas the Base Case drawdown indicates that it is close to equilibrium before 200 years, with the higher pumping rate in Exercise 2 that is not the case and drawdown is still increasing measurably. But in both cases the maximum drawdown is small compared to the average saturated thickness of the aquifer (150 m), so even with the higher pumping rate, additional long-term drawdown would not be a major concern to water managers.

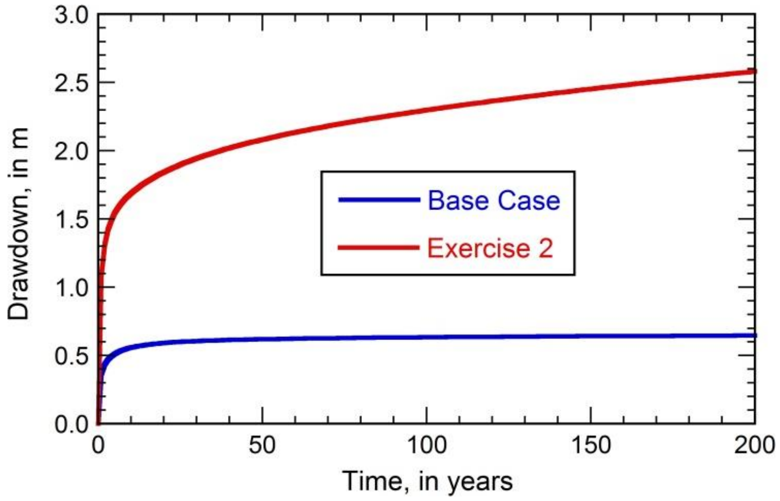

Figure ExSol 2-4 - Calculated drawdown in the pumping well for Base Case and Exercise 2 (with higher pumping rate) (Konikow and Bredehoeft, 2020).

Assessing Heads:

The head distribution calculated after 200 years of pumping (Figure ExSol 2-5) indicates that even though the pumping rate is higher and drawdown is greater than in the base case, all of the groundwater entering the well is ultimately derived from the upstream reaches of the river; the pumping well does not capture any of the water derived from mountain front recharge along the system's western boundary. Comparison to the base case heads (Figure 20b) indicates that the most visible differences include more drawdown, lower heads in most areas, stronger convergence of flow close to the pumping well, and head contours perpendicular to the river where it has gone dry (indicating a lack of exchange between the aquifer and the river along those dry reaches of the river).

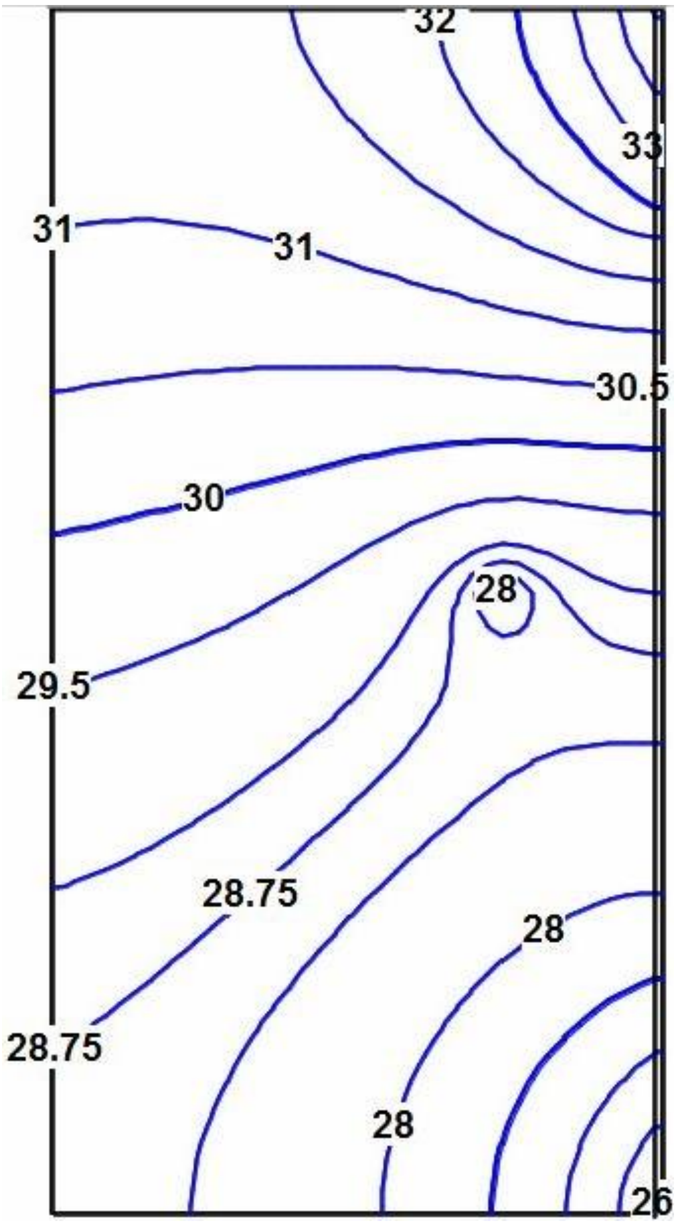

Figure ExSol 2-5 - Calculated heads (in m) in the aquifer after 200 years for conditions of Exercise 2 (with higher pumping rate). The contour interval is 0.5 m, with a supplementary contour at 28.75 m (Konikow and Bredehoeft, 2020).

Assessing the Water Budget:

The water budgets (Tables ExSol 2-1 and ExSol 2-2) have changed substantially from those of the Base Case (Tables 1 and 2 in the book), mostly because of the different assumptions about incoming streamflow and well pumpage. A comparison of water budgets for the two cases indicates that the biggest changes are (1) the much higher storage depletion rate and (2) the reduced outflow to the river at 200 years -- both of which result directly from the higher imposed pumping rate than in the Base Case.

Table ExSol 2-1 - Groundwater budgets for Exercise 2 for predevelopment conditions and after 200 years of pumping one well. All flux values are in m^3/d.

		Predevelopment	t = 200 Years
IN	Mountain Front Recharge	1,688	1,688
	Change in Storage	0	2,299
	Stream infiltration	5,842	6,667
	Total	**7,530**	**10,654**
OUT	Wells	0	6,078
	Outflow to stream	7,534	4,577
	Total	**7,534**	**10,655**

Table ExSol 2-2 - Streamflow budgets for Exercise 2 for predevelopment conditions and after 200 years of pumping one well. All flux values are in m3/d.

	Predevelopment	t = 200 Years
River Inflow	6,667	6,667
River Outflow	8,232	4,577

As in the Base Case (Figure 21), the components of the water budget change substantially during the 200-year transient simulation period (Figure ExSol 2-6). Compared to the Base Case, in this new simulation the increase in induced infiltration from the river to the aquifer (a type of recharge) has reached its maximum possible value at about 6 years, and then ceases to change after that. The maximum increase in induced infiltration from the river is the difference between the inflow to the river (6,667 m^3/d) and the stream infiltration (or seepage loss from the river) under predevelopment conditions (5,842 m^3/d) -- a difference of 825 m^3/d. Once that reaches its maximum, the growth in capture is derived solely from continued decreases in groundwater discharge to the river in its downstream reaches. An inflection in the capture and groundwater storage depletion curves occurs at 6 years also when the increase in recharge stabilizes. The system becomes capture dominated at about 100 years (compared to almost 20 years in the Base Case).

The fractional sources of water to the well under the conditions of Exercise 2 are shown in Figure ExSol 2-7. After 200 years, 38 percent of well pumping is balanced by groundwater storage depletion while 62 percent of pumping is balanced by capture -- substantially less than in the Base Case. This indicates that it will take a much longer time for the system to reach a new equilibrium.

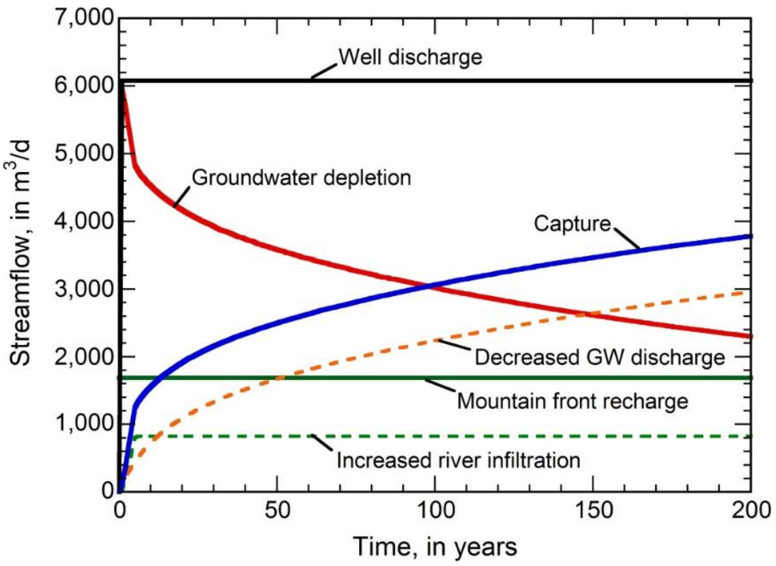

Figure ExSol 2-6 - Calculated changes in the water budget of the hypothetical desert-basin aquifer during the 200-year simulation period for the Exercise 2 case. Capture is the sum of its components represented by the dashed lines (Konikow and Bredehoeft, 2020).

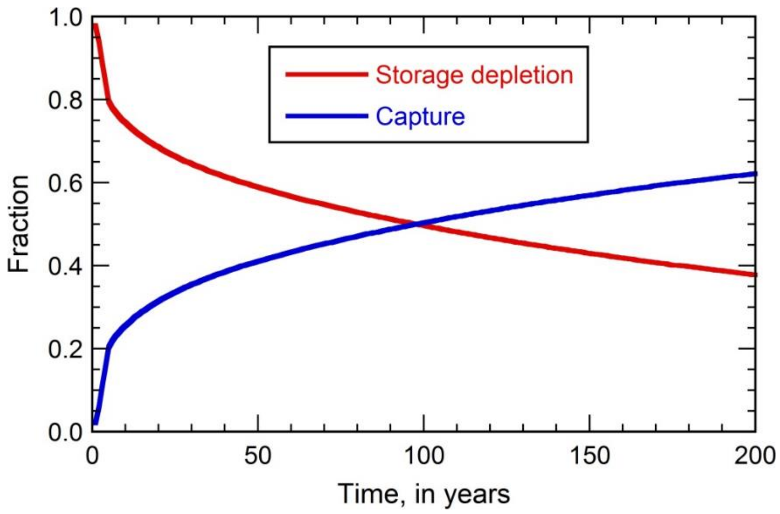

Figure ExSol 2-7 - Nondimensional sources of water being pumped based on annual rates for the conditions of Exercise 2 (Konikow and Bredehoeft, 2020).

<u>Assessing Sustainability:</u>

The well is still deriving water from increasing amounts of capture after 200 years of pumping. The water budget analysis and system responses indicate that from a hydraulic perspective solely of the aquifer, the pumping is sustainable. However, the capture rates are high relative to the flow in the river, and part of the river goes dry after just 5 years of pumping. The extent of the dry reach grows continually over time. This would likely constitute unacceptable surface water impacts and environmental/ecological damage to most hydrologists and water managers.

Return to exercise 2↥

Exercise 3 Solution: Analytical Solution for Streamflow Depletion

The problem:

Estimate streamflow depletion using Glover's analytical solution (rather than a numerical model). How do these results compare with those from the numerical model used in the Base Case and in Exercise 1? Explain any differences.

Approach:

Solve Equation 5 for each year of the 200-year simulation. There are a number of ways to do this, but one reasonable approach is to use formulas in an Excel spreadsheet. This exercise can be completed using information available in this book and in results for Exercise 1.

Analytical solution:

We solved Glover's analytical solution (Equation 5 of this book) using formulas in an Excel spreadsheet. We copied the spreadsheet from Exercise 1 ("RateBudgets.xlsx") and pasted a copy into the Exercise 3 subfolder "DataSpreadsheets". Then we deleted the worksheets for the closer well and further well, and added a new worksheet "Glover Soln.". We entered the known values for parameters in column B, and then solved for values of z and $Qs(t)$ for every year from 0 to 200 years in columns G and H using the *ERFC* function in Excel for complementary error function. We then plotted these results and the capture values from the Base Case analysis (Figure ExSol 3-1). The zip file "Exercise3.zip", including the spreadsheet, is available in the online Supplementary Information for this book↗.

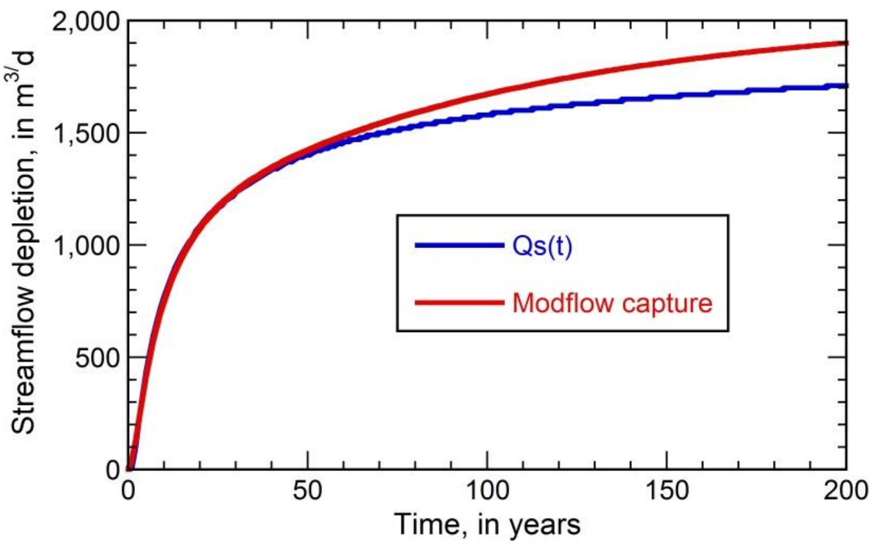

Figure ExSol 3-1 - Streamflow depletion estimates for analytical and numerical solutions for the Base Case scenario. The $Qs(t)$ (blue curve) is the streamflow depletion calculated using the analytical solution developed by Glover and Balmer (1954). The red curve represents capture calculated from the water budget data of the numerical solution using MODFLOW-NWT for the Base Case (Konikow and Bredehoeft, 2020).

Comparison of analytical and numerical solutions:

The two solutions are in close agreement for the first 50 years or so. After that they begin to diverge, with the analytical solution providing smaller values of streamflow depletion at later times. The largest difference, at 200 years, is about 200 m^{3}/d, with the analytical solution value being about 10 percent less than the numerically calculated value.

The analytical solution requires the application of several simplifying assumptions. The Base Case is designed to include many simplifying assumptions also, so the match should be good. The analytical solution assumes a semi-infinite aquifer (i.e., an assumption that the aquifer extends without end in a particular direction). However, the hypothetical aquifer is not semi-infinite in extent as it has impermeable boundaries on the north, west, and south sides (i.e., at finite distances from the well and river); that is, the numerical model assumes the aquifer has a limited areal extent, as defined by the outer impermeable or no-flow boundaries). Figure ExSol 1-7 shows that measurable drawdown occurs at the north, west, and south boundaries. The effect of such boundaries would be to eventually cause increased drawdown in the modeled aquifer compared to having no such boundaries as in a semi-infinite idealization. Increased drawdown would cause increased capture. This is consistent with the results shown in Figure ExSol 3-1, so it is likely that this one difference between the assumptions for the analytical solution and the boundary conditions of the numerical model can explain most or all of the underestimate by the analytical solution. However, the overall excellent agreement, especially for the first 50 years in this case, combined with the efficiency and ease of use of the analytical solution, indicates that this would be a valuable method to apply early in any study of groundwater development in a stream-aquifer system. It will also give you an expectation and basis of comparison for the results of a numerical model that encompasses more complex boundary conditions and heterogeneous properties.

Return to exercise 3↥

About the Authors

Dr. Leonard Konikow is a Scientist Emeritus with the United States Geological Survey, where he was a research hydrologist for more than 40 years. He is also the Editor-in-Chief of Groundwater journal. His research interests include the development and application of simulation models for groundwater flow and contamination problems, groundwater-surface water interactions, coastal submarine groundwater discharge processes, and groundwater depletion. He is a member of the National Academy of Engineering, Fellow of the American Geophysical Union, and has received the M. King Hubbert Science Award and Life Member Award from the National Ground Water Association and the O.E. Meinzer Award from the Geological Society of America. He also received the President's Award from the International Association of Hydrogeologists. He was the Birdsall Distinguished Lecturer for the Hydrogeology Division of GSA in 1986-87. He served on several committees of the National Research Council and in leadership positions in several professional societies. He is author or coauthor of numerous scientific articles and reports.

Dr. John Bredehoeft had a 32-year career with the United States Geological Survey (USGS) as a research geologist and as a senior manager in their Water Resources Division. After retiring from the USGS, he went on to found a major geotechnical consulting firm. He had also served as Editor-in-Chief of Groundwater journal. He is a member of the National Academy of Engineering and recipient of several major awards and honors, including the Penrose Medal of the Geological Society of America, the Horton Award from AGU, the Meinzer Award from GSA, and the Life Member Award from the National Ground Water Association. He taught at the University of Illinois, Stanford, University of California - Santa Cruz, and San Francisco State University. He served on numerous national advisory committees for the National Research Council, the National Science Foundation, and the Department of Energy. Bredehoeft is the author of more than 100 scientific papers in the refereed scientific literature. Together with George Pinder, they (1) developed and published the first widely used numerical groundwater flow model, and (2) the first widely used contaminant transport model.

Modifications to Original Release

Changes included in Modifying from Original Version to Version 2

Page numbers refer to page numbers in the original pdf.

pages i, ii, Removed small caps font from title to be consistent with current GWP book format.

page 84, last paragraph, "Figure ExSol 1-7" changed to "Figure ExSol 1-8"

page 85, figure caption, "Figure Ex1-7" changed to "Figure ExSol 1-8"

Changes from Version 2 to Version 3

page ii, updated version number

page 58, updated hyperlink to Kansas Department of Agriculture

page 70, updated hyperlink to software for GW_Chart

page 71, Figure Box 3-6, replaced with a higher resolution version

page 72, Figure Box 3-7, revised the style and axis labels to be consistent with other figures in the book

page 73, Figure Box 3-8, revised the style and axis labels to be consistent with other figures in the book

page 74, Figure Box 3-9 and Figure Box 3-10, revised the style and axis labels to be consistent with other figures in the book

page 75, updated hyperlink to software for GW_Chart

page 75, first full paragraph, second line, corrected the cell coordinates of the pumping well from (35,40,1) to (30,40,1) page 76, Figure Box 3-12, revised the style and axis labels to be consistent with other figures in the book

page 75, Figure Box 3-11, replaced with graph showing drawdown at the corrected the cell coordinates of the pumping well (30,40,1)

Supplemental Material revision:

> The zip file "Groundwater Resource Development: Case Study 1--Models.zip" was revised: the spreadsheet "Base.Case.Budget.Streamflow.Hydrograph.xlsx" within the subfolder "SpreadsheetForBaseCaseAnalysis" within the folder "Case Study 1--Models" was replace with updated. The revision contains drawdown at the correct location of the pumping well, cell(30,40,1)

Changes from Version 3 to Version 4

Version 3: July 6, 2023, Version 4: January 19, 2024

Page numbers refer to the Version 3 PDF.

page ii, added page requesting support of the Groundwater Project

page ii, now page iii, updated version number and date

page iii, now page iv, added "Any use of trade, firm, or product names is for descriptive purposes only and does not imply endorsement by the U.S. Government."

Changes from Version 4 to Version 5

Version 4: January 19, 2024, Version 5: September 16, 2024

Page numbers refer to the Version 4 PDF.

page iii, updated version number and date

page 67, clarified the explanation of the bottom and side views of Figure Box 3-1

page 79, 1st paragraph, corrected the sentence started with 'However,' to say 'streamflow is highest for the case of the closer well and lowest for the case of the further well.'