The influence of poorly interconnected fault zone flow paths on spring geochemistry

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ABSTRACT

Thermal springs commonly occur along faults because of the enhanced vertical permeability afforded by fracture zones. Field and laboratory studies of fault zone materials document substantial heterogeneities in fracture permeabilities. Modeling and field studies of springs suggest that spatial variations in permeability strongly influence spring locations, discharge rates and temperatures. The impact of heterogeneous permeability on spring geochemistry, however, is poorly documented. We present stable isotope and water chemistry data from a series of closely spaced thermal springs associated with the Hayward Fault, California. We suggest that substantial spatial variations observed in δ¹⁸O and chloride values reflect subsurface fluid transport through a poorly connected fracture network in which mixing of subsurface waters remains limited. Our measurements provide insight into the effect of fracture zone heterogeneities on spring geochemistry, offer an additional tool to intuit the nature of tectonically induced changes in fault zone plumbing, and highlight the need to consider local variations when characterizing fracture zone fluid geochemistry from spring systems with multiple discharge sites.

Key words: fault zones, mixing, permeability, springs, stable isotopes

INTRODUCTION

Thermal springs commonly occur in association with fault zones (Curewitz & Karson 1997) because of horizontal impedance and vertical enhancement of groundwater flow by the fault zone (Bredehoeft et al. 1992; Caine et al. 1996; Evans et al. 1997; Rawling et al. 2001). Such springs offer the potential to provide data on flow, pressure, and geochemical properties of fluids within fault zones.

Both field and laboratory studies document fracture networks within fault zones that create substantial permeability variations at scales ranging from centimeters to hundreds of meters (e.g. Antonellini & Aydin 1994; Geraud et al. 1995; Caine et al. 1996; Evans et al. 1997; Rawling et al. 2001; Forster et al. 2003; Kato et al. 2004). Nevertheless, most models of groundwater flow through fault zones treat faults as homogeneous conduits for fluid transport from depth to the surface (e.g. Goyal & Narasimhan 1982; Forster & Smith 1988a,b; Lopez & Smith 1995). Recent field investigations (e.g. Fairley et al. 2003; Fairley & Hinds 2004; Heffner & Fairley 2006), however, show that heterogeneity in fault zone permeability results in local variations in spring temperatures. Numerical models (e.g. Lopez & Smith 1996) indicate that permeability variations strongly influence spring location and discharge along fault traces. The effect of such heterogeneity on spring geochemistry is poorly documented. Heffner & Fairley (2006) document little variation in spring solute chemistry despite substantial temperature variations, while Satake et al. (2003) report distinct changes in ion concentrations across an active fault zone.

Here we document geochemical variations in water from a series of closely located (tens of meters) thermal springs discharging along a branch of the Hayward fault system in San Jose, California. Despite the close proximity of the spring outlets, a continuum of δ¹⁸O and chloride (Cl) values suggests that limited mixing of waters occurs in the...
fracture network feeding the springs. These results both highlight the potential of using changes in spring geochemistry to recognize changes in fault zone fracture plumbing (in response to seismic events) and the need to exercise caution when characterizing fault fluid geochemistry based on a subset of a larger population of springs.

SITE DESCRIPTION, DATA COLLECTION AND ANALYSIS

A series of springs located in the Alum Rock Park, San Jose, California, emanate from laminated cherts of the Claremont Formation near the western boundary of the Alum Rock structural block (Wentworth et al. 1999). Figure 1 presents the location of the springs relative to local faults and regional topography. The Alum Rock Block consists of steeply east-dipping strata bounded on the east by the Calaveras fault and to the west by transpressional, reverse-right-lateral strands of the Hayward fault system (Wentworth et al. 1999). The springs lie approximately 100 m up-dip (east) of a mapped Holocene strand of the Hayward fault (Wentworth et al. 1999; Graymer et al. 2006).

The springs occur in a canyon of Penitencia creek at an elevation of approximately 180 m, several meters above the creek and approximately 400 m below the adjacent hill crests. They discharge along both sides of the creek extending approximately 250 m north from the local range front (the inferred fault trace). Spring outlets vary from seeps to tunnels cut into the hill slopes. The tunnels were created in the early 20th century to direct spring flow to basins for drinking and baths for recreation (Waring 1915). The spring waters are no longer used for consumption or recreation and many of the outlets are in disrepair with flow no longer issuing from two of the four tunnels mentioned by Waring (1915). Waring (1915) noted that all springs were ‘sulphureted’ but varied in salt and mineral contents.

Between 2003 and 2007, we periodically made field measurements of flow, temperature, and alkalinity and collected samples for stable isotope analysis (δ18O and δD). We collected up to eight isotope samples at the high discharge springs (4 and 11), and two to five samples at the other springs. In addition, samples were analyzed for chloride at nine springs, major ions at two springs (4 and 11), and tritium (3H) at spring 4.

Fig. 1. Generalized map of the southern portion of the Hayward Fault system showing the locations of two sets of thermal springs relative to major faults and regional topography. Fault locations adapted from Andrews et al. (1993).
Samples were prepared for stable isotope analysis using the water–CO₂ equilibration method for oxygen (Epstein & Mayeda 1952) and the zinc-reduction method for deuterium (Coleman et al. 1982). Isotope analyses were performed on an isotope ratio mass spectrometer. Results are reported in the conventional delta notation as per mil deviations from the Vienna-Standard Mean Ocean Water (V-SMOW) reference standard. Analytical precision is approximately 0.2‰ on δ¹⁸O and 1‰ on δD values. Major dissolved cations and anions were analyzed on an ion chromatograph equipped with an electrochemical detector. Dissolved Si concentrations were analyzed by inductively coupled plasma atomic emission spectrometer (ICP-AES). Tritium analyses were made using the ³He accumulation method (Surano et al. 1992) with an analytical detection limit of 1 pCi l⁻¹.

RESULTS

Table 1 provides a summary of both field measurements of temperature, flow rate, pH, and alkalinity, as well as, the results of laboratory analysis for stable isotopes and chloride concentrations of the Alum Rock spring waters. The collective discharge from 12 springs with measurable discharge is 0.516 ± 0.026 l s⁻¹ [mean ± standard deviation (SD)] or 0.1% of the annual recharge to the Penitencia Creek basin. Two springs (no. 4 and 11; both tunnels), account for 75% of the total discharge with flows of 0.15 ± 0.03 and 0.23 ± 0.05 l s⁻¹, respectively. The discharge of the 10 remaining springs ranges from 0.003 to 0.04 l s⁻¹. Spring discharge appears insensitive to rainfall events, but does respond to precipitation trends on multi-year timescales (King et al. 1994). Since the early 20th century, overall spring discharge appears to have remained relatively stable. Waring (1915) reported that when first developed for recreation, few springs exceeded flows of 0.06 l s⁻¹ and none exceeded 0.32 l s⁻¹. Since that time, reported flow rates for spring 11 have ranged from 0.32 to 0.27 l s⁻¹ (Berkstresser 1968; King et al. 1994; Melchiorre et al. 1999) and in the 1960s Berkstresser (1968) reported a value of 0.13 l s⁻¹ for spring no. 4. We are uncertain whether the lower discharges measured at spring no. 11 in this study result from variations in sampling method, as it is difficult to collect all of the water flowing out of irregular tunnel orifices, or whether the measured decrease represents a real drop in spring discharge over the last decade.

Spring temperatures range from 15 to 29°C, with the lowest values at the low-flow seeps representing ambient temperatures. Temperatures at springs 4 and 11 averaged 28.9 ± 0.74 and 26.5 ± 1.5°C, respectively, during continuous monitoring from November 2002 to May 2003. Waring (1915) reported a temperature range of 20–30°C at the springs in the early 20th century.

<table>
<thead>
<tr>
<th>Spring</th>
<th>Temp (°C)</th>
<th>Flow (l s⁻¹)</th>
<th>δ¹⁸O</th>
<th>δD</th>
<th>pH</th>
<th>Alkalinity (ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>17.8 ± 2</td>
<td>0.03 ± 0.03</td>
<td>-1.6 ± 1.7</td>
<td>4</td>
<td>5.98 ± 0.17</td>
<td>707 ± 47</td>
</tr>
<tr>
<td>2</td>
<td>19.5 ± 0.7</td>
<td>0.01 ± 0.03</td>
<td>-1.5 ± 1.5</td>
<td>4</td>
<td>5.88 ± 0.02</td>
<td>530 ± 6</td>
</tr>
<tr>
<td>3</td>
<td>19.0 ± 1.7</td>
<td>0.00 ± 0.03</td>
<td>0.9 ± 0.9</td>
<td>4</td>
<td>5.94 ± 0.06</td>
<td>1657 ± 275</td>
</tr>
<tr>
<td>4</td>
<td>20.9 ± 0.7</td>
<td>0.01 ± 0.03</td>
<td>0.1 ± 0.1</td>
<td>4</td>
<td>6.18 ± 0.12</td>
<td>1667 ± 275</td>
</tr>
<tr>
<td>5</td>
<td>20.2 ± 1.3</td>
<td>0.00 ± 0.03</td>
<td>-1.0 ± 1.0</td>
<td>4</td>
<td>6.49 ± 0.04</td>
<td>1667 ± 275</td>
</tr>
<tr>
<td>6</td>
<td>20.5 ± 1.3</td>
<td>0.00 ± 0.03</td>
<td>-1.7 ± 1.7</td>
<td>4</td>
<td>6.49 ± 0.04</td>
<td>1667 ± 275</td>
</tr>
<tr>
<td>7</td>
<td>20.7 ± 2</td>
<td>0.00 ± 0.03</td>
<td>-2.5 ± 2.5</td>
<td>4</td>
<td>6.41 ± 0.02</td>
<td>1667 ± 275</td>
</tr>
<tr>
<td>8</td>
<td>21.0 ± 2</td>
<td>0.00 ± 0.03</td>
<td>-3.0 ± 3.0</td>
<td>4</td>
<td>6.29 ± 0.02</td>
<td>1667 ± 275</td>
</tr>
<tr>
<td>9</td>
<td>21.0 ± 2</td>
<td>0.00 ± 0.03</td>
<td>-3.0 ± 3.0</td>
<td>4</td>
<td>6.29 ± 0.02</td>
<td>1667 ± 275</td>
</tr>
<tr>
<td>10</td>
<td>21.0 ± 2</td>
<td>0.00 ± 0.03</td>
<td>-3.0 ± 3.0</td>
<td>4</td>
<td>6.29 ± 0.02</td>
<td>1667 ± 275</td>
</tr>
<tr>
<td>11</td>
<td>21.0 ± 2</td>
<td>0.00 ± 0.03</td>
<td>-3.0 ± 3.0</td>
<td>4</td>
<td>6.29 ± 0.02</td>
<td>1667 ± 275</td>
</tr>
<tr>
<td>12</td>
<td>21.0 ± 2</td>
<td>0.00 ± 0.03</td>
<td>-3.0 ± 3.0</td>
<td>4</td>
<td>6.29 ± 0.02</td>
<td>1667 ± 275</td>
</tr>
<tr>
<td>13</td>
<td>21.0 ± 2</td>
<td>0.00 ± 0.03</td>
<td>-3.0 ± 3.0</td>
<td>4</td>
<td>6.29 ± 0.02</td>
<td>1667 ± 275</td>
</tr>
</tbody>
</table>

n, number of measurements; *Temperature recorded by dataloggers at 12 min intervals.
Figure 2A shows that $\delta^{18}O$ values of the Alum Rock spring waters all plot to the right of the Global Meteoric Water Line (GMWL). Springs are enriched in $\delta^{18}O$ relative to local meteoric waters by 0.4‰ (spring 4) to 4.2‰ (spring 8). The Alum Rock springs appear to separate into two groups with statistically distinct ($P < 0.001$) mean $\delta D$ values ($-48_{-65}^{+77}$ and $-54_{-9}^{+3}$, Fig. 2A), which reflect differing local sources of meteoric waters. The $\delta D$ values of the isotopically heavier group (springs 1, 7, 8, 9, 10, 11, 12, and 13) appear to correspond to those of Penitencia Creek water sampled just upstream of the Alum Rock spring complex. The $\delta D$ values of the lighter group (springs 2, 4, and 5) resemble waters discharging from cold springs at an elevation of 1000 m in neighboring drainage basins to the east (Melchiorre 1998).

Neither $\delta D$-groupings of springs exhibit a statistically significant slope for the best-fit linear regressions of $\delta D$ versus $\delta^{18}O$. Some of the variability in $\delta D$ seen in Fig. 2A appears to have a seasonal dependence, with samples collected in spring (March to May) exhibiting a $1_{-5}^{+3}$‰ decrease relative to values measured in winter and fall.

The Alum Rock spring waters exhibit a linear relationship of increasing Cl content with increasing $\delta^{18}O$ values (Cl = 3238$^{18}$O + 2524, $r^2 = 0.91$ and $P < 0.001$; Fig. 2B). The apparent grouping observed in the $\delta D$ data is not reflected in the $\delta^{18}O$ versus Cl relationship.

Springs with high Cl and heavier $\delta^{18}O$ values are spatially clustered toward the center of the Alum Rock spring complex (Fig. 3A,B). A correlation between flow and geochemistry is not apparent (Fig. 3C); spring 4 with the second highest flow rate has the lowest Cl and $\delta^{18}O$ values, but spring 11 with the highest flow rate has Cl and $\delta^{18}O$ values which fall in the middle of the overall range for the spring complex.

A water sample collected from spring 4 during the dry season contained no detectable tritium (<0.2 pCi l$^{-1}$). From the absence of tritium we infer that the spring discharges waters which were recharged to the subsurface a minimum of 40 years before sampling. Of all the Alum Rock springs, the stable isotope composition of spring 4 lies the closest to the meteoric water line. Therefore, we infer that none of the springs contains a significant component of young recharge (<40 years).

Dissolved ion data from springs 4 and 11 are presented in Table 2. In an attempt to constrain the depth to which the spring waters circulate, we used a number of chemical geothermometers to estimate water–rock equilibrium temperatures using this dissolved ion data. Chalcedony (Fournier 1977), Mg-K (Giggenbach 1997), Na-K (Kharaka & Mariner 1989; Giggenbach 1997), Na-Li (Fouillac & Michard 1981; Kharaka & Mariner 1989), Na-K-Ca (Fournier & Truesdell 1973) and Mg-Li (Kharaka & Mariner 1989) methods resulted in equilibrium temperature estimates ranging from 33 to 200°C, with no consistent estimates between methods. The inconsistency of geothermometers is not surprising given the mixing of the waters during transport to the surface and the high CO$_2$ content of the springs suggested by the high HCO$_3^-$ and low pH of the waters (Table 2) (Fournier 1977). Therefore, based on water chemistry, we were unable to determine the circulation depth of the flow system feeding the springs.
DISCUSSION

Chloride-rich springs in the northern and eastern California Coast Ranges commonly show linear trends on δD–δ18O plots that represent mixing of tectonically expelled connate water or brines with local meteoric water (White et al. 1973; Donnelly-Nolan et al. 1993; Peters 1993; Davisson et al. 1994; Melchiorre et al. 1999). In many cases, the connate fluid is highly enriched in 18O and deuterium relative to local meteoric waters. Water–rock isotopic exchange at elevated temperatures can account for the observed δ18O enrichment (White et al. 1973; O’Neil & King 1981), whereas the deuterium enrichment may reflect an evolved marine pore water composition or metamorphic fluid (White et al. 1973; Peters 1993).

The composition of many coastal range waters suggest mixing with both evolved connate and metamorphically shifted nonmeteoric waters (White et al. 1973; Donnelly-Nolan et al. 1993). Both the Br/Cl and B/Cl ratios of Alum Rock waters, however, suggest that mixing of a 18O-shifted seawater or connate water with local meteoric waters may account for the range in isotopic concentrations observed at these springs. Br/Cl ratios of 0.002 and 0.003 (Table 2) for springs 4 and 11, respectively, are consistent with seawater ratios (White et al. 1973; Peters 1993). Previously reported B/Cl ratios of 0.02 to 0.06

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**Table 2** Major ion chemistry (mg l⁻¹) for springs 4 and 11.

<table>
<thead>
<tr>
<th>Spring no.</th>
<th>F</th>
<th>Cl</th>
<th>Br</th>
<th>I</th>
<th>NO₂</th>
<th>NO₃</th>
<th>SO₄</th>
<th>PO₄</th>
<th>Na</th>
<th>K</th>
<th>Mg</th>
<th>Ca</th>
<th>Li</th>
<th>Si</th>
<th>SiO₂</th>
<th>HCO₃</th>
</tr>
</thead>
<tbody>
<tr>
<td>4</td>
<td>1.7</td>
<td>138.8</td>
<td>0.3</td>
<td>0.07</td>
<td>0.0</td>
<td>38.9</td>
<td>60.0</td>
<td>0.00</td>
<td>469</td>
<td>17.0</td>
<td>16.4</td>
<td>11.2</td>
<td>9.8</td>
<td>21.0</td>
<td>1033</td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>1.6</td>
<td>398.7</td>
<td>1.3</td>
<td>0.40</td>
<td>41.9</td>
<td>1.5</td>
<td>285.1</td>
<td>0.00</td>
<td>599</td>
<td>16.5</td>
<td>31.8</td>
<td>3.0</td>
<td>0.35</td>
<td>10.5</td>
<td>22.5</td>
<td>675</td>
</tr>
</tbody>
</table>

Reported HCO₃ values were calculated based on the charge balance for all other ionic species. A water sample from spring 11 collected by one of the authors (TPR) in December 1995 had a measured dissolved inorganic carbon (DIC) concentration of 2061 mg l⁻¹. Using the carbonate equilibria data from Drever (1982) and measured pH (6.6) and T (27°C) values, the concentrations of DIC species were as follows: CO₂(aq) = 1300 mg l⁻¹; HCO₃ = 761 mg l⁻¹; CO₃ = 0.2 mg l⁻¹. The fact that charge balance requires only 675 mg l⁻¹ HCO₃ may indicate errors in field pH measurements, which are not uncommon in actively degassing systems. If the pH on the December 1995 sample is decreased to 6.04, the HCO₃ value required for charge balance is obtained. Errors in measured concentrations of other ionic species may also account for some of the discrepancy.
(Berkstresser 1968; Melchiorre 1998; Melchiorre et al. 1999) show a closer affinity to connate waters and brines of the Eastern Coast Ranges than to metamorphically shifted ‘Franciscan’ waters, more prevalent in the Western Coast Ranges (White et al. 1973; Donnelly-Nolan et al. 1993). Figure 4 presents the Alum Rock data along with data from Peters (1993) for both cool springs and Au-bearing thermal waters from the Clear Lake region of the California Coast Ranges. The range in isotopic compositions for all three waters can be explained by the variable mixing of local meteoric waters with 18O-shifted seawater (e.g. White et al. 1973; Peters 1993). Given the limited geochemical data available for the Alum Rock springs, we are unable to postulate a definitive ‘end-member’ composition for the nonmeteoric water component for these spring waters, but such an end-member may lie along the trend line depicted in Fig. 4. The local variability in δD observed at the Alum Rock springs may result from mixing of the 18O-shifted seawater with two sources of local meteoric water (higher altitude recharge or local surface waters) and/or possibly arises from modest δD enrichment because of evaporation. This evaporation may occur both during recharge and from limited ponding of the spring waters at the discharge points in Alum Rock park.

Figure 5 presents a conceptual model of the flow system feeding the Alum Rock springs. At depth, recharging meteoric waters enter the fracture network conveying connate waters expelled from depth to the surface. The mixing of waters results in a dilution of the connate waters and an isotopic shift in the meteoric water composition to heavier isotopic values. During transport to the surface, mixing of waters within the fracture network remains limited such that the springs do not individually integrate the geochemical characteristics of the fault zone. As a result, the spring water geochemistry exhibits considerable variability across a spatially limited area (Fig. 3).

Melchiorre et al. (1999) argued that the expulsion of saline-rich brines along the eastern front of the Coast Ranges resulted from the eastward propagation of shallow lithosphere compression by transpressive movement along local fault zones. Alum Rock data suggest that while connate waters in the western Coast Ranges may be significantly diluted relative to those in the east, such waters may not be entirely flushed from depth.

**POSSIBLE GEOCHEMICAL CONSTRAINTS ON SPRING RESPONSES TO EARTHQUAKES**

Changes in stream flow (Muir-Wood & King 1993), wells, and springs in response to earthquakes are well-documented phenomena (see for review Manga & Wang 2007). These changes include increases and decreases in flow and water levels, increases and decreases in water temperature, and changes in geochemistry. Proposed explanations for these changes include both permanent and temporary alterations of aquifer permeabilities and hydraulic heads occurring at both shallow and deep crustal levels (see for review Montgomery & Manga 2003). The nature, timing, and duration of observed hydrological responses to earthquakes all offer insights into the mechanics and depth of subsurface earthquake-induced changes. Geochemical observations, such as those from Alum Rock, offer an additional constraint for
understanding changes in fault zone permeability and connectivity in response to tectonic and seismic events.

Springs along the Hayward fault system have a long history of responding to distant earthquakes. Waring (1915) noted that the Warm Springs in Fremont, CA, located adjacent to the Hayward fault north of Alum Rock (Fig. 1) increased in both number and discharge immediately following the 1906 San Andreas earthquake. At Alum Rock, periodic monitoring of temperature and conductivity at springs 4 and 11, and flow at 11 from 1976 to 1991 by King et al. (1994) recorded short-term (days to months) flow increases and temperature decreases immediately following five local earthquakes (on other faults).

King et al. (1994) attributed the spring responses at Alum Rock to increased permeability, unclogged plumbing, and faster groundwater flow through the subsurface. The observed earthquake-induced flow changes at Alum Rock all show a clear distance–magnitude relationship with a threshold similar to that observed for liquefaction of unconsolidated materials (Montgomery & Manga 2003) (Fig. 6). The rapid response time (minutes to days), decrease in water temperature, transient nature and clear relationship between spring responses, and the distance and magnitude of the earthquakes suggest that the flow and temperature changes originated from shallow depths. If the observed flow changes were induced by aquifer changes at depth, it is likely that spring temperature would increase with faster transport rates from depth, the response time would be more delayed, and the response would be more persistent because of the structural changes in the aquifer fracture network (Montgomery & Manga 2003).

No stable isotope or major ion measurements were reported in conjunction with the King et al. (1994) study, and no earthquakes exceeding the threshold on Fig. 6 have occurred during the course of this study (2003 to early 2007). Shallow versus deep, transient stress-induced changes versus more permanent structural changes, however, should have predictable geochemical signatures. An influx of shallow waters to the spring system should cause a shift in isotopic composition to the left on Fig. 2A towards the GMWL and a decrease in the Cl content of the waters. An increase in permeability at depth in the system would cause the opposite shift in spring isotopic composition and Cl content. Structural changes in the fracture network feeding the springs, such as unlogging or rearranging the flow pathways, should alter the mixing of waters and result in a redistribution of individual spring chemistries. The persistence of such changes and transience in spring chemistries has the potential to provide some insights into rates of crack sealing (e.g. Lowell et al. 1993; Gratier et al. 2003). While this study was in review, a magnitude 5.6 earthquake occurred within 4 km of the Alum Rock Springs on October 30, 2007 and we are presently monitoring and sampling the springs. The results of ongoing monitoring of the spring responses to this event will provide the data necessary to test these hypotheses and ideally better constrain the origin of the waters and the controls on mixing of meteoric and connate waters within this fault zone.

**CONCLUSIONS**

At the Alum Rock mineral springs a range in $\delta^{18}$O and chloride concentrations of waters issuing from a number of discrete thermal springs located along a few hundred meter stretch of Penitencia Creek, just upgradient of a strand of the Hayward fault system, illustrate the effect of fracture/fault zone heterogeneities on fluid transport and geochemistry. The observed range in spring geochemistry may be explained by variable mixing of local meteoric waters with deep connate waters that have an $^{18}$O-shifted seawater composition. The observed range in meteoric–connate water mixing highlights the isolation of individual flow paths feeding the springs. Because the geochemistry observed at the spring outlets depends on the relative contribution of shallow and deep waters we propose that earthquake-induced changes in fault zone permeabilities and aquifer heads should result in observable changes in spring characteristics. From these changes it should be possible to distinguish both the relative

![Fig. 6. Relationship between spring response, earthquake magnitude and distance between the spring and earthquake. Filled and open symbols indicate earthquakes for which spring discharge changed or did not change, respectively. Circles are for the response of Alum Rock Springs monitored by King et al. (1994) between 1975 and 1991. Dashed line is the threshold line fitted to earthquakes resulting in Alum Rock spring response. Solid line is empirical limit of distance from epicenter for observed liquefaction; Montgomery & Manga (2003) found this distance–magnitude relationship to correspond well to limit of seismically induced changes in stream and spring flow.](image-url)
depth of aquifer response and help differentiate transient responses from more permanent changes in fault zone fracture networks.

ACKNOWLEDGMENTS

This work was supported through an IGPP grant to MM and TR. We thank the City of San Jose, California for granting access to the Alum Rock springs. We thank William Evans for valuable and insightful comments provided on an earlier draft of this manuscript. Additional insights and improvements were provided by two anonymous reviewers. A portion of this work was performed under the auspices of the U.S. Department of Energy by the University of California Lawrence Livermore National Laboratory under Contract W-7405-Eng-48.

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