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Absence of earthquake correlation with Earth tides: An indication of high preseismic fault stress rate

John E. Vidale
Department of Earth and Space Sciences, University of California, Los Angeles

Duncan Carr Agnew
Institute of Geophysics and Planetary Physics, University of California, San Diego

Malcolm J. S. Johnston and David H. Oppenheimer
U.S. Geological Survey, Menlo Park, California

Abstract. Because the rate of stress change from the Earth tides exceeds that from tectonic stress accumulation, tidal triggering of earthquakes would be expected if the final hours of loading of the fault were at the tectonic rate and if rupture began soon after the achievement of a critical stress level. We analyze the tidal stresses and stress rates on the fault planes and at the times of 13,042 earthquakes which are so close to the San Andreas and Calaveras faults in California that we may take the fault plane to be known. We find that the stresses and stress rates from Earth tides at the times of earthquakes are distributed in the same way as tidal stresses and stress rates at random times. While the rate of earthquakes when the tidal stress promotes failure is 2% higher than when the stress does not, this difference in rate is not statistically significant. This lack of tidal triggering implies that preseismic stress rates in the nucleation zones of earthquakes are at least 0.15 bar/h just preceding seismic failure, much above the long-term tectonic stress rate of 10^{-4} bar/h.

1. Introduction

Of all the possible triggers for earthquake rupture, a special place is occupied by the Earth tides, since these are, almost all of the time, the largest contribution to temporal variations in crustal stress. Gravitational interaction between the Earth, Moon, and Sun deforms the Earth [*Melchior*, 1983], with oscillatory stresses in the crust of up to 30 mbar at diurnal and semidiurnal periods; Figure 1 shows a typical example. Strain and seismic measurements show that the average stress drop in earthquakes is much higher than tidal stresses: 1 to 100 bars [*Johnston et al.*, 1987; *Kanamori and Anderson*, 1975; *Kikuchi*, 1992]. Since the time between seismic ruptures is typically 10 years (10^5 hours) or more, the implied long-term rate of stress loading is no more than 1 mbar/h, and often much less. The stress rates associated with the tides (Figure 1) can easily be much larger, and these higher tidal rates imply that any particular failure stress almost always be reached for increasing tides: tidal triggering would then be common. Measurement of the level of triggering (if any) thus provides a valuable clue to what conditions initiate earthquakes, and addresses the question of the predictability of earthquakes.

Given the plausibility of short-period oscillatory stresses in the crust triggering earthquakes, correlations between tides and earthquakes have been investigated repeatedly; *Emter* [1997] gives a good review of recent work, and *Cotton* [1922] of the older literature. While there are many claims that tidal triggering has been found, those papers which pay the most attention to careful statistical analysis are also those which find no evidence of triggering, for example *Heaton* [1982] and *Rydelek et al.* [1992]. In one of the most thorough of recent studies, *Tsuruoka et al.* [1995] found no evidence of triggering except for normal-faulting earthquakes near mid-ocean ridges, though, as *Emter* [1997] points out, their statistical evaluations may be flawed by the excessive subdivision of their dataset. In this paper we examine the level of tidal triggering using a large catalog of earthquakes from California, and discuss the implications of our results for earthquake nucleation.

2. Data: Earthquakes and Tides

In order to get reliable statistics on tidal triggering we need a large number of earthquakes; if we are to use actual stresses on the fault plane, we also need to know the focal mechanism and ideally the rupture plane. The earthquakes we use are 6796 events from the San Andreas fault near Parkfield and 6246 on the Calaveras fault, for a total of 13,042. We use events from 1969 through 1994 within 1 km of the fault surface; Figure 2 shows the spatial distribution. We selected simple segments of the faults and events very close to the fault plane to make the assumption of a known fault plane more accurate; 85% of the first motions (when determined) are consistent with right-lateral slip on the fault nearby, so we assume the catalog to be little contaminated by earthquakes with other mechanisms. We removed clustered events [*Reasenberg*, 1985] because it is likely that earthquake afterslip overwhelms the small tidal stress changes, and because the clustering complicates any later analysis [*Young and Zürn*, 1979]. The catalog includes events as small as magnitude 0, but most are above magnitude 1. (We

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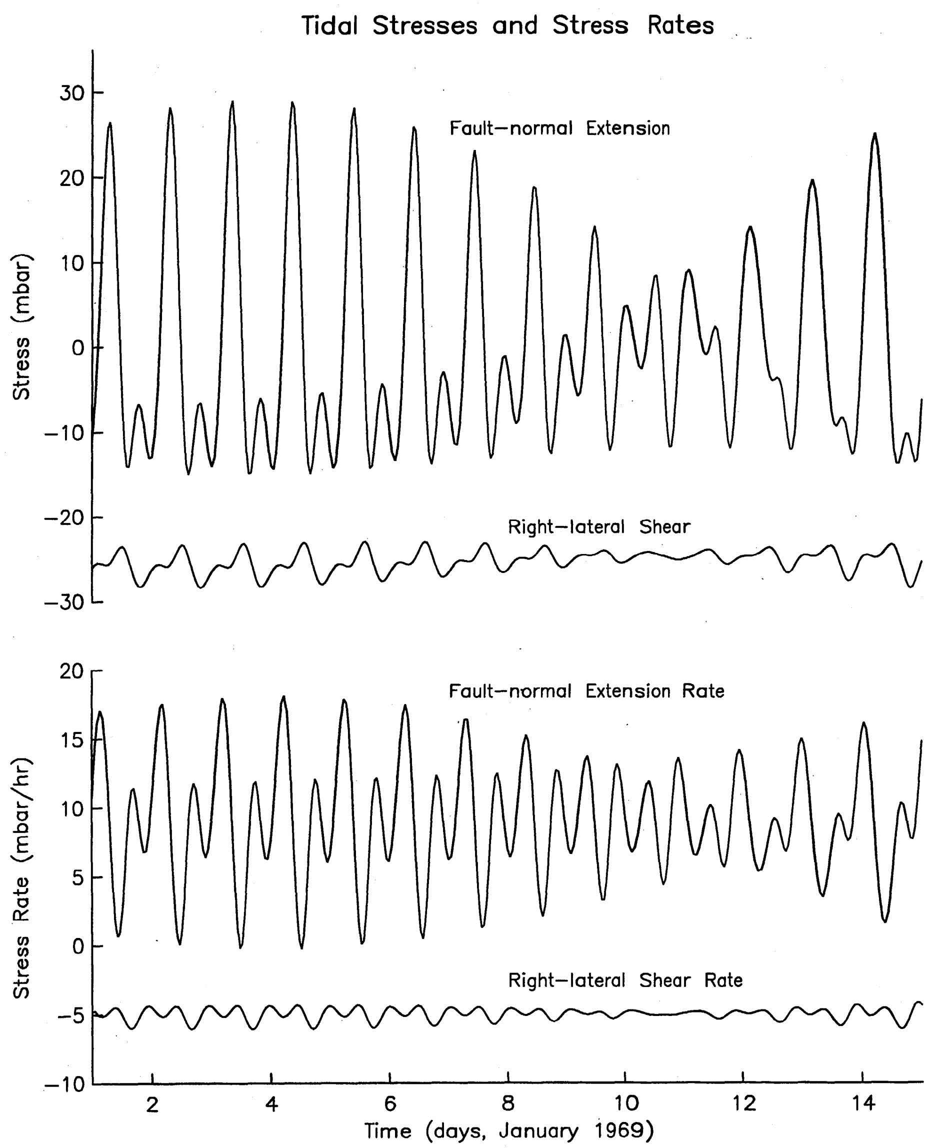

Figure 1. The shear and normal horizontal stresses and stress rates along the San Andreas fault for 15 days calculated for a spherical elastic Earth, with ocean loading included. The stresses are calculated at latitude 36.05° N, longitude 120.69°W, for a fault striking N48W, starting at 0:00 on January 1, 1969. Right-lateral stress is positive, and extensional stress is positive. The stress varies little with depth.

have also analyzed only the events of sufficient magnitude that the catalog is complete since the beginning of 1969, which reduces the number of events to 5297, and we reach the same conclusions.)

Given that we know the fault plane, we can compute the shear and normal tidal stress (σ_s and σ_n) on this plane; laboratory experiments [*Byerlee*, 1978] suggest that some combination of these stresses is what influences fault rupture. As Figure 1 shows, the tidal fluctuations of these two stresses have quite different amplitudes and are not in phase, so we must consider them separately; in addition, we consider the Coulomb stress, $\sigma_c = \sigma_s + 0.6\sigma_n$. We also consider the rate of change of these stresses ($\dot{\sigma}_s$, $\dot{\sigma}_n$, and $\dot{\sigma}_c$), since if failure is associated with reaching a critical stress level, we might expect most failures to start when the tidal stress is increasing rather than decreasing [see *Souriau et al.*, 1982].

We computed the tidal stresses and stress rates at the time and place of each of the 13,042 earthquakes in the catalog; this calculation included both the body and load tides, the load being computed for the CSR3.0 global-tide model plus a local model for the tides of San Francisco Bay [*Agnew*, 1997]. The fault planes were assumed to be oriented along the strike of the fault in the vicinity of each earthquake. To provide a comparison series, we also computed the stresses for the same location and fault plane at 20 times chosen randomly within the 26-year span of the catalog, giving a total of 260,480 comparison events.

3. Hypothesis Testing

To proceed further, we decide what we mean by the term "tidal triggering"; we need to formulate a specific (and testable) hypothesis. We take the the null hypothesis (no triggering) to be that the earthquakes occur as a Poisson pro-

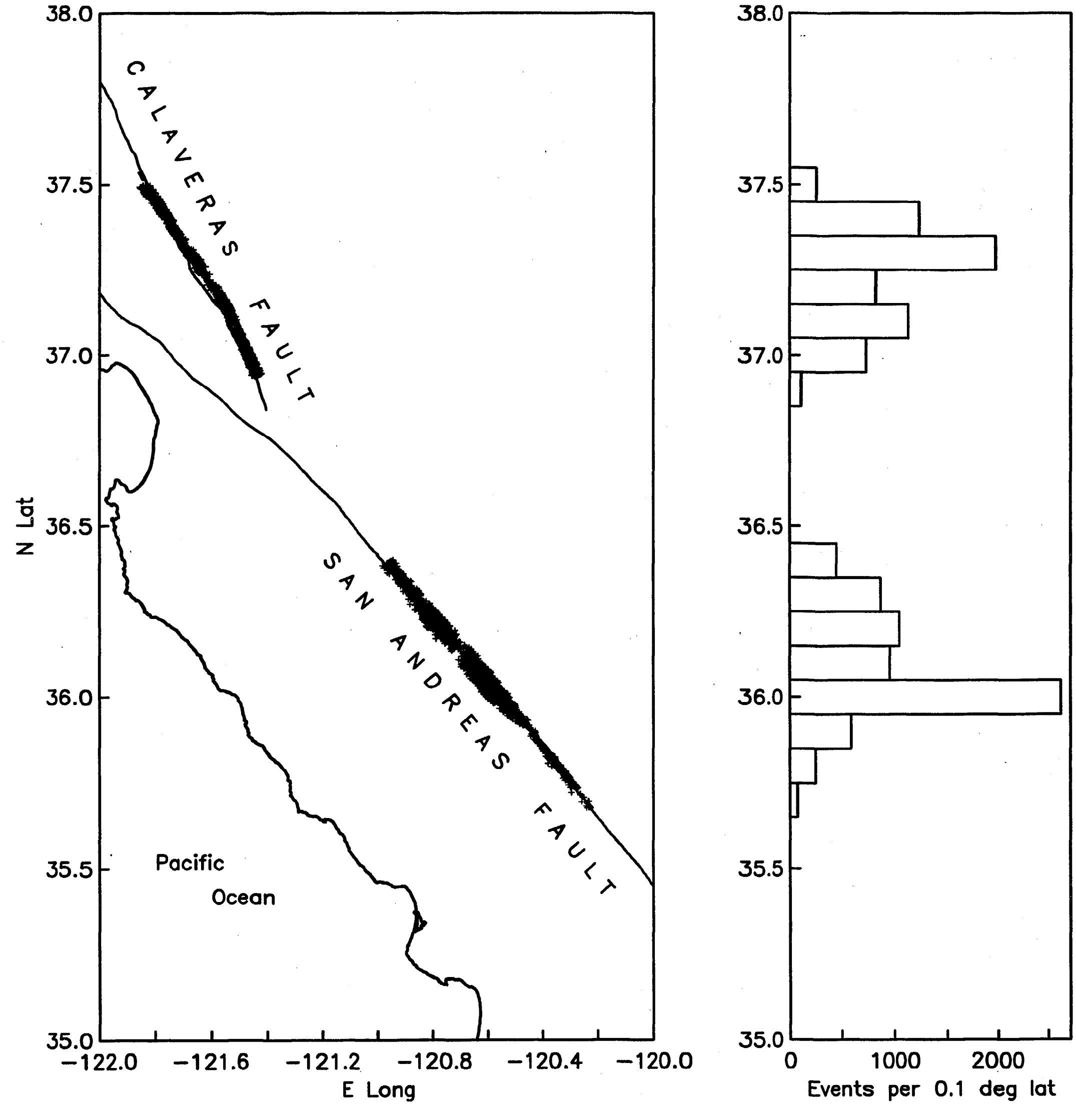

Figure 2. Distribution of the earthquakes along the San Andreas and Calaveras faults, shown in map view on the left and as a histogram on the right. All earthquakes were recorded by the U.S. Geological Survey Northern California Seismic Network; the events chosen had at least 8 P-wave arrivals from stations with a maximum azimuthal gap of 180°, and with root-mean-square misfit of 0.3 seconds or less, computed standard errors in location of 2.5 km or less in the horizontal, and 5.0 km or less in the vertical.

cess, with the probability of an earthquake in a short time dt being $\lambda_m dt$, where λ_m is the time-invariant rate of occurrence (intensity in the statistics literature). A fairly general hypothesis for tidal triggering would be that this rate is time-variable and can be written as

$$\lambda(t) = \lambda_m R(\boldsymbol{\sigma}(t))$$

where $\boldsymbol{\sigma}$ is a vector of those tidal stresses relevant to failure; this could include functions of the stress (such as stress rates). The function R gives the dependence of rate on these stresses; if R is constant (independent of the tidal stress), then there is no tidal triggering; R must have a mean value of 1.0 for the long-term rate to match the Poisson rate. Obviously, to see if R is constant we have to first decide on the components of $\boldsymbol{\sigma}$. Once this is done, the best estimate of $R(\boldsymbol{\sigma})$ is the ratio $p_E(\boldsymbol{\sigma})/p(\boldsymbol{\sigma})$, where $p(\boldsymbol{\sigma})$ is the probability density function (pdf) of $\boldsymbol{\sigma}$ for all times, and $p_E(\boldsymbol{\sigma})$ is the probability density function of $\boldsymbol{\sigma}$ at the times of earthquakes. (That is, we expect more or fewer earthquakes for some value of $\boldsymbol{\sigma}$ depending on $R(\boldsymbol{\sigma})$.) Even for the limited number of independent components of $\boldsymbol{\sigma}$ we consider here (σ_s, σ_n, $\dot{\sigma}_s$, and $\dot{\sigma}_n$), these pdf's would need to be estimated in four dimensions, and the reliable estimation of density functions in large numbers of dimensions requires many sample points [*Silverman*, 1986]. We settle instead for projecting $p(\boldsymbol{\sigma})$ and $p_E(\boldsymbol{\sigma})$ into univariate distributions as a function of σ_s, σ_n, σ_c, and their rates; such comparison of univariate distributions was developed by *Shudde and Barr* [1977] and *Young and Zürn* [1979]. Figure 3 shows the pdf's (approximated by histograms) for the stresses and stress rates. A visual comparison suggests, and a two-sample Kolmogorov-Smirnov test [*Press et al.*, 1992] confirms, that the distributions do not differ significantly.

This might be regarded as a nonparametric test for tidal triggering: any difference between the distributions would indicate triggering of some kind. We may develop more quantitative bounds if we make our hypothesis more specific. We next

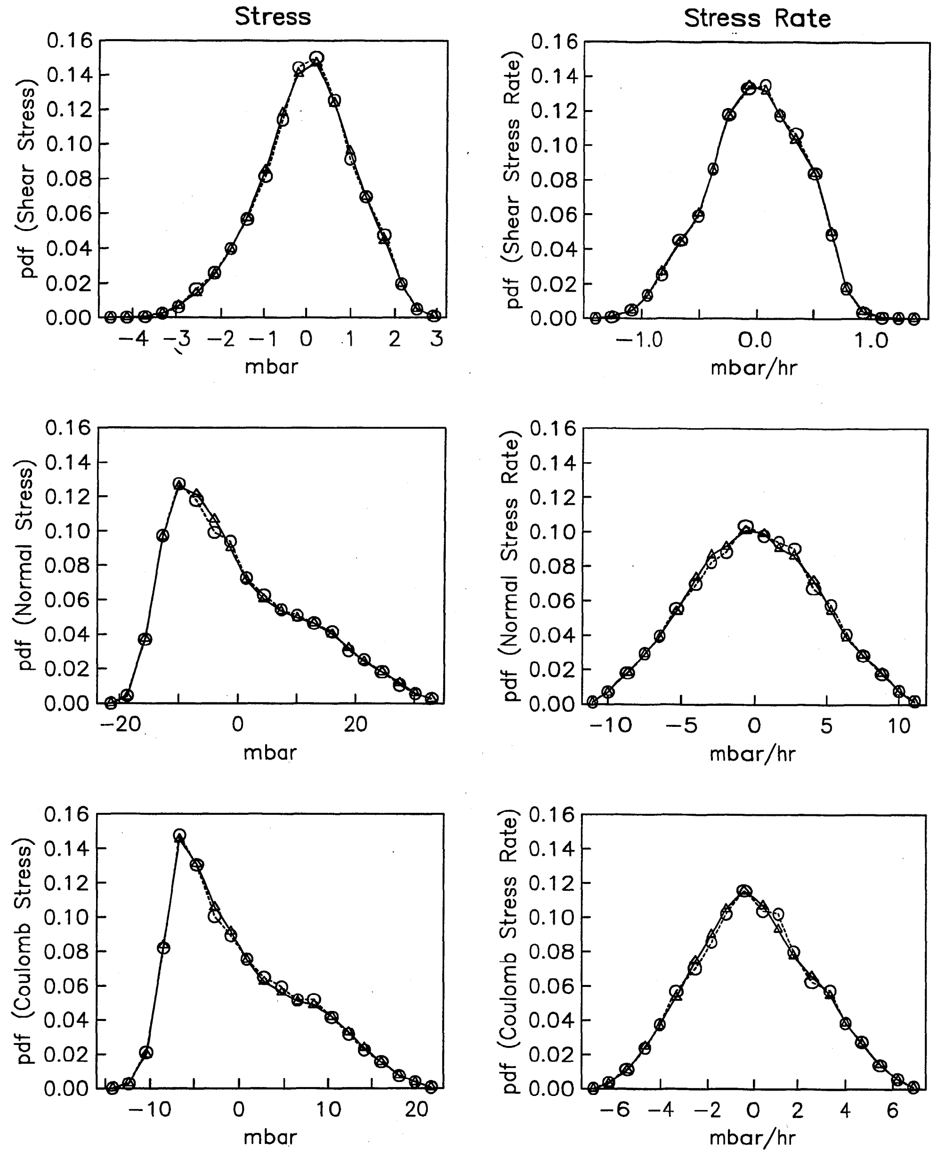

Figure 3. Comparison of the probability distribution of tidal stress and stress rate at 260,840 random times (solid line with triangles) and at the 13,042 times of the earthquakes on the Calaveras and San Andreas faults (dotted line with circles). The probability distribution function has been approximated by binning the data; the values are plotted at the center of each bin. In each plot positive stress or stress rate favors fault rupture.

suppose, partly following *Souriau et al.* [1982], that the rate λ takes on two constant values depending on σ. In the univariate case, take the relevant stress or stress rate to be σ; our model is then

$$R = \begin{cases} R^+ & \sigma \geq 0 \\ R^- & \sigma < 0 \end{cases}$$

and the level of triggering is given by the ratio R^+/R^-, which we denote by S_b: a "binomial" model, with S_b equal to 1 for the case of no triggering. We would expect on physical grounds that S_b would be greater than 1; stresses promoting failure would increase the probability of earthquakes.

As this is a two-rate Poisson model, the maximum likelihood estimate for each rate is given by the number of events divided by the time [*Cox and Lewis*, 1966]. The best estimate for S_b is thus

$$S_b = \frac{N^+}{N-N^+} \frac{1-f^+}{f^+} \quad (1)$$

where N^+ is the number of earthquakes which occur for $\sigma > 0$, N the total number of events, and f^+ the fraction of the time that $\sigma > 0$. (The distribution of σ for random times gives f^+.) To get confidence limits for S_b we assume that N^+ in equation (1) is a random variable $\hat{N}^+$, namely the number of successes in a binomial distribution with probability p, where $p = N^+/N$. For p close to 0.5 (as is true here) and N large, $\hat{N}^+$ is, to a good approximation, distributed as a normal variable with mean N^+ and variance $Np(1-p)$ [*Rice*, 1988], so the 95% confidence limits on S_b are given by substituting

$$N^+ \pm 1.96N^{1/2}(p(1-p))^{1/2} \quad (2)$$

into equation (1).

Table 1. Results for Binomial Model of Triggering

Stress	f^+	N^+	N_{ex}	S_b (low)	S_b	S_b (high)	T_m	T_h
σ_s	0.507924	6661	36	0.9772	1.0113	1.0467	177.8	43.9
σ_n	0.417722	5513	64	0.9857	1.0207	1.0567	97.7	36.3
σ_c	0.419236	5559	90	0.9939	1.0291	1.0654	69.7	31.6
$\dot{\sigma}_s$	0.511328	6709	39	0.9783	1.0124	1.0478	161.9	42.8
$\dot{\sigma}_n$	0.498217	6579	80	0.9906	1.0252	1.0610	80.3	33.8
$\dot{\sigma}_c$	0.485486	6428	95	0.9952	1.0300	1.0660	67.7	31.3

$N_{ex} = N^+ - f^+N$ is the "excess" number of earthquakes for positive stress over the number expected if the distribution is random.

Table 1 shows the results of this computation for 6 possibilities: shear, normal, and Coulomb stress, and their rates of change. While in all cases the best estimate of S_b gives a value slightly above 1, at the 95% confidence level we cannot say that S_b is different from 1: we have a suggestion of a slight amount of tidal triggering, but not clear evidence of it. Of course, if the extent of triggering is this slight, we would need a large number of events to conclusively prove its existence. For example, if S_b were really 1.02, we would need 68,000 events to be able to show that S_b was different from 1 at the 99% confidence level. With the restrictions we have applied in compiling this catalog, getting this many events would require another 110 years of observations.

4. Implications for Earthquake Nucleation

Given the observed upper bound to the amount of tidal triggering, S_b, we can estimate a lower bound on the stress rate that loads faults just prior to failure. If the loading rate were comparable to the tectonic rate (much less than the tidal rate), and if failure occurs soon after a critical stress has been attained, then we would expect earthquakes to occur at peak tides, which they clearly do not. For higher rates of loading, the effect of the tides would be less in the amplitude of stress than in the rate of stress change: the probability of crossing a threshold to failure would be proportional to the total rate of stress, which is the loading rate $\dot{\sigma}_b$ plus the tidal stress rate $\dot{\sigma}_t$ (Figure 4). Roughly speaking, the ratio of earthquake rates for positive and negative tidal stress rate would be the ratio of the average total rates of stress:

$$S_b = \frac{\dot{\sigma}_b + \langle\dot{\sigma}_t^+\rangle}{\dot{\sigma}_b - \langle\dot{\sigma}_t^-\rangle} \tag{3}$$

where $\langle\dot{\sigma}_t^\pm\rangle$ are the average of the positive and negative tidal stress rates. When, as in our case, $\langle\sigma_t^+\rangle = \langle\sigma_t^-\rangle$, (3) becomes

$$\dot{\sigma}_b = \left[\frac{S_b + 1}{S_b - 1}\right] \langle\dot{\sigma}_t^+\rangle \equiv T\langle\dot{\sigma}_t^+\rangle \tag{4}$$

The variable T gives the factor by which the prerupture stress rate must exceed the tidal stress rate to give as low a level of triggering as is observed. Table 1 gives the values of T for the best estimate and upper 95% bound of S_b; these are T_m and T_h. (The lower bound on S_b is consistent with no triggering, which would imply that T is infinite.) The lowest values of T fall in the range of 30-40; for the fault-normal stress $\langle\dot{\sigma}_t^+\rangle$ is about 3.2 mbar/h (Figure 3), implying that $\dot{\sigma}_b$ is at least 0.15 bar/h. Long-term tectonic stress rates are 1000 times slower than this rate. Thus, if the attainment of a critical stress instantaneously triggers earthquakes, the very low amounts of tidal triggering imply that the rate of loading just before failure is higher than the tidal rate, and thus much higher than the tectonic loading rate. Because the number of events examined here is much larger than in previous investigations, the constraint on the minimum loading rate is much stronger than in most earlier studies; many prior studies were not able to use the amplitudes of tidal strains and so could not constrain stress rates at all. *Rydelek et al.* [1992] did obtain a similar constraint to the one here, but had a much smaller ratio between long-term stress rate and inferred prefailure loading rate because they were looking at seismicity in the rapidly deforming Campi Flegri in Italy, rather than along a plate transform boundary.

Of course, our results apply to small earthquakes, although tests with larger events generally show similar results. Hazard mitigation is concerned with larger earthquakes. The conventional viewpoint is that all earthquakes start in a similar fashion, but some grow bigger than others [*Abercrombie and Mori*, 1994; *Mori and Kanamori*, 1996]. If this is true, our finding of no detectable tidal triggering would also apply to large events. There is some evidence that the approach to failure [*Knopoff et al.*, 1996] and the initial phase of seismic rupture [*Ellsworth and Beroza*, 1995] may differ between large and small events, which may complicate extrapolation of our research to the largest earthquakes.

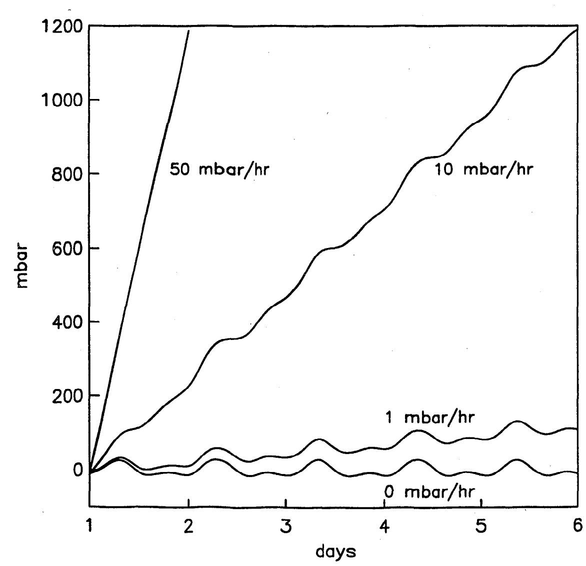

Figure 4. The total stress for different rates of loading stress, with the tidal stress (from Figure 1) included. As the steady rate grows, the effect of the tides is lessened.

Static stress changes of 0.1 bar or more have been shown to trigger seismicity [*Das and Scholz*, 1981; *King et al.*, 1994; *Reasenberg and Simpson*, 1992; *Stein and Lisowski*, 1983] and inferred to change the probability of future, damaging earthquakes [*Stein et al.*, 1994]. Two existing theories can explain the high stress rates just before failure. Dieterich's model of state- and rate-dependent friction predicts high stressing rates across earthquake nucleation zones [*Dieterich*, 1987], and changes in fluid plumbing of the fault system could conceivably be more rapid than tidal strains and may trigger failure [*Sibson*, 1973]. The scale length of the preseismic slip is of course not determinable from the methods used here; so even if this slip occurs, it may be invisible to current instrumentation [*Vidale*, 1996].

In summary, we have examined a large set of earthquakes for which the rupture planes can be assumed to be known; we find, consistent with many earlier studies, that there is no clear effect of the tidal stress change on when earthquake rupture starts. We infer from this that prior to such rupture, the stress rate near where the rupture begins must be much higher than the tidal stress rate, and hence considerably larger than the tectonic stress rate, even though it is the latter that creates most of the stress buildup that is released by earthquakes.

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D. C. Agnew, Institute of Geophysics and Planetary Physics 0225, University of California San Diego, La Jolla, California 92093-0225 (e-mail: dagnew@ucsd.edu)

M. J. S. Johnston and D. H. Oppenheimer, U.S. Geological Survey, MS 977, 345 Middlefield Rd, Menlo Park, California 94025 (e-mail: mal@andreas.wr.usgs.gov; oppen@alum.wr.usgs.gov)

J. E. Vidale, Department of Earth and Space Sciences, University of California Los Angeles, Los Angeles California 90095-1567 (e-mail: vidale@ess.ucla.edu)

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