A method to estimate groundwater depletion from confining layers

L. F. Konikow[1] and C. E. Neuzil[1]

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[1] Although depletion of storage in low-permeability confining layers is the source of much of the groundwater produced from many confined aquifer systems, it is all too frequently overlooked or ignored. This makes effective management of groundwater resources difficult by masking how much water has been derived from storage and, in some cases, the total amount of water that has been extracted from an aquifer system. Analyzing confining layer storage is viewed as troublesome because of the additional computational burden and because the hydraulic properties of confining layers are poorly known. In this paper we propose a simplified method for computing estimates of confining layer depletion, as well as procedures for approximating confining layer hydraulic conductivity (K) and specific storage (S_s) using geologic information. The latter makes the technique useful in developing countries and other settings where minimal data are available or when scoping calculations are needed. As such, our approach may be helpful for estimating the global transfer of groundwater to surface water. A test of the method on a synthetic system suggests that the computational errors will generally be small. Larger errors will probably result from inaccuracy in confining layer property estimates, but these may be no greater than errors in more sophisticated analyses. The technique is demonstrated by application to two aquifer systems: the Dakota artesian aquifer system in South Dakota and the coastal plain aquifer system in Virginia. In both cases, depletion from confining layers was substantially larger than depletion from the aquifers.

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1. Introduction

[2] An accurate water budget is an important basis for managing groundwater resources and for understanding hydrologic fluxes at local and regional scales. Even at national and global scales, estimates of groundwater budgets can provide useful input to water policy issues [*Galloway et al.*, 2003; *Konikow and Kendy*, 2005; *U.S. Geological Survey*, 2002; *Zekster*, 2000] as well as to understanding the human role in global changes in the hydrosphere, such as sea level rise [*Sahagian et al.*, 1994]. At a World Climate Research Programme workshop held in 2006, participants highlighted the need for data on changes in subsurface water storage resulting from groundwater use changes and aquifer mining, among other factors [*Church et al.*, 2007]. A simple method to help estimate the long-term large-scale changes in water stored in low-permeability confining layers would help meet this need.

[3] Confining layer storage is a significant source of water when aquifers are developed [e.g., *Theis*, 1940; *Jacob*, 1946; *Hantush*, 1960; *Bredehoeft et al.*, 1983]. Indeed, low-permeability sediments tend to be volumetrically dominant and more storative than confined aquifers, suggesting that they could often be the primary source of groundwater released from storage. The equations describing transient leakage of water out of low-permeability confining layers and analytical solutions for them, especially for flow to a single pumped well, have been well documented in the literature [e.g., *Jacob*, 1946; *Neuman and Witherspoon*, 1969a, 1969b; *Herrera*, 1970]. Confining layer response to pumping can also be represented numerically in simulations of large-scale aquifer development, using fully numerical or semianalytical approximations, as described by *Leake* [1990] and *Leake et al.* [1994], among others.

[4] With measurements or estimates of the hydraulic properties of confining layers and knowledge of the drawdown history in adjacent aquifers, the head changes in a confining layer can be calculated, thereby allowing direct computation of changes in storage within the confining layer as well as flow through it. *Herrera and Figueroa* [1969] and *Herrera* [1970] present integrodifferential equations that permit calculation of depletion for such a system. Development of numerical solutions to the integrodifferential equations was documented by *Herrera and Yates* [1977] and *de Marsily et al.* [1978] as well as by *Cooley* [1992], whose approach was incorporated into the widely used three-dimensional model MODFLOW [*McDonald and Harbaugh*, 1988] by *Leake et al.* [1994]. Depletion can thereby be calculated using a well-calibrated numerical simulation model. Alternatively, given the same information, analytical approaches can be used. *Neuman and Gardner* [1989] present convolution integrals from which the drawdown at any point and time in the confining layer can be calculated. This convolution approach makes it

[1]U.S. Geological Survey, Reston, Virginia, USA.

possible to calculate flow through the confining layer as well as changes in groundwater storage within it.

[5] Despite the availability of these tools, depletion of groundwater storage in low-permeability layers is still considered inconvenient to calculate, and is still all too rarely monitored or included in water budgets. For example, the simulations used in some regional aquifer system studies (RASAs) conducted by the U.S. Geological Survey were based on so-called quasi-three-dimensional conceptualizations. In this approach the confining layers are represented as vertical conductance terms and transient changes in storage within the confining layers themselves cannot be computed as a distinct component of the groundwater budget. Examples include the coastal plain of Virginia [*Harsh and Laczniak*, 1990], the Gulf Coast aquifer systems, south-central U.S. [*Williamson and Grubb*, 2001]; the coastal plain aquifer system in the southeastern U.S. [*Barker and Pernik*, 1994], bedrock aquifers in the Northern Great Plains [*Downey*, 1986], and the Cambrian-Ordovician aquifer system, northern Midwest U.S. [*Mandle and Kontis*, 1992].

[6] Undoubtedly, the additional computational burden, the paucity of data on confining layer properties and head changes, and the perception that it is relatively unimportant have all perpetuated the practice of ignoring confining layer storage. The objective of this paper is to present a simplified and widely applicable approach to estimating long-term decreases in the volume of water stored in low-permeability confining layers in response to withdrawals from adjacent aquifers. For situations in which the confining layer hydraulic properties are unknown or poorly defined, we offer geologically based guidelines to make appropriate order-of-magnitude estimates of the hydraulic conductivity (K) and specific storage (S_s) of low-permeability sedimentary media. We document the theory, evaluate its reliability, and demonstrate its application to actual field settings. The approach can be considered a scoping calculation, but the errors arising from the approximations will generally be smaller than the errors and uncertainty in defining the hydraulic properties of a confining layer, even when measurements of the latter are available. The results of the analyses and example applications demonstrate the utility of our simplified approach and also show the importance of confining layer depletion as a source of water to wells in confined aquifers.

2. Theory

[7] Drawdown in an aquifer alters the boundary conditions for adjoining confining layers, causing them to release water from storage to the aquifer as declines in hydraulic head propagate slowly into the low-permeability material. Although their low permeability mediates the rate of release, relatively large specific storage values in clay-rich confining layers can enable large quantities of water to "leak" into aquifers over human timescales. The amount released can be calculated by solving an equation for flow in the confining layer using (1) the drawdown history in the aquifer and (2) the thickness and hydraulic diffusivity (defined as the hydraulic conductivity, K, divided by the specific storage, S_s) of the confining layers above and below the aquifer. However, *estimates* of the amount released can be obtained more simply by (1) representing the aquifer drawdown with a roughly equivalent step function, (2) estimating K and S_s of the confining layers from their lithology, depth, and age (if no direct measurements are available), and (3) representing the actual drawdown in the confining layer with a roughly equivalent depth of penetration for the boundary head change.

2.1. Thick Confining Layer

[8] In this case it is assumed that the confining layers of interest are thick enough that drawdown effects do not reach either (1) the opposite confining layer boundary or (2) a drawdown effect from an aquifer on the opposite side. This assumption can easily be checked as the analysis proceeds, but it is unlikely to be an issue in many cases. Under this assumption, the drawdown in a semi-infinite domain is germane. Specifically, from the analogous solution for linear heat flow in a semi-infinite solid [*Carslaw and Jaeger*, 1959, p. 59], the head change in a semi-infinite confining layer initially having a constant and uniform initial head H_0 and subjected to a drawdown of $\Delta\overline{H}$ at the boundary ($z = 0$) at time $t = 0$ can be calculated in terms of the dimensionless drawdown in the confining layer ($\Delta h/\Delta\overline{H}$), given by

$$\left(\frac{\Delta h}{\Delta\overline{H}}\right) = \operatorname{erfc}\left[z/2(Kt/S_s)^{1/2}\right] \tag{1}$$

where z is distance into the confining layer from its boundary with the aquifer.

[9] This solution is plotted in dimensionless form in Figure 1. Assuming that the volume of water released from storage is linearly related to the drawdown, that is, that the parameters K and S_s are constant, the area below the curve in Figure 1 (area A plus area B) is proportional to the volume of depletion from the confining layer. Simple inspection suggests that area is reasonably approximated by the sum of area A plus area C, which can be thought of as a constant drawdown $\Delta\overline{H}$ that penetrates the confining layer to a dimensionless distance of 1 (Figure 1). In fact, the ratio of area A plus B to area A plus C is approximately 1.1. Thus for our analysis we assume that depletion can be represented by a uniform head decrease of $\Delta\overline{H}$ that penetrates the confining layer to a dimensionless distance of 1. Under this assumption, the virtual distance of penetration, z_d, is given by

$$z_d = (Kt/S_s)^{1/2}. \tag{2}$$

By knowing (or estimating) K and S_s, choosing a representative duration (t) of the drawdown, and choosing a representative area (A) over which the drawdown is effective, the volume of confining layer affected by drawdown $\Delta\overline{H}$ can be computed. Having specified S_s, the volume of water released from confining layer storage (V_w) can then be calculated as

$$V_w = S_s z_d A \Delta\overline{H}. \tag{3}$$

If the confining layer properties are the same on both sides of a confined aquifer, the total depletion volume is $2V_w$ because drawdown occurs in both adjacent confining layers.

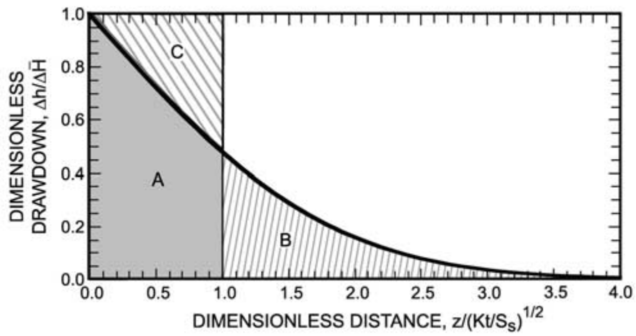

Figure 1. Dimensionless plot showing drawdown in a semi-infinite confining layer as a function of system properties and time in response to a step change in head in the bounding aquifer. Area $A + B$ is approximately equal to area $A + C$.

[10] In choosing a "representative" duration, t, attention should be paid to the aquifer's drawdown history. If the rate of drawdown has been relatively constant over time since development began, then a t of about half the total time since development started might generally be appropriate. In an analogous problem of estimating flux out of a confining layer, *Bredehoeft and Pinder* [1970, p. 887] assumed that the stepwise head change was applied at one half the elapsed time and they considered "... the results to be sufficiently accurate for engineering purposes." On the other hand, if the rate of withdrawal has been relatively constant over time, the rate of drawdown will decrease with time and the bulk of the drawdown will have occurred during the early phase of development; in this case, the appropriate duration might be the time since half the drawdown occurred. If the history of well drilling, observation well hydrographs, and (or) pumpage records indicate that most of the drawdown was relatively recent, a shorter t would be appropriate.

2.2. Thin Confining Layer

[11] When aquifer drawdowns release water from storage in confining layers, pressure changes may propagate through the entire thickness of the confining layer during the time of interest. This happens when the confining layer is relatively thin or its hydraulic diffusivity is relatively large. Highly lithified Paleozoic shales, for example, may be densely fractured and more hydraulically diffusive than younger, more plastic shales. In such cases the confining layer cannot be analyzed as a semi-infinite domain. A modified procedure, however, can be used for estimating the volume of water released from a confining layer under such conditions.

[12] First, it is necessary to define the conditions under which a confining layer of thickness l can no longer be treated as semi-infinite. When drawdown penetrates the entire confining layer, reaching its distal boundary, the response depicted in Figure 1 is interrupted. For a limiting case in which there is no drawdown in an aquifer on the other side of the confining layer—equivalent to a constant-head boundary condition—then $\Delta h = 0$ at that boundary. Referring to Figure 1, and setting $z = l$, we see that for thicknesses such that $l/(Kt/S_s)^{1/2} > 4$, we have $\Delta h/\Delta\overline{H} \approx 0$; in these cases the imposition of constant head at distance l (that is, at the far boundary) does not conflict with the assumption of a semi-infinite response. At smaller values of $l/(Kt/S_s)^{1/2}$ semi-infinite behavior becomes increasingly compromised. In the present context, however, the resulting errors are certainly small enough to be ignored until $l/(Kt/S_s)^{1/2}$ is smaller than 3. In fact, at $l/(Kt/S_s)^{1/2} = 2$, the actual volume of water released from the confining layer will approach the value calculated from equation (3) with time, whereas at $l/(Kt/S_s)^{1/2} = 1$, equation (3) will overestimate the depletion volume by a factor of about 2. Thus we can conservatively adopt $l/(Kt/S_s)^{1/2} < 2$ (or $l < 2\ z_d$) as a criterion for abandoning the semi-infinite assumption. When $l < 2\ z_d$, a modified method of estimating water released from a confining layer of finite thickness, described below, should be used.

[13] *Crank* [1964, p. 47, equation 4.22] presents the solution that describes flow in a finite layer with a uniform initial head and arbitrary constant heads imposed at both boundaries at $t = 0$. Analogously to equation (1), this solution applies exactly for step changes in head in the aquifer at the confining layer boundaries, but also like equation (1), it can be applied to approximate the response to a gradual drawdown in the bounding aquifers. We will first consider the case of head change in one aquifer only; that is, a step change in head at one confining layer boundary and no change from the initial value at the other boundary. According to *Crank* [1964, p. 47, equation 4.23], the volume of water released from confining layer storage by time t, V_{wt}, is given by

$$V_{wt} = V_{w\infty}\left\{1 - \frac{8}{\pi^2}\sum_{n=0}^{\infty}\frac{1}{(2n+1)^2}\exp\left[-K(2n+1)^2\pi^2 t/S_s l^2\right]\right\} \tag{4}$$

where $V_{w\infty}$ is the ultimate volume, or the volume of water released after infinite time. The latter quantity is easily determined. Figure 2 shows a schematic cross section of a confining layer of thickness l with a linear initial head gradient between the initial boundary heads of H_{01} at its lower boundary and H_{02} at its upper boundary. For simplicity the special case of $H_{01} = H_{02}$ is shown, but the same analysis applies when $H_{01} \neq H_{02}$. A head change (ΔH_0) is applied at the lower boundary at $t = 0$. A new equilibrium is attained at infinite time, and that equilibrium head distribution is shown by the dashed line. From it we can see that

$$V_{w\infty} = S_s l A \Delta\overline{H}_t \tag{5}$$

where A is the area over which the head decline has occurred and $\Delta\overline{H}_t$ is the average head change over the thickness of the confining layer. As seen in Figure 2,

$$\Delta\overline{H}_t = \frac{\Delta H_0}{2}. \tag{6}$$

Rewriting equation (4) with the quantity within braces represented by $f(t)$ yields

$$V_{wt} = f(t) S_s l A \Delta\overline{H}_t. \tag{7}$$

The remaining task is evaluating $f(t)$. This is readily done graphically using the plot of $f(t)$ versus $(Kt/S_s l^2)^{1/2}$ provided as Figure 3.

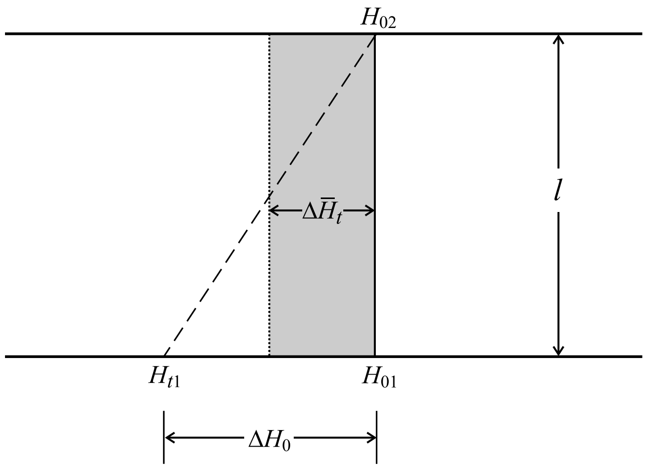

Figure 2. Schematic cross section through a hypothetical confining layer illustrating effect of a step change in head at the lower boundary (ΔH_0) on the equilibrium head gradient through the confining layer. The area of triangle $H_{t1} - H_{01} - H_{02}$ equals $\Delta \overline{H}_t \times l$.

[14] The previously established criterion of $l/(Kt/S_s)^{1/2} < 2$ for adopting the finite-thickness analysis is equivalent to $(Kt/S_s l^2)^{1/2} > 0.5$. This region of Figure 3 therefore will be the area of interest, although, as shown below, $(Kt/S_s l^2)^{1/2} > 0.25$ applies when there is drawdown on both sides of the confining layer. Note in Figure 3 that for $(Kt/S_s l^2)^{1/2} > 0.75$, $V_{wt} \approx V_{w\infty}$ (or $f(t) \approx 1$). The practical significance of this is that release of water from confining layer storage has essentially ceased; further releases of any significance from storage in the confining layer will occur only after additional aquifer drawdown.

[15] Thus far, drawdown in only one aquifer has been considered. Often, however, withdrawals occur in more than one aquifer and release of water from a confining layer will result from head changes at both its upper and lower boundaries. For determining whether or not the aquifer can be considered semi-infinite, this effectively halves the available thickness of the confining layer. As a result, the criterion for invoking the finite-thickness analysis presented here can be generalized as follows. If there has been drawdown on one side of the confining layer, the finite-thickness analysis should be used if

$$l/(Kt/S_s)^{1/2} < 2, \tag{8}$$

which is equivalent to $l < 2\, z_d$; if there has been drawdown on both sides of the confining layer, the finite-thickness analysis should be used if

$$l/(Kt/S_s)^{1/2} < 4, \tag{9}$$

which is equivalent to $l < 4\, z_d$.

[16] Calculating the volume of water released from storage when head changes have occurred on both sides of a confining layer is relatively straightforward. The linearity of the transient flow equation (when K and S_s are invariant) means that responses are simply additive. One simply computes the volumes of water released in response to each boundary head change using equation (7) and adds them. Because system behavior is approximated using the solution for a step change, the same strategy for choosing an appropriate time t and aquifer drawdown ΔH_0 as previously described should be used.

3. Estimating Hydraulic Properties of Confining Layers

[17] The accuracy of depletion estimates will obviously be affected by uncertainty and error in the values of the hydraulic properties of the confining unit. Site-specific field or laboratory measurements are clearly desirable, but reliable data from such tests frequently are not available. Inverse numerical models can be used as a basis for estimating the hydraulic properties of confining units but, in the case of specific storage, require knowledge of total water withdrawals – the quantity we wish to estimate. The simplified analysis technique described above is intended to provide approximations in cases where more thorough or sophisticated analyses are impractical, so it is appropriate to incorporate, when necessary, hydrogeologically estimated values for the hydraulic properties of the confining layers. In most groundwater budget analyses it is likely that, despite potential errors, using estimated confining layer properties is preferable to ignoring the contribution of confining layers altogether. Moreover, by estimating confining layer hydraulic properties it is possible to perform scoping calculations and to examine possible extremes. Finally, and perhaps most importantly, we note that even when site-specific measurements of confining layer properties are available, they may be more or less accurate and representative of the confining layer at the location and scale of interest. Thus we believe careful estimates of the properties can be highly useful and can even aid in the evaluation of confining layer data of uncertain quality and applicability. In this section we outline an approach to simply and quickly estimate K and S_s for confining layers. We emphasize, however, that in most cases these estimates will undoubtedly be the primary source of error in the calculation of V_w, whether they are used with our simplified method or a more sophisticated analysis of aquifer-confining layer interaction.

3.1. Hydraulic Conductivity

[18] A growing body of data, much of it obtained in the last decade or two, reveals that the hydraulic conductivity

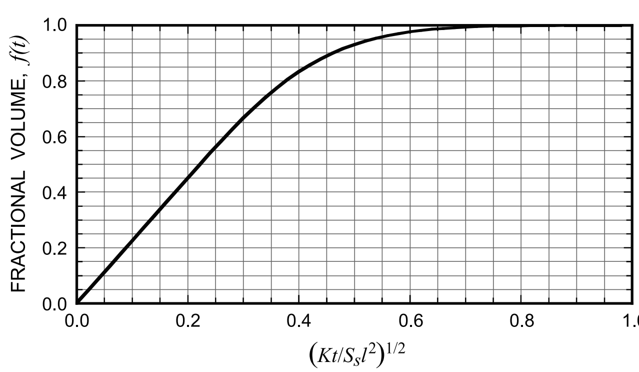

Figure 3. Dimensionless plot of volume of water released from storage in a confining layer of finite thickness in response to a step change in head in a bounding aquifer as a function of system properties.

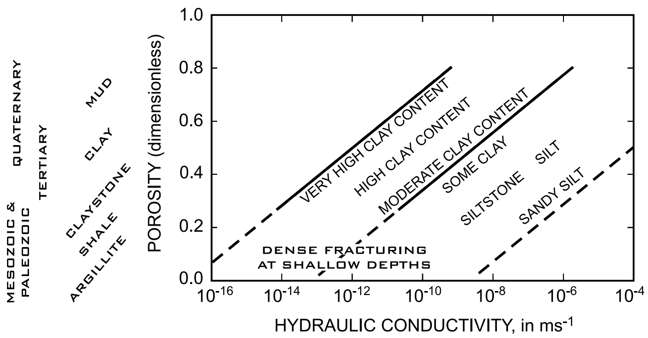

Figure 4. Generalized relation between porosity and hydraulic conductivity for argillaceous confining layers under a range of geologic conditions. Data synthesized in this plot represent a wide variety of argillaceous media and geologic environments and were obtained from a compilation of large- and experimental-scale data by *Neuzil* [1994]; data published by *Katsube et al.* [1996], *Dewhurst et al.* [1999], and *Lee and Deming* [2002]; and a number of unpublished data kindly supplied by D. Saffer (Pennsylvania State University), E. Screaton (University of Florida), K. Brown (Scripps Institute), and B. Dugan (Rice University).

(K) of clay-rich media can be related to porosity (ε) in a reasonably consistent fashion (see Figure 4). This generalized plot is a synthesis of numerous data from a wide variety of geologic environments; the sources of the data are a compilation of large- and experimental-scale data by *Neuzil* [1994], data published by *Katsube et al.* [1996], *Dewhurst et al.* [1999], and *Lee and Deming* [2002], and a number of unpublished data kindly supplied by D. Saffer (Pennsylvania State University), E. Screaton (University of Florida), K. Brown (Scripps Institute), and B. Dugan (Rice University).

[19] The range in K at each porosity value is wide but related to lithology. At contacts with coarse-grained aquifer sediments that represent transitions in depositional regimes, confining layers may contain significant admixtures of nonclay sediments, so the default choice, in the absence of information to the contrary, should be the central "moderate clay content" line. If information pointing to a lithology with more clay or less clay is available, it can be factored into the estimate via the descriptions on the plot. These lithological descriptive terms and likely geologic age ranges have been added along the porosity axis to help select an appropriate value of porosity. Some North American examples can be used as guidelines: Paleozoic Appalachian shales: $\varepsilon \approx 0.20$; shallow Tertiary-age coastal plain clays: $\varepsilon \approx 0.35$; deep Mesozoic-age coastal plain clays: $\varepsilon \approx 0.25$; midcontinent Cretaceous shales: $\varepsilon \approx 0.30$; deep midcontinent Paleozoic shales: $\varepsilon \approx 0.10–0.15$; and eastern U. S. Triassic basin shales: $\varepsilon \approx 0.05–0.25$, depending on depth. As a further guide, note that materials with $\varepsilon > \sim 0.35$ are plastic (moldable) and materials with $\varepsilon < \sim 0.20$ would usually be considered a lithified shale. As noted, shales and argillites that have low matrix porosities but have been uplifted to near the land surface (<~200 m) may have dense fracturing that would make them more permeable than indicated, typically by two to 4 orders of magnitude.

[20] Clay mineralogy also contributes to the spread in permeability seen in Figure 4 and, if known, can also help narrow the range of possible values. Generally, media rich in smectitic clays, such as bentonites, tend to have the lowest permeabilities, whereas those dominated by kaolinite have the highest.

3.2. Specific Storage

[21] Specific storage (S_s) in clay-rich media is also related to porosity in a reasonably consistent fashion. Figure 5, a synthesis of data for a wide variety of argillaceous media, shows the relationship. It was developed primarily from compressibility values from *Domenico and Mifflin* [1965], *Skempton* [1970], *Cripps and Taylor* [1981], *Tellam and Lloyd* [1981], and *Neuzil* [1993]. *Skempton*'s [1970] compilation of compressibility data for normally consolidated media is particularly comprehensive for argillaceous sediments with porosities greater than ~0.20 or 0.30.

[22] Although porosity explicitly appears in expressions for S_s, the decrease in porous matrix compressibility with compaction is the primary reason for the trend of decreasing specific storage with decreasing porosity evident in Figure 5. At any particular porosity, the primary consideration for estimating specific storage is whether the confining layer in question is overconsolidated or normally consolidated. "Overconsolidated" refers to sediments that have experienced a higher overburden load in the past. Thus it applies where there has been erosion of overburden. Sediments that are overconsolidated are sometimes said to be in the "elastic" deformation range, alluding to the fact that additional compactional strains can be largely reversed by removing the load that was added. "Normally consolidated" applies where sediments are at or close to their greatest burial depth. Normally consolidated sediments are sometimes described as being in the "plastic" deformation range because further compaction is largely irreversible. The distinction is important here because elastic range compressibilities, and S_s values, typically are substantially smaller

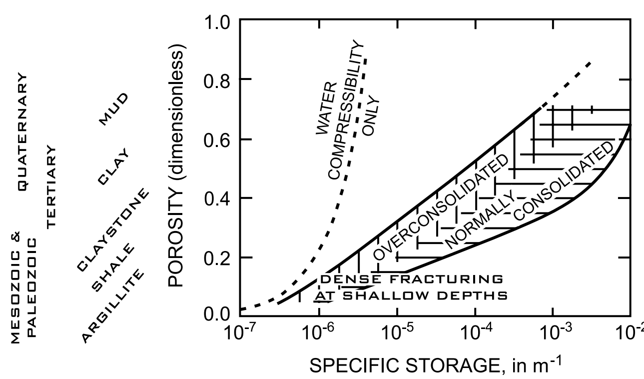

Figure 5. Generalized relation between porosity and specific storage for normally consolidated and overconsolidated argillaceous confining layers. "Water compressibility only" curve shows specific storage for a hypothetical medium with incompressible porous matrix and mineral grains. The plot synthesizes data for a number of different argillaceous media taken from *Domenico and Mifflin* [1965], *Skempton* [1970], *Cripps and Taylor* [1981], *Tellam and Lloyd* [1981], *Burland* [1990], and *Neuzil* [1993].

than plastic range compressibilities and S_s values. Thus a knowledge of a region's geologic history is important for using Figure 5 to estimate S_s.

[23] It is important to note that overconsolidation is a matter of degree. Mildly overconsolidated sediments (those at only slightly less than their maximum overburden load) tend to retain relatively large values of matrix compressibility and S_s, whereas highly overconsolidated sediments (a past overburden load that is multiples of the current one) tend to have small values. This accounts for much of the range in "overconsolidated" values in Figure 5.

[24] For normally consolidated, nonlithified media the range can be narrowed using the value of the so-called liquid limit. The "liquid limit" is standard geotechnical parameter that reflects the amount of water needed to liquify a sample [see, e.g., *Lamb and Whitman*, 1969]. The range in values of S_s shown in Figure 5 for normally consolidated media is correlated with liquid limit values between 30 (smallest S_s) and 140 (largest S_s) for porosities larger than $\sim$0.20. When highly lithified, overconsolidated clay-rich media can be densely fractured, which can increase its compressibility and S_s to higher than normal values for their porosity.

[25] An additional consideration, subsidence due to compaction, can provide a helpful bound on estimates of specific storage in many situations. Release of water from storage in both confining layers and aquifers is accompanied by compaction, which is generally manifested as a like amount of subsidence, or lowering of the ground surface. Subsidence can be apportioned between aquifers and confining layers on the basis of their thickness and the tendency of aquifers to be less compressible than confining layers. This is useful in areas where, for example, there are no data on confining layer specific storage, but subsidence is known or can be estimated. Indicators of the magnitude of subsidence include coastal inundation (or the lack of it), changes in surface drainage, damage to infrastructure, and so on. In some cases one may be able to place an upper limit on subsidence if no indications of subsidence are present. This can be used to advantage to check the reasonableness of estimates of S_s by computing the subsidence that would be expected in any particular situation.

[26] Water is released from confining layers because of the compressibility of both the porous medium and the water itself. Here we are interested in relatively compressible confining layers and can ignore the water compressibility, giving

$$S_s = \rho g \alpha, \tag{10}$$

where ρ is the water density, g is gravity, and α is porous medium compressibility. The latter can be equated to a medium's change in thickness per reference thickness divided by the change in water pressure. Equation (10) can then be solved for the change in thickness. In the case of a thick confining layer, this change in thickness, or the subsidence Δz_{sub}, is given by

$$\Delta z_{sub} = S_s z_d H_0. \tag{11}$$

Here we have equated the reference thickness to penetration depth z_d and the change in water pressure to $\rho g H_0$. In like manner, the subsidence in a thin confining layer would be calculated as

$$\Delta z_{sub} = S_s l f(t) H_0. \tag{12}$$

In this case the reference thickness is l, the confining layer thickness, and the change in water pressure is $\rho g f(t)\overline{H}_t$.

[27] Calculated values of Δz_{sub} that seem too large are a warning signal that the estimated S_s is also too large. This would be the conclusion, for example, if one calculates a large Δz_{sub} for a coastal area that lacks evidence of any appreciable subsidence. In such a circumstance S_s can be reduced until a more reasonable Δz_{sub} is obtained. Care should be exercised, however, because even significant subsidence may not be apparent if there are no stable reference elevations for comparison. Conversely, more or less direct measures of subsidence, such as failed well casings or INSAR data, may be available in some instances to help constrain S_s. If sufficiently accurate regional subsidence data are available, they can be used directly to estimate depletion from both low-permeability units and aquifers by equating the subsidence to the volume of depletion. It is important to note, however, that most monitoring is begun after significant subsidence has occurred so that the data capture only the most recent part of the total elevation change.

4. Test and Evaluation

[28] The proposed method for estimating the volume of groundwater depletion from confining layers was tested and evaluated by application to a hypothetical system with specified hydraulic properties and boundary conditions. A comparison between depletion estimated using the proposed method and depletion computed using a three-dimensional numerical simulation of the system provided the basis for evaluating the reliability of the method. The numerical simulation used the MODFLOW-2000 model [*Harbaugh et al.*, 2000].

[29] The hypothetical groundwater system was designed to represent a regional (100 km by 50 km) aquifer system in which a 200-m thick confining layer underlies a shallow water table aquifer and overlies a permeable confined aquifer (Figure 6). Substantial groundwater withdrawal from wells in the confined aquifer causes drawdown, which in turn induces leakage and depletion from the confining layer. Model parameters are listed in Table 1 and the geometry, grid, and boundary conditions of the test problem are illustrated in Figure 6. We chose specific storage values typical of normally consolidated sediments, and thus the problem is representative of areas such as coastal plains and alluvial valleys, which are among the most densely populated of geological terrains. Depletion in overconsolidated sediments, such as exhumed lithified sediments, can be substantially smaller.

[30] The predevelopment steady state head distribution was calculated first to provide the initial conditions from which drawdown (change in head) occurs in response to 26 pumping wells placed at various locations distributed throughout the confined aquifer. The steady state heads were controlled by areal recharge on the unconfined aquifer at a rate of 0.10 m/yr, topographic control by a representative dendritic network of rivers at the land surface (and

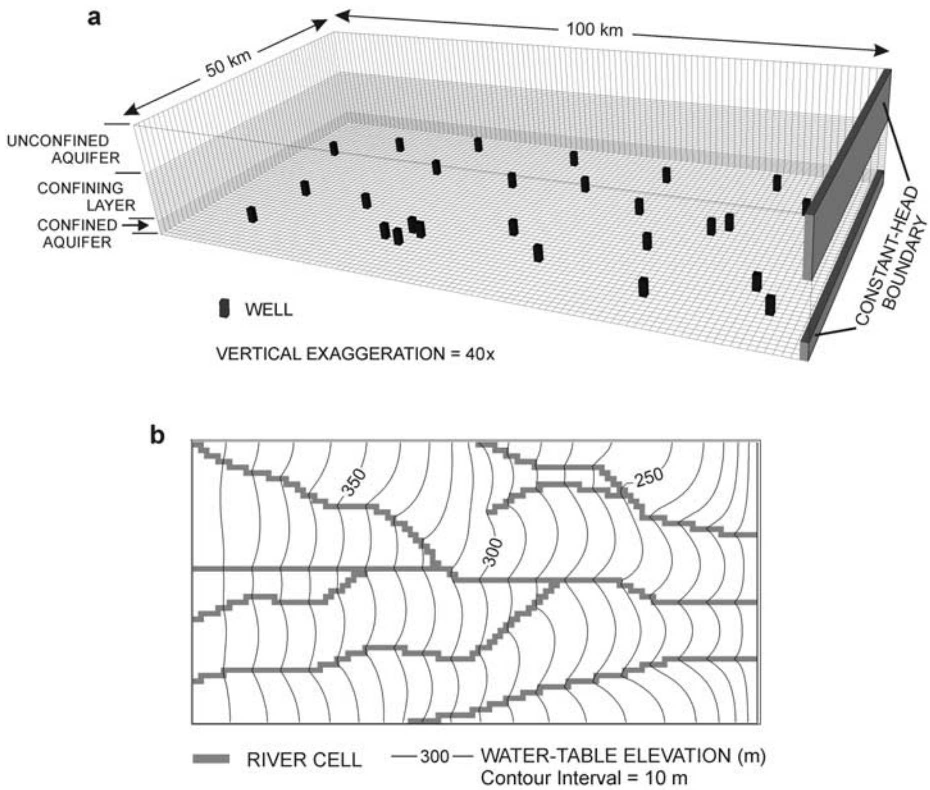

Figure 6. Grid used for three-dimensional simulation of hypothetical regional aquifer system in test problem, showing (a) three-dimensional perspective with locations of constant-head boundary conditions and wells and (b) map view with river locations and calculated water table elevations.

where river cells in the model, as mapped in Figure 6, represent head-dependent flux boundary conditions, and specified river elevations vary from 400 m on the left to 200 m on the right), and specified constant-head values of 200 m at the right edge of the unconfined aquifer and 210 m at the right edge of the confined aquifer (Figure 6). River cells can act as fluid sources or sinks, depending on local head gradients. The transient simulation calculated changes in head in response to 25 years of pumping from the 26 wells. The computed components of the water budget for both steady and transient conditions are listed in Table 2.

[31] The results show that the well pumpage (equivalent to a total volume of 4.10 km^3 over the 25-year period) was derived from or offset by (1) a reduction in groundwater storage (i.e., groundwater depletion), (2) decreased discharge (primarily at constant-head boundaries), and (3) increased recharge (from increased inflow from constant-head boundaries). About 71 percent of the pumpage is derived from groundwater storage. Over the 25-year period, a cumulative total of 2.92 km^3 of water has been removed from storage. Of that total, the bulk (2.88 km^3, or 98.5 percent) was derived from depletion in the confining layer (the remainder—0.04 km^3—was derived from depletion in the confined aquifer).

[32] After 25 years, the average drawdown at the top of the confined aquifer was 128.7 m. However, during the pumping period, the drawdown did not change in a stepwise fashion, nor in a linear fashion (Figure 7), so one issue in estimating depletion using the proposed method is selecting an appropriate value for t. First, we can check the criterion

Table 1. Model Parameters[a]

Hydrogeologic Unit	K_H, m/s	K_Z, m/s	S_y	S_s, 1/m	Thickness, m	Number of Model Layers	Number of Wells
Upper aquifer	10^{-4}	10^{-4}	0.2	10^{-6}	60–260[b]	1	0
Confining unit	10^{-8}	3×10^{-10}		10^{-4}	200	20	0
Deep confined aquifer	10^{-4}	10^{-4}		10^{-6}	60	10	26

[a]Total Q in transient simulation = -5.2 m^3/s = -4.1 km^3 over 25 years.
[b]Upper aquifer is unconfined, so thickness varies with water table elevation.

Table 2. Calculated Water Budgets From the Simulation Model[a]

Water Budget Element	Steady Flow		Transient	
	In	Out	In	Out
Storage[b]	0.0	0.0	2.92	0.0
Constant head	0.0	1.53	0.90	1.26
Wells	0.0	0.0	0.0	4.10
River leakage	0.89	11.74	0.89	11.73
Recharge	12.37	0.0	12.37	0.0
Total	13.27	13.27	17.09	17.09

[a]Values are cumulative volumes in km^3 over a 25-year period.
[b]Following the MODFLOW convention, water coming out of storage is considered an inflow to the groundwater system.

to decide whether the semi-infinite approach for a thick confining layer is applicable. In this case, for t = 25 years, z_d = 48.6 m. Because l = 200 m, $l > 2\ z_d$ and the approximation for the semi-infinite model is acceptable. The drawdown history shows that the response to constant pumping was a nonlinear rate of change in head, and that half of the drawdown (64.4 m) occurred after about 4.75 years. Thus, to estimate depletion, we would set t equal to the time since half of the drawdown occurred, or 20.25 years.

[33] We estimated the total depletion for the confining layer in this hypothetical groundwater system by solving equations (2) and (3). The results yield z_d = 43.5 m and V_w = 2.8 km^3. The depletion estimated using the approximate method proposed here is very close (about a 4 percent difference) to the actual depletion, as computed using the three-dimensional simulation model. If we did not have an accurate record of drawdown over time, and we had incorrectly set t to half the time since pumping started (t = 12.5 years), we would have underestimated the depletion at V_w = 2.2 km^3, or 75 percent of the actual total. In either case the errors are smaller than those likely to result from uncertainty in the values of K and S_s in the confining layers.

5. Field Applications

5.1. Thick Confining Layer Case: Dakota Aquifer

5.1.1. Background

[34] The Dakota and related sandstones in western-central North America comprise an extensive aquifer. The Dakota Aquifer is present throughout much of the Rocky Mountain foreland and in many areas is considered a classic example of an artesian aquifer system. This large aquifer system is extensively developed and has played a particularly important role in the settlement and subsequent economic development of South Dakota. Study of the aquifer system began with N.H. Darton around the turn of the last century [e.g., *Darton*, 1896, 1905, 1909], and helped shape current ideas about artesian aquifers [*Bredehoeft et al.*, 1983]. For over a century the Dakota Aquifer has been stressed by oil and gas exploration, mining activities, industrial expansion, and steadily increasing agricultural development [*Case*, 1984].

5.1.2. Hydrogeologic Setting

[35] The Dakota Aquifer underlies more than 171,000 km^2 of South Dakota [*LeRoux and Hamilton*, 1985]. In western South Dakota it consists of a lower sandstone unit called the Inyan Kara Group and an upper unit called the Newcastle Sandstone; these merge to form the Dakota Sandstone in the eastern part of the state [*Bredehoeft et al.*, 1983; *Schoon*, 1984] (Figure 8). Below the Inyan Kara is the Madison Limestone, an important carbonate aquifer. Low-permeability confining layers separate the major aquifers. The Dakota and Newcastle Sandstones are overlain by the relatively thick Cretaceous Shale confining layer [*Bredehoeft et al.*, 1983].

[36] Significant recharge to the Dakota and Madison Aquifers occurs where they crop out on the flanks of the Black Hills. The Dakota discharges at low elevations in the eastern part of the state. Discharge from pumped and flowing wells has become an important source of discharge from the aquifer system [*Case*, 1984].

[37] Substantial development of the aquifer system in the state had already begun by the early 1880s [*Bredehoeft et al.*, 1983]. By 1905, over 1,000 wells were flowing in the state 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].

[38] Large rates of head decline in the Dakota Aquifer occurred before 1915. The James River Lowland region of eastern South Dakota, for example, 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].

[39] Pumpage data presented by *Bredehoeft et al.* [1983], *Helgesen et al.* [1984], and *Case* [1984] indicates that the cumulative well discharge from the Dakota Aquifer system in South Dakota from predevelopment time to 1981 totaled about 19.7 km^3 of water. The history of development is incompletely documented, but *Bredehoeft et al.* [1983] estimate that well discharge in 1912 was 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.

[40] Uncontrolled flow from wells has been an important contributor to head declines [*Schoon*, 1971]. The casings in many early wells have been destroyed by corrosion and efforts to regain control of the flow have been largely unsuccessful. This unmetered flow has also made it difficult to accurately evaluate cumulative withdrawals from the Dakota system. J. Goodman (South Dakota Department of Environmental and Natural Resources, oral communication,

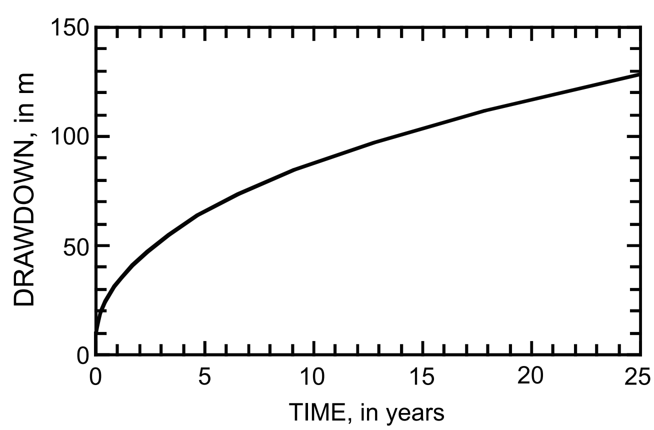

Figure 7. Areally averaged drawdown at the top of the confined aquifer in the synthetic groundwater system.

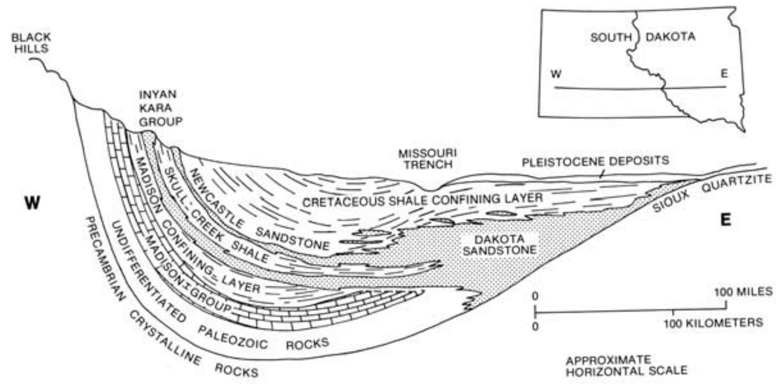

Figure 8. Schematic east-west cross section of major aquifers and confining layers in South Dakota (not to scale) (modified from *Bredehoeft et al.* [1983, Figure 1b]). Vertical scale is greatly exaggerated.

2003) stated that free flowing wells were one of the most significant sources of water loss to the system. The estimate of total withdrawals is probably reliable to within ±20 percent.

5.1.3. Depletion in the Aquifer

[41] Groundwater depletion from the Dakota Aquifer system in South Dakota was evaluated using potentiometric maps showing predevelopment [*Darton*, 1909] and 1980 conditions [*Case*, 1984, Figure 8]. 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.

[42] *Bredehoeft et al.* [1983, Table 3] summarize the estimates of dimensionless storage coefficient values for the Dakota Aquifer. Values ranged from 1.0×10^{-4} to 1.0×10^{-5}; we used the median value, 5.0×10^{-5}, to estimate depletion in this confined aquifer system.

[43] The area underlain by the Dakota Aquifer was discretized into a grid of square cells measuring 8.0 km on a side. The volume of groundwater depletion within each cell of the grid was estimated using GIS software as the product of the change in head times the area of the cell times the storage coefficient. Summation over all cells indicates 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. This represents about 2 percent of the estimated cumulative discharges of 19.7 km^3.

5.1.4. Depletion in the Confining Layers

[44] *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.

[45] In using our new approximation method to estimate long-term depletion from the confining units of the Dakota Aquifer system in South Dakota, we considered the overlying Cretaceous shale and the underlying Skull Creek Shale and Madison confining units. The Skull Creek Shale is essentially contained within the Dakota Aquifer, but it thins and pinches out in the central part of the state (Figure 8); thus, to simplify the analysis, we conservatively assumed that water derived from it was being removed from only one boundary.

[46] Because of uncertainty in the drawdown history of the aquifer system, there is also uncertainty in the time to apply the step drawdown. Thus depletions associated with three different effective times were calculated, namely 65, 50, and 35 years before 1980. These correspond to the years 1915, 1930, and 1945, and, if groundwater withdrawals began about 1880, they occur after 35, 50, and 65 years of development.

[47] *Neuzil* [1993] provides an estimate for the vertical hydraulic conductivity of the Cretaceous Shale confining unit of approximately 3×10^{-13} m/s, and model calibration analyses [*Bredehoeft et al.*, 1983, Table 3] indicated vertical hydraulic conductivity values of 4.6×10^{-12} m/s and 6.1×10^{-12} m/s for the Skull Creek and Madison confining units, respectively. A model calibration estimate for the value of specific storage was 1.6×10^{-4} m^{-1} and this was assumed to be a representative value for all confining units [*Bredehoeft et al.*, 1983, Table 3].

[48] The penetration distance (z_d) was calculated for each confining unit using equation (2) (Table 3), and the ratio of penetration distance to confining unit thickness was calculated. Because the z_d to l ratio was less than 0.25 in all cases, we can assume that semi-infinite conditions for a thick confining layer are appropriate for all three confining units.

[49] The total calculated volume of water removed from storage from the three confining units up to 1980 was

Table 3. Calculated Volume of Water Removed From Storage in Confining Units by 1980 for Various Effective Times, Assuming $S_s = 1.6 \times 10^{-4}$ m^{-1} and $\Delta\overline{H} = 47$ m

	Average Thickness, m	Area, km^2	z_d, m			Depletion, km^3		
			65 Years	50 Years	35 Years	65 Years	50 Years	35 Years
Cretaceous shale	476	190,000	2.0	1.7	1.4	2.9	2.6	2.1
Skull Creek confining unit	172	120,000	7.6	6.6	5.5	7.2	6.3	5.3
Madison confining unit	48	100,000	8.7	7.7	6.4	6.9	6.0	5.1
Total						17.0	14.9	12.5

Table 4. Sensitivity of Estimates of Total Depletion From Confining Units to Uncertainty in Key Hydraulic Parameters[a]

	Multiplication Factor for K		
S_s, m^{-1}	0.1	1.0	10.0
4.9×10^{-5}	2.6	8.2	25.8
1.6×10^{-4}	4.7	14.9	47.1
4.9×10^{-4}	8.2	25.8	59.0

[a]Here t = 50 years; depletion is in km^3.

estimated to range from 12.5 to 17.0 km^3 (Table 3), which represents a large fraction of the estimated cumulative discharge of 19.7 km^3. Because higher withdrawals prevailed at earlier times, it is quite likely that 35 years is too short a period of time in this case.

[50] Because of errors and uncertainty inherent in estimates of hydraulic properties, average effective drawdown, and the appropriate time, we evaluated the sensitivity of the estimates of total depletion to a reasonable range of values in these factors. The effects of varying the hydraulic properties (K and S_s) are shown in Table 4. Uncertainty for K is greater than that for S_s, reflecting the plausible ranges in values for these parameters. As a result, we considered a hundredfold range in K values versus a tenfold range in S_s values. Because the largest estimates of depletion are substantially greater than the 19.7 km^3 cumulative withdrawals estimated from well discharge data, they are probably unreasonably high. Conversely, reported discharge might be expected to underrepresent actual discharge so that the actual withdrawals might exceed the 19.7 km^3 estimate. Uncertainty in $\Delta\overline{H}$ and time induce a relatively small variation in the estimated total depletion.

[51] If we accept the central value of depletion (14.9 km^3) as being correct, it would imply that 76 percent of the withdrawals were derived from depletion in confining units. This can be compared with the independent estimate that approximately 2 percent of total withdrawals derived from depletion within the aquifer itself, which implies that about 98 percent of the water removed from storage was derived as leakage from (and depletion of) confining units. Because of limited potential in this deep system for increasing recharge in response to pumping, related to its relatively small outcrop area, we believe that the estimate of 76 percent derived from the confining units is probably an underestimate. Nevertheless, the results of this "real-world" application of the proposed approximation method indicate that the method can provide a reasonably reliable (within bounding limits) estimate of long-term groundwater depletion.

[52] In this example, independent estimates of confining layer properties were available from prior studies of the Dakota Aquifer and its confining shales. It is reasonable to ask how estimates of K and S_s one might obtain from the plots in Figures 4 and 5 compare with these relatively sophisticated earlier characterizations. The answer, of course, depends partly on the type and quality of other data that are available. In the case of the Pierre Shale, many other data are available. The Pierre is a late Cretaceous claystone with a total porosity that ranges between ∼0.28 and ∼0.33. Its clay content is as high as 80 percent, and a significant

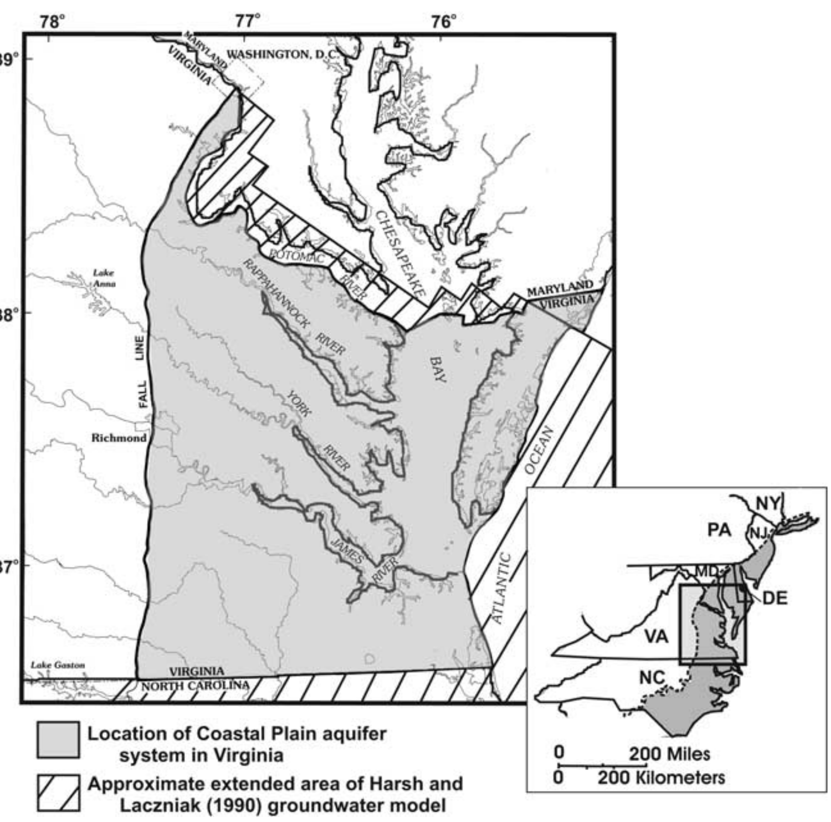

Figure 9. Location of coastal plain aquifer system in Virginia (modified from *Meng and Harsh* [1988, Figure 2]).

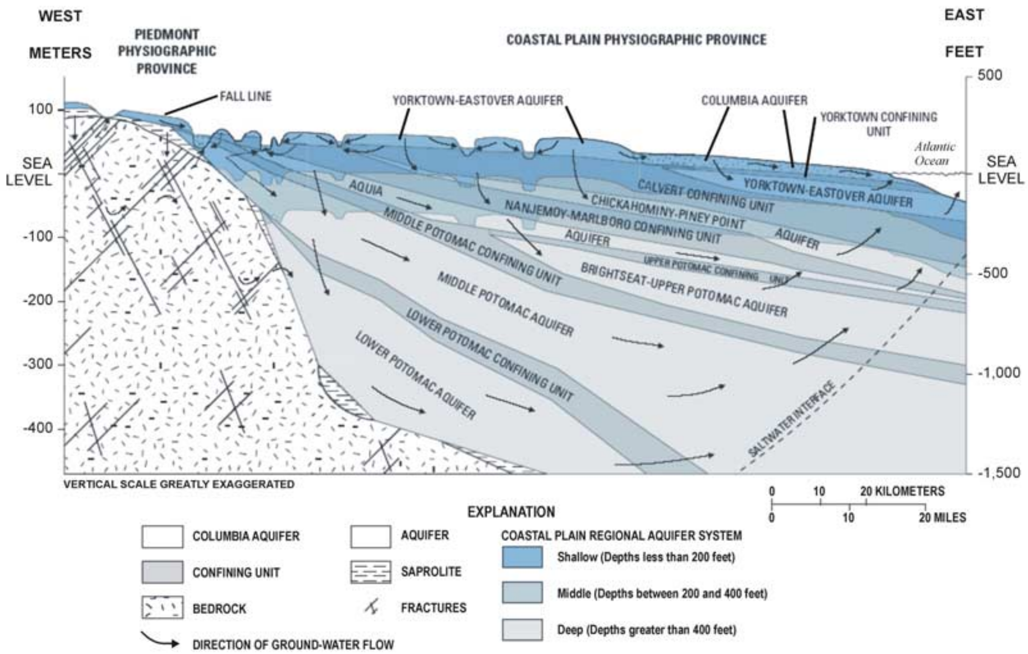

Figure 10. Generalized hydrogeological section of the Virginia coastal plain aquifer system [*Nelms et al.*, 2003].

fraction is smectite. At a porosity of 0.30, the prior determination of $K = 3 \times 10^{-13}$ m/s falls between "high" and "very high" clay content in Figure 4, a reasonable description of the Pierre lithology. The extreme values of K used in the sensitivity analysis (Table 4), 3×10^{-14} m/s and 3×10^{-12} m/s, fall on "very high" and between "high" and "moderate" clay contents respectively, which reasonably bracket the Pierre lithology. It is tempting to conclude that an estimate of K based on Figure 4 would be reasonably accurate, but a better judgment would require more formal tests. In addition, estimates would likely be poorer without accurate porosity and lithology data.

[53] At a porosity of 0.30, the prior estimate of S_s for the Pierre, 1.6×10^{-4} m^{-1}, falls in the "normally consolidated" region of Figure 5 despite the fact that the Pierre Shale is overconsolidated. Figure 5 suggests a value for S_s that is approximately one third of this value (at the high extreme of the "overconsolidated" range to account for its high clay content). This apparent underestimate may result from the unusually high smectite content of the Pierre Shale tending to make it deform viscoelastically.

5.2. Thin Confining Layer Case: Coastal Plain Aquifer

5.2.1. Background

[54] The coastal plain aquifer system of eastern Virginia and adjacent parts of the coastal plain of Maryland and North Carolina (Figure 9) covers approximately 44,000 km^2 [*Meng and Harsh*, 1988]. Groundwater from the system is an important source of water for municipal, domestic, industrial, and agricultural uses [*Harsh and Laczniak*, 1990].

[55] There are several major pumping centers with cones of depression in Virginia, and water levels have declined as much as 60 m from predevelopment levels. Many parts of the Virginia coastal plain are experiencing significant population growth and groundwater withdrawals are increasing as a result. Agricultural and industrial demands for groundwater are also growing. A groundwater model was developed by *Harsh and Laczniak* [1990] to help to better understand and manage this important aquifer system.

5.2.2. Hydrogeologic Setting

[56] The Virginia coastal plain aquifer system is an eastward thickening sedimentary wedge atop a basement surface that slopes gently eastward (Figure 10). The sediments are more or less compacted by overburden but not lithified and range in age from Cretaceous to Recent; their maximum thickness exceeds 1,800 m [*Meng and Harsh*, 1988]. The sedimentary wedge is composed of gravels, sands, silts, clays, and varying amounts of shell material. [*Meng and Harsh*, 1988] provide a detailed description of the hydrogeology of the study area. Although the hydrogeologic framework has recently been revised to some extent [*McFarland and Bruce*, 2006], we base our analysis on that of [*Meng and Harsh*, 1988] and on the related model conceptualization of *Harsh and Laczniak* [1990].

[57] The groundwater system consists of an unconfined aquifer underlain by a series of confined sand aquifers separated by clayey confining units. Most of the hydraulic properties of the aquifers were determined through aquifer tests. Data on the hydraulic properties of individual confining units are limited, although some K values have been estimated from laboratory tests of core samples [*Harsh and Laczniak*, 1990]. *Harsh and Laczniak* [1990] note that data defining the storage properties of confining units are generally not available, so they assessed the model's sensitivity to transient leakage by assuming two different values of S_s for the confining units: 3.3×10^{-6} m^{-1}, which is probably too low because it reflects just the compressibility of water and a minimal amount of matrix compressibility, and S_s =

Table 5. Penetration Distance z_d Relative to Thickness l of Confining Unit as a Function of Time Before 1980, Assuming $S_s = 1.6 \times 10^{-4}$ m^{-1}

Hydrogeologic Unit	z_d, m		z_d:l		Boundaries
	t = 20 Years	t = 25 Years	t = 20 Years	t = 25 Years	
Nanjemoy-Marlboro confining unit	27.0	30.2	0.88	0.99	1
Brightseat–Upper Potomac confining unit	24.5	27.4	2.29	2.56	2
Middle Potomac confining unit	23.5	26.3	1.93	2.15	2
Lower Potomac confining unit	21.1	23.6	2.77	3.10	2

3.3×10^{-4} m^{-1}, which would be characteristic of normally consolidated clayey sediments. *Hansen* [1977] reports a number of S_s values from consolidation tests on cores obtained from confining layers at nearby sites in Maryland. At effective stresses of interest here, his data show S_s values of $1.2–3.0 \times 10^{-4}$ m^{-1} for the Marlboro Clay and $1.5–2.5 \times 10^{-4}$ m^{-1} in the Brightseat–Upper Potomac confining unit. *Pope and Burbey* [2004] estimated specific storage values for skeletal compressibility in confining layers in the Virginia coastal plain from compaction data as great as $S_s = 1.0 \times 10^{-4}$ m^{-1} for the shallower confining units and as great as 1.5×10^{-5} m^{-1} for the deeper confining units. On the basis of this prior information, we assess depletion from confining layers in the Virginia coastal plain assuming a value of $S_s = 1.6 \times 10^{-4}$ m^{-1} and a range between 3.3×10^{-5} and 3.3×10^{-4} m^{-1}. These values approximately bracket those in Figure 5 for normally consolidated sediments at a porosity of ~0.30. This is consistent with the area's geologic history, which suggests that the sediments are not significantly overconsolidated.

[58] Groundwater use from the confined aquifers began in Virginia by the late 1800s, and the estimated annual discharge from flowing wells in the study area ranged from 15,000 to 38,000 m^3/d during the water years 1891 to 1945. Flowing wells were a significant source of supply until 1935, when water levels in deeper confined wells fell below land surface [*Harsh and Laczniak*, 1990]. Withdrawals substantially increased in the state after 1955 and were approximately 380,000 m^3/d by the water year 1980. In total, it was estimated from withdrawal data that about 4.5 km^3 of groundwater was withdrawn from the Virginia coastal plain aquifer system between 1891 and 1980 [*Harsh and Laczniak*, 1990].

5.2.3. Depletion in the Aquifers

[59] Simulations of this aquifer system conducted by *Harsh and Laczniak* [1990] used a quasi three-dimensional model; confining units were not directly simulated, but were represented instead by equivalent vertical conductances between aquifer layers. This approach inherently ignores horizontal flow, transient changes in head, and, most importantly in the present context, storage in the confining units. Each of the aquifers was represented as a layer in the model.

[60] The transient simulation was divided into ten "stress periods" during each of which all hydraulic stresses, including rates of groundwater withdrawal, remain constant. The stress periods spanned the time between 1891 (assumed to represent predevelopment conditions) and 1980. A dimensionless storage coefficient value of 1.0×10^{-4} was used for all confined aquifers in the model. Other hydraulic properties, boundary conditions, and calibration methods are described in detail by *Harsh and Laczniak* [1990].

[61] The calibrated transient simulation computed a water budget for the simulation over the 90-year period 1891–1980. It indicated that 0.18 km^3 was removed from storage in aquifers during the study period.

5.2.4. Groundwater Depletion in Confining Units

[62] Because large cones of depression occur in several of the groundwater system's aquifers, it is expected that decreases in storage in the intervening confining units has occurred and may represent a relatively large source of water to wells in the area. In this groundwater system, confining units are generally thinner than the aquifers (Figure 10). As a multiple-aquifer system, water may drain from a confining unit toward both its upper and lower boundaries because withdrawals and drawdown are occurring in aquifers on both sides of the confining units.

[63] To estimate depletion in the confining units, we considered the aquifers with the most development and greatest head declines (Lower and Middle Potomac, Brightseat–Upper Potomac, and Aquia Aquifers) and their adjoining confining units (Lower and Middle Potomac, Brightseat–Upper Potomac, and Nanjemoy-Marlboro confining units). Because of uncertainty in confining unit properties, we evaluated an order of magnitude range in specific storage values as noted above (3.3×10^{-5} m^{-1} to 3.3×10^{-4} m^{-1}) and values of K that are five times higher and lower than the average values used by *Harsh and Laczniak* [1990, Table 6]. We note that the nominal K value adopted corresponds, approximately, with the "moderate clay" line in Figure 4 for a porosity of ~0.30. There is uncertainty about the appropriate time over which drawdown is effective. The model analyses by *Harsh and Laczniak* [1990] indicate that about half the drawdown occurred 20 to 25 years before 1980, so we evaluated both values as effective times.

[64] Initially assuming a value of $S_s = 1.6 \times 10^{-4}$ m^{-1} (half of the maximum assumed value) for a base case analysis, the ratio of penetration distance (z_d) to confining unit thickness (l) was calculated for each of the confining units using equation (2) (see Table 5). In this case, because the z_d to l ratio was always greater than 0.25 (or 0.5 if assuming penetration at only one boundary), the analysis for a "thin" confining layer was applied and depletion was estimated using equation (7).

[65] Maps showing simulated hydraulic heads in 1891 and 1980 [*Harsh and Laczniak*, 1990, Figures 36, 38–40, 53, and 55–57] for the aquifers were used to compute values of H_0, or the average head changes over the area. GIS tools were used to construct maps of head changes for each of the aquifers in the major pumping zones between 1891 and 1980.

[66] The total volume of water removed from storage from the four confining units through 1980 was thereby calculated (as shown in Table 6) to be about 3.5 km^3, assuming $S_s = 1.6 \times 10^{-4}$ m^{-1} and average K values. Accounting for uncertainty in parameter values indicates that the depletion may be as small as about 0.7 km^3 or as large as about 6.9 km^3.

Table 6. Calculated Volume of Water Removed From Storage in Confining Units, Assuming S_s = 1.6 × 10^{-4} m^{-1} and t = 20 Years[a]

Hydrogeologic Unit	Average Thickness, m	K,[b] m/d	Area, km^2	$\Delta\overline{H}$, m	$(kt/s, l^2)^{1/2}$	$f(t)$	Depletion, km^3
Nanjemoy-Marlboro confining unit	30.5	1.6E-05	33,000	–	0.88	1.0	1.01
Aquia aquifer	–	–	–	12.2	–	–	–
Brightseat–Upper Potomac confining unit	10.7	1.3E-05	34,000	–	2.29	1.0	0.76
Brightseat–Upper Potomac aquifer	–	–	–	13.5	–	–	–
Middle Potomac confining unit	12.2	1.2E-05	38,300	–	1.93	1.0	1.07
Middle Potomac aquifer	–	–	–	14.4	–	–	–
Lower Potomac confining unit	7.6	1.0E-05	25,900	–	2.77	1.0	0.64
Lower Potomac aquifer	–	–	–	24.9	–	–	–
Total							3.47

[a]Thickness and K information for confining units from *Harsh and Laczniak* [1990, Tables 6 and 13].
[b]Read 1.6E-05 as 1.6 × 10^{-5}.

[67] The reasonableness of the S_s values was checked by calculating subsidence with equation (12). For S_s = 1.6 × 10^{-4} m^{-1}, one obtains an average total subsidence due to compaction in the four confining layers of ~21 cm. *Pope and Burbey* [2004] estimated total subsidence, from initiation of development to 1995, to be ~14 cm at two locations in the Virginia coastal plain. The estimates apply near pumping centers, which tends to make them larger than average, but are located toward the western boundary of the study area where the aquifers and confining layers are relatively thin, which tends to make them smaller than average. Because of the differences, a simple comparison is difficult, but the similarity of the estimates suggest to us that our chosen range of values for S_s is reasonable.

[68] The calculated volumes of confining layer depletion were much less sensitive to an order of magnitude uncertainty in K than in S_s because the volumetric calculation of equation (7) only depends on the value of K to the extent that K affects $f(t)$, and $f(t)$ approaches or equals 1.0 in most cases. For the same reason, the calculated volumes of depletion for the 20 and 25 year durations are essentially equal. The drawdown in all of the confining units penetrates more than half of the confining unit thickness; thus all confining units are under what might be termed "finite" conditions. Under finite conditions, the penetration distance (z_d) and elapsed time (t) are not used in equation (7) to calculate a final depletion volume.

[69] If the specific storage is 1.6 × 10^{-4} m^{-1}, the total depletion from aquifers and confining units in the Virginia coastal plain aquifer system would be on the order of 3.7 km^3, which represents about 80 percent of total withdrawals. For this condition, about 95 percent of the total water removed from storage was derived from the confining layers adjacent to the most developed aquifers. Of course, uncertainty in the properties of the confining units makes this estimate subject to error.

6. Concluding Remarks

[70] To the extent that our site analyses are representative, confining layers may contribute most to nearly all of groundwater withdrawn from storage in sedimentary terrains and probably will continue to do so for some time to come. This is perhaps unsurprising in view of the relatively large storativity of clays, shales, and similar media, but generally has been unappreciated and often ignored. Because confining layer depletion is so large, it must be considered in any attempt to estimate groundwater withdrawals or determine budgets for hydrologic systems. Confining layer depletion is also important because it is often largely irreversible. Unless significantly "overconsolidated," that is, under smaller stresses than in the past, argillaceous media can recover only a fraction (perhaps 10 percent typically) of the pore volume lost when water is released from storage and the pore structure collapses. This means that significant quantities of water are being removed from long-term (on the order of 10^3 to 10^7 years) subsurface storage and being added to other components of the hydrologic cycle. The change can be considered largely permanent in human terms.

[71] The rate at which the transfer occurs is likely to increase in coming decades. Global dependence on groundwater is growing, especially in developing countries, but evaluation of groundwater systems in developing areas is particularly challenging because of the sparsity of data. The methods for estimating confining layer depletion proposed in this paper are usable even where relatively few hydrogeologic data are available. Indeed, one can at least make preliminary estimates with only the most basic information on formation thickness, lithology, and water-level trends. If available, more specific data, particularly confining layer K and S_s, is easily incorporated to refine estimates.

[72] Our test of the method on a hypothetical confined regional aquifer system, for which total depletion was calculated using a numerical simulation, demonstrated that the new estimation method is highly accurate if the properties and boundaries of the system are known. In most applications the greatest source of error in the proposed method will almost certainly be in the estimation of confining layer properties. Typically, K is the most significant source of error although, as the Virginia coastal plain example showed for the case of thin confining layers, S_s is sometimes more important. Because of this, analyses of depletion may derive most benefit from efforts to determine confining layer K or S_s, depending on the local situation. Thus the proposed method can also help direct sparse investigative resources.

[73] The results of the South Dakota site analysis also reinforce a general observation regarding groundwater depletion. While significant amounts of water are being derived from confining layer storage, the flow regime is transient and evolving to a new steady state. Where recharge is limited, as in the Dakota aquifer system, well withdrawals are almost entirely offset by depletion of storage in confining layers, and the availability of groundwater may decrease with time as the confining layer storage is depleted.

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L. F. Konikow and C. E. Neuzil, U.S. Geological Survey, 431 National Center, 12201 Sunrise Valley Drive, Reston, VA 20192, USA. (lkonikow@usgs.gov)