A permeability and compliance contrast measured hydrogeologically on the San Andreas Fault

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Abstract Hydrogeologic properties of fault zones are critical to faulting processes; however, they are not well understood and difficult to measure in situ, particularly in low-permeability fractured bedrock formations. Analysis of continuous water level response to Earth tides in monitoring wells provides a method to measure the in situ hydrogeologic properties. We utilize four monitoring wells within the San Andreas Fault zone near Logan Quarry to study the fault zone hydrogeologic architecture by measuring the water level tidal response. The specific storage and permeability inferred from the tidal response suggest that there is a difference in properties at different distances from the fault. The sites closer to the fault have higher specific storage and higher permeability than farther from the fault. This difference of properties might be related to the fault zone fracture distribution decreasing away from the fault. Although permeability channels near faults have been documented before, the difference in specific storage near the fault is a new observation. The inferred compliance contrast is consistent with prior estimates of elastic moduli in the near-fault environment, but the direct measurements are new. The combination of measured permeability and storage yields a diffusivity of about $10^{-2}$ m$^2$/s at all the sites both near and far from the fault as a result of the competing effects of permeability and specific storage. This uniform diffusivity structure suggests that the permeability contrast might not efficiently trap fluids during the interseismic period.

1. Introduction

Elevated pore pressure weakens fault zones by reducing effective stress [Hubbert and Rubey, 1959]. During the interseismic period, the hydrogeologic properties of the fault damage zone determine the development of pore pressure inside the fault zone that may affect the fault stability [Rice, 1992]. During fault slip, the hydrogeologic properties of a fault zone can also influence hydromechanical dynamic weakening processes within the fault slip zone [Andrews, 2002; Byerlee, 1993; Segall and Rice, 1995]. The hydrogeologic properties of the fault zone often serve as an indicator of fracture density, and therefore the hydrogeologic structures of fault zones can reflect the history of deformation conditions [Caine et al., 1996]. Despite the importance of fault zone hydrogeologic structure, the hydrogeologic properties of fault zones are hard to measure and are generally poorly understood. Continuous water level responses to tidal forcing, however, provide a probe that enables us to efficiently determine in situ hydrogeologic properties of fault damage zones [Elkhoury et al., 2011; Hsieh et al., 1987; Xue et al., 2013]. This study utilizes four water wells with different distances to the fault and uses water level tidal response to study the hydrogeologic architecture of the San Andreas Fault near Logan Quarry.

2. Tectonic Background

The A. R. Wilson Quarry is owned by Graniterock and located on the southwestern side of the San Andreas Fault in Aromas, California (Figure 1a). The facility was historically known as the Logan Quarry and we use the older name here to be consistent with the existing geological literature on the site. The Quarry is harvesting the exposed sliver of quartz gabbro which is unconformably overlain by Eocene to Miocene sedimentary and volcanic rocks. The high fracture density inside the fault damage zone has enabled economic extraction of the gabbro as a source of construction material for over 100 years. The quartz gabbro at this site is correlated with the quartz gabbro at Gold Hill about 180 km to the southeast, which indicates that the long-term average slip rate of the San Andreas Fault is 25–35 mm/yr [Ross, 1970]. The site is located near the transition from the locked and creeping section of the San
Andreas Fault. During the 1906 San Francisco and 1890 Pajaro Gap earthquakes, surface cracking was observed adjacent to the Pajaro River near the San Andreas Fault. Historic seismic slip was documented by a railroad located between the Chittenden and the Pajaro Bridges near the Quarry that was bent in the 1906 San Andreas earthquake [Lawson and Reid, 1908] (Figure 1a). The most prominent strand of the San Andreas Fault inside the Quarry is a lithology boundary with quartz gabbro of the Logan Quarry on the southwest side and Tertiary sedimentary rocks on the northeast side [Wagner et al., 2002]. This major boundary is clearly the locus of the largest displacement, although other strands of the fault may be active at the present time. We will therefore refer to the lithologic boundary as the San Andreas Fault.

3. Observations

We utilize four wells (DW4, DW11, DW13, and DW2) at various distances to the San Andreas Fault to observe the water level tidal response (Figure 1a). All of the chosen wells are on the southwest side of the San Andreas Fault within the gabbro formation, and their screened intervals are in the fractured gabbro. Pressure sensors measured the height of the column of water above the sensor in each of the four wells as detailed in Table 1 and shown in Figure 2. The water levels have been observed from April 2014 to July 2015 with 10 min sample intervals. Originally, all the wells used a Onset HOBO pressure transducer (http://www.onsetcomp.com, model U20-001-01-Ti with depth range of 100 ft) with a resolution of 0.46 cm. The well DW4 has a small signal of water tidal response near the resolution of the HOBO logger, so we changed the pressure sensor inside DW4 on 18 May 2015 to an RBR SoloD (http://www.rbr-global.com, model SoloD with depth range of 20 m) which has a better resolution of ~0.03 cm. All of the pressure transducers measure the absolute pressure which includes both water pressure and barometric pressure. To evaluate the effects of the barometric pressure, we also deployed a pressure transducer in the air inside well DW4. The logger of DW13 was temporarily jammed on the well wall, so there was no recorded water level oscillation during August–November 2014. Otherwise, the time series are uninterrupted for the 15 month period.
The water levels were not disturbed by data retrievals (Figure 3). One mysterious feature of the data is the step change of water level that occurred on 12 April 2015 and was observed at all the wells except DW4. This water level jump is not related to any local earthquakes, quarry blast events, or surface hydrological events. The cause of this step change is not clear. There was a similar drop that occurred on 1 September 2015, and the cause of this event is also not clear. We speculate that both may be related to transient creep, but it requires further data to evaluate this possibility. Therefore, the study of the steps is beyond the scope of this paper. The tidal response analysis pursued here focuses on a specific harmonic and is therefore unaffected by these steps. The long-term trend of the observed water levels inside each well is not correlated to the trend in the height of the Pajaro River. The height change of the Pajaro River is correlated to the precipitation (Figure 3). There were no significant water level changes inside wells at the time when there were big river height changes (Figure 3), so the connection between the river and the wells is not significant.

Assuming the porous media is homogeneous and isotropic, tidal responses are sensitive to a cylindrical volume around the well with radius proportional to \( (D/f)^{1/2} \), where \( D \) is hydraulic diffusivity (m\(^2\)/s) and \( f \) is frequency (1/s). As will be shown below, the estimated scale of sensitivity for the four wells is much less than the distances between the wells, so each well provides an independent, local measure of the hydrogeologic properties. As the wells are distributed at different distances from the San Andreas Fault, the inferred hydrogeologic property of the formation surrounding each well gives the hydrogeologic properties at four different distances to the major fault.

### 4. Methods

#### 4.1. Tidal Response

Two different tidal response models are used in this study for different flow geometries. For most wells, we use a horizontal flow model as is standard in tidal response studies where the tidally driven flow is radially symmetric about the well. One well, DW13, shows evidence for a significant vertical flow requiring a different interpretative model. Well DW13 is close to the fault which could be a barrier for the horizontal flow, thus driving the flow to be predominantly vertical. An artesian well close to DW13 but not monitored as part of this study also exhibits vertical flow further suggesting that the vertical flow model is likely appropriate for DW13. In addition, DW13 is the only well with observed positive phase lag (water level oscillation leads the imposed tidal dilatation strain). The positive phase cannot be caused by a horizontal flow as be discussed below. We will apply a vertical flow model to calculate the equivalent hydrogeologic properties of the formation surrounding DW13, and apply
a horizontal flow to calculate the equivalent hydrogeologic properties for the three remaining wells. These three wells have significant negative phase response (less than $-1^\circ$) which is consistent with horizontal flow and cannot be generated by vertical flow.

4.1.1. Horizontal Flow Model

In a well-aquifer system, solid Earth tidal forcing generates dilatation strain in the aquifer formation resulting in water head disturbances which cyclically pump water in and out of a well [Hsieh et al., 1987]. Finite time is needed for pore pressure to diffuse through the porous media, so the water level inside the well oscillates at the same frequency as the imposed dilatation, but with a phase lag. Permeability is a parameter controlling how easily water can travel through porous media, and storage is proportional to the formation compressibility [Freeze and Cherry, 1977]. Permeability and storage therefore determine the phase and amplitude of water level response to tidal forcing. At the far field of the well-aquifer system, the water head responds to the imposed dilatation strain simultaneously, so the far field head response is the pore pressure response under an undrained condition which can be expressed as [Hsieh et al., 1988]:

$$S_i = \frac{e}{h} = \frac{\epsilon X}{X h}$$  

(1)
where $S_s$ is the specific storage under undrained conditions which is proportional to the media compressibility (1/m), $\varepsilon$ is the tidal dilatation strain, $h$ is the far field water head (m), and $x$ is the water head inside the water well (m). The Hsieh et al. [1987] solution computes the response of the water in an open well with the assumptions that the aquifer is confined, extends laterally indefinitely, and the aquifer media is homogeneous and isotropic. By combing (1) and Hsieh et al. [1987] solution, then the response of water level inside the well to the far field head perturbation can be expressed as:

$$A = \left| \frac{x_0}{e_0} \right| = \frac{1}{S_s} \left( E^2 + F^2 \right)^{-1/2}$$  \hspace{1cm} (2)

$$\eta = \arg \left( \frac{x_0}{e_0} \right) = -\tan^{-1} \left( \frac{F}{E} \right) \hspace{1cm} (3)$$

where $E$ and $F$ are defined as,

$$E = 1 - \frac{\alpha r_w^2}{2T} \{ \Psi \text{Ker}(x_w) + \Phi \text{Kei}(x_w) \}$$  \hspace{1cm} (4)

$$F = \frac{\alpha r_w^2}{2T} \{ \Phi \text{Ker}(x_w) - \Psi \text{Kei}(x_w) \}$$  \hspace{1cm} (5)

where

$$\Psi = \frac{-\left[ \text{Ker}_1(x_w) - \text{Kei}_1(x_w) \right]}{2^{1/2} x_w \left[ \text{Ker}_2^2(x_w) + \text{Kei}_2^2(x_w) \right]}$$

$$\Phi = \frac{-\left[ \text{Ker}_1(x_w) + \text{Kei}_1(x_w) \right]}{2^{1/2} x_w \left[ \text{Ker}_2^2(x_w) + \text{Kei}_2^2(x_w) \right]}$$

$$x_w = \left( \frac{\omega S}{T} \right)^{1/2} r_w$$

$$S = bs_w$$

and where $A$ is the amplitude response which is the ratio of amplitude of water level oscillation $x_0$ inside the well to the imposed tidal dilatation strain $e_0$. $S$ is the storativity of the aquifer equal to the multiplication of the undrained specific storage $S_s$ (1/m) and the well open interval $b$ (m), $T$ is the aquifer transmissivity (m$^2$/s), $r_w$ is the radius of the well (m), $r_c$ is the radius of the casing (m), and $\omega$ is the angular frequency of the oscillation (rad/s). Ker and Kei are Kelvin functions of order zero, and Ker$_1$ and Kei$_1$ are Kelvin functions of order one [Hsieh et al., 1987]. The range of the phase response is from $-90^\circ$ to $0^\circ$, and the phase response cannot be positive for the horizontal flow. The transmissivity $T$ and storativity $S$ can be inverted from the observed amplitude and phase response using equations (2) and (3) by minimizing the misfit between the observed and predicted response.

### 4.1.2. Vertical Flow Model

When the water head gradient is vertical, the water tidal response can be simulated by periodic loading applied at the surface of a half space [Wang, 2000]. The boundary condition at the water table is drained, and at infinite depth, the boundary condition is undrained. The response of water level to the tidal forcing under these conditions is [Wang, 2000]:

$$A = \left| \frac{x_0}{e_0} \right| = \frac{1}{S_s} \left[ 1 - 2 \exp \left( -\frac{z}{D} \right) \cos \left( \frac{z}{D} \right) + \exp \left( -\frac{2z}{D} \right) \right]^{1/2}$$  \hspace{1cm} (6)

$$\eta = \arg \left( \frac{x_0}{e_0} \right) = -\tan^{-1} \left( \frac{\exp \left( -\frac{z}{D} \sin \frac{\delta}{D} \right)}{1 - \exp \left( -\frac{2\delta}{D} \cos \frac{\delta}{D} \right)} \right)$$  \hspace{1cm} (7)

where, $\delta = \sqrt{\frac{z}{2D}}$ and $z$ is the depth from the surface; $D$ is the hydraulic diffusivity (m$^2$/s) which equals the division of transmissivity $T$ and storativity $S$. The other terms are defined as the same in section 4.1.1. The range of the phase response is from $-1^\circ$ to $45^\circ$ and the vertical flow model cannot cause negative phase lag beyond $-1^\circ$.

To measure the amplitude and phase response of water level to tidal forcing for the horizontal flow and vertical flow, we analyze the response in the time domain [Xue et al., 2013]. The four largest tidal constituents,
O1, K1, M2, and S2, are used to do the least squares fit for both water level oscillation and the synthetic tidal dilatation strain. The synthetic dilatational strain data are generated by a theoretical tidal code which includes both the solid Earth and ocean-loading tides [Agnew, 2012]. As a check, we used a local gravimeter to measure the tides in Logan Quarry, and found there is no phase shift between the theoretical and observed tides. The method and parameters of the tidal response analysis are the same as the process in Xue et al. [2013].

4.2. Error Estimation

To assess how reliable the range of the estimated phase and amplitude responses are, we estimate the model covariance matrix by using the data covariance matrix. The least squares linear inversion problem in this study can be expressed as:

\[ Gm = d + e \]  

and

\[ G = \begin{bmatrix} \sin \omega_1 t_1 & \cos \omega_1 t_1 & \ldots & \sin \omega_2 t_1 & \cos \omega_2 t_1 \\ \vdots & \vdots & \vdots & \vdots & \vdots \\ \sin \omega_1 t_j & \cos \omega_1 t_j & \ldots & \sin \omega_2 t_j & \cos \omega_2 t_j \end{bmatrix} \]  

where \( G \) is the operation matrix which is composed by synthetic harmonic functions, \( \omega_i \) is the angular frequency of the \( i \)th chosen tidal constituent, and \( t_j \) is the time, \( m \) is the model parameter matrix which is the coefficient for the combination of sine and cosine function, \( d \) is the water level oscillations, and \( e \) is the error vector of the water level data. It is a vector of \( \frac{r_1}{2} \ldots \frac{r_i}{2} \) , where \( r_i \) is the linear misfit of the \( i \)th recorded data point, and the data covariance matrix is \( \text{diag} \{ \frac{r_1}{2} \ldots \frac{r_i}{2} \} \). Since it is a linear inversion, the model covariance matrix can be expressed by the data covariance matrix as

\[ \text{cov} (m) = G' \text{cov} (d) G \]  

The errors of the model give the range of the coefficient of the combination of sine and cosine function. The errors of phase and amplitude response can be calculated by error propagations of the calculated coefficient errors. The resulting phase resolution is about 0.5° for all the data sets except for DW4, and amplitude response resolution is \( 1.8 \times 10^4 \) m/strain. The recorded water level oscillations of DW4 are near the resolution of the HOBO sensor, so the resulting error, which is about 1°, is larger than for the rest of the observations. After the logger of DW4 was changed to the RBR pressure sensor with 0.03 cm resolution, the estimated phase error of DW4 is about 0.7°.

The equivalent \( T \) and \( S \) is determined by the minimum misfit of the forward fitting to the observed phase and amplitude response by applying a large range of \( T \) and \( S \) to equations (2) and (3) when the flow is horizontal or to equations (6) and (7) when the flow is vertical. The errors of the equivalent \( T \) and \( S \) are constrained by the results corresponding to a reasonable range of the fitting residual. The fitting residual of the phase is allowed to be changed by 0.25° which is the phase resolution at the frequency of M2. The fitting residual of the amplitude response is allowed to vary within the error of the resolution of each pressure transducer. The upper and lower bounds of the equivalent hydrogeologic properties are determined by propagating the upper and lower bounds of the phase and amplitude response to the inversion of the equivalent hydrogeologic properties.

5. Results

5.1. Spectrum of Water Level Oscillations

To evaluate whether water level inside each water well responds to the imposed tidal forcing, we computed spectra of both water level data and barometric pressure data (Figure 4). The spectra of all the wells show clear tidal constituents at O1, K1, S2, and M2, which indicate excellent response to tidal forcing. However, the spectrum of the observed barometric data shows clear constituents at K1 and S2 which obscure the tidal response at these two frequencies. Therefore, we focus on the tidal response at the frequency of M2, which has the strongest component with little contamination from the barometric loading, to infer the hydraulic storage (S) and transmissivity (T). All of the wells other than DW4 have larger spectral amplitude
at M2 than S2, so the tidal response analyses of the M2 constituent of these three wells are reliable. The interpretation of the tidal response of DW4 at M2 requires more caution because of the low signal to noise ratio of M2.

5.2. Tidal Response

The phase responses are in the range of $217^\circ$ to $38^\circ$. The phase responses show an increasing phase lag (increasing absolute number) trend as a function of distance of the well away from the fault zone, with the exception of DW4. The phase response of DW4 is highly variable. This scatter is due to the strong effect of the barometric pressure of the S2 component in this site as discussed earlier and the limitation of the logger resolution. The amplitude of water level oscillation inside DW4 is close to the resolution of the HOBO, and therefore it is hard to resolve the phase information accurately for this data. The phase response of DW4 measured by the HOBO sensor therefore will not be interpreted, and we prefer to use the phase response of DW4 measured by the RBR sensor to constrain the hydrogeologic properties of the formation surrounding DW4. The phase response of DW4 measured by the RBR sensor is about $215^\circ$ (Figure 5, the dots after May, 2015). As discussed above, the small positive phase lag of DW13 might be due to the effect of a vertical flow [Roeloffs, 1996], and vertical flows were observed at the nearby artesian well, so we will use the vertical flow model to calculate the equivalent hydrogeologic properties of DW13 in the later section. The magnitude of the observed water level oscillations is $<1$ cm, so the corresponding ratio between the measured water level and the theoretical tidal dilatation strain is of the order of magnitude $10^4$ m/strain.

5.3. Hydrogeologic Properties

The hydraulic transmissivity ($T$) and storativity ($S$) determine the phase and amplitude responses, the in situ $T$ and $S$ therefore can be inferred from the measured phase and amplitude response. The inverted transmissivity and storativity depend on the length of the well open interval. To avoid the effect of different well completions (Figure 2), $T$ and $S$ are converted to permeability ($k$) and specific storage ($S_s$) by:
\[ k = \frac{\mu}{\rho gb} \]  
\[ S_s = S/\mu \]  

where \( k \) is the fluid dynamic viscosity, \( b \) is the length of the open interval of the well, and \( \rho \) is the density of fluid [Freeze and Cherry, 1977]. Using \( \mu = 10^{-3} \) Pa s at 20°C, \( \rho = 10^3 \) kg/m³, \( g = 9.8 \) m/s², the calculated permeability \( k \) and specific storage \( S_s \) of the formation surrounding these four wells are shown in Figure 6. Since the variations of the permeability and specific storage are not big, we take the average value for each well (Figure 7). The permeability of the wells closest to the fault (DW4 and DW13) is larger than those of the further wells (DW2 and DW11) by a factor of \( \approx 10 \) (Figure 7a). Similarly, the specific storage of the closer wells is larger than those of the further ones. The specific storage of DW4 is larger than that of DW2 and DW11 by a factor of \( \approx 10 \), and the specific storage of DW13 is larger than that of DW2 and DW11 by a factor of 2 (Figure 7b). Interestingly, the combined effects of the permeability and storage result in a hydraulic diffusivity of about \( 10^{-2} \) m²/s for all of the wells (Figure 7c).

In combination, the data suggest that there is heterogeneity in permeability and storage, but not hydraulic diffusivity. The apparent trend in permeability and storativity with distance to the fault may indicate the influence of fracturing associated with the fault zone.

5.4. Response to Earthquakes

Other recent studies have observed earthquake-well interactions [Elkhoury et al., 2006; Manga et al., 2012; Xue et al., 2013]. Even though the primary purpose of the study is to investigate hydrogeologic architecture, we took the opportunity to also examine the effects of earthquakes on the wells. We utilize a local broadband seismic station BK.SAO [NCEDC, 2014] as a reference station to select earthquakes with the largest ground motion at the Quarry Site. The record displacements are filtered at 0.5–15 Hz, and corrected to the displacement at the Quarry using the attenuation equation from Kanamori et al. [1993] and corrected by H. Kanamori (personal communication, 2004):

\[ \log(Ac) = \log(A) + 1.2178 \times \log(D_1/D_2) + 0.002302 \times (D_1 - D_2) \]  

where \( Ac \) is the corrected displacement at the study site, \( A \) is the recorded displacement, \( D_1 \) is the distance from the earthquake to the record station (km), and \( D_2 \) is the distance from the earthquake to the study site (km).
We found that a sequence of local earthquakes with magnitudes as large as 4.2 occurred within 30 km of the Quarry between 20 November 2014 and 6 December 2014 caused the largest ground motions at the Quarry site (Figure 1b). It is hard to separate the response to individual earthquakes in the sequence, because the time window of the phase response analysis is ~29.5 days. The continuous phase response of all wells except DW4 has subtle increases in that interval. As previously discussed, the temporal phase response change of DW4 is not reliable in this time interval.

To clearly separate the properties before and after the earthquakes, we invert the data omitting the segment overlap from 20 November to 6 December (Figure 8). The phase response had a slight increase right after the M\(_L\) 3.9 earthquake on 6 December, and recovered the background level within 2 months. The change of phase lag is measured by the difference between the first data point after the earthquake and the averaged phase value before the earthquake. The change of equivalent permeability is calculated from the changes of phase response. After the sequence of earthquakes, the permeability of DW13, DW2, and DW11 is increased by 38%, 160%, and 19%, respectively. The rapid height changes of Pajaro River at that time were related to the large rainfall (Figure 3) and unlikely related to the permeability enhancement which lasted for a longer time.

These earthquakes produced only one identified phase response during our observations. The largest regional earthquake during the observation period was the 24 August 2014 Napa earthquake. No observable permeability response occurred in the site during the Napa earthquake. The Napa earthquake even did not cause

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**Table 2. Hydrogeologic Properties of the Formation Surrounding Each Well**

<table>
<thead>
<tr>
<th>Well Name</th>
<th>( R^a ) (m)</th>
<th>( k ) (m(^2))</th>
<th>( S_s ) (l/m)</th>
<th>( D \times 10^{-2} ) m(^2)/s</th>
<th>( L^a ) (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>DW4</td>
<td>21.7</td>
<td>( 2.3 \times 10^{-14} )</td>
<td>( 1.5 \times 10^{-5} )</td>
<td>1.5</td>
<td>30.5</td>
</tr>
<tr>
<td>DW13</td>
<td>50.2</td>
<td>( 1.6 \times 10^{-14} )</td>
<td>( 2.8 \times 10^{-6} )</td>
<td>5.6</td>
<td>36.6</td>
</tr>
<tr>
<td>DW2</td>
<td>20.8</td>
<td>( 1.5 \times 10^{-15} )</td>
<td>( 1.6 \times 10^{-6} )</td>
<td>1.0</td>
<td>426.7</td>
</tr>
<tr>
<td>DW11</td>
<td>26.7</td>
<td>( 2.2 \times 10^{-15} )</td>
<td>( 1.4 \times 10^{-6} )</td>
<td>1.6</td>
<td>548.6</td>
</tr>
</tbody>
</table>

\( R^a \) is diffusion length, which is \( R = \sqrt{Dt} \), where \( D \) is the inferred diffusivity m\(^2\)/s and \( t \) is 12.42 h, which is the period of semidiurnal tidal component M\(_2\). \( R \) controls the radius of influence. \( L \) is the distance to the fault zone.
any observable disturbances to water levels (Figure 3). Local seismometer indicates that the ground motion of the local earthquakes in the 20 November is larger than that from the Napa earthquake.

6. Discussion

6.1. Hydrogeologic Properties

We have estimated the permeability of rock near the San Andreas Fault to be about $10^{-14}$ m$^2$ which is a relatively high number compared to the permeability of other fault zones. The permeability from the isotropic model gives the lower bounds on the highest permeability of the system and therefore is a conservative estimate [Xue et al., 2013]. The laboratory measured permeability from core samples of the Nojima fault zone indicates that the fault core permeability is $10^{-19}$–$10^{-18}$ m$^2$, and damage zone permeability is $10^{-17}$ and $10^{-16}$ m$^2$ [Lockner et al., 2000, 2009]. The in situ measured permeability of Wenchuan rupture zone is about $10^{-16}$ m$^2$ [Xue et al., 2013]. The laboratory measured permeability of granite generally has a range of $10^{-16}$–$10^{-21}$ m$^2$ [Brace, 1980]. Our measured permeability is at the high end of the laboratory values. This likely indicates that our measured media is dominated by mesoscale fractures. The measured hydrogeologic diffusivity of this site is about $\sim 10^{-2}$ m$^2$/s, which is similar to the value of $2.4 \times 10^{-2}$ m$^2$/s inferred from the Wenchuan Earthquake fault [Xue et al., 2013].

6.2. Hydrogeologic Architecture

The trend of variation between the wells along strike direction is not significant. DW13 and DW2 are almost identical along the strike direction, but they have different properties. The results therefore are interpreted as reflecting trends perpendicular to the fault. The fundamental observation is that the paired wells on the main strand of the San Andreas Fault have higher permeability than the pair farther away (Figure 7). As the wells move farther away from the fault, the observed amplitude response decreases and the phase lag except for DW4 increases (absolute number increases). To first order, the amplitude response is controlled by the storage $S$ and the phase lag is controlled by the transmissivity $T$. Therefore, a naive interpretation of the data might be that both storage and transmissivity increase close to the fault. However, the storage $S$ can contribute to phase lag and the transmissivity $T$ can also affect the amplitude response. Increasing transmissivity can also increase the amplitude response as the well drains more easily during each oscillation, and larger storage can also reduce the phase lag as the aquifer releases more water during each oscillation. Therefore, we need to assess whether the observed tidal response patterns could potentially be explained by varying only $S$, or $T$ without a contribution from the other factor.

For fixed specific storage for all the wells, the decreasing phase lag closer to the fault except DW4 suggests an increasing transmissivity $T$ closer to the fault. The increasing transmissivity $T$ would result in an increasing amplitude of water level oscillations. This is opposite to the observed trend of the amplitude response. Overall, both the trend of the amplitude and phase response cannot be caused by differences in transmissivity only.
The alternative case of specific storage varying requires more calculation. To test this, we want to investigate what value of $S$ is needed to obtain the observed phase and whether its resulting amplitude response agrees with the observations. We therefore first calculate the responses for a fixed transmissivity with only specific storage varying (Figure 9). The fixed transmissivity is the average of $T$ of $6.2 \times 10^{-6}$ m$^2$/s based on the inversion of $T$ and $S$ at the same time. We invert for the corresponding storativity for the observed phase lag (Figure 9a), then compare the synthetic amplitude response calculated from the inverted storativity and the fixed transmissivity to the observed amplitude response (Figures 9b and 9c). The variation of storage alone can result in the observed trend of the amplitude response, but the difference of the absolute value of the synthetic amplitude response is ~4 orders of magnitude larger than the difference of the observations. The absolute value of the synthetic amplitude response depends on the given $T$, but the large differences between the synthetic amplitude responses cannot be compensated by a chosen $T$. Overall, the comparison indicates that the observed systematic changes of phase and amplitude response between the sites have to be the result of changes in both $S$ and $T$.

The measured permeability $k$ indicates that sites far away from the fault have smaller $k$ than the places which are closer to the fault. The measured $S$ surprisingly shows a clear decreasing trend outward from the fault (Figure 7) which indicates that the formation has higher compliance when it is closer to fault zone. Based on the inferred diffusivity of the formation surrounding each well, the length scale surrounding each well over which the tidal response samples is several tens of meters (details see Table 2). The distances between any two wells are several hundreds to thousands of meters. The length scale of each well over which the tidal
response is sampled is much smaller than the distances between wells, therefore, the radius of the influence zone of each well is nonoverlapping regions. There is no trend in the sampled depth of wells (Figure 2), so the observed trend of hydrogeologic properties is unlikely caused by the heterogeneity of the depth to the surface. The sites close sampled within 40 m of the fault have different hydrogeologic properties than the sites sampled >400 m to the fault (Figure 7a,b). We infer that the different hydrogeologic properties of this localized zone might be related to the fault zone fracture density decaying with the distance away from the fault [Savage and Brodsky, 2011]. Inside the Long Quarry, the fracture density has been quantified but has not been mapped to vary systematically with distance to fault because of the intensely fractured nature of the gabbro.

The apparent contrast in specific storage across a fault zone affects the fault’s response to applied stress. The damage in a fault zone results in the fault having higher compliance and a larger elastic response than the surrounding formation. Similar compliance contrasts have been observed in the nearby faults of the 1999 Hector Mine earthquake [Fialko et al., 2002]. The observed induced deformation over a kilometers-wide zone on the nearby faults of the Hector Mine earthquake could be caused by a rigidity contrast of 0.4–0.6. Guided seismic waves also show that the Landers Fault zone has a 100–300 m wide weaker zone extending tens of kilometers deep with seismic wave velocities reduced up to 30%–50% [Li et al., 1994, 1998]. This velocity reduction suggests a corresponding modulus contrast of a factor of 0.5–0.7 assuming there is no change of rock density. Our observed hydrogeologic specific storage contrast indicates a bulk modulus contrast of a factor of 0.1–0.5 which is comparable to the results of the mentioned geodetic and seismic observations. All of these observations together indicate that the fault zone is mechanically weaker than the surrounding rocks. Barbour [2015] recently reported apparent spatial modulus reduction near major faults based on the seismic wave response of pore pressure. That work showed a multikilometer-scale architecture that was interpreted as indicative of strain rate control. We calculated a relative strain rate change of about 0.2% within 400 m to the fault by assuming a vertical plane in a half space with a slip rate of 30 mm/yr and a locking depth of 10 km [Segall, 2010]. For instance, if the strain rate on the fault is 0.4775 microstrain/yr, then the strain rate at 400 m to the fault is 0.4767 microstrain/yr. The subtle strain rate change unlikely affects the tidal response. Here we are therefore focusing on the subkilometer-architecture and see no direct evidence of strain rate control.

The measured diffusivity is about $10^{-2}$ m$^2$/s which is the same order of magnitude of that of the Wenchuan Earthquake fault zone ($2.4 \times 10^{-2}$ m$^2$/s). It is interesting that the high diffusivity at the location near the fault is the same at both other two unrelated sites, even though the permeability and specific storage have large contrasts (Figure 7). This indicates the diffusivity structure of the fault zone is unexpectedly uniform due to the competing effect of permeability and specific storage. The uniform diffusivity might suggest the accumulated pore pressure during the interseismic period distributes across a broad zone rather than within a narrow channel. The consistency across sites might also suggest that fault zones evolve to a narrow range of diffusivities.

6.3. Response to Earthquakes

The 24 November and 6 December earthquakes were accompanied by the largest rate change of permeability (Figure 8a), so we focus on the study of this most significant response to earthquakes. Other apparent variations of permeability do not vary as rapidly and do not correspond to any earthquakes. After the
earthquake sequences during 24 November and 6 December, all wells have a phase change of 1.7°–3.0° except DW4. Previous work suggests the contrast in specific storage may also cause a pressure gradient to drive flow though the fault zone and surrounding rock when seismic waves pass [Brodsky and Prejean, 2005]. This resultant flow can cause permeability enhancement as shown in laboratory oscillation experiments [Candela et al., 2015]. To assess whether the flow driven by seismic waves can cause permeability enhancement based on our observed in situ specific storage contrast, we utilize the same model as shown in the study of Brodsky and Prejean [2005] of two abutting media with distinct storages. This model calculates the induced hydraulic heads caused by dilatational seismic strain in faults and the surrounding rocks with different storages. Based on this model, the flow rate can be calculated by:

\[
q = -jK \left( \frac{d}{\frac{1}{\mathcal{S}_f} + i\omega} \right) \frac{1}{\mathcal{S}_r} \exp(-j\chi)
\]

where \( q \) is the flow rate (m/s), \( K \) is hydraulic conductivity (m/s), \( \mathcal{S}_f \) and \( \mathcal{S}_r \) are the specific storage for the fault zone and surrounding rock, respectively (m\(^{-1}\)), \( \omega \) is angular frequency of the imposed strain (rad/s), \( \theta_\infty \) is the imposed strain, \( \kappa_s \) is the diffusivity of the rock (m\(^2\)/s), and \( d \) is the fault zone width (m) [Brodsky and Prejean, 2005].

Assuming a fault zone has a 40 m wide weak zone with a specific storage of \( 1.47 \times 10^{-5} \) m\(^{-1}\), and the surrounding rock has a smaller specific storage of \( 1.36 \times 10^{-5} \) m\(^{-1}\) and a permeability of \( 10^{-15} \) m\(^2\). The resulting maximum flow rate is \( 9 \times 10^{-8} \) m/s near the edge of the fault zone by choosing a wave with period of \( 10 \) s and strain amplitude of \( 1.2 \times 10^{-6} \). The chosen seismic strain is calculated by using the largest recorded displacement of 0.37 cm and the general Rayleigh wave velocity of 3 km/s. The flow in this model is extremely local and decays to a negligible level within 10 diffusion lengths (~0.1 m) of the boundary for the high-frequency seismic waves. Permeability enhancement was observed in laboratory experiments for flow velocities as low as \( 5 \times 10^{-6} \) m/s [Candela et al., 2014]. The flow rate inferred by the observed storage architecture is lower than the experimental range resolvable in the laboratory experiments, which indicates that further experiments are needed to investigate the potential effects of lower flow rates.

7. Conclusions

The observed hydrogeologic architecture of the San Andreas Fault Zone in the Logan Quarry indicates a localized region near the fault zone which differs from the surrounding rock. Sites within ~40 m of the fault zone have a larger specific storage and a larger permeability than those that are more than 400 m away. This hydrogeologic property structure could be related to the fracture density distribution of the fault zone.

The measured specific storage structure is novel and has consequences for fault mechanics. The competing effects of the permeability and specific storage changes result in a relatively uniform, high value of diffusivity of \( 10^{-2} \) m\(^2\)/s. This uniform diffusivity structure suggests that the permeability contrast might not efficiently trap fluids during the interseismic period. Our study shows the utility of the tidal response for measuring the fault zone hydrogeologic architecture in situ and studying the response behavior of fault zone hydrogeologic properties to earthquakes. The data also hint that hydraulic diffusivity may evolve to a narrow range of values ~\( 10^{-2} \) m\(^2\)/s in fault zones.

References


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