Application of a Tidal Method for Estimating Aquifer Diffusivity: Swan River, Western Australia

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by

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ABSTRACT

A tidal method for estimating aquifer diffusivity is applied to water level data collected during a study of surface water-groundwater interaction between the Swan River, Western Australia, and the adjacent shallow groundwater system. Harmonic analysis of tidal fluctuations in the river and water level responses in nearby groundwater monitoring bores reveals that the dominant diurnal tidal frequencies (K1 and O1 constituents) propagate more than 340 m into the aquifer. Although the shallow groundwater system is considered to be regionally unconfined, the estimates of diffusivity \((T/S)\) obtained in this study suggest that near to the river in the study area the aquifer is locally confined or semi-confined.

Diffusivity values obtained from an analysis of tidal efficiency and lag vary in the range 13,500 to 268,000 m\(^2\)day\(^{-1}\). Assuming an aquifer transmissivity of \(T = 600\) m\(^2\)day\(^{-1}\), the aquifer storage coefficient is estimated to be between 0.002 and 0.044, which is characteristic of confined groundwater. It is improbable that, within a twenty-four hour tidal cycle, groundwater can flow the distances required to induce a response at the monitoring bores. A more likely explanation is that the response arises from elastic compression and expansion of the aquifer under confined conditions.
MATHEMATICAL NOTATION

Variables

\(F\)  Tidal forcing function (general)
\(H\)  Head at a prescribed tidal boundary \([L]\)
\(P\)  Period \([T]\)
\(R\)  Tidal response function (general)
\(S\)  Aquifer storage coefficient \([1]\)
\(S_o\) Aquifer specific storativity \([L^{-1}]\)
\(T\)  Aquifer transmissivity \([L^2T^{-1}]\)
\(t\)  Time \([T]\)
\(x\)  Horizontal spatial coordinate \([L]\)
\(y\)  Time series (general)
\(\delta t\) Time-series sampling interval \([T]\)
\(\Phi\) Fourier coefficient (complex)
\(\omega\) Angular frequency \([T^{-1}]\)
\(\nu\) Frequency \([T^{-1}]\)

Subscripts

\(a\)  Amplitude
\(i\)  Discrete value
\(j\)  Fourier harmonic
\(k\)  Tidal constituent
\(N\)  Nyquist harmonic
\(\theta\) Phase
INTRODUCTION

This report describes the application of a tidal method (Ferris, 1951) to estimate aquifer diffusivity at a study site on the Swan River in Western Australia (Figures 1 and 2). The work was initiated following the discovery that tidal responses could be detected in groundwater monitoring bores more than 340 m from the river. This indicated the possible confinement of shallow groundwater in the study area, which was contrary to the common assumption of an unconfined shallow aquifer system. An opportunity was presented to obtain estimates of the aquifer hydraulic properties by analysing tidal propagation in the aquifer.

The work described forms one component of a detailed study of surface water-groundwater interaction along a 20 km reach of the Swan River, between the Causeway and Guildford. This main body of research is reported by Linderfelt et al. (unpublished), Turner et al. (submitted) and Smith and Turner (submitted).

Figure 1 Location Map
Figure 2 Monitoring bore locations at Ron Courtney Island and Tonkin Overpass
THE TIDAL METHOD

The tidal method (Ferris, 1951; Carr and Van Der Kamp, 1969; Townley, 1995) is a simple technique for estimating aquifer diffusivity \( T/S \) based upon the response of an aquifer to tidal forcing at a boundary. Similar to conventional pump test techniques, in which the piezometric head response of an aquifer to artificial pumping is used as a basis for estimating aquifer hydraulic properties, the tidal method takes advantage of natural tidal forcing and obtains an estimate of diffusivity from the attenuation of a tidal signal as it propagates into an aquifer. The tidal signal can be expressed as a linear combination of sinusoidal terms or tidal constituents (Forman and Henry, 1989, p.109) that are differentially attenuated as they travel though the aquifer; the degree and rate of attenuation depends upon the aquifer hydraulic properties. There are two possible mechanism for tidal propagation: (1) flow of tidal water into and out of an unconfined aquifer, (2) compression and expansion of a confined aquifer due to the load of the incoming and outgoing tide. In an aquifer that has tidal forcing at a lateral boundary, the attenuation of the tidal signal is normally described by the two quantities tidal efficiency and lag.

Tidal efficiency, \( TE \), is a normalised measure that relates the amplitude of head fluctuations in the aquifer to the amplitude of fluctuations at the tidal boundary. In general,

\[
TE_k = \frac{R_{ak}}{F_{ak}}
\]  

(1)

where \( F_{ak} \) is the amplitude of forcing at the tidal boundary, \( R_{ak} \) is the amplitude of the response at a point in the aquifer and \( k \) denotes the tidal constituent or frequency.

Lag is a measure of the speed of propagation of a tidal constituent as it moves through the aquifer

\[
lag_k = F_{ak} - R_{ak}
\]  

(2)

where \( F_{ak} \) is the phase of tidal forcing and \( R_{ak} \) is the phase of the tidal response at a point in the aquifer.

Ferris (1951) presents analytical expressions that describe the tidal efficiency and lag in a one-dimensional, semi-infinite and homogeneous aquifer with uniform transmissivity (see
also Carr and Van Der Kamp, 1969). These are derived by solving the equation for 1D-transient groundwater flow, with a tidal or harmonic boundary condition at $x = 0$

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

(3)

$$h(0,t) = H_a \cos(\omega t - H_0)$$

(4)

where $H_a$ is the amplitude of head fluctuations at the tidal boundary [L] and $H_0$ is the phase measured in radians. Since (3) and (4) are both linear, the groundwater flow problem they describe can be decomposed into a steady-state flow problem and a harmonic flow problem that comprises one or more frequencies. The harmonic solution is rearranged to yield the following expressions for tidal efficiency and lag

$$TE_k = \exp\left(-x \frac{\pi S}{\sqrt{TP_k}}\right)$$

(5)

$$\text{lag}_k = x \frac{SP_k}{4\pi T}$$

(6)

Here, $x$ is the distance from the tidal boundary, $S$ is the aquifer storage coefficient [1], $T$ is a uniform aquifer transmissivity [L2T-1], $P_k$ is the period [T] of the tidal constituent $k$ and $\omega_k = 2\pi/P_k$ is the corresponding angular frequency [T-1]. Since a particular tidal signal may comprise several dominant frequencies, e.g. semi-diurnal, diurnal and monthly constituents, several independent estimations of diffusivity are possible from a single tidal signal.

From (5) and (6), efficiency-based and lag-based expressions for aquifer diffusivity are straightforward to derive

$$\frac{T}{S} = \frac{\pi x^2}{(\ln TE_k)^2 P_k}$$

(7)

$$\frac{T}{S} = \frac{x^2 P_k}{4\pi (\text{lag}_k)^2}$$

(8)

Equating (7) and (8) gives the non-dimensional relationship
\[
\frac{\text{lag}_k}{P_k} = -\left(\frac{1}{2\pi}\right) \ln TE_k
\]

where the left-hand term is a normalised lag. It follows that, on a semi-log plot, the normalised lag and tidal efficiency are related by the straight line with slope equal to \(-1/2\pi\). The tidal efficiency decreases exponentially with increasing distance from the tidal boundary, whereas the lag increases linearly.

**DISCRETE FOURIER ANALYSIS**

The term Fourier analysis refers to any data analysis technique that describes or measures fluctuations in a time series by comparing them with sinusoids. Well known methods include filtering, least squares regression on sinusoids and harmonic analysis (Bloomfield, 1976; Chatfield, 1975; James, 1995). Fourier techniques are commonly used to detect frequency components that are hidden within a 'noisy' signal.

Harmonic analyses fits uniformly sampled time series data to a set of harmonic frequencies; that is, a set of frequencies that are integer multiples of one another. The fitted harmonics are normally in the frequency range \(0 \leq \nu \leq \nu_N\), where

\[
\nu_N = \frac{1}{P_N} = \frac{1}{2\delta t}
\]

is the Nyquist frequency \([T^{-1}]\), \(P_N\) is the Nyquist period \([T]\) and \(\delta t\) is the data sampling interval \([T]\). The Nyquist condition simply states that the smallest period that can be reasonably fitted to the data is equal to twice the sampling interval. This corresponds to fitting, at most, one local maxima or minima between each pair of data point in the time series. From a practical point of view, the Nyquist conditions tells us that it is unreasonable to fit, say, a semi-diurnal frequency (12 hour period) to data that has only been sampled 24 hourly. The smallest period or highest frequency that should be fitted to diurnal data is 48 hours or 0.5 cycles per day.

**Discrete Fast Fourier Transform (FFT)**

The discrete FFT is a computationally efficient algorithm that enables regularly sampled time series data to be fitted by a set of harmonic frequencies (Chatfield, 1975; Bloomfield, 1976; James, 1995). For a time series of \(n\) data values, the FFT fits \(n/2\) frequencies over the
frequency range \( \frac{2v_n}{n} \leq v \leq v_n \). The FFT of the time series is the set of \( n/2 \) complex valued Fourier coefficients \( \Phi(v_j) \), where the moduli \( \left| \Phi(v_j) \right| \) are equal to the amplitudes of the fitted frequencies and arguments the \( \arg(\Phi(v_j)) \) are equal to the phases.

The Inverse Fast Fourier Transform (IFFT) is a reverse operation that converts a set of Fourier coefficients \( \Phi(v_j) \) back to the original time-series data \( y(t_j) \). This relationship is denoted symbolically as (James, 1995)

\[
y(t_j) \leftrightarrow \Phi(v_j)
\]  

(11)

A measure of the energy or power at each FFT frequency is provided by the Power Spectral Density (PSD) function

\[
S(v_j) = \Phi(v_j) \Phi^*(v_j)
\]  

(12)

Here, \( \Phi^*(v_j) \) denotes the complex conjugates of the Fourier coefficients. A large value of \( S(v_j) \) indicates more energy at frequency \( v_j \). This means that the dominant frequencies in a time series are represented as larger spikes on a PSD plot.

**FREMANTLE TIDE DATA**

The Swan and Canning river estuary system is tidally forced and in permanent contact with the Indian Ocean at the Port of Fremantle. Fremantle experiences a mixed tide that is characterised by diurnal (24 hour) and semi-diurnal (12 hour) constituents. Harmonic analysis of 1996 tidal data for Fremantle Fishing Boat Harbour (Department of Transport, Western Australia, unpublished) indicates that the K1 (period = 23.9345 hours) and O1 (period = 25.8193) diurnal tidal constituents are the dominant frequencies. The semi-diurnal frequencies vanish for part of each month, which make them unsuited to the present analysis.

**SWAN RIVER TIDE DATA**

Tide data collected in the Swan River near to the Ron Courtney Island and Tonkin Overpass transects (Figure 1) can be compared with Department of Transport tidal data for the Fremantle Fishing Boat Harbour and Barrack Street jetty at Perth. For consistency, all data sets are standardised to an hourly sampling interval; the corresponding 12 cycles per day Nyquist frequency is adequate to detect the expected diurnal or 1 cycle per day tidal constituents. The number of harmonics fitted in each analysis depends upon the number of
data values in the time series being considered. Since the discrete FFT fits \( n/2 \) harmonics to \( n \) data points, a larger number of data points result in more frequencies and a higher resolution PSD plot. When a standardised sampling interval is used, the number of data points, and hence Fourier frequencies, is determined by the total length of the time series.

Foreman and Henry (1989, Table 1) specify that the minimum total length of time series required for separation of the K1 and O1 tidal constituents in a harmonic analysis are 24 hours and 328 hours (13.67 days), respectively. These criteria are exceeded by all data sets used here. Harmonic analyses are performed using MATLAB (The Math Works Inc., 1995) by first detrending the data, and then implementing MATLAB’s intrinsic FFT function. The PSD function is calculated as \( S(v_j) = \Phi(v_j) \Phi^*(v_j)/n \).

Figures 2a and 3a depict water levels fluctuations recorded in the Swan River at Barrack Street jetty and Tonkin Overpass, and Figures 2b and 3b present the corresponding PSD plots. The PSD or 'spectral signatures' of the data sets are consistent and confirm that the K1 and O1 tidal constituents are the dominant frequencies. The smaller peaks at semi-diurnal frequencies represent harmonics of the K1 and O1 constituents, and are characteristic of the FFT rather than the data. In general, these harmonics represent the non-sinusoidal character of the dominant frequencies (Chatfield, 1975, p.132). Peaks at lower frequencies \( (v_j \leq 0.25) \) correspond to the passage of weather systems and associated fluctuations in barometric pressure; in the river, there may also be 'piling up' of water at the shoreline due to prevailing winds. These effects are non-stationary, however, and are complicated by the fact that barometric pressure fluctuation can effect the aquifer storage directly through atmospheric loading, as well as indirectly through tidal loading. They are therefore not suited for use with the tidal method.

Figure 5 depicts piezometric head fluctuations and PSD for the monitoring bore CRB22, which is located approximately 120 metres from the Swan River shoreline at Ron Courtney Island (see Figure 2 for location). The estuary spectral signature is evident at the monitoring bore and indicates that diurnal tidal forcing can propagate a significant distance into the aquifer. This suggests partial or semi-confinement of superficial groundwater in the study area.

Contrasting results are obtained for monitoring bore CTB22 (Figure 6). Though located a similar distance onshore from the river (approximately 200 metres), the estuary tidal signature is not detectable in the measured water levels. There is no separation of the K1 and O1 tidal constituents and barometric fluctuations are not apparent; these longer-period oscillations should propagate furthest from the river since they are less attenuated by the
The dominant diurnal and semi-diurnal peaks that are evident in Figure 4 are believed to be a result of nearby groundwater pumping, most likely for irrigation.

Within the study area, it seems likely that groundwater confinement occurs locally and is spatially variable. Nonetheless, the tidal spectral signature is evident in monitoring bores located more than 340 metres from the river. The most likely explanation is that river tidal fluctuations are propagated via elastic compression and expansion of the aquifer under confined conditions. It is improbable that groundwater can flow hundreds of metres into and out of the aquifer within a twenty-four hour tidal cycle.

Only two of the ten groundwater-monitoring locations in the study area, CRB2 and CRB3, are suitable for applying the tidal method. At each site there are three monitoring bores: one shallow bore screened just below the water table, one deep bore screened at the base of the aquifer, and one intermediate bore screened midway between the deep and shallow bores. At the other monitoring sites, either the bores are too close to the river shoreline to provide meaningful estimates of diffusivity (CRB1, CRB4, CRB5, CTB1, CTB3) or the tidal signal is indiscernible (CRB6, CTB2, CTB4). The five monitoring sites close to the river all have strong tidal responses, but applying a tidal analysis would only yield diffusivity estimates for small sections of the river bank, between the river and monitoring bores. Such estimates are not considered useful for the present assessment. At CRB6, CTB2 and CTB4, the tidal response—if one occurs—is apparently masked by groundwater pumping from local production bores that also has a strong diurnal component.
Figure 3  Tidal data and Power Spectral Density for Barrack Street Jetty, Perth

Figure 4  River tidal data and Power Spectral Density for Tonkin Overpass
Application of a Tidal Method for Estimating Aquifer Diffusivity

Figure 5  Piezometric head data and Power Spectral Density for groundwater monitoring bore CRB22; located approximately 120 metres from the Swan River shoreline

Figure 6  Piezometric head data and Power Spectral Density for groundwater monitoring bore CTB22; located approximately 200 metres from the Swan River shoreline
ESTIMATION OF AQUIFER DIFFUSIVITY

Results

At each of the six monitoring bores, four estimates of aquifer diffusivity can be made by applying the tidal method to the K1 and O1 tidal constituents; each constituent yields an efficiency-based and lag-based estimate of diffusivity. Results for the monitoring locations CRB2 and CRB3 are summarised in Tables 1 and 2.

Tidal efficiency and lag are calculated from FFT’s of the river tidal data and groundwater level responses in the monitoring bores. At each bore

$$TE_{K1} = \frac{\Phi(v_{K1})_{bore}}{\Phi(v_{K1})_{river}}$$

$$TE_{O1} = \frac{\Phi(v_{O1})_{bore}}{\Phi(v_{O1})_{river}}$$

(13)

and

$$lag_{K1} = \text{arg}\left(\frac{\Phi(v_{K1})_{river}}{\Phi(v_{K1})_{bore}}\right)$$

$$lag_{O1} = \text{arg}\left(\frac{\Phi(v_{O1})_{river}}{\Phi(v_{O1})_{bore}}\right)$$

(14)

Diffusivity is then calculated from equations (7) and (8) using the results from (13) and (14). Estimates of the aquifer specific storativity $S_o$ are also presented based upon an assumed transmissivity $T = KB = 600 \text{ m}^2/\text{d}$; where $B = 15 \text{ m}$ is the saturated thickness of aquifer and $K = 40 \text{ m/d}$ is the horizontal hydraulic conductivity. The values chosen are based upon stratigraphic interpretation of bore logs from the study site and calibration of a 2D-horizontal flow model (Linderfelt et al., unpublished).

Discussion

The specific storativity values in Tables 1 and 2 vary in the range 0.003 to 0.0001 m$^{-1}$ (average 0.0008 m$^{-1}$), which is indicative of confined or semi-confined groundwater. For a 15 metre thick aquifer, the aquifer storage coefficient $S = S_oB$ is calculated to be between 0.044 and 0.002 (average 0.012). Unconfined aquifers normally have a storage coefficient or effective porosity between 0.1 and 0.3. Supposing an average specific storativity value of 0.0008 m$^{-1}$, as calculated above, and a storage coefficient between 0.1 and 0.3, then the aquifer
transmissivity would need to be between 9,430 and 28,300 m$^2$d$^{-1}$ to satisfy the results of the tidal analysis. In a 15 metre thick aquifer, this equates to a hydraulic conductivity between 628 and 1,890 m d$^{-1}$. Such a large value of transmissivity is considered unrealistic. The calibrated value of horizontal hydraulic conductivity from regional-scale 2D-plan modelling is 40 m d$^{-1}$ (Linderfelt et al., unpublished) and is characteristic of medium to fine sand (Bouwer, 1978). The existence of dense clay layers, as described in the monitoring bore stratigraphic logs, provides additional evidence that the aquifer contains confining beds. Tables 1 and 2 also indicate that the degree of confinement increases with depth, as would be expected in a layered interbedded system. Between the shallow and deep bores the tidal efficiencies increase and the lags decrease.

The estimates of diffusivity and specific storativity discussed above are considered to be order of magnitude values that should be interpreted in light of the assumptions of the tidal method; that is, 1D flow, homogeneous aquifer and uniform transmissivity. The aquifer is known to be inhomogeneous in the study area, and comprises a complex sequence of interbedded sand, silt and clay made up of river channel fill and estuarine deposits. Close to the river the assumption of 1D flow is also improbable: firstly, groundwater from deeper in the aquifer must move upward to discharge at the riverbed, implying significant vertical flow; secondly, the horizontal pattern of groundwater flow is strongly influenced by the river meander geometry. In fact, the river meander pattern induces horizontal and vertical variation in the flow due to groundwater converging toward the outside of meander bends, which protrude further into the regional flow system (Linderfelt et al., unpublished; Smith and Turner, submitted).

Finally, Figure 7 presents a semi-log plot of normalised lag verses tidal efficiency, on which the estimated Swan River values from Tables 1 and 2 are compared against the theoretical relationship from equation (9). In general, there is evidence of a logarithmic relationship in the study site data, though the slope is clearly different to that of the theoretical model ($-1/2 \pi$). This difference is most likely a consequence of the modelling assumptions and reinforces the view that the calculated diffusivity values should be treated as order of magnitudes estimates.
### Table 1 Summary of tidal analysis for K1 tidal constituent

<table>
<thead>
<tr>
<th>Bore ID</th>
<th>Distance to River (m) (closest point on shoreline)</th>
<th>$TE_{K1}$</th>
<th>Efficiency-based Diffusivity ($m^2$/day)</th>
<th>$S_o$  $B=15m$ $K=40md^{-1}$</th>
<th>$lag_{K1}$</th>
<th>Lag-based Diffusivity ($m^2$/day)</th>
<th>$S_o$  $B=15m$ $K=40md^{-1}$</th>
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</thead>
<tbody>
<tr>
<td>CRB21 (deep)</td>
<td>120</td>
<td>0.544</td>
<td>113,000</td>
<td>0.0004</td>
<td>0.075</td>
<td>217,000</td>
<td>0.0002</td>
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<tr>
<td>CRB22 (int.)</td>
<td>120</td>
<td>0.377</td>
<td>44,200</td>
<td>0.0009</td>
<td>0.120</td>
<td>84,900</td>
<td>0.0005</td>
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<tr>
<td>CRB23 (shallow)</td>
<td>120</td>
<td>0.378</td>
<td>44,400</td>
<td>0.0009</td>
<td>0.116</td>
<td>91,300</td>
<td>0.0004</td>
</tr>
<tr>
<td>CRB31 (deep)</td>
<td>346</td>
<td>0.047</td>
<td>37,400</td>
<td>0.0011</td>
<td>0.367</td>
<td>76,200</td>
<td>0.0005</td>
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<tr>
<td>CRB32 (int.)</td>
<td>346</td>
<td>0.045</td>
<td>36,400</td>
<td>0.0011</td>
<td>0.326</td>
<td>96,500</td>
<td>0.0004</td>
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<td>CRB33 (shallow)</td>
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<td>0.012</td>
<td>17,900</td>
<td>0.0022</td>
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### Table 2 Summary of tidal analysis for O1 tidal constituent

<table>
<thead>
<tr>
<th>Bore ID</th>
<th>Distance to River (m) (closest point on shoreline)</th>
<th>$TE_{O1}$</th>
<th>Efficiency-based Diffusivity ($m^2$/day)</th>
<th>$S_o$  $B=15m$ $K=40md^{-1}$</th>
<th>$lag_{O1}$</th>
<th>Lag-based Diffusivity ($m^2$/day)</th>
<th>$S_o$  $B=15m$ $K=40md^{-1}$</th>
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<tr>
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<td>0.530</td>
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<td>268,000</td>
<td>0.0001</td>
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<td>CRB22 (int.)</td>
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<td>0.0008</td>
<td>0.076</td>
<td>211,000</td>
<td>0.0002</td>
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<td>CRB23 (shallow)</td>
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<td>47,900</td>
<td>0.0008</td>
<td>0.076</td>
<td>211,000</td>
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<td>CRB31 (deep)</td>
<td>346</td>
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<td>CRB33 (shallow)</td>
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<td>0.006</td>
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<td>0.0030</td>
<td>0.531</td>
<td>36,300</td>
<td>0.0011</td>
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</table>
Figure 7  Non-dimensional plot of normalised lag verses tidal efficiency for groundwater monitoring bores CRB2(1-3) and CRB3(1-3)
REFERENCES


