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Persistent water level changes in a well near Parkfield, California, due to local and distant earthquakes

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Abstract. Coseismic water level rises in the 30-m deep Bourdieu Valley (BV) well near Parkfield, California, have occurred in response to three local and five distant earthquakes. Coseismic changes in static strain cannot explain these water level rises because (1) the well is insensitive to strain at tidal periods; (2) for the distant earthquakes, the expected coseismic static strain is extremely small; and (3) the water level response is of the incorrect sign for the local earthquakes. These water level changes must therefore be caused by seismic waves, but unlike seismic water level oscillations, they are monotonic, persist for days or weeks, and seem to be caused by waves with periods of several seconds rather than long-period surface waves. Other investigators have reported a similar phenomenon in Japan. Certain wells consistently exhibit this type of coseismic water level change, which is always in the same direction, regardless of the earthquake's azimuth or focal mechanism, and approximately proportional to the inverse square of hypocentral distance. To date, the coseismic water level rises in the BV well have never exceeded the seasonal water level maximum, although their sizes are relatively well correlated with earthquake magnitude and distance. The frequency independence of the well's response to barometric pressure in the frequency band 0.1 to 0.7 cpd implies that the aquifer is fairly well confined. High aquifer compressibility, probably due to a gas phase in the pore space, is the most likely reason why the well does not respond to Earth tides. The phase and amplitude relationships between the seasonal water level and precipitation cycles constrain the horizontal hydraulic diffusivity to within a factor of 4.5, bounding hypothetical earthquake-induced changes in aquifer hydraulic properties. Moreover, changes of hydraulic conductivity and/or diffusivity throughout the aquifer would not be expected to change the water level in the same direction at every time of the year. The first 2.5 days of a typical coseismic water level rise could be caused by a small coseismic discharge decrease at a point several tens of meters from the well. Alternatively, the entire coseismic water level signal could represent diffusion of an abrupt coseismic pore pressure increase within several meters of the well, produced by a mechanism akin to that of liquefaction. The coseismic water level changes in the BV well resemble, and may share a mechanism with, coseismic water level, stream discharge, and groundwater temperature changes at other locations where preearthquake changes have also been reported. No preearthquake changes have been observed at the BV well site, however.

1. Introduction

Changes in groundwater levels and surface water discharge following earthquakes have been reported for decades [*Waller et al.*, 1965; *Whitehead*, 1985; *Whitehead et al.*, 1984]. There are several reasons to investigate these groundwater level changes. For example, groundwater level changes following earthquakes can affect water supplies. It would be helpful to be able to foresee these effects, and it is sometimes necessary to evaluate the causative role of an earthquake in insurance claims for loss of water supply (S. Wolf, *U.S. Geological Survey*, personal communication, 1994). Underground waste repositories are at risk from seismically induced water table rises. *Carrigan et al.* [1991] have calculated that water table rises resulting from poroelastic aquifer compression should be, at most, 2 to 3 m, but many observed water level changes exceed those predicted by poroelastic models and represent other mechanisms that should be evaluated. Seismic waves can affect oil well production, and it has been suggested that in some cases the application of artificial seismic waves can stimulate production [*Beresnev and Johnson*, 1994]. Earthquake-induced fluid pressure changes are hypothesized to control the timing and/or location of aftershocks and triggered seismicity [*Hill et al.*,

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1995; *Gomberg*, 1996]. Finally, these groundwater level changes could be related to the mechanisms of reported hydrologic earthquake precursors.

Several types of coseismic groundwater level changes are currently recognized. Step-like groundwater level changes in confined aquifers within a few source dimensions of an earthquake epicenter can often be quantitatively accounted for as the poroelastic response to the earthquake's static strain field [*Quilty and Roeloffs*, 1997; *Wakita*, 1975]. Seismic oscillations, due primarily to surface waves from distant events, occur in some wells tapping highly transmissive aquifers [*Leggette and Taylor*, 1935; *Liu et al.*, 1989; *Woodcock and Roeloffs*, 1996]. Strong ground shaking near the earthquake rupture may increase permeability and, consequently, change groundwater levels [*Rojstaczer et al.*, 1995] .

Recent observations confirm that there is an additional class of water level responses to large earthquakes as distant as several hundred kilometers. These responses are due to seismic waves but, unlike seismic water level oscillations, are monotonic and persist for days or weeks. The M_W 7.3 1992 Landers, California, earthquake produced water level changes of this type at several locations [*Roeloffs et al.*, 1995]. High-quality data sets document coseismic water level drops in two wells in Japan in response to earthquakes up to 740 km away, with some of the larger groundwater level changes preceded by smaller preearthquake changes [*Matsumoto*, 1992; *Matsumoto and Takahashi*, 1994; *Takahashi and Matsumoto*, 1994]. Seismic water level oscillations in a fractured-rock well in Oregon have been followed by persistent water level drops for sufficiently large or close earthquakes [*Woodcock and Roeloffs*, 1996]. Where available, long data records show that certain wells, in both alluvial and fractured aquifers, consistently respond this way and that for each such well, the water level always rises or always falls, regardless of the earthquake's focal mechanism or location.

Coseismic changes of groundwater temperature and discharge in Japan exhibit features similar to these coseismic water level changes and have sometimes been preceded by signals interpreted as earthquake precursors. *Mogi et al.* [1989] report coseismic rises in the temperature of water from an artesian thermal well at the Usami Hot Spring on the Izu Peninsula, Japan; possible precursory signals preceded some of the larger temperature increases. *Fujimori et al.* [1995] describe coseismic seep discharge increases at the Rokko-Takao site near Kobe, Japan, in response to earthquakes up to several hundred kilometers away. At this site, discharge gradually increased over the 2 months preceding the January 17, 1995, Hyogo-ken-Nanbu earthquake, which ruptured to within 20 km of the discharge point and produced a coseismic discharge increase of more than 1 order of magnitude. It is plausible that groundwater level, discharge, and temperature changes have a common underlying mechanism since discharge rate depends on subsurface fluid pressure and temperature changes can be caused by changes in aquifer mixing attributable to aquifer fluid pressure changes.

This paper describes the earthquake-induced water level rises in an alluvial well at Bourdieu Valley, near Parkfield, California, and considers possible explanations for them. A relation describing the response amplitude as a function of earthquake magnitude and distance is developed. The aquifer's hydraulic and poroelastic properties are inferred from its response to seasonal rainfall and barometric pressure and its lack of a response to Earth tides. Two classes of mechanisms are considered as follows: first, that earthquakes directly affect fluid pressure near the well site and, alternatively, that earthquakes change aquifer properties, with the observed water level changes representing equilibration to the modified properties. Viable explanations must account for the persistence of the response after seismic wave motion has decayed, for the relatively large responses to small or distant earthquakes, and for the observation that the water level always rises, regardless of the earthquake's location or focal mechanism. The characteristic time history of the water level rises is compared with expected responses to changes in hydraulic properties, local discharge rate, or a shaking-induced pressure increase similar to that associated with liquefaction.

2. Well and Data Collection Methods

The Bourdieu Valley (BV) well (Figures 1a and 1b) is 30 m deep. It is situated at an elevation of 851 m in a small catchment containing the headwaters of the Pancho Rico Creek (Figures 2a and 2b). The well was drilled in late 1986 to provide hydrologic information to supplement water level records from a 305-m-deep borehole drilled about 5 m away. Drillers' logs from the shallow well state only that the well is drilled in alluvium. The deeper borehole, drilled to assess the site's suitability for a borehole strain meter, was completely cased and later perforated to allow communication with the formation at depths of 280 to 282 m. The site is 1.8 km northeast of the San Andreas fault zone, and both wells have been monitored since 1987 as part of the Parkfield Earthquake Prediction Experiment [*Bakun and Lindh*, 1985; *Roeloffs and Langbein*, 1994].

The water level record from the deeper borehole did not respond to Earth tides, barometric pressure, or rainfall, but it was found to react to moving the transducer or inserting a measuring tape, suggesting poor, if any, communication between the well bore and the formation. No earthquake-related water level changes have been observed in the deep borehole. The hydrograph from this borehole is not considered representative of aquifer fluid pressure and therefore is not discussed further in this paper.

Water level in the BV well has been measured using submersible differential pressure transducers, which produce output voltages proportional to the difference between water pressure and atmospheric pressure. The transducer models used in the well, together with their respective full-scale ranges, are listed in Table 1. Resolution of the digital water level data is 0.03 cm of water for transducers with a full-scale range of 7 m and 0.05 cm of water for the transducer with a full-scale range of 10 m. A vent tube in the transducer cable transmits atmospheric pressure to the submerged sensing element; blockage of this vent tube impaired recording of the well's response to barometric pressure from April 1993 through November 1995.

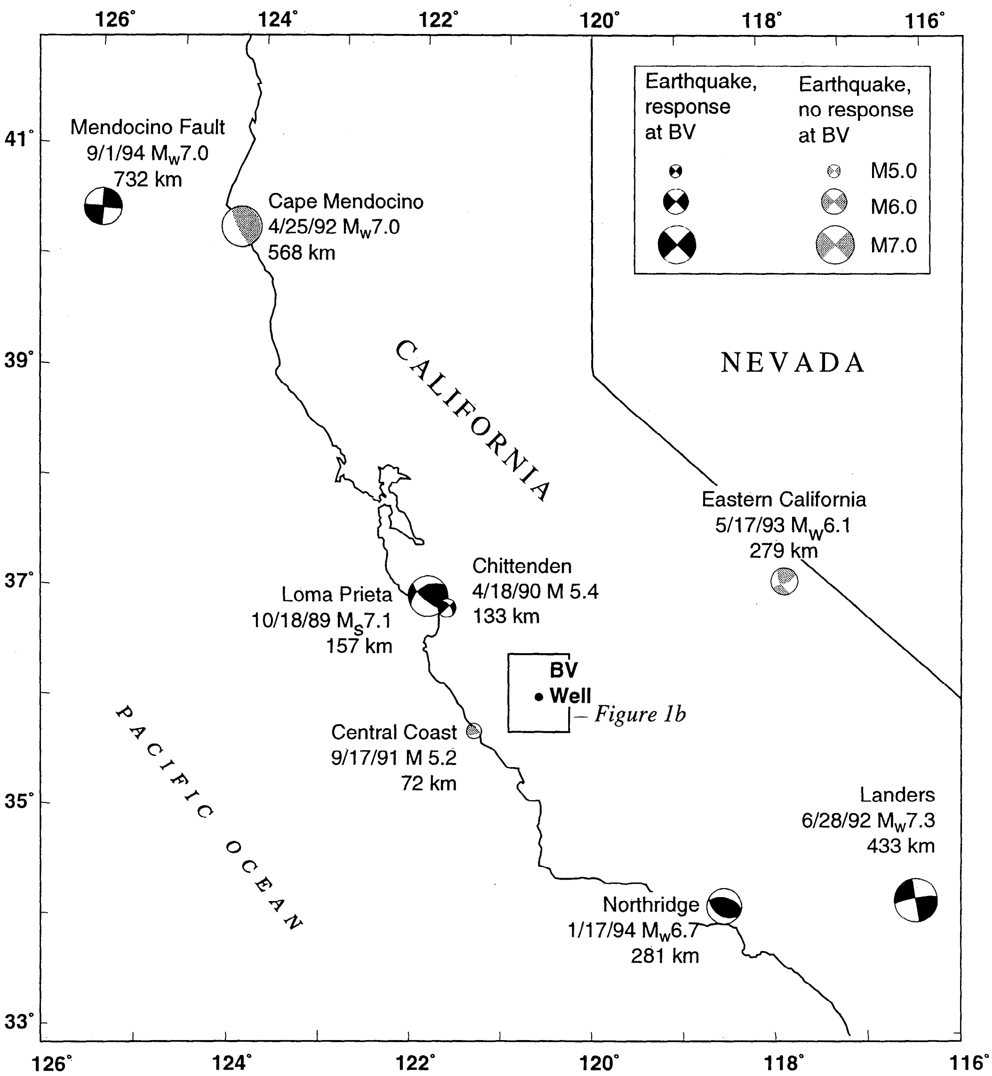

Figure 1a. Map of California showing locations of the Bourdieu Valley (BV) (120.6042°W, 36.0845°N) well, five earthquakes outside the Parkfield area that produced coseismic water level rises, and three earthquakes that might have been expected to cause water level changes but did not.

Water level is sampled once every 15 min and telemetered to the U.S. Geological Survey office in Menlo Park, California. Manual measurements of water level, accurate to the nearest 0.3 cm, are made two to four times per year to check against transducer drift. Barometric pressure and rainfall are also measured at the site every 15 min, and rainfall readings are compared with those at sites throughout the Parkfield area to identify any periods of gage malfunction. Several time periods exist for which no water level data are available owing to equipment failures; these data gaps are listed in Table 2.

Figure 3 shows a typical month of water level, barometric pressure, and rainfall data from the BV well. The raw water level data correlate inversely with barometric pressure. After correcting for the barometric pressure response, one can discern small water level rises following rainfall. Frequent abrupt drops in water level are attributed to gas bubbling, which is audible in the well. The typical size of these water level drops (2 cm) and the diameter of the well casing (6.4 cm) imply that about 64 cm^3 of gas escapes in each bubble. The rate of bubbling varies but averages one to two bubbles this size each day. The gas composition has not been determined.

3. Earthquake-Related Water Level Changes

Figure 4 shows hourly averages of water level in the BV well for the period January 1987 through September 1995. Water level fluctuated between 14 and 19 m below land surface during this period, with a seasonal variation of 1 to 3 m in amplitude. Superimposed on the seasonal variations and labeled in Figure 4 are six easily visible water level rises caused by earthquakes, whose parameters are listed in Table 3. The five largest changes are plotted with an expanded timescale in Figure 5a, and the smallest

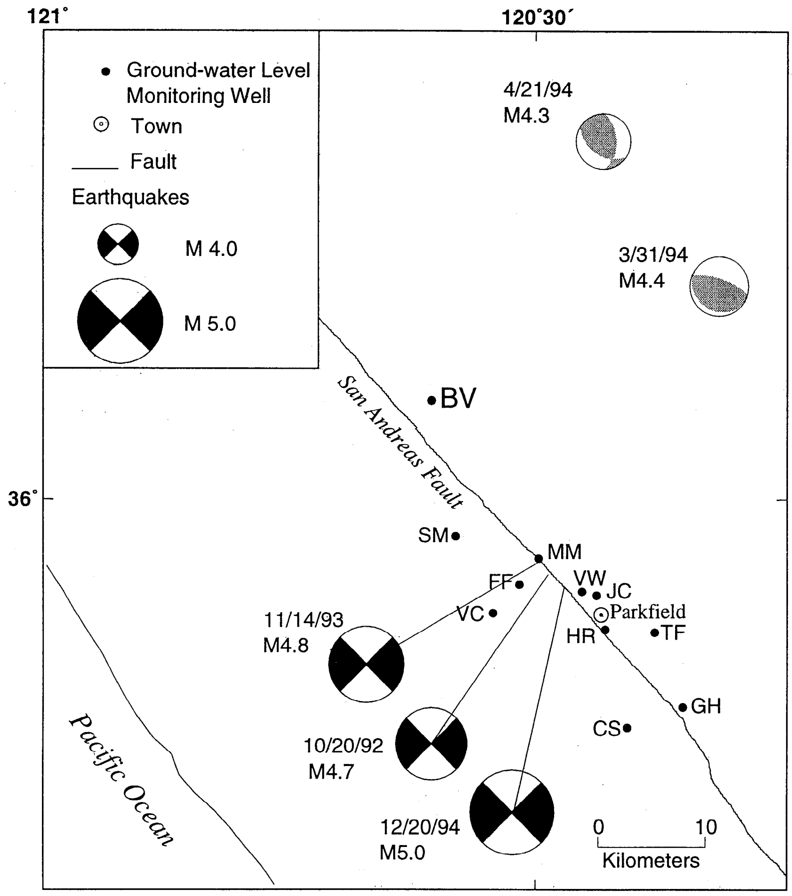

Figure 1b. Map of the Parkfield area showing locations of the BV well, other water wells monitored, three local earthquakes that produced water level changes, and two earthquakes that did not.

of the six is plotted in Figure 5b, together with water level rises in response to the Chittenden earthquake on April 18, 1990, and the Mendocino Fault earthquake on September 1, 1994, which are too small to be visible in Figure 4. The 5-cm rise associated with the Chittenden earthquake is judged to be the smallest earthquake-related change that could be distinguished from background water level fluctuations.

The first earthquake-related water level change occurred in response to the 1989 Loma Prieta earthquake, 157 km northwest of the well site. All of the earthquake-produced responses are rises in water level. Figure 4 shows that the hydrograph generally resumes its seasonal character 1 to 3 months after each earthquake, implying that the earthquake-induced disturbance decays to zero over this time interval.

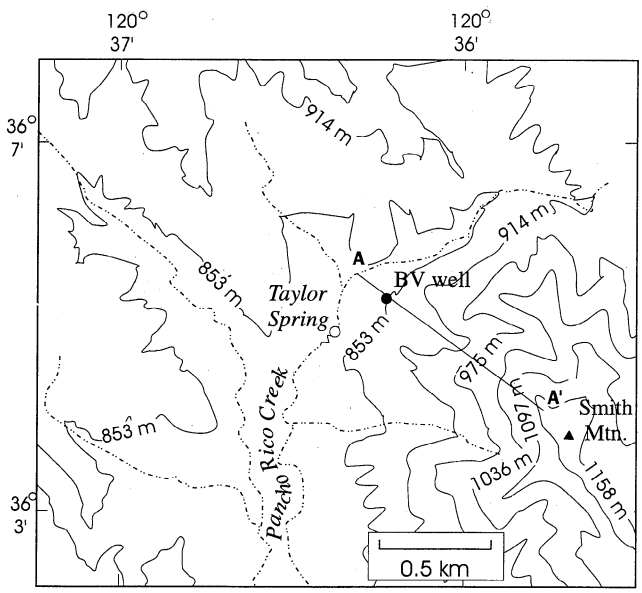

Figure 2a. Map showing BV well, Pancho Rico Creek, and local topography.

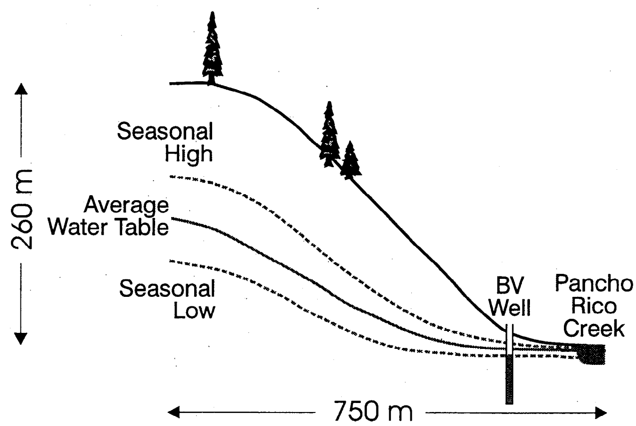

Figure 2b. Generalized cross section along A-A' in Figure 2a showing hydrologic setting of BV well.

Table 1. Transducer Types and Dates of Operation at Bourdieu Valley Well

Transducer Type	Full-Scale Range, m	Date Installed	Date Failed
Envirolab	7	Dec. 1986	July 9, 1987
Druck PDCR 830	7	1987	July 10, 1992
Druck PDCR 830	10.5	July 11, 1992	March 11, 1993
Druck PDCR 930 blocked vent tube	7	April 28, 1993	Dec. 26, 1994
Druck PDCR 830IP	7	Dec. 9, 1994	still operating

Ranges are given in meters of water.

To date, none of the coseismic water level rises appears to have exceeded the typical seasonal high water level.

Four additional water level rises were identified for which no associated earthquake could be found (Figure 5c). None of the other geophysical instruments near Parkfield recorded anomalous signals at these times.

For the five largest water level rises (those shown in Figure 5a), Figure 6 shows a graph superimposing the responses after subtracting a linear trend fit to the 10 days of data preceding the earthquake and then dividing by the maximum values. All five responses have similar time histories for the first 2 days following each earthquake. Differences among the decay rates as shown in Figure 6 may actually result from inaccuracy in removing the seasonal trend. In particular, the responses to the M_W 7.3 Landers earthquake and the M 4.8 Parkfield earthquake are essentially identical.

All of the water level changes shown in Figure 5 began in the 15-min sampling interval following each earthquake. In contrast to coseismic phenomena reported in Japan, no preearthquake changes are apparent. *Mogi et al.* [1989] noted that coseismic groundwater temperature rises at the Usami Hot Spring were preceded by unusual temperature variations prior to three earthquakes and two earthquake swarms, all within 31 km of the spring. The coseismic water level drops in the Haibara, Japan, well, corresponding to the five largest earthquakes, were preceded by preearthquake changes [*Matsumoto and Takahashi*, 1994; *Takahashi and Matsumoto*, 1994].

Among the 12 wells monitored near Parkfield, the BV well's responses to distant earthquakes are distinctive because of their large amplitudes and because the well is insensitive to strain at tidal periods. Other wells at Parkfield have shown water level changes associated with some of these distant earthquakes, most of which are small and can be observed only after removal of Earth tide and barometric pressure fluctuations. Two wells located in and close to the San Andreas fault zone have shown relatively large coseismic responses, however. In the Middle Mountain well (MM in Figure 1b), coseismic water level drops were observed in response to all five of the earthquakes that produced the largest responses at BV (those shown in Figure 5a). The MM well has also displayed small water level rises, as well as large water level drops, in response to creep events on the San Andreas fault, which have been interpreted as poroelastic responses to volumetric strain [*Roeloffs et al.*, 1989]. *Quilty and Roeloffs* [1997], however, have shown that the December 20, 1994, magnitude 5 Parkfield earthquake imposed contractional strain at the MM site, which should have resulted in a water level rise, rather than the observed water level drop. The Haliburton Ranch (HR) well also displayed a water level change opposite in sign from that expected in response to this earthquake, but it has not been monitored long enough to observe its responses to other events. These examples of coseismic water level changes of discrepant sign suggest that tectonic strain may affect the MM and HR wells, not only poroelastically, but also by a mechanism similar to that of the BV well. On the other hand, the proximity of both these wells to the fault zone may influence the nature of their coseismic responses.

Table 2. Time Periods for which Water Level Data in the BV Well are Unavailable

Beginning of Gap	End of Gap
Feb. 10, 1987	Feb. 20, 1987
Nov. 1, 1989	Nov. 15, 1989
Sept. 17, 1990	Oct. 11, 1990
Oct. 30, 1990	Nov. 6, 1990
March 18, 1991	July 4, 1991

4. Dependence on Earthquake Magnitude and Distance

The sizes of the water level rises in the BV well can be related to the magnitude of each earthquake and its distance from the well. A simple dependence upon magnitude and distance might not be expected, if earthquakes modify aquifer properties, because the size and speed of fluid pressure adjustment could depend on the time during the seasonal cycle when the earthquake occurs. For the Haibara well in Japan, however, it has been determined that water level drops occur in response to earthquakes of magnitude M if

$$M \geq 0.69 + 2.45 \log_{10} D \tag{1a}$$

where D is the well-hypocenter distance in kilometers [*Matsumoto and Takahashi*, 1994; *Takahashi and Matsumoto*, 1994]. In addition, *Mogi et al.* [1989] found that groundwater temperature increases at Usami Hot Spring were produced by earthquakes if

$$M \geq 1.3 + 2.2 \log_{10} D. \tag{1b}$$

To determine whether a similar relationship holds for the BV well, the sizes in centimeters, Δh_i, of the eight

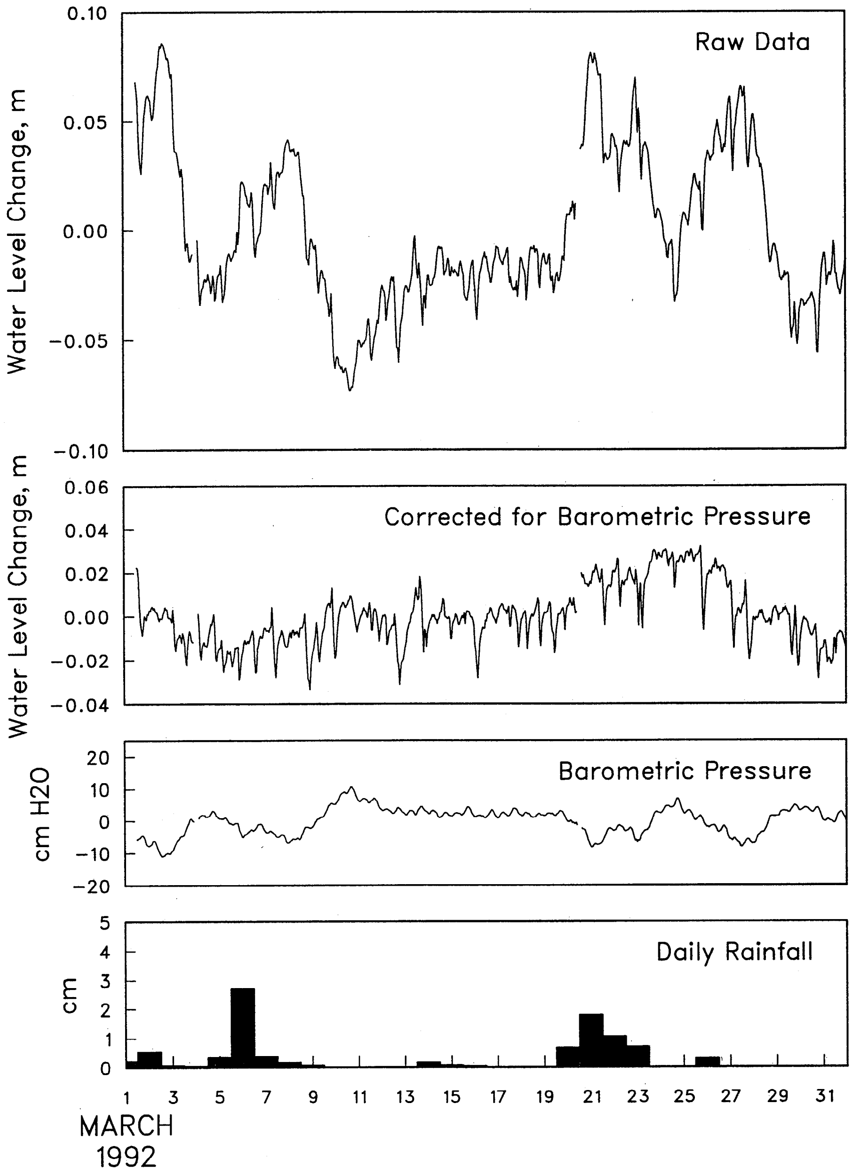

Figure 3. Water level in the BV well for March 1992, a period when no earthquake-related responses were observed. Barometric pressure and rainfall recorded at the site are also shown.

unambiguous coseismic water level rises (those shown in Figures 5a and 5b) were fit to an equation of the form

$$\log_{10}\Delta h_i = w_1 M + w_2 \log_{10} D + w_3 \qquad (2)$$

where w_1, w_2, and w_3 are constants and M is the magnitude listed in Table 3 (M_W where available). The resulting relationship estimates Δh_i as

$$\log_{10}\Delta\hat{h}_i = 0.89\, M - 1.63\, \log_{10} D - 0.69 \qquad (3)$$

The water level rises predicted by (3) are plotted versus those observed in Figure 7a. For the earthquakes that produced coseismic water level rises, (3) does a good job of predicting the size of the response. Some insight into the physical meaning of (3) can be gained by expressing earthquake magnitude as

$$M = \log_{10} A + C_1 \log_{10} D + C_2 \qquad (4)$$

where A is the seismic trace amplitude (proportional to ground velocity) at a specified frequency and the C_i are constants, with $C_1 = 1.66$ for the surface wave magnitude M_s [*Bullen and Bolt, 1985*]. Although the coefficients in (3) were determined using M_W and M_D rather than M_s, the differences between these magnitude scales should be, at most, a few tenths of a magnitude unit for the events listed in Table 3, so that substituting (4) into (3) should not lead to large errors. The resulting relation gives $\Delta\hat{h}_i$ proportional to $A^{0.89} D^{-0.15}$, implying that the size of the water level change is approximately proportional to the peak ground velocity at the well site and is not strongly dependent upon the distance to the earthquake. The ground velocities are, in turn, proportional to dynamic strain, but because the coefficients of proportionality depend upon the specific stress component as well as the type of seismic wave, no attempt is made here to derive a quantitative relationship of water level change to peak strain.

Setting $\Delta\hat{h}_i = 5$ cm, the smallest change that can be

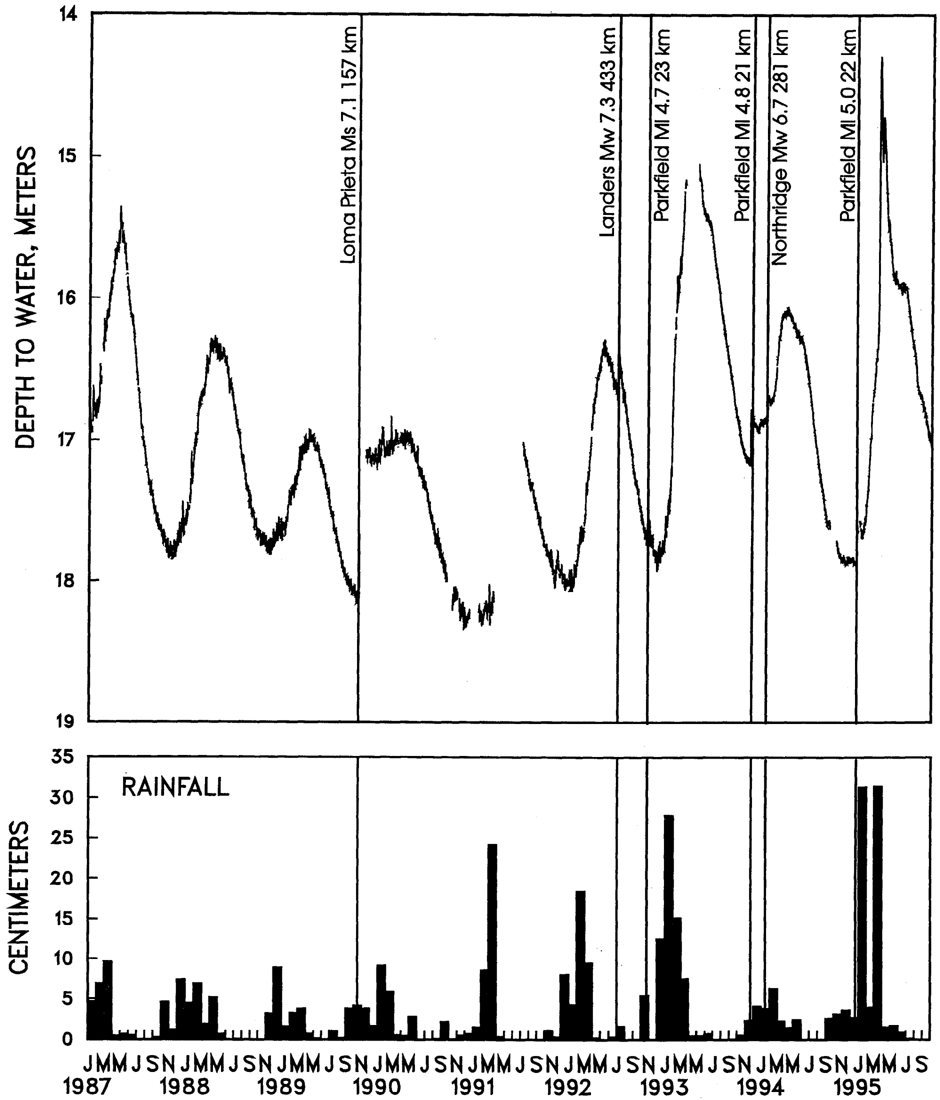

Figure 4. Hourly water level data and monthly rainfall data from the BV site. The times of six earthquakes producing water level changes are indicated.

discriminated from background noise, yields the threshold criterion

$$M \geq 1.55 + 1.82 \log_{10} D. \quad (5)$$

The equations describing the response thresholds for the Usami Hot Spring, Haibara well, and BV well all have log distance terms with coefficients near 2.

In Figure 7b, the criterion given by (5) is plotted together with the magnitudes and distances of earthquakes that did and did not produce water level rises in the BV well. Figure 7b shows that water level rises were observed for all earthquakes for which (3) predicts a change of 9 cm or more, but it also reveals five additional earthquakes for which water level changes between 4 and 9 cm would have been expected (Table 3 and Figures 1a and 1b). Four of these five earthquakes occurred when water level in the BV well was at or near its seasonal maximum, supporting the previous conjecture that this level limits the coseismic water level rises. On the other hand, the September 17, 1991, earthquake on the central California coast, for which the largest response was expected but none was observed, took place close to seasonal minimum water level. An alternative view is that all earthquakes that produced coseismic water level rises occurred in a NW-SE striking band along the plate boundary. The 1992 Cape Mendocino event is the only one fitting this description that might have been expected to cause a response but did not. This azimuthal dependence, if significant, could indicate that the water level response is sensitive to the orientation of the dynamic strains and therefore to the seismic wave radiation pattern. Of course, it is also possible that the magnitudes of the events that did not produce expected responses are overestimated by 0.1 to 0.2 units, so that, in fact, no response should have been observed.

The existence of responses to earthquakes as small as M 4.7 suggests that long-period waves are not required to produce a response. Further evidence for this idea comes from calculations of dynamic stress and strain at Parkfield [*Spudich et al.*, 1995], showing that dynamic stresses there caused by the Landers earthquake exceeded those produced

Table 3. Earthquakes and Water Level Responses in the BV Well

Location	Date and Time, UT	Magnitude	Latitude, Longitude	Depth, km	Distance From BV Well, km	Expected Water Level Change (cm)	Observed Water Level Change, cm
Loma Prieta	Oct. 18, 1989, 0004	M_S 7.1, M_W 6.9	37.04°N, 121.88°W	18.5	157	74.4	85
Chittenden	April 18, 1990, 1354	M_D 5.4	36.93°N, 121.66°W	5.8	133	4.7	5.5
Central Coast	Sept. 17, 1991, 2110	M_D 5.2	35.82°N, 121.33°W	6.0	72	8.6	0
Cape Mendocino	April 25, 1992, 1806	M_S 7.1, M_W 7.0	40.33°N, 124.23°W	10.2	568	4.3	0
Landers	June 28, 1992, 1157	M_W 7.3	34.20°N, 116.44°W	3.2	433	34.5	34.0
Parkfield	Oct. 20, 1992, 0528	M_D 4.7	35.93°N, 120.47°W	10.0	23.3	18.4	14.0
Eastern California	May 17, 1993, 2321	M_D 6.1	37.17°N, 117.79°W	7	279	6.0	0
Parkfield	Nov. 14, 1993, 1225	M_D 4.8	35.95°N, 120.49°W	11.5	20.9	26.9	36.0
Northridge	Jan. 17, 1994, 1230	M_W 6.7	34.21°N, 118.54°W	18	281	19.0	19.0
NE of Parkfield	March 31,1994, 2000	M_D 4.4	36.18°N, 120.31°W	11.5	30.8	6.6	0
NE of Parkfield	April 21, 1994, 1637	M_D 4.3	36.30°N, 120.43°W	9.6	30.2	5.5	0
Mendocino Fault	Sept. 1, 1994, 1515	M_W 7.0	40.44°N, 126.89°W	7.0	732	5.3	9.1
Parkfield	Dec. 20, 1994, 1027	M_D 5.0	35.92°N, 120.47°W	8.9	24.0	32.4	33.0

Events are listed if they either produced a water level response, or would have been expected to produce a response greater than 5 cm, based on (4). Moment magnitudes are listed for all of the earthquakes for which these are available; for the other earthquakes, M_D from the U.S. Geological Survey Preliminary Determination of Epicenters or Calnet catalog is listed.

by the Loma Prieta earthquake by factors of 2 to 5 for periods longer than 10 s. Only in the period band of 3-7 s were the dynamic stresses associated with the Loma Prieta earthquake consistently larger than those associated with Landers. Since the water level response to the Loma Prieta earthquake was more than twice as large as the response to the Landers earthquake, the dynamic stress calculation suggests that this period band may preferentially cause the water level changes.

5. Inferences of Aquifer Properties From Hydrographs and Implications for Strain Sensitivity

The water level records from the BV well can be used to infer aquifer hydraulic and mechanical properties that, in principle, govern the well's sensitivity to strain. The techniques used are summarized by *Roeloffs* [1996]. The well responses to barometric pressure and tidal strain yield constraints on vertical hydraulic diffusivity c_v and poroelastic properties, while the response to seasonal precipitation constrains the horizontal diffusivity c_h and, to a lesser degree, the horizontal hydraulic conductivity. These properties are estimated herein and used in later sections to evaluate the well response to hypothetical mechanisms for the coseismic water level changes.

5.1 Barometric Response

The BV well does not respond to Earth tides but does respond to barometric pressure (Figure 3). In general, the barometric response of water level in an open well is the sum of two contributions: the poroelastic fluid pressure change in the aquifer and the direct action of atmospheric pressure on the water surface in the well. The poroelastic fluid pressure change in the aquifer is diminished if the aquifer is poorly confined. More specifically, where the diffusivity with respect to air of the material above the water table is high enough to negligibly impede the communication of atmospheric pressure, the ratio of water level change to barometric pressure change at radian frequency ω is expected to follow the relationship

$$E_B(z,\omega) = -(1-\alpha)\ \left\{1-\exp[-(i\omega/c_v)^{1/2}(z-z_w)]\right\} \qquad (6)$$

[*Quilty and Roeloffs*, 1991], where $i=(-1)^{1/2}$. In (6), z is the depth at which the well is open to the aquifer, z_w is the depth to the water table, and α is given by

$$\alpha = B(1+\upsilon_u)/3(1-\upsilon_u) \qquad (7)$$

where B is Skempton's coefficient and υ_u is the undrained Poisson ratio. E_B is dimensionless, provided barometric pressure is expressed in units of equivalent height of water.

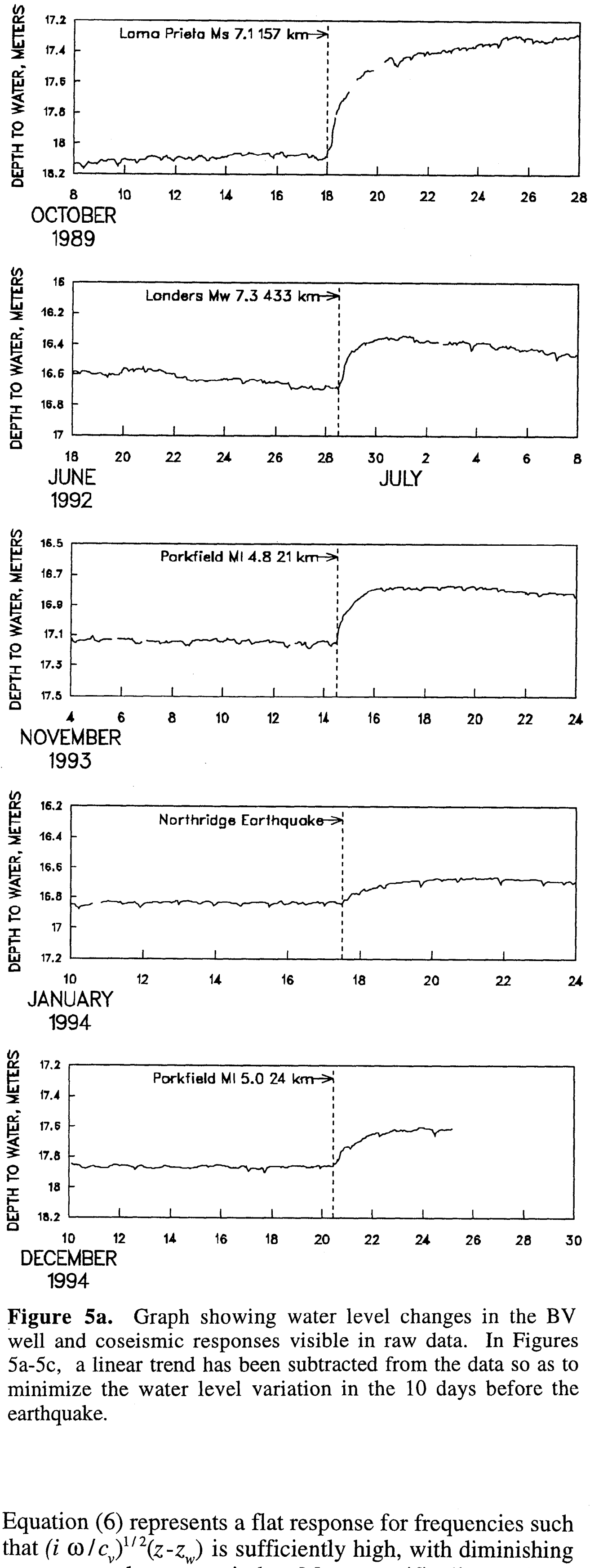

Figure 5a. Graph showing water level changes in the BV well and coseismic responses visible in raw data. In Figures 5a-5c, a linear trend has been subtracted from the data so as to minimize the water level variation in the 10 days before the earthquake.

Equation (6) represents a flat response for frequencies such that $(i\,\omega / c_v)^{1/2}(z-z_w)$ is sufficiently high, with diminishing response at longer periods. More specifically, at depth $z-z_w$ below the water table, volumetric strain variations of radian frequency ω produce pore pressure variations less than $1/e$ times the undrained response if

$$c_v > 5\omega(z-z_w)^2 \tag{8}$$

[*Roeloffs*, 1996]. This degradation of response applies not only to barometric pressure, but also to water level variations that might be induced by tectonic strain. Consequently, a well's barometric response as a function of frequency can yield information about the vertical diffusivity of the material overlying the aquifer as well as the aquifer's poroelastic properties [*Rojstaczer*, 1988; *Beavan et al.*, 1991; *Quilty and Roeloffs*, 1991], which can, in turn, illuminate the well's response to strain.

The barometric response of the BV well was estimated using the methods outlined by *Quilty and Roeloffs* [1991], and its amplitude is shown in Figure 8 for the frequency range 0.13 to 0.72 cpd, where the available data provide transfer function estimates with coherence >0.85. In this frequency range, the transfer function relating barometric pressure to water level is essentially flat, with amplitudes of 0.86 to 0.91 and phases of 180° ±10°. Transfer function estimates at frequencies lower than 0.13 cpd have low coherences, but it can be concluded that frequency

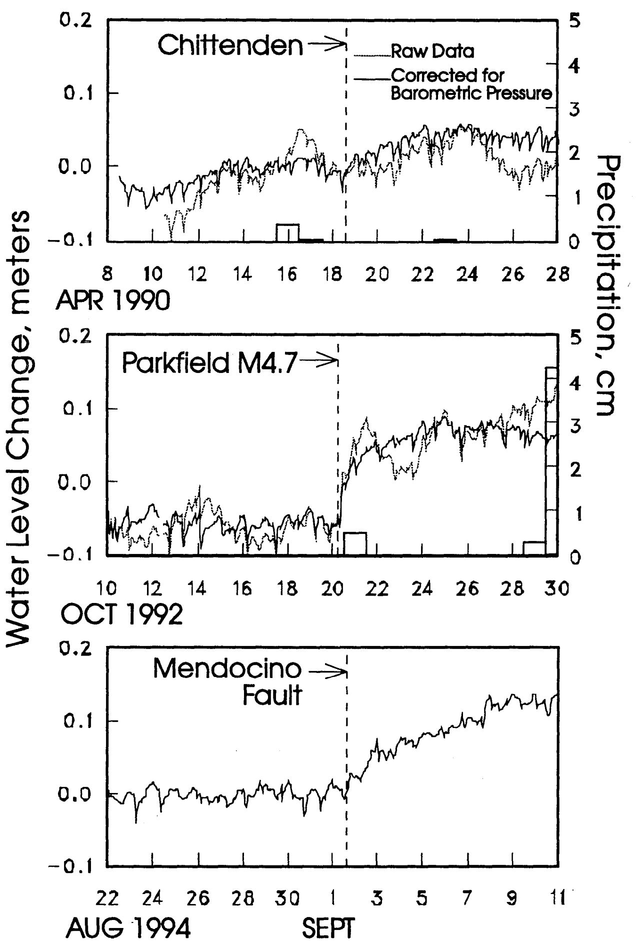

Figure 5b. Same as Figure 5a, but showing smaller coseismic responses that are not discernible in the presence of barometric pressure fluctuations. Data are shown with and without barometric pressure removed for the Chittenden and October 20, 1992, Parkfield, California, earthquakes. Raw data only are shown for the Mendocino Fault earthquake, which occurred while a blocked breather tube suppressed the well's barometric response.

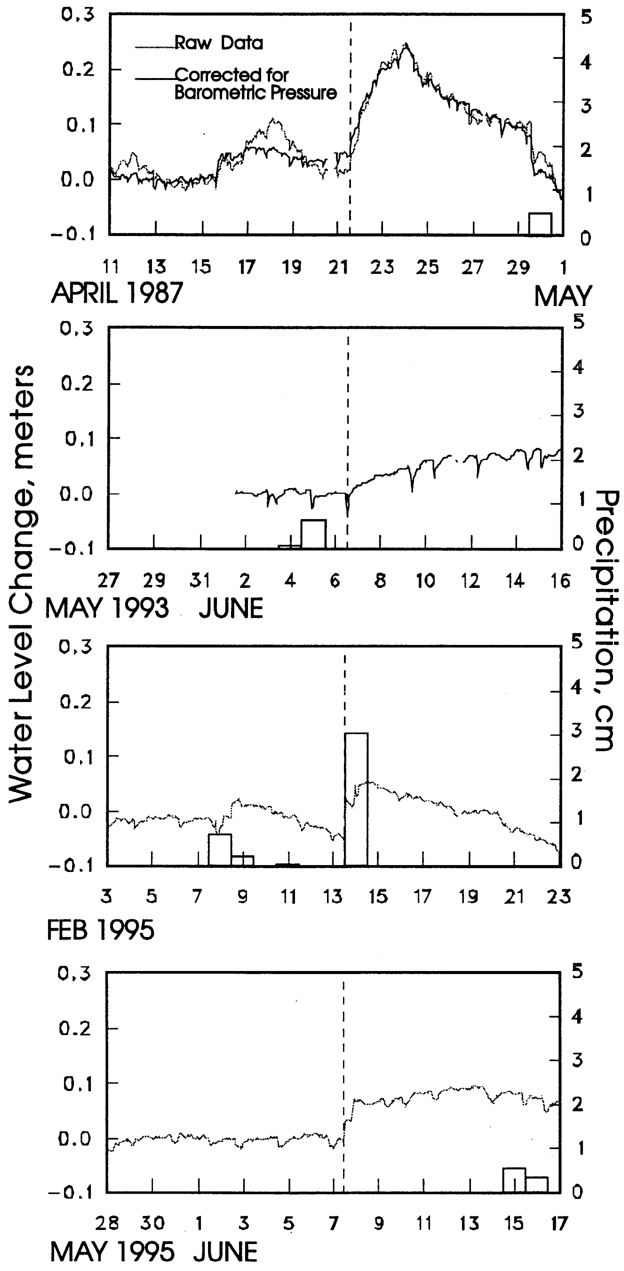

Figure 5c. Water level changes in the BV well for which no causative earthquakes could be identified.

dependence in the barometric response caused by flow to the water table is limited to periods longer than 0.13 cpd. Figure 8 shows that (6) matches the observed transfer function, with values of α between 0.13 and 0.19 and for $c_v \leq 3 \times 10^{-4}$ m²/s. The value of c may be lower, but it cannot be higher without introducing unobserved frequency dependence into the response. Including the effect of low pneumatic diffusivity in (6) would provide an even lower upper bound on c_v.

5.2 Implications of the Lack of a Response to Earth Tides

Earth tides in the Parkfield area produce volumetric strain fluctuations with a peak-to-peak amplitude of 0.06 x 10^{-6}, but no tidal water level variations occur in the BV well. The flat barometric response in the frequency range shown in Figure 8 implies that vertical flow does not degrade strain signals at tidal periods. The well's barometric response also shows no signs of frequency dependence at periods approaching 1 cpd, contraindicating poor communication between the aquifer and the well bore as the reason for the lack of tidal response. Consequently, the reason why no tidal fluctuations are observed must be that the aquifer has a high compressibility and/or low Skempton's coefficient. Under undrained conditions, the ratio of pore pressure change to volumetric strain in an isotropic, homogeneous, poroelastic aquifer is given by

$$E_T = \frac{2GB}{3}\frac{1+\upsilon_u}{1-2\upsilon_u} = B/\beta_u \tag{9}$$

where G is the shear modulus and $\beta_u = 3(1-2\upsilon_u)/\,2G(1+\upsilon_u)$ is the undrained compressibility [*Roeloffs*, 1996]. To be unobservable, the tidal fluctuations in the BV well must be smaller than 2 cm peak to peak, implying that E_T is less than 3.3×10^5 m of water or 3.3 GPa.

Inferences can be drawn about the aquifer's physical properties from the upper bound on E_T, together with the estimates of α obtained from the barometric response. For any value of the drained Poisson ratio υ, the following equations applied in sequence yield Skempton's coefficient, the undrained Poisson ratio, the shear modulus, and the drained compressibility:

$$B = \frac{3(1-\upsilon)\alpha}{(1+\upsilon)+2(1-2\upsilon)\alpha} \tag{10a}$$

$$\upsilon_u = (3\alpha - B)/(3\alpha + B) \tag{10b}$$

$$G = \frac{3(1-2\upsilon_u)\,E_T}{2B(1+\upsilon_u)} \tag{10c}$$

$$\beta = \frac{3(1-2\upsilon)}{2G(1+\upsilon)} \tag{10d}$$

Using the undrained compressibility computed from (10d), together with values of $\beta_f = 0.435$ GPa^{-1} for the compressibility of water and $\beta_s = 0.0303$ GPa^{-1} for the

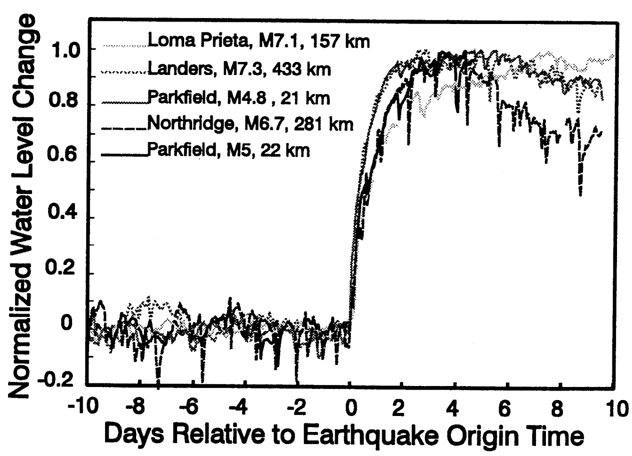

Figure 6. Graph of the five largest coseismic water level changes, superimposed after dividing by their peak values.

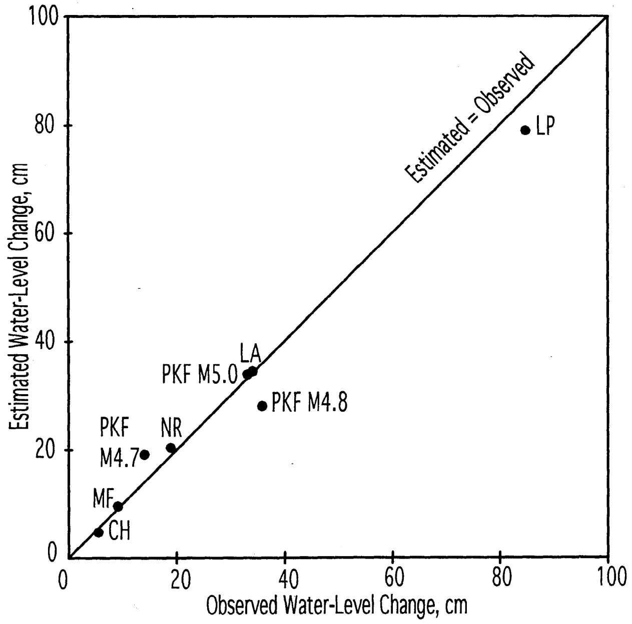

Figure 7a. Graph showing fit of observed water level changes to (4).

compressibility of quartz, then allows the aquifer porosity ϕ to be estimated from

$$\phi = \frac{(1-B)}{B}\frac{\beta - \beta_s}{\beta_f - \beta_s}. \tag{11}$$

Knowledge of E_T and α does not suffice to determine all these properties uniquely. An additional constraint, however, is that the porosity must be reasonable, say, $\phi <$ 0.4. Interestingly, there are no values of υ that satisfy this constraint for $\alpha = 0.19$. For $\alpha = 0.13$, porosities in the range $0.25 < \phi < 0.4$ are obtained only if $0.35 < \upsilon < 0.5$ (the upper limit). Porosities lower than 0.25 are consistent with the observed tidal and barometric response only if either the grains are more compressible than quartz or the pore fluid is more compressible than water. The latter possibility is favored by the observation of gas bubbling in the well, which must derive from a gas phase in the aquifer pore space.

5.3 Inference of Hydraulic Properties From Seasonal Variations

The seasonal water level variations in the BV well are presumably driven by seasonal precipitation variations. In this section, it is shown that there is an approximately linear and constant relationship between the amount of total annual precipitation and the seasonal water level variation, presumably governed by the hydrologic parameters of the drainage basin. The constancy of the relationship constrains the sizes of changes in these parameters that could be caused by earthquakes.

It is important to identify the hydrologic path by which rainfall affects piezometric level in the BV well. Precipitation falling on the drainage basin can evaporate, be used by vegetation, run overland to the Pancho Rico creek, flow downslope through shallow soils during and immediately following storms, and/or enter the groundwater system to flow downslope and, when head is high enough, provide base flow to the creek. Water that is evaporated or transpired is not available to affect the BV well or the creek. Discharge measurements from the Pancho Rico Creek would permit estimation of this part of

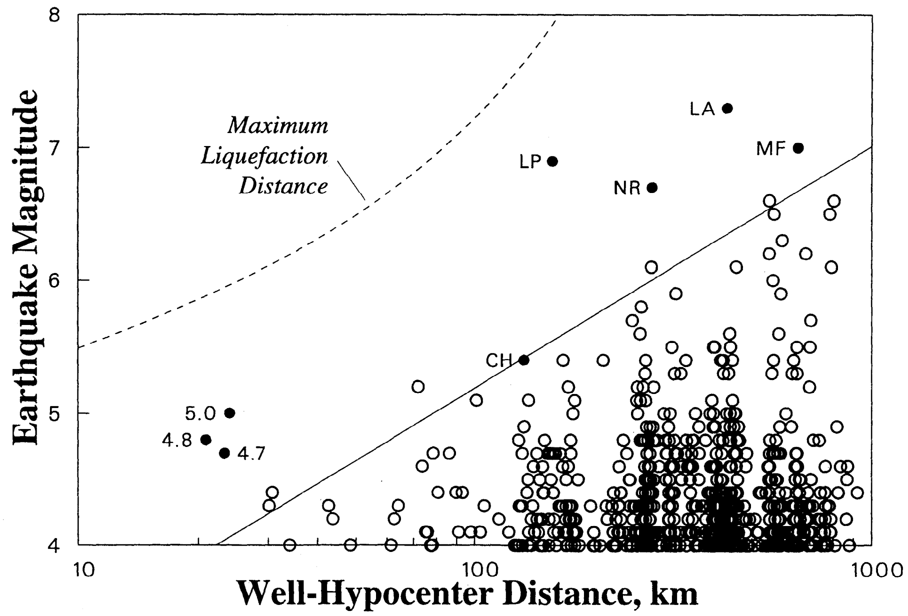

Figure 7b. Graph showing estimated threshold for a water level rise more than 5 cm in the BV well as a function of earthquake magnitude and distance, together with magnitudes and distances of earthquakes from the Calnet (for latitudes 35.5° and north) and Southern California (for latitudes south of 35.5°) catalogs for the period January 1, 1987, through September 30, 1995, excluding the periods listed in Table 2 for which no water level data are available (solid line). The dashed curve denotes the estimate by *Papadopoulos and Lefkopoulos* [1993] of the maximum distance of liquefaction.

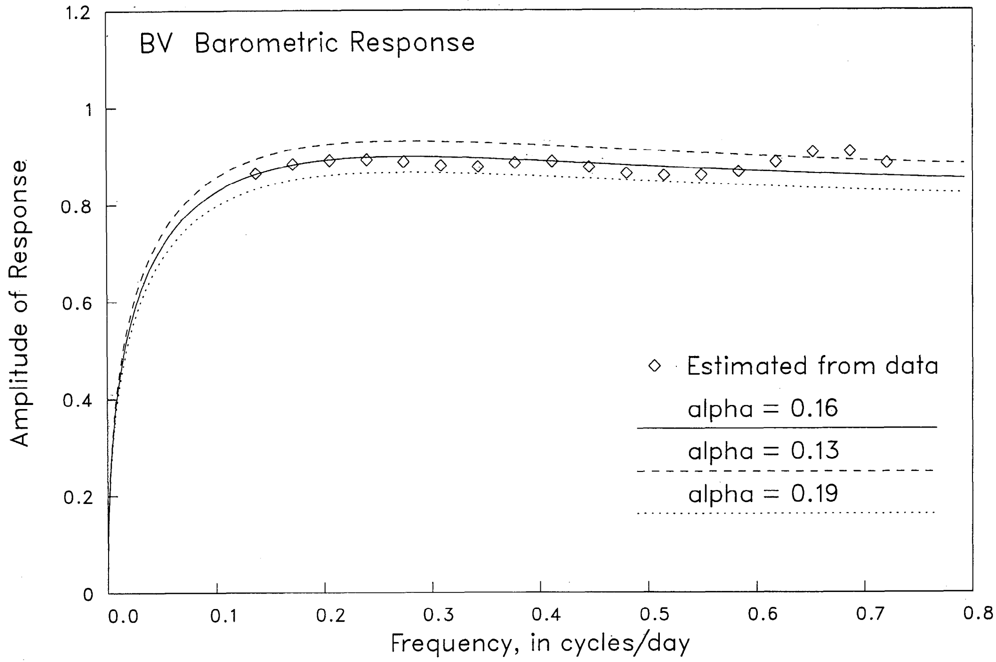

Figure 8. Amplitude of the barometric response of the BV well as a function of frequency.

the water budget, but it is simply assumed here that a constant proportion of the incident precipitation is evaporated or transpired each year.

Shallow subsurface storm flow is also unlikely to influence either the seasonal or the earthquake-induced water level changes in the well. Such flow has been observed to be significant only on convex hillslopes covered with extremely permeable soil, and it lasts only for hours to days following storms [*Freeze and Cherry,* 1979]. Water that reaches the creek via shallow storm flow will leave the catchment as streamflow, rather than percolate into the groundwater system, because at these times the fluid pressure gradient at the creek bed favors upward, rather than downward, flow.

Overland flow to the Pancho Rico Creek probably takes place during some storms and may percolate down into the groundwater system to increase piezometric level in the BV well. For example, 12 cm of rain fell on March 10, 1995, and water level in the BV well rose 2 m in the subsequent 4 days (Figure 4). The rate of rainfall likely exceeded the vertical hydraulic conductivity, temporarily saturating the surface, so that additional rain ran overland to the creek. This storm occurred before groundwater flow had caused water level in the BV well to reach its seasonal peak, so that the upward groundwater head gradient supplying flow into the creek bed was not fully developed, allowing water in the creek to then percolate downward to the well. Following this storm peak, water level in the well fell rapidly. The short time constants for both the rise and fall of water level in response to this intense storm are consistent with flow through a short, high-diffusivity path such as that between the well and the creek bed and enable such an event to be readily distinguished from the smoother, longer-period annual variations analyzed herein.

Flow from the creek to the well following storms seems unlikely to account for the earthquake-related water level changes. For example, it did not rain for 20 days prior to the Loma Prieta earthquake on October 18, 1989, so that the creek was almost certainly dry at the time the earthquake occurred and therefore unable to supply water to the well. Nevertheless, this earthquake produced the largest observed postseismic water level change.

The flow path most likely governing the seasonal behavior of the well is through the hillslope above the well. In this section, a mathematical model of head fluctuation in such an aquifer recharged by seasonal precipitation is fit to the observed seasonal variations. This mathematical model predicts that the amplitude of the seasonal water level variation is proportional to the amount of seasonal recharge, with a constant of proportionality depending on the horizontal hydraulic diffusivity, and that

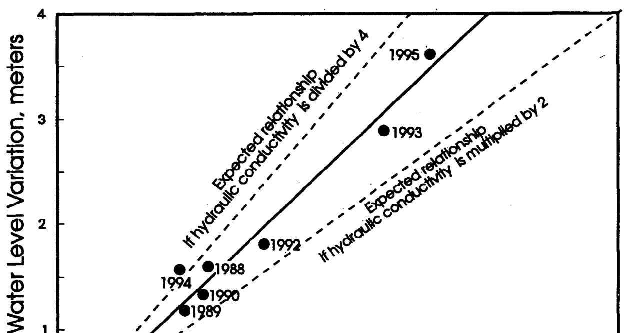

Figure 9a. Relationship between size of annual water level cycle and yearly amount of precipitation.

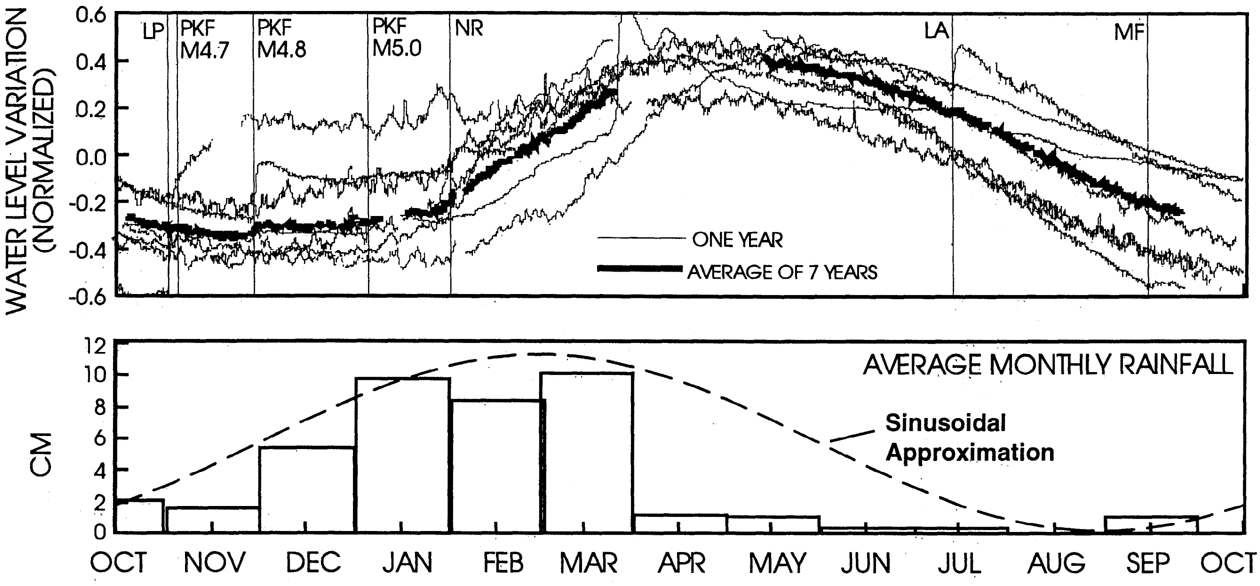

Figure 9b. Graphs of the seasonal variation in the BV well for the 1988 through 1995 water years, after dividing by their maximum values, together with the average of these curves and the average monthly rainfall totals.

the peak water level lags the peak precipitation by an amount of time that depends on diffusivity. Fitting this simple model to the data from the BV well permits estimation of the horizontal hydraulic diffusivity and places limits on how large a permanent diffusivity change can occur in this hillslope aquifer without modifying the relationship between water level and recharge.

In order to study the seasonal behavior of water level in the BV well, the hydrograph shown in Figure 4 was divided into annual segments, each beginning on October 1. The year beginning October 1, 1990, for example, will be referred to as the 1991 water year. The amplitude of the water level fluctuation for each year was determined as the difference between the maximum and minimum values observed during that water year; for the 1988 through 1995 water years, the amplitude of the fluctuation ranged from 1.1 to 3.6 m and averaged 2.0 m. For the same period, the total precipitation (virtually all rainfall) ranged from 25 to 80 cm, with an average of 41.3 cm/yr. Figure 9a shows a graph of the peak-to-peak amplitudes of the observed annual variations in water level and precipitation, and in Figure 9b these annual variations are superimposed after dividing by their peak-to-peak amplitudes. The average water level peaks on April 29 and reaches a minimum on November 3, almost exactly 6 months later, with maximum water levels in individual years occurring between March 15 and May 1. Peak precipitation occurs between January and March; mid-February is the center of this rainiest period, while August is almost always free of rain. Thus, on average, the peak water level lags the peak precipitation by 1 to 3 months, corresponding to 30°-90° in an annual cycle.

The relationship between the peak-to-trough amplitude h_0 of the water level cycle and the total annual precipitation P_0 is approximately linear and is estimated from the points in Figure 9a with linear regression as

$$h_0 = 4.19\, P_0 + 0.13 \text{ m.} \tag{12}$$

An expression is now derived for the relationship between recharge variation and water level variation that can be used to estimate the diffusivity for the aquifer tapped by the BV well, given knowledge of the aquifer length, the amount of recharge, and the amplitude and phase of the annual water level variation at the site. Consider a one-dimensional porous aquifer extending a distance L from a recharge area to a discharge point where the piezometric level is taken to be zero, as shown in Figure 10. The aquifer need not be confined, but it is assumed that the change in transmissivity and storage due to the seasonal head variation is negligible. The aquifer has horizontal hydraulic conductivity K_h and specific storage S_s, corresponding to horizontal hydraulic diffusivity $c_h = K_h / S_s$. The seasonal head variation in the aquifer is given as $h_s(x,t;\, K_h,\, S_s)$, where x represents distance from the recharge boundary and t represents time. Figure 9 shows that the seasonal variation of recharge $R(t)$ around its annual average R_{avg} can be approximated by a sinusoidal function

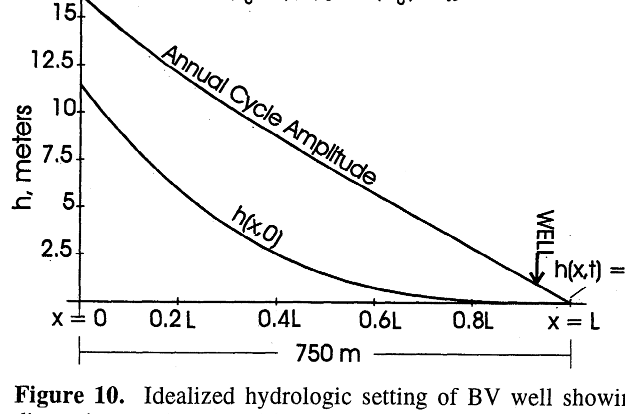

Figure 10. Idealized hydrologic setting of BV well showing dimensions and assumed boundary conditions.

$$R(t)-R_{avg} = \mathrm{Re}\left\{R_0 \exp\left[2\pi i\,(t-t_0)/T\right]\right\}$$
$$=R_0\cos\left[2\pi(t-t_0)/T\right] \qquad (13)$$

where T is the period of the seasonal cycle (in this case, 1 year), R_0 is the amplitude of the seasonal recharge variation, and t_0 is the time of peak seasonal recharge. $R(t)$, R_{avg}, and R_0 represent volumetric fluxes per unit area and therefore have the dimensions of length/time, with positive values indicating flow into the groundwater system. The boundary conditions are

$$h_s(L,t;\ K_h,\ S_s) = 0 \qquad (14a)$$

$$\left.\frac{\partial h_s}{\partial x}\right|_{x=0} = \frac{-[R(t)-R_{avg}]}{K}$$
$$= -\ \mathrm{Re}\left\{\frac{R_0}{K}\exp\left[2\pi i\,(t-t_0)/T\right]\right\} \qquad (14b)$$

The head, $h_s(x,t;K_h,S_s)$, satisfies the diffusion equation

$$\frac{\partial^2 h_s}{\partial x^2} = \frac{S_s}{K_h}\frac{\partial h_s}{\partial t} = \frac{1}{c_h}\frac{\partial h_s}{\partial t} \qquad (15)$$

(see, for example, *Freeze and Cherry* [1979]). Equation (15) is linear, so that $h_s(x,t;\ K_h,\ S_s)$ has the same periodic time variation as $R(t)-R_{avg}$ and can be expressed as

$$h_s(x,t;K_h,S_s) = \mathrm{Re}\left\{\hat{h}_s(x;K_h,S_s)\exp\left[2\pi i\,(t-t_0)/T\right]\right\} \quad (16)$$

The solution to the diffusion equation that satisfies these boundary conditions is

$$\hat{h}_s(x;\ K_h,S_s) = \frac{R_0}{K_h}\frac{\sinh\left[(2\pi i/c_hT)^{1/2}(L-x)\right]}{(2\pi i/c_hT)^{1/2}\cosh\left[(2\pi i/c_hT)^{1/2}L\right]} \qquad (17)$$

Here $\hat{h}_s(x;\ K_h,S_s)$ can alternatively be expressed in terms of dimensionless quantities $\gamma = c_hT/L^2$, $\kappa = K_hT/L$, $\xi = x/L$, and $\tau = t/T$, so that (17) becomes

$$h_s(\xi,\tau;\ \gamma,\ \kappa) = (R_0T/\kappa)\ \times$$
$$\mathrm{Re}\left\{\frac{\gamma^{1/2}}{(1+i)\pi^{1/2}}\frac{\sinh\left[\frac{(1+i)\pi^{1/2}}{\gamma^{1/2}}(1-\xi)\right]}{\cosh\left[\frac{(1+i)\pi^{1/2}}{\gamma^{1/2}}\right]}\exp\left[2\pi i\,(\tau-\tau_0)\right]\right\} \qquad (18)$$

For any position ξ, $\hat{h}_s(\xi;\ \gamma,\kappa)$ is a complex number representing the amplitude and phase of the water level variation relative to the recharge variation. Figure 11 is a plot of the amplitude and phase of $\hat{h}$ as a function of γ at the position ξ=0.93, corresponding to the location of the BV well, and shows that phase lags become small for γ larger than about 100. The observed 30° to 90° lag of peak water level behind peak precipitation corresponds to values of γ between 5.70 and 1.27; for T= 1 year and L = 750 m, these values of γ correspond to the diffusivity range $2.27 \times 10^{-2} < c < 1.02 \times 10^{-1}$ m²/s. This range of diffusivities is comparable to that expected for sand or

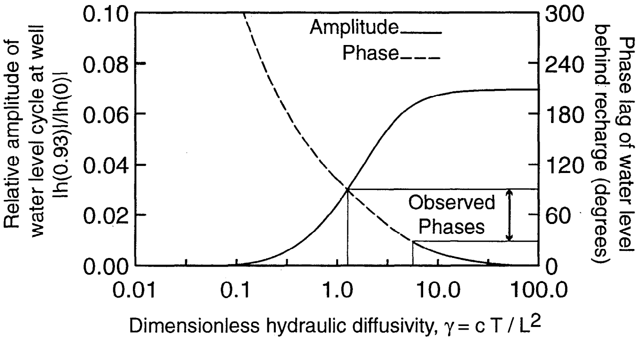

Figure 11. Phase of water level variation with respect to recharge variation as a function of dimensionless diffusivity, $\gamma = cT/L^2$, for the spatial coordinate x=0.93, which corresponds to the position of the BV well. A positive phase indicates lag of peak water level behind peak precipitation.

siltstone [*Roeloffs*, 1996] but is about an order of magnitude larger than the vertical diffusivity inferred from the barometric response. The inferred diffusivity does not depend on the amplitudes of either the water level variation or the recharge, only on their phase relationship.

With a range for c_h established, the factor R_0T/κ can be estimated from the amplitude of the observed average water level fluctuation in the well, which is 2.0 m peak to peak, corresponding to a true amplitude of 1.0 m. For this diffusivity range, evaluation of the quantity in brackets in (18) shows that the head fluctuation at the well site is 0.030 to 0.064 times R_0T/κ. The average annual precipitation is 0.413 m/yr; if all precipitation enters the groundwater system then R_0 is one half this rate (neglecting evapotranspiration or overland runoff) and a reasonable lower bound for R_0 is 20% of the upper limit (assuming that 80% of the precipitation does not enter the groundwater system). These assumptions constrain the hydraulic conductivity to the range 2.85×10^{-8} m/s < K_h < 3.20×10^{-7} m/s, an appropriate range for silt or silty sand [*Freeze and Cherry*, 1979]. Combining these estimates with the constraints on diffusivity places the specific storage in the range 2.8×10^{-7} m^{-1}< S_s < 1.4×10^{-5} m^{-1}.

This simple model accounts reasonably well for the seasonal variations at the BV well and allows a horizontal diffusivity to be inferred for the aquifer. The diffusivity that fits the data is constrained by the observed phase lag of peak water level behind peak precipitation to within a factor of 4.5. The hydraulic conductivity can be determined only to within a factor of 11, because the proportion of precipitation that enters the groundwater system is not known. The model further implies that changes in diffusivity or hydraulic conductivity would affect the amplitude and phase relationships between the water level fluctuations and the seasonal precipitation. In Figure 9a, lines have been plotted to show how these relationships would change in response to a change of hydraulic conductivity throughout the aquifer. The exact amount of change in the slope depends on the exact value of diffusivity, but for any diffusivity in the acceptable range, a factor of 2 increase or a factor of 4 decrease in hydraulic conductivity could be detected. Comparison of these lines with the scatter in the observation points suggests that if

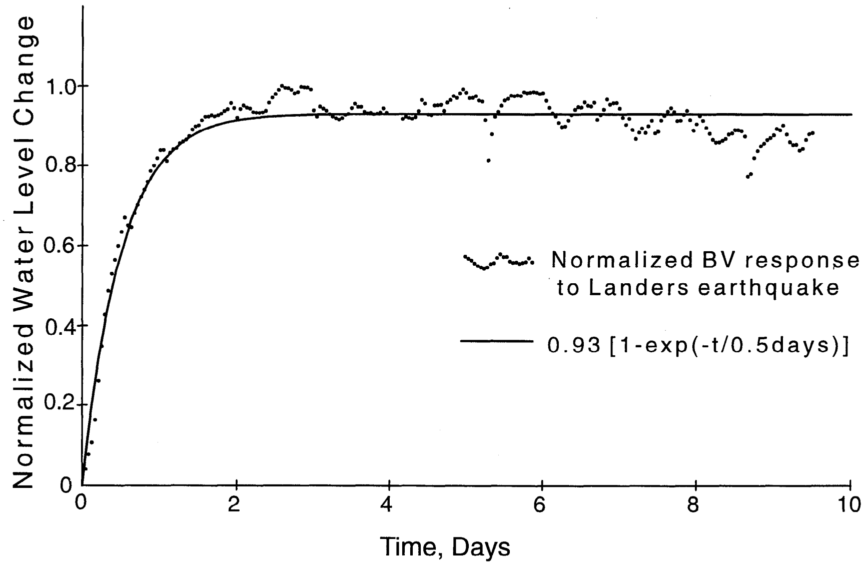

Figure 12. Fit of the BV response to the Landers earthquake, divided by its maximum value, to an exponential relaxation function.

earthquakes have permanently changed hydraulic properties at the BV site, then the change has been within these bounds. In section 6, the transient effects of such changes in aquifer properties are estimated.

6. Evaluation of Possible Mechanisms for Coseismic Water Level Changes

In this section, computed water level time histories based on three possible mechanisms for the coseismic water level changes are compared with the observed response to the Landers earthquake, which is typical of the common time history shared by the coseismic water level changes (Figure 6). This time history resembles the decaying exponential function

$$u(t) = u_0[1-\exp(-t/t_r)] \tag{19}$$

and Figure 12 shows that this function provides a good fit to the first 10 days of the response to the Landers earthquake for t_r=0.50 days. Equation (19) has been found to describe other coseismic phenomena. For example, *Mogi et al.* [1989] concluded that coseismic temperature increases at the Usami Hot Spring in Japan had decaying exponential time histories, but with a shorter time constant for nearby earthquakes (about 1 hour) than for distant earthquakes (2.5 to 3.3 hours). They attribute the coseismic temperature rises to increased subsurface conductance due to the passage of seismic waves but do not present a physical model giving rise to (19). For the Long Valley Caldera response to the 1992 Landers, California, earthquake, *Hill et al.* [1995] found that the first 6 days of tilt and borehole strain transients, as well as the cumulative number of triggered earthquakes, were all well described by (19) with t_r= 1.8 days. Although decaying exponential time dependence characterizes diffusive processes such as groundwater flow, (19), because it includes no spatial dependence, does not constrain the spatial extent of earthquake-induced fluid pressure changes. In subsequent sections, complete time- and space-dependent solutions for physical mechanisms are compared with the observed time history.

6.1 Response to an Earthquake-Induced Change of Hydraulic Properties

Rojstaczer et al. [1995] argued that coseismic stream discharge changes following the 1989 Loma Prieta earthquake were due to permeability enhancement by seismic shaking and proposed a simple model in which a permeability increase causes groundwater adjustment to a decreased hydraulic gradient. As the water table stabilizes at the new lower gradient, the base flow contribution to stream discharge returns to the preearthquake value in equilibrium with basin recharge.

The BV well responds to earthquakes too small or too distant to enhance permeability by opening partially cemented fractures, but low-amplitude shaking could change permeability in other ways. For example, gas bubbles lodged in the pore space may be expected to reduce the aquifer's permeability with respect to fluid flow [*Faybishenko*, 1995], and low-amplitude shaking may be able to mobilize the bubbles [*Beresnev and Johnson*, 1994], restoring permeability.

The response of groundwater level to a permanent change of hydraulic properties can be calculated independently of the reason for the change. Consider the same system shown in Figure 10, in which seasonal water level variations occur in response to seasonal recharge. Suppose that at time t=0, the hydraulic properties of the aquifer change. From the previous discussion of the steady state solution, it is apparent that if the horizontal diffusivity changes from c_0 to c_1, the amplitude and phase of the annual water level cycle will be affected. If the diffusivity and aquifer length are such that γ exceeds 100, then the phase will be unaffected but the amplitude will change because the head at the recharge boundary is

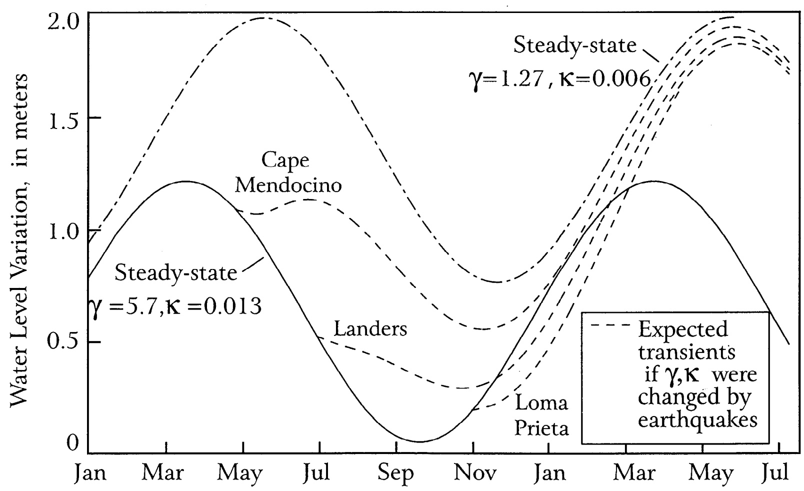

Figure 13. Steady state hydrographs for two pairs of aquifer properties consistent with the observed relationship between seasonal water level variation and peak precipitation and hypothetical curves showing the transient response expected if the change from one set of properties to the other occurred instantaneously at the time of the Cape Mendocino, Landers, or Loma Prieta earthquakes.

inversely proportional to hydraulic conductivity. For the BV well, however, the observed lag of peak water level behind peak precipitation implies diffusivities low enough that changes in the phase relationship would also be expected. A change of diffusivity may be associated with a change of hydraulic conductivity and/or specific storage. If the hydraulic conductivity and specific storage changed together in such a way as to preserve the same diffusivity, then the amplitude of the annual cycle would be affected but not its phase. The solution describing the transient response to such a change of hydraulic properties is developed in Appendix A. This solution (see (A10) through (A12)) depends upon the hydraulic properties before and after the earthquake, the position of the observation point in the aquifer, and the time during the seasonal cycle at which the earthquake occurs.

Figure 13 illustrates the calculated variation of head with time at the well site during 1.5 years. Seasonal cycles are shown for two dimensionless diffusivities $\gamma = 1.27$ and 5.70, which are the limits of the diffusivity range consistent with the observed phase relationship between the annual water level and precipitation cycles. It is also assumed that the total annual recharge is in the middle of the plausible range, specifically, that $R_0T = 0.12$ m and that $R_{avg} = R_0$, consistent with the usual situation of no rainfall during the dry season. Dimensionless hydraulic conductivities were chosen such that the amplitudes of the seasonal water level cycles approximate the observed value of 1 m and so that the water level corresponding to the higher diffusivity is always lower than that corresponding to the lower diffusivity.

Figure 13 also includes hypothetical water level responses from (A10)-(A12) if the aquifer properties were to change instantaneously at different times in the seasonal cycle owing to earthquakes. Specifically, it is assumed that γ decreases from 5.70 to 1.27 and that κ decreases from 0.013 to 0.006. Decreases in these parameters were chosen rather than increases in order to yield a postearthquake head that is higher than the preearthquake head at all times during the year, which seems most consistent with the observation of increased postearthquake water level. The calculated transients demonstrate that the head gradually adjusts to the seasonal cycle that is consistent with the new hydraulic properties. This adjustment, however, does not begin with an abrupt change in water level over a period of 1 or 2 days, as observed in the BV well. The solution will yield a faster risetime if higher diffusivities and higher contrasts between them are used, but these hydraulic properties would not be consistent with observed seasonal fluctuations in the BV well. Moreover, the calculated water level change at the time of the Loma Prieta earthquake is a water level drop, whereas a rise was observed. The seemingly counterintuitive result that the water level does not immediately change in the direction of the steady state solution corresponding to the new hydraulic properties is due to competition between the terms in (A10), arising from the sinusoidal and time-independent components of the transient solution. The equilibration of the time-independent component follows the same curve, regardless of the time during the year when the change takes place, but the sinusoidal component equilibrates at different rates and by different amounts at different times during the year. This feature is inconsistent with the observation that head in the BV well always rises following earthquakes. A change of properties at the time of the Cape Mendocino earthquake produces a change comparable in size to those at the times of the Landers or Loma Prieta earthquakes. Consequently, this mechanism does not explain why water level changes might be less likely to occur at or near the seasonal high water level. These features are also present when the transient solution is evaluated using different choices of hydraulic properties and suggest that the observed water level changes are not caused by permanent changes of aquifer properties in the hillslope above the well. If the earthquakes temporarily changed hydraulic properties, then the changes could be

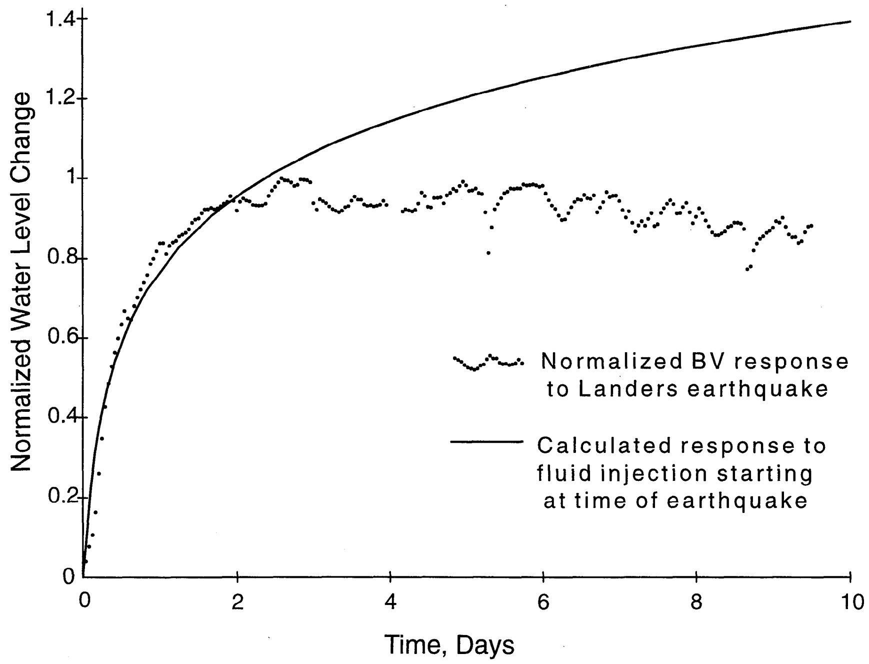

Figure 14. Fit of the first 2.5 days of the BV response to the Landers earthquake, divided by its maximum value, to (20), which gives the head response to an instantaneous decrease of discharge at a point.

larger without violating the constraints imposed by the seasonal variations, but the calculated changes would still not always be rises.

6.2 Response to a Coseismic Discharge Change

Coseismic changes in spring discharge [*Fujimori et al.*, 1995; *Koizumi et al.*, 1996] and in stream discharge [*Whitehead et al.*, 1984; *Muir-Wood and King*, 1993; *Rojstaczer and Wolf*, 1994] have frequently been reported. In two stream discharge cases where an investigation was made, the increased discharge was issuing from a single streambed spring [*Quilty et al.*, 1995; *Roeloffs*, 1993]. A creek bed spring has been mapped in the bed of the Pancho Rico Creek near the BV well, but its behavior has not been studied. The aquifer model shown in Figure 10 presumes that at least during part of the year the aquifer head distribution is maintained by discharge into the creek bed. Here I consider the possibility that the primary effect of an earthquake on the aquifer system is to decrease the discharge at a point instantaneously by an amount ΔQ. Regardless of the reason for the discharge change, the head change Δh in an infinite confined aquifer of thickness H is given approximately by the Theis equation

$$\Delta h(r,t) = \frac{\Delta Q}{4\pi HK} \int_{\frac{r^2 S_s}{4Kt}}^{\infty} \frac{e^{-y}}{y} dy \qquad (20)$$

[*Freeze and Cherry*, 1979], where r represents the distance from the discharge change to the observation well. Equation (20) is a monotonically increasing function of time and can therefore fit only the rising portion of the coseismic water level time history. Nonlinear regression yields a best fit of (20) to the first 2.5 days of the normalized response to the Landers earthquake when $\Delta Q/4\pi HK = 0.2745$ m and $r^2 S_s/4K = 0.0352$ days (Figure 14). Using the horizontal diffusivity range determined from the seasonal response places r in the range of 16.6 to 35.0 m, comparable to the distance between the well and the bed of the Pancho Rico Creek (50 m). The magnitude of ΔQ is not as well constrained because the aquifer thickness is unmeasured and because the hydraulic conductivity is less well known than the diffusivity. For aquifer thicknesses in the range of 15 to 45 m and hydraulic conductivities in the range $2.85 \times 10^{-8} < K < 3.2 \times 10^{-7}$ m/s, the flow rate required per meter of head change is $1.5 \times 10^{-3} < \Delta Q < 5.0 \times 10^{-2}$ L/s. Coseismic discharge changes of this magnitude could have escaped notice since discharge in the Pancho Rico Creek has not been measured.

Although a postulated coseismic decrease of discharge provides an acceptable fit to the first 2.5 days of a typical coseismic water level rise, for an amount of discharge that could have gone unnoticed, other features of this explanation are more equivocal. A gradual return of the discharge to the preearthquake value would be required to match the declining part of the coseismic water level record. Furthermore, most reported coseismic discharge changes are decreases, not increases.

6.3 Diffusion of a Localized Coseismic Pressure Increase

A third possible explanation for the observed coseismic water level rises is a localized pulse of elevated fluid pressure accompanying the earthquake. Such a pressure increase might originate by a mechanism similar to that of earthquake-induced liquefaction, wherein horizontal acceleration accompanying seismic shear waves densifies

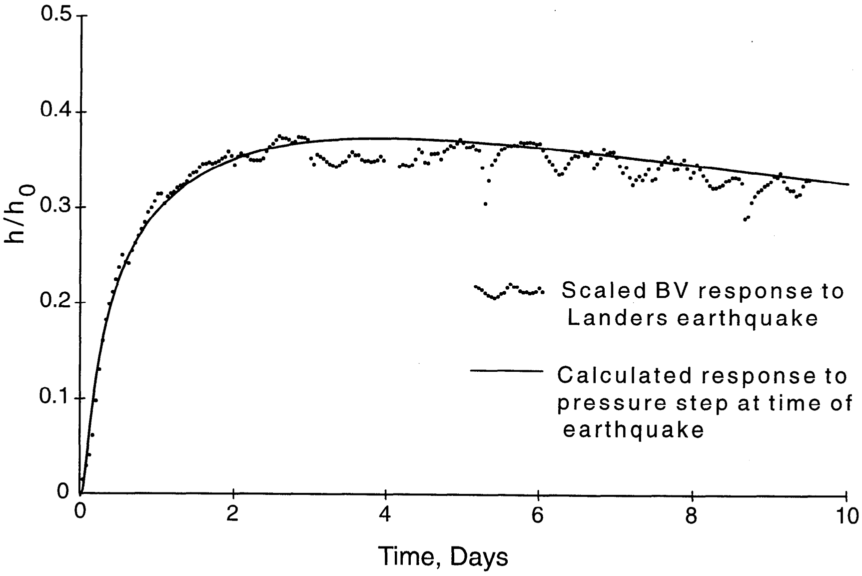

Figure 15. Comparison of the first 10 days of the BV response to the Landers earthquake with the expected pressure history resulting from diffusion of a localized instantaneous increase in aquifer pressure. The data have been scaled to the peak value of the curve.

unconsolidated deposits with an attendant increase of pore pressure [*Holzer*, 1989]. Liquefaction, which requires shaking-induced pore pressure to reach the overburden pressure, has not been observed in response to such distant earthquakes (*Papadopoulos and Lefkopoulos* [1993] and Figure 7b). Earthquakes can, however, cause pressure increases below this level [*Ishihara et al.*, 1981]. Like the coseismic water level changes in the BV well, the liquefaction mechanism always produces an increase of pore pressure, regardless of the earthquake's azimuth or focal mechanism.

In Appendix B, the solution is developed for the response to an instantaneous coseismic increase of pore pressure in the region $|x-x_0| < \varepsilon$ for some position x_0 within the aquifer. For this solution, pore pressure changes outside the region $|x-x_0| < \varepsilon$ are delayed, but for sources sufficiently close to the well, the delay can be less than the water level sampling interval. Figure 15 shows a possible fit of this solution to the Landers earthquake response, obtained with $c=10^{-3}$ m^2/s, x_0= 670 m, and ε=23 m. The edge of the area subjected to the pressure pulse is only 7 m from the well, but this proximity is required to avoid a delay between the earthquake and the onset of increased water level in the well. The diffusivity, which is constrained by the rate of pore pressure decay following the peak, is a factor of 20 lower than the lowest horizontal diffusivity estimated from the seasonal response and a factor of 3 higher than the highest vertical diffusivity estimated from the barometric response.

This solution has the attractive feature of accounting for the entire time history as the response to a phenomenon generated at the time of the earthquake. Moreover, the BV well is in shallow alluvium, where sufficiently strong shaking might be expected to cause liquefaction. If the mechanism is indeed the same as that of liquefaction, then this type of coseismic response may diagnose locations where future liquefaction is likely. On the other hand, the pore pressure increases in the BV well are as large as those recorded by *Ishihara et al.* [1981] in association with accelerations of 0.1 g, an order of magnitude higher than the estimated accelerations at the BV well from any of the earthquakes listed in Table 3. Moreover, the liquefaction-type mechanism cannot account for other coseismic hydrologic phenomena such as coseismic water level decreases [*Matsumoto*, 1992] or persistent water level changes in consolidated rock [*Woodcock and Roeloffs*, 1996]. *Hill et al.* [1995] rejected a localized coseismic pressure pulse as a mechanism for the Landers-induced transients at Long Valley Caldera because a localized source close to the strain instrumentation would have been required to avoid an unobserved delay.

Measurements of discharge in the Pancho Rico Creek would illuminate the mechanism of the coseismic water level changes. A mechanism related to that of liquefaction would produce coseismic discharge increases, as opposed to the discharge decrease mechanism considered in section 6.2. The estimated discharge changes are small and are at the lower end of the range that could be measured using a weir in the absence of a large component of steady flow.

7. Discussion and Conclusions

The coseismic water level rises in the BV well are similar to other coseismic water level changes reported in the literature that cannot be explained by water column resonance in response to seismic waves or by poroelastic response to coseismic static strain. Certain sites consistently exhibit this type of coseismic water level change, typified by a response that is always in the same direction, regardless of the earthquake's azimuth or focal mechanism, and by a magnitude threshold of occurrence roughly proportional to the inverse square of hypocentral distance. Of 12 wells monitored near Parkfield, only the BV well and possibly two other wells (MM and HR) respond predominantly in this manner. The other wells monitored at Parkfield respond to Earth tides, and the strain sensitivity inferred from their tidal response is a good predictor of their response to static coseismic strain.

To date, the coseismic water level rises in the BV well appear to have been limited by the seasonal water level maximum, and of the five earthquakes that would have been expected to cause water level rises but did not, four happened when the water level was at or very close to its seasonal maximum. On the other hand, the earthquakes that did not produce the expected responses occurred preferentially off strike of the plate boundary, suggesting a possible dependence on dynamic strain orientation. Nevertheless, the sizes of the water level changes are relatively well predicted as a function of earthquake magnitude and distance. The larger response to the Loma Prieta earthquake than the Landers earthquake, when compared with estimates of the peak dynamic stresses produced by these two earthquakes at Parkfield [*Spudich et al.*, 1995], is consistent with the water level being most sensitive to seismic waves with periods of 3 to 7 s. Responses to earthquakes of magnitude as low as 4.7 suggests that long-period surface waves do not cause the water level changes. Recording of ground motion or strain at the well site would clarify these issues.

The lack of frequency dependence in the response of the well to barometric pressure implies that the aquifer is fairly well confined, with vertical hydraulic diffusivity less than 3 x 10^{-4} m^2/s. High compressibility of the aquifer, probably due to a gas phase in the pore space, is the most likely reason why the well does not respond to Earth tides. The phase and amplitude relationships between the seasonal water level and precipitation cycles constrain the horizontal hydraulic conductivity to the range 2.27 x $10^{-2} < c_h <$ 1.02 x 10^{-1} m^2/s. These amplitude and phase relationships have remained constant enough to infer that any permanent earthquake-induced changes in the hydraulic properties of the hillslope above the well must be less than a factor of 4.

Within this range of aquifer properties, a permanent change of hydraulic conductivity and/or diffusivity throughout the hillslope aquifer cannot account for the coseismic water level changes because it does not reproduce their risetimes and, more compellingly, does not predict that the water level would change in the same direction at every time of the year. The first 2.5 days of a typical coseismic water level change could be caused by a coseismic discharge decrease at a point several tens of meters from the well, with the required amount of discharge undetectable by casual observation. Alternatively, the coseismic water level rises could represent diffusion of abrupt coseismic pore pressure increases in an area extending to within 7 m of the well, produced by a mechanism akin to that of liquefaction. Measurements of discharge would permit discrimination between these two possible explanations, as well as place tighter constraints on the hydrologic parameters of the basin.

Although no preearthquake changes have been observed at the BV well, it is possible that these coseismic water level changes represent a phenomenon closely related to coseismic water level, stream discharge, and groundwater temperature changes at other locations where preearthquake changes have also been reported. Further understanding of these phenomena would be facilitated by simultaneous measurements of discharge, temperature, water level, and chemical parameters at several such sites.

Appendix A: Transient Solution to an Instantaneous Change in Aquifer Properties

In this appendix the solution $h(x,t;\, K_0,\, S_{s0},\, K_1,\, S_{s1})$ is developed for the changes in groundwater level and head in a one-dimensional aquifer subject to a permanent, instantaneous change in diffusivity, hydraulic conductivity, and/or specific storage. The strategy is to find

$$h_t(x,t;\, K_0,S_{s0},K_1,S_{s1}) = h(x,t;\, K_0,S_{s0},K_1,S_{s1}) - [h_s(x,t;\, K_1,S_{s1})+(R_{\mathrm{avg}}/K_1)(L-x)] \quad \text{(A1)}$$

for $t>0$, where h_t represents the difference between the complete hydrograph, including the transient response, and a seasonal variation consistent with the aquifer properties K_1 and S_{s1}. Consider the same configuration shown in Figure 10 and assume that at time $t=0$ the hydraulic conductivity, specific storage, and diffusivity change from K_0, S_{s0}, c_0 to K_1, S_{s1}, c_1. Prior to time $t=0$, the head is given by (17), with c_0 replacing c_h, and for times much greater than $t=0$, the head is also given by (17), except that c_1 now plays the role of c_h. The solution developed here gives the difference, $h_t(x,t;\, K_0,S_{s0},K_1,S_{s1})$, between the head for $t>0$ and the seasonal solution, which depends on time and position, as well as on the hydraulic parameters before and after the change. For $t>0$, this function satisfies the diffusion equation

$$\frac{\partial^2 h_t}{\partial x^2} = \frac{S_{s1}}{K_1}\frac{\partial h_t}{\partial t}. \quad \text{(A2)}$$

The boundary conditions are constant head at the discharge boundary,

$$h_t(L,t;\, K_0,S_{s0},K_1,S_{s1}) = h(L,t;\, K_0,S_{s0},K_1,S_{s1}) - h_s(L,t;\, K_0,S_{s0}) = 0 \quad \text{(A3a)}$$

and, at the recharge boundary, zero difference between the transient and eventual steady state seasonal head gradients:

$$\left.\frac{\partial h_t}{\partial x}\right|_{x=0} = \left.\frac{\partial h(x,t;\, K_0,S_{s0},K_1,S_{s1})}{\partial x}\right|_{x=0} - \left.\frac{\partial h_s(x,t;K_1,S_{s1})}{\partial x}\right|_{x=0} + \frac{R_{\mathrm{avg}}}{K_1} = 0. \quad \text{(A3b)}$$

The initial condition is that

$$h_t(x,0;\, K_0,S_{s0},K_1,S_{s1}) = [h_s(x,0;\, K_0,S_{s0})+(R_{\mathrm{avg}}/K_0)(L-x)] - [h_s(x,0;\, K_1,S_{s1}) + (R_{\mathrm{avg}}/K_1)(L-x)]. \quad \text{(A4)}$$

The validity of these boundary conditions for times close to zero needs clarification. It is assumed that the change in hydraulic properties brought about by the passage of seismic waves takes place entirely during the time when the seismic wave amplitude is nonnegligible. Although the exact amplitude of shaking required to change the hydraulic properties is not known, it is reasonable to assume that shaking of sufficient amplitude lasts, at most, minutes for the earthquakes considered here. The solution developed below does not apply while hydraulic properties are changing. Once the change of hydraulic properties is complete, the boundary condition at $x=0$ presumes that the head gradient in an infinitesimal neighborhood of the recharge boundary has changed to a value consistent with the new hydraulic properties. The initial condition also specifies that the head at time zero is unchanged from its preearthquake value throughout the aquifer.

To solve (A2) subject to these boundary conditions, fundamental solutions of the form

$$h^n(x,t)=f^n(x)g^n(t) \tag{A5a}$$

are sought such that

$$\frac{d^2 f_n}{dx^2} = A_n f_n(x) \tag{A5b}$$

$$\frac{dg_n}{dt} = A_n(K_1/S_{s1})\, g_n(t) \tag{A5c}$$

where A_n are separation constants. The $f^n(x)$ and A_n are chosen to satisfy the boundary conditions exactly. Specifically,

$$A_n=-(2n+1)^2\pi^2/4\, L^2 \tag{A6a}$$

$$f_n(x) = \eta_n \cos[(2n+1)\pi x/2L] \tag{A6b}$$

where η_n are constants to be determined by matching the initial condition. With this choice of η_n, the time functions are proportional to

$$g_n(t) = \exp\left[\frac{-(2n+1)^2\pi^2 c_1 t}{4L^2}\right] \tag{A7}$$

so that

$$h_t(x,t;\, K_0,S_{s0},K_1,S_{s1}) =$$

$$\mathrm{Re}\left\{\sum_{n=0}^{\infty} \eta_n \cos\left[\frac{(2n+1)\pi x}{2L}\right] \exp\left[\frac{-(2n+1)^2\pi^2 c_1 t}{4L^2}\right]\right\} \tag{A8}$$

where $c_1 = K_1/S_{s1}$. The initial condition is that this function equals the difference between the steady state solutions for two different sets of hydraulic properties. Consequently, it suffices to find a_n such that

$$\sum_{n=0}^{\infty} a_n \cos\left[\frac{(2n+1)\pi x}{2L}\right] = h_s(x,0;\, K,S_s) + \frac{R_{\mathrm{avg}}}{K}(L-x). \tag{A9}$$

The a_n can be determined by multiplying both sides of (A9) by $\cos[(2m+1)\pi x/2L]$ and integrating over the interval from 0 to L to obtain

$$a_n(K,S_s) = \frac{2}{LK}\left\{\frac{R_0 \exp(-2\pi i t_0/T)}{2\pi i S_s/K + \left((2n+1)^2\pi^2/4L^2\right)} + \frac{R_{\mathrm{avg}}}{(2n+1)^2\pi^2/4L^2}\right\}. \tag{A10}$$

Equation (A8) then becomes

$$h_t(x,t;\, K_0,S_{s0},K_1,S_{s1}) = \mathrm{Re}\left\{\sum_{n=0}^{\infty}\left[a_n(K_0,S_{s0}) - a_n(K_1,S_{s1})\right]\times \cos\left[\frac{(2n+1)\pi x}{2L}\right]\exp\left[\frac{-(2n+1)^2\pi^2 c_1 t}{4L^2}\right]\right\} \tag{A11}$$

which is the solution sought. Equation (A11) represents the difference between the full transient solution and the steady state solution for the modified hydraulic properties, so that the total postearthquake head is given by

$$h_{\mathrm{trans}}(x,t;\, K_0,S_{s0},K_1,S_{s1}) = h_t(x,t;\, K_0,S_{s0},K_1,S_{s1}) + h_s(x,0;\, K_1,S_{s1}) + (R_{\mathrm{avg}}/K_1)(L-x). \tag{A12}$$

Appendix B: Solution for Pore Pressure Due to Localized Instantaneous Increase of Aquifer Pressure

It is assumed that the aquifer configuration is the same as shown in Figure 10. At time $t=0$, it is assumed that an instantaneous increase of pressure by an amount h_0 is imposed on the part of the aquifer defined by $|x-x_0| < \varepsilon$, with subsequent pressure changes throughout the aquifer governed by the diffusion equation (15). The boundary condition prescribing zero pressure change at $x=L$ (see (14a)) still applies, but the boundary condition at $x=0$ becomes

$$\left.\frac{\partial h_p}{\partial x}\right|_{x=0} = 0.$$

Solutions $h_p(x,t)$ can be expected to have the form

$$h_p(x,t) = h_0 \sum_{n=0}^{\infty} \lambda_n \cos[(2n+1)\pi x/2L]\exp[-(2n+1)^2\pi^2 ct/4L^2]. \tag{B1}$$

The initial condition is

$$h_p(x,0) = \begin{cases} h_0, & |x-x_0| < \varepsilon \\ 0, & \text{elsewhere.} \end{cases} \tag{B2}$$

The solution is given by (B1) with

$$\lambda_m = \frac{8}{\pi(2m+1)}\times \cos[(2m+1)\pi\, x_0/2L]\,\sin[(2m+1)\pi\varepsilon/2L)]. \tag{B3}$$

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