A permeability-change model for water-level changes triggered by teleseismic waves

Z. M. GEBALLE, C.-Y. WANG AND M. MANGA

Department of Earth and Planetary Science, University of California, Berkeley, Berkeley, CA, USA

ABSTRACT

Although many hydrologic changes induced by teleseismic waves have been reported, the mechanism(s) responsible for the changes are usually not known. Permeability changes induced by seismic strains are often invoked to explain changes in water level. Using water-level data in Taiwan after the 2008 Wenchuan earthquake, we show here that the observations cannot be properly explained by previously proposed models of postseismic permeability changes. A new model is required in which the postseismic permeability decreases exponentially as a function of time, with a time constant of \(<3\) min, which is appreciably shorter than inferred from earlier studies. The result may have important implications for pore-sealing mechanism(s).

Key words: dynamic stresses, permeability evolution

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Corresponding author: Zachary M. Geballe, Department of Earth and Planetary Science, University of California, Berkeley, Berkeley, CA, USA.

Email: zgeballe@berkeley.edu. Tel: +510 643 8325. Fax: +510 643 9980.

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INTRODUCTION

Hydrologic responses to teleseismic waves have been documented in several studies of water wells (Matsumoto 1992; Roeloffs 1998; Brodsky et al. 2003; Elkhoury et al. 2006; Doan & Cornet 2007). From these studies, it has become clear that significant changes in water level can be driven by moderate amplitude dynamic stresses. By compiling observations of water well responses to earthquakes, Wang & Chia (2008) found that seismic waves with energy densities as small as \(10^{-2}\) Jm\(^{-3}\) are capable of causing sustained water-level change. Additionally, water level can display a wide range of evolutions in time. For example, Wang et al. (2004) identify four types of responses among Taiwan’s wells to the 1999 Chi-Chi earthquake: (i) a coseismic increase in water level followed by a decline, (ii) a coseismic drop followed by an gradual increase, (iii) a coseismic drop followed by further decline, and (iv) a coseismic increase followed by steady or increasing water level. Roeloffs (1998), however, documents that the water level in one well in California increases coseismically in response to all sufficiently large regional earthquakes (i.e. the well always exhibits response type 1 or 4).

Such variety suggests that seismic waves may influence aquifers through a variety of mechanisms. Proposed mechanisms include unblocking fluid pathways (Brodsky et al. 2003), compaction of unconsolidated sediment (Roeloffs 1998) and dilatation of fractured rock (Doan & Cornet 2007). As a first step toward understanding which is typically responsible for hydrological changes, we must study the details of individual events and their seismic triggers.

Models of seismically induced permeability change have successfully explained several data sets with few free parameters (Brodsky et al. 2003; Wang et al. 2004; Elkhoury et al. 2006; Claesson et al. 2007). Moreover, unblocking of cracks by teleseismic waves has been shown to be physically plausible on theoretical grounds (Brodsky et al. 2003) and in laboratory experiments (Liu & Manga 2009; Elkhoury et al. 2011). Following an initial increase in permeability, several authors have inferred that cracks sealed over months (Elkhoury et al. 2006) to a couple years (Claesson et al. 2007) to more than a couple years (Manga & Rowland 2009). Here, we find that permeability returns to its pre-earthquake level in \(<3\) min, the shortest timescale documented to date.

We investigate the details of the 9-cm decline of water level at Liujia well in Taiwan following the 2008 Wenchuan earthquake (see also Wang et al. 2009). We show that a model of constant postseismic permeability, such as that proposed by Brodsky et al. (2003), does not fit the
transient water-level data in Taiwan. Instead, the data require a revised model in which the permeability decreases exponentially as a function of time. We further find that the moment of permeability increase coincides with the arrival of a approximately 20-s Love wave. Lastly, we interpret our fitted parameters in terms of the local hydrogeology and discuss possible mechanisms for rapid permeability changes.

**OBSERVATION**

Passing seismic waves from the 2008, Wenchuan earthquake caused both oscillations and a permanent drop in water level at Liujia well in southwestern Taiwan (Fig. 1) (Wang et al. 2009). Water level is determined from a pressure gauge that is suspended in the water from a cable. The gauge readings have precision of approximately 1 mm and are documented by a digital recorder at a rate of 1 Hz. The observed water level (bottom of Fig. 2) starts to drop 600 s after the earthquake, decreases approximately 9 cm in the subsequent 400 s, and levels off about 1200 s after the earthquake. No further changes are observable above the 0.3-cm tidal oscillation during the day after the earthquake. That surface waves from the Wenchuan earthquake arrive at southwestern Taiwan at 600–750 s can be seen from the seismic velocity time series at nearby TAIB station (top of Fig. 2). In addition to triggering the decline in water level, seismic waves also induce oscillations in water level (top of Fig. 2).

**MODEL OF PERMEABILITY EVOLUTION**

We interpret the monotonic drop in water level as a consequence of permeability changes. Following Darcy’s law, the source of deviations from steady-state flow can be divided into diffusivity changes (equivalent to permeability assuming viscosity and specific storage remain constant) and head changes. To maintain simplicity, previous studies have modeled step function changes in permeability (Brodsky et al. 2003) or in head (Doan & Cornet 2007) in unbounded 1 D aquifers. The latter are only reasonable for large strains (e.g. Jonsson et al. 2003) or at wells neighboring easily disrupted rock (e.g. Doan & Cornet 2007). Neither is true in our study; the teleseismic waves have a modest energy density of $2 \times 10^{-3} \text{J m}^{-3}$ (using equation 2 of Wang & Chia 2008), and Liujia well probes sedimentary rocks at 210 m depth. Therefore, we focus on the Brodsky et al. (2003) model of change in diffusivity.

Brodsky et al. (2003) model the coseismic water-level drop at a well near Grants Pass, Oregon, as the result of sudden increase in permeability. Specifically, they seek to explain the well’s response to two earthquakes: the 1994 Petrolia earthquake and the 1999 Oaxaca earthquake. A schematic of their model is shown on the left half of Fig. 3. They model the region probed by the well as a 1 D unbounded aquifer that is divided by an impermeable barrier located a distance $d$ away from the well. Before the earthquake, hydraulic head is distributed as a step function with step size $h$ at the barrier. As the seismic waves pass by, the diffusivity of the barrier jumps from zero to the value of the surrounding aquifer, $D_0$. No further changes in diffusivity occur. Brodsky et al. (2003) find an analytic solution for change in water level, $\Delta W$, as a function of time, $t$, after the diffusivity change:

$$\Delta W = -h \cdot \text{erf}\left(\frac{d}{\sqrt{4D_0 t}}\right)$$

**Fig. 1.** (Left) Location of Liujia well and TAIB seismic station in Taiwan. The angle of approach of the teleseismic waves is shown by arrows. (Inset) Liujia well (circle), broadband seismic stations (squares), GPS stations (plus symbols). (Right) Lithologic log of Liujia well, reproduced from Wang et al. 2009. The bar from 205 to 220 m depth marks the screened portion of the well. The Liujia well is located in layered sediments on top of a basement high, the Peikang High, which is a stable crustal block beneath western Taiwan. Sediments on the Peikang High are flat lying; thus, the well log is a close illustration of the local geologic conditions. No major active faults cut through the Peikang High, nor do they come near the well.
The Liujia data cannot be fit by the Brodsky et al. (2003) model, indicating there is additional complexity in the mechanism of water-level change.

Equation 1 does not come close to fitting the 2200 s of data starting from the water-level drop (Fig. 2 – blue curve). Rather, early water levels are higher and late water levels are lower than the best fit. Although the model can successfully explain the first 600 s after the decline starts (Fig. 2 – red curve), it overestimates the change at late times. We note that the same problem appears in the fit to data from the Petrolia earthquake in Fig. 5 of Brodsky et al. (2003), although another study found good fits (Sil 2006). The data from Liujia well, however, demand a different model. We expand on the original model by adding further time evolution to the permeability change that allows us to reproduce both the slow onset of water-level decline at early times and the rapid leveling off at late times.

Motivated by the observation that water-level changes level off more quickly than expected for a permanent change in permeability, we propose a new model with two extra parameters. A schematic illustration of our model is shown on the right half of Fig. 3. Our domain is two adjacent 1D aquifers of the same height and diffusivity, \( D_0 \), divided by a region of low permeability, labeled ‘aquitard’, whose permeability we assume to be zero for our computations. In a later section, we will interpret the aquitard as a horizontal layer of mudstone. The well is assumed to probe the water level at depth \( z = 2d \), at one edge of the aquifer, while the mudstone is centered at \( z = d \). Seismic shaking causes a sudden increase in diffusivity so that it becomes \( D_0 \) everywhere. That is, a flow in barrier is breached. Then, diffusivity decays exponentially in time back to zero in the barrier region with a characteristic time, \( t \), the assumed initial head is a high constant in the aquifer connected to the well, a low constant in the neighboring aquifer, and linearly distributed in the barrier of width \( f/2d \) that connects the two aquifers (Fig. 3A).

To simulate the time evolution of head, Darcy’s law is solved numerically. We use an explicit time-step algorithm to model the evolution of hydraulic head, \( h \), according to \( \frac{\partial h}{\partial t} = \frac{D_0}{\sqrt{\pi t}} \frac{\partial^2 h}{\partial x^2} \), for \( D = D_0 \exp[-(t - t_0)/\tau] \) with no-flow boundary conditions at each edge of the domain: \( \frac{\partial h}{\partial t} = 0 \). Because all values of \( d \) and \( D_0 \) that give the...
same value of \(d^2/D_0\) yield the same time evolution of \(h\), we consider \(d/D_0\) to be a single parameter. Because it is the same fitting parameter as in Brodsky \textit{et al.} (2003), we have only added two free parameters, \(\tau\), the permeability recovery time, and \(f\), the fraction of the domain occupied by the barrier. We will show that \(f\) is not constrained by the model, meaning we have effectively added one fitting parameter, \(\tau\). In Appendix A, we show that the addition of this new parameter is justified by the sevenfold decrease in misfit of the model to the water-level evolution at Liujia well.

The new model constrains both fitted timescales, \(\tau\) and \(d^2/D_0\), as well as the starting time, \(t_0\). Results are given in Table 1. We reject parameters that give misfits \(>0.15\) cm (Appendix B). Assuming \(t_0 = 600\) s, the 0.15-cm misfit contour in Fig. 4 shows the region of parameter space that we accept: \(\tau < 140\) s (2.3 min) and \(d^2/D_0 = 1500–3000\) s. Uncertainty in \(t_0\) (Fig. 5) causes negligible increases in the acceptable range of fitting parameters. The uncertainty in the fraction, \(f\), is so large that it can be anything between 0 and 1. Allowing any value of \(f\) in this range, we find that acceptable values of \(\tau\) are bounded above by 180 s while \(d^2/D_0\) can range from \(10^3\) to approximately \(10^6\) (Fig. 6). Typical values of the initial difference in head between the two aquifers, \(h\), were approximately 30–500 cm. Values \(>100\) m are required for \(\tau < 10\) s or \(f > \frac{1}{4}\), making these extreme parameter values unlikely. This constraint on \(f\) reduces the upper bound on \(d^2/D_0\) to \(10^4\) s (Fig. 6). Our final result is that \(\tau\) is between 10 and 180 s and \(d^2/D_0\) is between 700 and 30 000 s (12 and 500 min).

**SEISMIC TRIGGER**

By comparing the timing of the modeled permeability change to the seismic record, we find that Love waves must be responsible for initiating the water-level drop at

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**Table 1** Parameters and their fitted values.

<table>
<thead>
<tr>
<th>(D_0/d^2)</th>
<th>(\tau)</th>
<th>(t_0)</th>
<th>(f)</th>
</tr>
</thead>
<tbody>
<tr>
<td>700–30 000 s</td>
<td>10–180 s</td>
<td>550–625 s</td>
<td>0–0.75</td>
</tr>
</tbody>
</table>

**Fig. 4.** Contours of the rms misfit of the model from 600 to 2600 s, assuming \(f = \frac{1}{4}\) and \(t_0 = 600\) s. Defining the acceptable range of fitting parameters as those that provide misfits of <0.15 cm, \(\tau < 140\) s (2.3 min), and \(d^2/D_0 = 1500–3000\) s (25–50 min).

**Fig. 5.** Misfit of best fit versus assumed starting time. We reject starting times before 550 or after 625 s, shown in the shaded region.

**Fig. 6.** The range of \(d^2/D_0\) that give misfits below 0.15 cm, shown as bars, is plotted versus \(f\). The shaded region shows parameter values for which the initial head difference between aquifers is <100 m. We note that while the diffusion timescale shown here increases 2 orders of magnitude as \(f\) goes from 0.1 to 0.9, no such trend is observed in either \(\tau\) or the rms misfit over the same range of \(f\).
Liujiang well. Wang et al. (2009) also proposed that Love waves initiated the water-level change. We infer that the permeability change happened between 550 and 625 s by rejecting all starting times that give misfits >0.15 cm (Fig. 5). We confirm that early times are improbable because the seismic amplitude was an order of magnitude lower in the 100 s before 550 s than in the 100 s afterward (Fig. 7). By the same logic, it is most probable that the trigger of water-level decline happened sometime during the dramatic increase in seismic energy after 575 s. The dominant source of seismic energy in the time period from 575 to 625 s, a approximately 20-s-period Love (inset of Fig. 7), is therefore the most likely source of the change in permeability that precipitated the drop in water level at Liujiang well. To our knowledge, this is the first time that a persistent hydrologic change has been associated with a Love wave. Synchronized GPS timing of the well and seismic station made this new observation possible.

**GEOLOGICAL INTERPRETATION**

The fitted value of $\frac{d^2}{D_0}$ is consistent with the stratigraphy at Liujiang well (Fig. 1). The fitted value of $D_0$, meanwhile, reveals that permeability near the well can change much more rapidly than previously documented in field measurements.

That $\frac{d^2}{D_0}$ is more than 700 s but <30 000 s is consistent with reasonable values of $d$ and $D_0$ at Liujiang well. Possible low-head aquifers include the sandy layers shown in Fig. 1 at 230, 223, 192, and 172 m depth that are separated from the screened portion of Liujiang well by thin, silty layers. This would imply that fluid diffused through $d$ approximately 5–30 m of siltstone to reach the well. Given our fitted values of $\frac{d^2}{D_0}$, we find that $D_0$ approximately 0.0008 to 1.3 m$^2$ s$^{-1}$ which is a typical range of siltstone diffusivities (Roeloffs 1996). In future studies, possible ways to verify that the change in head is because of vertical fluid flow are analysis of water level or of isotope data from several nearby wells that probe aquifers at different depths, as has been demonstrated in Wang (2007) and Wang et al. (2005).

The fitted timescale of permeability recovery, $\tau < 180$ s, is much shorter than expected from previous studies. In cases where pre-earthquake permeability was found to recover, the timescale of recovery varied from 6 days to 6 years (Hiramatsu et al. 2005; Claesson et al. 2007; Ingebritsen & Manning 2010). In addition, many earthquake hydrology studies have found that permeability increases without evidence for a subsequent decay (Rojstaczer et al. 1995; Brodsky et al. 2003; Wang et al. 2004). The lack of previous evidence for permeability recovery over several minutes may be due to a sampling bias as capturing and documenting fast recovery requires high sampling frequency, a feature rarely available for wells. To determine whether the rapid timescale documented here is typical or atypical, further studies must be carried out at other sites where water level is monitored at high frequency.

**MECHANISM OF PERMEABILITY RECOVERY**

We hypothesize that the rapid permeability recovery documented here is because of reclogging of fluid pathways that were unclogged during the earthquake. First, we rule out precipitation of minerals as the reclogging mechanism because controlled laboratory studies have shown the timescale is much longer than minutes. For example, Tenthorey et al. (2003), Giger et al. (2007) and Kay et al. (2006) found that permeability decreased because of dissolution and re-precipitation of minerals over hours, even when temperatures were very high (700–900°C) or chemistry was far from equilibrium. Second, we agree with the suggestion of Elkhoury et al. (2011) that the simplest explanation of permeability increase and subsequent decay is a reversible process such as unblocking and reclogging of fluid pathways by entrainment and deposition of particles. Third, Wang et al. (2009) have shown that reclogging of fractures because of dynamic strains is physically reasonable for this event. Thus, we suggest that unblocking and reclogging could have caused the permeability changes seen in our field study while precipitation could not have.
At least one laboratory study supports our hypothesis, while most others do not reject it. Elkhoury et al. (2011) show that permeability decreases on minute timescales following permeability increases in fractured sandstone that are induced by dynamic stressing. On the other hand, a similar laboratory study found no healing on timescales of 10 s of minutes (Liu & Manga 2009). Most laboratory studies (Tenthorey et al. 2003; Kay et al. 2006 and Giger et al. 2007), though, do not include a period of dynamic stressing and therefore do not document the ability of pores to reclog during the minutes following an unclogging event. Further laboratory studies of permeability response to dynamic stressing would be useful to clarify the conditions under which rapid re-clogging can occur.

POSSIBILITY OF MULTIPLE LENGTH SCALES

We note that this study does not address changes in the permeability in the 100 s of meters surrounding wells that have been inferred in other aquifers in response to dynamic strain (Elkhoury et al. 2006). While Elkhoury et al. (2006) analyzed tidal amplification factors to determine changes in the well response to 12 h forcing, we can only describe the hydrologic response to the earthquake over the 10 s of minutes during which we see the water-level drop. It is possible that farther than approximately 100 m from the well, there were more permanent changes in permeability that may be manifest in the water-level evolution in the days following the earthquake.

CONCLUSION

After the coseismic permeability increase, we find that the model of a constant postseismic permeability fails to fit the water-level data at Liujia well in response to the 2008 Wenchuan earthquake. In order to fit the observed data, we propose a new model in which the postseismic permeability decreases exponentially with time. We suggest that flowing water unclogged some pathways in fractured mudstone in response to a 20-s Love wave and that it reclogged in <3 min. We find that a low-head sandy layer located 5–30 m from the screened portion of the well, with diffusivity of 0.0008–1.3 m² s⁻¹, drew down the well’s water level. Thus, our new model reveals that fractured mudstone can reclog on timescales much shorter than observed previously.

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We use Bayesian statistics to show that our model provides a more likely explanation for the Liuji data than the existing model of Brodsky et al. (2003). We justify the addition of a new parameter, τ, by the great improvement in fit. Assuming that the new model and old model have equal prior likelihoods, the ratio of posterior likelihoods of our model to the previous model is (Sivia & Skilling 2006)

\[
\frac{P(\text{new|data})}{P(\text{previous|data})} = \frac{P(\text{data|new})}{P(\text{data|previous})} = \exp(-\chi^2_{\text{new}}) \frac{d\tau}{d\tau_{\text{previous}}} \frac{df}{df_{\text{previous}}} \frac{\Delta f}{\Delta f_{\text{previous}}}
\]  
(A1)

where

\[
\chi^2 = \left( \frac{\sum (h_{\text{obs}} - h_0)^2}{\sigma^2} \right) = N \frac{\text{misfit}^2}{\sigma^2}
\]  
(A2)

and \( h_{\text{obs}} \) and \( h_0 \) are the observed and modeled water levels. The sum extends over all data points that we used, \( N \) is the number of data points, and \( \sigma \) is the uncertainty in each measurement. \( d\tau \) and \( df \) are the ranges of \( \tau \) and \( f \) that give acceptable misfits, while \( \Delta \tau \) and \( \Delta f \) are the prior ranges of these parameters. We have assumed that the analogous factors for the parameter \( d^2/D_0 \) in the two models are equal

\[
\left( \frac{d(d^2/D_0)}{\Delta(d^2/D_0)_{\text{new}}} \right) = \left( \frac{d(d^2/D_0)}{\Delta(d^2/D_0)_{\text{previous}}} \right)
\]

We now use the following values: \( N = 2200 \), misfit = 0.15 cm for the new model (Fig. 4), misfit = 0.74 cm for the previous model (Fig. 2), \( df \) approximately 0.75 and \( \Delta f = 1 \) (Fig. 6), and \( d\tau \) approximately 180 s. As it is debatable what the physically expected range of \( \tau \) should be, we choose a conservative value of \( \Delta \tau = 10^3 \) years (i.e. we assume a uniform prior probability density function of \( \tau = 0–10^3 \) years). Lastly, we estimate \( \sigma = 0.1 \) cm from the amplitude of the noise before and after the water-level decrease (Fig. 2). The resulting ratio of probabilities is:

\[
P(\text{new|data}) \approx \exp\left(-2.2 \cdot 10^5\right) \quad \frac{180}{0.75} \quad \frac{10^6}{1} = 10^{6000}
\]

That the \( \chi^2 \) term of equation 2 dominates the \( d\tau/\Delta \tau \) and \( df/\Delta f \) terms indicates that the increase in likelihood of the new model because of the improvement in fit far outweighs the reduction because of adding two new parameters.

**APPENDIX B: ACCEPTABLE RANGE OF PARAMETERS**

Assuming the uncertainty in each measurement is the scatter in the data (0.1 cm), the 1σ error bars of \( \tau \) and \( d^2/D_0 \) should only extend from the position of the best fit (misfit = 0.1 cm) to the misfit contour at \( \sqrt{0.1^2 + 0.1^2} = 0.1 + 10^{-8} \) cm. For practical reasons, we chose to overestimate the error bars by bounding the acceptable range of all parameters by the 0.15-cm misfit contour.

Our choice of 0.15 cm is governed by the following reasoning. In order to limit computational time, we use 100 spatial grid points. The discreteness of this grid causes uncertainty in the calculation of the rms misfit which we estimate as the approximately 0.01-cm scatter in the region of minimum misfit. Hence, our choice of the 0.15 cm contour is a conservative estimate of the range of acceptable parameters considering both the relatively small uncertainty in the data and the relatively large uncertainty in our computation of misfits.