ground water

New Insights from Well Responses to Fluctuations in Barometric Pressure

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Abstract

Hydrologists have long recognized that changes in barometric pressure can produce changes in water levels in wells. The barometric response function (BRF) has proven to be an effective means to characterize this relationship; we show here how it can also be utilized to glean valuable insights into semi-confined aquifer systems. The form of the BRF indicates the degree of aquifer confinement, while a comparison of BRFs between wells sheds light on hydrostratigraphic continuity. A new approach for estimating hydraulic properties of aquitards from BRFs has been developed and verified. The BRF is not an invariant characteristic of a well; in unconfined or semi-confined aquifers, it can change with conditions in the vadose zone. Field data from a long-term research site demonstrate the hydrostratigraphic insights that can be gained from monitoring water levels and barometric pressure. Such insights should be of value for a wide range of practical applications.

Introduction

For more than three centuries, scientists have known that changes in barometric pressure can produce changes in water levels in wells (Pascal 1973). Although the phenomenon has long been recognized, the underlying mechanisms have only been clarified much more recently (Jacob 1940; Weeks 1979; van der Kamp and Gale 1983; Rojstaczer 1988; Spane 2002). For a confined aquifer, a change in the barometric pressure load on the land surface is transmitted downward, grain to grain, near instantaneously to the interface between the confining unit and the aquifer. Part of the load is then borne by the pore water and part is borne by the aquifer framework (Figure 22B in Ferris et al. 1962). By contrast, the entire load is borne by the water column in a well open to the atmosphere. The resulting pressure difference induces water flow between the aquifer and the well, leading to the commonly observed inverse relationship between barometric pressure and water level (Figure 1). The magnitude of the water-level change primarily depends on how the load is shared between the pore water and the aquifer framework, although the properties of the aquifer and overlying units and the characteristics of the well (e.g., well diameter and degree of well development) can also play important roles. A different mechanism controls water-level responses in an unconfined aquifer. In that case, access to the free water surface minimizes pore-pressure changes produced by the grain-to-grain transmission of the surface load; the primary control on responses is the downward propagation of air pressure through the pores of the vadose zone (Figure 22A in Ferris et al. 1962). For shallow water tables, this propagation can occur so quickly that the pressure difference between the well and the aquifer is negligible and, as a result, there is virtually no flow between the two. If the propagation is delayed, due to the depth to water and/or conditions in the vadose zone, the inverse relationship of Figure 1 is observed (Weeks 1979; Hare and Morse 1997; Spane 2002).

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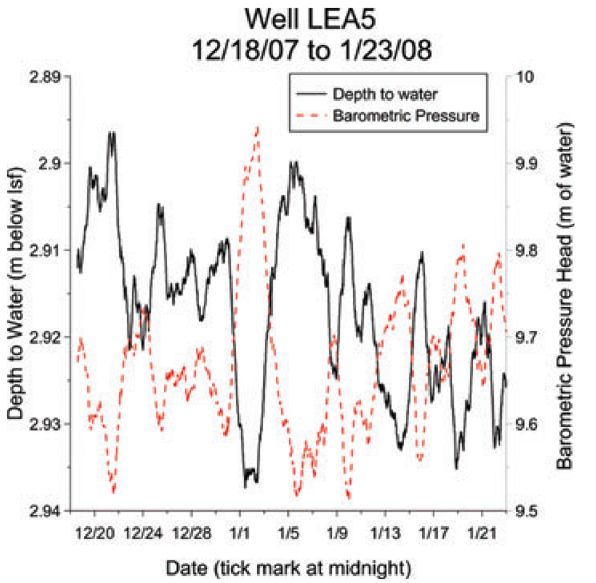

Figure 1. Depth to water from land surface and barometric pressure head for well LEA5 at the Larned Research Site for a period in the winter of 2007 to 2008. Depth to water is plotted increasing downward to display the inverse relationship between water level and barometric pressure; spans of the left and right *y*-axis differ by a factor of ten.

Hydrologists have traditionally characterized the relationship between barometric pressure and water level using the ratio of the change in water level to the change in barometric-pressure head, which is termed the barometric efficiency (BE) and, by sign convention, varies between zero and one (Jacob 1940). In a confined aquifer, a BE value near zero indicates that most of the load is borne by the pore water, while a value near one indicates most is borne by the aquifer framework. Although BE has proven to be an effective means of characterizing the short-term response of a well to a change in barometric pressure, the barometric response function (BRF) is a more effective means for characterizing the longer-term response (Rasmussen and Crawford 1997; Spane 2002). The BRF, which can be determined through a regression convolution procedure (Furbish 1991; Rasmussen and Crawford 1997; Toll and Rasmussen 2007), characterizes the water-level response over time to a step change in barometric pressure, essentially BE as a function of time since the imposed load. The BRF has been successfully used to remove the effect of barometric-pressure changes on water levels (Toll and Rasmussen 2007), a critical step, for example, in the interpretation of pumping tests when drawdown is small (Batu 1998). Rasmussen and Crawford (1997) and Spane (2002) discuss the impact of well conditions and site hydrogeology on BRFs and propose characteristic BRF forms for certain hydrogeologic settings (confined and deep unconfined aquifers). Spane (2002) reviews a number of time-domain (e.g., Toll and Rasmussen 2007) and frequency-domain (e.g., Quilty and Roeloffs 1991) approaches that have been proposed for removing the effect of barometric-pressure changes from water-level data series and concludes that the BRF (a time-domain method) is particularly effective for this purpose.

Water-level responses to fluctuations in barometric pressure have also been used to estimate subsurface properties. Specific storage can be determined from BE if estimates of aquifer porosity and pore-water compressibility are available (Jacob 1940; Batu 1998). Time- and frequency-domain methods, often in a type-curve format, have been developed for determining hydraulic properties from water-level responses to barometric-pressure changes (e.g., Weeks 1979; Rojstaczer 1988; Rojstaczer and Riley 1990; Evans et al. 1991; Furbish 1991). These methods, however, have yet to be widely adopted.

The purpose of this paper is to extend the earlier work to show how the BRF can be utilized to glean further hydrogeologic insights. Our primary emphasis is on gaining insights into the low permeability unit (henceforth, aquitard) that overlies a semi-confined aquifer. Given the utility of the BRF for removing the impact of barometric-pressure changes from water-level data, we also explore its value for estimating subsurface properties. We propose a new approach for estimating aquitard hydraulic conductivity (K) by fitting theoretical responses to field-determined BRFs. We demonstrate these concepts using field data from a long-term research site and discuss how the BRF can also be used to gain insights into conditions in unconfined aquifers and, potentially, in the vadose zone. Although BE is considered an invariant parameter of a well, we show that a BRF can change as a function of conditions in the vadose zone.

Field Site Overview

The field component of this study took place at the Larned Research Site (LRS; 38.2°N latitude, 99.0°W longitude) of the Kansas Geological Survey (Figure 2a). The primary focus here is on three LRS wells (LWC2, LEA5, and LEC2 in Figure 2a; all 0.102 m inner diameter) screened in the semi-confined High Plains aquifer (interval A in Figure 2b), with a secondary focus on adjacent wells screened at the bottom of the unconfined Arkansas River alluvial aquifer (interval B in Figure 2b). Each well has an integrated pressure-transducer and datalogger unit submerged in the water column (miniTroll, In-Situ, Inc., Fort Collins, Colorado); pressure readings are taken every 15 min. Gauge (relative to atmospheric pressure) and absolute pressure transducers were used in this work; Price (2009) describes how both types of transducers can be utilized for assessing water-level responses to barometric-pressure changes. Atmospheric pressure is recorded with on-site barometers at the top of wells (baroTroll, In-Situ, Inc.) and at a weather station (Onset Computing Corp., Bourne, Massachusetts); readings are also taken every 15 min. Groundwater in the vicinity of the LRS is primarily used for irrigation, so the vast majority of pumping is during the growing season (mid-March to mid-October). In most years, water levels recover from seasonal pumping by mid-December (Figure 3). The Arkansas River, which was flowing at the

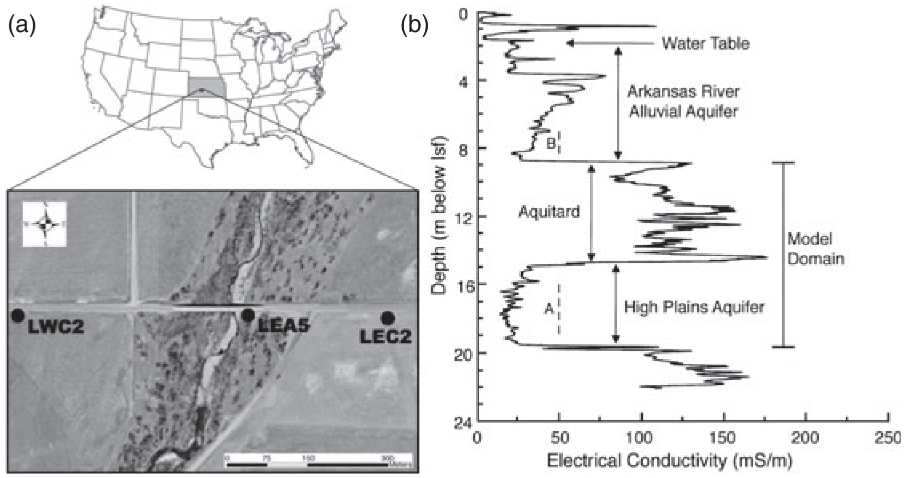

Figure 2. (a) Location map and aerial photo of the Larned Research Site (LRS). Aerial photo (year 2000) only shows the wells discussed in paper, watercourse in photo is the Arkansas River. (b) High-resolution direct-push electrical conductivity (EC) log from near the center of the LRS riparian zone. Wells in the High Plains aquifer are screened across the interval marked A, while adjacent wells in the lower portion of the Arkansas River alluvial aquifer are screened across the interval marked B. At this site, high EC values indicate clays and low values indicate sands and gravels. Bar on right side shows the vertical extent of the model discussed in the text.

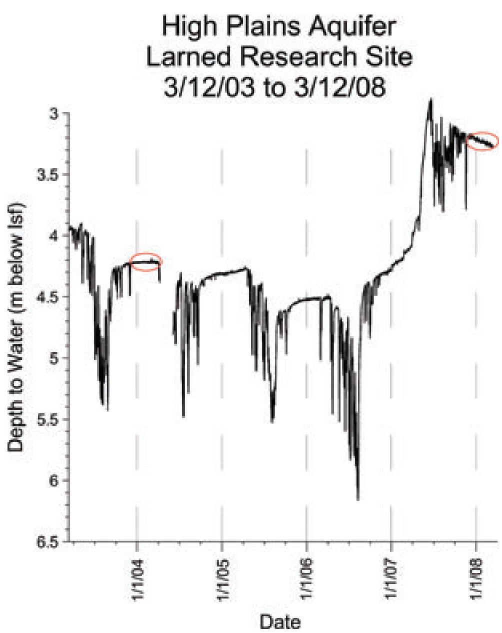

Figure 3. Depth to water versus time plot for LRS well LEC2 (see aerial photo in Figure 2a for location); well LEC2 has the most continuous record for this period of the three wells shown in Figure 2a. The ovals indicate the time intervals used for the analyses discussed in this paper. Note the pronounced seasonality of groundwater pumping in the vicinity of the LRS and the period of significant recharge beginning in the latter half of 2006.

time of the photograph in Figure 2a, is intermittent at the LRS and had little to no flow for the period of the analyses discussed here. Note that all of the LRS wells in the semi-confined High Plains aquifer display a pronounced water-level response to changes in barometric pressure (e.g., Figure 1). Small responses (a few mm or less) to precipitation loading (e.g., Rasmussen and Mote 2007) have been observed, but had no influence on the analyses. Water-level responses to stream-stage loading (e.g., Boutt 2010) have also been observed, but were negligible during the period of the analyses.

Methodology

BRFs were determined from the LRS water-level and barometric-pressure data using the regression convolution approach of Furbish (1991) and Rasmussen and Crawford (1997). This approach, which has been implemented in a spreadsheet format (e.g., Halford 2006; Toll and Rasmussen 2007), assumes that temporal changes in a detrended (removal of linear trend in this work) time series of water levels (equally spaced in time) can be represented as:

$$\Delta W(t) = \sum_{i=0}^{m} \alpha_i \Delta B(t - i\Delta t) + \sum_{i=0}^{n} \beta_i \Delta E(t - i\Delta t) \quad (1)$$

where $\Delta W(t)$ is the change in detrended water-level elevation (L) between time t and the previous time when a measurement was taken $(t - \Delta t)$; $\Delta B(t - i\Delta t)$ and $\Delta E(t - i\Delta t)$ are the changes in the detrended barometric-pressure head (L) and earth-tide gravity potential (L/T^2), respectively, between time $t - i\Delta t$ and the previous time when a measurement was taken $[t - (i + 1)\Delta t]$; α_i and β_i are the unit (impulse) barometric-pressure and earth-tide response functions at lag i, respectively; m is the maximum number of time lags for the barometric pressure response; n is the maximum number of time lags for the earth-tide response; and Δt is the time between adjacent measurements. The underlying assumption of this implementation of the BRF approach is that $\Delta W(t)$ is only a function of changes in barometric pressure and the earth-tide gravity potential, that is, the impact of other mechanisms on water levels is negligible or can be removed by detrending the water-level data. This assumption appears quite reasonable for systems such as the High Plains aquifer in western Kansas where recharge is very small and pumping is seasonal in nature. The earth-tide gravity potentials for the LRS are obtained with TSOFT, which generates synthetic earth-tide records for a given location (Van Camp and Vauterin 2005). Earth tides do have a small effect on water levels at the LRS, so they are incorporated in the analysis following the approach outlined by Toll and Rasmussen (2007). The focus of this paper, however, is on the much larger fluctuations induced by changes in barometric pressure and the insights that can be derived from them.

Ordinary least-squares linear regression is used to estimate α_i and β_i, and the BRF for lag j, A_j, is obtained by summing the α_i terms up to that lag:

$$A_j = \sum_{i=0}^{j} \alpha_i \tag{2}$$

with the standard error given as

$$\hat{\sigma}_{A_j} = \sqrt{\sum_{i=0}^{j}\sum_{k=0}^{j} C_{i,k}} \tag{3}$$

where C is the variance-covariance matrix for the α_i estimates (e.g., Abraham and Ledolter 1983). A small BE and finite transducer resolution can result in occasions when $\Delta W(t)$ is incorrectly truncated to zero. In order to reduce such truncation errors, the above approach can be extended to incorporate water-level and barometric-pressure changes over multiples of Δt.

We have developed theoretical BRFs for an aquifer system similar to that of Figure 2b using a semi-analytical solution for a one-dimensional (1-D) vertical representation of a two-layer (aquitard and aquifer—see vertical bar on Figure 2b) configuration. The governing equations, which are based on the development of van der Kamp and Gale (1983), are:

$$\frac{\partial h_1}{\partial t} - \gamma_1\, h_0\, \delta(t) = D_1 \frac{\partial^2 h_1}{\partial z^2} \tag{4a}$$

$$\frac{\partial h_2}{\partial t} - \gamma_2\, h_0\, \delta(t) = D_2 \frac{\partial^2 h_2}{\partial z^2} \tag{4b}$$

where h_0 is the change in barometric-pressure head at the land surface (L), $\delta(t)$ is the delta function (/T), and h_i, D_i, and γ_i are head deviation from static (L), hydraulic diffusivity (L^2/T), and loading efficiency (1 − BE) [−] for the aquitard (1) and aquifer (2), respectively, and z is depth ([L], 0 at aquifer-aquitard interface and increases downward). The hydraulic diffusivity is the ratio of hydraulic conductivity [K_i, (L/T)] over specific storage [S_{si}, (/L)]. The loading efficiency term [$\gamma_i h_0 \delta(t)$] represents the pressurization of the pore water via the near-instantaneous grain-to-grain transmission downward of the surface load. Groundwater flow is primarily driven by the boundary condition at the top of the aquitard, which is a function of the pressure propagation through the pores of the overlying vadose zone and unconfined aquifer.

The initial condition for the system is static heads in the aquifer and aquitard (i.e., h_i is zero); the boundary conditions are a constant head at the top of the aquitard (produced by the propagation of a step change in barometric-pressure head to the bottom of the overlying unconfined aquifer), zero flow at the bottom of the aquifer, and continuity of head and flow at the aquitard-aquifer interface.

A solution for the governing equations, 4a and 4b, and auxiliary conditions is obtained using standard integral-transform techniques. The system of equations is transformed into Laplace space and solved to yield the transform-space solution:

$$\overline{h}_i(z,p) = \frac{h_0}{p} f_i(z,p) \tag{5a}$$

where $\overline{h}_i$ is the Laplace transform of h_i, p is the Laplace-transform variable,

$$f_1(z,p) = \frac{\left(\frac{h_{UB}}{h_0} - \gamma_1\right) K_r \sqrt{\frac{D_2}{D_1}} \operatorname{sech}\left(\sqrt{\frac{p}{D_1}}\, l\right) - (\gamma_1 - \gamma_2) \tanh\left(\sqrt{\frac{p}{D_1}}\, l\right) \tanh\left(\sqrt{\frac{p}{D_2}}\, a\right)}{K_r \sqrt{\frac{D_2}{D_1}} + \tanh\left(\sqrt{\frac{p}{D_1}}\, l\right) \tanh\left(\sqrt{\frac{p}{D_2}}\, a\right)} \cosh\left(\sqrt{\frac{p}{D_1}}\, z\right)$$
$$- \frac{\left(\frac{h_{UB}}{h_0} - \gamma_1\right) \operatorname{sech}\left(\sqrt{\frac{p}{D_1}}\, l\right) + (\gamma_1 - \gamma_2)}{K_r \sqrt{\frac{D_2}{D_1}} + \tanh\left(\sqrt{\frac{p}{D_1}}\, l\right) \tanh\left(\sqrt{\frac{p}{D_2}}\, a\right)} \tanh\left(\sqrt{\frac{p}{D_2}}\, a\right) \sinh\left(\sqrt{\frac{p}{D_1}}\, z\right) + \gamma_1 \tag{5b}$$

for the aquitard ($-l \le z \le 0$),

$$f_2(z,p) = \frac{\left(\frac{h_{UB}}{h_0} - \gamma_1\right) \operatorname{sech}\left(\sqrt{\frac{p}{D_1}}\, l\right) + (\gamma_1 - \gamma_2)}{1 + \frac{1}{K_r}\sqrt{\frac{D_1}{D_2}} \tanh\left(\sqrt{\frac{p}{D_1}}\, l\right) \tanh\left(\sqrt{\frac{p}{D_2}}\, a\right)} \times \frac{\cosh\left[\sqrt{\frac{p}{D_2}}(z-a)\right]}{\cosh\left(\sqrt{\frac{p}{D_2}}\, a\right)} + \gamma_2 \tag{5c}$$

for the aquifer ($0 \leq z \leq a$), K_r is K_1/K_2, h_{UB} is the constant head at the top of the aquitard (L), a is aquifer thickness (L), and l is aquitard thickness (L). The derivation of Equation 5a is given in the Appendix.

The real-space form of Equation 5a is generated using the inversion algorithm of Stehfest (1970). The expression for the head in the semi-confined aquifer is

$$h_2(z,t) \approx h_0 \sum_{n=1}^{N} \frac{V_n}{n} f_2\left(z, \frac{\ln 2}{t} n\right) \tag{6}$$

where V_n is the coefficient for the Laplace inversion and N is the number of terms in the Stehfest summation (14 for this work). The BRF for a well in the semi-confined aquifer is

$$A(z,t) = 1 - \frac{h_2(z,t)}{h_0} \approx 1 - \sum_{n=1}^{N} \frac{V_n}{n} f_2\left(z, \frac{\ln 2}{t} n\right) \tag{7}$$

Equation 7 assumes a constant head (h_{UB}) at the top of the aquitard. Temporal variations in that head can be readily incorporated using superposition (convolution) procedures (e.g., Olsthoorn 2008) as shown in the Appendix. Although wellbore storage is ignored in this development because of the rapid (relative to the typical Δt used in practice) response of wells in aquifers of moderate to high K, the solution can be extended to incorporate wellbore storage following Furbish (1991) and Spane (2002). Similarly, the solution can be extended to incorporate the vadose zone following Weeks (1979) and others.

Application

The regression convolution approach was applied to data from three LRS wells (LWC2, LEA5, and LEC2—Figure 2a) and the site reference barometer (adjacent to LEC2). Winter 2003 to 2004 (henceforth, winter 2004) data were used because there was virtually no pumping then and well responses appeared to be representative of typical conditions observed in LRS High Plains aquifer wells (Figure 3). The winter 2004 BRFs (Figure 4) have three important characteristics. First, the agreement between the BRFs from the different wells is quite striking, despite the wells being separated by over 680 m, indicating that the character of the aquifer-aquitard system is not changing substantially between the wells. Second, the short-term (1 h) response is typical of what would be expected in a confined aquifer in which most of the load is borne by the pore water (BE $\approx$ 0.08), consistent with the near-surface, unconsolidated nature of the aquifer (e.g., Rasmussen and Mote 2007). Third, the longer-term (1 d) response is typical of what would be expected in a semi-confined (leaky) aquifer where water movement through the aquitard equilibrates heads, consistent with the results of a 4-d pumping test at the LRS in which drawdown stabilized as a result of leakage (Butler et al. 2004). The BRFs of wells screened at the

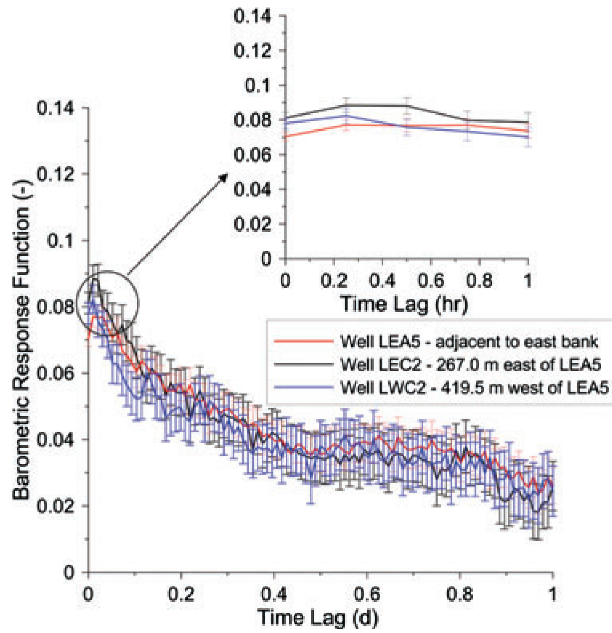

Figure 4. One-day and 1-h (inset) barometric response functions (BRFs) for three LRS wells in the High Plains aquifer in the winter of 2004. BRFs for winters 2005 and 2006 are similar in form. Agreement between the BRFs from these wells was observed in all years since monitoring began (2001 or 2002). Error bars indicate one standard error about the estimated functions; linear trend removed from data series prior to BRF calculation.

bottom of the unconfined aquifer (interval B of Figure 2b) were zero for this time period (winter 2004 curve of Figure 5), indicating that barometric-pressure changes propagated rapidly through the pores of the vadose zone and the unconfined aquifer. The rapid propagation across the unconfined aquifer (BRFs from LRS wells screened at the water table and those screened at the bottom of the unconfined aquifer always coincide) indicates that the apparent clay layers in the unconfined aquifer shown in the electrical conductivity log of Figure 2b are not laterally extensive enough to affect the hydraulic connection between the top and bottom of that aquifer for the temporal resolution ($\Delta t = 15$ min) of this analysis.

The BRFs presented in Figure 4 suggest the possibility of acquiring information about the aquitard from these functions. Theoretical response functions were computed using Equation 7 and fit to the field-determined BRFs to estimate the properties of the aquifer-aquitard system. Conditions at the top of the aquitard, which are required for the response function calculation, were obtained from wells screened at the bottom of the unconfined aquifer (interval B of Figure 2b). For winter 2004, the BRFs for those wells were essentially zero for all lags beyond the zero lag (winter 2004 curve of Figure 5). Using that condition at the aquitard top, an aquifer loading efficiency (1 – BE) of 0.92, and an estimate of aquifer diffusivity (2.9×10^6 m^2/d) based on previous estimates of aquifer K (88 m/d) and S_s (3.0×10^{-5}/m) obtained from the 4-d LRS pumping test (Butler et al. 2004), we fit a theoretical response function to the winter 2004 BRF for

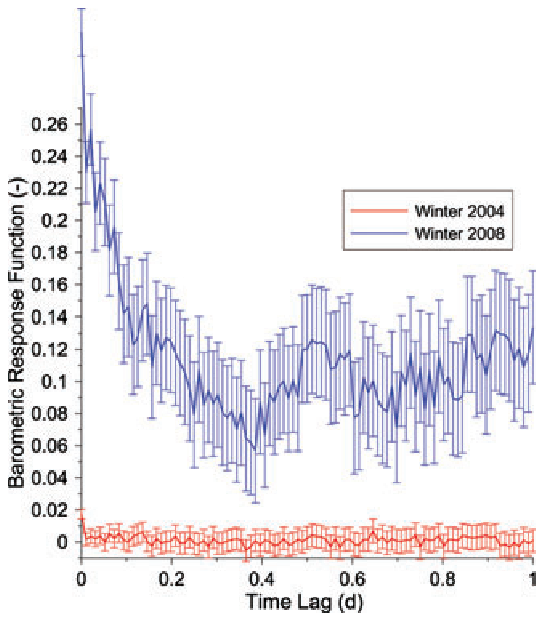

Figure 5. One-day barometric response functions for well LEA4 for winters 2004 and 2008, well LEA4 is adjacent to well LEA5 and screened across interval B of Figure 2b. Linear trend removed from the data series prior to BRF calculation.

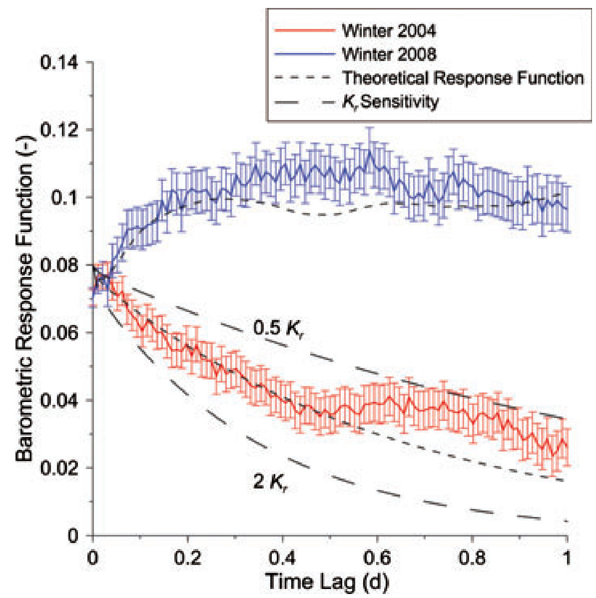

Figure 6. One-day barometric response functions for well LEA5 for winters 2004 and 2008, and the best-fit theoretical response function for the winter 2004 data. Hydraulic parameters from the winter 2004 fit were used to generate the winter 2008 theoretical response function. Estimated parameters were obtained from the first half-day of the 2004 BRF: $K_r = 1.8 \times 10^{-5}$ [−], $D_1 = 1.7 \times 10^2$ m^2/d, and $\gamma_1 = 0.97$ [−]. The aquifer loading efficiency (γ_2), the aquifer diffusivity (D_2), and the ratio of aquitard thickness to aquifer thickness were fixed at 0.92 [−], 2.9 × 10^6 m^2/d, and 1.0, respectively. Similar results were obtained for the other High Plains aquifer wells. The sensitivity of the response functions to K_r is shown for variations of a factor of two about the 2004 theoretical response function; similar variations in D_1 produced plots that were barely distinguishable from the response function, indicating the much smaller sensitivity to that parameter for these conditions (i.e., aquifer and aquitard characteristics). Linear trend removed from the data series prior to BRF calculation. Span of y-axis is half that of Figure 5.

well LEA5 (Figure 6). The fit, which was based on the first half-day of the BRF because additional mechanisms appear to be affecting the BRF at larger times, yielded estimates of the aquitard diffusivity ($D_1 = 1.7 \times 10^2$ m^2/d), the aquitard loading efficiency ($\gamma_1 = 0.97$), and the ratio between the aquitard and aquifer hydraulic conductivity ($K_r = 1.8 \times 10^{-5}$). Using the aquifer K estimate from the LRS pumping test, an aquitard K of 1.6×10^{-3} m/d, which is within 25% of the pumping-test value (2.1×10^{-3} m/d), is calculated from the K_r estimate, demonstrating the similarity of the K_r ratios obtained with the different approaches. Note that an estimate of aquifer diffusivity is required for the parameter estimation procedure because of the high degree of correlation between K_r and D_2. In the absence of the pumping-test information that was available at the LRS, the aquifer diffusivity could be estimated using the aquifer K from a slug test and the aquifer S_s determined from the BE. In this example, the target for comparison was the aquitard K from the LRS pumping test, so the aquifer K and S_s values from that same pumping test were used for the diffusivity estimate. Isotropy in aquifer hydraulic conductivity was assumed for the calculation of K_1. This assumption should be reasonable in unconsolidated aquifers with a hydrostratigraphic framework similar to that at the LRS (Figure 2b).

A check on the parameters calculated from the BRF fit was performed using the same parameters to generate a theoretical response function to compare with the field-determined BRF for winter 2008; this time was chosen because it followed an extended period of recharge (Figure 3). Although intuitively one might expect the BRF to be a characteristic function of a well, our results show otherwise. The BRF for well LEA5 in winter 2008 (Figure 6) is distinctly different from the 2004 BRF because of differing conditions at the top of the aquitard (bottom of the unconfined aquifer). In winter 2008, the wells screened at the bottom of the unconfined aquifer display a barometric response (e.g., winter 2008 BRF in Figure 5), indicating that the propagation of air pressure through the vadose zone was affected by a change of conditions in that zone (e.g., perched water table or layer of frozen soil). This difference in barometric responses at the bottom of the unconfined aquifer between 2004 and 2008 is analogous to the difference reported by Hare and Morse (1997) between a well below a low permeability landfill cap and one adjacent to the cap. Moreover, the 2008 response in the unconfined aquifer cannot be matched with the 1-D, uniform vadose-zone solution of Weeks (1979), indicating that the response must be a product of complicated flow paths or other phenomena. Using the parameters determined from the 2004 fit and the upper boundary condition based on the 2008 data (i.e., the winter 2008 BRF of Figure 5), we generated the 2008 theoretical response function for well

LEA5 shown in Figure 6. The agreement with the 2008 field-determined BRF is quite good, although no curve fitting was involved, and can be considered a strong verification of the parameters calculated from the 2004 analysis.

Discussion and Conclusions

This work demonstrates the utility of the BRF for gaining insights into site hydrostratigraphy. In semi-confined aquifers, the form of the BRF indicates the degree of aquifer confinement and it can be exploited to estimate aquitard properties using the approach developed here. A comparison of BRFs between wells can shed light on aquitard continuity. However, the generality of the conclusions that can be drawn from this comparison depends on the BRF averaging (support) volume, which is the subject of ongoing work. In unconfined aquifers, the similarity of BRFs from the wells screened at the water table and those at the base of the aquifer indicate that low-K layers within the aquifer are not laterally extensive enough to affect the hydraulic connection across the aquifer for the temporal resolution of this analysis. Differences between such BRFs could potentially be exploited to estimate the vertical K of an unconfined aquifer in a manner analogous to the frequency-domain approach of Rojstaczer and Riley (1990).

This work appears to be the first to show that BRFs are not necessarily an invariant characteristic of a well. The form of a BRF can depend on the nature of the pressure propagation through the vadose zone, even for wells in a semi-confined aquifer (Figure 6). This dependence on the vadose zone presents the opportunity to glean insights into changes in vadose-zone conditions, a subject of ongoing work. Spane (2002) and others have speculated that the barometric response of wells in unconfined aquifers could vary as a function of vadose-zone conditions. This work confirms that speculation and demonstrates that a similar dependence is found in semi-confined aquifers (Figures 5–6).

The BRF is a promising tool for gaining important insights through monitoring of water levels and barometric pressure. In this study, we demonstrated that BRFs can provide reasonable estimates of aquitard properties as well as valuable information about other aspects of site hydrostratigraphy. Thus, they can often be a cost-effective alternative/supplement to a conventional pumping test. Similarly, BRFs should be a useful tool for initial screening of potential shallow target zones for CO_2 sequestration and, more generally, for monitoring changes in formation and fluid properties (e.g., porosity and fluid compressibility) in the course of sequestration activities. Although we demonstrated the approach in a system in which the strong seasonality of pumping facilitated data processing, it is also applicable in aquifers that are pumped more continuously, although more involved processing is required to remove the impact of other mechanisms. Finally, we must emphasize that the approach for estimation of aquitard properties described here is best implemented with a well in the overlying aquifer. In the absence of such a well, considerable error may be introduced into the parameter estimates through uncertainty about head conditions at the top of the aquitard. This uncertainty can be particularly large at sites with thick vadose zones.

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Appendix

Solution Derivation

The Laplace-space expressions for the governing Equations 4a and 4b are

$$p\overline{h}_1 - \gamma_1 h_0 = D_1 \frac{d^2\overline{h}_1}{dz^2} \quad \text{(A1a)}$$

$$p\overline{h}_2 - \gamma_2 h_0 = D_2 \frac{d^2\overline{h}_2}{dz^2} \quad \text{(A1b)}$$

with the notation defined in the main text following Equation 5.

The Laplace-space expressions for the boundary conditions are

$$\overline{h}_1(-l, p) = \frac{h_{UB}}{p} \text{ for the constant-head condition } (h_{UB}) \text{ at the aquitard top} \quad \text{(A2)}$$

$$\frac{d\overline{h}_2(a, p)}{dz} = 0 \text{ for the no-flow condition at the bottom of the aquifer} \quad \text{(A3)}$$

and

$$\overline{h}_1(0, p) = \overline{h}_2(0, p) \quad \text{(A4)}$$

and

$$K_1 \frac{d\overline{h}_1(0, p)}{dz} = K_2 \frac{d\overline{h}_2(0, p)}{dz} \quad \text{(A5)}$$

for continuity of head and flow, respectively, at the aquifer-aquitard interface.

The general solution to Equation A1 is

$$\overline{h}_1(z, p) = A_1(p)\cosh\left(\sqrt{\frac{p}{D_1}}z\right) + B_1(p)\sinh\left(\sqrt{\frac{p}{D_1}}z\right) + \frac{\gamma_1 h_0}{p} \quad \text{(A6a)}$$

$$\overline{h}_2(z, p) = A_2(p)\cosh\left(\sqrt{\frac{p}{D_2}}z\right) + B_2(p)\sinh\left(\sqrt{\frac{p}{D_2}}z\right) + \frac{\gamma_2 h_0}{p} \quad \text{(A6b)}$$

where A_i and B_i are functions of p that are determined from the boundary conditions.

Using the boundary conditions of A2 to A5 and the D_i and K_r notation defined after Equation 5, expressions for A_1, A_2, B_1, and B_2 can be written as follows:

$$A_1(p) = \frac{h_0}{p}\,\frac{\left(\frac{h_{UB}}{h_0} - \gamma_1\right) K_r\sqrt{\frac{D_2}{D_1}}\operatorname{sech}\left(\sqrt{\frac{p}{D_1}}l\right) - (\gamma_1 - \gamma_2)\tanh\left(\sqrt{\frac{p}{D_1}}l\right)\tanh\left(\sqrt{\frac{p}{D_2}}a\right)}{K_r\sqrt{\frac{D_2}{D_1}} + \tanh\left(\sqrt{\frac{p}{D_1}}l\right)\tanh\left(\sqrt{\frac{p}{D_2}}a\right)} \quad \text{(A7)}$$

$$A_2(p) = \frac{h_0}{p}\,\frac{\left(\frac{h_{UB}}{h_0} - \gamma_1\right)\operatorname{sech}\left(\sqrt{\frac{p}{D_1}}l\right) + (\gamma_1 - \gamma_2)}{1 + \frac{1}{K_r}\sqrt{\frac{D_1}{D_2}}\tanh\left(\sqrt{\frac{p}{D_1}}l\right)\tanh\left(\sqrt{\frac{p}{D_2}}a\right)} \quad \text{(A8)}$$

$$B_1(p) = -\frac{A_2(p)}{K_r}\tanh\left(\sqrt{\frac{p}{D_2}}a\right)\sqrt{\frac{D_1}{D_2}} \quad \text{(A9)}$$

$$B_2(p) = -A_2(p)\tanh\left(\sqrt{\frac{p}{D_2}}a\right). \quad \text{(A10)}$$

Substituting Equations A7 to A10 into Equation A6 and simplifying yields the Laplace-space solution of Equation 5a in the main text.

Convolution Expression

Temporal variations in the head at the top of the aquitard (h_{UB}) can be incorporated using a standard convolution approach (e.g., Olsthoorn, 2008). In order to demonstrate the approach for the head in the semi-confined aquifer (h_2), Equation 5c can be rewritten as:

$$f_2(z,p) = F(z,p) + \frac{h_{UB}}{h_0} G(z,p) \qquad \text{(A11a)}$$

where

$$F(z,p) = -\frac{\gamma_1 \operatorname{sech}\left(\sqrt{\frac{p}{D_1}}l\right) + (\gamma_2 - \gamma_1)}{1 + \frac{1}{K_r}\sqrt{\frac{D_1}{D_2}} \tanh\left(\sqrt{\frac{p}{D_1}}l\right) \tanh\left(\sqrt{\frac{p}{D_2}}a\right)} \times \frac{\cosh\left[\sqrt{\frac{p}{D_2}}(z-a)\right]}{\cosh\left(\sqrt{\frac{p}{D_2}}a\right)} + \gamma_2 \qquad \text{(A11b)}$$

and

$$G(z,p) = \frac{\operatorname{sech}\left(\sqrt{\frac{p}{D_1}}l\right)}{1 + \frac{1}{K_r}\sqrt{\frac{D_1}{D_2}} \tanh\left(\sqrt{\frac{p}{D_1}}l\right) \tanh\left(\sqrt{\frac{p}{D_2}}a\right)} \times \frac{\cosh\left[\sqrt{\frac{p}{D_2}}(z-a)\right]}{\cosh\left(\sqrt{\frac{p}{D_2}}a\right)} \qquad \text{(A11c)}$$

The F function quantifies the dissipation of the pressure in the aquitard-aquifer system produced by the barometric surface loading, while the G term quantifies the head change produced by the boundary condition at the aquitard top. Only the G term is involved in the convolution.

The series expression for the convolution in real space is

$$A(z,t) \approx A(z,k\Delta t) = 1 - \left[\sum_{n=1}^{N} \frac{V_n}{n} F\left(z, \frac{\ln 2}{k\Delta t}n\right) + \frac{h_{UB}(0)}{h_0} \sum_{n=1}^{N} \frac{V_n}{n} G\left(z, \frac{\ln 2}{k\Delta t}n\right) + \sum_{i=1}^{k-1} \frac{\Delta h_{UBi}}{h_0} \sum_{n=1}^{N} \frac{V_n}{n} G\left(z, \frac{\ln 2}{(k-i)\Delta t}n\right) \right] \qquad \text{(A12)}$$

where $t = k\Delta t$, $h_{UB}(0)$ is the head at the top of the aquitard at $t = 0$,

$$\Delta h_{UBi} = h_{UB}(i\Delta t) - h_{UB}(i\Delta t - \Delta t)$$

is the change in head at the aquitard top over one time interval Δt.