INFERRING TIME-VARYING RECHARGE FROM INVERSE ANALYSIS OF
LONG-TERM WATER LEVELS

by

Jesse Ervin Dickinson

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APPROVAL BY THESIS DIRECTOR

This thesis has been approved on the date shown below:

T.P.A. Ferré, Ph.D.
Professor of Hydrology

7-24-03

Date

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ABSTRACT

Water levels in alluvial aquifers of the Southwest typically vary in response to time-varying rates of mountain-front recharge. This suggests the possibility of inferring time-varying recharge rates on the basis of long-term water-level records. Presumably, these recharge rates depend on variations in precipitation rates due to known climate cycles such as the El Nino/Southern Oscillation index and the Pacific Decadal Oscillation. This investigation examined the inverse application of a one-dimensional analytical model for periodic flow [Townley, 1995] to estimate periodic recharge variations on the basis of water-level variations in long-term water-level records. Time-varying water level records at various locations along a flow path were obtained by simulation of ground-water flow in numerical models of idealized basins. Time-varying recharge was represented as sinusoidal recharge of a single period or as a composite of multiple periods of length similar to known climate cycles. Periodic water level components, reconstructed using singular spectrum analysis (SSA), were used to calibrate the analytical model to estimate each recharge component. The results demonstrate that periodic, time-varying recharge rates can be estimated from hydrologic time series such as long-term water level records. The results also demonstrate that periodic recharge estimates were most accurate in basins with nearly uniform transmissivity, and that the accuracy of the recharge estimates depended on the monitoring well locations.

1.0 INTRODUCTION

Interest in quantifying recharge to regional aquifer systems in the arid and semi-arid Southwest has increased because of mandates to achieve sustainability of groundwater resources [Alley *et al.*, 1999]. As these regions experience rapid development and population increases, the demand on groundwater for water supply will continue to grow. Regional groundwater models are commonly used to assess the effects of groundwater development and management practices on the sustainability of these resources. However, these groundwater flow models are very sensitive to recharge rates, which are difficult to quantify [Lerner *et al.*, 1990]. In addition, many of these models have been used to simulate decades of historical or future groundwater development. Therefore, temporal variability of inflows and outflows plays an important role in achieving an adequate simulation of the developed resources.

Regional groundwater models commonly use long-term average recharge rates in transient simulations to assess the limits of sustainability in developed aquifers. Time-averaged estimates of recharge have been used where there is limited model calibration data, or in cases of extensive aquifer development when discharge from pumpage dominates aquifer response. While most simulations of regional groundwater flow have assumed recharge rates that are independent of climatic variability, time-varying estimates of recharge can improve the adequacy of simulations for groundwater systems that are in overdraft conditions.

The potential response of groundwater resources to climate variability is a significant issue for sustainability in the United States [Gleick and Adams, 2000]. Studies have reconstructed the past climate in the Southwest by indirect measures such as tree rings, pollen grains in sediments, and geologic features. These measures indicate complex and cyclical patterns with time scales of decades to centuries [Merideth, 2000]. Other studies have reconstructed climate variability from the previous century and past decades by analyses of streamflow [Lins, 1997, Dettinger *et al.*, 2001], precipitation and temperature [Dettinger *et al.*, 2001], and groundwater levels [Venencio and Garcia, 1998, 1999, 2000]. These reconstructions contain multiple frequencies that are partially coincident to known climate cycles such as the El Nino-Southern Oscillation (ENSO) and Pacific Decadal Oscillation (PDO) [Dettinger *et al.*, 2001]. Water levels with multiple periodic components that are correlated to ENSO and PDO [Venencio and Garcia, 1998, 1999, 2000] have been identified in unconfined aquifers by using spectral methods. The reconstructed cyclical components estimated by spectral frequency analysis have been used to infer time-varying climate controls on groundwater recharge.

Studies of the groundwater supply from aquifers typically use numerical groundwater flow models such as the U.S. Geological Survey's Modular Three-Dimensional Finite-Difference Ground-Water Flow Model (MODFLOW) [McDonald and Harbaugh, 1988]. These numerical models require widely distributed groundwater-level data and boundary condition data such as streamflow, recharge, and evapotranspiration for specified inflows and model calibration. Because water levels are typically measured at infrequent and irregular time intervals, most numerical models have

been limited in terms of comparing and constraining groundwater flow on the basis of water-level responses owing to temporally variable recharge.

In some basins, long-term water levels are measured monthly or bi-monthly over a long period, thus the effects of climate variability can be assessed. Because these extensive water level records are not often common, the high temporal resolution of variability cannot be used systematically in a regional flow model. However, we propose that if these data are distributed within a sub-region of the basin along the general direction of flow, an analytical flow model can be constrained within this sub-region to estimate time-varying recharge rates. It can then be assumed that recharge estimates can be systematically distributed within similar environments in the watershed to represent climatically driven inflows for groundwater flow simulations [*Hanson et al.*, 2002]. The advantages of using an analytical model, such as that of *Townley* [1995], include its simplicity to program and model, and its ability to provide an estimate of time-varying recharge using few water-level records without the construction of a large numerical regional flow model.

1.1 PURPOSE AND SCOPE

This thesis presents the results of a study of the effects of climate variability on groundwater flow in the southwest United States under the Southwest Ground-Water Resources Project of the U.S. Geological Survey (USGS). The study results have been prepared as a manuscript for submission to a refereed journal. This manuscript is currently in the review process of the USGS. This thesis contains the manuscript with expanded introduction and conclusion sections.

The objectives of this analysis were to (i) develop and test a methodology to estimate time-varying recharge to basins using a numerical model of an idealized flow system that served as a surrogate for a typical alluvial aquifer in the southwestern United States, (ii) evaluate inferred estimates of time-varying recharge of both a single period and a complex recharge pattern of multiple periodic components that are similar to known climate periods, by inverse analysis of the *Townley* [1995] model, and (iii) examine the effect of monitoring well locations along the flow path on the recharge estimates.

1.2 PREVIOUS INVESTIGATIONS

Observed periodic variations in water levels are valuable for determining boundary conditions and aquifer parameters. Many solutions for periodic groundwater flow have been developed that are analogous to the periodic heat flow solution of *Carslaw and Jaeger* [1959], notably those of *Jacob* [1950] and *Ferris* [1951] for tidal fluctuations at one boundary of a semi-infinite one-dimensional homogenous aquifer to estimate diffusivity. *Carr and Van der Kamp* [1969] developed a method to estimate the hydraulic conductivity and the storage coefficient separately in aquifers with tidal boundary fluctuations on the basis of amplitude and phase lag. *Carotenuto, et al.* [1980] and *Li, et al.*, [2002] extended these equations to two dimensions to determine aquifer properties in coastal aquifers.

Black and Kipp [1981] proposed a method to determine hydrogeologic parameters on the basis of sinusoidal pressure fluctuations in an excitation borehole for confined aquifers. Equations were developed for conditions of a deep point source in a saturated,

elastic aquifer, and for a fully penetrating line source. The equations describe aquifer responses based on pressure fluctuations as functions of distance, signal frequency, specific storage, hydraulic conductivity, and flow rates.

Numerical flow models have been used to simulate the effects of periodic boundaries on groundwater flow and heads. *Maddock and Vionnett* [1998] considered the effects of seasonal variations in natural recharge and discharge on groundwater capture processes. A numerical groundwater flow model for the Sierra Vista subwatershed [*Vionette and Maddock*, 1992] in southeastern Arizona was used to simulate the effects of seasonal streamflow and evapotranspiration variations for both steady-state and oscillatory steady-state initial conditions. The capture by evapotranspiration and stream reaches, which is either an increase in recharge or decrease in discharge, indicated that the use of steady-state initial conditions produced a groundwater flow solution that was not consistent with the conceptual model. The oscillatory steady-state initial solution that represented seasonal conditions resulted in acceptable capture calculations.

Zawadzki et al. [2002] considered periodic boundary conditions in the development of methodologies for determining well-capture zones, which delineate a spatial aquifer zone from which pumped water may originate. Those capture zone designs accounted for cyclical variations in hydraulic gradients owing to fluctuating tidal boundaries. Numerical groundwater flow models were used to simulate the transport of particles to determine flow paths and travel times under periodic boundary conditions. Steady-state simulations were determined to be adequate for conditions where the period

of the hydraulic-gradient cycle is short relative to the travel time. When travel time approaches the period of the hydraulic-gradient cycle, the water level fluctuations should be considered in the design of capture zones.

Johansson [1987] presented two methods based on groundwater fluctuations to estimate groundwater recharge. The first method used a one-dimensional soil water flow model. Boundary conditions were defined using submodels for precipitation, snow dynamics, interception, and evapotranspiration. The second method directly related groundwater level fluctuations with equal volumes of water using flow balance equations, recession curves, and specific yield.

1.3 GROUNDWATER FLOW MODEL CALIBRATION

Groundwater flow model calibration is performed to ensure that the model boundary and initial conditions and distributed properties adequately represent the flow system. For example, parameters, boundary conditions, and stresses can be adjusted until the simulated groundwater levels and fluxes match measured values within an acceptable margin of error. Model calibration is an inverse problem, in which groundwater levels are known, and the values of model parameters such as transmissivity, storativity, boundary conditions, and stresses (e.g. recharge rates) are adjusted to match simulated water levels with measured water levels.

Steady-state regional groundwater flow models are calibrated for model parameters that produce the average or long-term measured flow conditions. Average flow conditions may be used for calibration when calibration data are not adequate to

describe transient flow, or when the initial, or predevelopment, conditions for a transient simulation are needed. Due to natural variability of recharge, water levels are rarely constant in time. Therefore, definition of the steady-state conditions requires some analysis of transient water-level records to define the average condition. Care must be taken in determining representative average conditions from transient water records [Anderson and Woessner, 1992]. The time-averaging scheme should be selected on the basis of the model objectives. For example, a model may average over all seasonal variations to represent a long-term average. Or, a steady-state simulation may be used to represent the initial conditions for each season in a transient simulation [e.g. Maddock and Vionette, 1998].

In cases of large seasonal or climate-related variations in water levels, and when assumptions of steady-state flow are not appropriate, a transient groundwater flow model can be calibrated to measured time-varying water levels. Unlike steady-state simulations, models calibrated using transient data can be used to determine the aquifer storage properties or time-variable stresses on the flow system.

The magnitude of the uncertainty of the model calibration should be assessed and quantified. These uncertainties are reduced when heads and fluxes are measured at many locations throughout the model domain. Ideally, measured water levels would coincide with model nodes to reduce the calibration to interpolated head. However, this is often impractical because the model grid is usually regularly or systematically spaced, and calibration data are often irregularly distributed. Assumptions of steady or average conditions where transient effects are present may result in misrepresentation of the

system by the model. A transient calibration allows for a solution with temporal changes in stresses. In cases in which water levels are measured infrequently, the temporal variations and processes may be removed through averaging. The uncertainty of the transient simulation can be reduced by calibration to data sets with high temporal resolution.

1.4 THEORY

Among the conditions considered by *Townley* [1995] is one-dimensional flow within a confined aquifer:

$$S \frac{dh}{dt} = T \frac{d^2 h}{dx^2}, \quad (1)$$

where S is the aquifer storage coefficient [-], $h(x, t)$ is head [L] at position x and time t , and T is the aquifer transmissivity [L^2/T]. This model assumes recharge only at the boundary and, therefore, neglects a spatially distributed recharge source term. *Townley* expressed h as:

$$h(x, t) = h_s(x) + \text{Re}\{h_p(x) \exp(i\omega t)\}, \quad (2)$$

where $h_s(x)$ are values of head of the steady-state solution upon which periodic heads $h_p(x)$ are superimposed. The angular frequency ω of periodic recharge variation is equal to $2\pi/P$, where P is equal to period [T]. The $\text{Re}\{\cdot\}$ operator obtains the real part of a complex number. Substituting this equation into equation (1) leads to steady and periodic equations:

$$T \frac{d^2 h_s}{dx^2} = 0, \text{ and} \quad (3)$$

$$T \frac{d^2 h_p}{dx^2} - i\omega S h_p = 0. \quad (4)$$

The solutions to these differential equations are:

$$h_s = C_1 x + C_2, \text{ and} \quad (5)$$

$$h_p = D_1 \cosh ax + D_2 \sinh ax, \quad (6)$$

respectively, where C_1 [-] and C_2 [L] are coefficients that depend on the form and magnitude of the steady boundary conditions, D_1 [L] and D_2 [L] are complex integration coefficients determined for periodic boundary conditions, and $a = (i\omega S / T)^{0.5}$ [1/L].

In this study, the analytical model was used to estimate the effects of periodic recharge at a boundary superimposed over solutions of steady-state flow. The assumption is made that the aquifer system is linear in space and time. That is, the aquifer properties are uniform and constant and the individual time-varying recharge components are independent and can be summed to provide a composite time-varying recharge estimate. For this study, Townley's model is solved using values of D_1 and D_2 to represent a periodic specified head boundary at $x = 0$, and a constant head at the downstream boundary condition at $x = L$, where L is the length of the aquifer in the direction of flow. For these boundary conditions,

$$C_1 = \frac{H_s|_{x=L} - H_s|_{x=0}}{L}, \quad (7)$$

$$C_2 = H_s|_{x=0}, \quad (8)$$

$$D_1 = H_p, \text{ and} \quad (9)$$

$$D_2 = -H_p \left(\frac{\cosh aL}{\sinh aL} \right), \quad (10)$$

where H_s is the specified steady head at $x = 0$ or $x = L$, and H_p is the specified periodic head at $x = 0$. Equations (7) and (8) were substituted into equation (6) to describe the periodic flow component:

$$h_p = H_p \cosh ax - H_p \sinh ax \left(\frac{\cosh b}{\sinh b} \right), \quad (11)$$

where

$$b^2 = a^2 L^2 = 2\pi i(L^2 S / TP). \quad (12)$$

The periodic rate of inflow into the aquifer at $x = 0$ per unit width perpendicular to the direction of flow, Q_p [$(L^3/T)/L$], can be determined from the product of T and the derivative of h_p with respect to x , using equation (6), as

$$Q_p = aTH_p \left(\frac{\cosh aL}{\sinh aL} \right). \quad (13)$$

Townley [1995] notes that b [-] depends directly upon the non-dimensional ratio $R = L^2 S / TP$, which is related to *Theis*' [1935] well function $W(u)$. *Townley* interprets R as a non-dimensional response time that is small for large T and P , and small S and L . R is large if the size of the values of T , P , S , and L are reversed.

2.0 METHODS

The adequacy of the analytical model to infer time-varying recharge rates was evaluated for an idealized aquifer that may be considered as a surrogate for real alluvial-aquifer systems subject to time-varying recharge related to climate variability. The methodology to estimate time-varying recharge using long-term water levels was evaluated using the following steps, which can be applied to a real-world basin beginning with step 2.

1. A numerical model of the idealized aquifer system was used to simulate groundwater flow [MODFLOW, McDonald and Harbaugh, 1988] though an unconfined aquifer with gradational transmissivity values, and recharge at external boundaries varying periodically with periods of 4, 8, or 16 years, or with a composite of periods of 3, 4, 8, 16, and 100 years.
2. For simulations using a composite of recharge periods, resulting variations with the 4-, 8-, and 16-year periods in the simulated water levels were reconstructed by singular spectrum analysis [SSA; Vautard, *et al.*, 1992, Dettinger, *et al.*, 1995].
3. The periodic boundary H_p was determined from the simulated head variations at the boundaries of the idealized model.

4. The amplitudes of simulated head variations at selected points in the idealized aquifer were used to calibrate the analytical model to determine an effective transmissivity. The known parameters were L , P , and S . This optimization was achieved by coupling the analytical model with PEST to estimate T [*Parameter ESTimation; Watermark Computing, 1998*].
5. The recharge rate, or inflow rate Q_p of the analytical model, was calculated by equation (10) using the effective transmissivity. Q_p was compared to the known single and multiple recharge rates applied to the numerical model. This represents an approximation to Darcy's law for each periodic recharge component and results in an estimation of periodic recharge for one or more frequencies.

To apply the analytical model to a typical basin used for the evaluations, the following simplifying assumptions are made: 1) A groundwater flow line from mountain-front recharge areas to a stream, or other fixed head discharge zone at the basin center, can be approximated by the one-dimensional analytical model; 2) While groundwater flow in basins is actually three-dimensional, we assume that flow generally is horizontal; 3) Variations in water levels are small relative to the saturated thickness, thus changes in transmissivity through time are small; 4) The effects of convergence and divergence along the flow line are negligible.

2.1 HYDROGEOLOGIC FRAMEWORK

Groundwater flow in the Southwest generally occurs in unconfined to semi-confined alluvial aquifers. The aquifers consist of alluvial deposits that are derived from relatively impermeable mountain ranges that bound the basins (Figure 1). Groundwater typically flows from mountain-front margins toward lower land-surface elevations at the basin center or to adjacent basins. Recharge occurs by underflow from adjacent, up-gradient basins, by infiltration from streams at the basin center, and by infiltration at or through mountain fronts along the margins of the alluvial aquifers. Basin discharge may occur as underflow to adjacent, down-gradient basins, losses to low-altitude streams, or by evapotranspiration where the groundwater levels approach the land surface.

The idealized aquifer simulated here is consistent with a hydrogeologic framework that is typical of alluvial aquifers in the southwestern United States. The geometry of the idealized basin aquifer system (Figure 1) is rectangular with groundwater flowing from the upstream end and the mountain fronts, and toward the basin center and down-stream end. [*Anderson and Freethey, 1996*].

Figure 1 Idealized alluvial basin aquifer containing a one-dimensional groundwater flow path from the mountain-front recharge boundary to the axial stream. Recharge occurs by underflow from an adjacent basin and mountain front recharge. Discharge occurs by underflow to an adjacent basin and baseflow to the stream.

2.2 IDEALIZED MODELS

Groundwater flow models for several idealized basin geometries, using MODFLOW [McDonald and Harbaugh, 1988], were used to simulate the response of an unconfined aquifer to single and multiple-frequency periodic recharge rates. For each geometry, a one-dimensional flow path was chosen along a flow tube between the

recharge and discharge boundaries. The hydraulic conductivity was specified to be homogeneous and isotropic in each model, as was the storage coefficient. However, in each model, the basement that bounds the aquifer on the bottom slopes downward, in various configurations, so that the aquifer thickness increases toward the basin center. This gives rise to spatially variable transmissivity.

Figure 2 Upper figure shows the idealized aquifer cross-section consisting of a sloping, no-flow basement boundary simulated by a numerical groundwater flow model. Lower figure shows the flow domain of the analytical model with assigned periodic head boundaries defined by the water level amplitudes from the numerical model at boundaries $x/L = 0$ and $x/L = 1$. The effective transmissivity was determined by calibration of simulated water level amplitudes between the flow boundaries.

The adequacy of the analytical model was evaluated by comparing recharge estimates with the recharge applied to the MODFLOW models. Specifically, MODFLOW [McDonald and Harbaugh, 1988] was used to simulate flow within a one-dimensional cross-section of an unconfined aquifer with specified periodic recharge at the mountain front boundary $x = 0$ and with constant head at $x = L$ (Figure 2). The model grid consisted of 1 row of unit width and 501 columns of cells of length $\Delta x = 26.67$ meters. The hydraulic conductivity and specific yield were uniform, thus the transmissivity varied only due to changes in the saturated thickness. Variation of the saturated thickness is dominated by the angle θ that represents the slope of the impermeable bottom boundary near the recharge boundary (Figure 2). The time-averaged transmissivity was 464 meters-squared per day at the basin center for all geometries. A slope of $\theta = 0$ degrees resulted in a minor increase in transmissivity toward the recharge boundary to 472 meters-squared per day due to higher water table elevation in the unconfined aquifer near the recharge boundary. A maximum slope of $\theta = 13$ degrees, the steepest angle to retain saturation of all numerical model cells, resulted in a decrease of the time-averaged transmissivity toward the recharge boundary to 18.6 meters-squared per day. The specific storage was assumed to be small compared with the assumed specific yield of 0.1 used for single recharge periods, or 0.01 used for a composite of multiple recharge periods, and thus storage was effectively uniform in all cases.

The boundary condition applied at $x = 0$ can represent a recharge rate varying at just one frequency, or alternatively, several periods. The time-averages of both the single and composite recharge rates are 0.929 meters-cubed per day per meter, which retains the time-averaged saturated thickness and transmissivity for both analyses.

Single recharge components of periods of 4, 8, and 16 years and magnitudes varying between 0 meters-cubed per day per meter and 1.86 meters-cubed per day per meter were applied to basins with θ ranging from 0 degrees to 13 degrees, in the course of a total of seven model configurations. The recharge rates of only one period constitute all recharge into the models, and the recharge rate is always positive.

A complex recharge pattern consisting of a composite of five periodic components were applied to two model configurations with $\theta = 0$ degrees and 12.5 degrees. The steady-state transmissivities were 472 meters-squared per day and 30.1 meters-squared per day, respectively, at the recharge boundary. The composite recharge had components with periods of 3, 4, 8, 16, and 100 years. A random, red-noise component was also added (Figure 3). The 3- and 4-year components represent variations that might derive from the ENSO climatic cycle [Merideth, 2000]. The amplitudes of the 3- and 4-year components varied between 0 to 0.255 meters-cubed per day per meter, and 0 to 0.227 meters-cubed per day per meter, respectively, in the composite recharge rate. The interdecadal 16-year cycle might derive from the PDO [Dettinger *et al.*, 2001]. The 100-year component is similar to century-scale variations that have been identified in some tree-ring chronologies [Meko *et al.*, 1980]. The amplitudes of the 8-year, 16-year, and 100-year components each varied between 0 and 0.0283 meters-cubed per day per meter,

in the composite recharge rate. The random component was assigned a comparable magnitude that varied between -0.0142 meters-cubed per day per meter to +0.0142 meters-cubed per day per meter. The shorter 3-year and 4-year ENSO-like cycles with the random components contribute 95 percent of the recharge variation. It was expected that the longer 8-, 16-, and 100- year components would be preserved with greater distance from the boundary than the shorter components, which occurs in similar applications of head flow in solids. Therefore, the variation of the recharge rates of the shorter components were chosen to be greater than the longer components for preservation of water level variations due to the shorter components at large distances from the recharge boundary.

Figure 3 Composite recharge rate applied to the idealized aquifer model, which is the sum of recharge contributions with periods of 3, 4, 8, 16, and 100 years, and a random component. The 4-, 8-, and 16-year components, in bold, are inversely inferred by calibration of the analytical model.

2.3 VERIFICATION OF THE NUMERICAL MODEL

The adequacy of MODFLOW to simulate periodic flow was verified by replicating the analytical solution for aquifer properties for various values of L^2S/TP (Figure 4). The amplitude of water level variation, $\max(h_p) - \min(h_p)$ from equation (9), for $x/L = 0$ to $x/L = 1$, was compared to the simulated amplitude of water level variation using MODFLOW [McDonald and Harbaugh, 1988] for confined flow conditions. A specified head boundary was assigned at $x/L = 0$ and a head-dependent boundary was assigned at $x/L = 1$. The values of R , the non-dimensional response time, were achieved using parameter values of $L = 30.5$ meters, $T = 9.29$ meters-squared per day, $S = 0.002$ to 0.2 , and $P = 0.2$ days to 20 days, where S and P were varied by an order of 10 . The optimal number of finite difference cells of the numerical model to replicate the analytical solution was determined by comparison of the averages of squared-residuals between the numerical and analytical results for various numbers of grids (Figure 5). A total of 501 model cells were used for the analysis because additional cells did not significantly improve the numerical solution at the cost of additional simulation time.

Figure 4 Comparison of MODFLOW simulations to Townley solution simulations for a specified head boundary at $x/L = 0$, a head-dependent boundary at $x/L = 1$, and various values of R was calculated using parameter values of $Q_p = -0.093$ to 0.093 meters-cubed per day per meter, $L = 30.5$ meters, $T = 9.29$ meters-squared per day, $S = 0.002$ to 0.2 , $P = 0.2$ days to 20 days.

Figure 5 Average-squared residuals between the numerical and analytical solutions for various number of finite difference grids in the numerical model.

2.4 ANALYSIS OF WATER LEVELS

The SSA Toolkit [Dettinger *et al.*, 1995] was used to extract the periodic components from the water levels simulated by the MODFLOW models. The SSA Toolkit uses singular spectrum analysis – a principle-component analysis of lag correlations - to isolate and reconstruct significant periodic components in time series [Vautard *et al.*, 1992]. The time series frequency data is obtained from the power spectrum. Because measured water levels from real basins commonly span short time

scales, we analyzed only 50 simulated years of water level variations. The observation interval period was 0.2 years.

The amplitudes of the periodic water-level variations were determined after the simulation reached the periodic steady state condition. For composites of multiple recharge cycles, the SSA Toolkit [Dettinger, *et al.*, 1995] was used to reconstruct periodic components of water levels that corresponded to the applied recharge periods at specified distances from the recharge boundary. The periodic head boundary of the analytical model at $x = 0$ was specified using the water level amplitude at $x = 0$ of the numerical model. The amplitudes of the periodic reconstructed components of water levels and components were used as calibration points for the calibration of the analytical model. In addition, the locations of the calibration points were varied to determine whether the recharge estimates depend on the location of the calibration points.

The analytical model was calibrated to the idealized numerical model using PEST [Watermark Computing, 1998]. The optimization scheme searched among uniform values of transmissivity in the analytical model to reduce the sum of square error between the analytical- and numerical-model water-level amplitudes at shared observation points. The distribution of the observation locations was varied to determine how the locations affected the analytical-model recharge estimates. Water-level amplitudes from all numerical model cells were used in calibration when the single period recharge boundary was applied, which represents an ideal case of periodic water level response that is fully characterized along the flow path length. When the composite of recharge periods was simulated, water levels from only two model locations were used for calibration, which

may reflect conditions of limited data in a real basin. One calibration point was at $x/L = 0.62$; the other point was moved to $x/L = 0.1, 0.2, 0.3$, or 0.42 .

3.0 PERIODIC WATER-LEVEL RESPONSES IN IDEALIZED AQUIFERS

3.1 SINGLE RECHARGE PERIODS

In all cases, the observed water levels varied most near the mountain-front recharge boundary and were defined as constant at the axial discharge boundary. The variations of water levels, $\pm h_p$, around the steady-state water levels, h_s , increased as functions of recharge period length and basement-slope θ . Small θ resulted in small steady-state water-level gradients and small periodic water-level amplitudes ($\max(h_p) - \min(h_p)$) at the recharge boundary that attenuated slowly with distance into the basin (Figure 6). The small mountain-front transmissivity associated with large θ resulted in steep steady-state water level gradients and large amplitudes near the recharge boundary that attenuated rapidly with distance into the basin. The amplitude at the recharge boundary for the 4-year period increased from 4.1 meters when $\theta = 0$ degrees to 18.8 meters when $\theta = 13$ degrees. The amplitude at the recharge boundary for the 16-year period increased from 8.2 meters when $\theta = 0$ degrees to 33.1 meters when $\theta = 13$ degrees.

Figure 6 Range of water levels variations of the numerical forward model of geometries $\theta = 0$ degrees and 13 degrees between $x/L = 0$ and 1 for applied single-component recharge rates of 4- and 16-years. For $\theta = 0$, amplitudes of variations are 4.08 meters using the 4-year period and 8.20 meters using the 8-year period. For $\theta = 13$ degrees, amplitudes of variations are 18.8 meters using the 4-year period and 33.1 meters using the 8-year period.

3.2 MULTIPLE RECHARGE PERIODS

Water level variations due to recharge rates of multiple periods were largest at the recharge boundary and attenuated rapidly in overall amplitude with distance from the recharge boundary (Figure 7). The shorter periods, whose flow rates in the composite recharge varied most through time, represented most of the variation at $x/L = 0.0$ and 0.3 . The longer periods (8 years and greater), whose flow rates varied little through time, contributed little to the variation of the composite water level near the recharge boundary although small variations were attributed to the 8- and 16-year components at $x/L = 0.62$.

Figure 7 Water levels at $x/L = 0.00, 0.30$, and 0.62 simulated by the forward numerical model using the composite of multiple recharge periods. Amplitude was attenuated with distance from the recharge boundary.

4.0 ESTIMATION OF PERIODIC RECHARGE RATE

4.1 SINGLE RECHARGE PERIODS

The accuracies of estimates of periodic recharge variations from PEST calibration of the analytical models was tested by comparison to the recharge actually applied in the numerical model. For single recharge cycles, the ratios of estimated to known recharge were 0.977, 0.980, and 0.986 for 4-, 8-, and 16-year periods, respectively, when $\theta = 0$ degrees (Figure 8). The accuracy of the recharge estimates gradually diminished as θ increased until about $\theta = 10$ degrees, above which estimation errors increased rapidly. However, even for $\theta = 13$ degrees, the ratios of estimated to known recharge were 0.950, 0.946, and 0.929 for periods of 4-, 8-, and 16-years, respectively. This accuracy is comparable to the uncertainty for other hydrologic measurements such as streamflow separation [Lerner *et al.*, 1990].

An assumption of the Townley [1995] model is that transmissivity is constant in both time and space. However, the numerical model allows the saturated thickness to vary, which results in changes of transmissivity through time. These changes in transmissivity are significant when water level changes are large relative to the saturated thickness, especially near the recharge boundary in aquifers with large θ and small time-averaged transmissivity. The accuracy of the recharge estimate decreased as a function of the ratio of water level amplitude to average saturated thickness (Figure 8). The estimates are most accurate when the amplitude was small relative to saturated thickness. The accuracy diminished quickly as the ratio of amplitude to saturated thickness approached 0.1, after which the ratio diminished more slowly. For $\theta = 13$ degrees and 16

year period, the accuracy is 0.929, even when the ratio of amplitude to saturated thickness = 0.69.

Figure 8 Ratio of calibrated recharge to the applied recharge as a function of basement slope angle θ . Numerical model parameters are; *specific yield* = 0.1, L = 13.3 km, Q_p of the numerical model = 0 to 1.86 meters-cubed per day per meter.

Figure 9 The effects of water level variations with respect to the saturated thickness on the recharge estimate is shown at $x/L = 0$. The recharge estimates were close to 1 when the amplitude is small compared to the saturated thickness, in which case the transmissivity is nearly constant in time. The recharge estimates were less accurate when the amplitude was larger compared to the saturated thickness, in which the transmissivity varied significantly in time. The numbers along the curves are the angle θ of the slope of the bottom basement boundary. Large θ resulted in large amplitude relative to the saturated thickness at $x/L = 0$.

The differences in the ratios of estimated to known recharge for different θ were a consequence of differences in the shape of the amplitude curves of the analytical and numerical models. Specifically, the analytical model assumed constant transmissivity, while the numerical model incorporated variable transmissivity in space and time. For small θ (nearly-uniform transmissivity), the shape of the amplitude distribution of the analytical model was most similar to that of the numerical model. For large θ (small transmissivity near the recharge boundary), the numerical model predicted a steep increase of amplitude near the recharge boundary because of decreased transmissivity near the recharge boundary (Figure 10). The amplitude of the calibrated analytical model was greater than that of the numerical model near the recharge boundary, and less at locations of $x / L \geq 0.07$.

Figure 10 Water level amplitudes of forward model of $\theta = 13$ degrees, $P = 16$ years, and water level amplitudes predicted by the analytical model using the effective transmissivity determined by calibration to amplitude data from all forward model cells. Use of other values of transmissivity results in large differences of predicted amplitudes, especially near the recharge boundary, which results in biases of the calibration procedure.

The transmissivity of the analytical model was calibrated to minimize the square error between the analytical- and numerical-model water levels. While some regional models use harmonic, geometric, or arithmetic means of transmissivity, the optimized transmissivity was very similar to the time-averaged transmissivity only at the first calibration point nearest the recharge boundary (Figure 11). Because the greatest amplitude response existed near the recharge boundary, the least error between the numerical and analytical models was produced when the value of the transmissivity of the analytical model was equal to the transmissivity of the numerical model at the

calibration location nearest the recharge boundary. Therefore, other locations further from the recharge boundary had less impact on the calibration. For example, a transmissivity value of 1.5 meters-squared per day (Figure 10) resulted in amplitude at $x/L = 0.1$ of 0.0 meters, while transmissivity of 5,000 meters-squared per day resulted in amplitude at $x/L = 0.1$ of 27.5 meters. The same transmissivity values at $x/L = 0.62$ resulted in a smaller range of amplitudes between 0.0 meters and 10.1 meters. Because all calibration points were weighted equally, the calibration produced an effective transmissivity that was biased to the transmissivity at the point nearest the recharge boundary.

Figure 11 Effective transmissivity for 4-, 8-, and 16-year periods and harmonic, geometric, and arithmetic averages of transmissivity of the numerical model for various numerical model geometries

4.2 MULTIPLE RECHARGE PERIODS

4.2.1 Singular Spectrum Analysis of Water Levels

The amplitudes of water-level variation used for calibration of the analytical model to infer the 4-, 8-, and 16-year periodic recharge components were determined by singular spectrum analysis of 50 years of the simulated water-level record using the SSA Toolkit. Before decomposition, the 100-year water-level variation was considered to be a trend within the 50 years of record and was removed by a low-order polynomial fit. The

de-trended water-level variations from $x/L = 0.62$ are compared to the 4-, 8-, and 16-year RCs (Figure 12). The 3-year and the random components were omitted intentionally to demonstrate that the individual 4-, 8-, and 16-year recharge components can be determined independently.

Figure 12 Water level variation and 4-, 8-, 16-year reconstructed components at $x/L = 0.0$

4.2.2 Calibration Data Location

The calibration procedure forced the fitted amplitude curve to match the two boundary end points exactly. The amplitude curve nearly matched the first variable calibration data point at either $x/L = 0.1, 0.2, 0.3$, or 0.43 . Figure 13 presents amplitude distributions of the calibrated analytical model of $\theta = 12.5$ degrees using a first calibration point at $x/L = 0.1, 0.2, 0.3$, or 0.4 , and a second calibration point at 0.62 . Because the calibration procedure was biased to the transmissivity at the location of the first calibration point, using different calibration data locations within in the numerical model of gradually variable transmissivity resulted in different calibrated effective transmissivities. The analytical model, calibrated for a single recharge component and using the same boundary conditions, but different calibration point locations, resulted in different effective transmissivity and recharge estimates. Therefore, the choice of calibration data controlled the fit of the analytical model to the observed data and the estimate of recharge.

Recharge estimates of each component of the composite recharge rate were inferred independently using the RCs of water levels for each component. The recharge estimates of the 16-year component within the complex recharge pattern are presented in figure 14. Recharge was approximately equal to the known recharge of the numerical model of $\theta = 0$ degrees for all calibration point locations (Figure 14a), but were most similar using a calibration location at $x/L = 0.2$. Recharge was most underestimated for the numerical model of $\theta = 12.5$ degrees (Figure 14b) using a calibration point at $x/L =$

0.1, and was less underestimated as the calibration point was at locations further from the recharge boundary to $x/L = 0.3$ and 0.43 .

Figure 13 Comparison of numerically simulated (solid curve) and calibrated analytical-model water levels of the 16-year RC using a calibration point nearest the recharge boundary at either $x/L = 0.1$, 0.2 , 0.3 , or 0.43 , and a point at $x/L = 0.62$. The differences between the observed and calibrated amplitudes using locations of $x/L = 0.1$, 0.2 , 0.3 , or 0.43 were -0.143 meters, -0.118 meters, 0.0250 meters, and 0.0708 meters, respectively.

a.

b.

Figure 14 Estimates of the 16-year recharge component decomposed from the composite recharge series with numerical-model geometries of $\theta = 0$ degrees (a) and 12.5 degrees (b).

The ratios between the estimated and known recharge rates of the 4-, 8-, and 16-year periods varied by location of the first calibration point for model geometries of $\theta = 0$ degrees and 12.5 degrees (Figure 15). No single calibration point produced the best recharge estimate for all components or model geometry. In general, the ratios of recharge for the numerical model of geometry $\theta = 0$ degrees were close to 1.0 for all calibration point locations because the uniform transmissivity was comparable to that used in the numerical model. The recharge ratios of the 4-year component were consistently equal to 0.98 for all calibration point locations, while the recharge ratios of the 8- and 16-year components were varied between 1.0 and 1.11. For the numerical model of $\theta = 12.5$ degrees, the recharge ratio increased as a function of distance of the calibration point location from the recharge boundary. The recharge ratios were smaller as the period length increased. Recharge was underestimated for all periods when the first calibration point was located at $x/L = 0.1$ and 0.2 . The 4- and 8-year recharge periods were overestimated when the point was at $x/L = 0.3$ and 0.43 . The 16-year period was underestimated in all cases. However, using points at $x/L = 0.2$ or 0.3 resulted in the closest recharge estimates for the 4- and 8-year period, and a point at $x/L = 0.43$ produced the closest estimate for the 16-year period. Considering that observation wells have been in place for several decades, these recharge estimates may be appropriate on the basis of limited spatial water level data to characterize the amplitude distribution along the flow path.

Figure 15 Analytical-model recharge estimate compared to the recharge specified in the numerical models. For geometry $\theta=0$ degrees, the ratio is near 1.0 for any calibration point location. For geometry $\theta=12.5$ degrees, the recharge estimate varied on the basis of the location of the first calibration point.

5.0 DISCUSSION

Time-varying recharge estimates can be obtained for well-characterized basins with continual and widely distributed water-level observations by calibration of a two- or three-dimensional numerical groundwater flow model. However, for basins with few water level records, and poorly characterized distributions of transmissivity, the application of the one-dimensional analytical model by *Townley* [1995] may be used to infer time-varying recharge, provided a one-dimensional model with periodic boundary conditions can approximate the flow system.

The *Townley* [1995] model can be used to estimate aquifer properties from water-level responses to periodic-recharge boundary conditions, while previous methods such as the classical aquifer test methods rely on constant flow rates to determine aquifer properties or constant boundary conditions. The examples in this paper represent cyclostationary aquifer flow responses to periodic natural recharge rates owing to climate variability. Thus this method can be used to infer recharge rates in real basins that vary with period lengths and magnitudes similar to those in the examples.

The shape of the distribution curve of water-level amplitudes along the flow path allows for calibration of the analytic solution for periodic recharge using the non-dimensional response time [*Townley*, 1995] of the periodic water-level changes at multiple distances from the recharge boundary. The response time, $L^2 S / TP$, is a unique parameter that controls the shape of the amplitude distribution curve of the analytical model. Small response time, caused by large T or P in the presented examples, resulted in water level amplitudes that penetrated far from the recharge boundary. Large response

time, caused by small T or P , resulted in water-level amplitudes that attenuated quickly from the recharge boundary. The analytical model was calibrated by fitting the shape of the water level amplitudes by optimizing the transmissivity using known values of length between boundaries (L), storage coefficient (S), and period (P) from the numerical model. Thus, the analytical model was calibrated to the numerical model by optimizing one component of the response time.

This inverse application of *Townley's* [1995] solution could also be used to determine an effective S at the location of the first calibration point if L , P , and T are known. Because the optimized *Townley* [1995] model was biased toward parameter values at the first calibration point, a single value of T known for the first point may be considered an effective T for the flow path. A value of T may be obtained at the first observation well from well log data of aquifer thickness and hydrologic conductivity measurements of core samples. Thus, S can be determined by optimizing the response time.

Periodic recharge on flow systems results in periodic changes in water levels with amplitudes that attenuate away from the recharge boundary. The rate of attenuation of amplitudes for the simulations was dependent on specific yield, transmissivity, and period. Considering actual values of basin aquifers, rapid attenuation may occur in cases of large specific yield or small transmissivity at the recharge boundary (large θ), or small recharge period as discussed by *Heitjema* [1995]. In such cases, the monitoring wells should be close to the recharge boundary so that fluctuations of water levels are measurable. Attenuation may be gradual for small specific yield, large transmissivities at

the boundary, or long recharge period, in which fluctuations may be measurable in monitoring wells that are located midway between boundaries. However, if the water level response approaches steady-state conditions for the maximum and minimum recharge, as when transmissivity equals 3,000 meters-squared per day in figure 10, the amplitude distribution becomes linear between boundaries. In this situation, the simulation reaches steady-state condition at the maximum and minimum of recharge rates. Such water-level amplitudes are not unique to transmissivity or recharge rate and should not be used for calibration. In this case, prior knowledge of the transmissivity at the first calibration point may serve as a surrogate to the calibration process.

The water levels along the flow path should be well distributed between the inflow and outflow boundaries to adequately characterize the distribution of water level amplitudes. An accurate time series data set is desired near the recharge area in order to confidently specify a periodic head boundary using the amplitude of the RCs. The accuracy of this location is critical because the largest gradients in periodic head fluctuations exist near the boundary, which control the amount of calculated flow into the basin. High accuracy of the water level measurements further into the basin are important because the transmissivity is calibrated using the distribution of the amplitudes.

6.0 CONCLUSIONS

A methodology was developed for the estimation of time-varying recharge and was evaluated using a numerical model that served as a test case of typical alluvial aquifers of the southwest United States. The analytical model was evaluated for various aquifer geometries and for single- and multiple-period recharge rates and periods that are similar to expected climate forcings. Depending upon the domain geometry, the time-averaged transmissivity used in the numerical models was either nearly uniform or it decreased gradually toward the recharge boundary at the mountains.

The results demonstrate that periodic, time-varying recharge rates can be estimated from hydrologic time series such as long-term water level records. The ratios of estimated to known recharge were 0.97 when the basement is non-sloping, and decreased to 0.92 if the basement slopes at 13 degrees from the recharge boundary to the midpoint of the basin cross-section. That is, the analytical model produced more accurate recharge estimates of the individual components for the idealized basin with nearly uniform transmissivity. These accuracies compare well to the accuracies of other hydrologic measurements such as stream-gage measurements of stream flow.

The SSA Toolkit [Dettinger *et al.*, 1995] was used to reconstruct periodic components of water levels from two numerical models for differing domain geometries. The analytical model was calibrated using the amplitude of the reconstructed components at two locations, representing two monitoring wells. Ratios between the estimated and known recharge were approximately 1.0 for all individual recharge components for nearly uniform transmissivity, thus the locations of the monitoring wells were not

important. For the model of gradually decreased transmissivity toward the mountains (sloping basement), the ratios were less than 1.0 for all periods using a calibration point at $x/L = 0.1$ (the monitoring location was located one-tenth of the distance from the mountains to the center of the valley). The ratio was approximately 1.0 for the 4- and 8-year periods using calibration points at $x/L = 0.2$ or 0.3 . The 16-year period was underestimated for all calibration locations tested. Therefore, no single location of monitoring well calibration data produced the most accurate recharge estimates of all periods for the model representing a basin with a sloping basement. However, all of the monitoring locations gave reasonable estimates of multi-component, time-varying recharge.

The *Townley* [1995] model used in this study assumed one-dimensional flow conditions that are likely to be a simplification of groundwater flow in real environments. However, these simplifications may be necessary in cases of limited model input and calibration data. The extension of the *Townley* [1995] model by *Trefry* [1999] to continuous subaquifers of homogenous properties may be applied to basins with high characterization of aquifer properties that can be defined along the flow path. In cases where the assumption of one-dimensional flow is not valid, a two-dimensional formulation of the *Townley* [1995] model can be developed and calibrated to periodic water level data distributed two-dimensionally. Efforts to develop analytical solutions for three-dimensional periodic flow in unconfined aquifers have not been successful.

The methodology tested in this investigation can be applied to basins similar to those represented by the examples. Specifically, these results apply to basins that include

a large component of inflow at the mountain fronts and minimal convergent and divergent flow toward discharge at a constant head boundary. To extend the applicability of this approach, the boundary conditions of the model may be reassigned to be more suitable to basins with different geometries or hydrogeologic frameworks. For example, a head dependent boundary may be used at the stream and a specified flow boundary may be assigned at the mountain front. Long term water levels along a flow path are needed that can be spectrally analyzed for periodic components. The *Townley* [1995] model can be calibrated to basins with prior knowledge of L from groundwater flow maps and P from the spectral analysis, but unknown values of parameters T and S . These parameters, whose ratio T/S is often termed the aquifer diffusivity, can be determined concurrently by calibration of the *Townley* [1995] model for non-dimensional response time $L^2 S/TP$.

Further study may include the assessment of the effects of time-varying recharge estimates on existing calibrated numerical models of regional groundwater flow. The effects can be assessed by comparing simulated time-varying water levels to measured variability in observed water levels. Although the time-varying recharge estimates will have more detail in time than the calibrated rates, the averages of both rates over identical stress periods should be similar.

In practice, the analysis presented may be a useful water supply and demand analysis tool. Time-varying recharge estimates and reconstructed components of water levels may be correlated to the timing of water consumption rates and containment into reservoirs. The impact of recharge variability on time-dependent water resource decisions may be used to assess the efficiency of current policies and practices.

The methodology presented allows for simplified estimates of historical cyclical recharge in basins where monitoring well records of high temporal resolution (e.g. monthly) are available. Although recharge is estimated for a sub-region of a basin, these estimates may be adequate to systematically characterize historical or future variable recharge rates for similar environments on a regional basis [eg. *Hanson et al.*, 2002]. Incorporation of interannual to interdecadal temporal resolution of periodic recharge within regional groundwater flow models may improve simulations of long-term groundwater development that are subject to periodic natural recharge driven by climatic variability.

7.0 REFERENCES

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