ESTIMATING THE HYDRAULIC PARAMETERS OF THE ARBUCKLE-SIMPSON AQUIFER BY ANALYSIS OF NATURALLY-INDUCED STRESSES

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IN THE NAME OF ALLAH, THE COMPASSIONATE, THE MERCIFUL

We made from water every living thing. Will they not then believe?

(the Holy Qur'ān, 21:30)

Dedication

To the Memory of my parents;

and

to my wife Thuria, my son Mohammed, my daughter Russel, my daughter Roaa, my daughter Rukia, and my son Hasanian.

ACKNOWLEDGMENTS

First and foremost praise be to Allah, Lord of the Universe.

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CHAPTER I

INTODUCTION

GENERAL

Groundwater studies often involve the use of either an analytical or numerical approach to solve a particular problem. For the last several decades, groundwater numerical models have increasingly proved their value in analyzing and evaluating groundwater systems. Todd and Mays (2006) suggested that last two decades have resulted in tremendous changes in the employment of computers for groundwater management.

Almost all groundwater flow and transport models are designed to solve the relevant partial differential equation either analytically or using numerical techniques. The two-dimensional groundwater flow partial differential equation in Cartesian coordinates is:

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} = \frac{S}{T}\frac{\partial h}{\partial t} \qquad (1.1)$$

where

h is the hydraulic head (L)

S is storage coefficient (dimensionless)

T is transmissivity (L^2/T) and t is time (T)

The equivalent form of Equation 1.1 in plane polar coordinates is (Jacob, 1950; Todd and Mays, 2005)

$$\frac{\partial^2 s}{\partial r^2} + \frac{1}{r}\frac{\partial s}{\partial r} = \frac{S}{T}\frac{\partial s}{\partial t} \tag{1.2}$$

where

s is the drawdown at the well (L) in response to a discharge Q (L^3/T)

Analytical solutions for Equation 1.2 are available for isotropic, homogeneous and extensive confined aquifers (Theis, 1935; Hantush, 1964).

Most natural aquifers are anisotropic and heterogeneous in. For anisotropic heterogeneous aquifers, Equation 1.1 must be approximated using numerical techniques such as finite difference or finite element. The numerical solution is accomplished by replacing the flow domain of an area by a discretized model domain consisting of cells, blocks or elements depending on the technique that is being employed. In order to derive the finite difference form that is equivalent to Equation 1.1, the equation is rewritten as:

$$\frac{\partial}{\partial x}\left(T_x \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(T_y \frac{\partial h}{\partial y}\right) = S\frac{\partial h}{\partial t} \tag{1.3}$$

where

T_x and T_y are transmissivities in the x and y directions respectively. Todd and Mays (2005) gave the following finite difference form for Equation 1.3 for the cell (i, j) in Figure (1.1):

$$
\begin{aligned}
&-T_{x_{i-1,j}}\Delta y_j\left(\frac{h_{i-1,j,t}-h_{i,j,t}}{\Delta x_i}\right)+T_{x_{i,j}}\Delta y_j\left(\frac{h_{i,j,t}-h_{i+1,j,t}}{\Delta x_i}\right)\\
&-T_{y_{i,j+1}}\Delta x_i\left(\frac{h_{i,j+1,t}-h_{i,j,t}}{\Delta y_j}\right)+T_{y_{i,j}}\Delta x_i\left(\frac{h_{i,j,t}-h_{i,j-1,t}}{\Delta y_j}\right)\\
&=S_{i,j}\Delta x_i\Delta y_j\left(\frac{h_{i,j,t}-h_{i,j,t-1}}{\Delta t}\right)
\end{aligned}
\tag{1.4}
$$

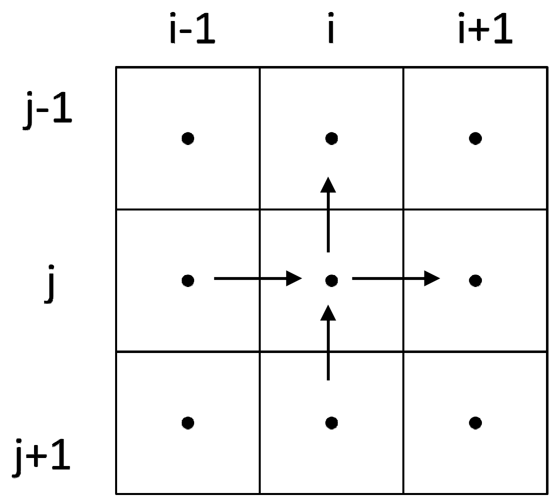

Figure 1.1. Finite difference grid representations.
The i and j are the index variables. The arrows indicate flow direction.

The numerical model as depicted by Equation 1.4 and Figure 1.1 requires, the assignment of a discrete value of the hydraulic parameters to each cell or block in the model (flow) domain. These discrete values are designated as the model hydraulic parameters. For a groundwater model to produce reasonable predictions, the model parameters should accurately represent the physical aquifer system. Knowledge of representative aquifer parameters is a necessity for accurate model calibration. If the calibration process were based on misrepresentative parameters, model prediction may be erroneous.

Aquifer hydraulic parameters needed to characterize the flow domain and are of concern to this research are transmissivity (T), storage coefficient (S), and porosity (η). These parameters are determined by field or laboratory methods, such us pumping and laboratory core tests. Pumping tests may be problematic especially when the aquifer water quality or the aquifer pollution is an issue (Mehnert, 1998). Pumping tests cannot be conducted when the aquifer system is undergoing a period of background monitoring, during which time human interference is prohibited (Ritzi, 1989). Furthermore, pumping tests are costly and have limited spatial extent, i.e. the results are site specific. Laboratory methods are of limited usefulness and hard to generalize since they represent disturbed core samples of small size.

An alternative option to determine the aquifer hydraulic parameters is warranted. Confine aquifers are subjected to natural stresses that produce measurable water-level fluctuations within a well. The ocean tides and atmospheric pressure changes produce corresponding fluctuations in water level within wells penetrating coastal aquifer. Inland aquifers are stressed by the solid earth tides as well as atmospheric-pressure changes. Analyzing natural stresses-induced water-level fluctuations for the purpose of parameter estimations could represent a significant cost reduction over pumping tests and other field methods. The purpose of this research was to analyze the naturally induced stresses on the aquifer surfaces along with the resulted water-level fluctuations to classify the aquifer and determine its hydraulic parameters. An integral part of the research is the introduction of an improved method to determine the barometric efficiency of confined aquifers.

The research was part of the hydrological study for the Arbuckle-Simpson aquifer coordinated by the Oklahoma Water Resources Board (OWRB) from 2003 to 2009. OWRB (2003) stated that

> *State and federal experts agree that the information garnered from previous studies of the Arbuckle-Simpson aquifer—concentrating primarily on its geology and hydrology at or near the surface—is inadequate to address the aquifer's complex geology and management issues confronting the local users. Investigation of the deeper part of the aquifer (greater than 1000 feet) is needed to understand the full extent of the fresh-water zone and the volume of water in storage in the aquifer. In addition, no sufficient information exists to predict the response of springs and streams to groundwater withdrawals. Critical to future study of the aquifer is an understanding of the formations "plumbing system" that controls the interactions between groundwater levels and spring flow."*

The goal of the hydrological study was to develop a management plan for the Arbuckle-Simpson to ensure the sustainability of the water resources of the aquifer and to preserve the environment and the ecosystem of the area.

NATURAL STRESSES

Aquifers are subjected to stresses from natural processes, such as mechanical forcing of the aquifer by ocean and earth tides and/or atmospheric pressure load (Todd, 1959; Walton, 1970; Freeze and Cherry, 1979; Domenico and Schwartz, 1990). Earth and ocean tides are the product of lunar and solar tidal (gravitational) forces. Changes in barometric pressure are induced by variations in temperature and circulation. Water levels in monitoring wells often reflect these stresses. Earth-tide induced water-level fluctuations generally have smaller amplitude compared to fluctuations caused by barometric-pressure fluctuations (Davis and De Wiest, 1966). The effect of earth tides can be observed in wells tapping confined aquifers (Bredehoeft, 1967) and in wells tapping deep, relatively stiff and low-porosity unconfined aquifers (Weeks, 1979: Rojstaczer and Agnew, 1989).

Earth tides force dilatation or compression of aquifer materials. A dilatation or compression cause proportional changes in the aquifer formation stress, which is balanced by an increase or decrease in pore fluid pressure. As the pore fluid pressure changes, a difference in pressure is introduced between the well bore storage and the aquifer, which results in flow into or out of the well (Lambert, 1940).

Barometric effects on aquifers result from stresses acting on the aquifer due to changes in atmospheric pressure. These changes are linked to periodic (diurnal and semidiurnal) and aperiodic atmospheric changes. The periodic changes are the result of atmospheric thermodynamic effects and the aperiodic are the result of long-term movements of air masses of low or high pressures. Atmospheric-pressure fluctuations reveal two lows at early morning and early afternoon and two highs at late morning and late afternoon. Clark (1967) suggested that atmospheric pressure lows occur at 4 am and pm and atmospheric highs occur at 10 am and pm. Merritt (2004) indicated that the timing of lows or highs is variable between 2-3 hours from day to day and it is difficult to decide the precise timing of these events.

High atmospheric pressure and high tides both lower the water level in wells. Water-level minima coincide with barometric maxima and high earth tides. High earth tides coincide with moon transit at which time the gravity pull on the earth crust is the highest, hence the crust is dilated and water level in wells is lowered.

Research Questions and the Organization of the Dissertation

This research evaluated the use of tidal analysis methods for the study of a complex thick carbonate aquifer. The emphasis of the research was on analyzing an aquifer-well system in which stress oscillations from tidal and atmospheric sources caused macroscopic water movement in and out of the well. The study area for the research was the Arbuckle-Simpson aquifer of south-central Oklahoma.

Water-level fluctuations may be analyzed for the estimations of specific storage, transmissivity, and porosity. The analyses are applicable to aquifers which generate sufficient potentiometric signals, usually confined aquifers. Previous researchers (Bredehoeft, 1967; Marine, 1975; Narasimhan and others, 1984; Hsieh and others, 1987; Merritt, 2004) have indicated several problems with earth-tide theory including: the compressibility of the solid parts of the aquifer, the type of aquifer being studied, and the interference between earth tides and atmospheric pressure changes.

An evaluation of this literature provided insight into areas of improvement for the theory that were possible. The first question that arose was whether tidal analysis could be used to evaluate the hydraulic response of an aquifer to define aquifer type using the response of the aquifer to the two major types of stresses, earth tides and atmospheric pressure changes. The second question resulted from determining that the Clark method (1967) performed poorly for determining *BE* for the study site. The *BE* is the ratio of the aquifer pressure head change to the atmospheric pressure change and was introduced by Jacob (1940). Clark (1967) presented a method to determine *BE* based on aperiodic long-term atmospheric-pressure changes that result from movement of air masses. An improved method of estimating *BE* was developed as part of this research. Finally, the third question evaluated the specific storage and porosity for the Arbuckle-Simpson aquifer.

Specific storage and porosity ultimately were the parameters evaluated as part of this research. Models to estimate transmissivity were surveyed and discussed but not used to determine the Arbuckle-Simpson aquifer transmissivity, due to their limited applicability. These models were applicable for aquifers with low transmissivities (less than 500 ft^2/day), which excludes the Arbuckle-Simpson aquifer which has much higher transmissivities.

This dissertation included eight chapters. Chapter one was the introduction. Chapters two through four provided background materials where Chapter Two evaluated previous studies, Chapter Three described governing equations, and Chapter Four described the study area. Chapters five, six, and seven were formatted as individual publications to evaluate the three questions described above. Chapter Eight provided a summary and conclusions of the dissertation.

CHAPTER II

PREVIOUS STUDIES

Analyzing natural-stress-induced water-level fluctuations for an aquifer characterization started as early as the beginning of the twentieth century. However, the phenomenon of water level fluctuations due to earth tides was observed and recorded in earlier times. Serious attempts have been initiated by the 1930s to monitor and analyze the natural stresses' effects. A review of the earth tide and barometric-induced water-level fluctuations theory and applications as well as brief introduction of the study area, the Arbuckle-Simpson aquifer is presented here.

NATURALLY-INDUCED GROUNDWATER-LEVEL FLUCTUATIONS: GENERAL

Wells are known to respond to earth tides and changes in atmospheric pressure. Bredehoeft (1967) cited Klonne (1880) who reported water-level fluctuations (of tidal nature) in a flooded coal mine near Duchov, the old Czechoslovakia. Grablovitz (1880), also cited in Bredehoeft (1967) attributed Klonne's (1880) observations to the dilatation produced by earth tides. Young (1913) reported periodical (of a period of approximately 12.5 h) water-level fluctuations in a well near Cradock, South Africa. Young's (1913) observations lasted for two two-week periods on May and June of 1905.

It appears that Young (1913) was the first to monitor periodical water-level fluctuations and accurately interpreted these fluctuations as the work of solid earth tides. Young (1913) concluded that “results seem to establish beyond question that the fluctuations in these wells are to be attributed directly or indirectly to extra-terrestrial causes.”

Robinson (1939) published several hydrographs of wells in New Mexico and Iowa which reflect the influence of earth tides on water level fluctuations. The author described the earth tide phenomena as including the following general components:

1) Two daily cycles of fluctuations where the average daily retardation of cycles agrees closely with that of the moon transit;
2) The daily troughs of the water level coincide with the transit of the moon at the upper and lower culminations;
3) Periods of large regular fluctuations coincide with periods of new and full moon, whereas periods of small irregular fluctuations coincide with periods of first and third quarters.

Theis (1939) working with Robinson’s data and other data from Carlsbad, New Mexico, recognized that the water-level fluctuations could only attributed to the dilation accompanies the tidal bulge. Jacob (1940) demonstrated how barometric and tidal effects can be used to determine the storage coefficient and porosity of an aquifer. He introduced the term “barometric efficiency” as an index of the elasticity of an aquifer system. Jacob (1940), also, described the mechanics of the ocean tidal fluctuations and introduced the term “tidal efficiency “ which is the ratio of the change of water level in a well to a change in tide stage. An “amplitude factor” was presented by Jacob (1950) and Ferris (1951) to describe the change of formation pressure caused by a spatially distributed change of pressure at land surface. The amplitude factor of Jacob's (1950) replaced the "tidal efficiency" that was introduced by Jacob (1940). Richardson (1956) reported water level fluctuations resulted from earth tides effect in a well at Oak Ridge, Tennessee. The well penetrates a water table aquifer.

Melchior (1960) performed harmonic analyses of tidal fluctuations reported by other investigators. He analyzed Robinson data from Iowa, Theis data from New Mexico, Richardson data from Tennessee, along with data from the old Czechoslovakia, Belgium, And the Congo. The author indicated that the comparison of the amplitudes of the major waves showed "reasonable agreement" with amplitudes predicted from the equilibrium tidal theory. The harmonic analyses of Melchior (1960) concluded that the water level fluctuations in these wells were linked to dilation produced by earth tides. Harmonic analyses of the earth tidal potential reveal that the potential includes a great number of harmonic tidal components. However, Melchior (1964) stated that, "only five of these components are of real importance for groundwater fluctuation". They constitute approximately 95 percent of the tidal potential. These five are: M2 , a lunar wave with a period of 12h 25m 14s; S2, a solar wave with a period of 12h 00m, N2, a lunar wave with a period of 12h 39m 30s; K1, a luni-solar wave with a period of 23h 56m 4s; and O1, a lunar wave with a period of 25h 49m 10s. Young (1913) harmonic analyses, also, identified five tidal components. However, Young's analyses did not reveal N2, instead revealed P1; a solar diurnal wave with a period of 24h 4 m.

A relationship between the motion of water within an open well bore (taking into account the storage of water within the well bore) and pressure-head oscillations in the confined aquifer was developed by Cooper et al. (1965). The resulting equation established a relationship between the amplitude and phase lag of oscillations of the water level in the well to the amplitude of oscillations of pressure-head in the aquifer. The amplitude ratio between the water level fluctuation within the well and the pressure head fluctuation within the aquifer is termed the amplitude response. Cooper et al. (1965) showed that the amplitude ratio is a function of the aquifer's hydraulic properties (the transmissivity and storage coefficient), the radius of the well casing, the period of the forcing pressure, and the inertial effects of the water in the well. Their analysis was developed for seismic disturbances within the aquifer. It should be mentioned that the earth tide and atmospheric-pressure fluctuations are considered exterior stresses

Gregg (1966) developed a modification of Jacob's tidal efficiency formula to compute the tidal efficiency adjusted for atmospheric pressure change. Gregg's study did not consider the earth tide effects on water level fluctuations. He stated that "Water-level fluctuations caused by earth tides are 180 degrees out of phase with those caused by ocean tides." Hence, any earth tide-caused fluctuations in the wells would be masked by the ocean tide caused fluctuations. An important finding of Gregg study was that the tidal efficiency decreased with depth when measured in the same location. This result suggests an increase in the *BE*, since these efficiencies add up to one (Jacob, 1940; Domenico and Schwartz, 1990).

Todd (1959) suggests that *BE* may be used as a measure of capability (competence) of the overlying confining layer to resist pressure changes. High *BE* values are associated with thick confined layers. On the other hand, tidal efficiency (TE), as was defined by Jacob (1950), is a measure of incompetence of the confining layer to resist pressure changes. Thicker layers are associated with small TE-values. Clark (1967) devised a method to estimate the *BE* based on aperiodic, long-term pressure variation rising from the movement of air masses and the corresponding measured head changes in the well. Clark found that the atmospheric pressure had a period of about 12 hour, being high at 10 a. m. and p. m., and low at about 4 a. m. and p. m. Davis and Rasmussen (1993) used linear regression technique to determine the *BE* and compared their approach with the Clark's method. The authors stated that Clark's method provides an unbiased consistent estimate of the *BE* when negative and positive changes in barometric pressure are equally likely. Davis and Rasmussen (1993) indicated that this conclusion holds for linear and nonlinear trend that may present within the atmospheric pressure data. However, when unequal numbers of positive and negative changes are present in the data and the trend is linear, Davis and Rasmussen (1979) suggested the use of iterative recursive technique to correct the estimated value of the *BE* as calculated by the Clark's method. Several researchers have cited problems with the Clark method to determine *BE* (Hsieh et al. 1987; Merritt, 2004). Marine (1975), obtained values of porosity of higher than 100 percent and he attributed these erroneous values to overestimated *BE* values. The author utilized The Clark method to determine *BE.*

Gonthier (2007) presented a graphical method to estimate *BE* that would minimize the influence of non-barometric pressure-induced water level changes. Rhoads and Robison (1979) employed graphical method and two arbitrary numerical approaches to determine the *BE* for three observation wells in Montgomery County, Virginia. Gonthier (2007) presented a graphical method to estimate *BE*. Gonthier (2007) suggested a long period of atmospheric-pressure and water-level changes monitoring (at least 60 days) to minimize the influence of barometric-pressure-independent water-level changes on the resulted *BE*.

NATURAL STRESSES-INDUCED GROUNDWATER-LEVEL FLUCTUATIONS: DEVELOPMENT OF THE THEORY

The first theoretical treatment of the response of a well to sinusoidal forcing function of within a confined aquifer was derived by Cooper et al (1965). Bredehoeft (1967) followed the analyses of Cooper et al. (1965) and developed a theory for the response of the water level in the well to earth tides. Cooper‘s analysis was developed for seismic disturbances (which is interior perturbation), but Bredehoeft applied it to earth tides (which are exterior forcing). Bredehoeft (1967) mentioned that the seismic forcing analysis can be applied equally well to the earth-tides forcing.

Bredehoeft (1967) showed that the inertial effects were negligible when the transmissivity of the aquifer was above 1 cm^2/s (0.001 ft^2/s or approximately 600 gpd/ft). The author considered the solid grains are incompressible so that volume changes in the formations, due to the effect of the earth tides, are assumed equal to changes in the pore volume. Bredehoeft (1967) presented a method for determining the specific storage of a confined aquifer if the Poisson’s ratio is known. He gave two possible approaches for analyzing observed water level fluctuations caused by earth tides, namely, (1) to compare the fluctuation in the well with fluctuation that one would expect from tidal theory or (2) to compare the amplitude of the various tidal components obtained by harmonic analysis of the hydrograph with the theoretical amplitude of the particular waves. Bredehoeft (1967) concluded that analyses of water level fluctuations caused by the earth tide within a well tapping a confined aquifer can be used to compute the specific storage and the porosity of the aquifer.

Robinson and Bell (1971) analyzed tidal fluctuation in wells and developed methods of obtaining aquifer parameters from these fluctuations. The authors conclude that tidal fluctuations in wells can be explained in terms of aquifer dilation caused by earth tides, ocean tides and barometric tides because reasonable values for aquifer parameters were obtained. However, they indicated that accurate calculation of aquifer parameters by analysis of tidal fluctuation in wells is difficult because independent knowledge of the values of several terms in the equations is usually lacking.

Bear (1972) indicated that barometric-pressure fluctuations induced water-level fluctuations are observed in wells tapping confined aquifers and are not observed in wells tapping unconfined aquifer. Bear (1972) explained that changes in barometric pressure are transmitted directly to the entire water table in the aquifer and at the same time to the water table in the well tapping this aquifer. Weeks (1979) investigated the affects of barometric-pressure fluctuations on the on water-level fluctuation within a well penetrating an unconfined aquifer. Weeks (1979) concluded that water levels in a well tapping an unconfined aquifer are influenced by changes in barometric pressure but the mechanism is “substantially different” from that of the confined aquifer. The author stated that “because of the pneumatic diffusivity of the unsaturated zone, pressure response in the soil gas at the water table lags that at land surface, whereas atmospheric pressure changes are transmitted instantaneously down the well pore”. Weeks (1979) attributed water-level changes in a well tapping an unconfined aquifer to the difference between the atmospheric pressure exerted on the water level within the well and the soil gas pressure exerted on water table elsewhere within the aquifer.

Maréchal et al. (2002) studied water-level fluctuation in a well tapping unconfined hard rock aquifer. The authors found strong correlations between synthetic earth tides and water-level observations signals. Maréchal et al. (2002) indicated that such findings indicate that water-level fluctuations were due to the influence of earth tides on the apparently unconfined aquifer which imply that the aquifer is of low porosity. Bredehoeft (1967) concluded that an unconfined aquifer will not show significant tidal response the aquifer is thick or it has low porosity.

Several researchers have found some variation of the earth tide and barometric fluctuations effects with depth. Gregg (1966) concluded that tidal efficiency decreases with depth. The author stated that “the decrease in tidal efficiency with depth is probably due to the heterogeneity of materials and the greater abundance of hard dense beds with depth resulting in a dissipation of energy with depth.” Gregg’s (1966) finding agrees with the TE characterization given by Todd (1959) as mentioned earlier. Melchior (1964) as cited by Bredehoeft (1967), found that the amplitude of the M2 (a lunar semidiurnal wave) fluctuations increase with depth. Bredehoeft (1967) suggested two causes may contribute to the amplitude increase: the decrease of porosity of the geologic formations and the decrease of the permeability of the confining layers with depth. Bredehoeft indicated that deeper confined aquifers more closely approximate ideal artesian conditions.

Marine (1975) compared crystalline rock aquifer parameters estimated from earth tides analysis with the results of pumping tests. He found that the specific storage calculated from earth tides, using Bredehoeft's (1967) model, was more than an order of magnitude higher than the specific storage determined from pumping tests. Marine (1975) also calculated porosity using the same model and found that the computed porosity of "this slightly fractured crystalline aquifer...would (reach) 100 percent, an absurd value." The author concluded that "the porosity is very sensitive to the *BE*, which is extremely difficult to calculate for wells whose predominant water level fluctuations are caused by earth tides." Bredehoeft (1967) points out that "the porosity would represent an average value of a large volume in the vicinity of the well, a quantity which interests hydrologists and which is difficult, if not impossible, to determine by other means." Marine (1975) indicated that the Bredehoeft (1967) statement is especially applicable for fractured rock where the overall porosity is difficult to estimate. Marine (1975) suggested that porosity computed by Bredehoeft model is very sensitive to both the specific storage and the *BE*. Marine's calculations revealed that a *BE* of 50 percent resulted in 100 percent porosity for a sandy aquifer. In his final remarks, Marine (1975) agreed with Robinson and Bell's (1971) conclusion that accurate calculation of aquifer parameters by analyzing tidal effect is difficult due to the lack of independent knowledge of several terms in the equations. Van der Kamp and Gale (1983) suggested that the high value of porosity obtained by Marine (1975) "may be due to the neglect (by Bredehoeft, 1967) of the compressibility of solids."

Van der Kamp and Gale (1983) presented an equation relating specific storage of an aquifer to earth-tide strain that includes the effect of compressibility of the solids' part of the porous media. The authors define the ratio of the change in pore pressure to the change in pressure loading at land surface as the "loading efficiency." It should be recalled here that Jacob (1940), considered the case of ocean-tide loading at the surface, referred to the change in pressure head within the formation, to the tidal stage change at the surface as "tidal efficiency."

The response to earth tides of an open well penetrating a confined aquifer was studied by Narasimhan and others (1984), who recognized the importance of well bore storage effects, the period of the tidal pulses, and the aquifer properties permeability and specific storage. They applied a numerical model of saturated flow to demonstrate the qualitative importance of these factors. Narasimhan et al. (1984) criticizes the Bredehoeft (1967) analysis and suggested that the analysis is "internally inconsistent". Their point was that the specific storage which is by definition a parameter defined only for drained condition cannot be determined form undrained response of the aquifer such as the aquifer response to earth tides.

Hsieh et al. (1988) discussed the questions raised by Narasimhan et al. (1984) regarding the analysis of Bredehoeft (1967) and showed that it is possible to directly determine specific storage from an undrained loading test. They proceeded to conclude "thus it is not unreasonable that one can determine the specific storage from earth tide response." The authors, also, showed that Van der Kamp and Gale's (1983) result reduce to Bredehoeft's result when the grains are assumed incompressible.

Ignoring inertial effects of water stored in the well bore, Hsieh and others (1987) determined the time-varying drawdown in the open well as a function of periodic pressure head oscillations in the aquifer. They derived an analytical solution expressing the phase shift between the tidal dilatation of the aquifer and the water level response in the well as a function of the aquifer transmissivity, storage coefficient, well radius, and the period of harmonic disturbance. Hsieh et al (1987) adapted a graphical procedure to estimate transmissivity once the phase shift was determined. They indicated that for phase analysis the concept of constant *BE* (as determined by Clark's method) is not sufficient for removal of barometric effects. However, their analyses showed that only K1 and S2 tidal constituents are contaminated by barometric fluctuation. Hence Hsieh and others (1987) restricted their phase analysis to the M2 and O1 tidal component in order to isolate the effect of the barometric pressure fluctuations. The N2 constituent was neglected by the authors due to its small amplitude. Results obtained by this study showed that the dominant tidal component in water level fluctuation in wells tapping the confined portion of the Arbuckle-Simpson aquifer was M1. The influence of the N2 tidal component was so small that it was neglected as the case with Hsieh et al. (1987).

Ritzi et al. (1991) indicated that the Earth tide influence occurs mainly at the four principal lunar and solar diurnal and semi-diurnal frequencies (O_1, K_1, M_1, and S_1). The authors analyzed the response of water level to earth tide, atmospheric pressure, and the combined effect of both stresses. Ritzi et al. (1991) found estimates of storativity are "nearly non-unique" therefore they recommended not estimating this parameter from tidal analysis. These finding of Ritzi et al. agreed with earlier conclusions by Hsieh et al. (1987) and Rojstaczer (1988). Ritzi et al. (1991) recommended using the well response to the combined effects of earth tides and atmospheric pressure variations for the purpose of transmissivity estimates, for this approach provides more usable frequencies (identifiability window) and more information would be available to estimate T.

Hobbs and Fourie (2000) monitored water-level fluctuation in a well penetrating a confined dolomite aquifer near the Vaal River in South Africa. Hobbs and Fourie (2000) found that the water level fluctuation demonstrate a cyclic, semi-diurnal behavior (with a period of 12.3 h) which the authors attributed to earth-tide effects. The water level-fluctuations reflected another semi-diurnal cyclic pattern which was attributed to barometric effects with a period of 11.3 h. Hobbs and Fourie (2000) found that the *BE* of the aquifer to be 63 percent on the average. However, some of their calculations revealed a *BE* as high as 400 percent. The authors used a simple model based on isolating a water-level change for a given time interval and divide it by the change in barometric pressure for the same time interval.

Merritt (2004) reviewed the research that has been done on the use of tidally-influenced and other naturally-induced head fluctuations for estimating the values of aquifer parameters. Based on this review, he determined which of these methods would be useful for the hydrologic environment of southern Florida. He then applied the selected methods to data from wells in the region. Merritt (2004) used the Bredehoeft (1967) approach to compute the specific storage. The author used a modified version of Clark's (1967) method to compute the *BE* to compute the porosity based on the work of Jacob (1940) and Bredehoeft (1967). Merritt (2004) concluded that "using naturally forced data to obtain estimates of aquifer properties has been found to provide generally useful transmissivity estimates and realistic estimates of specific storage and porosity." In terms of porosity calculation, which depends on the calculation of the *BE*, the author concludes that the Clark's (1967) method of calculating the *BE* can be effective when the head data are of high quality. But the method can provide values that are too low when the head data are noisy or have strong trend. Merritt (2004) went on to conclude that the method "provided values that were too low in data sets that did not have obvious problems of these kinds."

The Arbuckle-Simpson Aquifer

The Arbuckle-Simpson aquifer is located in south-central Oklahoma within the Arbuckle Mountain Physiographic Region. The areal extent of the aquifer is about 1286 km^2 (500 mi^2) (Fairchild et al., 1990). The Arbuckle Mountain Region covers more surface area than the aquifer. Ham (1951) described in details the geology of the Arbuckle Mountain area. He estimated the total surface area of the region to be about 800 mi^2. Ham (1951) suggested that the designation of the Arbuckle outcrop as the "Arbuckle Mountain" is misleading because about 80 percent of the area consists of "gently rolling hills." Suneson (1997) stated that

> "*The greatest relief is along U. S. Highway 77. In this area, the Washita River flows at an elevation of 770 feet, and 3 miles away is the top of the East Timbered Hills- the crest of the Arbuckle anticline and, with an altitude of 1377 feet, the highest point in the Arbuckle Mountains. This total relief of 607 feet is impressive only because it is some six times greater than that of any other topographic feature between Oklahoma City and Dalla*s."

The Arbuckle Mountain region includes three main anticlines: the Arbuckle, Tishomingo, and Hunton. The Arbuckle-Simpson aquifer is hosted in two rock groups, the Arbuckle and the Simpson. Each group is composed of several formations that may differ in their water-yielding capacity. Fairchild et al. (1990) disregarded these differences and treated the aquifer as composed of two lithological units, the Arbuckle and the Simpson. Fairchild et al. (1990) estimated the thickness of the Arbuckle Group between 4000 and 6700 ft, while that of the Simpson Group between 1000 and 2300 ft.

Rocks of the Arbuckle Group are mainly middle Cambrian to early Ordovician limestone and dolomite (Puckette et al., 2009; Fairchild et al., 1990). Sargent, 1969) indicates that the rocks of the Arbuckle Group of the Hunton anticline are mainly dolomites and thinner than the Arbuckle and the Tishomingo groups which are mainly limestones. Rocks of the Simpson Group include, primarily, sandstone and shale with some middle Ordovician carbonate (Puckette et al., 2009; Fairchild et al., 1990).

Groundwater movement and occurrence are governed by the lithology and structure of these two rock groups (Fairchild et al., 1990). Puckette et al. (2009) examined the records of 150 wells within the aquifer area along with the surface geologic map and concluded that the Arbuckle Group carbonates constitute the principal hydrostratigraphic unit and the Simpson Group sandstones constitute the secondary unit. The average water well yield of the Arbuckle carbonates is about 2000 gallons per minute (about 7570 liter per minute) and the yield for the Simpson sandstones is about 200 gallons per minute (750 liter per minute) (Puckette et al., 2009). Spring discharges reveal the same yield results. Puckette et al. (2009) estimated the average yield of large-volume Arbuckle springs such as Byrd’s Mill and the Washington group to be several thousand gallon per minute. In contrast, springs draining the Oil Creek sandstones (a Simpson Group formation) reported to have a maximum yield of less than 300 gallons per minutes and an average yield of about 55 gallons per minute.

CHAPTER III

GOVERNING EQUATIONS

The governing equations are presented in two sections. The first covers the analyses of the tidally induced fluctuation for the purpose of determining the aquifer specific storage and porosity. The *BE* equations and analysis are covered in this section. The second section covers the analyses of water-level fluctuation data for the purpose of computing the aquifer transmissivity. The specific storage computation is based on resolving the amplitude of groundwater fluctuation and the amplitude of the tide potential, while the analyses for the transmissivity determination requires the resolving of the phase angle in addition to the amplitudes.

SPECIFIC STORAGE

The partial differential equation that describes the saturated ground water flow, in a homogeneous, isotropic, and extensive confined aquifer, in radial coordinates, is (Jacob, 1950; Todd and Mays, 2005)

$$\frac{\partial^2 s}{\partial r^2} + \frac{1}{r}\frac{\partial s}{\partial r} = \frac{S}{T}\frac{\partial s}{\partial t} \tag{3.1}$$

where

s is the drawdown at the well (L) in response to a discharge Q (L^3/T),

S, T, and t are storage coefficient, transmissivity, and time as defined earlier,

r is the radial distance from the well (L).

The storage coefficient of the aquifer (S) is given by:

$$S = \gamma \eta d(\beta + \frac{\alpha}{\eta}) \tag{3.2}$$

where

γ is specific weight of water (N/m^3)

η is porosity (dimensionless),

d is aquifer thickness (L),

β is the compressibility of the water (m^2/N),

and α is the bulk compressibility of the formation (m^2/N).

The storage coefficient is specific storage (L^{-1}) multiplied by the aquifer thickness. Specific storage is given by

$$S_s = \gamma \eta(\beta + \frac{\alpha}{\eta}) \tag{3.3}$$

Since water compressibility (β) is known with reasonable certainty, then the specific storage may be calculated by Equation 3.3 if we know the porosity (η) and the aquifer bulk compressibility (α). But, both parameters are not known and not easily obtainable. The traditional approach to determine specific storage is through pumping test analysis. Unlike the pumping test designed to determine T, where water-level measurement within the pumping well suffice, a pumping test for determining S must include water-level measurement in at least one observation well in addition to the pumping well. This research is concentrated on the specific storage as index to the aquifer storativity.

The water level in an open well tapping a confined aquifer responds to pressure head disturbances caused by natural stresses. It fluctuates in response to earth tide or barometric-pressure changes. The degree to which water level fluctuates in response to these stresses is determined by the well dimensions, the transmissivity, storage coefficient, and porosity of the aquifer, Cooper et al. (1965) presented the following two equations to describe the harmonic pressure head disturbance in the aquifer (h_f) and the water level response in the well (x) (Figure 3.1):

$$h_f = h_o \sin(\omega t - \phi) \tag{3.4}$$

$$x = x_o \sin(\omega t) \tag{3.5}$$

respectively, where

h_o and x_o are the amplitudes of pressure head and water level fluctuations (L), respectively,

t is time (T),

$\omega = 2\pi/\tau$, it is angular frequency of the forcing function (T^{-1}),

τ period of fluctuation (T), and

ϕ is the phase angle.

Perturbations on the aquifer cause water to flow from the aquifer to the well and back to the aquifer. The velocity of water level fluctuation in the well casing is:

$$\frac{dx}{dt} = \omega x_o \cos(\omega t) \tag{3.6}$$

The amplitude factor (AF) as defined by Cooper et al. (1965) is given by:

$$AF = \frac{x_o}{h_o} = \frac{\rho g x_o}{p_o} \tag{3.7}$$

where

ρ is density of the water in the well (M/L^3),

g is acceleration due to gravity (L/T^2),

p_o is the forcing pressure amplitude (F/L^2).

Cooper et al. (1965) described the behavior of water level responding to a seismic event as that of a mechanical system subjected to forced vibration with viscous damping. Cooper at al. (1965) presented the following equation:

$$\frac{d^2x}{dt^2} + \frac{gr_w^2}{2TH_e} Ker(\alpha_w)\frac{dx}{dt} + \frac{g}{H_e}(1 - \frac{\omega r_w^2}{2T} Kei(\alpha_w)x = \frac{p_o}{\rho H_e}\sin(\omega t - \phi) \tag{3.8}$$

where

r_w is radius of the well (L),

$H_e = H + \frac{3}{8}d,$ is the effective height of the water in the well (L),

H is initial head in the aquifer (L),

d is thickness of the aquifer (L),

$\alpha_w = r_w(\omega S/T)^{1/2}$ (dimensionless),

$p_o/\rho H_e$ is amplitude of the forcing function (dimensionless),

Ker and *Kei* are the modified Bessel function of the second kind of order zero, sometimes called the Kelvin functions.

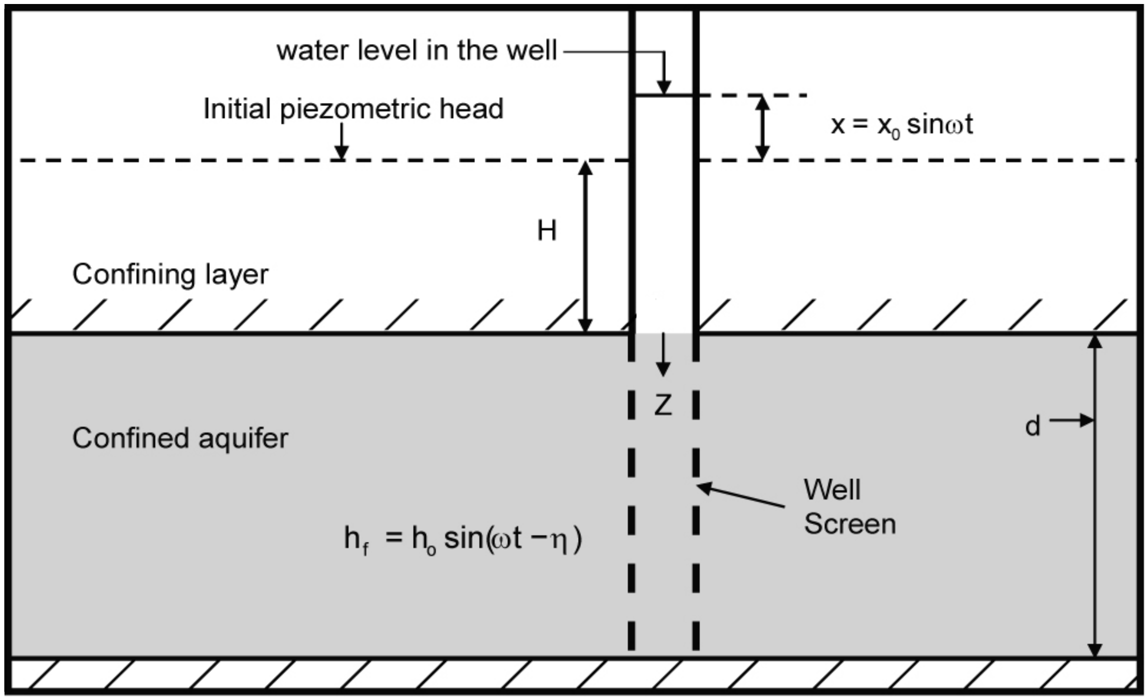

Figure 3.1. Idealized aquifer well system.

The idealization is assumed for the earth tidal effects on groundwater. Water move in and out of the well due to tidal influence causes the water level at the well to fluctuate with a magnitude of x.

The other terms were defined earlier. Several assumptions were considered in the development of Equation 3.8:

1. The well fully penetrates a homogeneous isotropic confined aquifer.
2. Inertial effects within the well were not neglected.
3. The forcing function on the aquifer is sinusoidal.
4. Drawdown is symmetric about the midpoint of the screen which is (½) *d*.
5. Flow from the aquifer to the well across the well screen is uniform.
6. The water velocity within the well screen is vertical and uniform across a horizontal section.
7. Friction forces due to flow within the well casing is negligible.

Equation 3.8 can be written in a reduced form as:

$$\frac{d^2x}{dt^2} + 2\beta\omega_w \frac{dx}{dt} + \omega_w^2 x = \frac{p_o}{\rho H_e}\sin(\omega t - \eta) \tag{3.9}$$

where

$$\omega_w^2 = \frac{g}{H_e}(1 - \frac{r_w^2\omega}{2T}Kei(\alpha_w)) \tag{3.10}$$

$$\beta = \frac{r_w^2 g}{4\omega_w T H_e}Ker(\alpha_w) \tag{3.11}$$

Equation 3.9 is analogous to the differential equation of motion of a mechanical system subjected to forced vibration with viscous damping. Cooper et al. (1965) solved Equation 3.7 for the pressure amplitude (p_o):

$$p_o = x_0 \rho H_e \left[\left(\frac{g}{H_e}\left(1 - \frac{r_w^2\omega}{2T}Kei(\alpha_w)\right) - \omega^2 \right)^2 + \left(\frac{\omega r_w^2 g}{2TH_e}Ker(\alpha_w) \right)^2 \right]^{1/2} \tag{3.12}$$

Letting ω = 2π/τ and substituting Equation 3.12 into Equation 3.7 yields the amplitude factor (*AF*) of Cooper et al. (1965):

$$AF = \left[\left(1 - \frac{\pi r_w^2}{T\tau} Kei(\alpha_w) - \frac{4\pi^2 H_e}{\tau^2 g}\right)^2 + \left(\frac{\pi r_w^2}{T\tau} Ker(\alpha_w)\right)^2\right]^{-1/2} \qquad (3.13)$$

Bredehoeft (1967) examined the amplitude factor equation (Equation 3.13) and stated "in aquifers with transmissivities in excess of about 1 cm^2/sec (0.001 ft^2/sec or approximately 600 gpd/ft) the change in pressure head (within the aquifer) due to the earth tide equals to the change in water level in the well." Bredehoeft (1967) presented the following equation for the change in head in a well produced by tidal dilatation (Δ_t):

$$-dh = \frac{\Delta_t}{S_s} \qquad (3.14)$$

where

$$\Delta_t = \left(\frac{1-2\nu}{1-\nu}\right)\left[\left(2\bar{h} - 6\bar{l}\right)\frac{W_2}{ag}\right] \quad , \qquad (3.15)$$

ν is the Poisson ratio(≈0.25, Bredehoeft, (1967)) of the aquifer material (dimensionless),

$\bar{h}$ and $\bar{l}$ are Love numbers at the surface of the earth (dimensionless),

and *a* is the radius of the earth (L).

Combining equations 3.15 and 3.14 and rearranging we obtain an expression for the specific storage (S_s) of the aquifer:

$$S_s = -\left[\left(\frac{1-2\nu}{1-\nu}\right)\left(\frac{2\bar{h} - 6\bar{l}}{ag}\right)\right]\frac{dW_2}{dh} \qquad (3.16)$$

The minus sign in equation 3.16 signifies the head (pressure) in the aquifer decreases as the tide-generating potential increases. In other words a pull caused by the transiting moon, for example, would expand the aquifer materials, hence reducing the pore pressure. The tide potential is determined from the equation:

$$W_2(\varepsilon,\phi,t) = gK_m bf(\theta)\cos[\beta(\varepsilon,t)] \tag{3.17}$$

where

K_m is the general lunar coefficient, taking into account the masses of the earth and moon,

the distance to the moon, and the earth's radius, it is equal to 53.7 cm (1.7618 ft),

b is an amplitude factor (dimensionless) that has a distinct value for each tidal component with a period τ,

$f(\theta)$ is the latitude function (dimensionless); and

$\beta\ (\varepsilon.,t)$ is a phase term that depends on the longitude ε and the Greenwich Mean Time (GMT) t.

The terms W_2 and ag in equation 3.16 have units of L^2/T^2. Therefore, S_s in Equation 3.16 has units of L^{-1}.

Merritt (2004) substituted the ratio of the tidal potential to changes in water level of Equation 3.16 by the ratio of their amplitudes. Merritt (2004) gave presented the following approximation for Equation 3.16:

$$S_s = 0.788\times10^{-12}\left(\frac{cm^2}{s^2}\right)^{-1}\frac{A_{w2}}{A_h} \tag{3.18}$$

where

A_{w2} is the amplitude of a harmonic component of W_2 and period τ.

A_{w2} is given by:

$$A_{w2} = gK_m bf(\theta) \tag{3.19}$$

and,

A_h is the amplitude of a component of the head change of period τ. The other terms have been defined earlier.

The dimensionless terms of b, $f(\theta)$, and $\beta\ (\varepsilon, t)$ were given by Merritt (2004), who correlated the work of Munck and McDonalds (1960) and Doodson and Warburg (1941) and present it in a form useful for hydrologists. Merritt (2004) Tables 4 and 7 are combined and presented for this study as Table 3.1 for seven tides. All the terms to compute A_{w2} are known and A_h can be determined from harmonic analysis if water-level fluctuations.

Van der Kamp and Gale (1983) stated "The compressibility of the solids may not be negligible for many common formations, especially if they have low compressibility and low porosity." They presented an equation for the specific storage that accounts for the compressibility of solids:

$$S_s = \rho g \left[\left(\frac{1}{K} - \frac{1}{k_s} \right)(1-\lambda) + \eta \left(\frac{1}{K_f} - \frac{1}{K_s} \right) \right] \tag{3.20}$$

where

$$\lambda = \frac{2(1-2\nu)}{3(1-\nu)} \left(1 - \frac{K}{K_s} \right) \tag{3.21}$$

K is the bulk modulus of elasticity of the formation (M/LT^2)

K_f and K_s are the bulk moduli of water and solid fraction including the effect of non-connected pores, and other symbols are as identified earlier.

Equation (3.20) was written, by Merritt (2004), in a form similar to Equation (3.16) plus a term to account for the compressibility of the solids as:

$$S_s = -\left[\left(1-\frac{K}{K_s}\right)\left(\frac{1-2v}{1-v}\right)\left(\frac{2\bar{h}-6\bar{l}}{ag}\right)\right]\frac{dW_2}{dh} \tag{3.22}$$

All symbols have been previously identified.

The bulk modulus is the inverse of compressibility and increases as the compressibility decreases. Therefore K is less than K_s and the first term in parentheses in equation (3.22) is always between 0 and 1. If the solids are incompressible, then this term is unity, and Equation (3.22) reduces to Equation (3.16).

Table 3-1. Harmonic components and some parameters of the main five tides.

These tides constitute about 95% of the tidal potential, and they are of importance to hydrogeological analyses.

Tidal Component	**Angular frequency (rad/h)**	**Frequency (cycles/day)**	**Period (h)**	**Amplitude factor (b)**	***f (θ)***	***β (ϕ, t)***
O1	0.24335189	0.92953573	25.819341	0.377	$\sin\theta\cos\theta$	$qt+\phi_s(t)-2\phi_m(t)-169.8^{\circ}+\phi$
K1	0.26251618	1.00273793	23.934469	0.531	$\sin\theta\cos\theta$	$qt+\phi_s(t)-10.2^{\circ}+\phi$
N2	0.49636693	1.89598200	12.658348	0.174	$0.5\cos^2\theta$	$2(qt+\phi_s(t)-1.5\phi_m(t)+0.5\phi_p(t)-79.8^{\circ}+\phi)$
M2	0.50586802	1.93227349	12.420602	0.908	$0.5\cos^2\theta$	$2(qt+\phi_s(t)-\phi_m(t)-79.8^{\circ}+\phi)$
S2	0.52359878	2.0000000	12.000000	0.423	$0.5\cos^2\theta$	$2(qt+\phi)$

Symbols: θ, latitude; q angular velocity of the earth relative to the mean sun (15 degrees per mean solar hour); $\phi_s(t)$, longitude of the mean sun (increasing by 0.0411degrees per mean solar hour); $\phi_m(t)$, mean longitude of the moon (increasing by 0.549 degrees per mean solar hour); $\phi_p(t)$, mean longitude of lunar perigee (increasing by 0.0046 degrees per mean solar hour); and ϕ, longitude of the observation point.

POROSITY

The porosity of a confined aquifer may be determined if the aquifer's storage coefficient and *BE* are known. Jacob (1940) presented the following equation for the relationship between, porosity, *BE*, and specific storage:

$$\eta = \frac{BE * S_s}{\rho g \beta} \tag{3.23}$$

where

β is compressibility of water.

The *BE* is given by (Jacob 1940):

$$BE = \frac{\rho g \Delta h}{db} \tag{3.24}$$

where

db is the change in barometric pressure.

The derivation of the relationship given by Equation 3.23 is presented here (Jacob, 1940; Batu, 1998; Todd and Mays, 2005).

The total stress on the top of a confined aquifer (σ_T) is balanced by the water pressure (p_w) and the compressive stress of the solid skeleton of the aquifer (σ):

$$\sigma_t = p_w + \sigma = b + \text{constant} \tag{3.25}$$

where

b is the atmospheric (barometric) pressure.

If db is the change in barometric pressure and dp_w is the change in hydrostatic pressure at the top of a confined aquifer, then

$$db = dp_w + d\sigma \tag{3.26}$$

The weight of the water column in the well above the upper boundary of the aquifer plus the change in atmospheric pressure is balanced by the hydrostatic pressure in the aquifer. Hence the change in the water pressure is given by:

$$dp_w = db + \gamma dh \tag{3.27}$$

Dividing equations 3.26 and 3.27 by γ and rearranging, two equations can be written, respectively, as:

$$\frac{db}{\gamma} = \frac{dp_w}{\gamma} + \frac{d\sigma}{\gamma} \tag{3.28}$$

$$dh = \frac{dp_w}{\gamma} - \frac{db}{\gamma} \tag{3.29}$$

By dividing Equation 3.29 by Equation 3.28 an expression for the *BE* is obtained:

$$\frac{dh}{\frac{db}{\gamma}} = -\frac{1}{\frac{dp_w}{d\sigma} = +1} \tag{3.30}$$

Since the aquifer grains are assumed incompressible, then as the aquifer is compressed the change in the bulk volume $d(\Delta V)$ is equal to the change in water volume $d(\Delta V_w)$. Note that $\Delta V_w = \eta \Delta V$. Based on these two relationships, one can write the following equation:

$$\frac{d(\Delta V_w)}{\Delta V_w} = \frac{d(\Delta V)}{\eta \Delta V} \tag{3.31}$$

The compressibility of water (β) is by definition:

$$-\beta = \frac{d(\Delta V_w)}{dp_w} \frac{1}{\Delta V_w} \tag{3.32}$$

and the compressibility of the aquifer skeleton (α), by definition is:

$$-\alpha = \frac{d(\Delta V)}{d\sigma} \frac{1}{\Delta V} \tag{3.33}$$

The negative sign accompany equations 3.32 and 3.33 signify decrease in volume with increase in stress. Substituting equations 3.32 and 3.33 into Equation 3.31 and introducing minor modification, the resulting equation is:

$$\frac{dp_w}{d\sigma} = \frac{\sigma}{\eta\beta} \tag{3.34}$$

Substituting Equation 3.34 into Equation 3.30 results in:

$$BE = \frac{dh}{\frac{db}{\gamma}} = -\frac{1}{1 + \frac{\alpha}{\eta\beta}} \tag{3.35}$$

The negative sign in Equation 3.35 indicates a decrease in water level in the well is accompanying an increase in barometric pressure. From equations 3.3 and 3.35, the specific storage (S_s) is:

$$S_s = \frac{\gamma\eta\beta}{BE} \tag{3.36}$$

Equation 3.36 is equivalent of Equation 3.23 which concludes the derivation.

Equations 3.18 and 3.23 may be used to determine the specific storage and porosity from water level fluctuation and barometric efficiency. Clark (1967) developed a method to calculate the barometric efficiency. The method employs observed changes in barometric pressure, Δb (given in height of water column units), and hydraulic head, Δh, for constant time increments. The Clark method assigns a positive sign to the barometric pressure or the hydraulic head when they are rising. The formulation involves the calculation of two sums, $\Sigma\Delta b$ and $\Sigma\Delta h$, according to the following rules:

1) when Δb is zero, neglect the corresponding value Δh in obtaining $\Sigma\Delta h$.
2) when Δb and Δh have dissimilar signs, add the absolute value of Δh in obtaining $\Sigma\Delta h$.

3) when Δb and Δh have similar signs, subtract the absolute value of Δh in obtaining $\Sigma\Delta h$.

4) $\Sigma\Delta b$ is the sum of absolute values of Δb. The barometric is calculated using:

$$BE = \frac{\sum \Delta h}{\sum \Delta b} \tag{3.37}$$

TRANSMISSIVITY

The partial differential equation that describes the saturated ground water flow in a confined aquifer, in two dimensions (Equation1.1) is rewritten:

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} = \frac{S}{T}\frac{\partial h}{\partial t} \tag{3.38}$$

All terms in Equation 3.38 were defined earlier.

Groundwater levels in coastal aquifers that are in direct hydraulic contact with oceans or confined aquifers that are intersected by a regulated surface stream are subject to fluctuation due to tidal or change of stage effects. Ferris (1951) and Todd and Mays (2005) described the propagation of these effects within a confined aquifer by the analysis of the one-dimensional flow equation:

$$\frac{\partial^2 h}{\partial x^2} = \frac{S}{T}\frac{\partial h}{\partial t} \tag{3.39}$$

where

h is the net rise or fall in the piezometric surface relative to the mean sea or stream water level, and,

x is the distance inland from the surface water body.

The boundary conditions are:

$$h = h_o \sin 2\pi t / t_0 \text{ at } x{=}0 \tag{3.40}$$

and,

$$h{=}0 \text{ at } x{=}\infty \tag{3.41}$$

where

h_o is the amplitude of the fluctuation, and t_o is the period of the ocean tide or river stage. The solution of Equation 3.39 with the applicable boundary conditions is (Ferris, 1951; Todd and Mays, 2005):

$$h = h_o e^{-x\sqrt{\pi S/t_0 T}} \sin\left(\frac{2\pi T}{t_0} - x\sqrt{\pi S/t_o T}\right) \tag{3.42}$$

Equation 3.42 defines a wave motion. The reduction of amplitude with distance is given by the factor $e^{-x\sqrt{\pi S/t_0 T}}$. Jacob (1950) indicates that when the aquifer response is due to loading effects rather than head changes at the outcrop, the amplitude factor becomes, $[\alpha/(\alpha+\eta\beta)]e^{-x\sqrt{\pi S/t_o T}}$, where α and β are as defined earlier.

From Equation 3.42 it follows that the range (twice the amplitude) h_x of ground water fluctuation at distance *x* from the shore line equals (Ferris, 1951):

$$h_x = 2h_o e^{-x\sqrt{\pi S/t_o T}} \tag{3.43}$$

Ferris (1951) studied the effect of surface stream stage fluctuation on groundwater level fluctuation. Groundwater fluctuations were measured in three observation wells in the City of Lincoln, Nebraska. Ferris (1951) defined the ratio $h_x/2h_o$ as the stage ratio: it is the ratio of groundwater fluctuation to the river stage fluctuation. Ferris (1951) adapted Equation 3.43 to compute the aquifer transmissivity in units of gallon per day per foot (7.48 gallon per cubic foot) as:

$$h_x = 2h_o e^{-4.8x\sqrt{S/t_oT}} \tag{3.44}$$

The stage ratio is given by:

$$\frac{h_x}{2h_o} = e^{-4.8x\sqrt{S/t_oT}} \tag{3.45}$$

or:

$$\ln(h_x/2h_o)/x = -4.8\sqrt{S/t_oT} \tag{3.46a}$$

Equation 3.46a is the equivalent of:

$$-\log_{10}(h_x/2h_o)/x = 2.1\sqrt{S/t_oT} \tag{3.46b}$$

When the distance *x* plotted against the range ratio on a semi-log paper, the left hand side of Equation 3.46b represent the slope of this plot. If the change in the stage ratio is selected over one log cycle, the slope would be $1/\Delta x$ where Δx is the change of the distance corresponding to one log cycle change of the range ratio. Ferris (1951) presented the following equation to compute *T* (in gallon per day per foot) from the measurement of river and groundwater fluctuations:

$$T = \frac{4.4S}{t_o\Delta x^2} \tag{3.47}$$

Hsieh and others (1987) developed an analytical method for estimating the aquifer transmissivity from the phase shift associated with each tidal component. The method is similar to Cooper et al. (1965) but neglects the inertial effect of the water stored in the well bore. Hsieh and others (1987) approach is as follows. The amplitude response (*A*) is:

$$A = \left| \frac{x_o}{h_o} \right| = \left(E^2 + F^2 \right)^{-1/2}, \tag{3.48}$$

and the phase shift ϕ is given by:

$$\phi = -\tan^{-1}(F / E), \tag{3.49}$$

where:

$$E = 1 - \frac{\omega r_c^2}{2T} Kei(\alpha_w), \tag{3.50}$$

$$F = \frac{\omega r_c^2}{2T} Ker(\alpha_w), \text{ and} \tag{3.51}$$

$$\alpha_w = \left(\frac{\omega S}{T} \right)^{1/2} r_w, \tag{3.52}$$

where

x_o is the complex amplitude of the water-level oscillation in the well (L),

h_o is the fluctuating pressure head in the aquifer (L),

r_w is the radius of the screened or open portion of the well (L),

r_c is the radius of the well casing (L),

$\omega = 2\pi/\tau$ is the frequency of the oscillations (T^{-1}); τ is the period of fluctuation (T), and

Ker and *Kei* are real and imaginary Kelvin functions of order zero.

The approach was developed for "single, laterally extensive aquifer that is homogeneous and isotropic" (Hsieh et. al., 1987). Equations 3.48 through 2.52 show that the amplitude response *A* and the phase shift ϕ are functions of two dimensionless parameters $\omega r_c^2 / T$ and α_w. Hsieh et al. (1987) used two other sets of dimensionless parameters that are “more convenient” for groundwater applications, $T\tau / r_c^2$ and Sr_w^2 / r_c^2.

The authors plotted ϕ and A, respectively, versus $T\tau / r_c^2$ for various values of Sr_w^2 / r_c^2.

Hsieh et al. (1987) plots were reproduced for this study (Figures 3.2 and 3.3). The *Ker and Kei* functions were calculated using the Maple 12 ® software.

Measured water level fluctuations and measured or calculated earth tide potential are regressed to determine the phase shift. The transmissivity is calculated by the graphs mentioned above. The method requires an independent estimate of the storage coefficient of the aquifer.

Mehnert et al. (1999) applied the results of Cooper et al. (1967) as given by Equation 3.8 to an externally forced aquifer (the external force is changes in atmospheric pressure). Mehnert et al. (1999) rewrote Equation 3.8 by expressing the amplitude of forcing function in terms of hydraulic head (x_o) and using a trigonometric identity as:

$$\frac{d^2x}{dt^2} + \frac{gr_w^2}{2TH_e} Ker(\alpha_w)\frac{dx}{dt} + \frac{g}{H_e}(1 - \frac{\omega r_w^2}{2T} Kei(\alpha_w)x = c_1 \cos(\omega t) + c_2 \sin(\omega t) \quad (3.53)$$

where

$$c_1 = \frac{-gx_o \tan\phi}{H_e\sqrt{\tan^2\phi + 1}}, \text{ and} \quad (3.54)$$

$$c_2 = \frac{-gx_o}{H_e\sqrt{\tan^2\phi + 1}} \quad (3.55)$$

Equation 3.53 has two-part solution (Mehnert (1998): homogeneous describing the transient-state and particular describing the steady-state. The steady-state solution as given by Mehnert (1998) and Mehnert et al. (1999) is presented here. The position of water level in a well *x(t)* is given by:

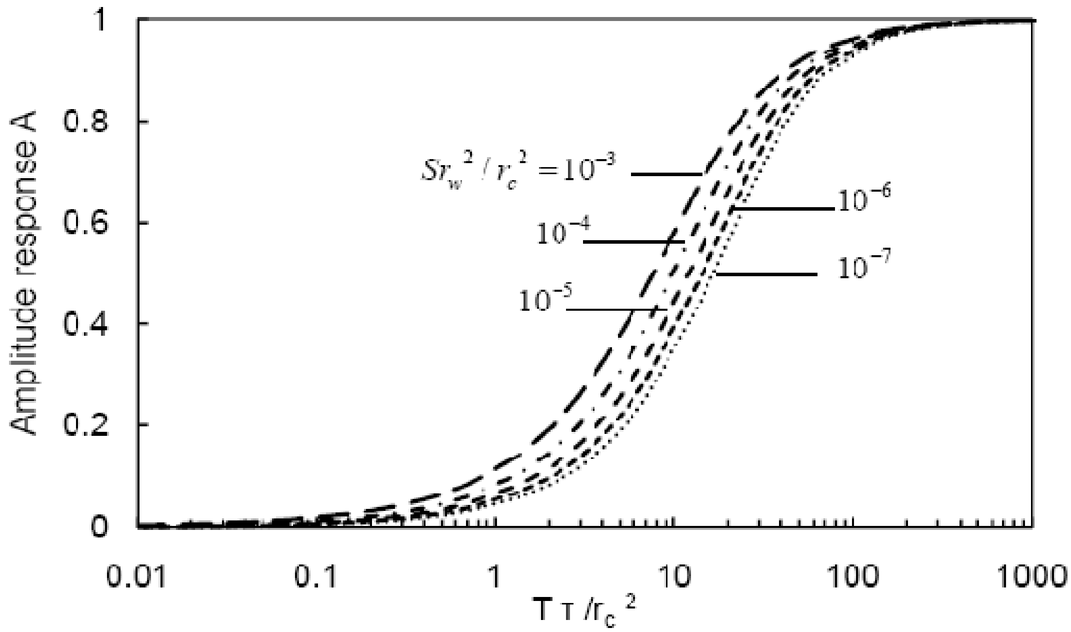

Figure 3.2. Amplitude response versus T' (after Hsieh et al., 1987)

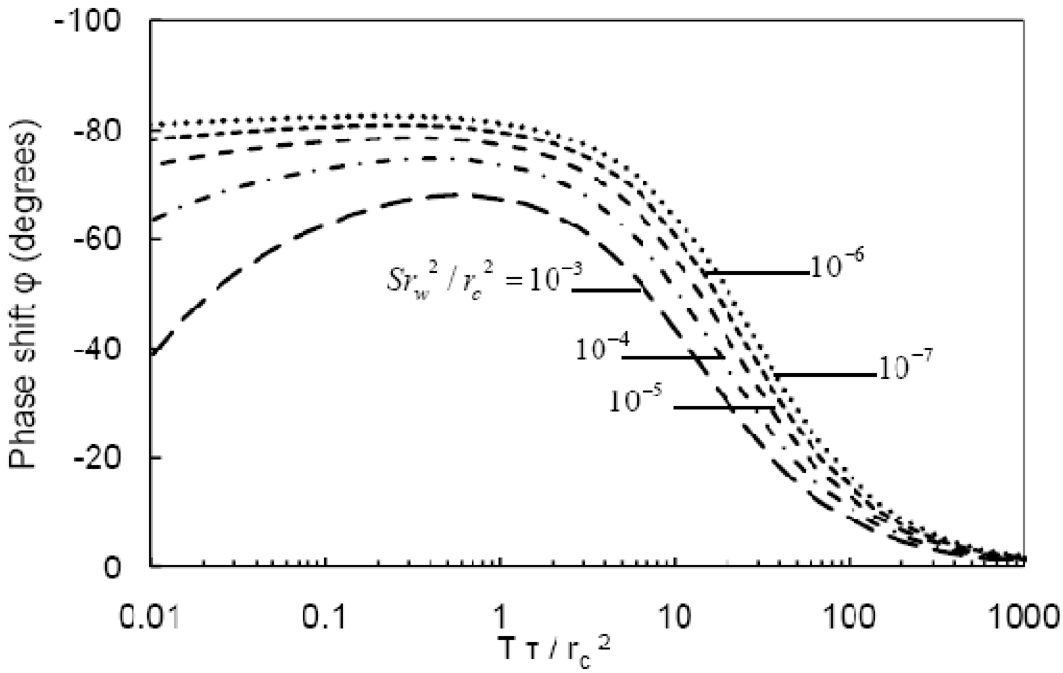

Figure 3.3. Phase shift versus T’ (after Hsieh et al., 1987)

$$x(t) = d_1 \cos(\omega t) + d_2 \sin(\omega t) \tag{3.56}$$

where

$$d_1 = \frac{-\omega^2 c_1 - \omega a c_2 + b c_1}{\omega^4 + \omega^2 a^2 - 2\omega^2 b + b^2}, \tag{3.57}$$

$$d_2 = \frac{-\omega^2 c_2 - \omega a c_1 + b c_2}{\omega^4 + \omega^2 a^2 - 2\omega^2 b + b^2} \tag{3.58}$$

$$a = \frac{g r_w^2 Ker(\alpha_w)}{2TH_e}, \text{ and} \tag{3.59}$$

$$b = \frac{g}{H_e}\left(1 - \frac{\omega r_w^2 Kei(\alpha_w)}{2T}\right) \tag{3.60}$$

Equation 3.56 can be written as a single sine term using the sine addition formula (Mehnert, 1999):

$$x(t) = AR \sin(\omega t - \phi) \tag{3.61}$$

where

AR is the amplitude ratio, it is given by:

$$AR = \frac{|x(t)|}{h_o} = \frac{\sqrt{d_1^2 + d_2^2}}{h_o} \tag{3.62}$$

and

ϕ is the phase angle given by:

$$\phi = \tan^{-1}\left(\frac{-d_1}{d_2}\right) \tag{3.63}$$

Mehnert et al. (1999) solved Equation 3.61, for known values of *T* and *S*, to generate plots of *AR* versus (*T*'') and ϕ versus (*T*''). *T'* is dimensionless transmissivity given by:

$$T' = T/\omega r_w^2 \tag{3.64}$$

Mehnert et al. (1999) designated these plots as “type curves” and recommended their use to determine *T*. The plots were regenerated for this study and are shown in figures 3.2 and 3.3. The *Ker and Kei* functions were calculated using the Maple 12 ® software. Once the *AR* is determined, the value of (T') is read from the “type curve” (Figure 3.2) and *T* is determined from the dimensionless transmissivity *T’* (Equation 3.64).

Bredehoeft (1967) and Merritt (2004) concluded that for transmissivity of more than 200 ft^2/day and the known frequencies of earth tides the, both Hsieh et al.’s (1987) and Mehnert et al.’s (1999) are becoming inapplicable. Merritt (2004) stated that “for sufficiently high values of aquifer transmissivity, the amplitude of earth-tide oscillations is not reduced in the well water-level, nor due water-level oscillations in the well exhibit a phase lag.” Bredehoeft (1967) indicated that for a transmissivity of about 85ft^2/day the change in pressure head within the aquifer due to earth tide is equal to the water level change in the well and the motion in the well is in phase with the pressure head change in the aquifer. Merritt (2004) has concluded that the method of Hsieh et el. (1987) is probably applicable to aquifers of transmissivities of less than 500 ft^2 /d. This conclusion is applicable to the Mehnert et al. (1999) method as well. The transmissivity of the Arbuckle-Simpson aquifer is very high and estimated to be in the order of 15000 ft^2/ day (Fairchild et al., 1990). Therefore neither methods discussed above can be used to estimate *T* for the aquifer. No further consideration will be given to *T* determination in this study.

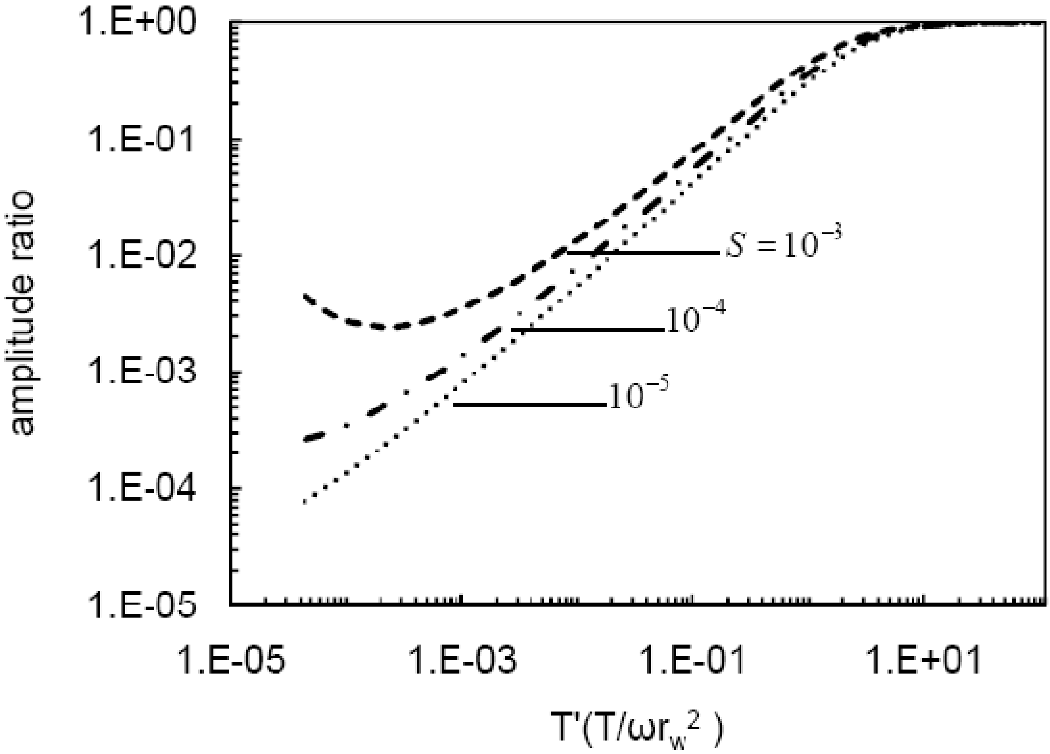

Figure 3.4. Amplitude ratio versus T' (after Mehnert et al., 1998)

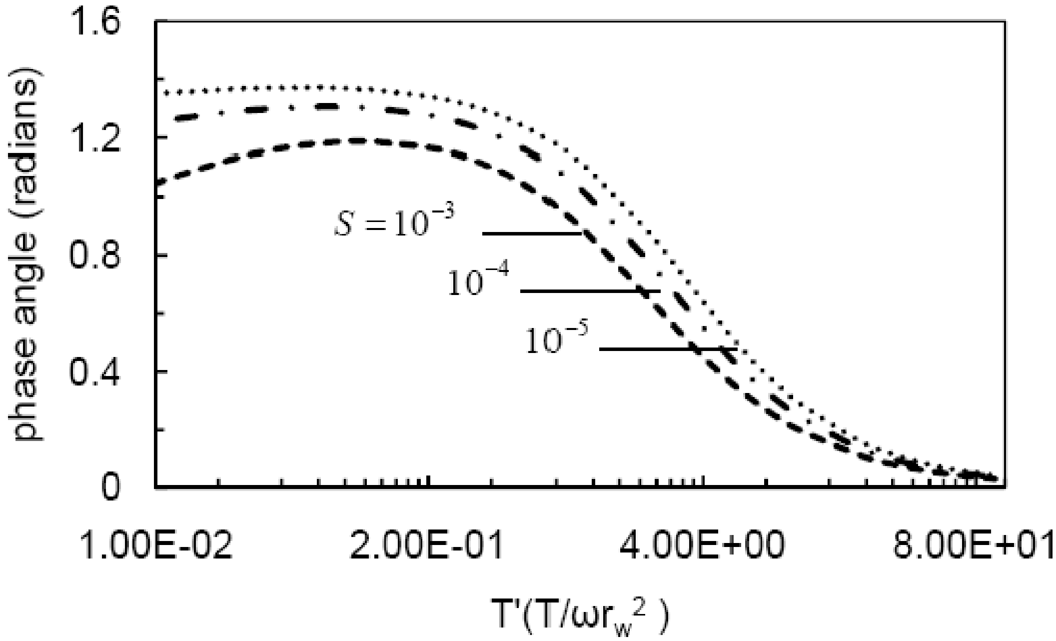

Figure 3.5. Phase angle versus T’ (after Mehnert et al., 1998)

CHAPTER IV

THE ARBUCKLE-SIMPSON AQUIFER AND MONITORED WELLS

THE ARBUCKLE-SIMPSON AQUIFER

The Arbuckle-Simpson aquifer is the study area for implementing this research. The aquifer is located in south-central Oklahoma within the Arbuckle Mountain physiographic province (Figure 4.1). The areal extent of the aquifer is greater than 1286 km^2 (500 mi^2) (Fairchild et al., 1990). The Arbuckle-Simpson aquifer is the principal source of water supplies for about 39,000 people in Ada, Sulphur, and several other towns and communities in south-central Oklahoma (OWRB, 2003).

Geology of the Arbuckle-Simpson Aquifer

The topography of the Arbuckle-Simpson aquifer area consists of two general physiographic parts: gently rolling hills to the west and generally undulating plain to the east (Ham, 1951). The two parts are separated by the Washita River. West of the Washita River the mountainous area, referred to as the Arbuckle Hills, consists of a series of northwest-trending ridges formed on resistant rocks that are characterized by intensive fold and fault system (Fairchild et al., 1990).

Ham (1951) described the Arbuckle anticline as being large anticlinal fold with an axial trend of northwest and of an approximate area of 170 mi^2. The eastern part of the mountain, referred to as the Arbuckle Plains, is characterized by a gently rolling topography formed on relatively flat-lying, intensively-faulted limestone beds with only local relief (Fairchild et al., 1990; Sargent, 1969).

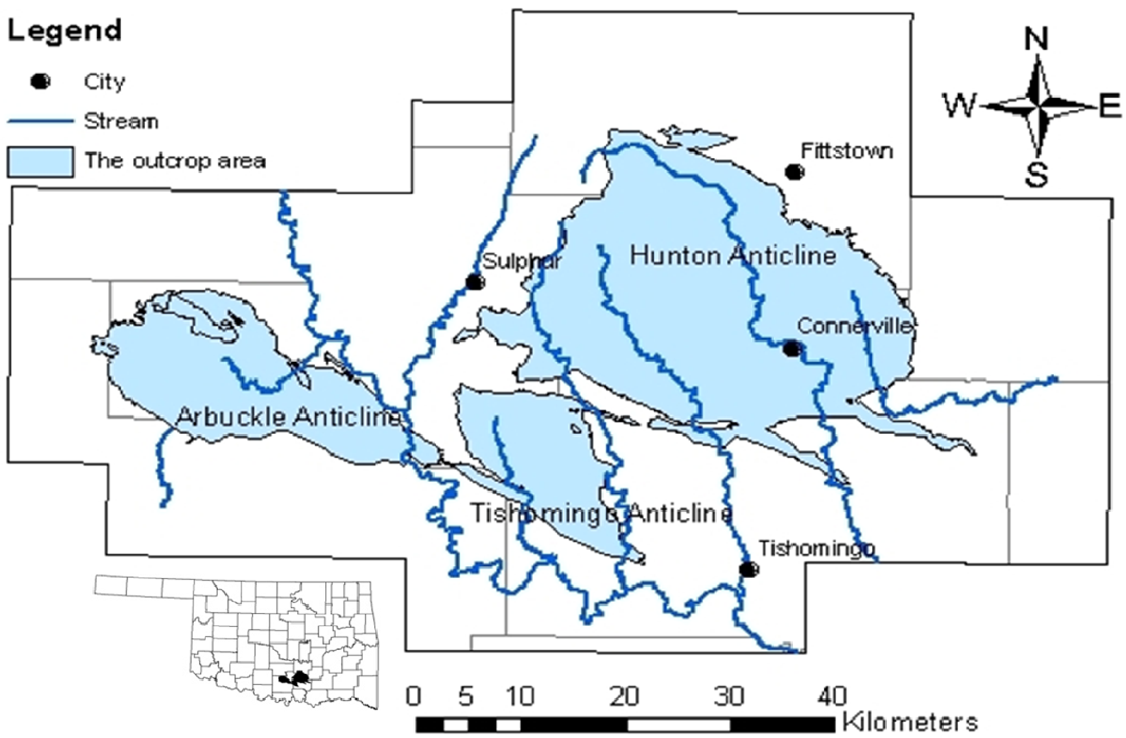

Figure 4.1. The Arbuckle-Simpson aquifer location map.

The aquifer is located in south-central Oklahoma and covers an area of about 1280 km^2 (500 mi^2)

Rocks of the Arbuckle-Simpson aquifer crop out in three anticlines within the Arbuckle Mountain physiographic province: the Arbuckle in the west and northwest, the Hunton in the east and northeast, and the Tishomingo in the south-central (Figure 4.1). The study area is restricted to the eastern parts of the Arbuckle-Simpson aquifer, namely the Hunton anticline (Figures 4.1 and 4.2).

The Arbuckle-Simpson aquifer comprises two main rock groups, the Arbuckle and the Simpson groups (Figure 4.3). The Arbuckle Group includes about eight, mainly carbonate rocks, formations. The Simpson Group consists of about five formations of mainly sandstones. The Arbuckle Group consists of limestone and dolomites that were deposited between 520 to 480 million years ago in Late Cambrian and Early Ordovician. Rocks of the Simpson group consist of sandstone, shale, and limestone that were deposited 480-460 million years ago in Middle Ordovician time (Fairchild et al., 1990; OWRB, 2003). Fay (1989) and OWRB (2009) considered the Arbuckle-Simpson aquifer to be composed of two aquifers, the Arbuckle aquifer, and the Simpson aquifer. The Honey Creek limestone, which is a formation of the Timbered Hills Group, was considered as part of the Arbuckle aquifer (Fay, 1989; OWRB, 2009).

Hydrology of the Arbuckle-Simpson Aquifer

The hydrology of the Arbuckle-Simpson aquifer was investigated for the first time by Fairchild et al. (1990). OWRB (2003) has initiated a comprehensive hydrological study which is ongoing at the time of the preparation of this dissertation. Both Fairchild et al. (1990) and OWRB (2003) studies were designed to develop an integrated approach of the surface and groundwater resources of the aquifer.

Surface Water Resources

The aquifer surface area is drained by several perennial streams (Figure 4.1). Blue River, Pennington Creek, Mill Creek, Rock Creek, Delaware Creek, and Oil Creek, are the principle streams draining the Hunton anticline and flow generally toward the south-southeast into the Washita River and Red River. Colbert, Hickory, Honey, Falls, Henryhouse, Cool, and Spring Creeks are the principle streams draining the western parts of the Arbuckle Mountains into the Washita River. The flow of the western part streams is sustained year around by spring discharges.

The total annual runoff from the eastern part of the aquifer (the Hunton outcrop, surface area of 398 mi^2) is estimated to be 7.6 in/y. The base flow mounts to about 4.7 in/y or about 60 percent of the total annual runoff. Seventy percent of this annual runoff is carried out by three streams: Blue River near Connerville (watershed area of 162 mi^2), Mill Creek near Mill Creek (watershed area 46.4 mi^2), and Pennington Creek near Reagan (watershed area of 65.7 mi^2) (Fairchild et al., 1990). Fairchild et al. (1990) analyzed the stream flow hydrographs for several years to estimate the mean annual discharge for each one of the three streams. The estimated average annual discharges are 83, 15, and 56 cubic feet per second (cfs) for Blue River, Mill Creek, and Pennington Creek respectively. The majority of runoff comes from surface runoff during wet seasons and from base flow during long-lasting dry periods.

Groundwater Resources

Groundwater regime in the Arbuckle-Simpson aquifer is affected by the complex geologic features of the aquifer. Fairchild et al. (1990) suggested that occurrence and movement of groundwater in the aquifer are strongly controlled by lithology and structure. The aquifer is a carbonates-rock aquifer exhibiting karst features, such as solution channels, especially in the western parts. Features such as folds, faults, fractures, and solution channels control groundwater flow rates and movements (OWRB, 2003). Fairchild et al. (1990) suggested that the association of springs with faults in the Arbuckle-Simpson aquifer indicates the strong structural control on groundwater movement. Flow rates can vary greatly depending on the flow paths. Groundwater moves slowly through fine fractures and pores and rapidly through solution-enlarged fractures and solution channels.

Recharge to the aquifer comes mainly from precipitation. The long-term annual average precipitation in the area is 38.2 in/y. Fairchild et al. (1990) estimated the recharge to the Arbuckle-Simpson aquifer based on the total average annual base flow from streams that drain the area to be about 4.7 in/y. Fairchild et al. (1990) indicated that discharge from the aquifer was through springs and seeps from groundwater to surface water streams. Discharge by evapotranspiration and pumping was minor. Fairchild et al. (1990) estimated that about 55 percent of the mean annual discharge of 154 cubic feet per second (cfs) from Blue River, Mill Creek, and Pennington Creek is base flow. Add to this amount the estimated annual average flow of 15 cubic feet per second from Byrds Mill Spring, and about 25 cfs which is the estimated base flow of 10 small streams in the area, the estimated annual mean discharge of the eastern part (Hunton anticline) of the Arbuckle-Simpson aquifer amounts to 125 cfs. The total drainage area of the main three creeks, the 10 small streams, and the Byrds Mill Spring, is estimated to be 359 mi^2 (Fairchild, 1990). Therefore, the annual mean discharge in units of depth amounts to 4.7

in/y. Since, the discharge is equal to the recharge, it can be concluded that the aquifer is under steady-state flow conditions.

The general flow direction in the Hunton Anticline is from northwest to southeast (Fairchild et al., 1990; OWRB, 2005). A groundwater contours map published by Fairchild et al. (1990) suggest that recharge occurs in the northwestern part of the aquifer, while discharge occurs within the southeast. More studies are needed to characterize the groundwater flow regime. The storativity of the aquifer was estimated to be 0.008, and transmissivity was 15,000ft^2/day (Fairchild et al., 1990). Porosity values of the Simpson Group sandstones ranged from <10-26 percent (Puckette et al., 2009). No porosity values estimates are available for the Arbuckle Group of rocks.

MONITORED WELLS

Instrumentation

Groundwater fluctuations in open wells were monitored and recorded at 15-minute intervals using pressure transducers. The Solinst Levelogger® transducers were used for the Oklahoma State University (OSU) monitored wells, and In Situ MiniTrolls and LevelTrolls®"were used the OWRB wells. The United States Geological Survey (USGS) Fittstown well utilized a KPSI Series 500 pressure transducer. The atmospheric-pressure fluctuation was recorded for the same time sequences in situ using a Solinst Levelogger attached in the upper two meters of the well OWRB 101246.

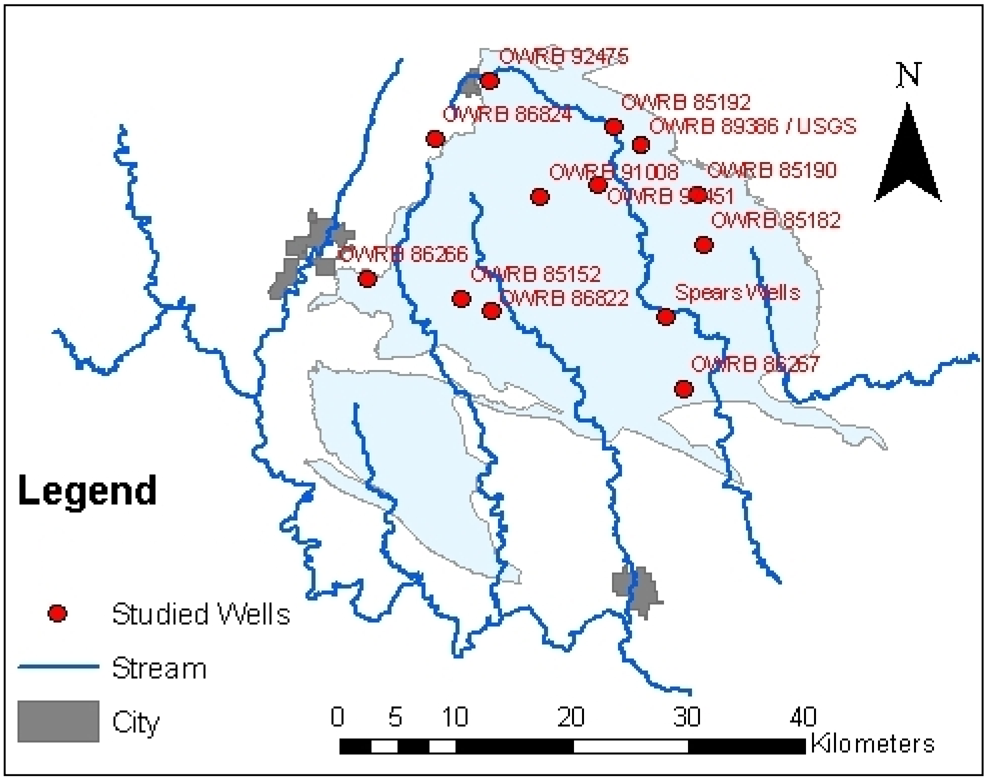

Figure 4.2. Wells location map of the Arbuckle-Simpson aquifer.

The wells were within the Hunton Anticline. The aquifer is located in south-central Oklahoma. The Hunton Anticline is the eastern part of the aquifer.

<table>
<tr><th>System</th><th colspan="3">Group and Formation</th></tr>
<tr><td rowspan="3">ORDIVICIAN</td><td>Viola Group</td><td colspan="2">Fernvale Formation
Viola Springs Formation</td></tr>
<tr><td>Simpson Group</td><td>Bromide Formation
Tulip Creek Formation
McLish Formation
Oil Creek Formation
Joins Formation</td><td>Simpson Aquifer</td></tr>
<tr><td rowspan="2">Arbuckle group</td><td>West Spring Creek Formation
Kinblade Formation
Cool Creek Formation
McKenzie Hill Formation
Butterly Dolomite</td><td rowspan="3">Arbuckle Aquifer</td></tr>
<tr><td rowspan="4">CAMBRIAN</td><td>Signal Mountain Formation
Royer Dolomite
Fort Sill Limestone</td></tr>
<tr><td rowspan="2">Timbered Hill Group</td><td>Honey Creek Limestone</td></tr>
<tr><td colspan="2">Reagan Sandstone</td></tr>
<tr><td colspan="3">Colbert Rhyolite</td></tr>
</table>

Figure 4.3. Idealized stratigraphic section for the Arbuckle-Simpson aquifer (after Fay, 1989; Puckette et al., 2009).

The atmospheric-pressure data are needed to compute the *BE* and to correct the water level readings when the water-level transducer is not vented as the case with the Levelogger. OWRB and USGS transducers were vented and did not need the barometric compensation. The KPSI Series 500 transducer had a 70-meter full range and an accuracy of 0.05 percent of the full scale (3.5 cm). The Solinst Levelogger transducers had a higher sensitivity with a 5-meter full range and an accuracy of 0.05 percent of full scale (0.25 cm). The resolution (precision) of the Solinst Levelogger is +/- 0.05 mm. All of the data analyzed for this study utilized the Solinst Levelogger, as the other loggers were not found to be sufficiently precise for the analyses performed.

Studied Wells

Fourteen wells distributed over the eastern part of the Arbuckle-Simpson aquifer were used for this study. Of the 14 wells, two, OWRB 101246 (Spears test 1) and OWRB 101247 (Spears test 2), were drilled as part of the Arbuckle-Simpson aquifer hydrological study program. The monitoring and data collection within these wells were performed by the OSU, Boone Pickens School of Geology. They were monitored for a period of more than one year. The monitoring program started in January of 2007 and ended in February of 2008. A Solinst Barologger® (a pressure transducer to log the atmospheric-pressure changes) was installed at the site of these two wells. The atmospheric pressure was monitored for the same period as water-level monitoring. The remaining 12 wells were installed prior to this study. One well, OWRB 89386 (USGS Fittstown), was owned and administered by the USGS. The rest of the wells were either private or owned by the OWRB, but all of these wells data were collected by the OWRB. Data collected by the OWRB and the USGS were also analyzed for this study.

The locations of the study wells are shown in Figure 4.2. Table 4.1 presents the wells and some of their parameters. Two of the studied wells are open in the Simpson Group (mainly sandstone) and the remaining wells are open completion wells in the Arbuckle Group (mainly carbonate rocks). The depths of most of the wells were taken from the records of the OWRB. The question mark on the depth of the well OWRB 86824 signifies doubts of the author about the reliability of the reported depth. Data analysis (as will be seen later) indicates that the well may much deeper than reported.

Water-level and barometric-pressure fluctuations were studied and analyzed to determine the specific storage and the porosity. The *BE* was determined and used with the specific storage values to determine the porosity of the aquifer. All monitored wells were not subjected to full analysis. Some wells were not considered for analysis because the data collected did not reveal earth-tides influence. The measured water-level data must demonstrate periodicity as an index of the tidal influence. Preliminary analyses revealed the lack of periodic behavior as shown in Figure 4.4; the spikes in water level measurement which oscillate between +7 to -7 cm every 24 hours may result from an equipment malfunction.

Data Processing and Analysis

Transducers employed for water-level monitoring were not vented. Therefore, they recorded a pressure that included water and atmosphere loads. The first step in data processing was to remove the barometric pressure component from the water-level fluctuations data. Atmospheric pressure components were removed by subtracting the monitored atmospheric pressure from the raw water-level data for each time step. A sample of raw data is shown in Figure 4.5 along with the moon phases. Figure 4.5 shows the water-level fluctuations, the fluctuations compensated for barometric pressure, and the barometric pressure fluctuation for the OWRB 101246 (Spears 1 test well).

After removing the atmospheric-pressure component, the raw data were filtered and smoothed to remove the trend that may be present. The trend may be caused by regional flow, evapotranspiration, remote pumping, and seasonal changes. Filtering and smoothing are achieved using two smoothing techniques: differencing and/or moving average. The difference filter is given by (Shumway and Stoffer, 2006):

$$y_t = x_t - x_{t-1} \tag{5.1}$$

where

y_t is the differenced head at time t, and

x_t is the measured head fluctuation at time t.

The symmetric moving average is given by:

$$y_t = \sum_{j=-k}^{k} a_j x_{t-j} \tag{5.2}$$

where

$a_j = a_{-j} \geq 0$, and

$\sum_{j=-k}^{k} a_j = 1$

Differencing is an example of a high-pass filter because it retains or passes the higher frequencies. The moving average is a low-pass filter because it passes or retains the lower (slower) frequencies (Shumway and Stoffer, 2006). Equations 5.1 and 5.2 were both applied for smoothing the collected data.

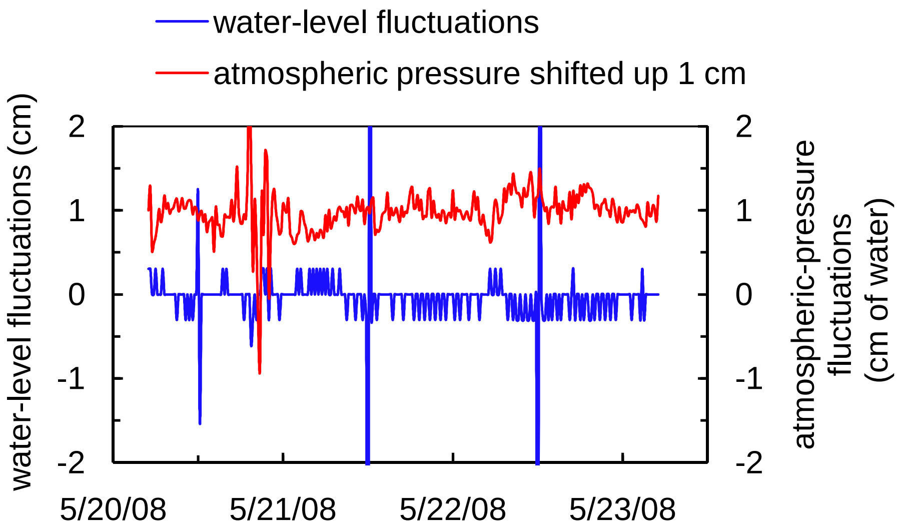

Figure 4.4. Water-level and atmospheric-pressure fluctuations for the well OWRB 91008.

The well was located in the Arbuckle-Simpson aquifer. Atmospheric pressure shifted downward one centimeter for the clarity of the figure to avoid having the data lay directly over the water-level data. Equipment malfunction may be caused the spikes within the water-level data. No other explanation can be offered.

Table 4-1. Studied wells and their depths.

The study area was the Arbuckle-Simpson aquifer.

Well Designation	Total Depth [m]/[ft]	Latitude [degrees]	Longitude [degrees]	Geological Formation
OWRB 92475	33.8/111	34.63067375	-96.82099024	Simpson
OWRB 86266	34.1/112	34.47692512	-96.93631684	
OWRB 85182	16.1/53	34.50518221	-96.61764855	Arbuckle
OWRB 85190	25.3/83	34.54411280	-96.62234170	
OWRB/USGS 86267	23/75	34.39340823	-96.63553401	
OWRB 85152	36.3/119	34.46265495	-96.84539322	
OWRB 91008	46/151	34.54196359	-96.77272160	
OWRB 86822	61/200	34.45271294	-96.81835462	
OWRB 85192	61.3/201	34.59666273	-96.70333808	
OWRB 86824	76/250?	34.58555358	-96.87247790	
OWRB 97451 Fittstown mesonet	78/257	34.55205563	-96.71793300	
OWRB 89386/USGS Fittstown	121/396	34.58288903	-96.67951376	
OWRB 101246 (Spears test 1)	183/600	34.449633	-96.6526158	
OWRB 101247 (Spears test 2)	548/1800	34.4494431	-96.6521400	

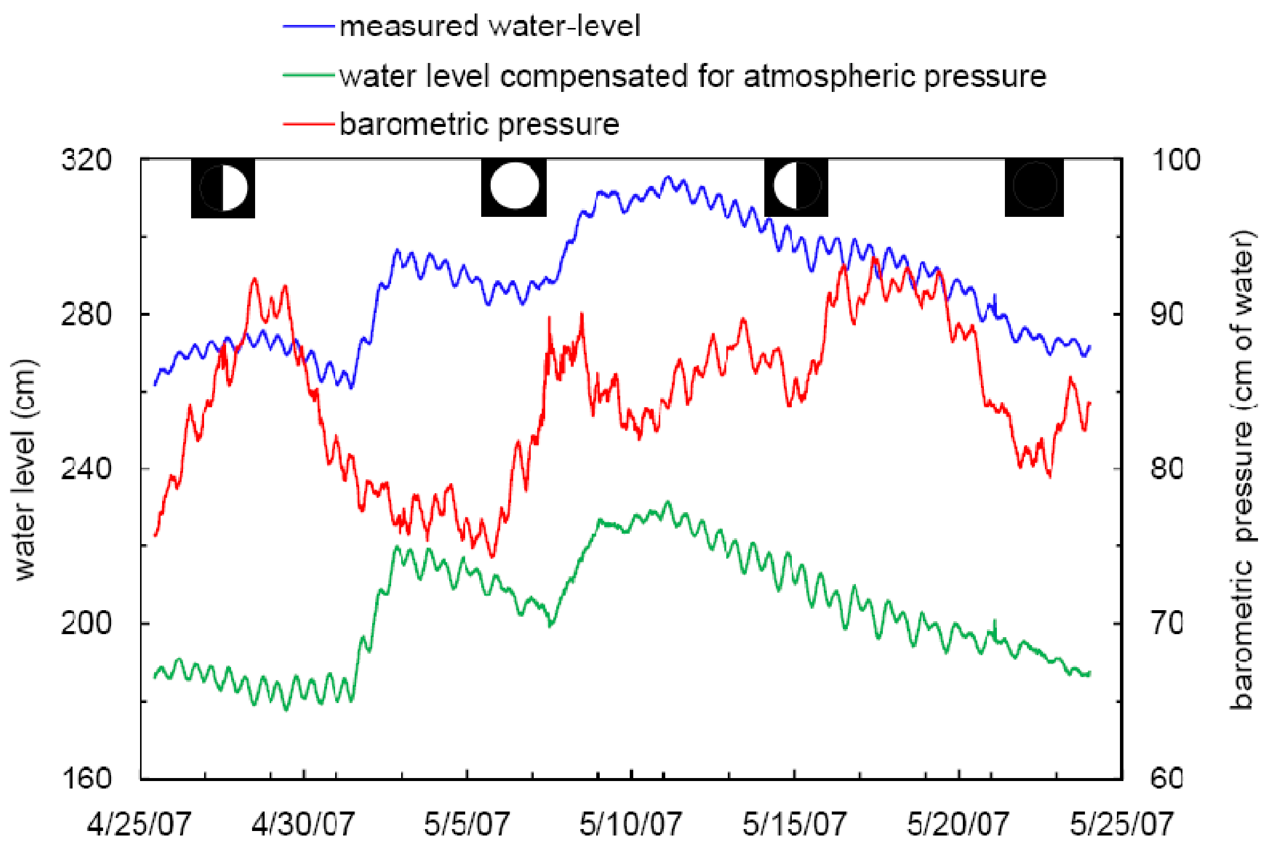

Figure 4.5. Measured water levels, water level compensated for atmospheric pressure, and barometric pressure for the well OWRB 101246 (Spears 1 test well). The well was located within the Arbuckle-Simpson aquifer. The data are for one lunar month and the four moon phases are shown in the figure.

CHAPTER V

ANALYZING WATER-LEVEL TIME SERIES TO IDENTIFY AQUIFER TYPE

ABSTRACT

Aquifer type, confined, unconfined, or semi-confined, may be identified by drilling or performing pumping tests. Both methods are costly, involve complex field issues, and may yield inconclusive results. Earth tides are known to influence water levels in wells penetrating confined aquifers or unconfined thick, low-porosity aquifers. Water-level fluctuations in wells tapping unconfined aquifers are also influenced by changes in barometric pressure. Time-series analyses of water-level fluctuations of the Arbuckle-Simpson aquifer were utilized in nine wells to identify aquifer type by evaluating the influence of earth tides and barometric pressure variations. The Arbuckle-Simpson is a thick (~1000 m) carbonate aquifer located in south-central Oklahoma. Two types of harmonic analyses were employed to determine aquifer type: 1) signal identification and 2) amplitude and phase angle determination. Based on the results, portions of the Arbuckle-Simpson aquifer responded as each type of aquifer. The results demonstrated that the technique was an accurate and low-cost method to determine aquifer type.

INTRODUCTION

Earth tides are known to influence water level in wells penetrating confined aquifers or unconfined thick, low porosity aquifers (Bredehoeft, 1967; Weeks 1979). Water-level fluctuations in wells tapping unconfined aquifers are also influenced by changes in barometric pressure (Weeks, 1979). Time series analyses of water-level fluctuation of an aquifer may be utilized to identify its type, i.e. confined, unconfined, and semi confined.

Traditionally, aquifer type is identified by drilling or performing pumping test on existing wells. Both methods are costly and involve tedious field work. Analyzing water-level time series to identify the aquifer type constitutes an appealing alternative to the existing methodology. The purpose of this paper is to identify the signals that were present in the water-level fluctuations data for several wells and, hence, identify the aquifer types for several wells located in the same sequence of lithology that may behave differently hydraulically. Two types of harmonic analyses were employed: signal identification and amplitude and phase angle determination. Based on the identified tides, aquifer types were identified in several localities of the Arbuckle-Simpson area.

THE ARBUCKLE-SIMPSON AQUIFER

The Arbuckle-Simpson aquifer is located in south-central Oklahoma within the Arbuckle Mountain Physiographic Region (Figure 5.1). The Arbuckle Mountain region includes three main anticlines: the Arbuckle, Tishomingo, and Hunton. The aquifer is hosted in two rock groups, the Arbuckle and the Simpson. Each group is composed of several formations that may differ in their water-yielding capacity (Figure 5.2). Fairchild et al. (1990) disregarded these differences and treated the aquifer as composed of two lithological units, the Arbuckle and the Simpson. The areal extent of the aquifer is estimated to be 1280 km^2 (500 mi^2). The estimated thickness of the Arbuckle Group I between 1220 and 2040 m (4000 and 6700 ft), while that of the Simpson Group is between 305 and 700 m (1000 and 2300 ft) (Fairchild et al., 1990).

Rocks of the Arbuckle Group are mainly middle Cambrian to early Ordovician limestone and dolomite (Puckette et al., 2009; Fairchild et al., 1990). Sargent (1969) indicated that the rocks of the Arbuckle Group of the Hunton anticline are mainly dolomites and thinner than the Arbuckle and the Tishomingo groups which are mainly limestones. Rocks of the Simpson Group include, primarily, sandstone and shale with some middle Ordovician carbonate (Puckette et al., 2009; Fairchild et al., 1990).

Groundwater movement and occurrence are governed by the lithology and structure of these two rock groups (Fairchild et al., 1990). Puckette et al. (2009) examined the records of 150 wells within the aquifer area along with the surface geologic map and concluded that the Arbuckle Group carbonates constitute the principal hydrostratigraphic unit and the Simpson Group sandstones constitute the secondary unit Fairchild et al. (1990).

Fairchild et al., (1990) suggested that the aquifer is not entirely confined; it is confined in some areas, semi-confined and unconfined in others. Fairchild et al. (1990) have not identified the localities at which the aquifer was confined, unconfined, or semi-confined. A recent study (OWRB, 2009) has considered the entire aquifer as confined. As the majority of the Hunton anticline area is composed of kilometer-thick scale dolomite, determining the aquifer type using lithology is not possible; other methods must be employed.

METHODOLOGY AND ANALYSES

Groundwater-level fluctuations collected from nine wells were analyzed to verify the presence of the tidal components (Figure 5.3 and Table 5.1). Water-level fluctuations were monitored by the Solinst Levelogger ®. The barometric-pressure changes were monitored in three well sites: OWRB 101246 (Spears test 1), OWRB 86267, and OWRB 86824. Out of the infinite number of harmonic components of the tide potential, only five are of importance to groundwater studies (Bredehoeft, 1967; Melchior, 1964). These components are: O1, K1, M2, S2, and N2 (Table 3.1). The five components account for about 95 percent of the tidal potential influencing groundwater levels. The frequencies of these tides are known precisely from the field of astronomy (Merritt, 2004).

Water-level data were detrended by differencing to remove long term trends and to insure that the time series was stationary. Difference filter is given by (Shumway and Stoffer, 2006):

$$y_t = x_t - x_{t-1} \tag{5.1}$$

where

y_t is the differenced head at time t, and

x_t is the measured head fluctuation at time t

Analysis of time series to determine the type of aquifer was accomplished by applying the discrete Fourier transforms technique (Shumway and Stoffer, 2006). The technique regresses measured data on sine and cosine functions. If these data contain a periodic component corresponds to the studied frequency, a Fourier transform value that is much higher than zero is obtained.

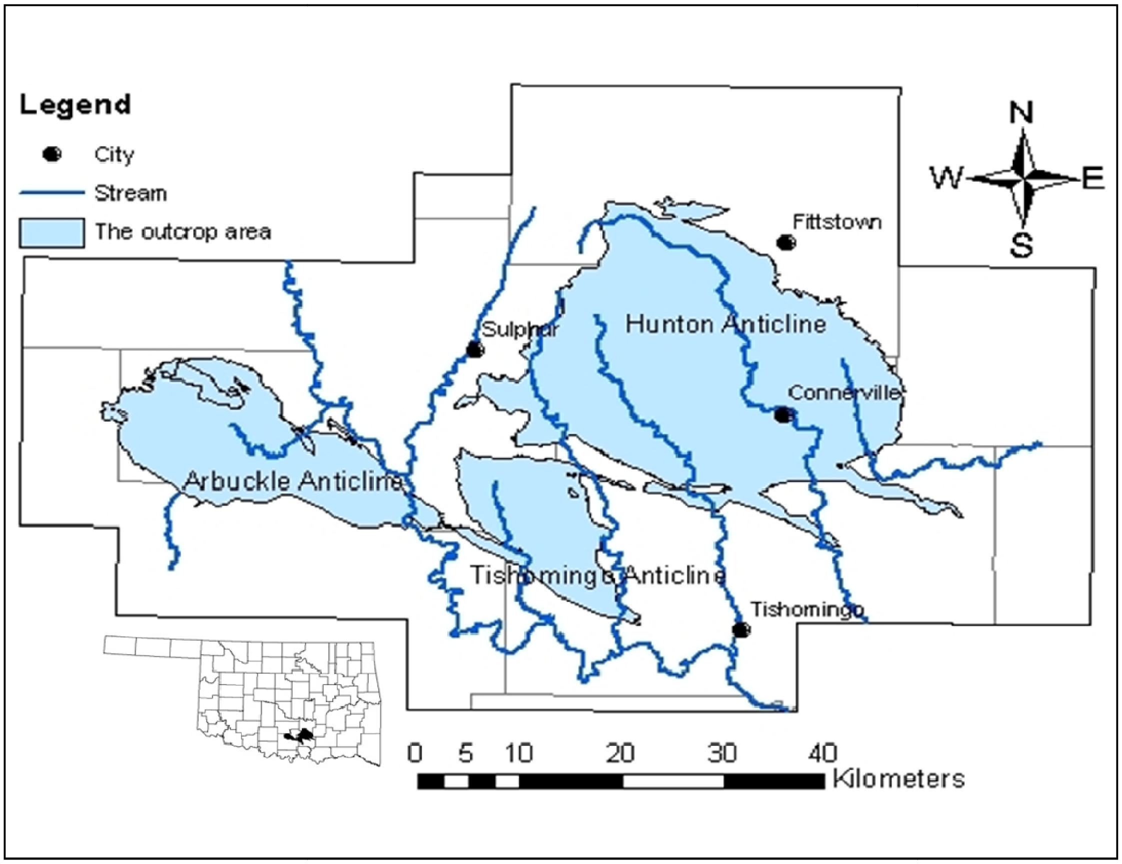

Figure 5.1. The Arbuckle-Simpson aquifer location map.

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-Cool Creek Formation
-McKenzie Hill Formation
-Butterly Dolomite</td><td rowspan="3">Arbuckle Aquifer</td></tr>
<tr><td rowspan="4">CAMBRIAN</td><td>-Signal Mountain Formation
-Royer Dolomite
-Fort Sill Limestone</td></tr>
<tr><td rowspan="2">Timbered
Hill
Group</td><td>-Honey Creek Limestone</td></tr>
<tr><td colspan="2">-Reagan Sandstone</td></tr>
<tr><td colspan="3">Colbert Rhyolite</td></tr>
</table>

Figure 5.2. Idealized stratigraphic section for the Arbuckle-Simpson aquifer (after Fay, 1989; Puckette et al., 2009).

Mathematically, the harmonic analysis may be summarized as follows. Let $x_1,\ldots,x_n$ be a time series observations of size n. Then, the series can be represented by the Fourier series as a regression model (Shumway and Stoffer, 2006):

$$x_t = a_0 + \sum_{j=1}^{(n-1)/2} \left[a_j \cos(2\pi tj/n) + b_j \sin(2\pi tj/n)\right] \tag{5.2}$$

where

j/n = period of the j^{th} harmonic component.

The Fourier coefficients a_j and b_j may be estimated by:

$$a_j = \frac{2}{n}\sum x_t \cos(2\pi tj/n) \tag{5.3}$$

$$b_j = \frac{2}{n}\sum x_t \sin(2\pi tj/n) \tag{5.4}$$

The coefficient a_0 simplifies to the mean of the time series:

$$a_0 = \frac{1}{n}\sum_{t=1}^{n} x_t \tag{5.5}$$

The Fourier coefficients a_j and b_j are regression coefficients for each *j* harmonic. They are, in essence, the correlation of these data with the sinusoidal oscillating at *j* cycles in n time points. A measure for the presence of a frequency of oscillation of *j* cycles in *n* points of time in the measured data may be given by (Shumway and Stoffer, 2006):

$$P(j/n) = a_j^{\,2} + b_j^2 \tag{5.6}$$

where

$P(j/n)$ is the periodogram (i.e. the signal indicator).

Tide period in (hour/tide) plotted against the resulted Fourier transforms to produce a periodogram for each studied well. Signal analysis was done for water-level and atmospheric-pressure fluctuations data. The computation package that was utilized is the MATLAB® version of 2009.

The least squares regression method was applied to the time series to determine the amplitude and the phase angle of the various tidal components. The regression process was outlined by Nowroozi and others (1966) and encoded in FORTRAN by Hsieh et al. (1987) and presented here. The least squares method, which minimizes the sum of squares of a set of residuals (SSR), is presented in the following paragraphs:

Let x_i be the i^{th} measured pressure head fluctuation corresponding to time t_i:

$$SSR = \frac{1}{n}\sum_{i=1}^{n}\left[x_i - \frac{a_0}{2} - \sum_{j=1}^{p}(a_j \cos\omega_j t_j + b_j \sin\omega_j t_j)\right]^2 \tag{5.7}$$

where:

n is number of measured pressure head points,

t_i is the time of the i^{th} pressure head measurement,

ω_i is frequency of the i^{th} tidal component,

$P=5$, it is the number of tidal components considered for the calculations, and

a_0, a_j and b_j are the $2P+1$ unknown coefficients to be determined by the least square method.

Equation 5.7 is differentiated with respect to the $2P+1$ unknown coefficients to produce $2P+1$ equations. Each equation is set to equal zero to minimize the sum of the squares. The resultant system of linear equations is solved to obtain the $2P+1$ unknown coefficients. The amplitude of the j^{th} tidal component (A_j) is computed by:

$$A_j = (a_j^2 + b_j^2)^{1/2}, \tag{5.8}$$

and its phase angle (ϕ_j) is given by:

$$\phi_j = \tan^{-1}(b_j / a_j) \tag{5.9}$$

Table 5-1. Studied wells for tidal detection and aquifer identification.

OWRB (Oklahoma Water Resources Board) and USGS (United States Geological Survey). The study area was the Arbuckle-Simpson aquifer.

Well Designation	Total Depth [m]/[ft]	Latitude [degrees]	Longitude [degrees]	Geological Formation
OWRB 86266	34.1/112	34.47692512	-96.93631684	**Simpson**
OWRB/USGS 86267	23/75	34.39340823	-96.63553401	**Arbuckle**
OWRB 85152	36.3/119	34.46265495	-96.84539322	
OWRB 91008	46/151	34.54196359	-96.77272160	
OWRB 86824	76/250?	34.58555358	-96.87247790	
OWRB 97451 Fittstown mesonet	78/257	34.55205563	-96.71793300	
OWRB 89386 USGS Fittstown	121/396	34.58288903	-96.67951376	
OWRB 101246 (Spears test 1)	183/600	34.449633	-96.6526158	
OWRB 101247 (Spears test 2)	548/1800	34.4494431	-96.6521400	

RESULTS AND DISCUSSION

Two types of harmonic analyses were performed on the water-level data. The first analysis was tidal signal identification, and the second was amplitude and phase angle determination. The two types of analyses were meant to supplement each other.

Tidal Signal Identification

Results of signal analysis indicated that three of the studied wells demonstrated the characteristics of a confined aquifer. Figure 5.4 shows water-level fluctuations for the well OWRB 101246 (Spears test 1) as an example of the three wells. The variation of amplitude seen in Figure 5.4 indicates that the water-level changes were influenced by tidal forces of lunar origin. The resulted periodogram is shown in Figure 5.5. The periodogram shows the presence of four tidal components: two diurnals K1 and O1, and two semidiurnal M2 and S2. Figure 5.5 reveals that M2 (main lunar semidiurnal) and S2 (main solar semidiurnal) are the dominant tides on the well water levels. N2 tide was not detected on the water-level data (Figure 5.5). The well depth is 183 m (600 ft). The strong signal of M2 tide along with the presence of O1 is an indication that this well is tapping a confined portion of the Arbuckle-Simpson aquifer.

Similar results were obtained for the well OWRB 101247 (Spears test 2) and the well OWRB 86824. The wells are 548 m (1800 ft) and 76 m (250 ft) deep respectively. OWRB 101247 (Spears 2) is located about 50 meters east of the well OWRB 101246 (Spears test 1). OWRB 86824 is located in the northwest corner of the studied area, 24 km from the Spears wells. Geologic data (Fairchild et al., 1990) indicate that the well OWRB 86824 penetrates the same geologic unit that is penetrated by the Vendome Well of Sulfur, Oklahoma. The Vendome is a flowing artesian well. This accentuates the conclusion that the well OWRB 86824 is taping a confined portion of the Arbuckle-Simpson aquifer.

The results for the three wells indicated no variation with depth. This finding contradicts previous results reported by other researchers. Bredehoeft (1967) indicated that increase of amplitude with depth may be attributed to a decrease in porosity of the aquifer in its deeper portions. Bredehoeft (1967) also suggested that the permeability of the confining layers decreases with depth, "so that deeper aquifers more closely approximate ideal artesian conditions." The constant response to tidal stress throughout the aquifer thickness is a unique property for the Arbuckle-Simpson aquifer and it was noticed by other researchers. Two other aquifer studies found no variation with depth in temperature or chemical composition (Puckette et al. (2009); Christenson et al. (2009)).

Harmonic analyses of the wells OWRB 89386 (also known as USGS well/Fittstown) and OWRB 97451 (also known the Fittstown mesonet well) indicate that the wells were completed in a semi-confined portion of the aquifer. The wells were located in the northeast part of the studied area. The periodograms of the well 89386 is presented in Figures 5.6 as a typical example of the two wells. The dominant tidal components were S2 (solar semidiurnal) and K1 (lunar-solar diurnal) as observed in the periodogram. The frequencies of these tides are the same as those of the atmospheric-pressure fluctuations (Figure 5.7). The results suggest that water-level fluctuations within these two wells were influenced primarily by the atmospheric-pressure fluctuations. This result indicated that these wells were penetrating an aquifer that was not confined. The presence of M2 tide, however (Figure 5.6) indicated that earth tides are also influencing the water-level fluctuation but to a lesser degree compared to the atmospheric pressure. If M2 was present and dominated the spectrum, the aquifer is confined (Bredehoeft, 1967). If M2 was not present and the dominant tide was S2, the aquifer is unconfined (Weeks, 1979). If both are present, the dominant S2, it is considered by this study as the case of semi-confined. It was concluded that the two wells were located in semi-confined portions of the Arbuckle-Simpson aquifer.

Signal analysis revealed that two wells (OWRB 86266 and OWRB 86267) water-level fluctuation were dominated by S2 and K1 tides. These two tides are the product of atmospheric pressure changes. The two wells are located in the southwest and southeast of the Hunton anticline respectively. The depth of the well OWRB 86266 is 34.1 m (112 ft) and that of OWRB/USGS 86267 is 23 m (75 ft). Water level within the two wells was monitored the barometric-pressure for two months. The atmospheric pressure was monitored at the site of the well OWRB 86267.

Results of the harmonic analyses of the water-level data for the well OWRB 86266 are shown in Figure 5.8. The figure reveals that the water level fluctuations influenced mainly by the barometric-pressure changes and no sign of other earth tides was present. Water-level fluctuations produce S2 and K1 harmonic components along with one unknown lower frequency component that occurs every 36 hours. The well OWRB 86266 is penetrating the Simpson Group of rocks which is composed sandstones, some limestones and shales. In particular, the well is penetrating a group of rocks that include three formations: the Bromide, Tulip Creek, and McLish. The geologic section available for the area indicates an unconfined aquifer conditions (Fairchild et al., 1990).

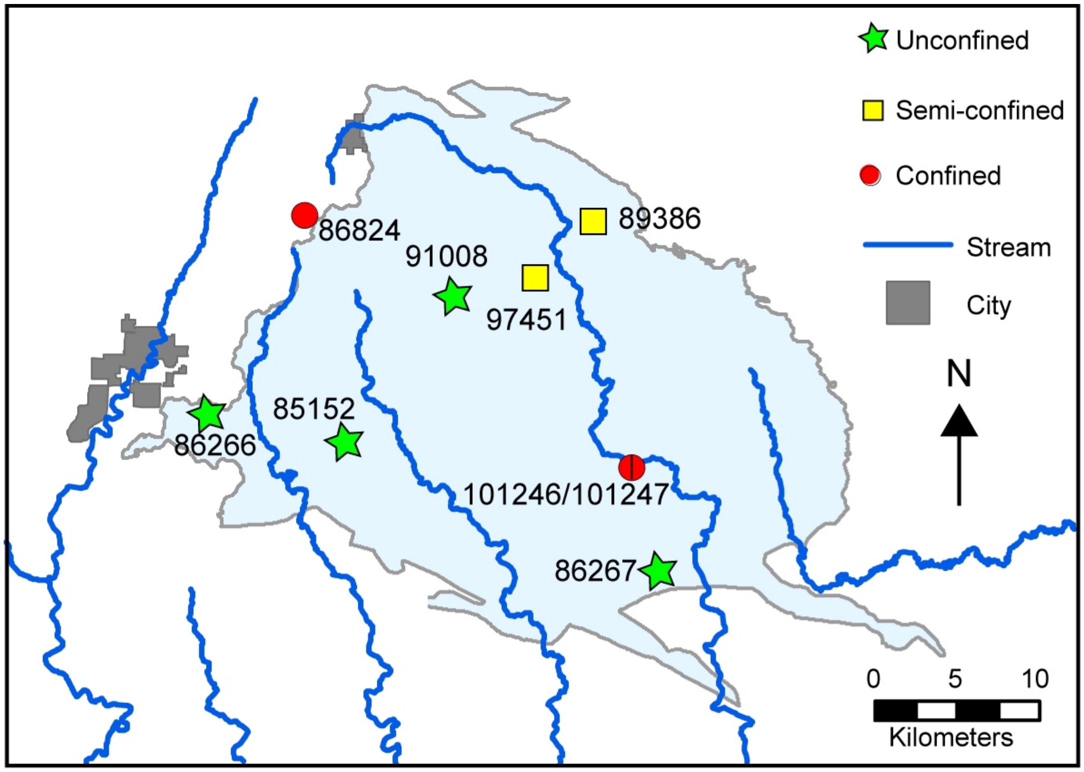

Figure 5.3. The Study area and well locations.

The area was the eastern part of the Arbuckle-Simpson aquifer. The wells symbolized an colored to reflect the aquifer type that was inferred from spectral analyses.

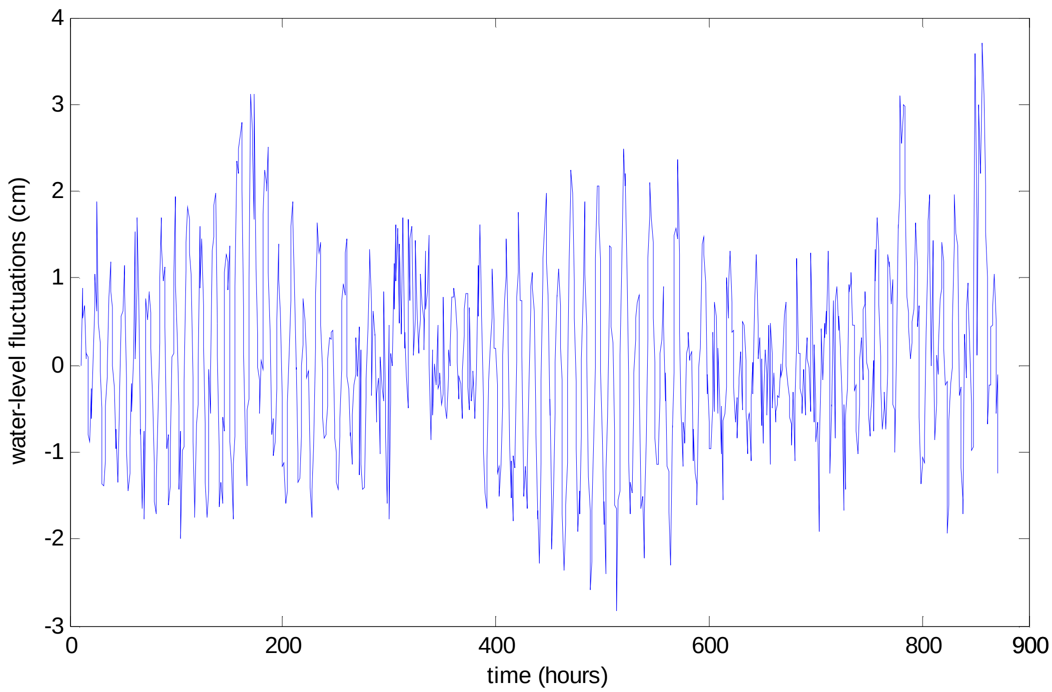

Figure 5.4. Observed water-level fluctuations at the well OWRB 101246 (Spears 1).

The data are for a period of one month and it shows the Spring and Neap tides. The well was located in the Arbuckle-Simpson aquifer.

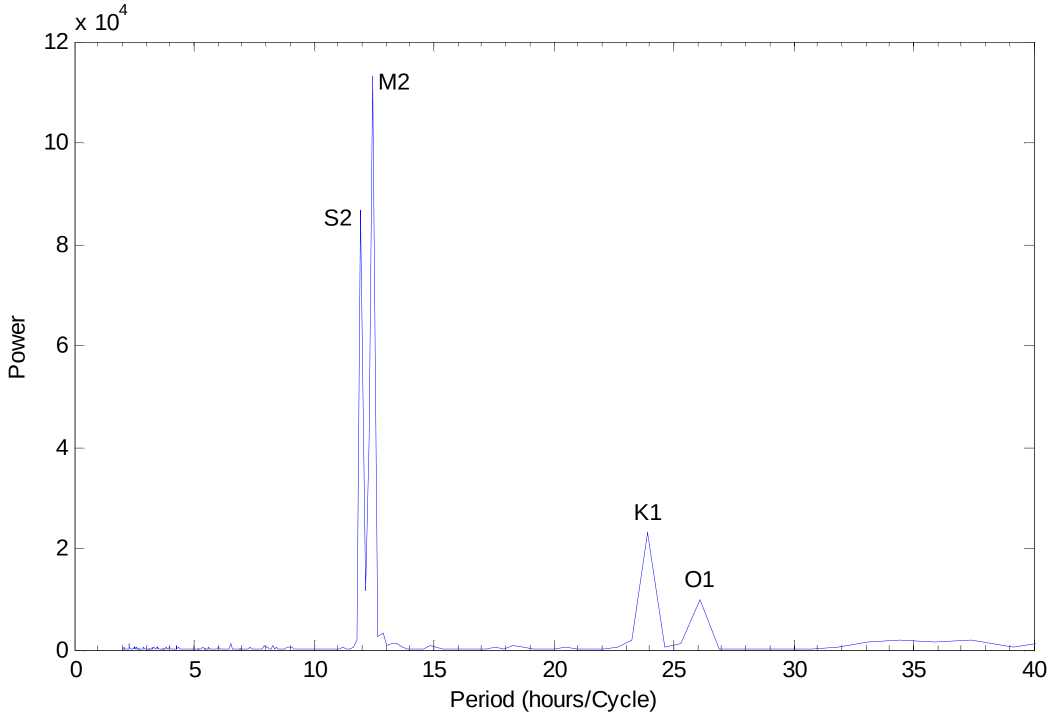

Figure 5.5. Power spectrum calculated from water-level fluctuations (well OWRB 101246 (Spears 1)).

The dominated tide is M2 (Lunar semi-diurnal). This is the characteristics of confined aquifer. The well was located in the Arbuckle-Simpson aquifer.

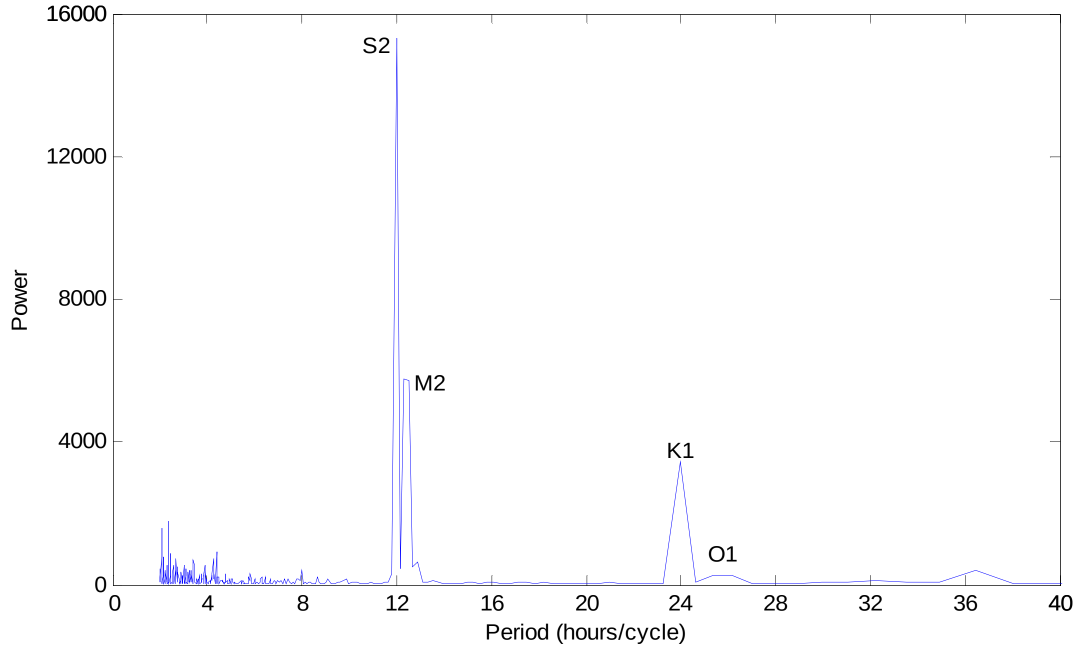

Figure 5.6. Power spectrum calculated from water-level fluctuations at the well OWRB 89386 (USGS Fittstown).

The dominant tide is S2, but M2 is also present. This is the characteristics of a semi-confined aquifer. The well was located in the Arbuckle-Simpson aquifer.

The periodogram of the well OWRB 86267 show the same results of the well OWRB 86266. The well OWRB 86267 is tapping Cool Creek and McKenzie Hill formations. The formations are part of the Arbuckle Group of rocks and they are mainly limestones. Since the well is relatively shallow (23 m (75 ft)), it is reasonable to assume that the well is tapping a local unconfined aquifer. Groundwater depth within the well increased from 4.29 m (14.08 ft) on 4/15/08 to 5.4 m (17.74 ft) on 5/20/08. Part of this large decline may be attributed to evapotranspiration losses. Loss of groundwater to evaporation is another indication that the well is penetrating an unconfined aquifer. The results of signal analyses of the wells OWRB 86266 and 86267 suggest that the two wells are tapping unconfined portions of the Arbuckle-Simpson aquifer.

Wells discussed so far indicate the presence of two or more tidal components within their water-level fluctuations. Other wells however revealed no tide signals within their water-level data. Figure 5.9 shows the spectral analysis of the well OWRB 85152. The well is located in the southwestern quarter of the studied area. The water-level data for this well revealed no tide signal within the known frequencies. The well total depth is 36 m (119 ft) and it is expected to tap an unconfined portion of the aquifer. The other well that revealed similar response as the well OWRB 85152 is OWRB91008. The depth of the well OWRB 91008 is 46 m (121 ft) and it is located in an area of dolomite and sandstones. Since the well is relatively shallow, it is more likely that it tapping an unconfined aquifer.

No explanation as to why these data did not show any tidal components can be given. However, it is not expected that all types of aquifers or water-bearing formation will response to earth tides or barometric pressure changes (Bredehoeft, 1967). If the upper boundary of an unconfined aquifer is in direct contact with atmosphere, it is not expected that water level within a well will be affected by atmospheric changes. Hence, it is reasonable to conclude that the wells OWRB 85152 and 91008 are tapping unconfined portions of the Arbuckle-Simpson aquifer.

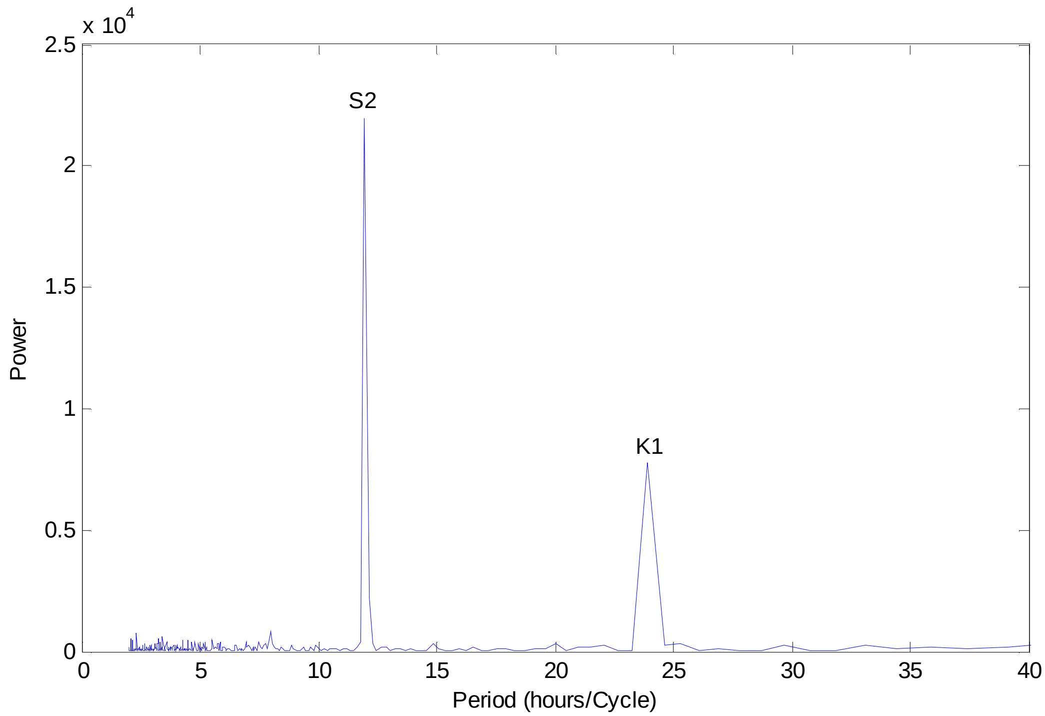

Figure 5.7. Power spectrum calculated from atmospheric pressure fluctuations (well OWRB 101246 (spears test 1)).

The atmospheric pressure spectrum includes S2 and K1 and shows no presence for M2 or O1. The well was located in the Arbuckle-Simpson aquifer.

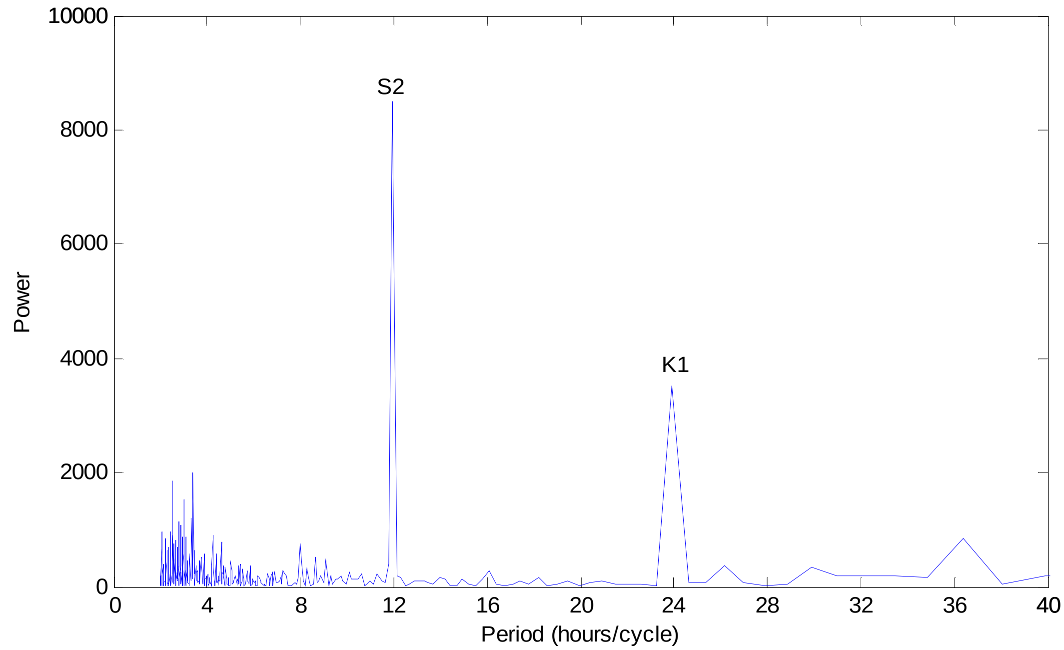

Figure 5.8. Power spectrum calculated from water-level fluctuations at the well OWRB 86266.

This spectrum shows only S2 and K1 and no M2 or O1. This is a characteristic of an unconfined aquifer. The well was located in the Arbuckle-Simpson aquifer

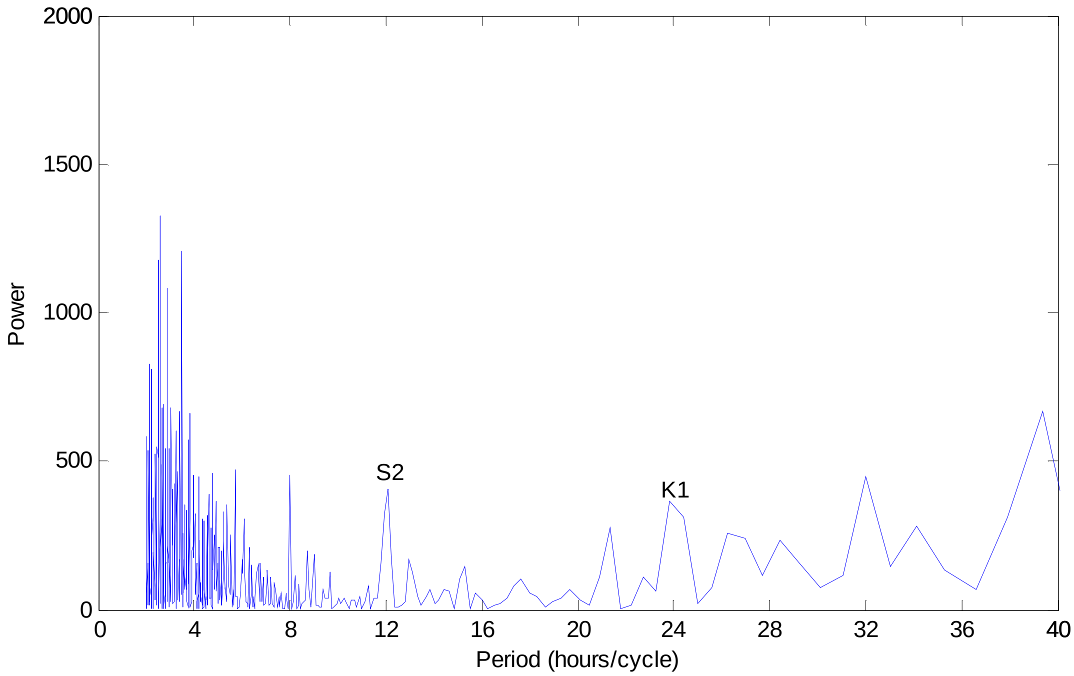

Figure 5.9. Power spectrum calculated from water-level fluctuations at the well OWRB 85152.

The well located in the Arbuckle-Simpson aquifer. No tide signal is clearly distinguished, but slight indication of S2 and K1. The well was unconfined.

Amplitude and Phase Angle Determinations

The least squares regression method was applied to the time series to determine the amplitude and the phase angle of the various tidal components. This type of harmonic analysis was meant to enhance the results of signal identification analysis that was covered in the previous section. Results obtained for the three artesian wells (OWRB 101246, OWRB 101247, and OWRB 86824) showed water-level fluctuation significantly influenced by earth tides, particularly the M2 tide. Results for the well OWRB 101246 are shown in Table 5.2 as an example of the confined portions of the aquifer. Table 5.2 shows that M2 tidal component has the highest percent of variance (61.2 percent). The percent of variance is a statistical test that signifies the influence of a given tide on water-level fluctuations. The relatively high percent of variance indicates that M2 tide dominated water-level fluctuations at this well (and the other two wells as well). Therefore, the conclusion that the wells OWRB 101246, OWRB 101247, and OWRB 86824 are tapping confined portions of the Arbuckle-Simpson aquifer is reemphasized.

Amplitude analysis of two wells (OWRB 97451 and OWRB 89386 USGS well) showed a strong presence of the tidal components S2 and K1. The analysis also revealed strong influence of the M2 tide. The sum of percent of variances is about 65 percent which comes from M2 and S2 (Table 5.3). The presence of S2 and M2 is considered a characteristic of the semi-confined aquifer. Hence the amplitude analysis agreed with the finding of the signal analysis. The amplitude analysis results confirmed that these two wells were located in semi-confined portions of the Arbuckle-Simpson aquifer.

Table 5-2. Results of the harmonic analyses for the well 101246 (Spears 1) 14/25-5/31/07.

The dominant tied is M2, an indication that the aquifer is confined. The well was located in the Arbuckle-Simpson aquifer.

Wave Component	Angular Frequency [rad/hr]	Amplitude [cm]	Phase [degrees]	Percent Variance [%]
O1	0.243352	0.05871	77.03	4.3
K1	0.262516	0.05343	-28.91	3.5
M2	**0.505868**	**0.22214**	**-18.25**	**61.2**
S2	0.523599	0.07838	116.88	7.6
N2	0.496367	0.04884	96.62	3

Table 5-3. Results of the harmonic analyses for the well OWRB 89386 (USGS) (04/15-05/20/08).

The domination of M2 and S2 tides suggest that the aquifer is semi-confined. The well was located in the Arbuckle-Simpson aquifer.

Wave Component	Angular Frequency. [rad/hr]	Amplitude [cm]	Phase [degrees]	Percent Variance [%]
O1	0.243352	0.01307	101.52	1.3
K1	0.262516	0.03379	-100.43	8.6
M2	**0.505868**	**0.06196**	**-8.66**	**29**
S2	**0.523599**	**0.05752**	**67.41**	**24.9**
N2	0.496367	0.0129	77.62	1.3

The well OWRB 86266 which penetrates the Simpson formation showed no signal of M2 or O1 tides within its water-level fluctuations. The results are shown in Table 5.4. The percent variance of M2 harmonic is zero which means that this tide had no influence on the water-level fluctuations. Table 5.4 shows a relatively strong influence on water-level fluctuations by the tides S2 and K1. The sum of percent of variances of all tidal components is 37.9 with most of the contribution comes from S2 and K1. Similar analyses were performed to the well OWRB 86267. Results of the analysis revealed (as was the case with the well OWRB 86266) no influence of M2 and O1 tides on water-level fluctuations. But the influence of the barometric-pressure changes as manifested by the presence of the tides K1 and S2 is clear. The finding of the amplitude analysis coincided with the signal analysis. Therefore, the earlier conclusion that these two wells are penetrating unconfined parts of the Arbuckle-Simpson aquifer is justified.

Results of amplitude analysis of the well OWRB 85152 are shown in Table 5.5. No tidal components, within the studied frequencies, can be identified from the table. The total percent of variances is about 2.5 percent. The low percent of variance is indicator that water-level fluctuations of this well were not influenced by any of the studied tides.

Table 5.6 summarizes the relationship between tides and type of aquifer. Water-level fluctuations in wells drilled in confined or semi-confined aquifers must show some type of tide signals. The difference is that M2 dominants confined aquifer signal, while S2 dominants semi-confined aquifers signals. Water level fluctuates within unconfined may or may not reveal tide signals. When signals are present within a well in unconfined aquifer, they are most likely S2 or K1. M2 and O1 are not anticipated to occur in wells of unconfined aquifers.

Table 5-4. Results of the harmonic analyses for the well OWRB 86266 (4/15-5/20/08). K1 and S2 tides dominated the spectrum, which is an indication that the aquifer is unconfined. The well was located in the Arbuckle-Simpson aquifer.

Wave Component	Angular Frequency [rad/hr]	Amplitude [cm]	Phase [degrees]	Percent Variance [%]
O1	0.243352	0.00847	-166.92	0.7
K1	**0.262516**	**0.03441**	**-141.31**	**11.2**
M2	0.505868	0.00076	-139.74	0
S2	**0.523599**	**0.05232**	**46.34**	**26**
N2	0.496367	0.00134	-20.09	0

Table 5-5. Results of the harmonic analyses for the well OWRB 85152. The well is located in the Arbuckle-Simpson aquifer (05/20-07/02/08). The M2 tide dominated the tidal spectrum, an indication of a confined aquifer.

Wave Component	Angular Frequency [rad/hr]	Amplitude [cm]	Phase [degrees]	Percent Variance [%]
O1	0.243352	0.00522	125.38	0.2
K1	0.262516	0.01064	-94.83	0.8
M2	**0.505868**	**0.002**	**159.33**	**0**
S2	0.523599	0.01439	62.02	1.4
N2	0.496367	0.00324	150.01	0.1

Table 5-6. Table 5.6 Types of aquifers and the expected response to various tides.

If var<3%, then the presence of the tide is very small and may be neglected. K1 and S2 tides may or may not be present in the case of unconfined aquifer, but M2 and O1 will not be present.

Wave component	**Confined**	**Semi-confined**	**unconfined**
O1	Present	Var< 2%	Var< 2%
K1	Present	Present	May be present
M2	Present and dominant	Present	Var< 2%
S2	Present	Present and dominant	May be present
N2	var< 2%	var< 2%	var< 2%

SUMMARY AND CONCLUSIONS

Aquifer type, confined, unconfined, or semi-confined may be identified by drilling or performing pumping tests. Both methods are costly, involve complex field issues, and may yield inconclusive results. Earth tides are known to influence water levels in wells penetrating confined aquifers or unconfined thick, low-porosity aquifers. Water-level fluctuations in wells tapping unconfined aquifers are influenced by changes in barometric pressure. Time-series analyses of water-level fluctuations of the Arbuckle-Simpson aquifer were utilized in nine wells to identify aquifer type by evaluating the influence of earth tides and barometric pressure variations.

The Arbuckle-Simpson is a thick (~1000 m) carbonate aquifer located in south-central Oklahoma. Previous studies categorize the aquifer as confined in some localities and semi-confined to unconfined in others. Harmonic analysis for water-level time series was employed to identify the aquifer types and their locations. The time series is the natural stresses-induced water-level fluctuations. The stresses are solid earth tides and barometric pressure changes.

The analyses involved the determination of signal strength and amplitude for each tidal component. Signal identification analyses preformed using Fourier transforms within the in MATLAB software, and amplitude and phase angle determination was done by least squares regression. Based on the results of the harmonic analyses, three types of the Arbuckle-Simpson aquifer were identified. The wells OWRB 101246, OWRB 101247, and OWRB 86824 were tapping confined portions of the aquifer. The wells are located in two locations: northwest and southeast of the Hunton Anticline. Wells OWRB 97451 and OWRB/USGS 89386 may be considered as tapping semi-confined portions of the aquifer. These two wells are located in the north western parts of the Hunton Anticline. The remaining wells (four of them) were tapping unconfined parts of the aquifer. The Arbuckle-Simpson aquifer was described as of a complex geology (Fairchild et al., 1990). More intensive spatial studies are needed to classify the aquifer thoroughly.

CHAPTER VI

IMPROVED METHOD FOR DETERMINING THE BAROMETRIC EFFICIENCY OF AQUIFERS

ABSTRACT

Barometric efficiency is an indication of the competence of a confining layer to resist atmospheric pressure changes. An important application of barometric efficiency is to determine the porosity of a confined aquifer if its specific storage is known. The barometric efficiency is commonly determined by the Clark method. The Clark method has been used for a number of aquifers and can give inconsistent and, in many instances, physically unrealizable results. A new method The Rahi method) is presented that is more consistent and overcomes the shortcomings of the Clark method. The Rahi method was tested using data from the Arbuckle-Simpson aquifer in Oklahoma, USA and evaluated against data from other research and produced consistent results within the expected range of the aquifers' barometric efficiencies.

INTRODUCTION

Barometric effects in aquifers result from stresses acting on aquifers due to changes in the atmospheric pressure. These changes are linked to periodic (diurnal and semidiurnal) and aperiodic atmospheric changes (Todd, 1959). The periodic changes are two lows, at early morning and early afternoon and two highs, at late morning and late afternoon. Clark (1967) suggested that the periodic changes can be approximated as atmospheric pressure lows at 4 am and pm and atmospheric highs at 10 am and pm. Merritt (2004) indicated that the timing of lows or highs is variable between 2-3 hours from day to day and it is difficult to decide the precise timing of these events. The aperiodic atmospheric-pressure changes are the result of long-term movements of air masses. Atmospheric perturbations change the external load on the aquifer resulting in changes in water levels within wells. This results in an inverse relationship between water-level fluctuations and barometric pressure changes with an increase in atmospheric pressure producing a decrease in water level.

The barometric efficiency (*BE*) is the ratio of the aquifer pressure head change to the atmospheric pressure change and was first introduced by Jacob (1940). Todd (1959) suggested that *BE* may be used as a measure of capability (competence) of overlying confining layer to resist pressure changes; the thicker the confining layer the higher the *BE*. The *BE* is utilized to compute porosity of confined aquifers if the storage coefficient is known, or to compute specific storage if the porosity is known. Clark (1967) developed a method to estimate *BE* based on aperiodic, long-term pressure variations resulting from the movement of air masses and the corresponding measured head changes in the well. Davis and Rasmussen (1993) extended the Clark method using a linear regression technique to determine the barometric efficiency and compared their approach with the Clark method. The authors stated that the Clark method provides an unbiased consistent estimate of the barometric efficiency when negative and positive changes in barometric pressure are equally likely. Davis and Rasmussen (1993) indicated that this conclusion holds for linear and nonlinear trends that may be present within the atmospheric pressure data. However, when unequal numbers of positive and negative changes are present in the data and the trend is linear, Davis and Rasmussen (1993) suggested the use of an iterative recursive technique to correct the estimated value of *BE*.

Two attempts at data filtering to improve the estimate of *BE* have been made. Rhoads and Robinson (1979) have filtered their water-level data and used only segments which indicate times when the barometric effects are dominant and other effects are small to determine the *BE*. Hobbs and Fourier (2000) calculated *BE* using an alternate method. The authors used a simple model based on isolating a water-level change for a given time interval and dividing it by the change in barometric pressure for the same time interval. Hobbs and Fourie (2000) monitored water level-fluctuations which were attributed to barometric effects with a period of 11.3 h, and found that the barometric efficiency of the aquifer averages 63 percent. However, some of their calculations revealed a barometric efficiency as high as 400 percent.

Several researchers have used the Clark method using field data and have found problems with obtaining realistic values of *BE*. Marine (1975) calculated porosity using *BE* values as computed by the Clark method and found that the computed porosity of "this slightly fractured crystalline aquifer...would (reach) 100 percent, an absurd value." The author concluded that "the porosity is very sensitive to barometric efficiency, which is extremely difficult to calculate for wells whose predominant water level fluctuations are caused by earth tides. Hsieh et al (1987) adapted a graphical procedure to estimate transmissivity using tidally-influenced water-level fluctuations. They indicated that for phase analysis, the concept of constant barometric efficiency, as determined by the Clark method, is not sufficient for removal of barometric effects. However, their analyses showed that only K1 (diurnal) and S2 (semidiurnal) tidal constituents are contaminated by barometric fluctuation. Hence Hsieh and others (1987) restricted their phase analysis to the M2 (semidiurnal) and O1 (diurnal) tidal component in order to limit the effect of the barometric pressure fluctuations on their transmissivity analysis. Merritt (2004) concluded that the Clark method of calculating the barometric efficiency can be effective when the head data are of high quality. However, the method can provide values that are too low when the head data are noisy or have strong trends. Merritt (2004) went on to state that the method "provided values that were too low in data sets that did not have obvious (data quality) problems..."

In general, research using field data has demonstrated that the Clark method provides a wide range of values of BE with some values that are physically unrealizable. The purpose of this paper is to examine the Clark method, identify its shortcomings, and introduce an improved algorithm designed to overcome those shortcomings. The Clark method and the Rahi method for calculating *BE* are presented. Both methods are compared for performance using short (one moon phase) and long (up to one year) data sets, comparison between wells in a single aquifer, and evaluation with a data set that is previously known to have difficulties using the Clark method. Field data for four wells are presented and the two methods are used to calculate *BE* in those wells.

GOVERNING EQUATIONS

The mathematical representation and derivation of the barometric efficiency and its relation to porosity and specific storage presented here. The derivation is based on Jacob, 1940; Batu, 1998; and Todd and Mays, 2005. Total stress on the top of a confined aquifer (σ_T) is balanced by the water pressure (p_w) and the compressive stress of the solid skeleton of the aquifer (σ). If db is the change in barometric pressure and dp_w is the change in hydrostatic pressure at the top of a confined aquifer, then (Batu, 1998; Todd and Mays, 2005):

$$\frac{db}{\gamma} = \frac{dp_w}{\gamma} + \frac{d\sigma}{\gamma} \tag{6.1}$$

The weight of the water column in the well above the upper boundary of the aquifer plus the change in atmospheric pressure is balanced by the hydrostatic pressure in the aquifer. Hence the change in the water level in the well is given by (Batu, 1998; Todd and Mays, 2005):

$$dh = \frac{dp_w}{\gamma} - \frac{db}{\gamma} \tag{6.2}$$

If Equation 6.2 is divided by Equation 6.1, keeping in mind that $dp_w - db = -d\sigma$, an expression for the BE is obtained (Batu, 1998):

$$\frac{dh}{\frac{db}{\gamma}} = -\frac{1}{\frac{dp_w}{d\sigma} + 1} \tag{6.3}$$

The first part of the denominator of the right hand side of Equation 6.3 is given by:

$$\frac{dp_w}{d\sigma} = \frac{\sigma}{\eta\beta} \tag{6.4}$$

When Equation 6.4 is substituted into Equation 6.3, the following equation is obtained (Batu, 1998):

$$BE = \frac{dh}{\frac{db}{\gamma}} = -\frac{1}{1+\frac{\alpha}{\eta\beta}} \tag{6.5}$$

The negative sign in equation 6.5 indicates that when atmospheric pressure increases, water level in a well in direct contact with the atmosphere decreases. The specific storage is given by (Jacob, 1940):

$$S_s = \gamma\eta(\beta + \frac{\alpha}{\eta}) \tag{6.6}$$

From Equations 6.5 and 6.6, the specific storage (S_s) is (Jacob, 1940):

$$S_s = \frac{\gamma\eta\beta}{BE} \tag{6.7}$$

Equation 6.7 defines the relationship among the *BE*, porosity, and specific storage of a confined aquifer.

BAROMETRIC EFFICIENCY DETERMINATION

The Clark method

The Clark method (1967) utilizes observed changes in barometric pressure, Δb, and hydraulic head, Δh, with constant time increments for determining *BE*. The method assigns a positive sign to the barometric pressure or the hydraulic head when they are rising. The scheme involves the computations of two sums: $\Sigma\Delta b$ and $\Sigma\Delta h$, according to the following rules (Davis and Rasmussen, 1993):

1) when Δb is zero, Δh is not added to $\Sigma\Delta h$.
2) when Δb and Δh have opposite signs, add the absolute value of Δh to $\Sigma\Delta h$.

3) when Δb and Δh have similar signs, subtract the absolute value of Δh from $\Sigma\Delta h$.

4) $\Sigma\Delta b$ is the sum of absolute values of Δb.

Starting from time step (t_i), concurrent sums (S) of the barometric pressure (b_i) and water pressure head (h_i) changes are computed using the algorithm developed by Merritt (2004):

$$\Delta b_i = b_i - b_{i-1} \tag{6.8}$$

$$\Delta h_i = h_i - h_{i-1} \tag{6.9}$$

$$index = \Delta b_i * \Delta h_i \tag{6.10}$$

$$S_b^i = S_b^{i-1} + \left|\Delta b_i\right| \tag{6.11}$$

$$S_h^i = S_h^{i-1} - \left|\Delta h_i\right| \text{ if } index > 0 \tag{6.12}$$

$$S_h^i = S_h^{i-1} + \left|\Delta h_i\right| \text{ if } index < 0\text{, and} \tag{6.13}$$

$$S_h^i = S_h^{i-1} \text{ if } index = 0 \tag{6.14}$$

The barometric efficiency is calculated by:

$$BE = \frac{\sum \Delta h}{\sum \Delta b} = \frac{S_h^n}{S_b^n} \tag{6.15}$$

The Rahi Method

Previous investigators (Marine, 1975; Hsieh et al., 1987; Hobbs and Fourie, 2000; Merritt 2004) have showed that the Clark method is inconsistent and may overestimate the barometric efficiency. Therefore, a new method (the Rahi method) that overcomes the deficiencies of the Clark method is presented here. The Rahi method computes the two adjusted sums of water-level changes ($\sum \Delta h_a$) and atmospheric pressure changes ($\sum \Delta b_a$) according to the following rules:

1) when Δb and Δh have opposite signs, and the absolute value of Δh is less than the absolute value of Δb, add the absolute value of Δh to $\Sigma\Delta h_a$ and the absolute value of Δb to $\Sigma\Delta b_a$;
2) Otherwise, Δh and Δb are not added to their respective sums.

Starting from time step (t_i), adjusted concurrent sums (AS) of the barometric pressure (b_i) and pressure head (h_i) changes are computed according the following scheme:

$$\Delta b_i = b_i - b_{i-1} \tag{6.16}$$

$$\Delta h_i = h_i - h_{i-1} \tag{6.17}$$

$$index = \Delta b_i * \Delta h_i \tag{6.18}$$

$$AS_b^i = AS_b^{i-1} + |\Delta b_i|\text{, if index <0 and } |\Delta h| \langle |\Delta b| \tag{6.19}$$

$$AS_b^i = AS_b^{i-1}\text{, otherwise} \tag{6.20}$$

$$AS_h^i = AS_h^{i-1} + |\Delta h_i|\text{, if index <0 and } |\Delta h| \langle |\Delta b| \tag{6.21}$$

$$AS_h^i = AS_h^{i-1}\text{, otherwise} \tag{6.22}$$

$$BE = \frac{\sum \Delta h_a}{\sum \Delta b_a} = \frac{AS_h^n}{AS_b^n} \tag{6.23}$$

The Rahi model differs from the Clark method in that it subjects the water-level data to two tests before adding it to the sum $\Sigma\Delta h_a$. The first test is the sign, and the second test is when the data point complies with the sign test it will be subjected to the magnitude test (Equation 6.21). In addition, the atmospheric pressure data are filtered by the Rahi model using the same filtering approach used for water-level data (Equation 6.19). The Clark method employs the sign test only (Equations 6.12 through 6.14) to filter the data.

METHODS

Two sets of water-level data were analyzed for the study. Water-level and barometric-pressure data from three wells in the Arbuckle-Simpson aquifer and one well in the Floridan aquifer were used in this study to evaluate the two methods for calculating the barometric efficiency. This allows a comparison between three wells in a single aquifer and an evaluation of well data used in previous research that was problematic for the Clark method (Merritt, 2004).

The Arbuckle-Simpson Aquifer

The study area for three wells was the Hunton Anticline of the Arbuckle-Simpson aquifer, south-central Oklahoma (Figure 1). The aquifer is composed of thick sequences of Late Cambrian and Early Ordovician dolomite and limestone. The aquifer saturated thickness is approximately 900 m (3000 ft) and its surface outcrop area is about 1300 km^2 (500 mi^2) (Fairchild et al., 1990). The Hunton Anticline covers the eastern part of the Arbuckle-Simpson aquifer and contains more than half of the total aquifer area.

Water-level fluctuations and barometric-pressure changes were monitored and recorded for a one year period (Jan 2007-Jan 2008) for two wells: OWRB 101246 and 101247 (Figure 1). Well OWRB 101247 has a data gap generated by high water levels during the summer period. The records before and after the data gap were treated as two separate records. Water-level data were measured for two months (Oct – Dec 2008) for the well OWRB 86824. Well OWRB 101246 is 183 m (600 ft) deep and well 101427 is 548 m (1800 ft) deep and they are located approximately 30 m (100 ft) apart. Well 86824 is 76 m (250 ft) deep and located about 21.4 km (13.4 mi) to the northwest of the other two wells (Figure 1). All wells demonstrate a confined aquifer tidal response (Rahi and Halihan, 2009).

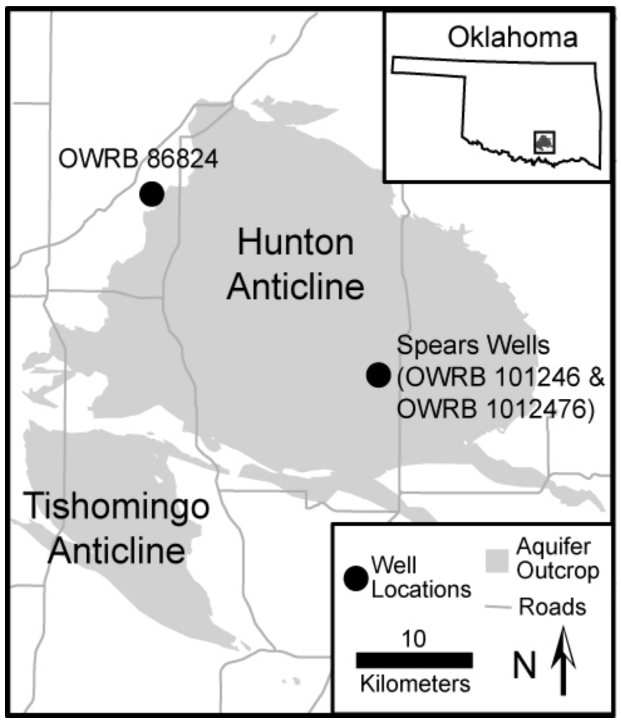

Figure 6.1. Location map for the Arbuckle-Simpson aquifer.

The aquifer includes three outcropped anticlines. This study was concentrated on the Hunton Anticline.

The Floridan Aquifer

Merritt (2004) studied the Floridan aquifer and reported unsatisfactory results when he applied the Clark method to a data set from a monitoring zone in well HE-1087 (1400-1810 ft below ground surface). Water-level at the well and atmospheric data nearby were collected between September 1998 and December 1999. Merritt (2004) calculated that this data have a *BE* of about 43 percent until mid-April of 1999 and a negative *BE* thereafter (Merritt, 2004). These well data are used as a comparison with a well where the Clark method has had difficulty.

Data Analysis

Data for the wells in the Arbuckle-Simpson aquifer were compensated for the barometric pressure using barometric pressure data collected at the locations of the wells. The data from well HE-1087 was collected using a vented transducer. Data for all the wells were then detrended using the following (Shumway and Stoffer, 2006):

$$y_t = x_t - x_{t-1} \tag{6.24}$$

where

y_t is the differenced head at time t, and

x_t is the measured head fluctuation at time t.

After detrending, the Clark and Rahi methods were applied to the data as described above. The two cumulative sums (water-level and atmospheric-pressure changes) were calculated based on equations 6.8 through 6.14, for the Clark, and 6.16 through 6.22 for the Rahi method. The cumulative sums of water-level changes were plotted against the cumulative sums of barometric-pressure changes. A linear best fit line was calculated for the data with the slope of the line being equal to the *BE*. The *BE* was calculated for the entire dataset, but additional estimations were performed by evaluating the *BE* for single moon phases of data (quarterly data, approximately 7 days) for the long period records available from wells OWRB 101246 and OWRB 101247.

Results

The Clark Method

BE as calculated by the Clark method for the well OWRB-101246 is shown in Figure 6.2. The slope of the line that was fitted to the data is 0.87 which indicates a *BE* of 87 percent. Average *BE* values for the three Arbuckle-Simpson wells, as determined by the Clark method range from 76 to 100 percent (Table 6.1). Table 6.1 also shows the periods of observation and the depth of each well for each well.

BE was calculated as a function of moon phases for wells OWRB 101246 and 101247. Rahi and Halihan (2009) reported *BE* values as high as 196 percent when the computation was made for individual moon phases. Figure 6.3 shows the results obtained for the well OWRB 101246. Similar results were obtained for the well OWRB 101247. The Clark method results in *BE* values between 9 and 196 percent and an average of about 91 percent. The highest barometric efficiencies calculated were associated with the last quarter and the second highest are associated with the first quarter. During these times, the moon is perpendicular relative to the earth-sun axis. In this position the moon and the sun tides partially cancel each other out. The influence of the M2 tidal response is small during the first and last quarters and water-level fluctuations are dominated by S2 and K1 tides. *BE* associated with full and new moons are more consistent and never exceeded 100 percent. The average *BE* for these two phases was 54 percent for the two wells, while the average for the first and last quarters was 126 percent.

For the well from the Floridan aquifer, Merritt (2004) calculated that these data have *BE* of about 43 percent until mid-April of 1999 and a negative *BE* thereafter (Merritt, 2004) using the Clark method. Merritt suggested a value of 75 percent by trial and error. The data from Merritt (2004) were reanalyzed as part of this study. The results were not conclusive. Two best fit linear lines were tried. When the intercept was fixed at zero the best fit line gave *BE* of 26 percent but negative R^2. The other best fit line was determined without fixing the intercept at zero. This line gave *BE* of 9 percent with R^2 of 16 percent. Thus the Clark method failed to produce realistic BE value in this study as was the case with Merritt (2004) (Figure 6.4). The Clark method produce a negative slope for the late portion of the data as was reported by Merritt (2004).

Table 6-1. Barometric efficiency results for the Arbuckle-Simpson.

The values are the average of time period for each well. Values of shorter periods varied considerably as shown in Figure 6.3.

Well Name	**Period of Observation.**	**Well depth (m/ft)**	**Barometric efficiency %**	
			Clark method	**Rahi method**
OWRB 86824	10/08-12/08	76.2/250	86	50
OWRB 101246	01/07-01/08	183/600	87	54
OWRB 101247	7/07-11/07	549/1800	100	59
OWRB 101247	11/07-2/08	549/1800	74	56
HE-1087	09/98-12/99	552/1810	9?	56

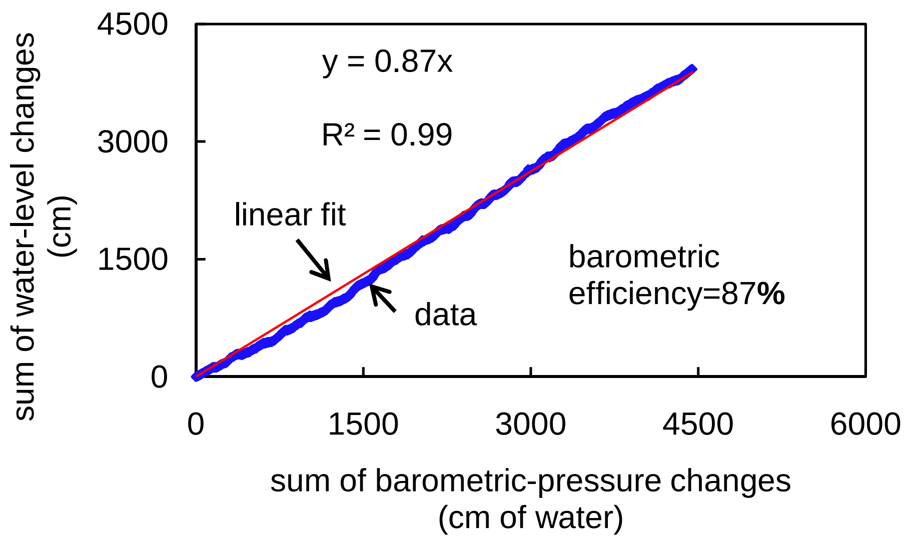

Figure 6.2. Barometric efficiency by the Clark method for the well OWRB-101246. The well is located in the Arbuckle-Simpson aquifer.

The Rahi Method

The barometric efficiency was recalculated for the three Arbuckle-Simpson wells using the Rahi model (Equations 6.16 through 6.23). Figure 6.5 show the results of the well 101246 which is representative for the other two Arbuckle-Simpson wells. The slope of the linear fit line is 0.54 which translates to a *BE* of 54 percent. The rest of the results are shown in Table 1. The *BE* as calculated by the Rahi model ranged from 50 to 59 percent. The average *BE* for the three wells is 55 percent.

Figure 6.3 shows *BE* as function of the moon phase for the well OWRB 101246. Results obtained from the well OWRB 101247 were similar. *BE* values ranged from 31 to 75 percent with an average value of 54 percent. The results for individual moon phases do not vary from those of longer periods. For the well in the Floridan aquifer, the *BE* was found to be 56 percent when computed by the Rahi model (Figure 6.6). The Rahi model R^2 value was 0.99.

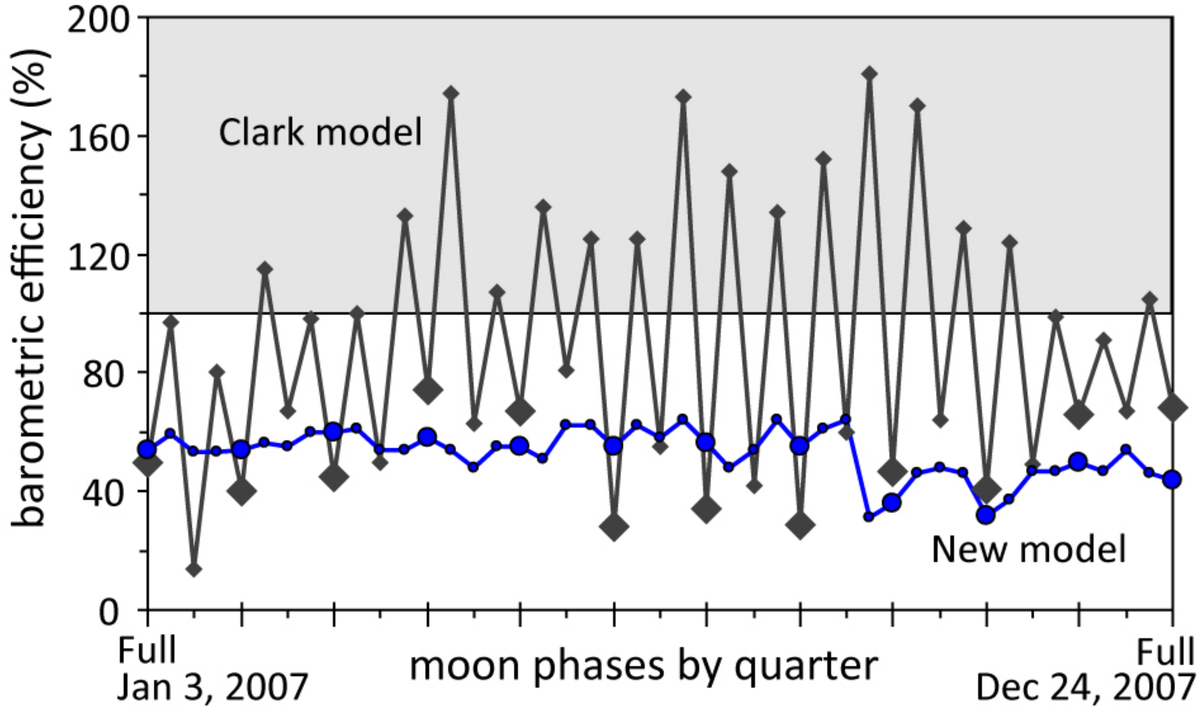

Figure 6.3. Barometric efficiency for the well OWRB 101246 as computed by the Clark and the Rahi models. Points for the full moon are enlarged to make the full moon quarter easier to observe. The gray area indicates barometric efficiencies that are physically unrealizable.

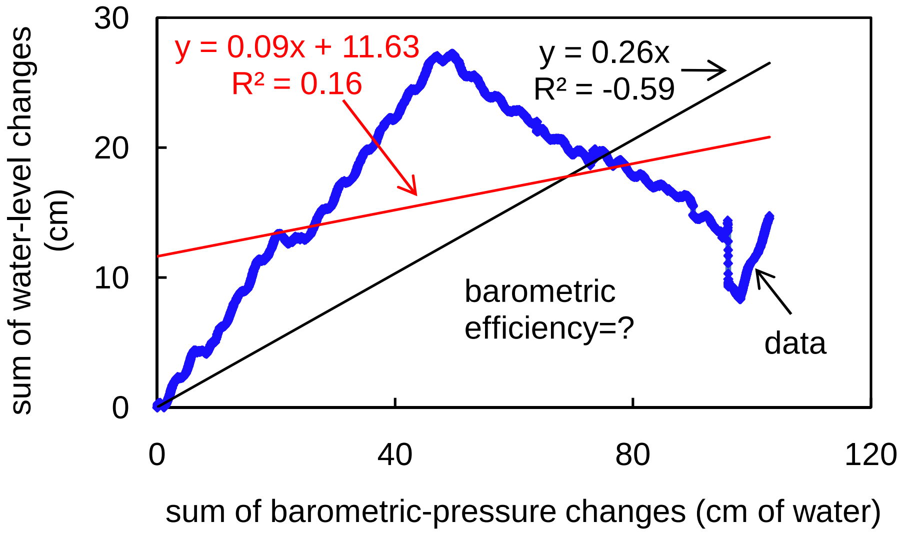

Figure 6.4. Barometric efficiency calculated by the Clark method using data from the well HE-1087 (Merritt, 2004).

The Clark method failed to determine the barometric efficiency. The same thing happened for Merritt (2004) analysis, so he implemented a trail and error approach. The *BE* value obtained by the trial and error was 75% (Merritt, 2004).

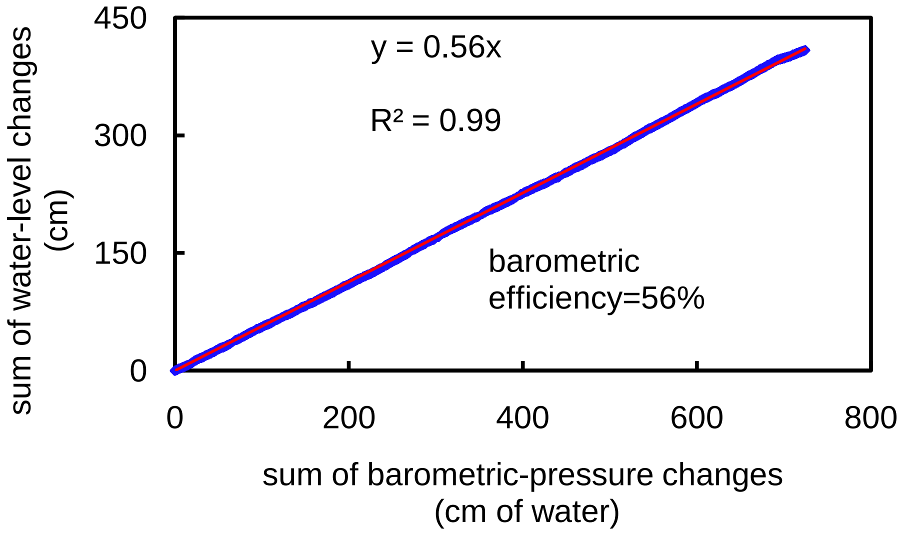

Figure 6.5. Barometric efficiency by the Rahi method for the well OWRB-101246. The barometric efficiency is the slope of the best fit line. The regression line and the data are too close to each other so could not be identified by pointing to each one separately.

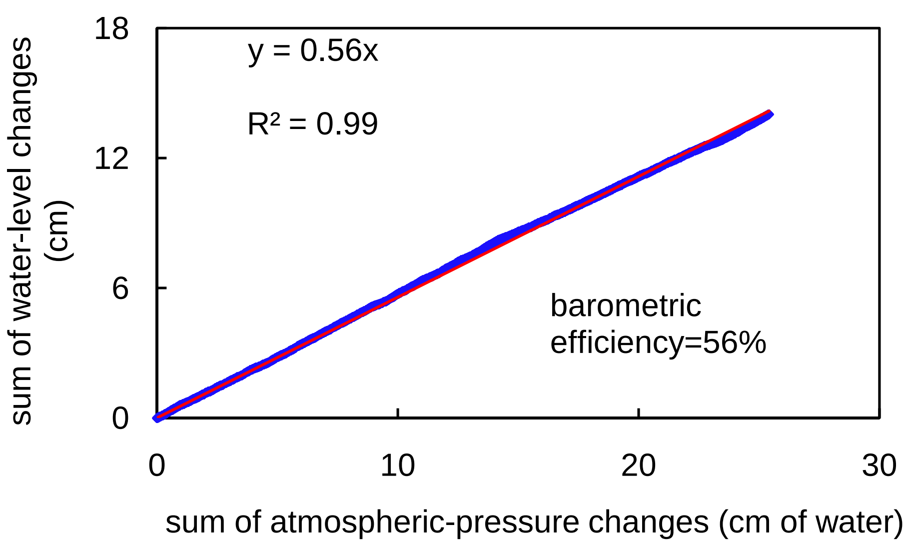

Figure 6.6. Barometric efficiency by the Rahi method using data from the well HE-1087 (Merritt, 2004). Merritt (2004) obtained a *BE* value of 75% by trial and error, because the Cark method failed to produce a reasonable value.

DISCUSSION

The Clark method algorithm was designed to add any water-level changes associated with tides that are on phase with barometric-pressure fluctuations. The method subtracts water-level changes that were produced by tides that are out of phase with barometric-pressure changes. S2 and K1 tidal components are in-phase with the atmospheric-pressure changes most of the time. The M2 and O1 tidal components alternate between in-phase and out-of-phase continuously. Clark (1967) indicated that for long records the in-phase tides balance the out-of-phase. For short records however, either the in-phase or the out-of-phase will dominate (Figure 6.3). Therefore, the Clark method often fails to determine *BE* for short records of water-level data, and can accumulate errors on longer records.

Further examination of the mathematics of the Clark method is helpful to understand why it can fail. The sum of water-level changes ($\Sigma\Delta h$) is actually two components (assuming data were detrended):

$$\sum \Delta h = \sum \Delta h_t + BE \sum \Delta b \tag{6.25}$$

where

$\sum \Delta h_t$ is sum of changes in water level changes that resulted from earth-tides influence.

Dividing Equation 6.25 by ($\Sigma\Delta b$) gives:

$$BE = \sum \frac{\Delta h - \Delta h_t}{\Delta b} \tag{6.26}$$

The Clark method assumes the second term in the numerator of the right hand side of the Equation 6.26 equals zero (Davis and Rasmussen, 1993). However, water-level fluctuations induced by earth tides are not zero. Accurate *BE* results cannot be obtained unless tidally-influenced water-level fluctuation are removed. The Clark method is not efficient in removing earth-tides effect, hence it overestimates the *BE*. Moreover, Clark (1967) stated

> "*A plot of $\Sigma\Delta W$ (sum of water-level changes) against $\Sigma\Delta B$ (sum of atmospheric-pressure changes), however, will not always be a straight line. Whenever the fluctuations of the water-level resulting from influences other than atmospheric pressure are in phase with atmospheric pressure, the fluctuations will be reinforced; and whenever they are out of phase, they will be dumped. Unless these other fluctuations are somehow closely related to the atmospheric pressure, it is doubtful that they would have the same period.*"

Clark (1967) went on to suggest that a large number of water-level incremental changes are needed so that the number of in-phase will statistically balance the number of out-of-phase. Clark's (1967) assertion for large amounts of data explains the failure of the Clark method to produce consistent *BE* values when applied to individual moon phases (Figure 6.3). Furthermore, water-level fluctuations are the product of a complex combination of tides, atmospheric pressure changes, and other natural influences. Therefore, it is highly unlikely that the in-phase number and magnitude will balance those of the out-of-phase. As a result, the Clark method may overestimate, underestimate, or even produce negative *BE* depending on the dominant influence; the in-phase or the out-of-phase.

Marine (1975) indicated that *BE* may be overestimated when computed using the Clark method. Results obtained in this study showed that the Clark method produced *BE* values of more than 100 percent in many instances (Figure 6.3). A value of 100 percent is impossible for natural aquifers. A 100 percent *BE* implies that the aquifer is rigid and the skeleton compressibility is zero and that the tidal efficiency is zero. Merritt (2004) reported inconsistency in his calculations of *BE* using the Clark method. Merritt's shows values of *BE* ranging from zero to about 90 percent. When Merritt's data were used for this study, the Rahi method resulted in *BE* values close to those obtained for the Arbuckle-Simpson aquifer. The results are consistent since both aquifers are composed of thick mainly carbonate formations. Merritt (2004) stated that the Clark method does not give acceptable results in several instances. In fact, the author's demonstrates the Clark method generating a negative *BE*. Hobbs and Flourie (2000) calculated values of *BE* using the Clark method can range from 100-400 percent due to the influence of earth tides on water level.

BE values obtained by the Rahi method were consistent and did not vary greatly according to the phase of the moon. The Rahi model filters the water-level as well as the barometric pressure data to remove the effects of the earth tides from the water-level fluctuations. Hence, the Rahi method attempts to include only the water-level fluctuations generated by barometric pressure changes and the barometric fluctuations that produce them for the calculation of the *BE*. Rhoads and Robinson (1979) have filtered their water-level data and used only segments which indicate times when the barometric effects are dominant and other effects are small to determine the *BE*. The same approach was followed by Hobbs and Fourie (2000). These approaches provide a similar logic to the Rahi method, besides, the method is a systematic algorithm that can be applied to a wide range of datasets.

The Rahi model produced consistent *BE* values for all the time intervals (Figures 6.3). The model produced no negative or other unrealistic values. The *BE* values ranged from 31 to 75 percent for the Arbuckle-Simpson aquifer (Figure 6 3). Accordingly, the short recording periods have no significant influence on the *BE* results as was computed by the Rahi model. When comparing the results of the two models, the Clark method as applied to water level fluctuations during the full and new moons produces results that were closer to the results obtained by the Rahi model (Figures 6.3). The Rahi method applied to data from two geographically different aquifers and show consistency and reasonable results. The method should be tested using data from aquifers that are expected to provide different values of *BE*.

SUMMARY AND CONCLUSIONS

The *BE* is the ratio of the aquifer pressure head change to the atmospheric pressure change. The *BE* can be used as an index of the capability (competence) of overlying confining layer to resist pressure changes; the thicker the confining layer the higher the *BE*. Another important application of the *BE* is that it can be used to calculate the aquifer porosity or the specific storage when one of the quantities is known. Commonly, *BE* is determined by the Clark method. The results of this study and other previous studies indicate that the Clark method may produce unrealizable and inconsistent *BE* values. A new method (designated the Rahi Method) that overcomes the Clark method problems and produces consistent, physically realizable results was introduced. The Rahi model was applied to three wells within the Arbuckle-Simpson aquifer and produced *BE* values ranging from 31 to 75 percent with an average of 54 percent, whereas the Clark method gave a range of 9 to 196 percent with an average of 91 percent for the same wells. Additionally, the *BE* was calculated by the two methods for a well in the Floridan aquifer using data from Merritt (2004). The Rahi method gave satisfactory results (56 percent), while the Clark method failed to produce acceptable results (it gave BE of 9%) and produced a negative slope for portion of the data.

CHAPTER VII

SPECIFIC STORAGE AND POROSITY OF THE ARBUCKLE-SIMPSON AQUIFER

ABSTRACT

Water yielding capacity of a confined aquifer is represented by its storage coefficient. The storage coefficient and porosity are the essential hydraulic parameters required as input for groundwater management and flow models. The standard method for determining the aquifer storage coefficient is the pumping test or aquifer test. Pumping tests are costly and site specific. Mechanical forcing of the aquifer by ocean tides, earth tides, and pressure fluctuations in the atmosphere stress the aquifer and cause water level fluctuations. The well water-level fluctuations were analyzed to determine the specific storage and porosity of the Arbuckle-Simpson aquifer. Data from seven wells of variable depths and variable time intervals were employed for the analyses. Harmonic analysis utilized to identify the tides that were present in the data. Harmonic analysis also employed to determine the amplitude and phase of each harmonic component. Resulted amplitude was utilized to determine the specific storage. The porosity was determined using the barometric pressure changes and the specific storage data. The average value of the specific storage was 5.6E-8 cm^{-1} (1.7E-6 ft^{-1}). The storage coefficient for the aquifer was 6.3E-3 which is comparable to values obtained by previous researchers. The porosity was approximately 15 percent.

INTRODUCTION

Groundwater studies usually involve the use of either an analytical or a numerical approach to solve a particular problem. Groundwater numerical models have increasingly proved their value in analyzing and evaluating groundwater systems. Todd and Mays (2006) indicated that last three decades have resulted in tremendous changes in the employment of computers for groundwater management. The numerical model requires the assignment of a discrete value of the hydraulic parameters to each cell or block in the model (flow) domain. The discrete values are designated as the model hydraulic parameters. For a groundwater model to produce reasonable predictions, the model parameters should accurately represent the physical aquifer system.

The aquifer hydraulic parameters that are needed to characterize the flow domain include transmissivity (T), storage coefficient (S), and porosity (η). The concentration of this paper is on S and η. Traditionally, these parameters are determined by field or laboratory methods such us pumping and laboratory core tests. Pumping tests might be problematic especially when the aquifer water quality or aquifer pollution is an issue (Mehnert, 1998). Pumping tests cannot be conducted when the aquifer system is undergoing a period of baseline monitoring during which time human interference is prohibited (Ritzi, 1989). Furthermore, pumping tests are in general expensive to conduct, but more expensive when conducted for S determination. A pumping test meant to determine T can be conducted within the pumping well only, while a pumping test for determining S needs at least one observation well in addition to the pumping well. Laboratory methods are of limited usefulness and hard to generalize since they represent disturbed core samples of small size. An alternative option to determine the aquifer hydraulic parameters is warranted.

Aquifers are subjected to stresses from natural processes such as mechanical forcing of the aquifer by ocean and earth tides and/or atmospheric pressure loading (Todd, 1959; Walton, 1970; Freeze and Cherry, 1979; Domenico and Schwartz, 1990). Earth and ocean tides are the product of lunar and solar tidal (gravitational) forces. Changes in barometric pressure are induced by variations in temperature and circulation. Water levels in monitoring wells often reflect of earth tides and barometric pressure changes. The effect of earth tides can be observed in wells tapping confined aquifers (Bredehoeft, 1967) and in wells tapping deep, relatively stiff and low-porosity unconfined aquifers (Weeks, 1979: Rojstaczer and Agnew, 1989). Earth tides are not known to influence water-level fluctuations in wells tapping unconfined aquifer. Weeks (1979) concluded that water levels in a well tapping an unconfined aquifer are influenced by changes in barometric pressure but the mechanism is “substantially different” from that of the confined aquifer.

Earth tides cause a periodical dilatation of the aquifer materials. A dilatation produces proportional changes in the aquifer formation stress, which is balanced by an increase or decrease in pore fluid pressure. As the pore fluid pressure changes, a difference in pressure is introduced between the well bore storage and the aquifer, which result in flow into or out of the well (Lambert, 1940). Barometric effects on aquifers result from periodic stresses acting on the aquifer due to changes in atmospheric pressure. Atmospheric-pressure fluctuations data reveal two lows at early morning and early afternoon and two highs at late morning and late afternoon. Clark (1967) suggested that atmospheric pressure lows occur at 4 am and pm and atmospheric highs occur at 10 am and pm. Merritt (2004) indicated that the timing of lows or highs varies between 2-3 hours from day to day and it is difficult to determine the precise timing of these events.

This paper addresses the use of natural stresses to determine aquifer specific storage as index to the aquifer storativity, and porosity. The study comprised field investigations and time series analyses. Water levels and atmospheric pressure were recorded over a periods of two months for some wells to one year for others.. Harmonic analyses were utilized to decompose the water-level and atmospheric-pressure data into its harmonic (tidal) components. Harmonic analyses were also used to resolve each tidal component into its amplitude and phase angle. The amplitude and the phase angle are functions of the aquifer hydraulic properties, among other parameters. The amplitude and phase angle are applied to the computation of the specific storage (Bredehoeft, 1967; Merritt, 2004). Barometric-pressure fluctuations were analyzed along with water-level time series to determine the barometric efficiency (*BE*) to determine the porosity.

GOVERNING EQUATIONS

The Specific Storage

The water level in an open well tapping a confined aquifer responds to pressure head disturbances caused by natural stresses. It fluctuates in response to earth tides and barometric-pressure changes. The degree to which water level fluctuates in response to these stresses is determined by the well dimensions, the transmissivity, storage coefficient, and porosity of the aquifer. Bredehoeft (1967) presented the following equation which relates the change in head in a well produced by tidal dilatation (Δ_t) to S_s:

$$-dh = \frac{\Delta_t}{S_s} \tag{7.1}$$

where

$$\Delta_t = \left(\frac{1-2\nu}{1-\nu}\right)\left[\left(2\bar{h} - 6\bar{l}\right)\frac{W_2}{ag}\right] \quad , \tag{7.2}$$

ν is the Poisson ratio (≈0.25, Bredehoeft, (1967)) of the aquifer material (dimensionless),

$\bar{h}$ and $\bar{l}$ are Love numbers at the surface of the earth (dimensionless),

W_2 is the tide potential,

a is the radius of the earth (L), and

g is the gravitational constant (L/T^2).

Combining equations 7.1 and 7.2 and rearranging we obtain an expression for the specific storage (S_s) of the aquifer:

$$S_s = -\left[\left(\frac{1-2\nu}{1-\nu}\right)\left(\frac{2\bar{h} - 6\bar{l}}{ag}\right)\right]\frac{dW_2}{dh} \tag{7.3}$$

The terms W_2 and ag in equation 7.3 have units of L^2/T^2. Therefore, S_s in Equation 7.3 has units of L^{-1}.

The minus sign in equation 7.3 signifies the head (pressure) in the aquifer decreases as the tide-generating potential increases. In other words, a pull caused by the transiting moon, for example, would expand the aquifer materials, hence reducing the pore pressure.

The tide potential (W_2) is determined from the equation:

$$W_2(\varepsilon, \phi, t) = gK_m bf(\theta)\cos[\beta(\varepsilon, t)] \quad (7.4)$$

where

K_m is the general lunar coefficient, taking into account the masses of the earth and moon, the distance to the moon, and the earth's radius, it is equal to 53.7 cm (1.7618 ft),

b is an amplitude factor (dimensionless) that has a distinct value for each tidal component with a period τ,

$f(\theta)$ is the latitude function (dimensionless); and

$\beta(\varepsilon,t)$ is a phase term that depends on the longitude ε and the Greenwich Mean Time (GMT) t.

Tide potential includes an infinite number of tidal components. However, five tidal components are important for geophysical applications (Bredehoeft, 1967). These components constitute 95% of the tide potential. These five are: M2 , a lunar wave with a period of 12h 25m 14s; S2, a solar wave with a period of 12h 00m; N2, a lunar wave with a period of 12h 39m 30s; K1, a luni-solar wave with a period of 23h 56m 4s; and O1, a lunar wave with a period of 25h 49m 10s. Young's (1913) harmonic analyses also identified five tidal components. However, Young's analyses did not reveal N2, instead revealed P1; a solar diurnal wave with a period of 24h 4 m (Bredehoeft, 1967; Merritt, 2004).

Merritt (2004) substituted the ratio of the tidal potential to changes in water level of Equation 7.3 by the ratio of their amplitudes. Merritt (2004) gave presented the following approximation for Equation 7.3:

$$S_s = 0.788 \times 10^{-12} \left(\frac{cm^2}{s^2} \right)^{-1} \frac{A_{w2}}{A_h} \tag{7.5}$$

where

A_{w2} is the amplitude of a harmonic component of W_2 and period τ.

A_{w2} is given by:

$$A_{w2} = gK_m bf(\theta) \tag{7.6}$$

and,

A_h is the amplitude of a component of the head change of period τ. The other terms have been defined earlier. The dimensionless terms of b, $f(\theta)$, and β (ε, t) were given by Merritt (2004), who correlated the work of Munck and McDonalds (1960) and Doodson and Warburg (1941) and presented it in a form useful for hydrologists. Merritt (2004) Tables 4 and 7 are combined and presented for this study as Table 7.1 for five tides. All the terms to necessary to compute A_{w2} are known, and A_h can be determined from harmonic analysis of water-level fluctuations.

Porosity

The porosity of a confined aquifer can be estimated if the S_s and *BE* of the aquifer are known. Jacob (1940) presented the following equation for the relationship between porosity, *BE*, and specific storage:

$$\eta = \frac{BE * S_s}{\rho g \beta} \tag{7.7}$$

where β is compressibility of water.

BE is given by (Jacob 1940):

$$BE = \frac{\rho g \Delta h}{db} \tag{7.8}$$

where *db* is the change in barometric pressure.

BE traditionally has been calculated by the Clark method (Clark, 1967). Rahi and Halihan (2009) examined the Clark method and found it inconsistent and may give *BE* values of higher than 100 percent. Rahi and Halihan (2009) suggested a new algorithm to calculate *BE* that is self-consistent and overcomes the shortcomings of the Clark method. The new method is designated the Rahi Method n this study. The Rahi method was used to calculate *BE* for this study.

STUDY AREA AND STUDIED WELLS

The Arbuckle-Simpson aquifer is located in south-central Oklahoma within the Arbuckle Mountain Physiographic Region (Figure 7.1). The Arbuckle Mountain region includes three main anticlines: the Arbuckle, Tishomingo, and Hunton. The aquifer is hosted in two rock groups, the Arbuckle and the Simpson. Each group is composed of several formations that may differ in their water-yielding capacity. Fairchild et al. (1990) disregarded these differences and treated the aquifer as composed of two lithological units, the Arbuckle and the Simpson. The surface area of the aquifer is about 1280 km^2 (500 mi^2). The estimated thickness of the Arbuckle Group between 1220 and 2040 m (4000 and 6700 ft), while that of the Simpson Group between 305 and 700 m (1000 and 2300 ft) (Fairchild et al., 1990).

Table 7-1. Parameters of the main diurnal and semidiurnal equilibrium tides.

These five tides constitute about 95% of the tide potential and they are of importance for groundwater studies.

Tidal Component	Angular frequency (rad/h)	Frequency (cycles/day)	Period (h)	Amplitude factor (b)	$f(\theta)$	$\beta\ (\phi, t)$
O1	0.24335189	0.92953573	25.819341	0.377	$\sin\theta\cos\theta$	$qt + \phi_s(t) - 2\phi_m(t) - 169.8^{\circ} + \phi$
K1	0.26251618	1.00273793	23.934469	0.531	$\sin\theta\cos\theta$	$qt + \phi_s(t) - 10.2^{\circ} + \phi$
N2	0.49636693	1.89598200	12.658348	0.174	$0.5\cos^2\theta$	$2(qt + \phi_s(t) - 1.5\phi_m(t) + 0.5\phi_p(t) - 79.8^{\circ} + \phi)$
M2	0.50586802	1.93227349	12.420602	0.908	$0.5\cos^2\theta$	$2(qt + \phi_s(t) - \phi_m(t) - 79.8^{\circ} + \phi)$
S2	0.52359878	2.0000000	12.000000	0.423	$0.5\cos^2\theta$	$2(qt + \phi)$

Symbols: θ, latitude; q angular velocity of the earth relative to the mean sun (15 degrees per mean solar hour); $\phi_s(t)$, longitude of the mean sun (increasing by 0.0411degrees per mean solar hour); $\phi_m(t)$, mean longitude of the moon (increasing by 0.549 degrees per mean solar hour); $\phi_p(t)$, mean longitude of lunar perigee (increasing by 0.0046 degrees per mean solar hour); and ϕ, longitude of the observation point.

Rocks of the Arbuckle Group are mainly middle Cambrian to early Ordovician limestone and dolomite (Puckette et al., 2009; Fairchild et al., 1990). Sargent (1969) indicated that the rocks of the Arbuckle Group of the Hunton anticline are mainly dolomites and thinner than the Arbuckle Group of the Tishomingo anticline which are mainly limestones. Rocks of the Simpson Group include, primarily, sandstone and shale with some middle Ordovician carbonates (Puckette et al., 2009; Fairchild et al., 1990).

The groundwater regime in the Arbuckle-Simpson aquifer is affected by the complex geologic features of the aquifer. Fairchild et al. (1990) suggested that occurrence and movement of groundwater in the aquifer are strongly controlled by lithology and structure. The aquifer is a carbonates-rock aquifer exhibiting karst features, especially in the western parts. Features such as folds, faults, fractures, and solution channels control groundwater flow rates and movements (OWRB, 2003). Fairchild et al. (1990) suggested that the association of springs with faults in the Arbuckle-Simpson aquifer indicates the strong structural control on groundwater movement. The general flow direction in the Hunton Anticline is from northwest to southeast (Fairchild et al., 1990; OWRB, 2005). A groundwater contours map published Fairchild et al. (1990) suggest that recharge occurs in the northwestern part of the aquifer, while discharge occurs within the southeast.

Water-level data from nine wells were used for this study. The water-level time series were subjected to harmonic analysis to verify the presence of tides. The five studied tides are listed in Table 7.1. The harmonic analysis results showed that tidal components were present in seven wells. The seven wells and some of their characteristics are listed in Table 7.2.

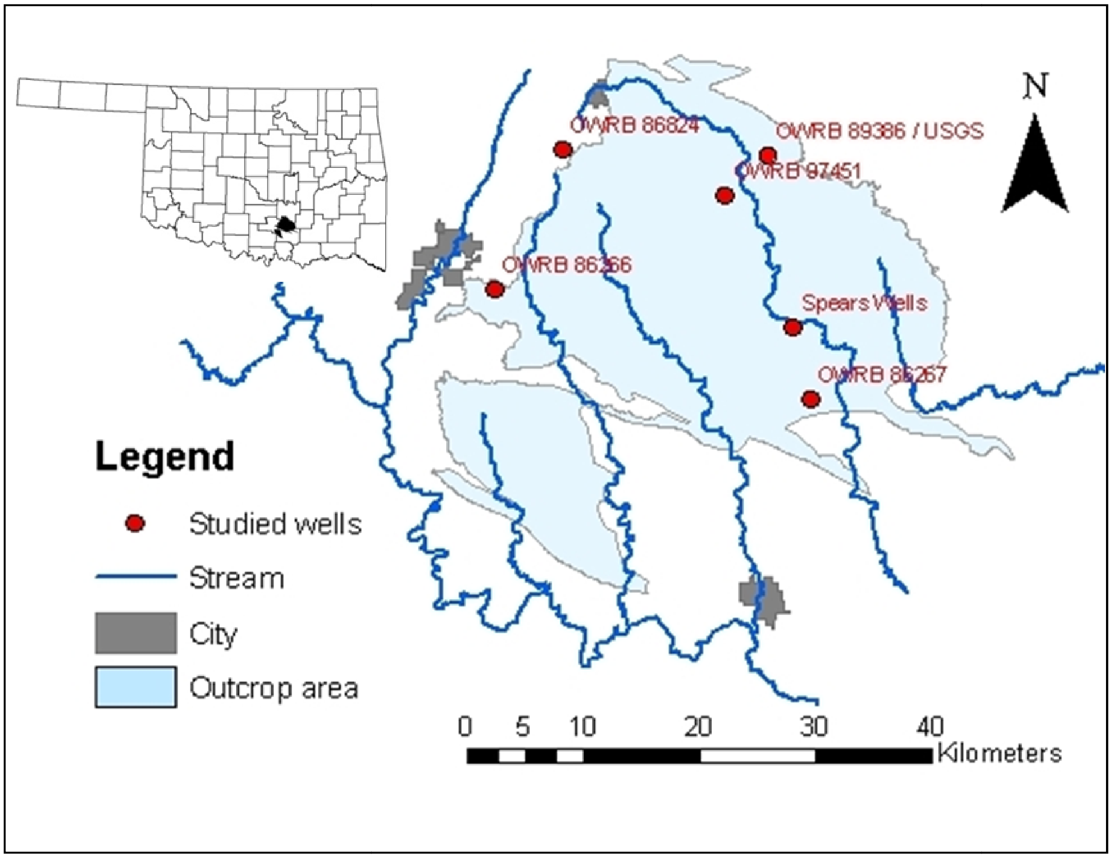

Figure 7.1. Location map of the study area. The area is the eastern part of the Arbuckle-Simpson aquifer which is called the Hunton Anticline.

Table 7-2. The studied wells are located in the Arbuckle-Simpson aquifer.

One well was in Simpson Group of rocks and the remaining wells are in the Arbuckle Group.

Well Designation	Total Depth [m]/[ft]	Latitude [degrees]	Longitude [degrees]	Aquifer Type (from Ch. 5)	Geological Formation
OWRB 86266	34.1/112	34.476925	-96.936317	Unconfined	Simpson
OWRB/USGS 86267	23/75	34.393408	-96.635534	Unconfined	Arbuckle
OWRB 86824	76/250?	34.585554	-96.872478	Confined	
OWRB 97451 mesonet	78/257	34.552056	-96.717933	Semi-confined	
OWRB 89386 USGS Fittstown	121/396	34.582889	-96.679514	Semi-confined	
OWRB 101246 (Spears test 1)	183/600	34.449633	-96.65262	Confined	
OWRB 101247 (Spears test 2)	548/1800	34.449443	-96.65214	Confined	

Bredehoeft (1967) approach to determine S_s from tidally-induced water-level fluctuation is applicable for confined aquifers. The wells 86266 and 86267 revealed no signal of earth tides; rather their water-level fluctuations were the product of barometric pressure changes (Rahi and Halihan, 2009). Earth tides signals were the dominant in three wells (101246 (Spears 1), 101247 (Spears 2), and 86824), and were present but not dominant in the remaining two wells (Rahi and Halihan, 2009). Rahi and Halihan (2009) concluded that the wells 101246, 101247, and 86824 are tapping the confined portion of the Arbuckle-Simpson aquifer. The wells 89386 and 97451 were tapping the semi-confined portion of the aquifer. The two remaining wells were tapping the unconfined aquifer. The wells that were tapping the confined and the semi-confined portion used to determine S_s and η. The wells that were studied are shown in Figure 7.1 and Table 7.2. One well (OWRB 86266) was tapping the Simpson Group (mainly sandstone) and the rest tap the Arbuckle Group (mainly carbonate rocks). The depths of most of the wells were taken from the records of the OWRB. The question mark on the depth of the well OWRB 86824 signifies doubts of the author about the reliability of the reported depth.

The tide component that was utilized for the calculation is M1. M1 is pure lunar semidiurnal and not influenced by the atmospheric-pressure changes. The other two lunar tides, O1 and N2, were neglected because their influence on water-level fluctuation was very small. The tides S2 and K1 were neglected because they were affected by barometric pressure (Hsieh et al., 1987).

METHODOLOGY

Groundwater fluctuations in open wells were monitored and recorded at 15-minute intervals using Solinst Levelogger® pressure transducers. The atmospheric pressure was recorded for the same period using a Solinst Barologger ® attached in the upper two meters of the well OWRB 101246 and OWRB 86824.

Transducers employed for water-level monitoring were not vented. Unvented transducers record a pressure that includes water and atmosphere components. The first step in data processing was to remove the barometric pressure component from the water-level data. Atmospheric pressure components were removed by subtracting the monitored atmospheric pressure from the raw water-level data for each time step. Figure 7.2 shows the water-level fluctuations, the fluctuations compensated for barometric pressure, the barometric pressure fluctuation for OWRB 101246 (Spears test 1 well), and moon phases.

After removing the atmospheric-pressure component, the raw data were detrended and smoothed to remove the trend that may be present. Detrending and smoothing were achieved using two smoothing techniques: differencing and/or moving average. The difference filter is given by (Shumway and Stoffer, 2006):

$$y_t = x_t - x_{t-1} \tag{7.17}$$

where

y_t is the differenced head at time t, and

x_t is the measured head fluctuation at time t.

The symmetric moving average is given by:

$$y_t = \sum_{j=-k}^{k} a_j x_{t-j} \tag{7.18}$$

where

$a_j = a_{-j} \geq 0$, and

$\sum_{j=-k}^{k} a_j = 1$

Differencing is an example of a high-pass filter because it retains or passes the higher frequencies. The moving average is a low-pass filter because it passes or retains the lower (slower) frequencies (Shumway and Stoffer, 2006).

A least squares fitting as presented by Hsieh et al. (1987) was employed to decompose water level fluctuations into its harmonic components. The least squares method, which minimizes the sum of squares of a set of residuals (SSR), is presented in the following paragraphs:

Let x_i be the i^{th} measured pressure head fluctuation corresponding to time t_i:

$$SSR = \frac{1}{n}\sum_{i=1}^{n}\left[x_i - \frac{a_0}{2} - \sum_{j=1}^{p}(a_j \cos\omega_j t_j + b_j \sin\omega_j t_j)\right]^2 \quad (7.19)$$

where:

n is number of measured pressure head points,

t_i is the time of the i^{th} pressure head measurement,

ω_i is frequency of the i^{th} tidal component,

$P=5$, it is the number of tidal components considered for the calculations, and

a_0, a_j and b_j are the $2P+1$ unknown coefficients to be determined by the least square method.

Equation 7.19 is differentiated with respect to the $2P+1$ unknown coefficients to produce $2P+1$ equations. Each equation is set to equal zero to minimize the sum of the squares. The resultant system of linear equations is solved to obtain the $2P+1$ unknown coefficients. The amplitude of the j^{th} tidal component (A_j) is computed by:

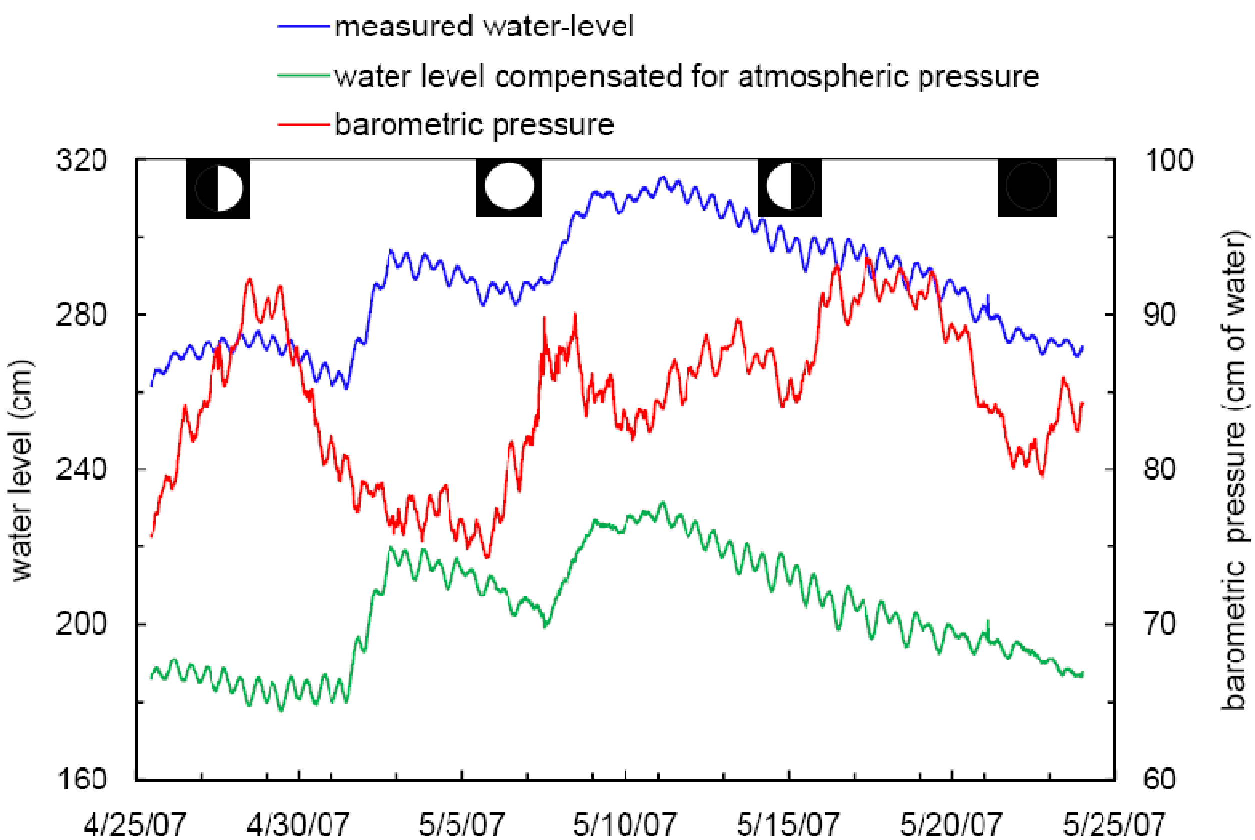

Figure 7.2. Measured water levels, water level compensated for atmospheric pressure, and barometric pressure for the well OWRB 101246 within the Arbuckle-Simpson aquifer (Spears 1 test well). The moon phases are shown in the figure to because the tide amplitude is influenced by the position of the Moon around the Earth.

$$A_j = (a_j^2 + b_j^2)^{1/2}, \tag{7.20}$$

and its phase angle (ϕ_j) is given by:

$$\phi_j = \tan^{-1}(b_j / a_j) \tag{7.21}$$

The calculations were restricted to the amplitude and phase angle of the five major tidal components (O1, K1, M2, S2, and N2).

The percent of variance (pv) for each tidal component is an indicator of the relative significance of the j^{th} tidal component in the tidally influenced water-level fluctuations. The percent of variance is half the amplitude (A_j) of the tidal component (j) squared divided by the variance of the data (Var). The percent of variance is given by (Hsieh, et al., 1987; Merritt, 2004):

$$pv_j = \frac{A_j^2}{2Var} \tag{7.22}$$

and Var is given by:

$$Var = \frac{1}{n}\sum_{i=1}^{n}(x_i - \frac{\alpha_o}{2})^2 \tag{7.23}$$

The value (pv_j) is indicator of the relative significance of the j^{th} tidal component in the tidally influenced water-level fluctuations. The sum of percent of variances (pv_j) is the mean square of fit (MS) divided by the Var of the data and it is given by (Merritt, 2004):

$$\frac{MS}{Var} = \sum_{j=1}^{p} pv_j = \frac{1}{2Var}\sum_{j=1}^{p} A_j^2 \tag{7.24}$$

The sum of percent of variances (MS/Var) is an indicator of the degree of the success with which the harmonic components identified by the regression process explain the variance in the data set. Results revealed that M2 tide had the highest *pv*. Hence M2 was considered for the S_s determinations.

Equation 7.5 was utilized to determine the specific storage of the Arbuckle-Simpson aquifer. The amplitude of the water level fluctuations (A_h) along with the amplitude of the harmonic component (mainly M2) of W_2, (A_{w2}), were substituted into equation 7.5 to compute the specific storage. A_{w2} was computed using equation (7.6) and the entries of Table (7.1). The constant of gravity (g) that was utilized for equation (7.6) is 797 cm/sec^2.

The porosity can be determined using Equation 7.7. The specific storage was determined as outlined above. The barometric efficiency was calculated by computing the cumulative sum of water-level and barometric-pressure changes according to the algorithm detailed in Section 7.2.2. Then, the cumulative sums of water-level changes were plotted against the cumulative sums of barometric-pressure changes. The slop of the linear fit to the plot is the barometric efficiency.

RESULTS AND DISCUSSIONS

The Specific Storage of the Arbuckle-Simpson Aquifer

Water-level fluctuation data were decomposed into its tidal components. The analyses were restricted to five tidal waves (O1, K1, M2, S2, and N2). The percent of variance (*pv*) for each component was determined along with the sum of percent of variances for the studied tides. Results of the harmonic analyses are listed in Table 7.3. Table 7.3 reveals that the dominant tide within the wells of the confined portion of the aquifer is M2. The dominant tide within the wells of the semi-confined aquifer is S2. Figure 7.3 shows a periodogram for water-level fluctuations within the well OWRB 101246 (Spears 1). Since the well is tapping the confined aquifer, its water-level fluctuations were dominated by the tide M2. An example of S2-dominated water-level data is the periodogram of the well OWRB 89386 (USGS) wells (Figure 7.4). The tide M2 was used to determine the specific storage because it was the dominant tide on groundwater fluctuations.

Results of the specific storage calculations are shown in Table 7.4. The specific storage values range from 5.4E-8 to 2.1E-7 cm^{-1}. The average value for the entire aquifer is about 1.056E-7 cm^{-1} (3.22E-6 ft^{-1}). The average specific storage value for the wells that are tapping the confined portions of the aquifer is 5.6E-8 cm^{-1}. The average specific storage for the wells which tap the semi-confined part of the aquifer is 1.8E-7 cm^{-1}. Since the theory adapted for the specific storage determination, followed in this research, was derived for confined aquifers (Bredehoeft, 1967), the storage coefficient resulted from the confined parts for the aquifer is considered. Therefore the specific storage of 5.6E-8 cm^{-1} (1.7E-6 ft^{-1}) is adapted for the Arbuckle-Simpson aquifer. After all, specific storage is a confined aquifer parameter and has no meaning for other types of aquifers.

Bredehoeft (1967) suggested that amplitude of tides in water-level fluctuations increases with depth. Increase of the tidal amplitude means decease in the specific storage. Results obtained by this study did not reveal variation of specific storage with depth. The specific storage was plotted against well depth for the five studied wells. Results suggested no well-defined trend of variation of the specific storage with depth (Figure 7.5). For the specific storage obtained from the wells of the confined aquifer, the results indicated no significant variation with depth (figure 7.5). These findings deviated from the previous research. However, they are consistent with the results of other researchers who studied the Arbuckle-Simpson aquifer.

Christenson et al. (2009) and Puckette et al. (2009) found no changes of temperature or chemistry of the waters of the aquifer as depth increases.

The Storage Coefficient of the Arbuckle-Simpson Aquifer

The storage coefficient (storativity) of the aquifer was calculated for the wells of the confined and the semi-confined portions of the aquifer. Storage coefficient is the product of specific storage and aquifer thickness. The aquifer thickness were obtained from the results of the 3-D EarthVision® model of the Arbuckle-Simpson aquifer developed by Faith and Blome (2008). Table 7.4 shows the storativity values along with the aquifer thickness. The average storage coefficient for the entire aquifer is 0.011. The average storage coefficient for the confined portions is 6.3E-3, and for the semi-confined portions is 1.8E-2.

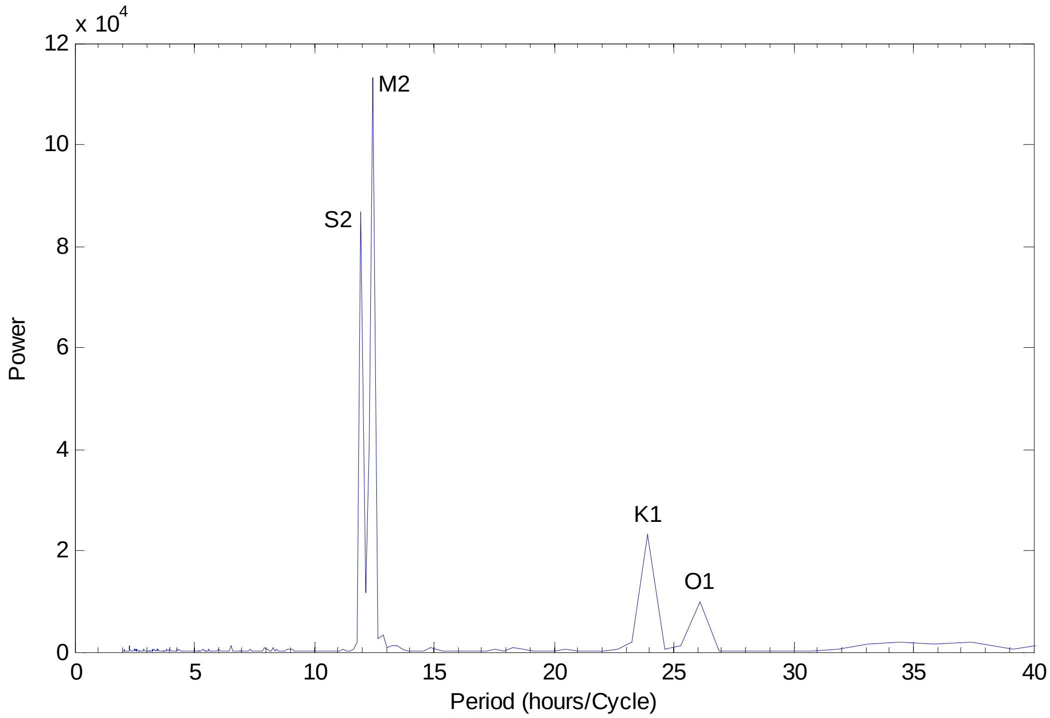

Figure7.3. Power spectrum calculated from water-level fluctuations for the well OWRB 101246 (Spears test 1).

The well was located within the Arbuckle-Simpson aquifer. M2 tide dominated the spectrum which indicates a confined aquifer conditions.

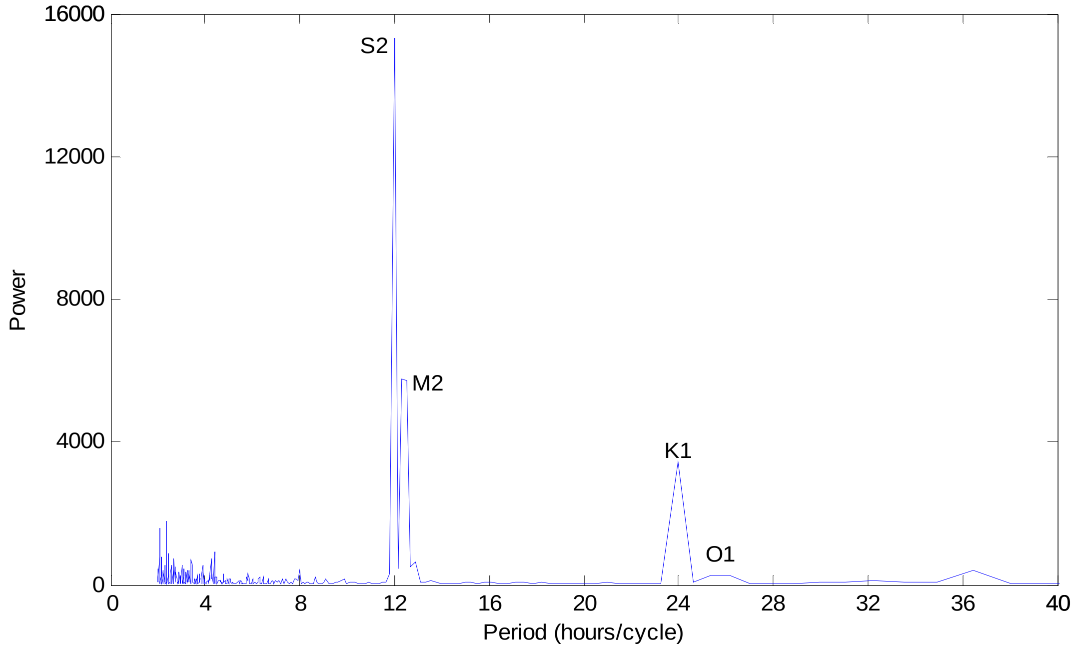

Figure7.4. Power spectrum calculated from water-level fluctuations at the well OWRB 89386.

The well was located within the Arbuckle-Simpson aquifer (USGS Fittstown). S2 dominated the spectrum and M2 present but with less power which indicates a semi-confined aquifer conditions.

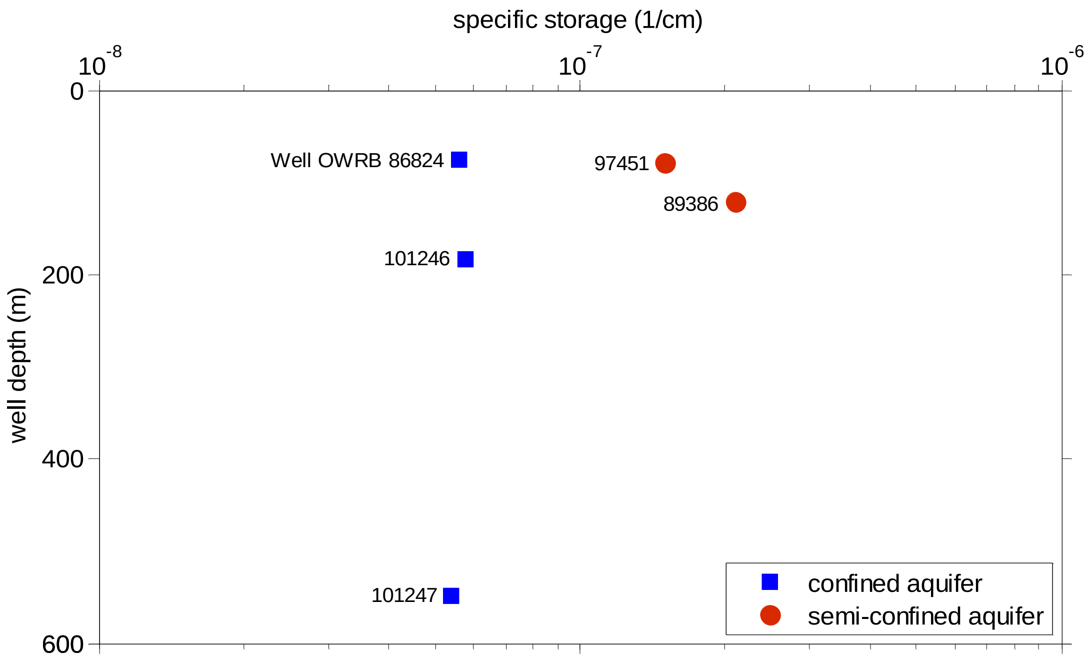

Figure 7.5. Specific storage vs. well depth for five wells for the Arbuckle-Simpson aquifer.

Previous research showed a decrease of specific storage with depth. Results of this study did not reveal such behavior. The Arbuckle-Simpson is uniform with depth.

Table 7.3. Harmonic components and some parameters of the main five tides.

These tides constitute about 95% of the tidal potential, and they are of importance to hydrogeological analyses.

Wave Component	**Well OWRB 101246 (Spears test 1)**			**Well OWRB 101247 (Spears test 2)**			**Well OWRB 86824**			**Well OWRB 97451**			**Well OWRB 89386 (USGS)**		
	Amplitude	**Phase**	**Percent Variance**	**Amplitude**	**Phase**	**Percent Variance**	**Amplitude**	**Phase**	**Percent Variance**	**Amplitude**	**Phase**	**Percent Variance**	**Amplitude**	**Phase**	**Percent Variance**
Tide	[cm]	[deg]	[%]	[cm]	[deg]	[%]	[cm]	[deg]	[%]	[cm]	[deg]	[%]	[cm]	[deg]	[%]
O1	0.058	77.03	4.3	0.070	-176.3	5.8	0.067	115.8	3.9	0.021	55.75	2.5	0.013	101.5	1.3
K1	0.053	-28.9	3.5	0.080	129.3	7.6	0.106	151.9	9.9	0.006	3.24	0.2	0.034	-100.4	8.6
M2	**0.222**	**-18.3**	**61.2**	**0.235**	**-74.4**	**65**	**0.227**	**176.5**	**45**	**0.086**	**-20.4**	**43.1**	**0.062**	**-8.66**	**29**
S2	0.078	116.9	7.6	0.112	119.4	15	0.135	72.19	15.9	0.018	87.81	1.9	0.058	67.41	24.9
N2	0.049	96.62	3	0.049	-16.2	2.8	0.041	-2.22	1.5	0.021	69.19	2.6	0.013	77.62	1.3
Mean square of fit/variance of signal			**79.6**			**96.2**			**76.2**			**50.3**			**64.5**

Table 7-4 Storativity and Porosity Calculations for the Arbuckle-Simpson aquifer.

Well Name	Tidal Comp.	Well Depth [ft]/[m]	Latitude [degrees]	Longitude [degrees]	Specific Storage [cm^{-1}]	Aquifer Thickness [m]	Storativity [dim]	Porosity [dim]
OWRB 86824	M2	250/76.2	34.5855535	-96.8724779	5.6E-08	1120	6.27E-03	0.14
OWRB 97451		257/78.1	34.5520556	-96.4179330	1.5E-07	1010	1.5E-02	0.42
OWRB 89386 USGS Fittstown		396/120.7	34.5828890	-96.6795137	2.1E-07	1028	2.15E-02	0.48
OWRB 101246 (Spears test 1)		600/183	34.449633	-96.6526158	5.8E-08	1126	6.53E-03	0.15
OWRB 101247 (Spears test 2)		1800/549	34.4494431	-96.65214	5.4E-08	1127	6.08E-03	0.15
Average							**0.011**	**0.26**

Fairchild et al. (1990) reported average storage coefficient of 0.008 based on pumping tests analyses. The estimated average storativity for the entire aquifer from this study is about 30 percent higher than that of Fairchild et al. (1990). However, the storage coefficient obtained for the confined portions of the aquifer by this study is very close to the value obtained by Fairchild et al.'s (1990) study. It is most likely that the value obtained by Fairchild et al. (1990) also represent the confined aquifer parts because it was calculated by the modified Theis method. The Theis method and its modifications were derived for confined aquifers (Theis, 1939). Evidently, the storage coefficient values obtained by this study are representative of the aquifer and they are close to what haven reported by previous researches.

Porosity of the Arbuckle-Simpson Aquifer

The porosity was determined for the wells OWRB 101246, OWRB 101247, OWRB 86824, OWRB 97451, and OWRB 89386. The porosity was computed using Equation 7.7. BE was determined by equations 7.9 through 7.16. Values used for the computations are: density of water is 1 gm/cm^3, acceleration due to gravity (g) is 979 cm/sec^2, and the compressibility of water is 2.1*10E-10 cm-sec^2/gm (Marine, 1975).

The porosity values for the Arbuckle-Simpson aquifer are shown in Table 7.4. Porosities from the wells OWRB 89386 and OWRB 97451 are too high (45 percent) compared to the other three wells. This may be attributed the high storage coefficient of these two wells. The high porosity was not attributed to high *BE*; rather it was attributed to the noticeably high specific storage. The specific storage obtained from these wells was not considered representative of the aquifer. Accordingly, the porosity values of these two wells should not be weighted equally along those of the other three wells. The porosities of the wells OWRB 101246 and 101247 (the Spears test wells) and OWRB 86824 are considered representative of the Arbuckle-Simpson aquifer. Those porosities more likely reflect the aquifer characteristics. The average porosity from the three wells is 15 percent. It is comparable to the values obtained by Puckette et al. (2009) which was obtained by well-log analysis.

SENSITIVITY ANALYSIS

A sensitivity analysis was conducted to examine the response of the calculated porosity to changes in the input variables. The two independent variables of equation 7.7 are *BE* and the S_s. The sensitivity analysis was carried out by fixing the value of one independent variable and changing the other. The changes on the variables were based on the values obtained during the course of the study, and fixing the upper and lower limits based the nature of each variable. For example, values of *BE* higher than 100 percent were not used for the analysis because they are unreasonable.

First, the value of S_s was fixed at 5.6E-8 cm^{-1} and *BE* was changed from 20 to 100 percent. The porosity increased about 5 times (from 5 to 27 percent) as a result of increasing *BE* five times (from 20 to 100 percent). It is concluded that porosity varies linearly as the *BE* changes. The obtained porosities were reasonable as long as *BE* was kept reasonable. Of course, *BE* value cannot exceed 100 percent or fall below zero percent. In this study, *BE* values higher than 100 percent were obtained by Clark's method as was discussed earlier. The sensitivity of porosity to *BE* values of more than 100 (though unreasonable) was examined. The highest *BE* used was 200 percent which produce a porosity of 0.54 which is still not unrealistic. The finding of this study differs from those reported by Marine (1975). Marine (1975) expected a porosity of 100 percent would be resulted from a *BE* of 50 percent. Figure 7.6 shows the results of the sensitivity analysis for the porosity to changes in *BE*.

The second analysis was to examine the sensitivity of porosity values to changes on S_s. For this analysis *BE* was fixed at 50 percent and S_s altered between 5.6E-8 and 1.1E-6. The range of S_s was chosen based on the range of values obtained during the course of the study for different wells. Some of the S_s values used for the sensitivity analysis were not considered characteristic of the Arbuckle-Simpson aquifer. These were disregarded in later stages because the signal analyses revealed that the aquifer from which the values obtained was unconfined. The response of porosity to changes in S_s is shown in Figure 7.7. The figure shows that if an estimation of S_s was altered one fold, a reasonable value of porosity cannot be obtained. This result requires that the modeler must have reasonable experience regarding the aquifer so he or she can recognize any unreasonable values of S_s.

SUMMARY AND CONCLUSION

The premise of this research is to analyze a confined aquifer-well system problem in which natural stresses cause macroscopic water movement into and out of the well to estimate the hydraulic parameters. The specific storage was determined by analyzing changes in water level caused by tidal potential. Porosity was determined using the specific storage and the barometric efficiency.

The study area is the Arbuckle-Simpson aquifer located in south-central Oklahoma. The aquifer is a carbonate aquifer exhibiting karst features, such as solution channels, especially in the western parts. Field work included monitoring water-level and atmospheric-pressure fluctuations. Measured water-level fluctuations were analyzed for

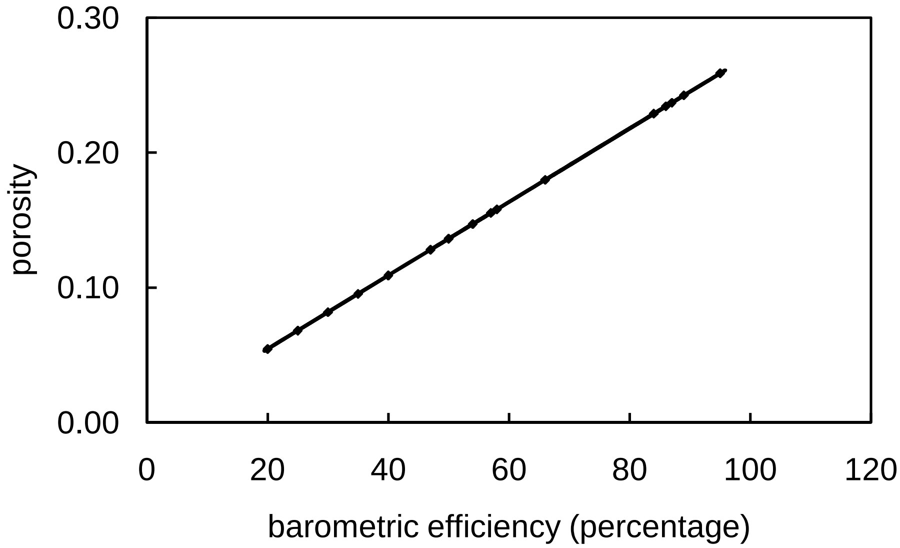

Figure 7.6. Porosity sensitivity to changes in the barometric efficiency.

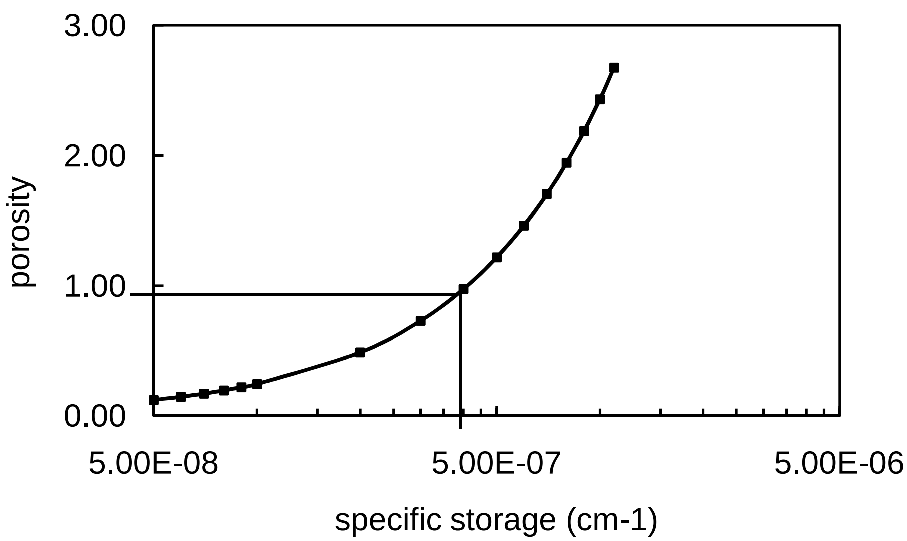

Figure 7.7. Porosity sensitivity to changes in the specific storage.

signal detection and frequency determination using fast Fourier transforms. Harmonic analyses were also used to determine the amplitude and phase angle of the water-level fluctuations. The analyses were restricted to the amplitude and phase angle of the five major tidal components (O1, K1, M2, S2, and N2). The K1 and S2 tides were not considered for specific storage calculations because they are influenced by atmospheric-pressure changes. The N2 and O1 tides contribution to water-level fluctuations was very small and was neglected. M2 dominated the water-level fluctuations and was used to calculate the specific storage.

The specific storage ranged from 5.4E-8 to 2.1E-7 cm^{-1}. The average storativity for the confined portions of the aquifer was 6.3E-3. The storage coefficient obtained by Fairchild et al. (1990) was 8.0E-3. The storage coefficient values obtained by this study are considered characteristic of the Arbuckle-Simpson aquifer, not because they are close to values reported by previous researches, but because when they were used to estimate porosity the results were reasonable.

The porosities for the Arbuckle-Simpson aquifer ranged from 15 percent (for the confined portions) to 45 percent (for the semi-confined portions). The porosity as calculated by Equation 7.7 is very sensitive to *BE* values. Overestimating *BE* may result in erroneously high porosity for the aquifer. Marine (1975) found that *BE* of 50 percent produces a porosity of 100 percent for crystallized hard rocks. Marine (1975) indicated that actual porosity of the rock formation he studied was less than 1 percent. Sensitivity analysis results of this study did not agree with the findings of Marine (1975). Changing *BE* from 20 percent to 200 percent increased porosity from 5 percent to 54 percent.

CHAPTER VIII

SUMMARY AND CONCLUSIONS

Groundwater studies usually involve the use of either analytical or numerical approach to solve a particular problem. A numerical model requires the assignment of a discrete value of the flow (hydraulic) parameters to each cell or block in the model (flow) domain. The discrete values are designated as the model parameters. For a groundwater model to produce reasonable prediction or output, the model parameters should accurately represent the physical aquifer system. The aquifer parameters, which are needed to characterize the flow domain and are transmissivity (T), storage coefficient (S), and porosity (η).

The Oklahoma Water Resources Board (OWRB) has conducted a hydrological study for the Arbuckle-Simpson aquifer. The study objectives are to evaluate the water resources of the aquifer and put forward a management plan to utilize the resources in a manner that will preserve the ecological and environmental balance. In particular the utilization of the resources should have no adverse effects on the discharge of the springs and streams. One objective was to "characterize the Arbuckle-Simpson aquifer in terms of geologic setting, aquifer boundaries, hydraulic properties…" This research is concerned with the aquifer hydraulic parameters.

Traditionally, the aquifer parameters are determined by field or laboratory methods such us pumping and laboratory core tests. Pumping tests might be problematic especially when the aquifer water quality or the aquifer pollution is an issue, besides it is a time consuming and costly experiment. In addition, field methods are of limited spatial extent (i.e. the results represent the specific site not the entire aquifer). The laboratory results are of limited usefulness and hard to generalize since they represent disturbed core samples. Thus, alternative methods of parameters' estimation that are less costly, would not interfere with any water quality issue, and could reflect the behavior of the entire aquifer (or large portions of it) are appealing. Several analytical and modeling approaches to estimate the aquifer hydraulic parameters are available. One approach is the use of the naturally induced stresses on the aquifer surfaces which cause water level fluctuation within open wells. These natural stresses are solid-earth tides and atmospheric-pressure changes.

Earth tides forces result in dilatation or compression of the aquifer materials. A dilatation or compression cause proportional in the aquifer formation stress, which is balanced by an increase or decrease in pore fluid pressure. As the pore fluid pressure changed, a difference in pressure is introduced between the well bore storage and the aquifer, which result in flow into or out of the well. Barometric effects on aquifers result changes in atmospheric pressure. Atmospheric changes are periodic, diurnal and semidiurnal, and aperiodic. The premise of this research is to analyze a confined aquifer-well system problem, in which pressure oscillation causes macroscopic water movement into and out of the well to estimate the hydraulic parameters.

The specific storage is determined by analyzing changes in water level caused by tidal potential. The five tides that account for about 95 percent of tide potential are M2, S2, N2, O1, and K1. The amplitude of the tidal wave and the amplitude of water-level changes were determined for each tide and the Specific storage was determined using equation 3.18. Porosity was determines by Equation 3.23 after *BE* was determined. *BE* was determined by a new method that is introduced as part of this research. Transmissivity was not determined due to inability of the existing methods to determine transmissivities higher than 500 ft^2/day. Nevertheless, methods to estimate transmissivity by the analysis of earth tides and atmospheric pressure changes were examined extensively.

The Arbuckle-Simpson aquifer is located in south-central Oklahoma. The aquifer covers an area of about 1280 km^2 (500 mi^2) in the Arbuckle Mountains physiographic province of south-central Oklahoma. The topography consists of gently rolling hills separated from the plains by the Washita River. West of Washita River the mountainous area, referred to as the Arbuckle Hills, is characterized by a series of northwest-trending ridges. The eastern part of the mountain, referred to as the Arbuckle Plains, is characterized by a gently rolling topography. Rocks of the Arbuckle-Simpson aquifer outcrop in three anticlines within the Arbuckle Mountain physiographic province: the Arbuckle in the west and northwest, Hunton in the east and northeast, and Tishomingo in the south-central. The study area is the Hunton anticline.

The Arbuckle-Simpson aquifer surface area is drained by several perennial streams. Blue River, Pennington Creek, Mill Creek, Rock Creek, Delaware Creek, and Oil Creek, are the principle streams draining the Hunton anticline and flow generally toward the south-southeast into the Washita River and Red River. The aquifer is considered a carbonates-rock aquifer exhibiting karst features, such as solution channels, especially in the western parts.

Geologic features such as folds, faults, fractures, and solution channels control groundwater flow rates and movements. Recharge to the aquifer is from precipitation that falls within the area. The long term annual average precipitation in the area is 38.2 in/yr. The general flow direction in the Hunton anticline potion of the aquifer is from northwest to southeast.

Field work included monitoring water-level and atmospheric-pressure fluctuations. Water-level monitored and recorded within 14 wells. The depth of the studied wells ranges from 16 m (75ft) to 550m (1800ft). Atmospheric-pressure monitored at three locations. Time periods of monitoring vary from location to the other.

The resulted water-level data were filtered and smoothed using two smoothing techniques, differencing and/or moving average techniques. Measured data were analyzed for signal detection and frequency determination using the fast Fourier transforms. The fast Fourier transform integrates the sine and cosine functions with the data, for each frequency, over the period of record. If the data contain a periodic component corresponding to the frequency, a Fourier transforms value result that is substantially different from zero. The graph of these results with frequency plotted as its abscissa is called a periodogram and shows the frequencies present in water-level fluctuations.

Harmonic analysis was also used to determine the amplitude and phase angle of the water-level fluctuations. The least squares method was employed for this purpose.

The analyses were restricted to the amplitude and phase angle of the five major tidal components (O1, K1, M2, S2, and N2). The K1 and S2 tides even though analyzed but were not considered for specific storage calculations for they are influenced by atmospheric-pressure changes. N2 contribution to the water-level fluctuations was very small and was neglected. The remaining two harmonic components, O1 and M2, were considered to calculate the specific storage. The contribution of O1 to the water-level fluctuations was very small and this tide was eventually neglected. The omitting of O1 leaves M1 as the only tide utilized for the specific storage determination. Porosity was determined from the specific storage and the barometric efficiency.

Signal identification process served as a tool to classify aquifers. The signal detection results showed that water-level fluctuations reveal the presence of four tidal components (O1, M2, K1, and S2), some reveal the strong presence of two components (S2 and K1), and several wells show no presence of any tidal component. The results of the study include three parts as summarized here:

ANALYZING WATER-LEVEL TIME SERIES TO IDENTIFY AQUIFER TYPE

Aquifer type, confined, unconfined, or semi-confined, may be identified by drilling or performing pumping tests. Both methods are costly, involve complex field issues, and may yield inconclusive results. Earth tides are known to influence water levels in wells penetrating confined aquifers or unconfined thick, low-porosity aquifers. Water-level fluctuations in wells tapping unconfined aquifers are influenced by changes in barometric pressure. These two findings were employed as a new approach to identify the aquifer type. Time-series analyses of water-level fluctuations of the Arbuckle-

Simpson aquifer were utilized in nine wells to identify aquifer type by evaluating the influence of earth tides and barometric pressure variations.

The Arbuckle-Simpson is a thick (~1000 m) carbonate aquifer located in south-central Oklahoma. Previous studies categorize the aquifer as confined in some localities and semi-confined to unconfined in others. Harmonic analysis for water-level time series was employed to identify the aquifer types within the Arbuckle-Simpson aquifer. The analyses involved the determination of signal strength and amplitude for each tidal component. Signal identification was achieved using MATLAB and amplitude determination was done by least squares regression. Based on the results of the harmonic analyses, three types of the Arbuckle-Simpson aquifer were identified. The wells OWRB 101246, OWRB 101247, and OWRB 86824 were tapping confined portions of the aquifer. The wells are located in two locations: northwest and southeast of the Hunton Anticline. Wells OWRB 97451 and OWRB/USGS 89386 may be considered as tapping semi-confined portions of the aquifer. These two wells are located in the north western parts of the Hunton Anticline. The remaining studied wells (four of them) were tapping unconfined parts of the aquifer. The Arbuckle Simpson is has a complex geology. More intensive spatial studies are needed to classify the aquifer thoroughly.

IMPROVED METHOD FOR DETERMINING THE BAROMETRIC EFFICIENCY OF AQUIFERS

The *BE* is the ratio of the aquifer pressure head change to the atmospheric pressure change. The *BE* can be used as an index of the capability (competence) of overlying confining layer to resist pressure changes; the thicker the confining layer the higher the *BE*. Another important application of the *BE* is that it can be used to calculate the aquifer porosity or the specific storage when one of the quantities is known.

Commonly, *BE* is determined by the Clark method. The results of this study and other previous studies indicate that the Clark method may produce unrealizable and inconsistent *BE* values. A new model (the Rahi method) overcomes the Clark method problems and produces consistent, physically realizable results was introduced. The Rahi Method was applied to three wells within the Arbuckle-Simpson aquifer and produced *BE* values ranging from 31 to 75 percent with an average of 54 percent, whereas the Clark method gave a range of 9 to 196 percent with an average of 91 percent for the same wells. Additionally, the *BE* was calculated by the two methods for a well in the Floridan aquifer using data from Merritt (2004). The Rahi method gave satisfactory results (56 percent), while the Clark method failed to produce acceptable results and produced a negative *BE*.

SPECIFIC STORAGE AND POROSITY OF THE ARBUCKLE-SIMPSON AQUIFER

The specific storage was determined by analyzing changes in water level caused by tidal potential. Porosity was determines by Equation 7.7. The study area is the Arbuckle-Simpson aquifer. The aquifer is located at south-central Oklahoma. The Arbuckle-Simpson aquifer covers an area of about 500 mi^2 in the Arbuckle Mountains Physiographic Province of south-central Oklahoma. The aquifer is considered a carbonate aquifer exhibiting karst features, such as solution channels, especially in the western parts. Field work included monitoring water-level and atmospheric-pressure fluctuations. Measured water-level fluctuations were analyzed for signal detection and frequency determination using the fast Fourier transforms. Harmonic analyses were also used to determine the amplitude and phase angle of the water-level fluctuations. The analyses were restricted to the amplitude and phase angle of the five major tidal components (O1, K1, M2, S2, and N2). The K1 and S2 tides were analyzed but were not considered for specific storage calculations because they are influenced by atmospheric-pressure changes. The N2 contribution to water-level fluctuations was very small and was neglected. The O1 tide was neglected because its contribution to water level was too small. M2 tides were used to calculate the specific storage. Porosity was determined from the specific storage and the barometric efficiency.

The specific storage values range from 5.4E-8 to 2.1E-7 cm^{-1}. The average specific storage values for the wells that were tapping the confined portions of the aquifer was 5.6E-8 cm^{-1}, and that for the wells tapping the semi-confined portions was 1.8E-7 cm^{-1}. The average storativity for the aquifer was 6.3E-3. The storage coefficient obtained by Fairchild et al. (1990) was 8E-3. The storage coefficient values calculated in this study were considered characteristic of the Arbuckle-Simpson aquifer, not because they were close to values reported by previous researches, but because when they were used to estimate porosity the results were very reasonable.

The porosities for the Arbuckle-Simpson aquifer ranged from 15 percent (for the confined portions) to 45 percent (for the semi-confined portions). The porosity as calculated by Equation 7.7 was very sensitive to *BE* values. Overestimating *BE* may result in erroneously high porosity for the aquifer. Marine (1975) found that *BE* of 50 percent resulted in a porosity of 100 percent for crystallized hard rock's aquifer. Marine (1975) indicated that actual porosity of the rock formation he studied was less than 1 percent. Sensitivity analysis results of this study did not agree with the findings of Marine (1975). Changing *BE* from 20 percent to 200 percent increased porosity from 5 percent to 54 percent.

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