Aquifer Storage Properties from Groundwater Fluctuations induced by Seismic Rayleigh Waves

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Abstract

In confined aquifers, observation wells can be sensitive to large global earthquakes and commonly produce hydroseismograms with centimeter scale water level fluctuations. Regional broadband seismographs can generally provide excellent estimates of Rayleigh wave (LR) displacement. To account for the sparse data collection interval of most well recorder data compared to the seismic data, we propose that the mean squared water level fluctuation as measured in the time domain in each well be compared to the mean spectral density (after adjusting to the wavelengths) of the passing Rayleigh wave.

For each individual well, Specific Storage (S_s) can then be calculated from the ratio of these quantities. For aquifers with multiple observation wells, this provides a method to estimate storage properties from an event based stress that can occur uniformly over the areal extent of the aquifer.

Project Objectives

1. Devise a method to estimate storativities using earthquake induced water level fluctuations sampled at five-minute intervals.

2. Estimate the storativities for the observational wells in the Grande Ronde Aquifer of the Columbia River Basalt Group for comparison with previous estimates from aquifer tests and barometric efficiencies (Table 1).

Background

If, after large earthquakes, complete time sequences of LR displacements and groundwater fluctuations are available, the S_s of a confined aquifer can be found. The theoretical basis of the method is the following equation (Shih, 2009):

h(t) = (1 / Ss) (5.77 w(t) / λ) [1]

where: t is time, w(t) is the vertical displacement of the Rayleigh wave of wavelength λ, and h(t) is the water head fluctuation due to the passing of the wave.

However, in practice, the situation is not so simple. The passing Rayleigh waveform is a combination of many wavelengths. With phase velocity being dependent on frequency, one cannot simply compare a single head deflection measurement with a simultaneous LR displacement in the time domain. In practice, spectral methods need to be employed to get useful results.

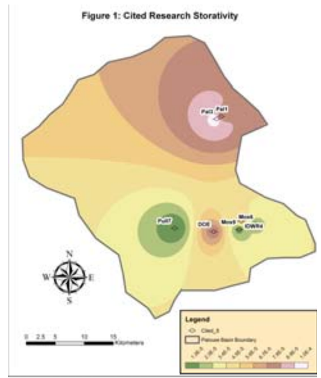
Shih (2009), for example, used the cross spectral density of the two data sets to identify a narrow frequency band of highest coherence. Then, only the autospectral densities and seismic wavelength at that narrow frequency band were used to calculate the storage. The key to Shih's method is having complete spectra of both time series available so that the most coherent period can be identified.

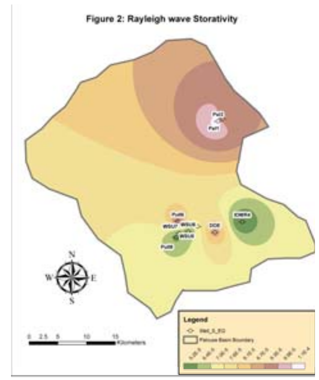
However, the sparse amount of hydraulic head data provided by the most well recorders relative to seismological data makes the direct application of Shih's method impossible because the autospectral density of the water deflection cannot be calculated. We suggest a novel approach that makes use of the mean-squared water level deflection as measured at whatever sampling period is available during the passage of the Rayleigh wave.

Table 1: Aquifer S from Barometric Efficiency	
Porosity	400 m. Thickness
0.05	8.9×10^{-5}
0.1	1.8×10^{-4}
0.15	2.7×10^{-4}

Data Requirements

- A complete time sequence of the LR wave displacements from which the $E\{A_k\}$ can be calculated.
- Knowledge of LR wave phase velocities so wavelengths can be calculated.
- Figure 4 is an example of the sufficient number of water level deflection measurements during the passage of the LR wave to get a reasonable estimate the mean squared deflection.





Methodology

Taking the Fourier Transform *F* of each side of equation [1]

$$F\{ h(t) \} = (1 / Ss)\ F\{5.77\, w(t) / \lambda\} \qquad [2]$$

Now find the complex conjugate of [2].

$$F^*\{ h(t) \} = (1 / Ss)\ F^*\{5.77\, w(t) / \lambda\} \qquad [3]$$

Multiplying [2] times [3], and dividing by the number of samples N in the Rayleigh wave time sequence:

$$F\{ h(t) \}\ F^*\{ h(t) \} / N = (1 / Ss^2)\ A_k$$

where: $A_k = F\{ (33.3 / \lambda^2)\ w(t) \}\ F^*\{ (33.3 / \lambda^2)\ w(t) \} / N$ [4]

The left side of [4] is the autospectral density of h(t) (Figure 3); the right side, exclusive of the Ss term, is A_k, the spectral density of w(t) adjusted for the wavelength and the constant.

However, by Parseval's Theorem, the area under an autospectral density curve is equal to the sum of the squared deviations of the entire time sequence. Dividing both sides of [4] by N yields:

$$E\{ h^2(t) \} = (1 / Ss)^2\ E\{ A_k \} \qquad [5]$$

where: the expected value operator *E* represents an average value over a sample interval.

Our formula for a storage estimate becomes:

$$\mathbf{Ss} = [\ E\{ A_k \} / E\{ h^2(t) \}\ ]^{1/2} \qquad [6]$$

Thus by comparing time domain measurements of water level fluctuations to spectral measurements of the Rayleigh wave displacement, an estimate of specific storage can be obtained for each observed seismic event. This provides data points for the use of the Arc GIS 9.0 inverse distance weighting tool to get an extrapolated storativity map for the basin(Figures 1 and 2).

Figure 3: Displacement and Autospectral Density

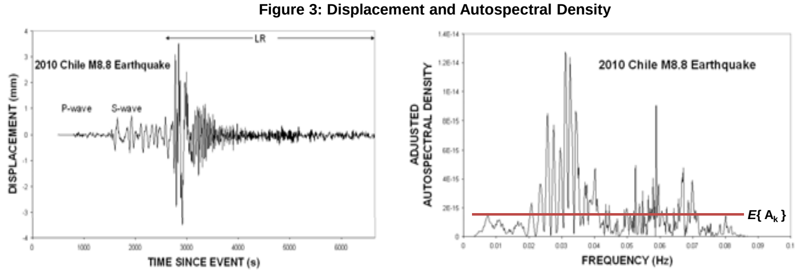
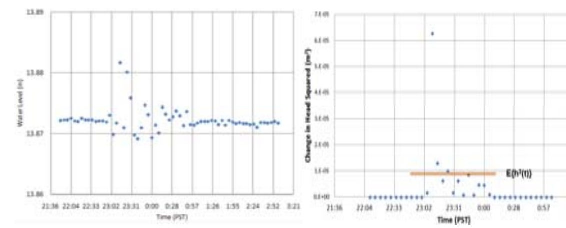


Figure 4: M8.8 Chile Earthquake signature in WSU5 well (Figure 2) and Mean Squared Water Level Deflection ($E\{h^2(t)\}$)



Results and Conclusions

- Estimation of specific storage can be accomplished by comparing sparse time domain measurements of water level fluctuations to spectral measurements of the Rayleigh wave displacement.
- Standard statistical techniques can be employed to estimate the uncertainty in the mean squared water level deflection which will be the biggest source of error due to the sparseness of the data from most water level records.
- The generated storage coefficients can be compared to previous estimates and used as additional evidence.
- As suggested in Table 2, our results confirm that our method provides estimates of storativity consistent with the very low values found by other methods in this confined aquifer.

Table 2: Comparison of the Methods			
Previous Research	Well	Storativity from Aquifer Tests	Storativity from Rayleigh waves
Moran 2011	Palouse 1	6.7×10^{-5}	$8.4 \times 10^{-5} \pm 5.0 \times 10^{-5}$
Ralston 2000	Palouse 3	1.0×10^{-4}	$1.1 \times 10^{-4} \pm 1.6 \times 10^{-4}$
Fielder 2009	IDWR4	5.2×10^{-5}	$5.2 \times 10^{-5} \pm 3.8 \times 10^{-5}$
McVay 2007	DOE	7.5×10^{-5}	$8.3 \times 10^{-5} \pm 6.1 \times 10^{-5}$

Acknowledgements

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References

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Did the 2011 Japan Earthquake Cause a "Groundwater Tsunami" in Moscow, Idaho?

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In confined aquifers, observation wells can be sensitive to large global earthquakes and commonly produce hydroseismograms with centimeter scale water level fluctuations. The travel time for the magnitude 9.0 Japanese earthquake seismic waves to reach the Palouse Basin was approximately 1.5 hours while the Japanese Tsunami didn't reach the U.S. coast for approximately 10 hours.

Regional broadband seismographs can generally provide excellent estimates of Rayleigh wave (LR) displacement. To account for the sparse data collection interval of most well recorder data compared to the seismic data, we propose that the mean squared water level fluctuation as measured in the time domain in each well be compared to the mean spectral density (after adjusting to the wavelengths) of the passing Rayleigh wave.

For each individual well, Specific Storage (S_s) can then be calculated from the ratio of these quantities. Specific storage is defined as the amount of water that a portion of the aquifer releases from storage (unit volume of aquifer per unit hydraulic head change). For aquifers with multiple observation wells, this provides a method to estimate storage properties from an event based stress that can occur uniformly over the areal extent of the aquifer.

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Figure 1: Grande Ronde Aquifer Water Level Fluctuation from LR

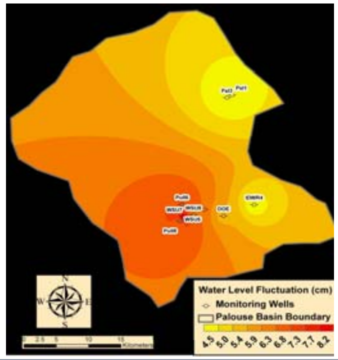
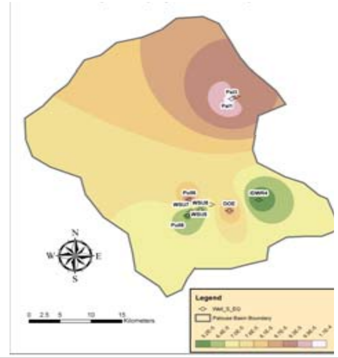


Figure 2: Grande Ronde Aquifer Storativity from LR



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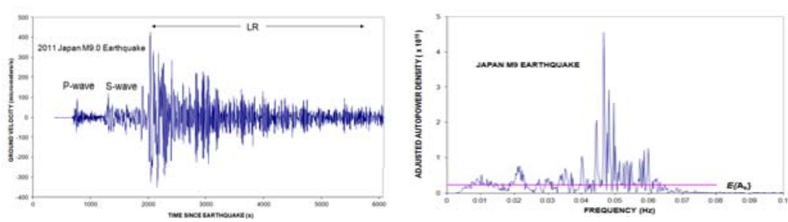

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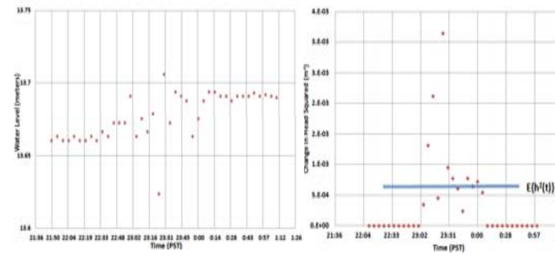

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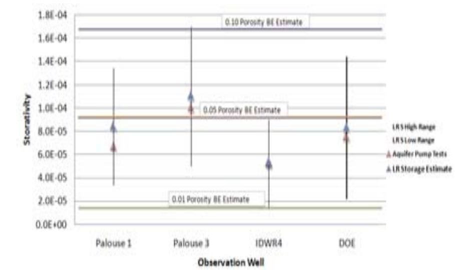
Figure 5: Storativity Method Comparison with seismic method error bars



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