Induced Seismicity in Oklahoma Affects Shallow Groundwater

by Chi-Yuen Wang, Michael Manga, Manoochehr Shirzaei, Matthew Weingarten, and Lee-Ping Wang

ABSTRACT

Documentation and analysis of groundwater responses to induced earthquakes are important to better understand their influence on shallow groundwater systems and hydrogeological properties and processes. Here we show that induced seismicity in Oklahoma can cause changes of groundwater level over distances \( > 150 \) km from the epicenter. We test existing models for the cause of the observed responses and find that the model most consistent with observations is enhanced crustal permeability produced by seismic waves, changing aquifer recharge. Simulation suggests that the sources of this recharge are close to the responding wells and have lateral dimensions of \( \sim 100 \) m. Continuous monitoring of pressure and temperature in wells, installing clustered wells to monitor multiple water levels near injection sites, and isotopic and chemical analysis of groundwater near injection sites are required to better understand and quantify the recharging sources.

Electronic Supplement: Figures of barometric pressure and water level in wells, coseismic static strain, and water-level change.

INTRODUCTION

A sharp increase of seismicity has occurred in the central United States in recent years, often near "high-rate injection sites" where billions of gallons of coproduced water from hydrocarbon extraction are injected into the subsurface for disposal or to enhance production (e.g., Frohlich, 2012; Ellsworth, 2013; Keranen et al., 2013, 2014; Hornbach et al., 2015; McGarr et al., 2015; Walsh and Zoback, 2015; Weingarten et al., 2015). Three \( M_w \geq 5 \) earthquakes occurred in Oklahoma alone since the beginning of 2016 (Fig. 1). A growing body of evidence suggests that induced seismicity and surface deformation are coupled in complex and unexpected ways (Shirzaei et al., 2016). A great concern is whether injection and hydrofracturing processes may impact shallow groundwater systems (e.g., Vidic et al., 2013; Stokstad, 2014; Vengosh et al., 2014). Natural earthquakes are known to cause a wide spectrum of hydrologic responses (e.g., Wang and Manga, 2010); thus, it is not surprising that earthquakes induced by wastewater injection may also cause similar changes. Documentation and analysis of groundwater responses to induced earthquakes, however, are important to better understand how induced seismicity may affect hydrogeological processes and shallow groundwater systems in the United States and elsewhere. After the 2016 \( M_w \) 5.8 Pawnee earthquake, an increase in stream discharge occurred near the epicenter of the earthquake (Manga et al., 2016). It was shown that earthquake-enhanced recharge may have provided the increased discharge, but the source of the excess water was not identified.

Here, we report changes of groundwater level over distances \( > 150 \) km from the epicenter following the 2016 \( M_w \geq 5.0 \) earthquakes in Oklahoma (Fig. 1). We discuss the mechanism of the changes by testing the existing hypotheses against observations and show that the observed changes are most consistent with the model of enhanced permeability produced by seismic waves. We simulate the observed changes and show that the changes of groundwater level were due to coseismic recharge from pre-existing sources near the wells. Finally, we highlight some important but unanswered issues.

OBSERVATION

Two types of aquifers are found in Oklahoma, in the Quaternary alluvial and terrace deposits and in the Paleozoic bedrock, with the former being the most important supplier for agricultural, municipal, and domestic use. The alluvial and terrace deposits consist of subhorizontal lenticular beds of sand, silt, clay, and gravel, which vary greatly in thickness within short lateral distances (Wood and Burton, 1968). All wells used in this study are open to aquifers in this formation.

The U.S. Geological Survey (USGS) manages 39 wells in Oklahoma; these were installed for monitoring groundwater level and equipped with automated recording equipment that continuously takes data at fixed intervals of 15–60 min. Data are transmitted to the USGS every hour. Most USGS wells are located away from the injection sites (Fig. 1); only two wells (364821 and 364831, Fig. 1 and Table 1) are located among many, and within 5 km from some injection wells. The
Oklahoma Water Resource Board (OWRB) manages numerous groundwater wells for irrigation, municipal, and domestic use. The vast majority of measurements are manually collected, but some OWRB wells have pressure transducers that provide hourly measurements of water level.

The three 2016 \( M_w \geq 5 \) earthquakes in Oklahoma all have strike-slip mechanisms (Fig. 1) and relatively shallow focal depths (Table 1). Aftershocks are not plotted in Figure 1 in order not to clutter the figure. The 13 February \( M_w \) 5.1 earthquake near Fairview ruptured a 70° dipping, southwest–northeast (SW–NE)-trending buried fault (Yeck et al., 2016); the 3 September \( M_w \) 5.8 earthquake near Pawnee probably ruptured a previously unmapped northwest–southeast (NW–SE)-trending fault (Bennett et al., 2016); and the 6 November \( M_w \) 5.0 earthquake near Cushing probably ruptured an SW–NE-trending fault (USGS, 2016).

Two USGS wells (364821 and 364831, Fig. 1), located 159 and 156 km from the epicenter, respectively, responded to the 6 November \( M_w \) 5.0 earthquake near Cushing. Water levels in the wells started to rise immediately after the earthquake (Fig. 2a,b) and continued to rise for 10–20 hrs before reaching their respective maxima of 4–7 cm above the pre-earthquake levels; afterward water levels slowly declined. Barometric pressure was stable at the time of the earthquake (see Fig. S1, available in the electronic supplement to this article), and no notable precipitation occurred several days before or after this earthquake. Furthermore, no similar change occurred in the records at other times. Thus, these water-level changes are likely to be directly related to the \( M_w \) 5.0 earthquake. Two OWRB wells (18699 and 171706, Fig. 1) responded to the 13 February \( M_w \) 5.1 Fairview earthquake. Well 18699 is 2.6 km from the epicenter and documented a coseismic rise of water level that continued to rise for several hours to reach a maximum of \( \sim 10 \) cm above the pre-earthquake level (Fig. 2c). Well 171706 is 81 km from the epicenter and showed a coseismic decrease of water level with continued decrease until being interrupted by a rapid rise of water level a day later (Fig. 2d). No rainfall occurred in the period several days before or after this earthquake. During the 3 September \( M_w \) 5.8 Pawnee earthquake, a change of water level was documented in OWRB well 127105 (Fig. 2e). Substantial rainfalls occurred both before and after this earthquake but none at the time of the earthquake. Rainfall leads to immediate rise of water level in the wells, sug-

![Figure 1. Locations of the epicenters (yellow stars) and focal mechanism of three 2016 \( M_w \geq 5 \) earthquakes in Oklahoma; smaller earthquakes are not plotted in order not to clutter the diagram. The locations of U.S. Geological Survey (USGS) wells (large triangles) and Oklahoma Water Resource Board (OWRB) wells (small triangles) are also plotted. The responding wells are labeled with identification numbers. Injection wells operating in the Arbuckle formation, which dispose waste fluid from oil and gas production, are shown with squares. The color version of this figure is available only in the electronic edition.](image-url)

<table>
<thead>
<tr>
<th>Table 1: Earthquakes and Wells in This Study</th>
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<tr>
<td><strong>Name of Earthquake</strong></td>
</tr>
<tr>
<td>Fairview</td>
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<td>Pawnee</td>
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<td>Cushing</td>
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<table>
<thead>
<tr>
<th><strong>Name of Well</strong></th>
<th><strong>Location (°)</strong></th>
<th><strong>Elevation (m)</strong></th>
<th><strong>Screen Depth (m)</strong></th>
<th><strong>Lithology</strong></th>
<th><strong>Epicenter Distance (km)</strong></th>
</tr>
</thead>
<tbody>
<tr>
<td>USGS 364821</td>
<td>36.806, -98.247</td>
<td>348</td>
<td>0–8.8</td>
<td>Alluvial</td>
<td>Fairview: 54 Pawnee: 125 Cushing: 159</td>
</tr>
<tr>
<td>USGS 364831</td>
<td>36.809, -98.201</td>
<td>346</td>
<td>0–6.9</td>
<td>Alluvial</td>
<td>Fairview: 58 Pawnee: 121 Cushing: 156</td>
</tr>
<tr>
<td>OWRB 18699</td>
<td>36.496, -98.881</td>
<td>458</td>
<td>15–21</td>
<td>Alluvial</td>
<td>Fairview: 2.6 Pawnee: 157 Cushing: 178</td>
</tr>
<tr>
<td>OWRB 171706</td>
<td>36.056, -97.981</td>
<td>347</td>
<td>0–16</td>
<td>Alluvial</td>
<td>Fairview: 81 Pawnee: 103 Cushing: 107</td>
</tr>
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suggesting surface runoff entering the well; the water-level rise after each rainfall was followed by a rapid water-level decline. The coseismic change of water level was followed by a sharp reversal in water level a few days later. A similar effect of rainfall may also have influenced water level in well 364821 during this time interval (top right panel), but no daily precipitation data are available near this well.

Despite the differences in polarities and epicenter distances, all responses discussed above share one common characteristic: water level following the coseismic change continued to change in the same direction for several hours to several days to reach maxima or minima before it gradually declined. This is similar to the water-level response of the Bourdieu Valley well near Parkfield, California, to natural earthquakes (Roeloffs, 1998).

DISCUSSION

Several mechanisms have been proposed to explain water-level changes during earthquakes; these include static strain due to fault rupture (Wakita, 1975; Muir-Wood and King, 1993; Ge and Stover, 2000; Chia et al., 2008), coseismic liquefaction (Roeloffs, 1998; Manga, 2001), and earthquake-enhanced permeability by dynamic stresses (Roijstaczer et al., 1995; Fleeger and Goode, 1999; Brodsky et al., 2003). The location of the earthquakes (Fig. 1) shows that all wells are located in areas of postseismic static extension, inconsistent with the observed increases of water level at the two USGS wells (364821 and 364831) and the OWRB well 18669. Furthermore, the calculated static volumetric strain (Okada, 1992; see Table S1) at the OWRB well 171706 due to the 13 February $M_w$ 5.1 earthquake is too small to explain the coseismic decrease of water level (Fig. 2d). For these reasons, we do not favor the static strain hypothesis for the observed coseismic change of water level.

The occurrence of liquefaction after the 3 September $M_w$ 5.8 earthquake (Kolawole et al., 2016; Manga et al., 2016) adds weight to the model of liquefaction by dynamic strain. We test this hypothesis for the present case with the magnitude of the dynamic strain that may be estimated from the peak ground velocity (PGV) in the vicinity of the wells (see Data and Resources). Seismographic records from several stations
near the wells show PGV of 0.2–0.3 cm/s. Assuming a shear velocity of 500 m/s for wet sands and gravels (Press, 1966), the peak dynamic shear strain is ~5 × 10⁻⁶, much smaller than the threshold amplitude of cyclic shear strains (10⁻¹⁻) required to initiate undrained consolidation in saturated sands, according to experiments on various kinds of sands under different confined environments (Vucetic, 1994). Furthermore, the occurrence of earthquake-induced liquefaction is delimited by a threshold of seismic energy density of ~0.1 J/m³; that is, the liquefaction limit that is shown as a straight line on a plot of hypocentral distance versus earthquake magnitude (Wang, 2007; Fig. 3). Plotting the responding wells on this diagram shows that most wells fall beyond the liquefaction limit (Fig. 3) at distances where the seismic energy density is below that required for liquefaction. For these reasons, we also do not favor the liquefaction model.

We next test the model of enhanced permeability by dynamic stresses. We consider a one-dimensional aquifer of length \( L \) (Fig. 4a); recharge is assumed to occur at the time of the earthquake along a section of the aquifer (between \( L_1 \) and \( L_2 \)). The differential equation for the coseismic change of hydraulic head \( h \) in a confined aquifer and the linearized Boussinesq equation for an unconfined aquifer have the same form:

\[
\frac{\partial h}{\partial t} = D \frac{\partial^2 h}{\partial x^2} + w(x,t),
\]

in which \( D \) is the hydraulic diffusivity and \( w(x,t) \) is the coseismic change of water level per unit time due to recharge. In the present context, we use equation (1) as the linearized Boussinesq equation for an unconfined aquifer. Thus, \( D = T/S_y \), in which \( T \) is the transmissivity and \( S_y \) is the specific yield of the aquifer; these parameters may be evaluated from well tests.

For boundary conditions, we consider no flow on one end of the aquifer to represent the presence of faults that block groundwater flow and zero head on the other end to represent discharge to local creeks. Thus,

\[
\frac{\partial h}{\partial x} = 0 \quad \text{at} \quad x = 0, \quad \text{and} \quad h = 0 \quad \text{at} \quad x = L.
\]

We further assume that, during the earthquake, \( w \) is finite for a short time, with the cumulative change being \( W_o \) between \( L_1 \) and \( L_2 \) along the aquifer but zero elsewhere. The solution for \( h \) is (see the electronic supplement)

\[
b(x,t) = \frac{4W_o}{\pi} \sum_{n=1}^{\infty} \frac{1}{n} \cos \left( \frac{n\pi}{2L} \right) e^{-\frac{n^2\pi^2}{4L^2} t} \left[ \sin \frac{n\pi L_2}{2L} - \sin \frac{n\pi L_1}{2L} \right].
\]

(3)

We limit the simulation to the responses of the two USGS wells (Fig. 2a,b); the records in the OWRB wells were either too noisy to simulate (Fig. 2c) or were interrupted by additional changes (Fig. 2d). The nonlinear least-square method is used to fit equation (3) to the increased water level after the earthquake. Equation (3) has five independent parameters \( x/L, L_1/L, L_2/L, W_o, \) and \( D/L^2 \). We tried different values of \( D/L^2 \) while keeping the rest as free parameters. The model with \( D/L^2 = 3 \times 10^{-7} \) s⁻¹ yields the smallest root mean square residuals and is chosen as the best model. Table 2 lists the best-fitting parameters and their uncertainties. Figure 4b,c shows the best-fitting curves together with the data for increased water levels.

\[\text{Figure 3.} \text{ Occurrences of earthquake-induced water-level changes and liquefaction plotted on a diagram of hypocentral distance versus earthquake magnitude. Straight lines show contours of equal seismic energy density; the uppermost line corresponds to a seismic energy density of } 10^{-6} \text{ J/m}^3; \text{ the threshold energy density required to trigger water-level change; the lower line marks the threshold energy density of } 30 \text{ J/m}^3 \text{ required to trigger liquefaction in sensitive soils (Wang, 2007). Responding wells of this study (filled triangles) are labeled with identification numbers; wells 364821 and 364831 overlap on this diagram. Nonresponding USGS wells are plotted as open triangles; nonresponding OWRB wells are not plotted because many wells do not have data of the required quality to be characterized. Data for global water-level changes are taken from a compilation (Wang and Manga, 2010; dark squares) and a recent study at Devils Hole (Weingarten and Ge, 2014; light squares). Liquefaction occurrences (black dots) are taken from a global compilation (Wang, 2007), with updates from New Zealand (S. Cox, personal comm., 2015). Responding wells of this study (filled triangles) are labeled with identification numbers; wells 364821 and 364831 overlap on this diagram. The plot shows that the seismic energy density at wells 364821 and 364831 during the 6 November } M_w 5.0 \text{ earthquake and those at well 171706 during the 13 February } M_w 5.1 \text{ earthquake are all less than the threshold energy required to trigger liquefaction but greater than that required to trigger water-level change. It also shows that the seismic energy density at well 171706 during an } M_w 2.7 \text{ aftershock was greater than the threshold energy required to trigger water-level change. The color version of this figure is available only in the electronic edition.} \]
Well tests and hydrogeological simulation of groundwater flow in the alluvial and terrace aquifer (Havens, 1989) yield $T = 0.0075 \text{ m}^2/\text{s}$ and $S_y = 0.15$, which give $D = 0.05 \text{ m}^2/\text{s}$. Using this $D$ together with $D/L^2 = 3 \times 10^{-7} \text{ s}^{-1}$, we obtain $L \sim 400 \text{ m}$. This value of $L$ is qualitatively consistent with geologic observations that the alluvial and terrace deposits consist mostly of lenticular beds of sand, silt, clay, and gravel, which vary greatly in thickness over short lateral distances (Wood and Burton, 1968). The simulation also suggests that the water-level increases in the two wells were caused by independent sources with widths of $\sim 100 \text{ m}$ ($L_2-L_1$) and $\sim 20 \text{ m}$ from the respective wells ($L_1-x$).

An important question is why some USGS wells closer to the 6 November $M_w$ 5.0 earthquake than the responding wells (Fig. 1) did not show a coseismic response. Different lithologies may not explain the different responses because many nonresponding wells are installed in similar alluvial sediments as the responding wells in this study. Another possible reason is that the responding USGS wells are closer to a large number of injection wells than the nonresponding wells (Fig. 1), but the nature of the recharging source remains unclear. It is also difficult to explain why the two USGS wells that responded to the 6 November $M_w$ 5.0 earthquake did not respond to the much closer 13 February $M_w$ 5.1 earthquake or the much larger 3 September $M_w$ 5.8 earthquake. We speculate that the first two earthquakes may have primed the aquifer, bringing it closer to the threshold for permeability change, and that the last earthquake, although smaller and further away, was the last increment that pushed the aquifer over the threshold to increase permeability. In addition, several studies have shown that earthquake triggering of seismicity and water-level changes depend on the frequency of seismic waves (e.g., Brodsky and Prejean, 2005; Wong and Wang, 2007; Guilhem et al., 2010; Rudolph and Manga, 2012). Thus, the answer to the question about why wells only sometimes respond to earthquakes may require a thorough analysis of the spectral content of seismic waves recorded at stations near the wells following each earthquake.

Finally, as noted earlier, the coseismic decrease of water level in well 171707 was followed a day later by a sharp increase of water level, and the coseismic increase of water level in well 127105 was followed a few days later by a sharp decrease of water level. It is noteworthy that the reversal after the 3 September $M_w$ 5.8 Pawnee earthquake was preceded immediately by an $M_w$ 2.7 aftershock near the well, and the reversal after the 13 February $M_w$ 5.1 Fairview earthquake was preceded by three aftershocks of $M_w > 3.5$ near the well on the same day. Although we cannot rule out the possibility that the coincidences in time between the reversals of water-level changes and aftershocks were accidental, laboratory study of permeability of fractured rocks (e.g., Liu and Manga, 2009) shows that cyclic loading can either increase or decrease permeability; thus, it is not unlikely that aftershocks can reverse the permeability changes induced by the mainshock, causing a reversal in the coseismic change of water level.

### CONCLUDING REMARKS

In this study, we document coseismic changes of groundwater level in Oklahoma after three induced earthquakes with magnitude greater or equal to 5.0. We showed that the observed changes can be explained neither by the static strain hypothesis nor by the liquefaction hypothesis. On the other hand, the model of enhanced crustal permeability produced by seismic waves, altering recharge of shallow aquifers, is consistent with observations. Simulations based on this model suggest that the sources of this recharge are close to the responding wells and
have lateral dimensions of $\sim 100$ m. Further testing of this model and better understanding and quantifying the influence of induced earthquakes on shallow groundwater systems require continuous monitoring of pressure and temperature in wells, installing clustered wells to monitor multiple water levels near injection sites, tidal analysis of water level in wells, and isotopic and chemical analysis of groundwater near injection sites.

**DATA AND RESOURCES**


**ACKNOWLEDGMENTS**

We thank Mark Belden of the Oklahoma Water Resource Board (OWRB) for answering queries about wells in Oklahoma, Inez Fung of University of California, Berkeley, and Jeff Robel of National Oceanic and Atmospheric Administration (NOAA) for providing links to the hourly surface pressure data. We also thank Associate Editor and Andrew Barbour for their constructive and thoughtful reviews that helped improve the article. M. M. and C. Y. W. were supported by National Science Foundation (NSF) Grant Number EAR1344424, and M. S. and M. W. were supported by Stanford Center for Induced and Triggered Seismicity.

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![Table 2](http://www.earthquake.usgs.gov/earthquakes/eventpage/us100075y8#map?ShakeMap)

<table>
<thead>
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<th>Best-fitting parameters in equation (3)</th>
<th>USGS Well 36482109</th>
<th>USGS Well 36483109</th>
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<tr>
<td>$L_1/L$</td>
<td>0.256 ± 0.056</td>
<td>0.114 ± 0.227</td>
</tr>
<tr>
<td>$L_2/L$</td>
<td>0.476 ± 0.035</td>
<td>0.446 ± 0.079</td>
</tr>
<tr>
<td>$x/L$</td>
<td>0.202 ± 0.018</td>
<td>0.067 ± 0.064</td>
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<tr>
<td>$H_2$ (m)</td>
<td>0.434 ± 0.015</td>
<td>0.032 ± 0.014</td>
</tr>
<tr>
<td>rms residuals (m)</td>
<td>0.0197</td>
<td>0.0024</td>
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<tr>
<th>Derived parameters with $L = 400$ m (see the Discussion section)</th>
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<th>USGS Well 36483109</th>
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</thead>
<tbody>
<tr>
<td>$L_1$ (m, left boundary of coseismic recharge in Fig. 4a)</td>
<td>102 ± 22</td>
<td>46 ± 91</td>
</tr>
<tr>
<td>$L_2$ (m, right boundary of coseismic recharge in Fig. 4a)</td>
<td>190 ± 14</td>
<td>178 ± 32</td>
</tr>
<tr>
<td>$x$ (m, well location in Fig. 4a)</td>
<td>81 ± 7</td>
<td>27 ± 26</td>
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rms, root mean square.