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Stress transferred by the 1995 M_w=6.9 Kobe, Japan, shock: Effect on aftershocks and future earthquake probabilities

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Abstract. The Kobe earthquake struck at the edge of the densely populated Osaka-Kyoto corridor in southwest Japan. We investigate how the earthquake transferred stress to nearby faults, altering their proximity to failure and thus changing earthquake probabilities. We find that relative to the pre-Kobe seismicity, Kobe aftershocks were concentrated in regions of calculated Coulomb stress increase and less

common in regions of stress decrease. We quantify this relationship by forming the spatial correlation between the seismicity rate change and the Coulomb stress change. The correlation is significant for stress changes greater than 0.2-1.0 bars (0.02-0.1 MPa), and the nonlinear dependence of seismicity rate change on stress change is compatible with a state- and rate- dependent formulation for earthquake occurrence. We extend this analysis to future mainshocks by resolving the stress changes on major faults within 100 km of Kobe, and calculating the change in probability caused by these stress changes. Transient effects of the stress changes are incorporated by the state-dependent constitutive relation, which amplifies the permanent stress changes during the aftershock period. Earthquake probability framed in this manner is highly time-dependent, much more so than is assumed in current practice. Because the probabilities depend on several poorly-known parameters of the major faults, we estimate uncertainties with Monte Carlo simulations. This enables us to include uncertainties on the elapsed time since the last earthquake, the repeat time and its variability, and the period of aftershock decay. We estimate that a calculated 3-bar (0.3-MPa) stress increase on the eastern section of the Arima-Takatsuki tectonic line (ATTL) near Kyoto causes two- to seven-fold increase in the 30-year probability of a subsequent large earthquake near Kyoto; a 2-bar (0.2-MPa) stress decrease on the western section of the ATTL results in a reduction in probability by a factor of forty to two-thousand. The probability of a M=6.9 earthquake within 50 km of Osaka during the next 10 years is estimated to have risen from 5-6% before the Kobe earthquake to 7-11% afterwards; during the next 30 years, it is estimated to have risen from 14-16% before Kobe to 16-22%.

Introduction

The 17 January 1995 Kobe, Japan, shock was the most destructive earthquake to strike urbanized Japan since 1946. The Kobe area is dominated by the young Philippine Sea plate subducting beneath the Eurasian plate at 40 mm/yr, generating great subduction earthquakes with repeat times of ~100 years (Figure 1a). In addition, the inner zone of central to southwest intraplate Japan is distinguished by the most dense distribution of active faults in Japan, composed principally of NE-striking right-lateral and NW-striking left-lateral faults, and N-striking thrust faults [*Research Group for Active Faults in Japan*, 1991] (Figure 1b). Most of these faults slip at 0.1 to 1.0 mm/yr, except for the Median Tectonic Line (1.0 to 9.0 mm/yr), generating shallow destructive earthquakes with repeat times _1,000 years. More destructive intraplate earthquakes have been documented in the Kyoto-Osaka corridor than in any other area of Japan [*Usami*, 1987], due in part to the density of active faults and the region's long recorded history.

We seek to answer two questions raised by the Kobe earthquake: what is the earthquake hazard in southwest Japan, and how did the 1995 earthquake affect it? The effects of the Kobe earthquake demonstrate the need for earthquake hazard assessment in populated areas worldwide, and underscore the imperative to prepare for the next large shock in Japan. But the brevity of the instrumental seismic record, relative to the thousand-year repeat times for large intraplate earthquakes, limits our ability to estimate earthquake probabilities.

Recent work on how earthquakes transfer stress and trigger subsequent shocks encourage us to evaluate the earthquake potential both before and after the Kobe event. Several large earthquakes in California and Nevada, such as the 1954 Rainbow Mountain-Fairview Park-Dixie Valley earthquakes [*Caskey and Wesnousky*, 1997; *Hodgkinson et al.*, 1996], 1979 Homestead Valley [*Stein and Lisowski*, 1983], 1984 Morgan Hill [*Oppenheimer et al.*, 1988], 1992 Landers [*Gross and Kisslinger*, 1997; *Hardebeck et al.*, 1997; *Harris and Simpson*, 1992; *King et al.*, 1994; *Stein et al.*, 1992], 1989 Loma Prieta [*Gross and*

Bürgmann, 1997; *Reasenberg and Simpson*, 1992] and 1994 Northridge [*Hardebeck et al.*, 1997; *Stein et al.*, 1994] shocks, reveal a good correlation between the calculated Coulomb stress change and the distribution of aftershocks. Sequences of moderate to large earthquakes were interpreted to result from stress triggering as well. *Stein et al*. [1992] and *King et al*. [1994] described several moderate earthquakes during the 17 years before the 1992 M=7.3 Landers, California, event that transferred stress to the future Landers rupture zone. *Stein et al*. [1994] argued that the 1971 San Fernando earthquake raised stress on the nearby 1994 Northridge rupture plane. *Deng and Sykes* [1996] suggested that the great 1812 Wrightwood earthquake on the San Andreas fault may have triggered a large earthquake near Santa Barbara 13 days later. *Harris et al*. [1995] found that in the 1.5-yr period following M_5 earthquakes in southern California, any subsequent nearby M_5 earthquake almost always ruptured a fault that had been loaded towards failure by the previous earthquakes. In an extensional tectonic setting in southern Italy, *Nostro et al*. [1997] found that most of the large historical earthquakes occurred in areas of static stress increase. *Simpson and Reasenberg* [1994], *Harris and Simpson* [1996] and *Jaumé and Sykes* [1996] found a lack of earthquakes in areas of reduced Coulomb stress ~ the so-called "stress shadows"~ produced by the great 1857 and 1906 shocks along the San Andreas fault. Other studies furnish evidence that the static stress change may affect long-term earthquake probabilities. *Stein et al*. [1997] calculated earthquake probability gains associated with M_6.7 earthquakes that occurred along the Northern Anatolian fault between 1939 and 1992, using state- and rate-dependent friction constitutive relations from *Dieterich* [1994] and *Dieterich and Kilgore* [1996].

The Kobe earthquake affords an excellent opportunity to examine the relationship between the static stress change and seismicity because the earthquake slip is well constrained by teleseismic, strong motion, and geodetic data; regional seismicity was recorded continuously during the 20-year period before the earthquake; and the historical catalog is complete to M~7.5 for the past 1,000 years. We find that the stress change strongly affected the spatial distribution of the Kobe aftershocks. The calculated coseismic stress change may have increased the earthquake probability on the Arima-Takatsuki Tectonic Line, the fault closest to the Kobe rupture, by nine-fold during the next decade. In contrast, for most faults in the region the estimated changes in the 10-year earthquake probability on the order of 10-30%.

Stress Changes caused by Kobe Earthquake

Rupture Model

We used the variable slip model of *Wald* [1996] because it is derived from joint inversion of near-source ground motions, teleseismic body waveforms, and geodetic displacements (GPS and leveling). We smoothed Wald's 3.3 • 2.5 km slip patches to 5 • 5 km patches to speed calculations, but retained his two planar fault segments, the Nojima (dipping 80° SE) and Rokko (dipping 85° NE) faults. The slip model has a hinge or offset near the epicenter. Although the modeled rupture is 60 km long, the slip is concentrated at shallow depth on the Nojima fault, where prominent surface ruptures were produced (Figure 2).

Coulomb Failure Stress Change

The static Coulomb failure stress change D s_f we calculate is

D s_f = D *t* - *m D s¢* (1)

where D *t* is the shear stress change on a given failure plane (positive in the direction of fault slip), D *s¢* is the effective normal stress change (positive in compression) and *m* is the coefficient of friction. *m D s¢* Å *m*

¢ D s, where D *s* is the normal stress change, and *m¢* is the apparent coefficient of friction, which includes the effects of pore-pressure changes. Here we set *m¢*=0.4 to minimize the uncertainty in *m¢* (0.0-0.75) to ±25% as discussed by *King et al*. [1994]. Positive values of the Coulomb stress change promote failure; negative values inhibit failure. We compute the Coulomb stress change in an elastic halfspace [*Okada*, 1992] by assuming a shear modulus of $3.2 \cdot 10^{11}$ dyne-cm^{-2} (3.2×10^{10} Nm^{-2}) and a Poisson's ratio of 0.25. We make two complementary calculations. First, we resolve stress changes on planar surfaces of the known active faults. Second, we calculate the stress change at every point in the medium on planes oriented to maximize the Coulomb stress change, following *King et al*. [1994] and *Stein et al*. [1992].

Stress Changes on the Active Faults

We modeled the regional active faults as either vertical strike-slip or 35°-dipping thrust faults, extending from the surface to a depth of 20 km. The dip of thrust faults in southwest Japan are not well known; the few data show surface dips of 20-70°. Focal mechanisms of thrust events in the Kobe area are consistent with an average dip of 35°, although it is not possible to distinguish between nodal and fault planes, leaving a large uncertainty. Here we use 35° because, in the absence of better geologic or seismic data, it is close to the theoretical optimum dip of 34.1° for a horizontal axis of maximum compression and *m¢*=0.4. Stress changes were calculated on the model faults every 10 km along strike (Figure 3a). Because the depths of the aftershocks are not well constrained and because the depth-dependence of the Coulomb stress change is small at distances more than one fault depth from the source rupture (Figure 3b), we will make all further calculations at a representative depth of 10 km. Better hypocentral precision would have permitted us to examine several depths. Figure 3a shows the result of the calculation.

The Coulomb failure stress on most faults near Kyoto and along the west bank of Lake Biwa is calculated to have risen, regardless of the fault type. In particular, stress on the eastern Arima-Takatsuki Tectonic Line (ATTL) is calculated to have increased by up to 3 bars (0.3 MPa). Stress on the Yamasaki fault and part of Median Tectonic Line (MTL) is calculated to have grown by up to 0.2 bar (0.02 MPa). In contrast, stress on the western portion of ATTL and central section of MTL is calculated to have dropped by up to 2 bars (0.2 MPa). Stress on the many thrust faults under the Osaka and Nara region is calculated to have decreased by as much as 0.3 bar (0.03 MPa). Later we will infer from these calculated stress changes the earthquake probability changes, but first we seek to validate our stress calculations by comparing them to the observed changes in regional seismicity.

Stress Changes throughout the Crust

In order to examine the relationship between aftershocks and stress changes, we assume that small faults are distributed regionally with all orientations, and that the faults optimally oriented for failure are the ones most likely to slip in the aftershocks. To calculate the stress changes on these optimally oriented faults, we assume a regional stress field consisting of uniaxial E-W compression, consistent with the pre-Kobe focal mechanisms in the region compiled by *Tsukahara and Kobayashi* [1991] (Figure 4), and the focal mechanisms of the larger Kobe aftershocks [*Katao et al*., 1997]. We select a regional stress magnitude of 100 bars (10 MPa). The stress change on optimally oriented planes is sensitive to the orientation of the regional stress, but insensitive to its magnitude. *King et al*. [1994] showed that as long as the regional stress is larger than the earthquake stress drop, the stress change on optimally oriented planes is little affected; in contrast, a regional stress magnitude less than the earthquake stress drop is highly unlikely. The largest Coulomb stress change on optimally oriented faults between the surface and 20 km depth is shown for strike-slip (Figure 5a) and thrust faults (Figure 5b). By this we mean that we have calculated the stress change at 5 km increments between 0 and 20 km depth, and select the most positive stress change found at any depth. Thus if an aftershocks most likely occur at the locations, depths, and orientations on which the Coulomb stress change is most positive, then there will be a positive correlation between sites of aftershocks and stress increase.

The stress change on optimally oriented faults is positive along the rupture (Figure 5), despite the fact that on average, the stress dropped on the fault plane; these regions are excluded from our corelations. The first reason for the stress increase along the rupture plane is that the optimal planes or the receiver faults are rotated with respect to the rupture plane, because the earthquake stress change is nearly as large as the assumed regional stress close to the rupture plane. In addition, the fault slip is not smooth (Figure 2), and the slip irregularities cause local peaks in stress. If we included these observations, the correlation between stress change and seismicity rate change we pursue next would be influenced by observations in the region when the largest modeling errors are expected. Thus, we consider sites with stress changes less than ±8 bars (±0.8 MPa), which excludes regions within 7 km of the rupture from the correlation (see contour in Figure 9a). Had we included the sites along the rupture plane, the correlation would have been higher.

Calculated stress increases of more than 0.5 bar (0.05 MPa) are located near the ends of the Nojima fault (the rupture extending southwest of the epicenter), and on the northwest side of the Nojima fault (Figure 5). The Kyoto region is in a zone of increased stress. The Osaka-Nara region is in a zone brought farther from failure by up to 0.3 bar (0.03 MPa). East (near Osaka and Nara) and west of the 1995 source, stress is calculated to have diminished on both strike-slip and thrust faults. North and south of the source fault, stress on strike-slip faults drops and on thrust faults rises. *Hashimoto* [1996] and *Hashimoto et al*. [1997] calculated the Coulomb stress changes throughout the crust on several specified strike-slip and thrust planes, using a simpler source function than the model we use here. Their results are similar to ours except close to the fault where the results are sensitive to the detailed slip function.

Throughout most of the region, focal mechanisms and observed surface displacements along the active faults are dominated by strike-slip mechanisms [*Research Group for Active Faults in Japan*, 1991]. Exceptions include the Osaka-Kyoto-Lake Biwa corridor and the Wakayama region, both of which show thrust and strike-slip mechanisms. To evaluate the relationship between the stress change and seismicity, we thus consider the stress changes on strike-slip faults everywhere except within the two dashed zones of Figure 4, where we use the larger of the stress changes calculated on strike-slip and thrust faults.

Effect of Stress Change on Seismicity

The post-Kobe M_1 seismicity is shown with the Coulomb stress changes resolved on major faults, and calculated throughout the region on optimally oriented faults, in Figure 6. Detection of M<2 events by the JMA network is incomplete in parts of this region, so the activity in some areas shown is under-reported. The majority of the aftershocks (63%) are found where the calculated Coulomb stress change for optimally oriented faults increased by >0.1 bar (0.01 MPa), such as northeast of the main shock rupture in an area surrounding Kyoto and Lake Biwa.

In other areas, however, the relationship between the calculated stress change and the post-Kobe seismicity is not as clear. North of the main shock, Coulomb stress increased on thrust faults but decreased on strike-slip faults (Figure 5). South of the main shock, many aftershocks occurred in an area of decreased stress on optimally oriented faults. Because both of these areas experienced a high rate of seismicity before the Kobe earthquake, a plot of aftershocks alone cannot reveal whether the rate of seismicity has in fact increased or decreased. The Wakayama area, for example, has been the site of extensive earthquake swarm activity composed of thrust fault events at shallow depth [*Earthquake Research Institute University of Tokyo*, 1995; *Earthquake Research Institute University of Tokyo*, 1996a; *Earthquake Research Institute University of Tokyo*, 1996b; *Watanabe and Maeda*, 1990]. Thus, to quantify the stress-seismicity correspondence, we must take into account the earthquake activity before the Kobe earthquake.

Minimum Magnitude of Completeness

To quantify the relationship between the Coulomb stress change and the seismicity rate change, one must use a catalog with the largest number of earthquakes and the longest possible observation period for which the detection level was constant in time. Strictly, such a condition does not require that the catalog is complete, but rather than the detection level is constant. In practice, the network was changed in 1994, altering detection levels at nearly all locations. We thus estimated the minimum magnitude of completeness within the area shown in Figure 6 during successive 1-year periods, starting 10 years before the Kobe earthquake (Figure 7). This was found by noting the minimum magnitude for which a power law relationship between the number of earthquakes and magnitude apparently holds. The minimum magnitude of completeness decreased as the JMA seismic network grew, and particularly when the network was upgraded in October 1994. Between 1985 and 1993, the minimum magnitude of complete coverage decreased from M=2.8 to M=2.4. After the network change in 1994, the catalog is complete for M_2.2. Our catalog ends June 30, 1996, 1.46 years after the main shock. Thus, the catalog provides complete coverage at the M_2.2 level between October 1, 1994 and June 30, 1996 (a 1.7-year period, including 4593 mostly post-mainshock events) and at the M_2.6 level between January 17, 1987 and June 30, 1996 (a 9.5-year period, including 3463 events, equally split between pre- and post-Kobe shocks; Figure 8a and 8b). We used both the M_2.6 and M_2.2 data in our measurements of seismicity rate change. The M_2.2 catalog provided more sensitivity to small rate increases, while the M_2.6 catalog was better able to resolve rate decreases owing to its longer background period. The overall results were similar with both catalogs. We present the results obtained with the M_2.6 catalog in Figure 8c.

Seismicity Rate Change

We represent the coseismic change in seismicity rate by the ratio of the average postseismic rate R to the average preseismic rate r. We calculated R/r for events within overlapping 10-km-square cells with centers located on a grid with 1-km spacing, smoothed with a 2D Gaussian filter (Figure 8c), following *Reasenberg and Simpson* [1992]. After the Kobe shock, the seismicity rate increased within 10-15 km from the mainshock rupture, near Kyoto and Wakayama, on parts of the Yamasaki fault, and beyond the southwest end of the mainshock rupture. The seismicity rate decreased southwest of Wakayama and northwest of Tokushima.

Spatial Correlation between the Stress and Seismicity Rate Changes

To test the spatial correlation between the calculated Coulomb stress changes and the pattern of seismicity rate change, we smoothed the stress change with the same spatial filter used to calculate R/r (Figure 9a), and sampled values of the stress change at every point on the grid for which the seismicity rate change could be measured (Figure 9b). Values of $R/r < 1$ were made negative. We then multiplied corresponding values of each field (R/r x Coulomb stress change) at each grid point. The result, which is positive where the stress and seismicity both changed in the same sense (both increased or both decreased) and negative in areas of discordant change in stress and seismicity rate, is presented in Figure 9c. Areas for which a calculation of

R/*r* was not possible, due to an insufficient number of pre- or post-Kobe earthquakes, are light gray in Figure 9b and 9c. Areas of agreement in Figure 9c dominate, with 63% of the area for which the calculation was possible experiencing concordant stress and seismicity changes (here, an agreement of 50% would be expected by chance).

Comparison of the calculated values of *R*/*r* and Coulomb stress change at each point (Figure 10) reveals a nonlinear dependence of seismicity rate change on stress change. Most points lie in the upper right quadrant of the plot and correspond to increases in both seismicity rate and Coulomb stress. In this comparison, we excluded regions where the calculated stress change exceeded 8 bars (0.8 MPa). This effectively eliminates the area within ~7.5 km from the source fault, which may include large errors due to uncertainty in the 5 • 5 km-smoothed slip model.

To estimate the significance of the correlation between stress changes and seismicity rate changes, we calculated the *c2* statistic for various four-fold tables (see *Reasenberg and Simpson* [1992]). A four-fold table compares the number of points falling within each quadrant in Figure 10, and enables us to assess the significance of the clustering of points in the upper-right and lower-left quadrants of the plots, and to test for the dependence of the seismicity rate on the calculated stress changes. By focusing on the counts in each quadrant, we use only the sign of the stress and seismicity rate changes, not their amplitudes; all points are given equal weight. (Later, we will model the functional relationship between *R*/*r* and the stress change in Figure 10.) To correct for the redundancy in sampling introduced by spatial smoothing, the numbers of points in the quadrants of Figure 10 was reduced by a factor of 78.5 (the weighted number of grid points covered by the smoothing kernel) to yield an effective number of independent observations. The results are summarized in Table 1.

Using all data shown in Figure 10 (in other words, all regions outside of the ±8 bar contour in Figure 9a), we are able to reject the null hypothesis that the stress and seismicity rate changes are independent at the 98% confidence level. When we restrict our tests to points with lower absolute levels of stress change (generally corresponding to areas farther from the main shock), the significance of the correlation decreases, but remains significant (96% confidence) down to absolute stress change levels of 1.0 bar (0.1 MPa; see contour in Figure 9a) (Table 1). A similar comparison using the M_2.2 catalog (3 months pre-Kobe with respect to 17.4 months post-Kobe) indicated a significant correlation (96% confidence) for absolute levels of stress change _0.2 bar (0.02 MPa, which excludes all red and purple regions in Figure 9a). We also carried out similar four-fold tests using stress changes on optimally-oriented strike-slip faults only (Figure 5a), since strike-slip faults and focal mechanisms dominate the region (Figure 4). In these tests, the confidence levels for both the M_2.6 and M_2.2 correlations were the same as the ones listed in Table 1. These results are comparable to other studies in which a minimum level of stress change of ~0.1 bar (0.01 MPa) was found to correlate with seismicity changes associated with the Loma Prieta [*Reasenberg and Simpson*, 1992], Landers [*Stein et al,* 1992; *Hardebeck et al*, 1997] and Northridge [*Harris et al*, 1995] earthquakes.

We conducted a second set of tests to compare the ability of our our stress transfer model to explain the seismicity rate changes with a control or alternative model. In the or contro, the stress change is positive and proportional to 1/(*d*+1), where *d* is the distance (in km) from the nearest point on the Kobe dislocation. This model simulates the possible effect of shaking (or dynamic stress) on the seismicity rate in the region [e.g. *Boore et al.,* 1997]. Because this stress field is everywhere positive, it is impossible to construct a fourfold table and carry out a *c2* test, as we did for the static stresses. Instead, we compare the two models in terms of the percentage of observations for which the seismicity rate change and the modeled stress change agree. For the control, this is simply the percentage of points at which seismicity rate increased. Both models explain approximately 60% of the seismicity rate observations in areas in which the absolute stress change was 8 bars (0.8 MPa) or less (Table 1). As the distance from the fault grows, the static stress model fits the seismicity rate change observations better. At distances corresponding to absolute stress changes of approximately 0.5 bar (0.05 MPa) or less, the static stress model agrees with the areas of seismicity rate decrease in its relaxed-stress lobes, while the dynamic stress model always disagrees with areas of seismicity rate decrease. These observations suggest that both shaking and static stress changes affect the seismicity rate, with the shaking contributing at close range and stress dominating farther from the Kobe earthquake. We find this despite the fact that the dynamic stresses are much larger than the static stresses at large distances, whereas their magnitudes are comparable close to the fault plane.

Differences in significance ascribed to the correlation measured by the c_2 and binomial tests (Table 1) reflect the fact that the c_2 test discounts the (coincidentally) preferential location of seismicity rate observations in areas of increased stress, while the binomial test does not. The c_2 test is the preferred one for assessing the overall significance of the correlation, while the binomial test is useful for comparing models.

In summary, our tests show a significant correlation between the sign of the calculated Coulomb stress changes and the sense (increase or decrease) of the seismicity rate changes. This correlation was primarily detected in areas in which Coulomb stress and seismicity rate both increased. Having carried out this exercise to validate our calculations of stress changes in the Kobe earthquake, we will next use the Coulomb stress changes resolved on major active faults (as shown in Figure 3) to calculate probability changes for large earthquakes in the region.

Probability of Future Earthquakes around the Kobe Region

Here we develop a method for estimating earthquake probability that accounts for both permanent and transient effects of the stress transferred by a nearby earthquake, expanding on *Dieterich* [1994], *Dieterich and Kilgore* [1996] and *Stein et al*. [1997].

Probability Models

Two statistical models of earthquake occurrence are commonly used to estimate earthquake probability: a stationary Poisson model and a conditional quasi-periodic model, both of which take into account uncertainties in the mean earthquake repeat time. The conditional probability model is time-dependent. It includes knowledge of the time since the last event and its uncertainty, and may also include the effects of a given stress change. Studies using a conditional probability model have employed Gaussian, lognormal and Weibull probability density functions of recurrence time [e.g. *Hagiwara*, 1974; *Nishenko and Buland*, 1987; *Sykes and Nishenko*, 1984]. Large data sets have been better fit with a Weibull function [e.g. *Sieh et al*., 1989; *Ward*, 1992].

Following *Working Group on California Earthquake Probabilities* [1988], the probability that an earthquake will occur at some time T in the interval (t, $t+Dt$) is

$$P(t \le T \le t + \Delta t) = \int_t^{t+\Delta t} f(t)\,dt \quad (2)$$

where $f(t)$ is the probability density function for the earthquake's recurrence. The probability conditioned on the fact that the earthquake has not occurred prior to t is

$$P(t \le T \le t + \Delta t \mid T > t) = \frac{P(t \le T \le t + \Delta t)}{P(t \le T \le \infty)} \quad . (3)$$

To include the permanent effect of a stress change D s $_f$ we assume that a sudden stress increase (or decrease) linearly shortens (or lengthens) the time until the next earthquake (Figure 12a). Thus the expected recurrence time changes from *T* to *T* ¢ [*Working Group on California Earthquake Probabilities*, 1990], where

$$T' = T - \frac{\Delta \sigma_f}{\dot{\tau}} \quad (4)$$

and $\dot{\tau}$ is the secular rate of stress loading on the fault, and D s $_f$ is positive for an increase in Coulomb stress.

State-dependent Formulation of Stress Change on Earthquake Occurrence

We now incorporate the state-dependent constitutive formulation for earthquake occurrence given by *Dieterich* [1994]. Dieterich argues that several types of widely observed earthquake phenomena, including aftershocks, earthquake clustering, and foreshocks, are perturbations of background seismicity caused by stress changes from a previous shock or set of shocks. The essential nucleation characteristic needed to model such clustering is a nonlinear dependence of the time to instability on stress change (Figure 11a). Seismicity is viewed as a sequence of earthquake nucleation events (squares in Figure 11a and 11b), in which the ‘state’ depends on the slip, slip rate, and elapsed time since the slip occurred. A key consequence of such a formulation for our work is that small changes in stress result in large perturbations of earthquake activity (circles in Figure 11b) at all earthquake magnitudes, followed by a 1/*t* decay to background seismicity rates. Such perturbations are more pronounced when the time to the instability is long (compare the time-change for points A and B in Figure 11a).

Dieterich’s formulation assumes that in the absence of stress perturbations, the seismicity rate is independent of time. The seismicity rate after a sudden stressing event, *R*, is

$$R = \frac{r}{\gamma \dot{\tau}_r} \quad (5)$$

where *r* is the steady-state seismicity rate in a region subjected to a shear stressing rate $\dot{\tau}_r$. The seismicity state variable is *g* with the dependence

$$d\gamma = \frac{1}{A\sigma}\left[dt - \gamma d\tau + \gamma\left(\frac{\tau}{\sigma} - \alpha\right)d\sigma\right] \quad (6)$$

where *A* is a fault constitutive parameter, which in laboratory experiments is found to be 0.005-0.02, *t* is the shear and *s* is the normal stress, *a* is a constitutive parameter governing the dependence of state on the normal stress, and *t* is time.

The effect of a sudden increase in the shear stress *t* (or a sudden decrease in the normal stress *s*) is to decrease *g*, which in turn leads to an increased rate of earthquake production, *R*. The opposite obtains for a sudden shear stress decrease or a normal stress increase. *Dieterich* [1994] derives the seismicity rate as a function of time *t* after a sudden stress step

$$R = \frac{r\dot{\tau}/\dot{\tau}_r}{\left[\frac{\dot{\tau}}{\dot{\tau}_r}\exp\left(\frac{-\Delta\tau}{A\sigma}\right)-1\right]\exp\left[\frac{-t}{t_a}\right]+1}, \dot{\tau}\neq 0 \qquad (7)$$

where $\dot{\tau}$ is the stressing rate before the stress change and t_a, the duration of the transient effect, is equal to As/$\dot{\tau}$. The merit of the state-dependent constitutive formulation is that it is consistent with the spatial and temporal behavior of aftershocks and subsequent main shocks, requisite of our probability calculations. Indeed, equation (7) has the form of Omori's law, as explored by *Dieterich* [1994].

Because the transient effect of the stress change is expressed as a change in expected rate of earthquakes, it is convenient to represent the earthquake occurrence as a non-stationary Poisson process in which the probability of an earthquake occurring in the interval (t, t+D t) is given by

$$P(t,\Delta t) = 1-\exp\left\{-\int_t^{t+\Delta t} R(t)dt\right\} = 1-\exp(-N) \qquad (8)$$

where $R(t)$ is the earthquake rate following the stress step and N is the number of earthquakes expected in the interval. To evaluate (8), we obtain N by integrating the solution for $R(t)$ following a stress step [*Dieterich*, 1994]. For the interval t = 0 to t=Dt, this yields

$$N = r_{perm}\left\{\Delta t + t_a \ln\left[\frac{1+\left[\exp\left(\frac{-\Delta\sigma_f}{A\sigma}\right)-1\right]\exp\left[\frac{-\Delta t}{t_a}\right]}{\exp\left(\frac{-\Delta\sigma_f}{A\sigma}\right)}\right]\right\} \qquad (9)$$

where Ds_f is the calculated Coulomb stress change and r_{perm} is the rate of seismicity associated with the permanent effect of the stress change.

The equation for earthquake occurrence following a stress change (9) is the key element of the probability analysis; it is composed of a permanent effect of the stress change, r_{perm}Dt, and a transient effect (r_{perm} times the second term in the braces). The amplitude of the transient effect is proportional to the ratio of the Coulomb stress change to As, while its decay period is governed by the aftershock duration, ta.

Next we equate the value of the conditional probability to a stationary Poisson probability to solve for r_{perm}.

$$r_{perm} = (-1/\Delta t)\ln(1-P) \qquad (10)$$

where P is the previously obtained conditional probability for the fault segment following the stress step, eqn (8). Figures 12c and 12d show the total probability as a sum of the probability before the stress step and the permanent and transient probability increases due to the main shock. The transient increase is largest immediately after the stress change but rapidly diminishes, so that the permanent increase is reached exponentially with time.

The number of earthquakes expected in the interval (t_0, t_1), as shown in Figure 12, is

$$N = r^{c}_{perm}(t_1 - t_0) + r^{1}_{perm} t_a \ln\left[\frac{1 + \left[\exp\left(\frac{-\Delta\sigma_f}{A\sigma}\right) - 1\right]\exp\left[\frac{-(t_1 - t_E)}{t_a}\right]}{\exp\left(\frac{-\Delta\sigma_f}{A\sigma}\right)}\right] - r^{0}_{perm} t_a \ln\left[\frac{1 + \left[\exp\left(\frac{-\Delta\sigma_f}{A\sigma}\right) - 1\right]\exp\left[\frac{-(t_0 - t_E)}{t_a}\right]}{\exp\left(\frac{-\Delta\sigma_f}{A\sigma}\right)}\right] \quad (11)$$

The permanent component of the earthquake rate change depends on the time interval, so that r^{c}_{perm}, r^{1}_{perm} and r^{0}_{perm} become

$$r^{c}_{perm} = \left[-1/(t_1 - t_0)\right]\ln(1 - P_c)$$

$$r^{1}_{perm} = \left[-1/(t_1 - t_E)\right]\ln(1 - P_1) \quad (12)$$

$$r^{0}_{perm} = \left[-1/(t_0 - t_E)\right]\ln(1 - P_0)$$

where P_c, P_1, and P_0 are the conditional probabilities following the stress step for the intervals (t_0, t_1), (t_E, t_1), and (t_E, t_0), respectively. Here we set t_0 to 1997 to reflect the 2 years that have elapsed without any large earthquake occurring along the active faults (Figure 13).

Parameter Estimation

In order to estimate the aftershock duration, t_a, we use the M_2.6 JMA catalog to project the time when the Kobe aftershock rate will return to the background rate experienced during the 8 years before the main shock. This calculation is sensitive to the area considered, as explored in detail by *Dieterich* [1994]. We used an 80 • 30 km rectangular area aligned parallel to and centered on the Kobe rupture enclosing 90% of the aftershocks, and find t_a = 23 yr (±7.7 yr at 1s uncertainty). In other words, we estimate that ~23 years will elapse before the seismicity rate returns to the pre-1995 rate on the Kobe rupture surface (Figure 14). Following *Dieterich* [1994], we take t_a to be a property of the region independent of mainshock magnitude. While aftershock decay period we estimate for Kobe is longer than that found for California and Alaska aftershock sequences [*Dieterich*, 1994] and the 10.2-yr global average found in the Harvard CMT catalog, which is dominated by subduction events, it may be plausible in view of the relatively long repeat times and low stressing rates associated with Japanese interplate earthquakes.

Since the time advance or delay caused by the stress change is $Ds_f / \dot{\tau}$, we must estimate the secular stressing rate $\dot{\tau}$ on each active fault. Here we model the stressing rate from geological estimates of the fault slip rate and compare this result to independent estimates based on a seismological model of the Kobe main shock and on the geodetic data. We estimate $\dot{\tau}$ from a deep slip model based on the geological slip rate, following the method applied to the North Anatolian fault by *Stein et al*. [1997]. The geological slip rate (Table 2, *col. 3*) is assumed to occur on the fault beneath the seismogenic zone (20-100 km depth) and the corresponding stressing rate at 10 km depth is calculated. We obtain values for ten faults ranging from 0.003 to 0.04 bars/yr (0.0003 to 0.004 MPa/yr) (Table 2, *col. 10*).

We can compare this range to an estimated 0.04 bar/yr (0.004 MPa/yr) stressing rate derived from the Kobe static shear-stress drop (15 bars from the *Wald* [1996] slip model) divided by the estimated earthquake repeat time (400 yr), corresponding to the interval between the Kobe and the previous Keicho-Fushimi event in 1596 [*Awata et al*., 1996; *Kanaori et al*., 1993; *Sangawa et al*., 1996]. The 0.04 bar/yr (0.004 MPa/yr) estimate is larger than the mean rate of 0.012 bar/yr (0.0012 MPa/yr) estimated on the Nojima and Rokko faults from the deep slip model, but it is less than the modeled rate on the MTL, consistent with the higher MTL slip rate (Table 2, *col. 10*). Another check on our modeled stressing rates can be made from the observed rate of shear strain. Geodetic data analyzed by *Hashimoto and Jackson* [1993] indicate 3.3±3.0 mm/yr of right-lateral motion distributed over ~180 km centered on the MTL. Assuming $3.15 \cdot 10^5$ bar ($3.15 \cdot 10^4$ MPa) shear modulus, the shear stressing rate becomes 0.004±0.003 bars/yr (0.0004±0.0003 MPa/yr), about one-third of our modeled rate, but suffering from large uncertainty.

Other parameters we have to determine are *As*, T_{ave}, the coefficient of variation of the recurrence times (*n*), and the elapsed time since the last earthquake. We estimate *As* by fitting the observed dependence of the seismicity rate change (*R*/*r*) on stress change predicted for rate- and state-dependent fault properties [*Dieterich*, 1994]. The seismicity rate for the interval 0 to *Dt* is given by

$$\frac{R}{r} = \frac{1}{\Delta t}\left\{\Delta t + t_a \ln\left[\frac{1 + \left[\exp\left(\frac{-\Delta\sigma_f}{A\sigma}\right) - 1\right]\exp\left[\frac{-\Delta t}{t_a}\right]}{\exp\left(\frac{-\Delta\sigma_f}{A\sigma}\right)}\right]\right\} \quad (13)$$

We use the 1.46-yr period for which we have aftershock data (Figure 10), and set t_a = 23 years (Figure 14). The observations in Figure 10 are best fit by *As* = 0.35. It is important that the model satisfies stress changes <2 bars (0.2 MPa), since the calculated stress changes on the major faults (Figure 3) do not exceed this value. For laboratory values of *A* [*Dieterich*, 1994], *As* = 0.35 yields an effective normal stress of 5-20 bars (0.5-2 MPa). Note that the stressing rate should be equal to As/t_a [Dieterich, 1994]. Our inferred values of *As* and t_a yield a stressing rate of 0.015 bars/yr (0.0015 MPa/yr), in accord with the mean fault stressing rate of 0.012 bars/yr (0.0012 MPa/yr) inferred from fault slip rates (Table 2).

For earthquake repeat times and elapsed times along the major faults, we rely on the historical and paleoseismological records (Figure 15). The most recent faulting along the Yamasaki fault is regarded as the A.D. 868 Harima earthquake from geological evidence [*Okada et al.*, 1987; *Toda et al.*, 1995] and historical documents [*Usami*, 1987]. The Tokushima section of the MTL is inferred to have ruptured in the A.D. 1596 Keicho-Fushimi earthquake (M=7.6) [*Tsutsumi and Okada*, 1996], which probably also ruptured the east section of the ATTL and the Kobe source region [*Sangawa et al.*, 1996]. The Hanaore fault has been locked since the last earthquake occurred there 12 centuries ago [*Togo et al.*, 1994]. The Naruto and Wakayama sections of the MTL both lack historical records and geological surface ruptures, and appear to have been inactive during at least the past 1,000 years.

We seek the probability of earthquakes equivalent to the Kobe event (Mw=6.9), although some active faults in this region have experienced larger earthquakes in the past. The mean recurrence time, T_{ave}, depends on the magnitude of the earthquake in question. In order to determine T_{ave} for active faults, we counted M_7 earthquakes in the historically populated Osaka-Kyoto corridor during the past 1,000 years (Figure 15) and found on average 143 years between events. We then converted this areal average to a mean recurrence time for faults with the potential to produce a Mw_6.9 earthquake. We divide the combined 330 km of major active faults in this corridor by 50 km, the typical length of a Mw=6.9 earthquake [*Wells and Coppersmith*, 1994]. We obtain 6.6 potential ruptures, giving a mean recurrence time for any given fault section of ~944 years. We thus use 1000 years as the mean Mw=6.9 recurrence time T_{ave} for all faults except the MTL, which we assign a much shorter recurrence time corresponding to its faster slip rate. Because of inadequate paleoseismological data, the coefficient of variation *n* cannot be estimated directly, so we let *n* = 0.75. Ward [1994] argued on the basis of numerical experiments that *n* = 0.5+0.5(M_{max} - M), where M is the magnitude of interest and M_{max} is the maximum earthquake expected on the fault. In our case, M=6.9 and the historical catalog suggests that M_{max}™ 7.4 (Figure 15), and thus n ™ 0.75. The Working *Group on California Earthquake Probabilities* [1990] used n =0.50. Thus we adopt a more conservative estimate than used by the Working Group, ensuring that we do not underestimate the uncertainty of our repeat times. Elapsed times are obtained from the historical record or are set to 1000 year if the historical evidence favors the absence of a large events. Clearly, there is large uncertainty of the earthquake recurrence and elapsed times. We examined the sensitivity of the probability change to the parameters T_{ave}, n, and elapsed time. The probability change (but not the probability) is most sensitive to the coefficient of variation, and is insensitive to the mean recurrence time and elapsed time.

To estimate the total uncertainty in the calculated probabilities given in Table 2 resulting from the combined uncertainties associated with each parameter, we use a Monte Carlo technique (e.g. *Savage*, 1991; *Savage*, 1992). The parameters whose uncertainties we consider are Ds_f, $\dot{\tau}$, As, t_a, T_{ave}, and t (elapsed time). We assume that the error in each parameter is normally distributed about its mean. We estimate plausible mean values and standard deviations for As and t_a directly from the plots of the Kobe observations and obtain 0.35±0.15 (Figure 10) and 23.3±7.7 yr (Figure 14), respectively. Standard errors for the calculated Coulomb stress changes Ds_f correspond to the range of frictional values (0.1_ m' _ 0.7) used in the calculation. The parameters $\dot{\tau}$, T_{ave}, and t are assigned standard deviations equal to 0.25 times their respective estimated values.

In the Monte Carlo technique, each probability is calculated 1000 times, with the value for each parameter drawn randomly from the distributions described above. From the resulting suite of probability calculations, we report the 15.9% and 84.1% quantiles, which correspond to ±1 standard deviation range (parentheses in Table 2). We then repeated the calculations an additional 1000 times, again drawing parameters at random from their error distributions, and confirmed that the resulting 1 standard deviation probability range was essentially unchanged.

Results for the Earthquake Probabilities

An important feature of the state-dependent formulation of earthquake occurrence is that the both Weibull and Poisson probabilities include the transient effect of the stress changes on faults, and thus both become highly time-dependent. The conditional probabilities are changed by both the permanent and transient effects of the stress changes, whereas the Poisson probabilities are changed only by the transient effects. Figure 16 and Table 2 show the results of our computations for earthquake probability. Despite the long elapsed time for the earthquakes on most of the major faults, the 10- and 30-yr Poisson and Weibull probabilities on these faults are relatively low, owing to their long recurrence times. The ratio of the conditional probability calculated with and without the effect of the stress change is the probability gain or loss (Table 2, *col. 19* and *20*, and Figure 16c).

We now focus on the Weibull probability calculation, although the Poisson values are also listed in Table 2. An extraordinary change of probability appears for the ATTL. The probability of a Kobe-size earthquake on the east section jumps by a factor of 6-12 and 4-7 for periods of 10 and 30 years, respectively, and the probability for the west section is calculated to decrease by a factor of more than 40 and 20 for the 10- and 30-year periods. The Tokushima and Wakayama sections of the MTL and Yamasaki fault exhibit more modest 7-22%, 7-18%, and 9-27% probability gains for the next 30 years, respectively, and the Naruto section of MTL also shows a 30-year probability drop by a factor of 2.5-100. Apart from the ATTL and the Naruto section of the MTL, the probability gains are almost entirely due to the transient effects of the small (<0.1-bar) calculated stress changes. Only where the stress change is large enough to advance or delay the next earthquake by a significant fraction of the earthquake repeat time does the permanent effect of the probability dominate (Table 3).

In order to evaluate the probability of a Mw_6.9 earthquake near Osaka, we consider the ATTL, the Uemachi fault, the Ikoma fault and the Wakayama section of the MTL, all within 50 km of the city. The combined probability for an event on one of these faults is given by

$$P = 1 - (1 - P_a)(1 - P_b)(1 - P_c)\cdots \quad (14)$$

where P_a, P_b, and P_c are the individual conditional probabilities for earthquakes on assumed non-interacting faults a, b, and c, respectively [*Working Group on California Earthquake Probabilities*, 1988]. The probability of a Kobe-size earthquake in the Osaka metropolitan area is estimated to be 7-11% during the next 10 years, and 16-22% during the next 30 years, up from 5% and 14-15% before the Kobe earthquake and corresponding to probability gains of 1.4-2.2 and 1.1-1.5 (Table 4). The west section of the ATTL and Wakayama section of the MTL contribute most strongly to the heightened earthquake probability near Osaka.

Limiting Assumptions

Myriad assumptions are embedded in this study, with the number mounting as the analysis proceeds from stress to probability. At the outset, we make stress calculations in a homogeneous elastic halfspace, adopting an earthquake source model based on independent measurements (Figure 2). Calculation of the Coulomb stress change (Figure 3) requires an apparent friction coefficient and, in the case the of optimally oriented stress changes (Figure 5), a regional stress orientation (Figure 4). The seismicity rate change (Figure 8) is a smoothed function calculated from an earthquake catalog. The seismicity rate change over half the study area cannot be defined, principally because the background rate is too low to measure in the years before the earthquake. Despite these limitations, a correlation is evident between stress change and seismicity rate change (Figure 9), and this relationship is consistent with earthquake nucleation relations for state- and rate-dependent friction (Figure 10).

To pass from stress change to probability invokes a second set of assumptions drawn from historical-earthquake, paleoseismic, and geodetic data. These are needed to calculate the permanent probability change caused by hastening or delaying the time until the next large earthquake. Both the Poisson and Weibull models depend on the mean earthquake repeat time and its coefficient of variation, and the Weibull model also depends on the elapsed time. To capture the dominant transient effect of the stress changes on earthquake probability, we rely on rate- and state-dependent constitutive relations, which require measures of the aftershock duration and a state parameter. While we are able to estimate these parameters directly from the Kobe observations, we must assume that the aftershock decay period, *ta*, and the state variable, *A*s, inferred from the Kobe aftershocks apply to the surrounding region. Furthermore, while the rate and state description of friction is consistent with the seismicity rate change findings reported here and elsewhere, it requires further validation.

The Monte Carlo technique allows us to explore the sensitivity of the probabilities to plausible variation of the model parameters. We find that for most of the faults within 100 km of the Kobe main shock, earthquake probabilities change significantly as a result of the Kobe event.

Conclusions and Implications

The transfer of stress by fault slip is a fundamental feature of earthquakes. We sought to validate the role of Coulomb stress changes on subsequent earthquake occurrence by rigorously assessing whether stress changes on small faults correlate with the observed seismicity rate change associated with the Kobe earthquake. To make this test, we assumed that small faults of various orientations are present throughout the crust, and that those faults optimally oriented for failure slip in aftershocks. Such a general relation was found by *King et al*.[1994], *Stein et al*.[1994], *Gross and Bürgmann* [1997], and *Gross and Kisslinger* [1997]. We found that the correlation between stress change and seismicity rate change is statistically significant in areas experiencing absolute stress changes greater than 1.0 bar (0.1 MPa) (for the longer M_2.6 catalog) and 0.2 bar (0.02 MPa) (for the more sensitive M_2.2 catalog). From this we conclude that stress increases promote earthquakes, causing the rate of seismicity to rise above the background rate (termed 'aftershocks'), and that stress decreases inhibit earthquakes, causing the rate of seismicity to drop. We posit that regions of seismicity rate decrease accompany all earthquakes, but because such decreases are hard to detect, they have escaped notice.

With evidence of the effect of stress change on seismicity in hand for Kobe, we further argued that stress changes influence the occurrence of both small and large shocks. This is a logical extension of our observations. Large and small shocks have the same stress drops [*Abercrombie*, 1995]; they differ principally in that the fewer large shocks are restricted to large faults. Thus, for our probability analysis we resolved the Coulomb stress changes on the major faults.

The calculation of a transient amplification of the earthquake probability (Figure 12c), if correct, has profound implications for seismic hazard analysis. First, each large earthquake strongly alters the probability of subsequent events over a distance of several fault-lengths from the source for a period of several times the aftershock duration. Thus a thirty-year probabilistic earthquake hazard assessment is valid only until the next large shock occurs, at which time it must be revised. Second, earthquake probabilities—even under the Poisson assumption—are highly time-dependent. If this description of earthquake nucleation is correct, the time-advance and time-delay estimates presented in many recent stress-triggering papers greatly underestimate the effect of the stress change during the period of aftershock decay. They also ignore the probability changes associated with stress changes of less than a few tenths of bars, which are controlled by the transient effect. After a large earthquake, such stress-based forecasts could thus be useful for identifying those areas with enhanced probability for large and damaging aftershocks or further mainshocks.

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FIGURE CAPTIONS

Figure 1. (a) Tectonic setting of the Japanese islands. The Kobe earthquake was an intraplate event in the Eurasia Plate. Bold-lined square corresponds to the area in (b). (b) Epicenter (star) and surface projection (hatched area) of the two model fault planes [*Wald*, 1997] of the 1995 Kobe earthquake.

Figure 2. Distribution of modeled fault slip for the Kobe earthquake from *Wald* [1997], derived from strong motion, teleseismic and geodetic data. This cross-section is viewed from the southeast. The Nojima and Rokko (also called 'Suma or Suwayama') faults dip 80° southeast and 85° northwest, respectively. The fault slip is concentrated at shallow depth on the Nojima fault, where surface ruptures were observed.

Figure 3. (a) Coulomb stress changes resolved onto active faults caused by the Kobe earthquake. Faults are assumed to be either vertical strike-slip or 35°- dipping thrusts extending from the surface to a depth of 20 km, and divided into 10 km-long sub-patches. Calculations are made in an elastic halfspace following *Okada* [1992], assuming a shear modulus of 3.2 ¥ 10^{11} dyne-cm^{-2}, Poisson's ratio of 0.25 and apparent coefficient of friction m ¢ =0.4. The Coulomb stress change caused by the Kobe mainshock is calculated at the center of each subpatch. (b) Cross-section showing Coulomb stress changes resolved on the Median

Tectonic Line. Stress changes are calculated at the centers of the 10 km-long, 2.5 km-deep rectangular fault patches.

Figure 4. Maximum regional compression directions estimated from earthquake focal mechanisms (bold lines) and *in-situ* measurement by hydraulic fracturing method (bold lines with small dot) [*Tsukahara and Kobayashi*, 1991]. Although some fluctuations are seen, an assumption of E-W oriented uniaxial compression appears appropriate for our calculations. Focal mechanisms of moderate earthquakes (4.0_M_6.0, depth_30 km, 1983-1994) by the Japan Meteorological Agency (JMA) show that most events are strike-slip except in the Wakayama region and south of Kyoto, where reverse mechanisms are also seen. Dashed zones isolate areas with both strike-slip and thrust focal mechanisms or active faults. Regions outside the dashed zones are dominated by strike-slip faulting.

Figure 5. The largest Coulomb stress changes between 0 and 20 km depth caused by the Kobe earthquake on optimally oriented faults. The calculation assumes an E-W regional compression of 100 bars (10 MPa) and an apparent friction coefficient m ¢ of 0.4. (a) Stress changes on optimally oriented strike-slip faults. Strike-slip faults are shown as bold black lines, thrust faults as gray lines. (b) Stress changes on optimally oriented thrust faults. Thrust faults are shown as bold black lines, strike-slip faults as gray lines.

Figure 6. Spatial relationship between calculated stress changes and aftershocks. Displayed Coulomb stress changes are on optimally oriented faults as in Figure 4, and on specified active faults shown in Figure 3. M_1.0 aftershocks during the first 1.5 years are shown.

Figure 7. The minimum magnitude of catalog completeness as a function of time. The magnitude of completeness is found by noting the minimum magnitude for which a power law relationship between the number of earthquakes and magnitude apparently holds. Many stations were replaced and added in October, 1994 when the detection level abruptly improved. A slight degradation after the Kobe earthquake is probably an artifact of processing the huge number of small events after the main shock. We used two JMA catalog sets, M_2.2 (pre-3.5 months and post-1.5 years), and M_2.6 (pre-8 years and post-1.5 years) to measure seismicity rate changes; we show only the M_2.6 results in figures 8-10.

Figure 8. Seismicity from the JMA catalog before and after the Kobe earthquake, and a smoothed estimate of the corresponding seismicity rate change. (a) Background seismicity for M_2.6 earthquakes during the 8-year period before the mainshock. (b) Seismicity for M_2.6 earthquakes during the 1.5-year period after the mainshock. (c) Estimate of seismicity rate change expressed as ratio *R*/*r* of mean post- and pre-mainshock rates. Both the Kyoto and Wakayama regions are sites of abundant pre-Kobe seismicity and aftershocks. Seismicity rates in these regions increased after the Kobe earthquake.

Figure 9. (a) To compare the Coulomb stress change with the seismicity rate change *R*/*r*, the calculated stress change is smoothed by a Gaussian function as in the *R*/*r* calculation of Figure 8c. Points outside the 8-bar (0.8 MPa) contour are plotted in Figure 10. (b) Seismicity rate change *R*/*r* for M_2.6 earthquakes from Figure 8c. (c) Spatial correlation between the calculated stress change (a) and observed seismicity rate change (b). The correlation is represented by the agreement between the sign of the smoothed stress change and the sense of the seismicity rate change. Areas of agreement are red; areas of disagreement are blue. Undefined areas in the *R*/*r* calculation are gray. Positive correlation occurs in 63% of the defined area.

Figure 10. Observed seismicity rate change *R*/*r* and calculated Coulomb stress change for the points corresponding to the non-gray areas in Figure 9c, and theoretical curves predicted by equation (13). Points with Coulomb stress changes larger than 8 bars (see Figure 9a) were excluded. Values of $R/r > 1.0$

correspond to seismicity rate gains; $R/r < 1.0$ correspond to losses. Points with $R/r<0.1$ are set to 0.1. These data are well fit by the rate- and state-dependent friction relations for from *Dieterich* [1994] for As= 0.35±0.15.

Figure 11. Graphical illustration of the principal features of rate- and state- dependent fault behavior, after *Dieterich* [1994]. (a) Distribution of slip speeds against time to instability. The instability, a stick-slip event, will occur at time = 0. The black curve is the state evolution toward instability, given an initial slip speed of the nucleation source. Solid squares give the distribution of slip speeds at a constant or 'background' seismicity rate. Circles give the distribution of slip speeds after a stress step (in this example, a stress increase due to a nearby earthquake). The effect of a stress step is to increase slip speed of each source so that the time to instability is shortened. For example, square A moves to circle A', square B moves to circle B'. (b) Change of distributions of slip speeds due to the sudden stress increase in (a). Increasing the slip speed decreases the time to instability, and results in a higher density of nucleation points at short times, but does have little effect on the density at longer times before the instability. Thus, the initial distribution is compressed into the short time zone to instability, and the seismicity rate rises.

Figure 12. Schematic illustration of sudden stress increase and decrease on faults near an earthquake rupture. (a) A stress increase advances the time to the next rupture. If the fault were closer to failure at the time of the stress step, the stress step might trigger an earthquake immediately. (b) A stress decrease delays the time to the next rupture. The sudden stress decrease produces a "stress shadow" [*Harris and Simpson*, 1996] inhibiting earthquakes. (c) Long-term (permanent) and short-term (transient) probability change [*Dieterich and Kilgore*, 1996] associated with the stress increase. (d) Permanent and transient probability change associated with the stress decrease.

Figure 13. Calculation of the probability of an earthquake occurring between time t_0 and t_1, following stress transfer by a nearby earthquake at t_E. 10- and 30- year periods starting 2 years after the Kobe earthquake are used for this study.

Figure 14. Estimation of the aftershock duration t_a, defined as the time until the seismicity rate returns to the background rate before the 1995 Kobe earthquake. The observed rate of earthquakes during the aftershock sequence is plotted as open circles. The background rate of seismicity during 1987-1994 is shown as a horizontal line. The intersection of these lines gives t_a=23±7 years.

Figure 15. The distribution of destructive historical earthquakes [*Usami*, 1987] and major active faults [*Research Group for Active Faults in Japan*, 1991]. Historical earthquakes have been documented or recorded around Kyoto and Osaka region during the past 1,000 years, but the rate during the past century exceeds that in the historical record, suggesting that the 1,000-year record is incomplete.

Figure 16. Conditional probability for the occurrence of Mw_6.9 earthquakes on the major active faults in the 10-year interval 1997 to 2006 (left-hand panels) and in the 30-year interval 1997 to 2026 (right-hand panels). (a) Conditional probabilities without the effect of the stress change caused by the Kobe earthquake. (b) Conditional probabilities with the effect of the Kobe shock included. (c) Probability changes caused by the Kobe earthquake.