Annual Temperature Variations and Seasonal Seismicity in California and Nevada

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Abstract

Earthquake catalogs declustered of foreshocks, aftershocks and triggered seismicity for California and Nevada show a small but statistically significant annual variation in the rates of mainshocks above magnitude 2. The lowest earthquake rates are observed in the late winter and early spring, while the highest rates take place in the summer and fall. This annual variation in the seismicity of California and Nevada is statistically significant at a very high confidence level. Evidence for similar seasonal variations in seismicity rate have been reported at some of the Cascades volcanoes and in the Himalaya, while the opposite pattern with the greatest seismicity rate in the late winter and spring has been found at the eastern south moat of the Long Valley caldera. Previously proposed explanations of the possible causes of annual seismicity rate variations are increases at seismogenic depths in pore-fluid pressures from groundwater recharge due to the seasonal rainfall and snowmelt, and increases in the surface vertical load due to seasonal variations in air pressure, rainfall or snowfall. The former explanation is difficult to reconcile with the long delay time (about 6 months) between the rainfall and the time of the maximum seismicity. The latter explanations are generally not satisfactory since there is relatively little rainfall and no snowfall in much of California and Nevada, and the annual air pressure changes in California and Nevada are only about 2 kPa. We suggest that the observed seasonal seismicity variations in California and Nevada are generated by thermoelastic strains in the upper crust associated with seasonal surface temperature changes. Spatial variations of the surface temperatures yield surface tractions that cause stress variations many kilometers into the crust. This mechanism can produce upper crustal stresses up to 7 kPa with maximum and minimum stresses in autumn and spring.
Introduction

The identification of repeatable temporal patterns in regional seismicity is an important goal in studies of earthquake physics and forecasting. If the temporal behavior of the seismicity of a region follows a Poisson distribution that is unchanging with time, then each earthquake is uncorrelated with every other event and the past earthquake activity contains no clues about what future earthquake activity will occur. On the other hand, if deviations from a stationary Poisson process can be found in the earthquake activity of an area, then these deviations can be investigated for their causes, which can lead to the identification of physical models upon which improved earthquake forecasting can be based. In particular, identification and modeling of the interaction of seismicity with regular external loadings can be used to derive information on aspects of earthquake physics, crustal strength, and time-dependent susceptibility of a region to the occurrence of large events. Examples of regular external loadings include solid earth tides, seasonal precipitation, variations of groundwater, barometric pressure changes and seasonal variations of the surface temperature.

Foreshocks and aftershocks represent well-known deviations from a stationary Poisson process (e.g., Utsu, 2002). Foreshocks reflect accelerating brittle deformation that may occur on a fault in preparation for a coming mainshock, while aftershocks represent brittle relaxation of the crust to rapid stress loading generated by previous earthquakes (e.g., Scholz, 1990; Ben-Zion, 2008). Additional possible deviations of seismicity from a stationary process include quiescence and activation over a broad region around the eventual rupture (e.g., Mogi, 1981) and triggering patterns associated with static and dynamic stress changes (e.g., Harris, 1998; Hill and Prejean, 2007). In California, when foreshocks and aftershocks are removed from the earthquake catalog, the remaining catalog of mainshocks closely follows a Poisson distribution (Gardner and Knopoff, 1974). Non-Poissonian elements may exist in the seismicity of the Earth (Ebel and Kafka, 2002), but in most places the non-Poissonian component is small (Kagan and Jackson, 1991).

In this paper, we present evidence that there are small seasonal variations in the rate of mainshocks above magnitude 2.0 in California and Nevada. In order to detect and analyze these small variations we use declustered catalogs where likely candidates for foreshocks and aftershocks are removed. The seasonal variations are statistically significant at a very high confidence level. We also investigate the possible cause of the observed temporal seismicity rate variations. For the analyzed California and Nevada earthquake data set, we suggest that the best explanation is given by thermoelastic stresses generated by the surface temperature variations and transmitted from the surface to seismogenic depth.

Annual Seismicity Variations in California and Nevada

The spatial area of this study extends from 32.0°N to 40.0°N and from -114.0°E to -126.2°E (Figures 1 and 2). This study area covers most of the state of California, although it excludes the seismicity associated with the subduction of the Juan de Fuca
plate beneath northern California. The study area also covers most of the state of Nevada. Seismicity from the ANSS earthquake catalog for two different time periods was analyzed, where the time periods were selected to assure completeness of the catalogs for all magnitude levels. For earthquakes with M \geq 2.0, the time period analyzed is from 1990 through 2007 (Figure 1). For earthquakes with M \geq 4.0, the time period analyzed is from 1932 through 2007 (Figure 2). According to Engdahl and Rinehart (1991), the M \geq 2.0 earthquake data set and the M \geq 4.0 earthquake data set should each be complete for the respective time periods that were used in the analyses. For each data set, foreshocks and aftershocks were removed using the time and distance windows as a function of magnitude defined by Gardner and Knopoff (1974) in order to obtain declustered earthquake catalogs that consist only of mainshocks. Because there are distant triggered earthquakes that were reported following the 1992 Landers, California and other earthquakes (e.g., Hill and Prejean, 2007), we decided to remove all earthquakes within the first 24 hours of each earthquake of M \geq 6.0 in the catalogs. While this simple definition of triggered seismicity is arbitrary, most of the events that were removed as triggered seismicity followed the Landers earthquake. Since relatively few earthquakes following all of the other M \geq 6.0 earthquakes in the catalogs were removed as triggered events, it was judged that this definition has an insignificant effect on the results of the analysis described below. The declustered M \geq 2.0 earthquake data set from 1990-2007 is shown in Figure 1, and the declustered M \geq 4.0 earthquake data set from 1932-2007 is shown in Figure 2.

To look for seasonal variations in the earthquake rates, a histogram of the cumulative number of earthquakes on each day of the year was created for each of the examined catalogs. The histograms were then smoothed with moving average filters to emphasize the long-term trends in the data. The data from one end of the year was wrapped around to the other end to remove edge effects in the smoothing process. Figure 3 shows the raw histogram and the histogram after a 90-point moving average filter was applied to the 1990-2007 M \geq 2.0 earthquakes data set. Also shown in Figure 3 are the raw and filtered histograms for the 1990-2007 data including only earthquakes with M \geq 3.0. In the filtered histograms for both M \geq 2.0 and M \geq 3.0 data sets, there is a clear annual variation in the seismicity rate, with higher seismicity rates in the second half of the year and lower seismicity rates in the first half of the year. The smoothed M \geq 3.0 histograms show a single cycle of lower-than-average seismicity between about days 30 and 170 (midwinter to late spring) with higher-than-average seismicity between about days 190 and 340 (mid summer to fall). The smoothed M \geq 2.0 histograms also show a lower-than-average seismicity rate between about days 350 and 140, but there are two peaks in the time of higher seismicity in the second half of the year with a short period of near-normal seismicity rate between them. The first peak occurs at about day 180, while the second peak occurs about day 330. In the M \geq 2.0 data, there is an increase of about 3 earthquakes each day between the times of the year of lowest and highest activity. For the M \geq 3.0 histogram, the annual rate variation is of the order of 1 earthquake each day.

Figure 4 shows the raw and 90-day smoothed histograms for the 1932-2007 earthquakes. This data set was analyzed for mainshocks with M \geq 4.0 and for mainshocks with M \geq 5.0. The smoothed M \geq 4.0 histograms show that a lower-than-average seismicity
rate occurs between about days 1 and 180, while the seismicity rate is above average between about days 250 and 350. The rate increase is about 1 earthquake per day between the least and most active times. For the smoothed M≥5.0 data set, there is a minimum in the smoothed histograms at about day 100, and the maximum is at about day 180. The seismicity rate stays at or above average from about day 150 to day 340.

To test the statistical significance of the seasonal seismicity rate variations found in Figures 3 and 4, the number of mainshocks in the first and second halves of the year were counted, and the statistical significance of the deviation from a uniform annual seismicity rate was computed under the assumption that the earthquakes follow a stationary random Poisson process (Table 1a). For the M≥2.0, M≥3.0 and M≥4.0 data sets, the differences in seismicity rates between the first and second halves of the year are different from stationary Poisson seismicity at a statistical significance of better than 99% using a standard P test. The statistical significance of the seismicity rate difference between the first and second halves of the year in the M≥5.0 data set is better than 94%. From 1932-2007 for M≥6.0, there were 18 mainshocks in the first half of the year and 16 mainshocks in the second half of the year. Thus, the strong seasonal variations seen below M 5.0 in these earthquake catalogs disappear for earthquakes above about M 6.0.

An analysis similar to that in Table 1a was also carried out using the earthquake data from Figures 1 and 2 but the events from the Coso and Long Valley areas removed from the data set. This second analysis was made because Gao et al. (2000) reported a seasonal modulation in the rates of very small earthquakes (M≤1.5) primarily in volcanic areas in southern and central California following the 1992 Landers earthquake. Table 1b summarizes the seismicity rates in the first and second halves of the year in this reduced data set, and it shows seismicity rate variations similar to those reported for the full data set in Table 1a. Thus, in this study the seasonal seismicity variations for earthquakes with M≥2.0 in California and Nevada do not appear to be confined to hydrothermal/volcanic areas.

Other Studies of Seasonal Seismicity Variations

The analysis in the previous section is not the first to find seasonal variations in the seismicity rate in California. Furthermore, studies of other areas in Japan, the Western U.S., and in the Himalaya have reported seasonal variations in seismicity rates. The main results of such studies are summarized below and compared with the seasonal variations found in this work.

Gao et al. (2000) reported an annual modulation of the microseismicity in southern California following the occurrence of the 1992 Landers earthquake. They looked at the rate of M≤1.5 earthquakes in southern and central California before and after the Landers mainshock. During the 4.5 years before the Landers event, the rate of the seismicity in their data set showed no systematic change throughout the year. However, following the Landers mainshock they found an annual cycle in the microseismicity rate, with the minimum seismicity rate in April and the maximum in August. This annual cycle decayed exponentially with time in amplitude during the 4.5
years of the analyzed data following the Landers event. They reported that most of the annual cycle was associated with seismicity rate variations at hydrothermal or volcanic regions in southern and central California. They attributed the annual cycle to seasonal variations in surface atmospheric pressure and argued that these seasonal pressure changes altered the vertical (normal) stress and the fluid pore-fluid pressures in the volcanic areas, leading to the annual seismicity fluctuations that they observed.

Saar and Manga (2003) examined the seismicity at Mt. Hood, Oregon, and reported a seasonal variation in the cumulative number of earthquakes and cumulative seismic moment for earthquake with $M \geq 1$ from 1980 to 2001. The largest earthquake in the data set was $M_{\text{max}} 4.5$. They found that both the number of earthquake and the total seismic moment peaked during September and October, with another peak in January and February. They attributed the annual variations to groundwater recharge from spring snow melt, which took about 151 days to diffuse down to seismogenic depths. Christiansen et al. (2005) studied 20-year data sets of seismic activity from ten volcanic areas in the western U.S. At five of these areas, they found statistically significant seasonal variations in the seismicity of the top 3 km of the crust. In four of those areas (Mt. St. Helens, Hebgen Lake/Madison Valley, Yellowstone Lake, and Mammoth Mountain), the highest rates of seismicity were observed in the summer and autumn. In the eastern south moat of the Long Valley caldera, the peak seismicity rate was observed in the winter and spring. They favored snow unloading and groundwater recharge as the physical mechanisms that best explained their observations of seasonal seismicity variations.

Wolf et al. (1997) reported on an annual cycle at a site of persistent earthquake clusters near Mt. Ogden on the Alaska/British Columbia border. For the time period 1975 to 1991, they reported a strong annual variation in the cluster seismicity rates, with almost no earthquakes between February and April, increasing seismicity rate peaking in June, a decrease in the seismicity rate in August, an increase to the highest seismicity rate in October, and then a sharp decrease in seismicity in November. The pattern of the average monthly seismicity rate throughout the year that Wolf et al. (1997) show is rather similar to the seismicity rate plot for $M \geq 2.0$ earthquakes for California and Nevada (Figure 3) found in this study. They suggested that groundwater recharge due to spring snowmelt and summer rains is the best explanation of the annual seismicity variations that they found, and Lee and Wolf (1998) estimated the range of hydraulic permeability necessary for pore-fluid pressures to diffuse to the seismogenic depths of the earthquakes observed at Mt. Ogden. Oike (1978), Ogata (1983) and Costain and Bollinger (1996) detected seasonal periodicities in the rate of shallow earthquakes in Japan, which they attributed like the above studies to rainfall, snowmelt and groundwater recharge.

Bollinger et al. (2007) and Bettinelli et al. (2008) discussed seasonal seismicity rate variations for the Himalayan region of Nepal. Bollinger et al. (2007) analyzed a declustered catalog of earthquakes with local magnitudes $2 < M_I < 4$ from the Nepal seismic network for the time period 1995-2000 and found that the rate of earthquake mainshocks in that area is approximately twice as great in the winter as it is in the summer. Their explanation of this annual pattern is that summer monsoon rains in
northern India add a surface load that suppresses the thrust earthquakes in the area, and that the drainage of this surface load during the fall allows the seismicity rate to increase. Bettinelli et al. (2008) did further modeling and showed that the crustal stresses associated with the surface load changes due to the rainfall is expected to be 2-4 kPa.

**Model of Thermoelastic Strains and Stresses in the Crust**

In this study we propose that thermoelastic strains and stresses of the type described by Ben-Zion and Leary (1986) provide a likely candidate explanation of the seasonal variations in seismicity rate seen in the California and Nevada data discussed in the previous section.

In the model of Ben-Zion and Leary (1986), thermoelastic strain in the crust arises from spatial variations of the surface temperature field $T$ and/or variations in the thermal properties of the rocks. The latter may be mapped to the former, and the first order effects in a given region may be calculated by assuming a single wavelength $\lambda$ that characterizes the dominant length scale of the surface temperature variations. The dominant variations of surface temperature may arise from differential heating on the north and south sides of mountains, due to surface temperature contrasts between water and rock bodies, due to variations of insulation associated with different flora coverage, etc.

The spatial temperature variations lead to crustal strains that are described by the equation

$$
\varepsilon_{xx} = \sum_\omega [A_1(\sigma, k, \gamma, y)e^{-ky} + A_2(k, \gamma)e^{-\gamma y}]bT_\omega \cos(kx)e^{i(\omega t - \phi)},
$$

where $\varepsilon_{xx}$ is the horizontal plane strain, $\sigma$ is Poisson’s ratio, $k$ is the spatial wavenumber of the temperature field, $\gamma$ is the thermal decay constant, $x$ is the horizontal coordinate where the stress is being calculated, $y$ is the depth where the stress is being calculated, $b$ is the coefficient of thermal expansion, $T_\omega$ is the amplitude of the temperature time series at frequency $\omega$, and $\phi$ is the phase lag of the temperature time series at frequency $\omega$. Equation (1) contains an equivalent thermal surface traction term $A_1$ and an equivalent thermal body force term $A_2$. The surface traction function $A_1$ arises from the spatial variation of temperature field at a given depth level and is given by

$$
A_1(\sigma, k, \gamma, y) = \frac{(1 + \sigma)k}{(1 - \sigma)(\gamma + k)}[2(1 - \sigma) + \frac{k}{(\gamma - k)} - ky].
$$

The function $A_2$ represents the thermal expansion associated with the direct heating at a given point and is given by

$$
A_2(\sigma, k, \gamma, y) = \frac{(1 + \sigma)k}{(1 - \sigma)(\gamma + k)}[-\frac{k}{(\gamma - k)}].
$$

The thermal expansion term decays proportional to Re($\gamma$) with depth, with $\gamma$ defined as
\[ \gamma = k[1 + i\omega/\kappa k^2]^{1/2} \]  \hspace{1cm} (4)

where \( \kappa \) is the thermal diffusivity of the surficial materials. Because of the low thermal diffusivity of Earth materials, the second term in Equation (1) decays to negligible values within a few hundred meters of the Earth’s surface. On the other hand, the traction term in Equation (1) decays with depth over a length scale proportional to \( k=2\pi/\lambda \). For spatial temperature variations of the order of tens of kilometers, the strains and stresses associated with this first term of Equation (1) can be significant over most of the seismogenic zone below the Earth’s surface.

In applying Equation (1), Ben-Zion and Leary (1986) recommend using a composite model in which there is a thin, unconsolidated cover layer over a rock half-space (Figure 5). The purpose of the top layer is to account for soils and unconsolidated surficial materials through which the surface temperature field must diffuse before it can produce tractions on the cohesive rock below. This surface layer acts to delay, attenuate and low-pass filter the strain signal in the rock half-space. Ben-Zion and Leary (1986) showed that strains computed with the model, using local measurements of atmospheric temperature and unconsolidated layer of about 0.6 m that delay the strain signal by 23 days, were in good agreement with high-resolution strain observations at a tunnel site in the San Gabriel Mountains, California. Prawirodirdjo et al. (2006) showed that the model with unconsolidated layers that are about 0.5-1 m thick provide good fit to the seasonal variations in clusters of GPS instruments in three regions of southern California.

For an area like California and Nevada, there can be strong spatial variations in temperature at all times of the year. In summer, the coastal areas generally experience temperatures that are cooler than at the inland valleys and deserts, while the opposite situation exists in the winter. Lower valleys tend to experience warmer temperatures than higher elevation mountains. Lakes and other areas covered by water are subjected to smaller temperature variations than the adjacent land masses. Thus, there is a spatially heterogeneous temperature field throughout California and Nevada that is persistent with time. Through Fourier analysis, this spatial temperature field can be decomposed into a spectrum of temperature variations as a function of wavelength \( \lambda \). In implementing Equation (1), \( T_\omega \) is found by Fourier transforming the temporal changes of the temperature field observed at a given site that is representative of the annual temperature variations of the region. One can then superpose contributions from different spatial Fourier components, or simply consider an effective dominant spatial wavelength that is representative for the region under consideration, as was done by Ben-Zion and Leary (1986). In this study, we follow the latter approach.

To illustrate how deep thermoelastic strain may extend into the crust, the strain in Equation (1) was computed using realistic surface values of the model parameters. The average annual daily temperatures for a one-year period at Twenty-Nine Palms, California were used as the representative input temperature data. The other parameter values were set as follows: unconsolidated layer thickness = 1.2 m, \( \sigma=.33, b=.00001 /^\circ C \), and \( \kappa=.0864 \text{ m}^2/\text{day} \). Figure 6 shows the computed absolute value of plane strain \( e_{xx} \) at \( x \).
= 0 and $t = 0$ as a function of depth for four different values of spatial wavelength of the horizontal temperature field. As the wavelength of the temperature field increases, the maximum value of the strain at the surface decreases, but also the strain falls-off slower with depth. For wavelengths of 60 km and longer, there is significant strain to below 5 km depth. Thus, a large part of the seismogenic upper crust can experience seasonal strain variations due to temperature changes with spatial wavelengths of 30 km and longer, and these strain variations may have some effect on the seasonal seismicity rates.

Evaluating the Cause of Seasonal Seismicity Rate Variations in California and Nevada

As mentioned earlier, a number of different mechanisms have been proposed to explain why crustal seismicity exhibits annual variations in rate. Aside from the thermoelastic strain model that is discussed in the previous section, the causes of seasonal seismicity rate variations fall into two categories: (1) increases in the surface vertical load due to seasonal variations in air pressure, rainfall or snowfall, and (2) increases at seismogenic depths in pore-fluid pressure from groundwater recharge due to seasonal rainfall and snowmelt. The seasonal variations in seismicity rate should occur in-phase with the variations in the surface loads since any change in surface load is transmitted at the elastic wave speed into the rock below. On the other hand, because it takes some time for the pore-fluid pressure due to an input of surface water to diffuse down into the crust to seismogenic depths, seismicity rate increases should take place with some delay after the groundwater gets recharged. Furthermore, this delay is likely to vary from place to place due to differences in rock porosity and permeability between the surface and the depth at which the seismicity is taking place. As mentioned earlier, the thermoelastic strain generated at seismogenic depths in California and Nevada by the regional temperature variations is expected to be delayed by a few weeks, since the temperature field has to propagate first through an unconsolidated surface layer of about 1 meter thickness on average.

Table 2 evaluates the annual stress variations due to different vertical loads calculated in previous studies where annual seismicity rate variations have been reported. Also included in Table 2 is an estimate of the horizontal stress change due to the thermoelastic elastic strain model described in the previous section and annual stress produced by the solid earth tides. As seen in the table, the thermoelastic strain model, the snow load and the monsoon surface water recharge all give comparable peak stresses at seismogenic depths. The air pressure peak load is less than 50% and the annual Earth tide peak load is an order of magnitude smaller than the other considered causes. For the California and Nevada seismicity, the seismicity rate peaks sometime between August and November, while it reaches its minimum in February and March. The air pressure load is out of phase with this observed seismicity variation. Likewise, the annual rainfall in California occurs almost exclusively between October and April, with the peak rainfall taking place in the winter months (January and February). This also is out of phase with the observed seismicity. Furthermore, many of the seismically active areas of California and Nevada receive little or no snow, so there is no snow load in these areas. Thus, of the causes listed in Table 2 the thermoelastic strain model is the load that best explains the
observed seismicity rate variations in California and Nevada.

Table 3 evaluates the proposed pore-fluid pressure increases due to groundwater recharge from seasonal rainfall events and springtime snowmelt. There are large differences in the peak pore-fluid pressure changes that are reported for the different areas in Table 3. These differences arise due to differences in the rain/snowmelt volumes, in hydraulic permeability and diffusivity, and in depths to which the pore-fluid pressure changes must diffuse. The data in Table 3 fall into two categories, one where the lag between the groundwater recharge event and the seismicity is of the order of 1 month or less, and the other where the seismicity lags the groundwater recharge event by 5-6 months. The data set that best illustrates the correlation of seismicity rate variations and several heavy rainfall events is that from Mt. Hochstaufen in Germany by Hainzl et al. (2006). The results from that study show evidence that strong rainfall events trigger increases in near-surface seismicity and that the increase in pore-fluid pressure diffuses down to depths of about 4 km over a period of about 10 days following a large rainfall. This downward diffusion of the pore-fluid pressure is accompanied by a downward migration of earthquake hypocenters. However, the hydraulic parameters used by Hainzl et al. (2006) and other studies assuming hydrologic triggering of seismicity may be orders of magnitude larger than estimated values of hydraulic diffusivity for crustal rocks (Christiansen et al., 2007).

For California and Nevada, the period of peak rainfall is January and February, while the seismicity rate starts to increase in April or May and peaks somewhere between August and November (Figures 3 and 4). The time lag between the peak in the seasonal rainfall and the peak in the seismicity rate in California and Nevada is between 6 and 10 months. Furthermore, many of the earthquakes in California and Nevada occur in desert areas that receive relatively little rainfall even during the rainy winter months. Given the data in Table 3, it is difficult to ascribe the annual seismicity changes in California and Nevada to groundwater recharge. Bollinger et al. (2007) and Bettinelli et al. (2008) looked at the possibility that groundwater recharge from the annual monsoon rains in the Himalaya might be causing the seasonal seismicity rate variations that they observed. However, they expressed doubts that groundwater recharge is the cause of the seasonal rate variations because of the long delay time between the rainfall and the seismicity rate change (6 months; Table 3) and the lack of depth dependence in the time when the seismicity rate increases. Their favored model is one where the increased surface load due to the summer monsoon suppresses the thrust earthquakes in the summer, while the seismicity rate increases in the fall and winter after the summer rainfall has drained to lower elevations.

If thermoelastic strains are the cause of the annual seismicity variations in California and Nevada, then there should be some depth dependence in the seismicity rates throughout the year. As is clear from Figure 6, the greatest thermoelastic strains, and therefore the greatest annual stress changes, take place closest to the Earth’s surface. Therefore, the annual variation in seismicity rate should be greatest for shallow depth earthquakes and should decrease with depth into the crust. Figure 7 gives the distribution of event depths in the 1990-2007 M≥2.0 earthquake catalog used in this analysis (Figure
1), and it also shows the distribution with depth of the change in number of events between the second half of the year and the first half of the year for this data set. We note that the depths of many of the events in the catalog are uncertain by probably several kilometers. For example, both distributions in Figure 7 show large peaks for event depths between 4 km and 6 km. This peak likely arises because the starting depth used for locating many earthquakes in the region is 5 km, and if the depth could not be well constrained by the arrival-time data for an earthquake, the location algorithm may not have perturbed the depth from its starting value. Keeping this caveat in mind, the bottom histogram in Figure 7 shows that there is an unusually large difference in the number of earthquakes during each half of the year for events with focal depths less than 2 km. This difference is disproportionately greater than the differences for earthquakes at deeper depths, and therefore it provides some evidence that the shallowest seismicity exhibits the greatest annual variation in California and Nevada. This observation is consistent with the assertion that changing thermoelastic strains caused by annual temperature variations in California and Nevada are generating the annual changes in seismicity rates that are observed.

Discussion

It has long been suspected that there are seasonal variations in the rate of earthquake activity in California and other locations. This may be responsible in part for the popular speculations about the existence of “earthquake weather”. Richter (1958) noted this when he wrote, “In California, minor earthquake activity often shows a perceptible increase toward the end of the year, about the beginning of the rainy season, when large air masses are being shifted and the load on the earth’s surface is changing.” Richter’s perceptive comment indicates that he observed an increase in the rate of earthquake activity around October (the beginning of the rainy season), and like Gao et al. (2000) he speculated that this rate increase was due to seasonal variations in atmospheric pressure. Most of the data used to document the seasonal seismicity variations in California and Nevada in this study include earthquakes that took place after Richter (1958) published his book. Thus, annual seismicity rate variations in California and Nevada seem to be a long-term characteristic of the earthquake activity.

Of the causes of seasonal seismicity rate variations that have been proposed so far, the thermoelastic strain model is the one that best fits the data for California and Nevada. Nevertheless, it is quite possible that the seasonal seismicity rate variations seen in other areas may be due to different causes. For example, the monsoon in India and Nepal (Bollinger et al., 2006) give much greater amounts of summer rainfall than the total amount of rainfall that is experienced annually in California. Also, significant snow accumulates on the volcanic peaks studied by Christiansen et al. (2005) in the western U.S., and this snow load might affect the local rate of seismicity beneath those volcanoes. Thus, groundwater recharge and rain/snow loads (e.g., Saar and Manga, 2003; Christiansen et al., 2005; Bollinger et al., 2006; Hainzl et al., 2006) may be significant factors in controlling variations in the rate of seismicity in some areas even if they are not factors that best explain the annual seismicity rate variations in California and Nevada.
Our estimated stress level that may be generated by thermoelastic strain, and the other information in Tables 2 and 3, suggest that long-term stress perturbations of 10 kPa or less may be sufficient to trigger or suppress some earthquake activity. These values are comparable to the thresholds that have been assumed for tidal and various other dynamic and static triggering mechanisms (e.g., Hardebeck et al., 1998; Cochran et al., 2004; Fischer et al., 2008).

Beeler and Lockner (2003) analyzed triggered stick-slip events in laboratory experiments with loadings consisting of a constant background rate plus a superimposed small sinusoidal component with frequency in the range $10^{-4}$-$10^{-1}$ Hz. They found that the number of triggered events depends on both the amplitude and frequency of the sinusoidal stress component. The observations discussed in this paper extend the range of loading periods by many orders of magnitudes. The results have a simple explanation in terms of delayed nucleation processes, since longer periods allow more time for nucleation processes to mature and may hence trigger instabilities at lower loading amplitudes. In this explanation the lowest threshold for triggering is associated with static stress changes. Dieterich (1987) concluded that there is no dependency between the loading period and triggering involving nucleation processes in rate- and state-dependent friction. However, the bulk of seismicity occurs away from large faults with well-developed sliding surfaces, and may be associated with static fatigue (e.g., Scholz, 1990) and various other nucleation processes of rock fracturing (e.g., Ben-Zion, 2008) rather than frictional sliding.

Further clarifications of the relations between the amplitude and period of stress changes and resulting number of triggered events can improve the understanding of rock rheology and provide useful information for earthquake forecasting. The thermoelastic strain and other seasonal loadings may also be relevant for some of the periodicities that are observed for non-volcanic tremors (e.g., Miller et al., 2002) and other triggered geophysical phenomena.

Data and Resources

The seismic data used in this analysis was obtained from the Advanced National Seismic System (ANSS) web site. The declustered earthquake catalogs can be obtained by request from the first author.

Acknowledgments

We thank Alan Kafka for discussions concerning the statistical analyses done in the paper. The manuscript benefited from comments by two anonymous referees and associate editor Lorraine Wolf.
References


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Table 1a. Earthquake Statistics for California and Nevada (All Mainshocks)

<table>
<thead>
<tr>
<th>Time Period</th>
<th>Magnitude Range</th>
<th>Total # EQs</th>
<th># EQs 1st Half of Year</th>
<th># EQs 2nd Half of Year</th>
<th>% Difference</th>
<th>Significance of Difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>1990-2007</td>
<td>M≥2</td>
<td>16037</td>
<td>7903</td>
<td>8134</td>
<td>2.92%</td>
<td>99.48%</td>
</tr>
<tr>
<td>1990-2007</td>
<td>M≥3</td>
<td>3595</td>
<td>1746</td>
<td>1849</td>
<td>5.90%</td>
<td>99.17%</td>
</tr>
<tr>
<td>1932-2007</td>
<td>M≥4</td>
<td>1224</td>
<td>599</td>
<td>674</td>
<td>12.52%</td>
<td>99.81%</td>
</tr>
<tr>
<td>1932-2007</td>
<td>M≥5</td>
<td>196</td>
<td>91</td>
<td>108</td>
<td>18.68%</td>
<td>94.91%</td>
</tr>
<tr>
<td>1932-2007</td>
<td>M≥6</td>
<td>34</td>
<td>18</td>
<td>16</td>
<td>-11.11%</td>
<td>--</td>
</tr>
</tbody>
</table>

Table 1b. Earthquake Statistics for California and Nevada (Without Coso and Long Valley Mainshocks)

<table>
<thead>
<tr>
<th>Time Period</th>
<th>Magnitude Range</th>
<th>Total # EQs</th>
<th># EQs 1st Half of Year</th>
<th># EQs 2nd Half of Year</th>
<th>% Difference</th>
<th>Significance of Difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>1990-2007</td>
<td>M≥2</td>
<td>14494</td>
<td>7144</td>
<td>7350</td>
<td>2.88%</td>
<td>99.19%</td>
</tr>
<tr>
<td>1990-2007</td>
<td>M≥3</td>
<td>3299</td>
<td>1607</td>
<td>1692</td>
<td>5.29%</td>
<td>98.06%</td>
</tr>
<tr>
<td>1932-2007</td>
<td>M≥4</td>
<td>1139</td>
<td>532</td>
<td>607</td>
<td>14.10%</td>
<td>99.88%</td>
</tr>
<tr>
<td>1932-2007</td>
<td>M≥5</td>
<td>177</td>
<td>83</td>
<td>94</td>
<td>10.84%</td>
<td>87.17%</td>
</tr>
<tr>
<td>1932-2007</td>
<td>M≥6</td>
<td>29</td>
<td>17</td>
<td>12</td>
<td>-29.41%</td>
<td>--</td>
</tr>
</tbody>
</table>
Table 2. Estimates of Annually Varying Surface Loads

<table>
<thead>
<tr>
<th>Load</th>
<th>Area</th>
<th>Earthquake Depths</th>
<th>Peak Load Stress</th>
<th>Time of Peak Stress</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air Pressure</td>
<td>California</td>
<td>1-15 km (a)</td>
<td>2 kPa</td>
<td>Dec.-Jan.</td>
<td>Gao et al. (2000)</td>
</tr>
<tr>
<td>Surface Thermelastic Strains</td>
<td>California</td>
<td>1-10 km</td>
<td>7 kPa</td>
<td>August</td>
<td>This Study</td>
</tr>
<tr>
<td>Snow</td>
<td>Western U.S. Volcanoes</td>
<td>2-8 km</td>
<td>6 kPa</td>
<td>February</td>
<td>Christiansen et al. (2005)</td>
</tr>
<tr>
<td>Annual Earth Tides</td>
<td>Western U.S. Volcanoes</td>
<td>2-8 km</td>
<td>.2 kPa</td>
<td>(b)</td>
<td>Christiansen et al. (2005)</td>
</tr>
<tr>
<td>Monsoon Surface Water</td>
<td>Nepal Himalaya</td>
<td>5 km</td>
<td>6 kPa</td>
<td>July</td>
<td>Bollinger et al. (2006)</td>
</tr>
</tbody>
</table>

(a) Estimated depth range of seismicity.
(b) Not reported in paper.

Table 3. Estimates of Pore-Fluid Pressure Changes and Earthquakes

<table>
<thead>
<tr>
<th>Area</th>
<th>Earthquake Depths</th>
<th>Peak Pore-Fluid Pressure Changes</th>
<th>Seismicity Lag</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mt. St. Helens</td>
<td>3 km</td>
<td>64 kPa</td>
<td>1 month</td>
<td>Christiansen et al. (2005)</td>
</tr>
<tr>
<td>Long Valley – Eastern South Moat</td>
<td>8 km</td>
<td>2 kPa</td>
<td>5 month</td>
<td>Christiansen et al. (2005)</td>
</tr>
<tr>
<td>Mammoth Mt.</td>
<td>6 km</td>
<td>123 kPa</td>
<td>1 month</td>
<td>Christiansen et al. (2005)</td>
</tr>
<tr>
<td>Yellowstone Lake</td>
<td>5 km</td>
<td>12 kPa</td>
<td>1 month</td>
<td>Christiansen et al. (2005)</td>
</tr>
<tr>
<td>Hebgen Lake/Madison Valley</td>
<td>2 km</td>
<td>36 kPa</td>
<td>1 month</td>
<td>Christiansen et al. (2005)</td>
</tr>
<tr>
<td>Mt. Hood</td>
<td>4.5 km</td>
<td>10 kPa</td>
<td>5 month</td>
<td>Saar and Manga (2003)</td>
</tr>
<tr>
<td>Mt. Ogden</td>
<td>(a)</td>
<td>(a)</td>
<td>1 month</td>
<td>Wolf et al. (1997)</td>
</tr>
<tr>
<td>Nepal Himalaya</td>
<td>10 km</td>
<td>10 kPa</td>
<td>6 month</td>
<td>Bollinger et al. (2006)</td>
</tr>
<tr>
<td>Mt. Hochstaufen</td>
<td>4 km</td>
<td>1.3 kPa</td>
<td>.3 month</td>
<td>Hainzl et al. (2006)</td>
</tr>
</tbody>
</table>

(a) Not reported in paper.
Figure Captions

Figure 1. \(M \geq 2.0\) seismicity from the Advanced National Seismic System (ANSS) catalog for the region from \(32^\circ N\) to \(40^\circ N\), \(-114.0^\circ E\) to \(-126.2^\circ E\) for the time period from 1990 through 2007. Foreshocks, aftershocks and triggered events (see text) have been removed from the catalog.

Figure 2. \(M \geq 4.0\) seismicity from the Advanced National Seismic System (ANSS) catalog for the region from \(32^\circ N\) to \(40^\circ N\), \(-114.0^\circ E\) to \(-126.2^\circ E\) for the time period from 1932 through 2007. Foreshocks, aftershocks and triggered events (see text) have been removed from the catalog.

Figure 3. (top left) Cumulative number of earthquakes with \(M \geq 2.0\) versus day of year for California and Nevada for the time period 1990-2007 (Figure 1 data set). (top right) The data to the left after smoothing with a 90-day moving average filter. (bottom left) Cumulative number of earthquakes with \(M \geq 3.0\) versus day of year for California and Nevada for the time period 1990-2007 (Figure 1 data set). (bottom right) The data to the left after smoothing with a 90-day moving average filter. The dashed lines on the right-hand plots show the average annual seismicity rates.

Figure 4. (top left) Cumulative number of earthquakes with \(M \geq 4.0\) versus day of year for California and Nevada for the time period 1932-2007 (Figure 2 data set). (top right) The data to the left after smoothing with a 90-day moving average filter. (bottom left) Cumulative number of earthquakes with \(M \geq 5.0\) versus day of year for California and Nevada for the time period 1932-2007 (Figure 2 data set). (bottom right) The data to the left after smoothing with a 90-day moving average filter. The dashed lines on the right-hand plots show the average annual seismicity rates.

Figure 5. The composite model for thermoelastic strain of the crust. The model has two layers: a thin unconsolidated surficial layer overlying a rock half-space. The source of the strain is a spatial and temporal variation in temperature, as indicated in the equation above. In this equation, \(T\) is the temperature as a function of horizontal position \(x\) and time \(t\), \(T_\omega\) is the amplitude of the temperature at frequency \(\omega\), \(k\) is the spatial wavelength of the temperature, and \(\phi\) is the phase shift of the temperature. (From Ben-Zion and Leary, 1986).

Figure 6. Variations of thermoelastic strain with depth for a model with a \(1.2\) m thick unconsolidated surface layer as calculated using Equation (1). Each plot was generated with a different value of \(\lambda\), which is the spatial wavelength of the surface temperature variation. The average daily high and low temperature readings from Twenty-Nine Palms, California were used as the input temperature time series.

Figure 7. (top) Distribution of the focal depths of the earthquakes with \(M \geq 2.0\) for California and Nevada for the time period 1990-2007 after foreshocks, aftershocks and triggered events have been removed from the catalog (Figure 1). (bottom) Distribution
with depth of the differences between the number of events in the second half of the year and the number of events in the first half of the year for the $M \geq 2.0$ data set.
Figure 1.
Figure 2.
Figure 3.
Figure 4.
Figure 5.
Figure 6.
Figure 7.