REVIEW

Streamflow and Water Well Responses to Earthquakes

David R. Montgomery[1] and Michael Manga[2,3]

Earthquake-induced crustal deformation and ground shaking can alter stream flow and water levels in wells through consolidation of surficial deposits, fracturing of solid rocks, aquifer deformation, and the clearing of fracture-filling material. Although local conditions affect the type and amplitude of response, a compilation of reported observations of hydrological response to earthquakes indicates that the maximum distance to which changes in stream flow and water levels in wells have been reported is related to earthquake magnitude. Detectable streamflow changes occur in areas within tens to hundreds of kilometers of the epicenter, whereas changes in groundwater levels in wells can occur hundreds to thousands of kilometers from earthquake epicenters.

Effects of earthquakes on the direction, quantity, and rate of surface and subsurface water flow have been noted for centuries. Over the past several decades, measurements of hydrological response to numerous earthquakes have quantified changes in both surface water (streamflow) and groundwater levels in wells (*1*, *2*). As a result of these observations, a variety of mechanisms have been proposed to explain hydrological responses to earthquakes (Fig. 1). Changes in streamflow and water levels in wells have been attributed to expulsion of fluids from the seismogenic zone (*3*), pore-pressure diffusion after strain occurs in the upper crust (*1*, *4*, *5*), compression of shallow aquifers (*1*, *6*–*8*), increased permeability of surficial materials resulting from either shaking of near-surface deposits (*9*–*11*) or opening of bedrock fractures (*12*–*14*), and decreased permeability resulting from consolidation of surficial deposits (*15*–*17*). Although often relegated to the status of strange or curious phenomena, pore-pressure responses to distant earthquakes may trigger local earthquakes (*18*), and chemical and thermal changes in hydrological systems have been suggested as having potential use in earthquake prediction (*19*, *20*). Moreover, interactions between seismological and hydrological processes offer the potential for new insights into the temporal and spatial variability of hydrological properties and processes at scales ranging from pores to continents.

The great variety of reported hydrological responses to earthquakes may be systematized by considering near-field versus far-field, transient versus sustained, and rapid versus delayed responses. More specifically, we suggest that it is helpful to distinguish between near-field effects within about one rupture length of the fault, intermediate-field effects from 1 to 10 fault-rupture lengths away, and far-field effects at greater distances. It is also useful to distinguish the transient response to passage of seismic waves and the effects of liquefaction from more sustained or permanent responses arising from aquifer compression or dilation. Transient response ranges from rapid oscillations in water levels in wells to increased stream flows that last for weeks, whereas sustained response can persist for months or reflect permanent changes in aquifer properties. Rapid response initiates during ground shaking (coseismic), whereas delayed response arises after ground shaking ceases (postseismic). These different styles of response reflect different mechanisms, as well as proximity to the epicenter and geological context.

Different mechanisms likely characterize responses at various distances from epicenters. In the nearfield, changes in properties of the fault zone itself can influence both groundwater levels and stream flow (*21*). Liquefaction and consolidation also can occur in the nearfield to the intermediate field (*22*), and far-field effects most likely result from the interaction of aquifer properties and transient strain during the passage of seismic waves. Here, we review observations from studies of hydrologic responses to earthquakes and compile observations on the relations between surface and subsurface response, distance from the epicenter, and earthquake magnitude.

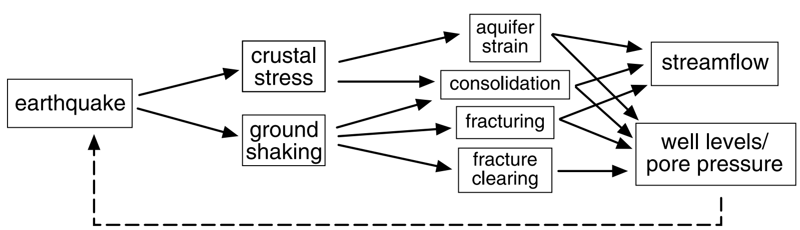

Fig. 1. Interactions between earthquakes and hydrological processes.

Groundwater Response

The development of seismographs and automated water level monitors in the early 20th century led to the now common observation that seismically induced oscillation in water levels in wells provides a form of hydroseismograph. Both near-field and far-field well responses can include high-frequency oscillations (*23*–*27*) and step responses that may be either transient or sustained (*28*, *29*). Analytical solutions have been developed for both the high-frequency (ranging from seconds to minutes) oscillation of groundwater levels and pore-pressure fluctuations in response to the passage of seismic waves (*2*, *30*, *31*), as well as the strain amplification that leads to well level oscillations. Such a response depends on the interaction between the flow in the well and the flow into and out of the aquifer (*32*). The particular amplitude of transient high-frequency well response at any given site is a result of the site-specific effect of the interaction of the aquifer properties and the volumetric strain associated with propagating seismic waves (*33*).

Sustained changes in water levels in wells depend on both the structure of the aquifer and the direction and magnitude of local seismically induced strain (*2*). A good example of such effects is provided by postseismic water level decreases in wells situated on bedrock ridges, changes that have been attributed to water table lowering resulting from the opening of bedrock fractures (*13*, *14*, *34*). Changes in water levels have also been attributed to local strain, such as in cases in which water levels in wells rise in zones of compression on the hanging walls of

[1]Department of Earth and Space Sciences, University of Washington, Seattle, WA 98195, USA. [2]Department of Earth and Planetary Science, University of California, Berkeley, CA 94720, USA. [3]Earth Sciences Division, Lawrence Berkeley National Laboratory, Berkeley, CA 94720, USA.

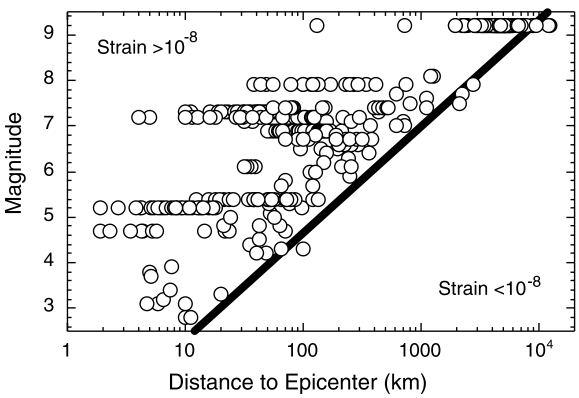

Fig. 2. Distance from epicenter (or hypocenter if reported) versus earthquake magnitude (moment magnitude M_w, if reported) for locations reported to have exhibited seismically induced changes (high-frequency to persistent) in well levels or groundwater-controlled springs. The line represents the theoretical limit to coseismic strain $>10^{-8}$ derived by Dobrovolsky *et al.* (*37*). Compiled data were reported directly in previous publications or measured off of maps presented in the original sources (*5*, *10*, *13*, *26*, *28*, *34*, *41–55*), with the exception of observations from the 1964 Alaska earthquake (*23*), which were corrected for the chord length from the point of observation to the earthquake epicenter, as were data for the 2002 Denali earthquake (*29*). Data for the 1906 San Francisco earthquake (M_w = 7.9) are based on accounts reported in Lawson (*56*). Each data point represents the response of a single well; response to a single earthquake may include many wells (such as that from the M_w 9.2 1964 Alaska earthquake).

normal faults (*7*, *35*). Water levels in wells drilled in unconsolidated valley-bottom deposits have exhibited seismically triggered rises that have been attributed to aquifer compaction (*35*). A sustained change in well level reflects nonrecoverable deformation, such as from changes in aquifer storage capacity (*36*). Different styles of sustained changes in near-field groundwater levels can arise from the interaction of seismic deformation, site-specific geologic structure, and the topography.

Dobrovolsky *et al.* (*37*) derived theoretical relations between earthquake magnitude, distance from the epicenter, and volumetric strain and reported that, in general, detectable seismically induced strain exceeds 10^{-8} (*38*). In a typical aquifer, compressive strain of 10^{-6} could produce a pressure rise of 1 m of water (*2*), whereas strain of 10^{-8} would produce a 1-cm rise barely detectable with high-precision instrumentation. We compiled data from the response of 912 wells to more than 44 earthquakes and show in Fig. 2 that the distance to which earthquake-induced changes in water levels in wells have been reported increases systematically with earthquake magnitude. Moreover, the limit to which water well response has been reported corresponds to the distance at which a strain of 10^{-8} would be expected. Hence, the coseismic response of wells to earthquakes extends as far from earthquake epicenters as we are able to reasonably detect such changes in typical near-surface aquifers.

The causes of delayed far-field changes in water levels in wells hundreds to thousands of kilometers from an epicenter are not well understood. Brodsky *et al.* (*29*) proposed that local ground shaking can loosen fracture-blocking colloidal material, which in turn results in changes in water pressure.

Streamflow Response

Streamflow changes observed after earthquakes include near-field and intermediate-field transient responses and sustained changes that altered the low discharges that define base flows in the periods between storm events. Most studies reporting streamflow changes record increased discharge, and three types of mechanisms have been proposed to explain changes in stream flow after earthquakes: (i) transient streamflow changes resulting from aquifer deformation, (ii) sustained changes in stream discharge resulting from changes in hydraulic conductivity or the opening or closing of near-surface fractures, and (iii) transient changes resulting from consolidation of surficial deposits. In some cases, the absence of evidence for changes in the rate of change of stream discharge during periods between storm events has been used to argue that the properties of aquifers feeding surface flows did not systematically change in response to seismic events (*15–17*).

Liquefaction of valley-bottom deposits provides a mechanism for rapid increases in stream flow, because compaction or settling of saturated near-surface deposits could yield large volumes of runoff as a result of the high specific storage that is typical of shallow aquifers. Moreover, liquefaction of valley-bottom deposits occurred in areas that experienced increased stream flow because of the 1983 Borah Peak (*6*) and the 2001 Nisqually (*17*) earthquakes. Although consolidation of surficial deposits could produce rapid runoff response, in some cases it may be difficult to explain large volumes of sustained streamflow increases by this mechanism (*16*). Papadopoulos and Lefkopoulos (*39*) reported an empirical limit to the maximum distance to which liquefaction has been documented from earthquake epicenters as a function of earthquake magnitude (*40*). We compiled data for 221 gauging stations where streamflow response was reported for more than 16 earthquakes, and we show in Fig. 3 that the limiting distance to which liquefaction has been reported also defines the upper limit to the envelope of reported streamflow responses to earthquakes. Although this compilation suggests that the maximum extent of streamflow responses is limited by the ability of liquefaction to occur, this does not mean that liquefaction is the exclusive cause of increased flow. Whereas other mechanisms likely affect near-field streamflow response, we suggest that liquefaction generally may be the limiting control on the distance to which such response occurs.

In some cases, near-field–to–intermediate-field streamflow responses appear more closely related to structural changes. In the 1983 Borah Peak earthquake, increases in stream flow occurred on the down-dropped hanging wall of the fault, whereas no response was observed on the footwall. Streamflow responses to the 1999 Chi Chi earthquake in Taiwan showed a pattern of increased and decreased stream flow, corresponding to areas of compressional and extensional deformation (*35*). Areas with the greatest increase in stream flow in response to the 2001 Nisqually earthquake were located in the Seattle Basin on the down-dropped side of the deep seismogenic normal fault (*17*). Hence, streamflow response to earthquakes can reflect a range of mechanisms that give rise to sustained versus transient and rapid versus delayed response. Consequently, a key challenge for studies of hydrological response to earthquakes is to elucidate the circumstances that favor each mechanism.

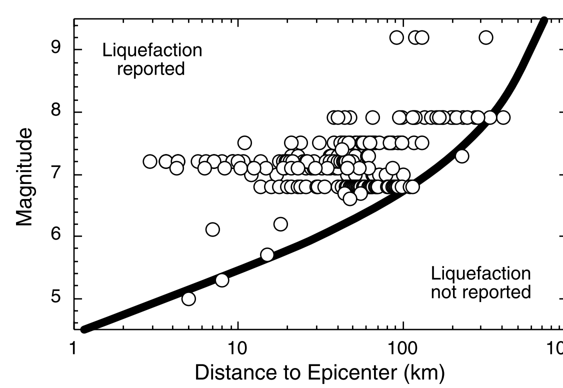

Fig. 3. Distance from epicenter versus earthquake magnitude, M_w, for gauging stations that exhibited seismically induced streamflow response. The line represents the empirical limit to the distance from the epicenter, beyond which liquefaction has been observed (*39*). Data represent increases in stream flow reported for the Nisqually earthquake (*17*), observations compiled by Manga (*15*), and additional data from published reports or taken from published maps (*9–12*, *14*, *16*, *50*, *51*, *53*, *54*, *57*). Data for the 1906 San Francisco earthquake (M_w = 7.9) are based on accounts of changes in stream flow and spring flow reported in Lawson (*56*).

Time Scale of Response

The time scale over which a change in water levels in wells occurs can be used to estimate the maximum depth at which the subsurface structure or fluid pressure has been altered by an earthquake. Roeloffs (*2*) derived a relation to estimate the time scale of vertical pore-pressure diffusion to the water table in near-surface aquifers as a function of hydraulic diffusivity D and depth below the water table that a hydrologic response originated. For typical hydraulic diffusivities of unconsolidated sands ($D \approx 1$ to 10^2 m^2 s^{-1}), seismically triggered pore-pressure response originating within 100 m of the water table would reach the water table within hours to days of an earthquake. For a hydraulic diffusivity more typical of fractured rock (10^{-2} to 10 m^2 s^{-1}), the same pore-pressure signal would take from hours to a year to reach the water table. Consequently, near-instantaneous high-frequency water level changes in a well must result from stress or changes in the vicinity of the well, whereas delayed or sustained changes may involve deeper or more distant sources.

Different mechanisms of hydrologic response to earthquakes should have different ranges of characteristic response times, and one also should expect to see a wide range of time scales for hydrologic response to earthquakes, depending on the local geological and geomorphological contexts. Streamflow response to the Nisqually earthquake, for example, occurred within 12 hours, indicating a shallow source, and therefore precluding expulsion of overpressured fluids in the seismogenic zone and pore-pressure diffusion after coseismic strain in the upper crust as mechanisms for the observed streamflow changes associated with the earthquake (*17*). Sustained streamflow response, such as the greatly increased stream flow that persisted for many months after the Loma Prieta earthquake (*13*, *14*), indicates a deeper source, whereas rapid transient streamflow response to earthquakes may reflect the expulsion of water from surficial deposits in response to consolidation during ground shaking. Distinguishing among the potential mechanisms of hydrological response to earthquakes should incorporate consideration of time scales, volumes of water, spatial patterns of response and their relation to the distribution of surficial deposits, and patterns of seismically induced strain.

Conclusions

Our comparison of the maximum distance from earthquake epicenters for stream flow and well response highlights differences in the spatial extent of surface versus subsurface hydrologic response to earthquakes. Models for the spatial limits of liquefaction and detectable response to strain characterize the maximum extent of areas over which susceptible sites respond, but the magnitude of hydrologic response to earthquakes is inherently site-specific because of local geological conditions. The variety of interactions between hydrological and seismological processes provides opportunities for insight into controls on the development of crustal-scale permeability and aquifer properties, fault behavior and strength, and the nature and distribution of seismic hazards. Moreover, the different spatial scales over which surface and subsurface hydrologic responses to earthquakes occur suggest different general mechanisms for near-surface sources of stream flow and deeper groundwater response.

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22. Liquefaction resulting from earthquakes generally occurs in loose sandy soils that densify during ground shaking, reducing the pore volume. The reduction in pore space increases pore water pressures and reduces the soil strength by reducing the effective normal stress. Liquefaction occurs when the shear strength of the soil is reduced enough that it no longer can resist failure and the soil behaves as a viscous liquid. Increased pore water pressures that result from liquefaction force water upward toward the water table.
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32. Specifically, a response function derived for the amplification A of seismic waves in water wells by assuming both radial flow into the well and that pore-pressure changes in the aquifer are proportional to dilational strain associated with passage of the seismic wave predicts
$$A = \{[1 - (\pi r_w^2/T\tau)\,\mathrm{Kei}\,\alpha_w - (4\pi^2 H_e/\tau^2 g)]^2 + (\pi r_w^2/T\tau)\,\mathrm{Ker}\,\alpha_w)^2\}^{-0.5}$$
where r_w is the radius of the well, T is the transmissivity (units of length^2/time), τ is the period of the seismic wave, α_w is a dimensionless constant given by $r_w(\omega S/T)^{0.5}$, ω is the angular frequency of the seismic wave (radians), S is a dimensionless aquifer storage coefficient, H_e is the effective height of the water column, and Kei and Ker are Kelvin functions (*30*).
33. Large response can reflect either high transmissivity (the product of hydraulic conductivity and aquifer thickness) or high specific storage, S_s, which is the volume of water released by a unit volume of an aquifer in response to a unit decline in hydraulic head, and is given by $S_s = \rho g\,(\alpha + n\beta)$ where ρg is the unit weight of water, α is the matrix compressibility, n is porosity, and β is the fluid compressibility.
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36. Storativity is the product of S_s (*33*) and aquifer thickness, and it characterizes the volume of water released by an aquifer from storage per unit surface area of the aquifer in response to a unit decline in the hydraulic head normal to the surface.
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58. This review was partially supported by the director, Office of Science, of the U.S. Department of Energy under contract no. DE-AC03-76SF00098. Figure 1 is modified from an illustration by D. Mays, and H. Greenberg determined the chord distances for the response to the 1964 Alaska earthquake used in Fig. 3. The comments of three anonymous reviewers are greatly appreciated.