

Response of the San Andreas Fault to the 1983 Coalinga-Nuñez Earthquakes: An Application of Interaction-based Probabilities for Parkfield

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Abstract. The Parkfield-Cholame section of the San Andreas fault, site of an unfulfilled earthquake forecast in 1985, is the best monitored section of the world's most closely watched fault. In 1983, the $M=6.5$ Coalinga and $M=6.0$ Nuñez events struck 25 km northeast of Parkfield. Seismicity rates climbed for 18 months along the creeping section of the San Andreas north of Parkfield, and dropped for 6 years along the locked section to the south. Right-lateral creep also slowed or reversed from Parkfield south. Here we calculate that the Coalinga sequence increased the shear and Coulomb stress on the creeping section, causing the rate of small shocks to rise until the added stress was shed by additional slip. But the 1983 events decreased the shear and Coulomb stress on the Parkfield segment, causing surface creep and seismicity rates to drop. We use these observations to cast the likelihood of a Parkfield earthquake into an interaction-based probability, which includes both the renewal of stress following the 1966 Parkfield earthquake and the stress transfer from the 1983 Coalinga events. We calculate that the 1983 shocks dropped the 10-year probability of a $M\sim 6$ Parkfield earthquake by 22% (from $54\pm 22\%$ to $42\pm 23\%$), and that the probability did not recover until about 1991, when seismicity and creep resumed. Our analysis may thus explain why the Parkfield earthquake did not strike in the 1980's, but not why it was absent in the 1990's. We calculate a $58\pm 17\%$ probability of a $M\sim 6$ Parkfield earthquake during 2001-2011.

1. Introduction

Recent efforts to explain earthquake interaction by stress transfer have drawn support from the association between Coulomb stress changes and seismicity rate changes [Reasenberg and Simpson, 1992; Toda *et al.*, 1998; Stein, 1999; Wyss and Wiemer, 2000]. Stress increases are seen to be followed by increases in seismicity rate followed by a decay toward the background rate. Although seismicity rate increases are readily detectable, measurement of decreases is best achieved on the rare faults with very high rates of background seismicity. Probability models incorporating earthquake interaction by stress transfer [Dieterich and Kilgore, 1996; Parsons *et al.*, 2000] suffer from the need to average stress changes and constitutive parameters over fault surfaces much larger than the site of earthquake nucleation. Study of the Coalinga-Parkfield interaction helps to overcome these obstacles: First, the creeping section has among the highest rates of microearthquakes of any fault in the U.S., making detection of seismicity rate decreases much easier [Miller, 1996; Poley *et al.*, 1987; Wiemer and Wyss, 1997; Wyss *et al.*, 1990] (Fig. 1b). Second, the 1983 events were large enough to impart significant stress to the San Andreas, but far enough away that the unknown details of the 1983 fault slip have a negligible impact on the stress transfer (Fig. 1a). Third, synthesis of the rich assemblage of seismic, geodetic, and surface creep data at Parkfield permit an assessment of the parameters needed for an interaction probability analysis. Finally, the 1934 and 1966 Parkfield shocks nucleated within a 5 x 5 km fault patch [Bakun and McEvilly, 1979; Bakun and McEvilly, 1984], and the earthquake slip in 1934 and 1966 was similar north of Cholame [Segall and Du, 1993], so one can focus on the site of past—and perhaps future—earthquake nucleation and rupture.

2. Calculation of Seismicity Rate Change

Seismicity within a 120-km-long by 10-km-wide rectangle aligned with the San Andreas fault with endpoints at 36.50°N/121.08°W; 35.66°N/120.20°W was extracted from the Northern California Seismic Network (NCSN). $M \geq 1.3$ earthquakes were relocated by

Felix Waldhauser using the current NCSN Parkfield velocity model and a double-difference algorithm [Waldhauser *et al.*, 1999]. Relocation reduces the errors in depth and in location (Fig. 2a), enabling us to examine the seismicity rate change in both map view and cross-section. The cumulative number of $M \geq 1.3$ earthquakes as a function of time is shown in Fig. 2b; relocation introduces no obvious temporal artifacts, and only 4% of the earthquakes could not be relocated. We measure the seismicity rate starting from May 1980, because a shift in USGS magnitudes relative to U.C. Berkeley magnitudes occurred in 1978-1980 [Wiemer, *written comm.*, 2000; Wiemer and Wyss, 1997], which would produce systematic errors in rate calculations before this time.

Unlike the seismic moment rate, the seismicity rate counts earthquakes irrespective of magnitude. Because the smallest included shocks are most abundant, they will be most influential. The minimum magnitude of complete reporting, M_c defines the ideal lower magnitude limit for inclusion in seismicity rate calculations. To determine M_c for 1980-1990, we plot the departure of a power-law fit of the frequency-magnitude distribution in Fig. 3a, following Wiemer and Wyss [2000]. At the 90% confidence level used by Wiemer and Wyss [2000], $M_c = 1.3$, a result consistent with the histogram in Fig. 3c, which shows that the number of shocks climbs as the magnitude is decreased through 1.3. To guard against systematic errors arising if M_c changed with time, we calculate M_c for the both pre- and post-Coalinga periods in Fig. 3b; $M_c = 1.3$ in both cases.

3. Observations of Seismicity Rate Change

The rate of seismicity along the San Andreas fault increased after Coalinga for about 18 months in the creeping zone (Fig. 2c), whereas it decreased and remained low for more than 3 years in the locked zone (Fig. 2d). The seismicity rate change is plotted as a function of depth in Fig. 4. During 1985-1990, the seismicity rate returned roughly to normal along the creeping section, but remained low along the Parkfield section (Fig. 4e). After 1990, the seismicity rate returned approximately to normal everywhere (Fig. 4f). In Figs. 1 and 4, the number of earthquakes in cylindrical volumes of 5-km radius with

centers spaced 1 km apart is computed for pre- and post-Coalinga periods. The rate change is smoothed with a Gaussian filter for every volume in which there are at least 6 shocks in total, and for which there is at least 1 pre-Coalinga shock [Matthews and Reasenber, 1988; Reasenber and Simpson, 1992]. Because more accurately relocated post-1990 Parkfield seismicity using waveform cross-correlation shows few, if any, earthquakes in the uppermost 2 km [Rubin *et al.*, 1999], we regard rate changes in the upper 2 km in the 1980-1990 data as unreliable, and they are not considered.

To test whether the seismicity rate changes are unduly influenced by the Gaussian filter, we also calculated rate changes with just 15% of the smoothing. The same trends are evident, but with greater spatial variation. (Both smoothed and unsmoothed observations are carried into the regression analysis in the next section.) To test the significance of the rate changes, we plot the Z-statistic of Habermann [1983] for several time periods in Fig. 5b-d. These plot the statistical significance of a rate change, rather than the value of the change, relative to expected random errors. The largest seismicity rate changes in Fig. 4 correspond to the largest Z-values (exceeding ± 3) in Fig. 5, suggesting that, except at the southern end of the 1966 Parkfield rupture zone where the lower rate of seismicity precludes calculation of Z, the rate changes are significant. The 5 x 5 km Parkfield hypocentral zone exhibits a Z-value of 2.0-2.5 for the longer intervals (Fig. 5c-d).

4. Correlation of Seismicity and Stress Changes

Several studies [Parsons *et al.*, 1999; Reasenber and Simpson, 1992; Stein, 1999; Toda *et al.*, 1998] have found that seismicity rate change is correlated with the calculated Coulomb stress change, ΔCFF

$$\Delta CFF = \Delta \tau + \mu \Delta \sigma \quad (1)$$

(where τ is the shear stress and σ is the normal stress, positive for unclamping). If correct, not only should aftershocks be more prevalent in regions of Coulomb stress increase, but the rate of earthquakes should drop in regions of stress decrease. The calculated shear stress change on the San Andreas fault imparted by the Coalinga-Nuñez events resembles the observed seismicity rate change along the fault at mid-crustal depths in map view (Fig. 1b) and in cross-section (compare Fig. 4b and Fig. 4d). This correspondence suggests a causal relationship: San Andreas earthquakes became nearly twice as frequent (a log rate change of +0.3) where the stress increased by ~ 0.5 bars, and roughly half as frequent (log rate change of -0.3) where the stress decreased by the same amount. That the shear stress rose on the creeping section and dropped on the locked Parkfield-Cholame section (Fig. 1) is simply an accident of Coalinga's location; had the 1983 earthquakes struck north or south of Coalinga, the stress distribution in the San Andreas would have differed.

The seismicity rate change and Coulomb stress change are statistically correlated, but the extent to which the correlation is driven by the shear stress change is equivocal. Visual inspection of Figs. 2b and 2d suggests the shear stress controls the seismicity rate change (e.g., $\mu \sim 0$ in eqn 1). However, a spatial regression of the Coulomb stress change on seismicity rate change (Fig. 6b and 6c) indicates that the regression coefficient, R , increases with μ . A regression plot for $\mu = 0.4$ is shown in Fig. 6a. The dependence of the seismicity rate change on stress change (i.e., the slope of the regression), on the other hand, decreases as μ grows from 0.2 to 0.8. Thus, in our judgment, the data lack the sensitivity to suggest more than $0.2 \leq \mu \leq 0.8$. Regardless of the amount of smoothing or the assumed value of friction, the y -intercept is about -0.1, suggesting that the seismicity rates are biased toward negative values (i.e., in the absence of a stress change, the seismicity rates appear to decrease after May 1983). The bias is probably an artifact of the magnitude shift in the catalog that was not fully removed by using data starting from May 1980. (Using only later data does not circumvent this problem because the pre-Coalinga period would become too short to adequately measure seismicity rates.)

5. Analysis of Surface Creep Changes

Data from fault creepmeters (Fig. 1a) permit independent analysis of the Coalinga stress transfer, and together with the seismic data, enable us to gauge parameters for a probability estimate. Surface creep slowed, stopped, or reversed on all creepmeters for 1-4 years after the Coalinga earthquake [Schulz *et al.*, 1990], with the southern sites taking longest to recover (Fig. 7a).

The duration of retarded or reversed creep is correlated with the longterm creep rate: the faster the creep rate, the shorter the recovery time (Fig. 7b). A linear correlation, for which $R=0.99$, suggests that there would be no retardation where the creep rate equals the longterm San Andreas slip rate of ~ 23 mm/yr, and that the recovery would last approximately 4-5 yr south of Cholame, where the fault is fully locked. (A power law fit, in which the retardation period becomes infinite as the creep rate goes to zero, fits the data less well, with $R = 0.84$.) Coseismic offsets of 0.1-1.8 mm also accompanied the Coalinga shock [Mavko *et al.*, 1985]. In what follows, we model the observed creep series by calculating the amount a frictionless San Andreas would slip in order to shed the stress imposed by the 1983 shocks, and subtract this induced slip from the longterm creep rate during the observed period of the creep retardation.

To find the distribution of San Andreas slip needed to relieve the stress imposed by the Coalinga earthquakes, we treat the crust as an elastic halfspace, and represent the San Andreas fault as a planar grid of boundary elements [Crouch and Starfield, 1983] free to slip except where the fault is locked. We identify the locked region, where the fault slip rate is effectively zero, from the slip-rate inversion of GPS and longterm creep data by Murray *et al.* [2001] who discretized the fault into 2×3 km patches (Fig. 8c). Such an elastic two-state model (the fault is either free to slip or fully locked) matches the decay of the longterm or secular creep rates toward the southeast (Fig. 8a). Using the fault slipping/locked model of Fig. 8c, we then find the slip needed to shed the stress

imposed solely by the 1983 Coalinga-Nuñez shocks (Fig. 8b). The effect of the 1983 shocks is to impose 6-25 mm of left-lateral slip, except at XSC1 where 9 mm of right-lateral slip is imposed. Finally, we subtract the imposed slip from the expected longterm creep over the observed period of creep retardation (Fig. 7a). The result is that creep reverses where the ratio of the imposed slip over the retardation period is larger than the longterm creep rate. For example, at XMM1 the Coalinga shocks removed a calculated ~25 mm of San Andreas slip in one year; because this is greater than the 16 mm/yr creep rate, the creepmeter moved left-laterally at 9 mm/yr for a year. The model matches most creep records; the largest departure is in the creeping section (XSC1), where accelerated creep is predicted but not observed.

6. Calculation of Earthquake Probability

The time-dependent response of seismicity and creep to the stress imposed by the Coalinga-Nuñez sequence can be incorporated into an earthquake probability calculation [Stein, 1999]. In Coulomb failure theory, a positive or negative stress change on the San Andreas fault causes an advance or delay to the time until failure is reached (Fig. 9, *upper panels*), resulting in a modest but permanent change in earthquake probability (Fig. 9, *thin solid lines in lower panels*). If this were a complete description of the process, then the Parkfield earthquake would be delayed by the stress change (-0.15 bars) divided by the stressing rate (~0.1 bar/yr), or 1-2 yr. But such an approach fails to explain strong earthquake interactions on other faults. Examples include the order-of-magnitude decrease in $M \geq 6$ seismicity in the San Francisco Bay area during the 75 years after the great 1906 earthquake [Bakun, 1999; Harris and Simpson, 1998], and the 12 progressive $M \geq 6.7$ earthquakes along 1000 km of the North Anatolian fault during 1939-99 [Barka, 1996; Stein et al., 1997]. In such cases, the calculated static stress changes of several bars would at most advance or delay subsequent earthquakes by decades, and could not explain seismicity rate changes persisting for 60-75 years.

Our solution to this conundrum is to incorporate rate and state friction into the probability model, in which the transient effect of a stress decrease strongly amplifies the permanent decrease, because the fault slips at a lower rate, causing a lower rate of earthquake nucleation (Fig. 9, *lower panels*). The same phenomenon would apply to stress increases. Here seismicity is viewed as a sequence of nucleation events in which the state depends on the fault slip, slip rate, and elapsed time since the last event [Dieterich, 1994; Dieterich and Kilgore, 1996]. The seismicity rate equation is

$$R(t) = \frac{r}{\left[\exp\left(\frac{-\Delta CFF}{A\sigma_n}\right) - 1 \right] \exp\left(\frac{-t}{t_a}\right) + 1} \quad (2)$$

in which R is the seismicity rate as a function of time, t , following a Coulomb stress change, ΔCFF . A is a constitutive parameter, σ_n is the total normal stress, t_a is the aftershock duration, and r is the seismicity rate before the stress perturbation, or the background seismicity rate.

We seek the time-dependent seismicity rate $R(t)$ following a stress change ΔCFF on a fault, relative to the measured background rate, r . To evaluate (2), one calculates the stress change, and estimates two of the following three properties: the fault stressing rate $\dot{\tau}$, the aftershock duration t_a , or $A\sigma_n$. To build the interaction into a renewal probability, one further assumes that with the passage of time from the last $M \sim 6$ shock in 1966, another such earthquake becomes more likely. For this one must also assume a probability density function, and estimate the elapsed time since the last earthquake, the inter-event time, and the coefficient of variation for such events. Despite inevitable uncertainty in these assignments, the short repeat time, similar size, and long historical record for Parkfield earthquakes make estimates more reliable here than for most faults.

Aftershock Duration. The transient decay is proportional to the aftershock duration, t_a , the time elapsed until the rate of seismicity returns to the rate that prevailed before the mainshock occurred. In rate and state friction, the aftershock duration t_a is related to $A\sigma$ through

$$t_a = A\sigma / \dot{\tau} \quad (3)$$

where $\dot{\tau}$ is the fault stressing rate [Dieterich, 1994; Dieterich and Kilgore, 1996]; t_a is independent of mainshock magnitude. We use the four largest mainshocks on the San Andreas in the NCSN catalog to estimate aftershock duration as a function of position along the fault (Fig. 10a-d). There is an increase in aftershock duration toward the southeast, with durations growing from ~ 0.6 yr in the creeping section to ~ 5 yr in the locked region. Because we are limited by the small size of earthquakes in the creeping section, we can not eliminate the possibility that the aftershock duration is a function of magnitude. However, the measured aftershock durations are comparable to the observed creep retardation periods (shown in the map panel in Fig. 10), suggesting that the seismic and creep observations are manifestations of the same process: transient recovery to sudden stress changes. Based on the creep retardation periods of Fig. 7 and the aftershock durations of Fig. 10, we take t_a to be ~ 0.5 yr in the creeping section and ~ 4.0 yr in the locked Parkfield section.

Although far from proven, these t_a assignments are consistent with another means to gauge the aftershock duration. The stressing rate, $\dot{\tau}$, can also be approximated by the mainshock shear stress drop, $\Delta\tau$, divided by the inter-event time, t_r , then from (3)

$$t_a = t_r (A\sigma / \Delta\tau) \quad (4)$$

In the Parkfield section, the maximum observed earthquake magnitude $M_{\max} \sim 6$ and $t_r \sim 22$ yr. *Nadeau and Johnson* [1998] find that in the creeping section, $M_{\max} \sim 4$ and $t_r \sim 2.4$ yr. Thus if earthquake stress drops and $A\sigma$ were constant in both locations, one would

expect t_a to be roughly an order of magnitude larger in the locked section, consistent both with the observed creep retardation periods and aftershock durations.

Fault stressing rate. We need to estimate the fault shear-stressing rate $\dot{\tau}$ at sites where earthquakes occur, as shown schematically in the upper panels of Fig. 9. Three approaches to estimation of $\dot{\tau}$ lead to different answers. (1) If the San Andreas were vertical, straight and subject to the same plate boundary tractions—or alternatively, the same deep slip rate and locking depth—throughout the region of Fig. 1, then its tectonic shear stressing rate should be uniform along strike. At a mid-crustal depth of about 8 km, the shear stressing rate would be about 0.1 bar/yr. (2) If one instead assumes that where the fault creeps stress is relieved, then the stressing rate would be near-zero in the creeping zone and much higher than 0.1 bar/yr at the north end of the locked zone, where dislocations would continuously pile up (Fig. 8c). (3) *Rubin et al.* [1999] and *Waldhauser et al.* [1999] find that seismicity in the creeping section is concentrated along isolated streaks, with the patches between streaks undergoing steady, aseismic creep. The stressing rate in the streaks would be higher than in the Parkfield locked patch, where the load is more uniformly distributed. (Streaks are not evident in Fig. 4d-g because pre-1984 data cannot be relocated by waveform cross-correlation.)

Because we can not confidently choose from these alternatives, we use the stressing rate that matches the observed seismicity rate as a function of time (Fig. 11), and satisfies the creep retardation (Fig. 7) and aftershock durations (Fig. 10). This estimate is compatible only with the third alternative in the foregoing paragraph, and thus a rate appropriate for seismic streaks. The creeping section (Fig. 11a) sustained a calculated mean 0.3-bar Coulomb stress increase, and t_a there is 0.5-1.0 yr on the basis of Fig. 10. The decay of the seismicity rate as a function of time is best fit by a stressing rate of 0.5 bar/yr, about five times higher than the average San Andreas stressing rate. This stressing rate is also compatible with the spatial regression of seismicity rate on stress change shown in Fig. 6 (*gray curve*). Both the temporal and spatial seismicity-rate data are fit by a high stressing rate in the creeping section. In the Parkfield section (Fig. 11b), the mean calculated

Coulomb stress change is -0.15 bars, and the data are satisfied by a stressing rate of 0.1 bar/yr for $t_a = 2-4$ yr. Thus, given one degree of freedom (the stressing rate), we can satisfy the temporal decay of the seismicity rate following the Coalinga-Nuñez earthquakes.

Parkfield probability. The probability follows naturally from the seismicity rate change plot of Fig. 11. We assume that the seismicity rate is altered not just for microearthquakes but for all magnitudes. Because $M \sim 6$ earthquakes are infrequent, there is only a chance that the rate change will result in a detectable change in the occurrence of a $M \sim 6$ event after 1983. We performed a Monte Carlo analysis of 1,000 runs in which the tested values were drawn from a Gaussian distribution of the input parameters, and plot the mean value and uncertainty as a function of time (Fig. 12). The calculated stress change at the 5×5 km hypocentral site of the 1934 and 1966 earthquakes is -0.3 ± 0.1 bar (Fig. 4b-c). The fault stressing rate (0.1 ± 0.025 bar/yr), aftershock duration (4 ± 1 yr) are estimated from the preceding analysis of the creep and seismicity data. Calculations were made by running half the cases with lognormal and half with Brownian Passage Time [Matthews *et al.*, 2001] probability density functions, with mean inter-event time of 22 years, and a coefficient of variation of 0.5 (given a range of 0.35 [Savage, 1993] to 0.70 [Roeloffs and Langbein, 1994]). The Brownian passage time is a renewal function with superposed Brownian noise; it has been used in the most recent Working Group probability analyses [Working Group on California Earthquake Probabilities, 1999]. The Monte Carlo analysis tests the full distribution of uncertainty, not simply the quoted 1σ error range.

The calculated 10-year probability (Fig. 12) in 1983 decreased from $54 \pm 22\%$ to $42 \pm 23\%$, and does not return to the pre-Coalinga probability until about 1991. October 1992 marked the beginning of a period of heightened seismic activity at Parkfield [Michael and Jones, 1998; Fletcher and Guatteri, 1999], including a $M=4.3$ shock in 1992, a $M=4.6$ shock in 1993, and a $M=4.7$ shock in 1994). The occurrence of the renewed activity is consistent with our probability calculation. Although the probability undergoes a fractional drop of

22% in 1983, it is not statistically significant. The rate/state effects of the stress change could be added to a Poisson model, as illustrated by *Toda et al.* [1998]. This would yield smaller nominal uncertainties because the coefficient of variation of the inter-event time (0.5) would not enter into the calculations. However, we regard renewal as a better description of earthquake occurrence at Parkfield.

The 1992-1994 earthquakes altered the stress at the site of the 1966 Parkfield hypocenter several kilometers away, and thus changed the succeeding probability of future Parkfield earthquakes. We calculate the Coulomb stress changes associated with the 1992-94 shocks on the hypocentral patch (for $\mu=0.4$) using sources simplified from *Fletcher and Spudich* [1998], in which the slip is assumed to be pure right-lateral on vertical co-planar faults (Fig. 8d). The left panel shows an increase in stress at the 1966 hypocenter by 4.4 bars if the sources are themselves coplanar with the 1966 hypocenter. If they were located 1 km to the northeast (right panel of Fig. 8d), the stress would drop by 0.8 bars. The mean stress change in a 3 x 3 km Parkfield hypocentral patch is +4.1 bars if on the fault, and -1.0 bar if the 1992-94 shocks were just 0.5 km off the fault (Fig. 8d). The main trace of the San Andreas lies 0.5-1.0 km northeast of the 1992-94 shocks [*Fletcher and Guatteri*, 1999], and the location and depth of the 1966 hypocenter is uncertain by at least 1 km. Borehole tensor strainmeters [*Gwyther et al.*, 1996] may provide additional evidence that the 1992-94 shocks changed the conditions for future Parkfield earthquakes. The right-lateral shear strain rate increased starting in 1993 at two of three borehole instruments at Parkfield, although rainfall may influence or account for these signals, as discussed by *Gwyther et al.* [1996] and *Roeloffs* [2001]. Thus, while the effect of the 1992-94 shocks on the 1966 hypocenter is likely large, the probability change depends on unknown features of the geometry of the patch and fault.

7. Comparison with Other Studies

Several studies previously identified the seismicity rate decrease at Parkfield [Miller, 1996; Poley *et al.*, 1987; Wyss *et al.*, 1990], but the rate increase along the creeping section had escaped notice, perhaps because it is briefer and spatially restricted. Wyss *et al.* [1990] described a seismicity rate decrease in the Parkfield region that began 2 years after the Coalinga shock. We find that the rate decrease southeast of Parkfield instead began at the time of the Coalinga shock, but this was masked by the rate increase north of Parkfield, which did not end until about 18 months later. Thus the seismicity rate decrease that Wyss *et al.* ascribed to a process precursory to the next Parkfield earthquake we instead interpret as a response to the Coalinga shocks. Wiemer and Wyss [1997] found that the a - and b -values of the frequency-magnitude relation in the Parkfield hypocentral zone are anomalous with respect to the rest of the creeping and locked zones, and attributed this to its role as a highly stressed asperity. They did not explore the change in these parameters after the Coalinga earthquake.

Simpson *et al.* [1988] fit the creepmeter records to a model that is, like ours, driven by stressing from the deeper San Andreas fault, and modulated by the stress change associated with the Coalinga earthquake. In their model, the San Andreas stressing rate below 5 km is 0.75 bar/yr at XMM1 and 0.5 bar/yr at XGH1. They let the upper 5 km of the fault respond in a linear viscous manner to the stress changes, producing a left-lateral excursion at XMM1, and smaller left-lateral excursions on creepmeters to the south, all of about one-year duration. There is no tendency for creep retardation periods increasing toward the south, but the magnitudes of the creep changes resemble the observations. It is unclear, however, whether their modeled creep reversal occurred because they doubled the Coalinga coseismic slip to make the changes in the creep rates more apparent on their plots.

Examining the 1983 coseismic offsets in the creep series, Mavko *et al.* [1985] found a rough match between the creepmeter offsets and calculated Coulomb stress changes for the Coalinga-Nuñez shocks, using $\mu=0.6$. Analysis of the response of creepmeters to rainfall and earthquakes by Roeloffs [2001] indicates, however, that re-centering of the

instruments during shaking contaminates the coseismic displacements, whereas the creep rates before or after earthquakes suffer fewer such problems. Thus we offset the modeled creep series at the time of the Coalinga earthquake in Fig. 7a.

Miller [1996] proposed that unclamping of the Parkfield section of the San Andreas by the Coalinga earthquake, as shown in Fig. 4a, decreased the pore fluid pressure in the fault zone, lowering the rate of fault weakening and inhibiting large earthquakes for an extended period. Unlike our explanation, in *Miller* [1996] the shear stress changes (Fig. 4b) have no role in changing the conditions for failure. Lacking any direct measurements of pore pressure or pore fluid flow, *Miller* supported his model by observations of seismicity rate, which he cites as rising from 1969 to 1983, dropping after the Coalinga event and remaining low through 1997. We instead regard the apparent increase in $M \geq 2.5$ seismicity rate from zero in 1969 as an artifact of inadequate reporting in the NCSN catalog. The *Meagher and Weaver* [2000] catalog, complete to $M \geq 3.5$ from 1934, shows a constant seismicity rate from 1940-1966. Further, the observed increase in seismicity rates during 1983-85 in the creeping section (Fig. 2c and Fig. 4d) is incompatible with *Miller's* model, because this occurred where the unclamping was greatest. *Miller* does not explicitly consider the creep data, except to remark that changes in creep rates after 1983 are reconciled by the strengthening effect of homogenizing the shear stress along the fault. But why creep rates reversed near Middle Mountain for up to a year and slowed or stopped farther south for 3-4 years is not explained. (*Miller* also predicted a Parkfield earthquake 34 years after the 1966 event (in 2000), which has not occurred.)

8. Conclusions

We have sought to integrate the diverse observations at Parkfield in order to develop an understanding of the process by which stress contributes to or modulates the occurrence of earthquakes. We have argued that stress increases and decreases associated with nearby earthquakes influence subsequent seismicity. Seismicity and creep rates are

easily observed at Parkfield; they change at the time of the 1983 Coalinga-Nuñez shocks in a manner that resembles the calculated Coulomb stress change imparted by the 1983 shocks. The seismicity rate changes are robust, and the correlation of the changes with the Coulomb stress changes is statistically significant. The change in surface creep is also compatible with stress interaction. From creep and aftershock observations, we infer that the San Andreas takes up to ten times longer to recover from a stress perturbation where it is locked than where it creeps, consistent with rate/state friction.

We argue that a probability model governed by rate and state friction and driven by steady stress buildup and stress transfer from nearby earthquakes can satisfy most of the Coalinga-Parkfield observations. We find the effect of the stress decrease on earthquake probability at Parkfield was larger and lasted longer than previously supposed, and may have contributed to the absence of a Parkfield earthquake since 1983. Looking forward, we calculate that the probability of a $M \sim 6$ Parkfield shock today is higher than it was before 1983; during 2001-2011, we estimate a $58 \pm 17\%$ probability of a Parkfield earthquake. These prospective results are severely tempered, however, by uncertainty associated with the stress imparted by the 1992-94 shocks near the site of the 1966 Parkfield nucleation.

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Figure Captions

Fig. 1 (a) Shear stress transferred by the 2 May 1983 $M=6.5$ Coalinga (CO) and 11 June-22 July $M=6.0$ Nuñez (NZ) earthquakes on vertical right-lateral planes parallel to the San Andreas fault at 8 km depth. Source parameters for CO are from *Stein & Ekström* [1992] for the coseismic period: 150° strike, 15° W dip, 4.7 m reverse slip, 10 km upper depth, 1.5-4.0 km width; for NZ they are based on *Eaton* [1990] and *Rymer et al.* [1990]: 178° strike, 65° E dip, 0.22 m right-lateral and 0.65 m reverse slip, 5.4 km length, 2 km upper and 8.3 km width. Excluded because of their negligible impact on the stress are the 25 October 1982 $M=4.8$ New Idria (NI) shock and the 4 August 1985 $M=6.0$ Kettleman Hills (KH) shock [*Ekström et al.*, 1992]. (b) The observed seismicity rate change is superimposed along the San Andreas fault, also at 8 km depth. Relocated $M \geq 1.3$ seismicity is from the NCSN in a 10-km-wide box with endpoints at $36.50^\circ/-121.08^\circ$; $35.65^\circ/-120.20^\circ$.

Fig. 2. Seismicity along the San Andreas fault in the vicinity of the 1983 Coalinga-Nuñez earthquakes. (a) Northern Calif. Seismic network (NCSN) catalog locations (*left panel*) and double-difference relocations (*right panel*) following *Waldhauser and Ellsworth* [2000]. Endpoints of the box in the 'Locked' zone are $35.978^\circ/-120.537^\circ$; $35.749^\circ/-120.306^\circ$; width=10 km, depth=3-12 km. For the box in the 'Creeping' zone they are $36.368^\circ/-120.934^\circ$; $36.096^\circ/-120.657^\circ$; width=10 km, depth=2-10 km. (b) The number of earthquakes as a function of time is little different for the NCSN and relocated shocks. (c) The seismicity rate is seen to rise for about 18 months after Coalinga in the creeping section. (d) Within several months after the Coalinga shock, the seismicity rate drops in the 1966 Parkfield rupture zone. Relocated earthquakes are shown in (c) and (d).

Fig. 3. Calculation of the minimum magnitude of complete reporting, M_c , following *Wiemer and Wyss* [2000]. (a) $M_c=1.3$ for both the NCSN and relocated earthquakes. (b)

$M_c=1.3$ for both the 3-yr pre-Coalinga and 7-yr post-Coalinga periods. (c) Histogram of the number of earthquakes in each magnitude band, the data used to construct (a-b).

Fig. 4 Calculated normal (a), shear (b), and Coulomb (c) stress change and observed seismicity rate change within 5 km of the San Andreas fault (d-f) associated with the 1983 Coalinga-Nuñez shocks, with earthquakes during the indicated post-Coalinga period superimposed in black. The seismicity rate change for the post-Coalinga 1.5-year period 830502-841101 (yr/mo/dy) (Fig. 4d) and the 5.5-yr period 841102-90501 (Fig. 4e) are calculated relative to the 3-year pre-Coalinga period 80502-830501. The stress changes are similar to those of *Simpson et al.* [1988]. Note the association between the calculated shear stress change (b) and the observed seismicity rate change during the first 1.5 yr after the Coalinga sequence (d). The rate increase in the creeping section (km = -50 to 0) disappears after 1-2 years (e). The seismicity rate cannot be reliably determined in areas rendered gray.

Fig. 5 Calculation of the significance of the seismicity rate changes shown in Fig. 4, using the *Z*-statistic of *Habermann* [1983] and *Zuñiga and Wyss* [1995] (Performed using *Zmap* 5.0 of S. Wiemer, using a variable smoothing radius with $r_{\min}=7.5$ km, $n=100$, and the ‘rubberband’ function.) Strong rate increases are evident in the creeping section and somewhat weaker seismicity rate decreases are evident in the Parkfield hypocentral zone. *Z*-values exceed ± 3 in several key areas, although the paucity of earthquakes precludes calculation of the *Z*-statistic southeast of the Parkfield hypocentral zone.

Fig. 6 (a) Spatial regression of the calculated Coulomb stress change (with $\mu=0.4$), from Fig. 4c, on the observed seismicity rate change during the first 18 months after the Coalinga shock, from Fig. 4d. Some 69% of the variance is accounted for by an exponential fit to the data (*black line*). The positive Coulomb stress change data are also well fit by rate-state parameters appropriate for the creeping section (*gray curve*). Dependence of the regression coefficients on friction, μ , is shown in (b) for the data as smoothed in Fig. 4, and in (c) when it is minimally smoothed.

Fig. 7 Response of creepmeters to the Coalinga earthquake, with records displayed north to south (top to bottom). (a) Observed and modeled surface creep; creepmeter locations are shown in Figs. 1a and 2. Rainfall is responsible for the rate increases in early 1983, but longterm creep rates and the creep retardation after Coalinga are judged reliable by *Roeloffs* [2001]. A 0.6-mm offset occurred on the day of the 4 August 1985 Kettleman Hills shock on XMM1, and the creep rate reversed on XPK1 5-6 days later. No other changes are evident at the time of the Kettleman Hills shock. The surface slip needed to shed the stress imposed by Coalinga, from Fig. 8b, is also indicated, with right-lateral slip positive. Subtracting the slip imposed by Coalinga from the longterm slip rate over the period of observed creep retardation produces the modeled creep series (*dashed*). (b) The linear relationship between the creep retardation and the longterm creep rate suggests a maximum retardation in the locked Parkfield section of ~ 4.4 yr; we equate this with the aftershock duration, t_a' , for the locked section.

Fig. 8 Calculated response of the San Andreas fault to secular loading (a) and to stresses imposed by the 1983 Coalinga-Nuñez earthquakes (b). The fault (c) is represented by a grid of freely slipping boundary elements except where it is fully locked [*Murray et al.*, 2001]. The observed longterm creep rates (open circles in a) are slightly lower than the model, appropriate if the creepmeters do not span the full width of the fault zone. The calculated slip values in (b) are reproduced in Fig. 7a. (d) Stress changes in the 1966 hypocentral patch caused by the 20 Oct 1992 M=4.3, 14 Nov 1993 M=4.6, and 20 Dec 1994 M=4.7 shocks [*Fletcher and Spudich*, 1998], which lie 0.5-1.0 km southwest of the main trace of the San Andreas fault near the 1966 hypocenter. The values at the '1966 epicenter' correspond to the star; the 'mean' values are for the 3×3 -km area centered on the star.

Fig. 9 Schematic illustration of the effect of negative (at point x) and positive (at point y) stress changes on earthquake probability along a strike-slip fault, given an arbitrary off-fault earthquake source (*rupture*). The 10-yr probability is the integral over the period in

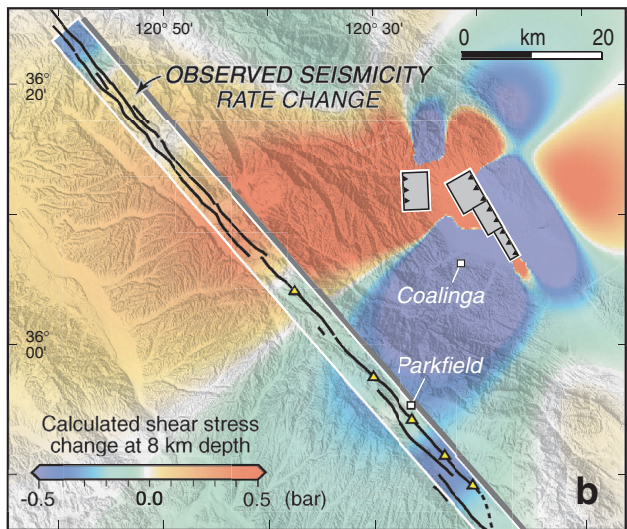
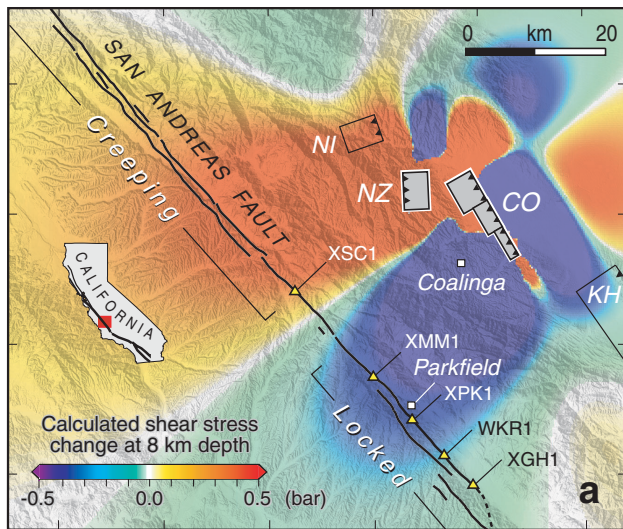
question. Rate/state effects of the shift to a time earlier or later in the earthquake cycle are neglected in this example. The lower panels can be compared to the observed seismicity rate changes in both the creeping and Parkfield sections (Fig. 11) and calculated probabilities for Parkfield (Fig. 12).

Fig. 10 Estimated aftershock durations, t_w , along the San Andreas fault, with features labeled in (a) and earthquake and creepmeter locations shown in the map panel at bottom. To permit the best possible calculation of the aftershock durations, the largest earthquakes along the creeping and Parkfield sections were selected at the four sites. Aftershocks were extracted from the NCSN catalog in the boxed regions; the background seismicity rate was estimated in the same boxes from 1980 until the time of the mainshocks. For the 1966 earthquake, the catalog of *Meagher and Weaver* [2000] was used, with the background rate estimated for 1937.5-1966.5. There is a general trend toward longer durations in the more fully locked part of the fault, consistent with the creep retardation periods in Fig. 7.

Fig. 11 Observed and theoretical response of San Andreas seismicity to the 1983 Coalinga-Nuñez earthquakes. The observed seismicity rate changes are averages for the rectangles shown in mapview (Fig. 2a) and in cross-section (Fig. 4d-f). Seismicity rates have been shifted by a log rate change of -0.1 , because of the negative seismicity rate bias discussed in the text, and indicated in Fig. 6 as the y intercept. The mean Coulomb stress changes are from Fig. 4c. (a) For $t_a=0.1$ yr, $A\sigma=0.05$ bar; for $t_a=0.5$ yr, $A\sigma=0.25$ bar; for $t_a=2.0$ yr, $A\sigma=1.0$ bar. (b) For $t_a=1$ yr, $A\sigma=0.1$ bar; for $t_a=2$ yr, $A\sigma=0.2$ bar; for $t_a=4$ yr, $A\sigma=0.4$ bar. In (c) and (d), the stressing rate is varied instead. The first interval covers 0.5-1.0 yr after the Coalinga shock to avoid the time period of the swarm at $\text{km}=22$ in Fig. 4d. The seismicity rate change data are approximately fit by $t_a\sim 2-4$ yr, in accord with creep (Fig. 7b), and seismic data for the 1966 aftershock duration (Fig. 10d).

Fig. 12 Ten-year probability of $M\geq 6$ earthquakes at Parkfield, calculated by Monte Carlo analysis following *Parsons et al.* [2000]. The 10-yr probability drops from $54\pm 22\%$ to

$42 \pm 23\%$ as a result of the stress decrease imposed by the Coalinga earthquake, and does not recover to pre-1983 values of probability until 1991. The 10-yr probability of such an earthquake for 2001-2011 is calculated to be $58 \pm 17\%$.



Modeled		Not included	
CO	1983 M=6.5 Coalinga	NI	1982 M=4.8 New Idria
NZ	1983 M=6.0 Nuñez	KT	1985 M=6.0 Kettleman

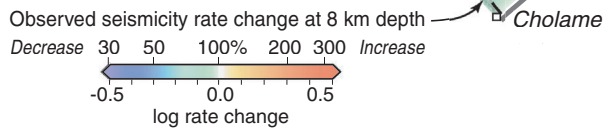


Fig. 1

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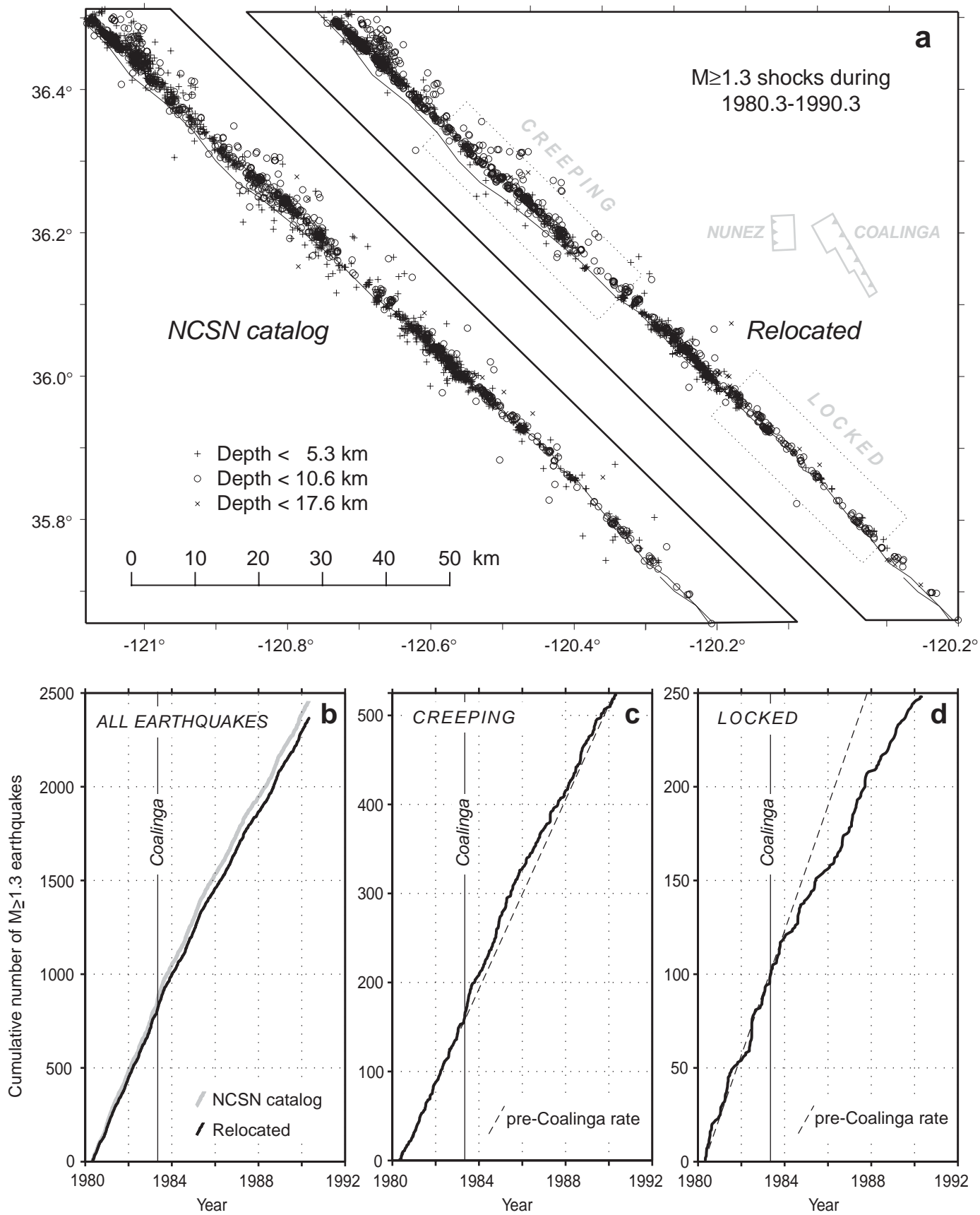


Fig. 2
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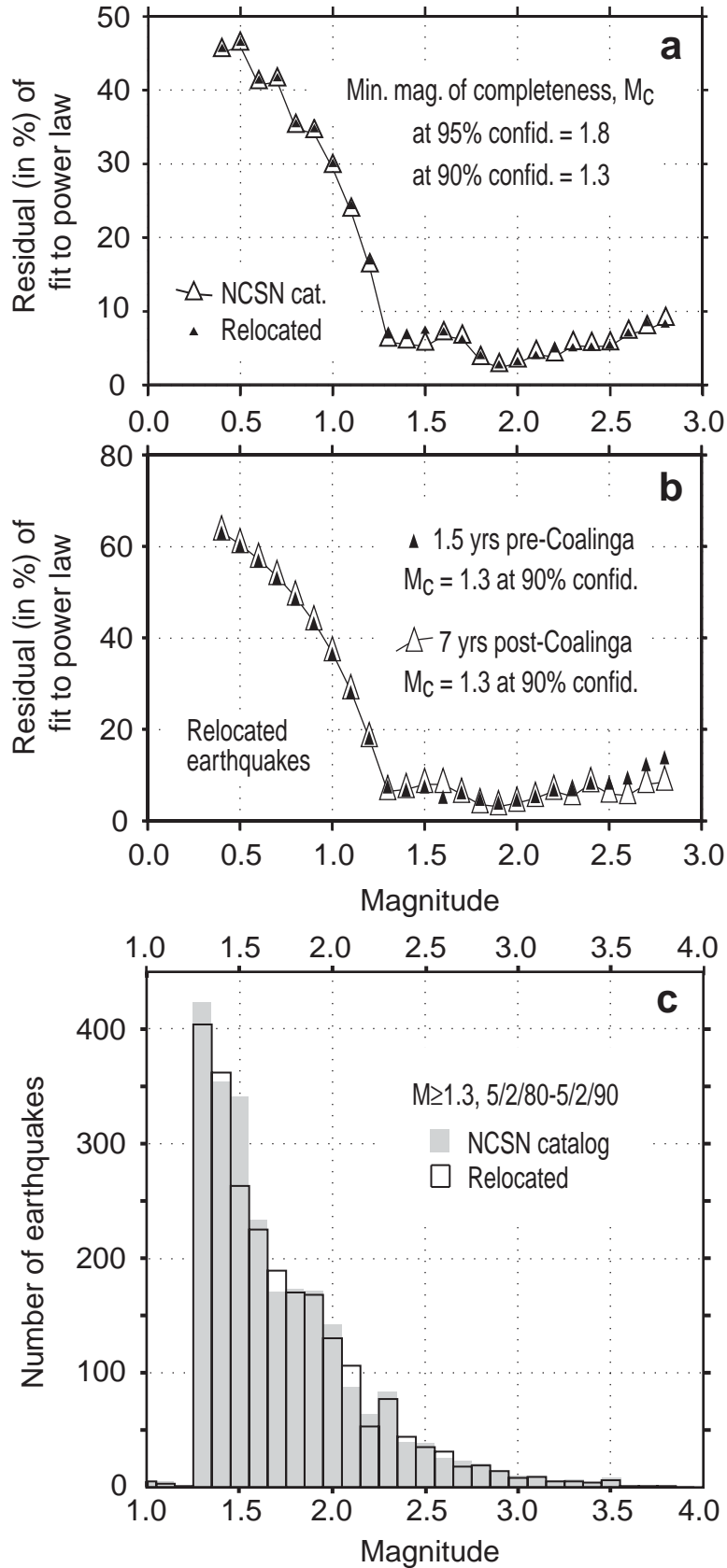


Fig. 3
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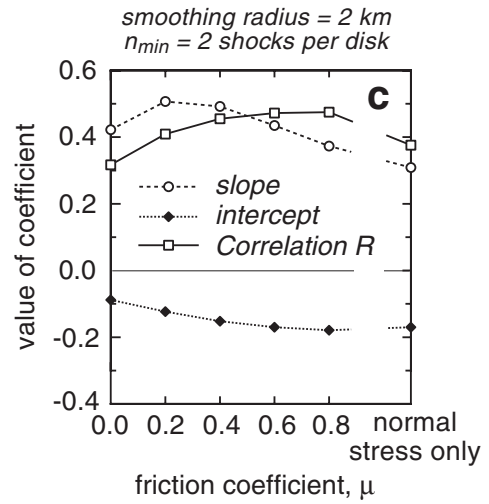
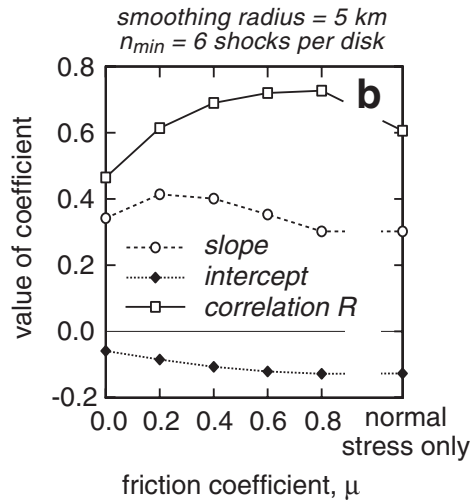
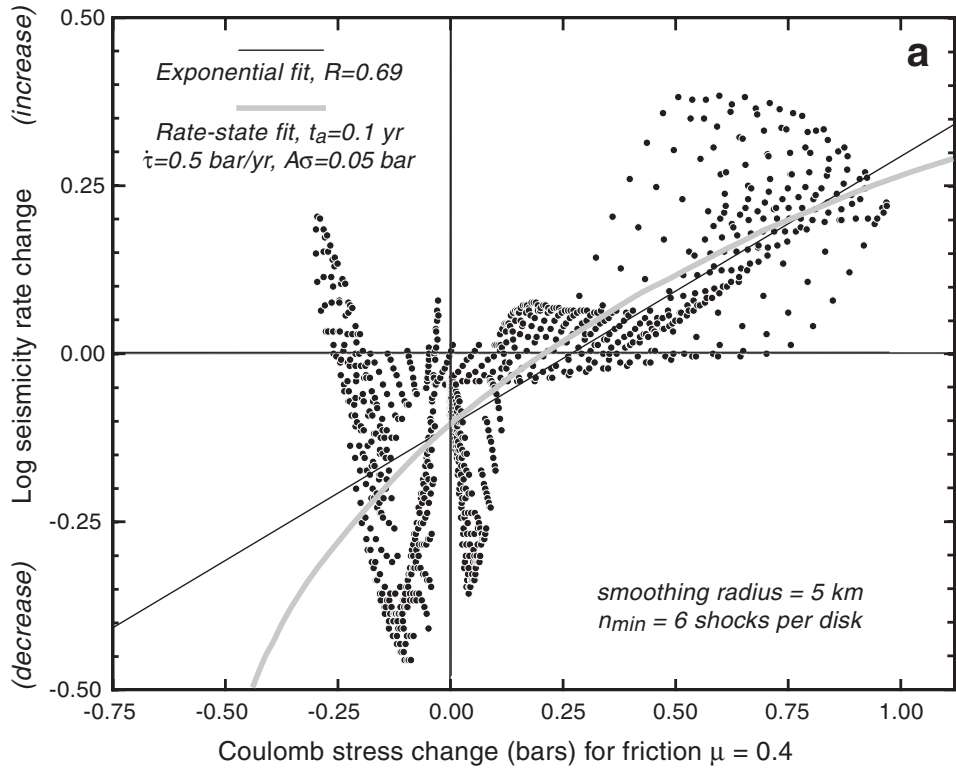


Fig. 6

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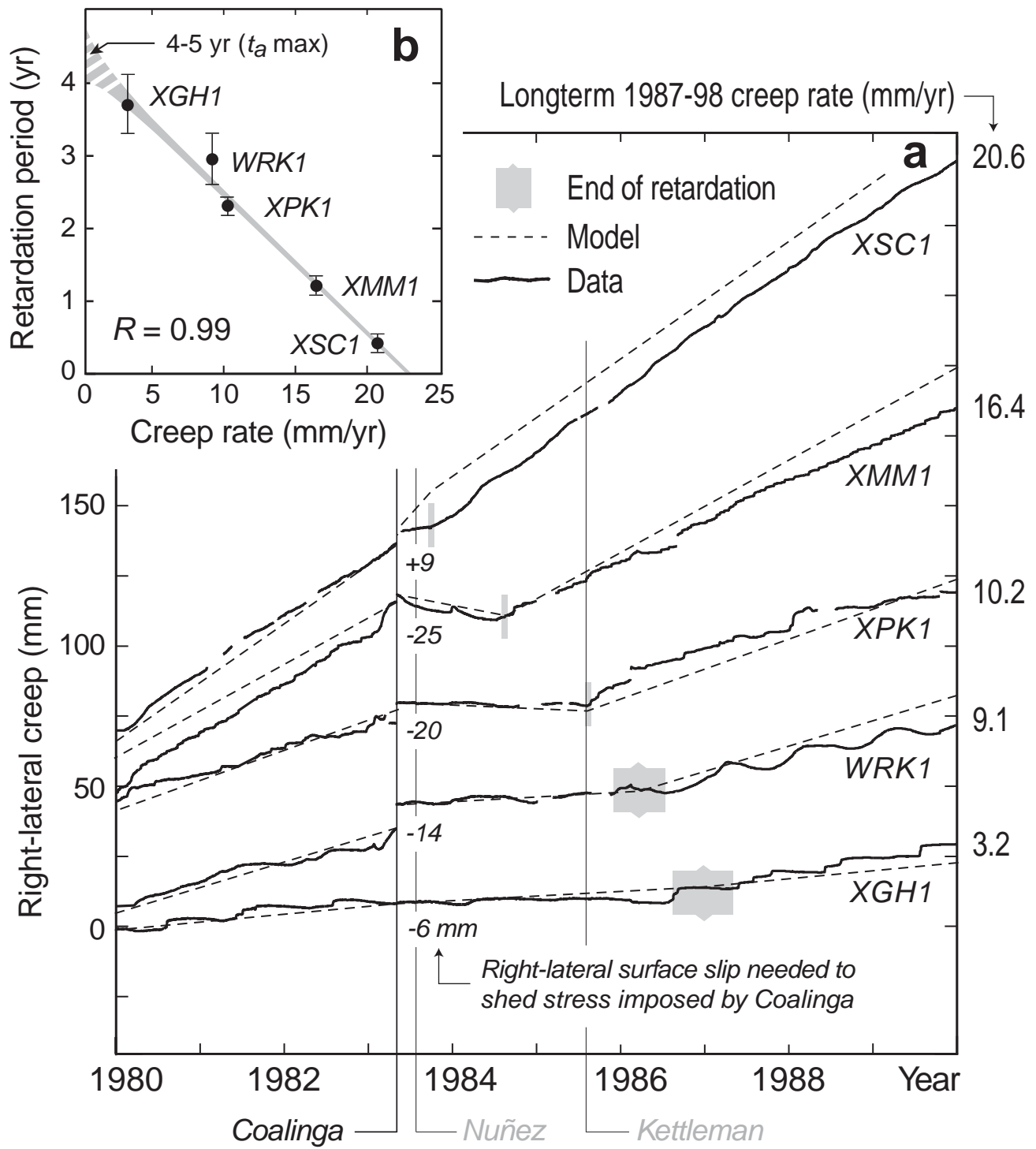


Fig. 7
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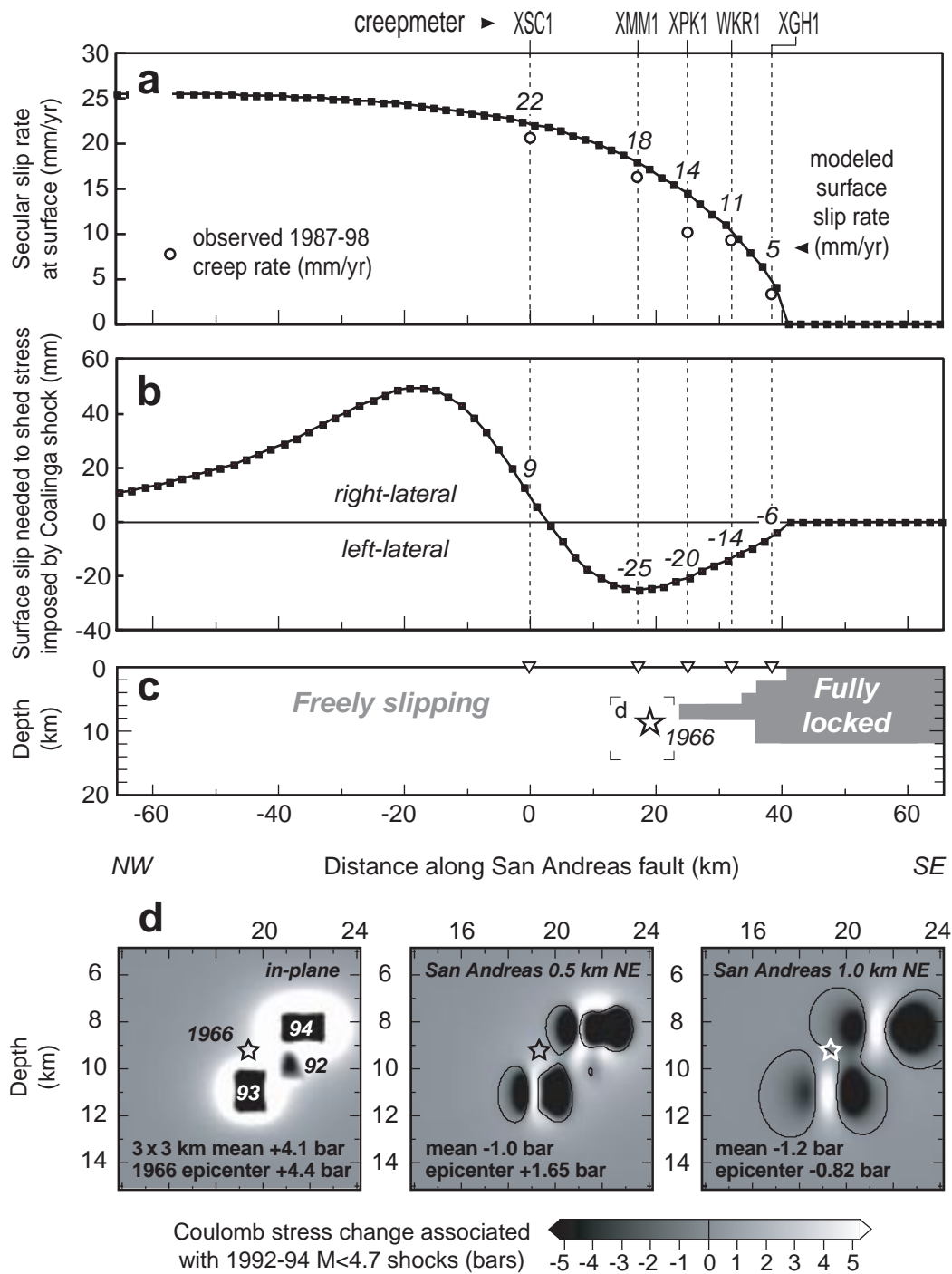


Fig. 8

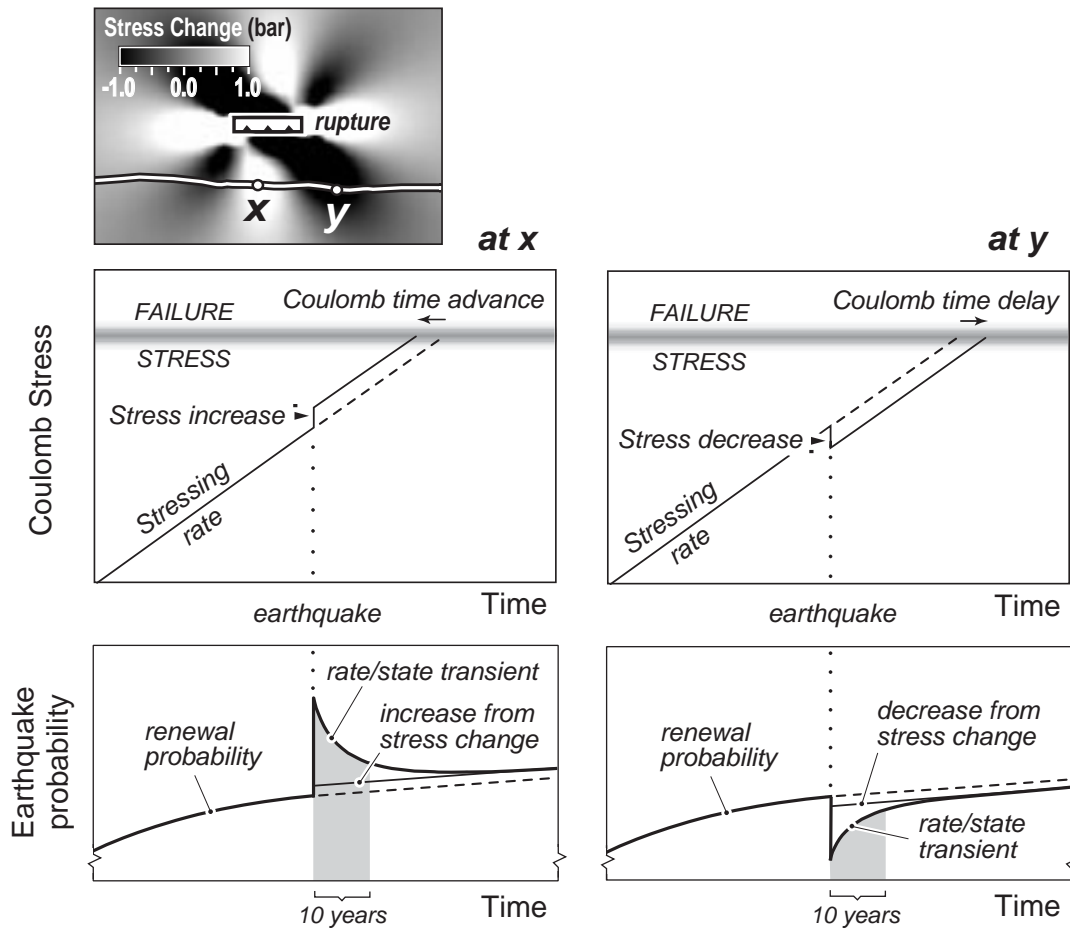


Fig. 9
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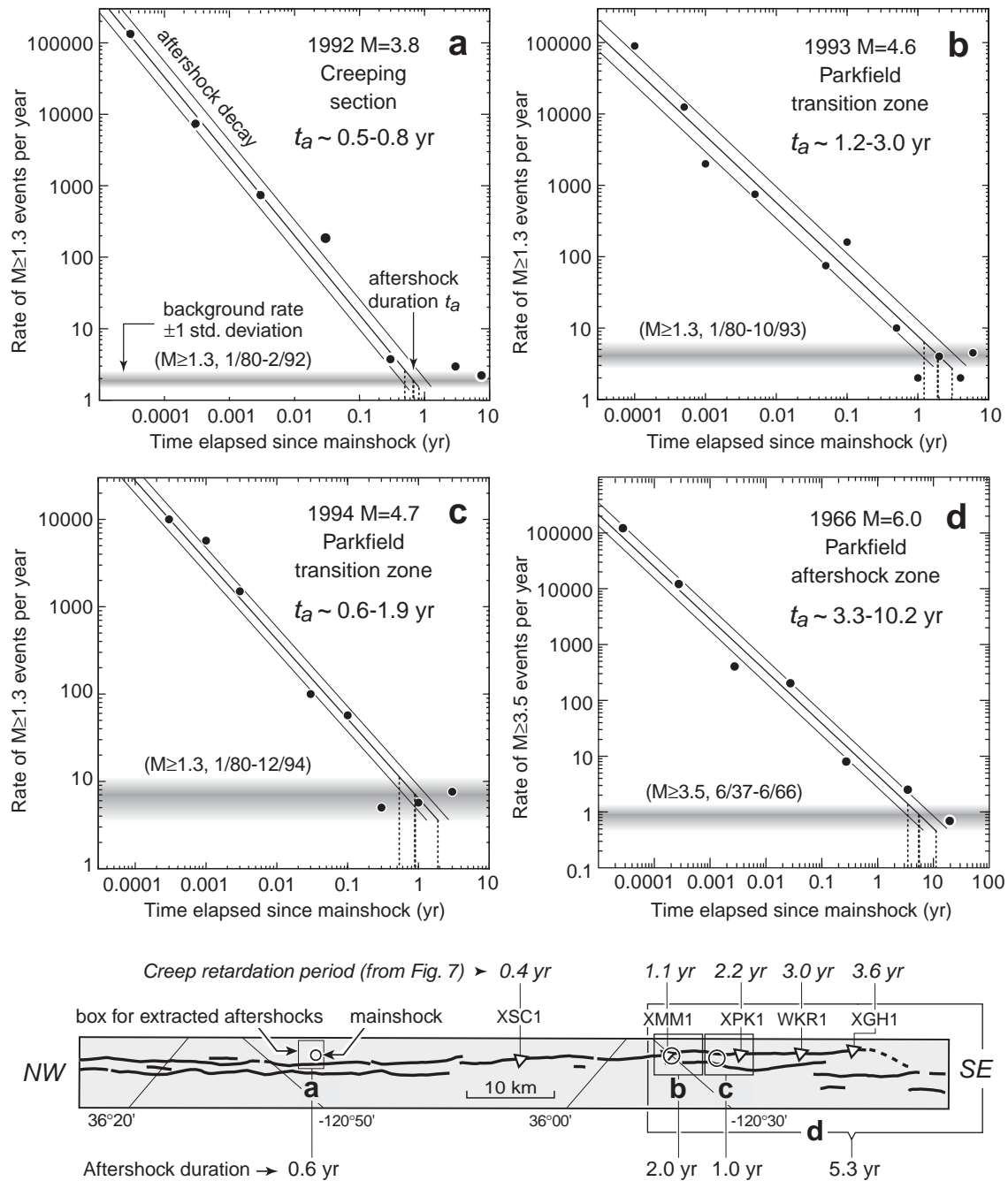


Fig. 10

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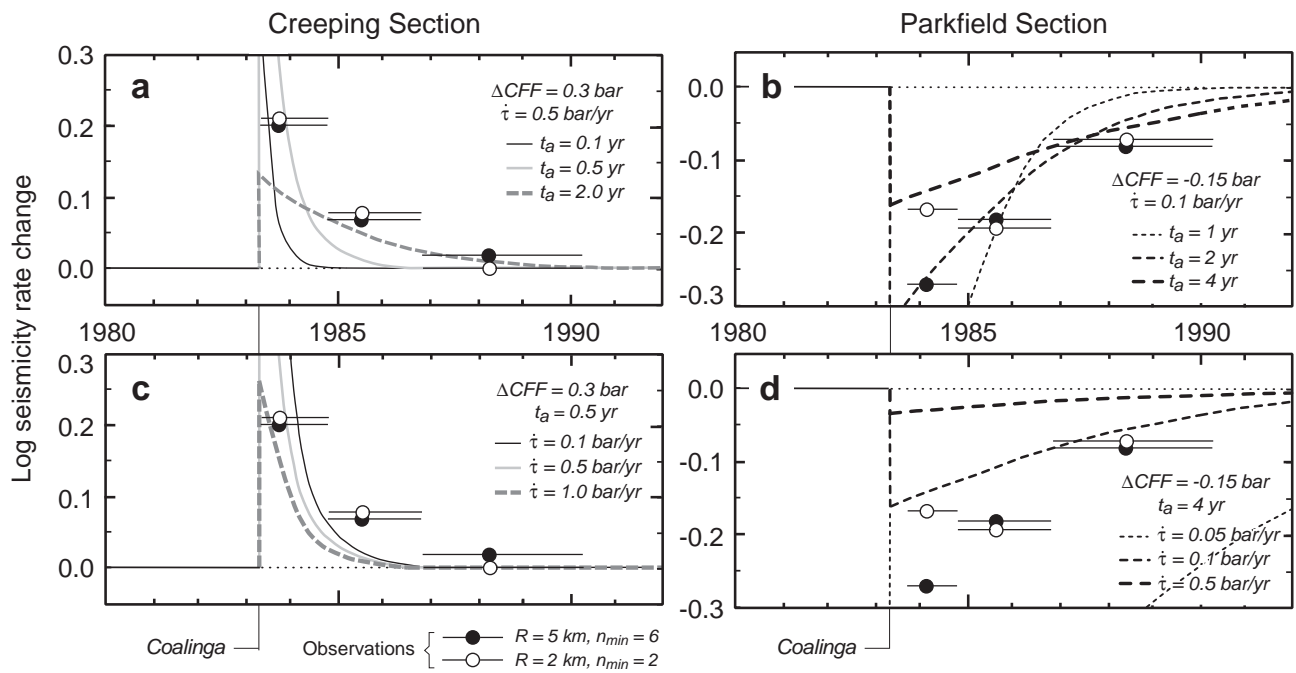


Fig. 11
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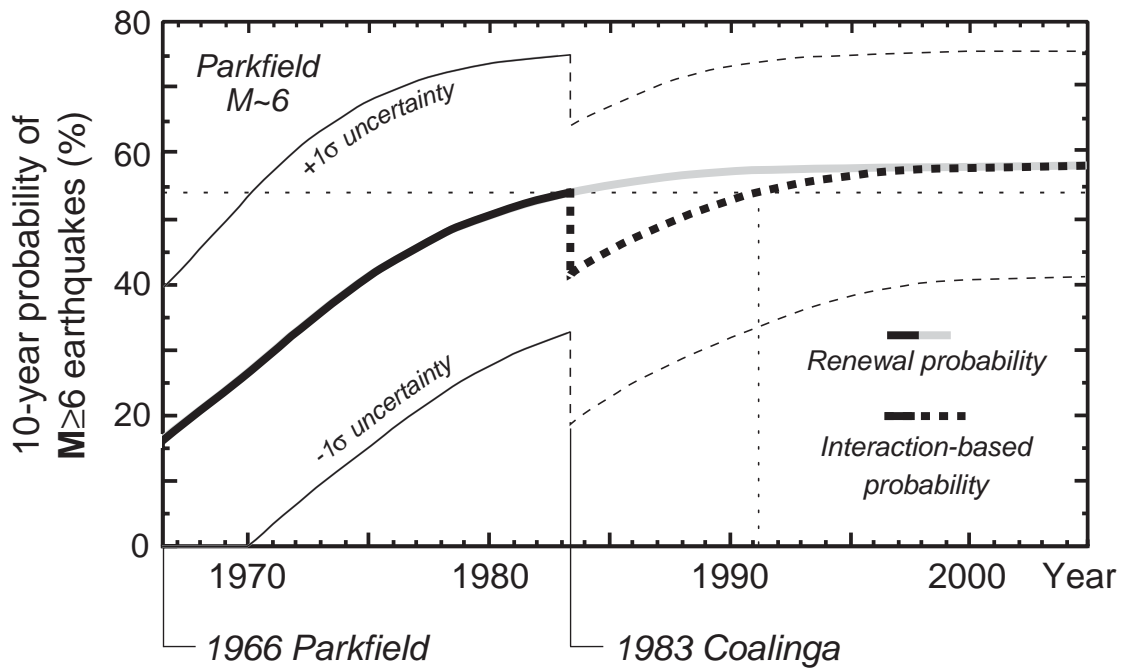


Fig. 12
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