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Foreshock Occurrence Rates before Large Earthquakes Worldwide

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Abstract—Global rates of foreshock occurrence involving shallow $M \geq 6$ and $M \geq 7$ mainshocks and $M \geq 5$ foreshocks were measured, using earthquakes listed in the Harvard CMT catalog for the period 1978–1996. These rates are similar to rates ones measured in previous worldwide and regional studies when they are normalized for the ranges of magnitude difference they each span. The observed worldwide rates were compared to a generic model of earthquake clustering, which is based on patterns of small and moderate aftershocks in California, and were found to exceed the California model by a factor of approximately 2. Significant differences in foreshock rate were found among subsets of earthquakes defined by their focal mechanism and tectonic region, with the rate before thrust events higher and the rate before strike-slip events lower than the worldwide average. Among the thrust events a large majority, composed of events located in shallow subduction zones, registered a high foreshock rate, while a minority, located in continental thrust belts, measured a low rate. These differences may explain why previous surveys have revealed low foreshock rates among thrust events in California (especially southern California), while the worldwide observations suggest the opposite: California, lacking an active subduction zone in most of its territory, and including a region of mountain-building thrusts in the south, reflects the low rate apparently typical for continental thrusts, while the worldwide observations, dominated by shallow subduction zone events, are foreshock-rich.

Key words: Foreshock, foreshock rates, earthquake clustering.

Introduction

Short-term earthquake clustering, including the occurrence of foreshocks and aftershocks, is a widely observed phenomenon in shallow crustal seismicity. Nearly half of all earthquakes (most of them aftershocks) are included in short-term clusters (REASENBERG, 1985; DAVIS and FROHLICH, 1991; OGATA *et al.*, 1995). Moderate or strong earthquakes are occasionally followed by stronger shocks nearby within a few days. This clustering makes possible short-term probabilistic earthquake forecasts after any earthquake based on triggered stochastic models (KAGAN and KNOPOFF, 1987) in which the occurrence times of the triggered events are Poissonian (with varying rate), while their rate, magnitude distribution and geographic distribution are empirically determined from the clustering behavior observed in the historic seismicity. After a moderate earthquake in an urbanized

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area, the increased probability of an imminent, larger earthquake is of great concern. For example, after a moderate ($M = 5 \sim 6$) earthquake in either San Francisco or Los Angeles areas, where large earthquakes are considered likely in the long term, there follows a transient probability gain for a ($M \geq 7$) earthquake of $10^2 \sim 10^3$, relative to the respective regional long-term probabilities. In absolute terms, the conditional probability of such an earthquake ranges from a few tenths of a percent to a few percent, and has a half-life of about 1 day. The corresponding, long-term daily probabilities in these areas are approximately 0.01% and 0.004% per day, respectively.

Two approaches have been taken to develop triggered stochastic models of earthquake occurrence. The first is based directly on the observed frequency of foreshock-mainshock (*fs-ms*) pairs. Empirical estimates of the *fs-ms* pairing rate and the distribution of *fs-ms* magnitude differences have been made with numerous earthquake catalogs (JONES and MOLNAR, 1979; VON SEGGERN, 1981; JONES, 1984; BOWMAN and KISSLINGER, 1984; LINDH and LIM, 1995; ABERCROMBIE and MORI, 1996; MICHAEL and JONES, 1998). These measurements, which characterize the clusters' transient, pair-wise interactions, are analogous to the a-value and b-value in the Gutenberg-Richter (time-independent, noninteracting events) seismicity model. Such foreshock studies provided the basis for conditional probability estimates for the next Parkfield, California, earthquake (MICHAEL and JONES, 1998), and for characteristic earthquakes on selected fault segments in California (AGNEW and JONES, 1991).

In another approach, REASENBERG and JONES (1989, 1994) introduced a stochastic model of aftershock occurrence based on the observed rate (a-value), magnitude distribution (b-value) and temporal decay (p-value) of 62 aftershock sequences following $M \geq 5$ mainshocks in California. They assumed self-similarity of clustering in their "California generic model," and allowed the modeled "aftershocks" to take on magnitudes larger than their "mainshock"—in effect, applying an aftershock model to foreshock-mainshock pairing. This approach was advantageous in that the model and its uncertainty were readily determined (even in a limited region such as California) owing to the large number of recorded mainshock-aftershock sequences. To estimate the expected numbers of $M > 3$ aftershocks following the 1994 Northridge, California earthquake, the California generic model worked reasonably well (REASENBERG and JONES, 1994). However, in the type of application for which the interest is greatest—the case of potential $M > 5$ foreshocks prior to a larger earthquake in California—the model has not been directly tested through comparisons to observed earthquake occurrence, because insufficient numbers of such pairs have been recorded. Recently, the question was raised as to whether the California generic model accurately calculates the average, short-term conditional probability of large earthquakes in California (CALIFORNIA EARTHQUAKE PREDICTION EVALUATION COUNCIL, 1977).

Here, I use a worldwide earthquake catalog to estimate rates of *fs-ms* pairing and compare these rates to the corresponding probabilities calculated with the California generic model. Because the goal is to test a model derived from (and applied to) California seismicity, which is largely confined to the upper crust, I use only shallow earthquakes in the global catalog. The empirical foreshock rates are determined separately for mainshocks with $M \geq 6$ and $M \geq 7$, for mainshocks with thrust, normal and strike-slip mechanisms. The observed rates are compared to the California generic model probabilities and their ranges of uncertainty. Then, I examine the magnitude differences among the foreshock-mainshock pairs involving $M > 7$ mainshocks in the Harvard catalog, and attempt to use them to constrain a model for their distribution.

Data

Data for this study were taken from the Harvard catalog of centroid moment tensor solutions for large earthquakes (DZIEWONSKI *et al.*, 1987). For this study, I use M_s for magnitude when it is given (90% of the shallow events), and M_b when M_s is not given (10% of the shallow events), and refer to this magnitude simply as magnitude (M) (Fig. 1). The Harvard CMT catalog is complete from 1977–1996 for $M_w \geq 5.3$ or $M_w \geq 5.5$ events (KAGAN and KNOPOFF, 1980, and this study). Within the magnitude range $M = 5$ to $M = 5.5$, M and M_w are in close agreement, having a maximum cumulative difference of less than 0.1, so that the estimated levels of completeness for M_w may be assumed for M. For this study I selected earthquakes with $M \geq 5.0$, below the completeness levels of the catalog. This choice requires justification. The motivation for it stems from the ultimate aim of the study, which is to test the assumption that an aftershock model (REASENBERG and JONES, 1989) may be used to forecast larger earthquakes (i.e., may be used as a foreshock-mainshock model). Because such an application is currently carried out by the USGS for all $M \geq 5$ earthquakes in California, I tailored the study to this magnitude range. The expected effect of choosing the foreshock magnitude cutoff below the completeness level is to underestimate the actual foreshock rates. I estimated with a graphical method (straight-line fit to log cumulative counts between $M = 5.5$ and $M = 6.2$) that approximately 28% of the earthquakes ($M \geq 5.0$) are missing from the Harvard catalog. Hence, the retrospective foreshock rates presented below may be underestimated by this amount.

For the 5695 events ($M \geq 5.0$; depth ≤ 50 km) taken from the Harvard CMT catalog, I used the hypocentral locations listed as having originated from either NEIS or ISC. From 1977–1992 I used only ISC hypocenters (SMITH, 1995). Between 1993 and 1996, I used the combination of ISC and NEIC hypocenters listed in the Harvard catalog, which consists primarily of NEIS locations. Because many of the depths are fixed or poorly determined, depth information was used only in the selection of events to be included in the study; calculated distances between events are epicentral distances. The choice of a 50-km depth cutoff, while somewhat arbitrary, accommodates our desire to work with "shallow" events and includes the fixed depths of 10 and 33 km assigned to numerous events in the Harvard catalog for which more precise depth determination was impossible. Aside from the cut at 50 km, depths were not considered in this study.

Definition of Foreshocks

Foreshock-mainshock pairs used in this study were defined according to the following protocol:

1. The magnitudes of the first and second events in each pair (M_f and M_m, respectively) exceed fixed thresholds: $M_f \geq M_f^{\min}$ and $M_m \geq M_m^{\min}$.
2. The second event is not smaller than the first: $M_f \leq M_m$.

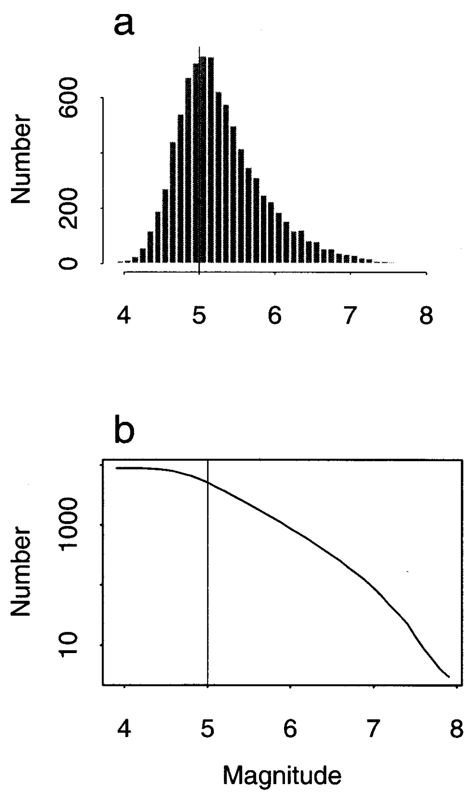


Figure 1
Magnitude distribution of earthquakes used in this study and taken from the Harvard CMT catalog, 1977–1996, 0–50 km depth. (a) Interval distribution; (b) cumulative distribution.

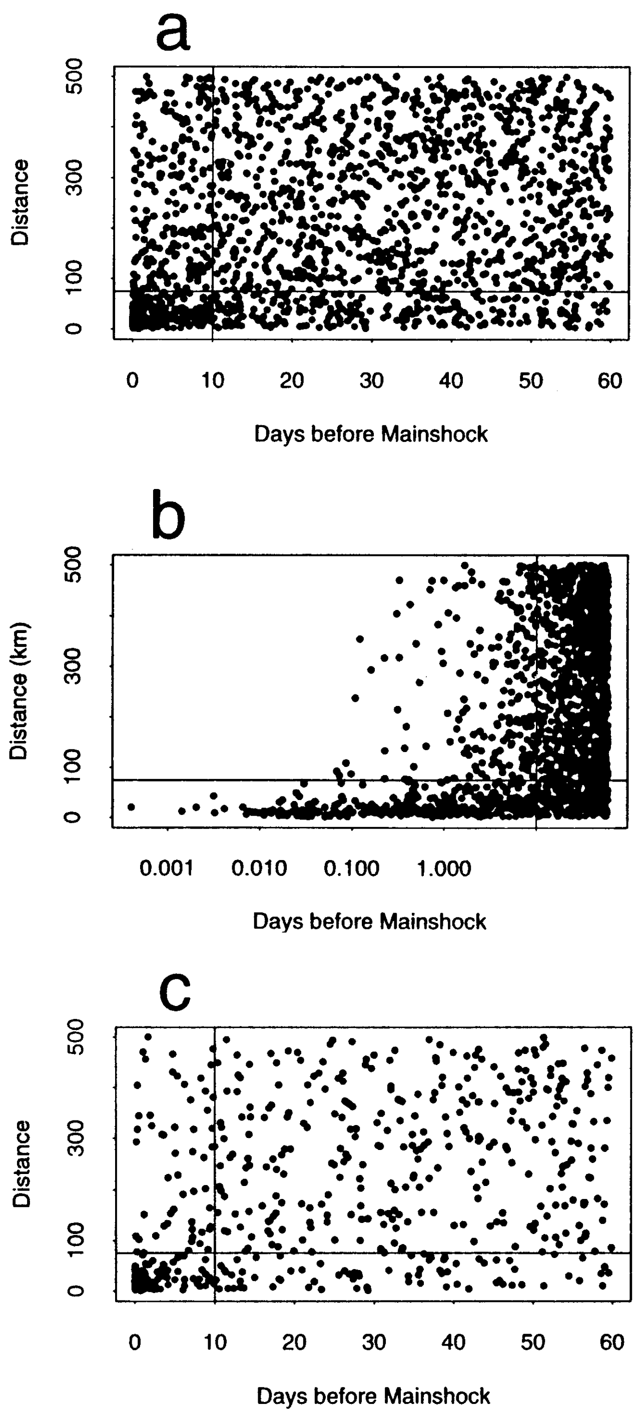
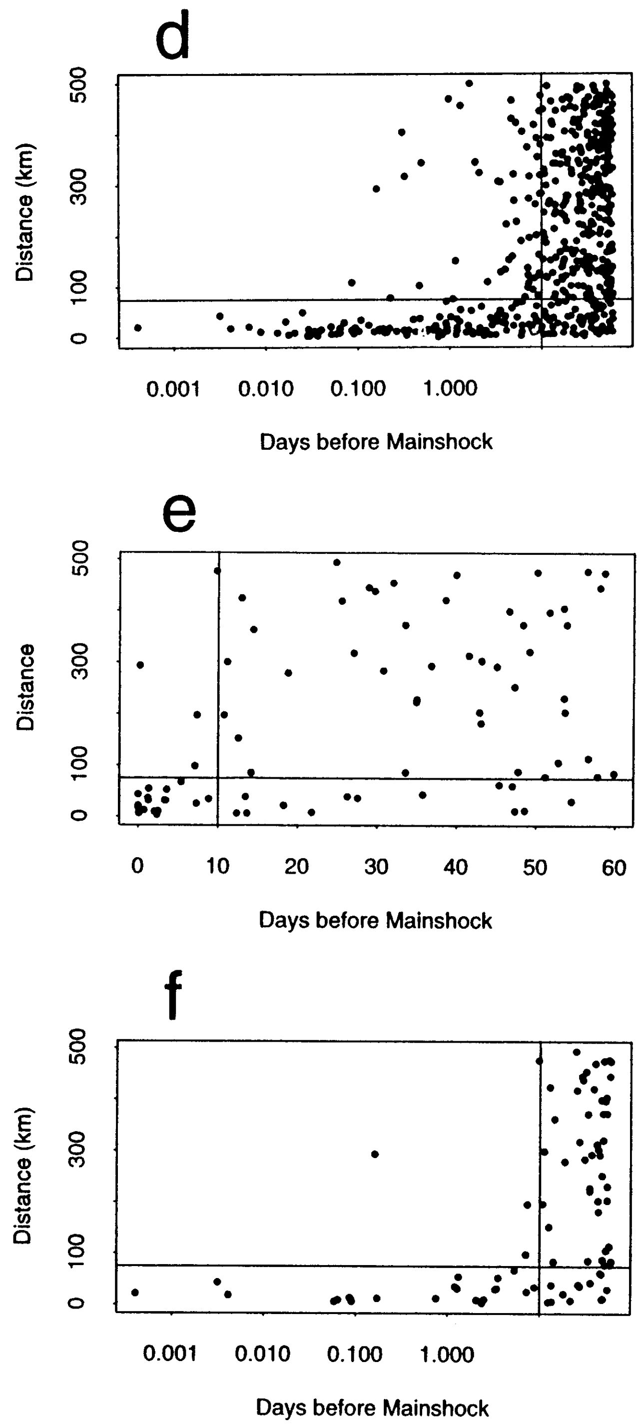

a
Distance
0
100
300
500
0
10
20
30
40
50
60
Days before Mainshock
b
Distance (km)
0
100
300
500
0.001
0.010
0.100
1.000
Days before Mainshock
c
Distance
0
100
300
500
0
10
20
30
40
50
60
Days before Mainshock

Fig. 2.


d
Distance (km)
0
100
300
500
0.001
0.010
0.100
1.000
Days before Mainshock
e
Distance
0
100
300
500
0
10
20
30
40
50
60
Days before Mainshock
f
Distance (km)
0
100
300
500
0.001
0.010
0.100
1.000
Days before Mainshock

3. Interevent epicentral distance in a pair does not exceed dX and the interevent time difference does not exceed dT.
4. When two or more mainshocks are themselves clustered within dX and dT of each other, only the largest is considered a mainshock.

In the estimates of foreshock rates below, tied magnitudes are allowed in the above definition, in conformance with the definition in REASENBERG and JONES (1989, 1994), so that the present results should be directly comparable to their model. However, in the subsequent analysis of the distribution of magnitude differences among foreshock-mainshock pairs, I require that $M_f < M_m$, in conformance with the definition used by AGNEW and JONES (1991), LINDH and LIM (1995) and MICHAEL and JONES (1998), which facilitates comparison with those studies. For each qualifying mainshock, all qualifying events within the time dT and distance dX of it are identified as potential foreshocks. If more than one foreshock is thus associated with a given mainshock, only the largest foreshock is counted.

Choice of Spatial and Temporal Windows

The spatial and temporal windows used to define the foreshock-mainshock pairs were selected so as to capture the strongest and most obvious foreshock clustering activity in the Harvard CMT catalog. Figure 2 shows the earthquake pairing corresponding to three choices of M_f and M_m, with interevent separations reaching 500 km and 60 days. Figure 3 presents the corresponding cumulative distribution of interevent time and distance among the pairs. At interevent distances greater than approximately 75 km and interevent times greater than approximately 10 days, the density of pairs apparent in Figure 2 is approximately uniform with respect to both time and distance and the cumulative curves in Figure 3 are approximately linear, consistent with noninteracting earthquakes occurring uniformly in time and epicentrally distributed uniformly along fault zones (see, for example, KAGAN and KNOPOFF, 1980). At interevent distances and times less than 75 km and 10 days, the dense clustering of points in Figure 2 and steep assent of the cumulative distributions in Figure 3 reveal the foreshock clustering process. The apparent

Figure 2

Foreshock-mainshock pairs in the Harvard CMT catalog (1977–1994) for various mainshock and foreshock magnitude thresholds, shown as a function of the pair's interevent distance and interevent time. (a, c, e): linear time axis; (b, d, f): same data shown with log time axis. (a and b) Pairs with foreshock magnitude $M_f \geq 5$ and mainshock magnitude $M_m \geq 5$; $N = 1966$. (c and d) Pairs with $M_f \geq 5$ and $M_m \geq 6$; $N = 562$. (e and f) Pairs with $M_f \geq 5$ and $M_m \geq 7$; $N = 84$. The large ranges of interevent time and distance shown here reveal both the foreshock-related clustering and background or "chance" clustering. Pairs with interevent times less than 10 days and distances less than 75 km (the lower left quadrants formed by the solid lines) were used in the subsequent foreshock rate analysis.

range of foreshock-mainshock distance appears to be independent of the magnitude of the mainshock; the median and 90%-ile interevent distances are approximately 15 km and 30 km, respectively, for pairs involving $M > 5$, $M > 6$ and $M > 7$ mainshocks (Fig. 4). The consistency of these ranges suggests that they are essentially controlled by location errors and should be considered upper bounds. BOWMAN and KISSLINGER (1984) found an apparent interevent distance range (which includes an unstated location uncertainty) of 20–40 km for events located near Adak Island. OGATA *et al.* (1995) observed a range of apparent interevent distances

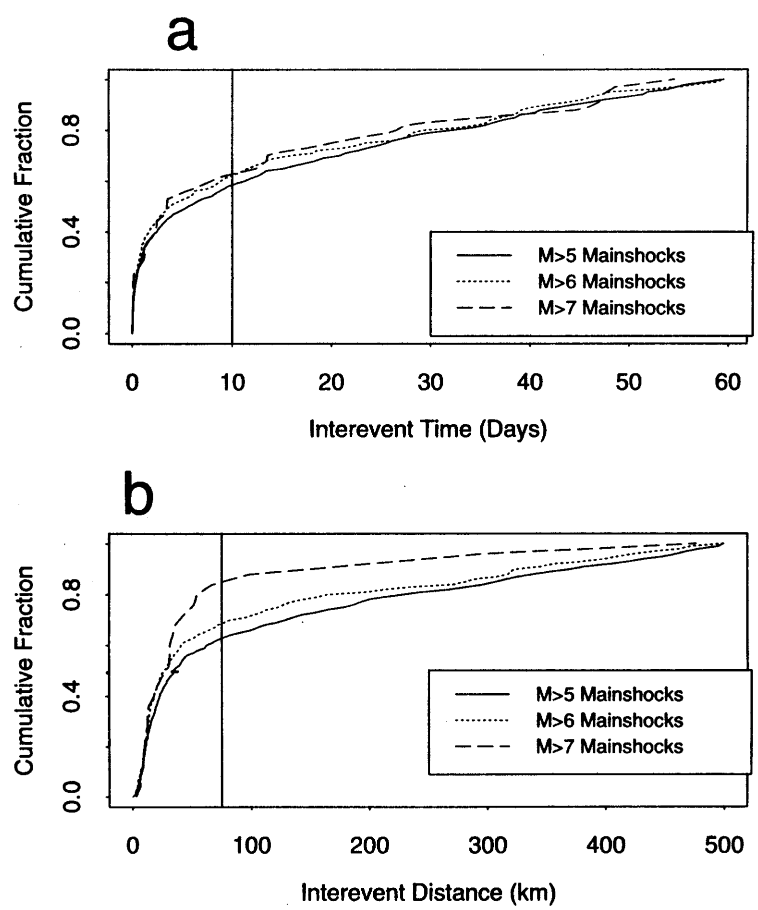


Figure 3
Interevent distances and times among foreshock-mainshock pairs found in the Harvard CMT catalog. (a) Cumulative distribution of interevent times among foreshock-mainshock pairs with interevent distances of 75 km or less. (b) Cumulative distribution of interevent distances among foreshock-mainshock pairs with interevent times of 10 days or less. Background seismicity, which contributes to the linear trends at large interevent times and distances, is not removed in this figure. Vertical lines indicate windows of 10 days and 75 km used in calculating foreshock rates.

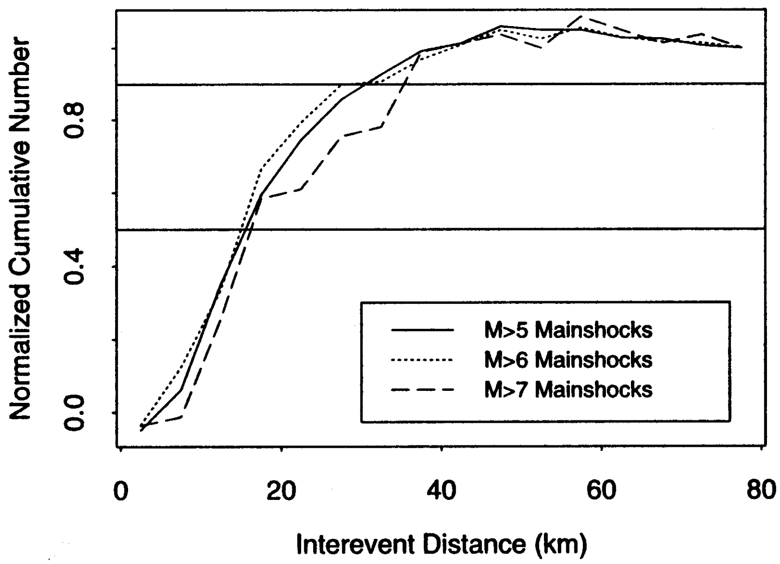


Figure 4

Normalized cumulative distributions of interevent distances among foreshock-mainshock pairs shown in Figure 3b, after correcting for "background rate." Background rate is defined as the average spatial density of pairs observed in each data set over the interevent distance range 200–500 km (and corresponds to the slopes of the flat portions of curves in Fig. 3b). Removal of these constant background rates accounts for the decreases in "cumulative" number seen in the upper tails of the distributions. Horizontal lines indicate the median and 90%-ile levels. For all three data sets (mainshocks with $M > 5$, $M > 6$ and $M > 7$), the median and 90%-ile interevent distances are approximately 15 km and 30 km, respectively. Numbers of pairs represented are 518, 161 and 26 for the $M > 5$, $M > 6$ and $M > 7$ mainshocks, respectively.

between $M \geq 4$ foreshocks and mainshocks in the JMA catalog of 10–20 km, but they warned that event location errors may be contributing to these distances. Among observations of well-located events, there is evidence that foreshocks often occur much closer to their mainshocks. Using standard catalog locations, JONES (1985) found that most foreshocks ($M > 3$) in southern California were located within approximately 1 km of the epicenters of their respective mainshocks. Using precise relative event locations (uncertainties of 0.1 to 0.3 km) obtained with a waveform correlation technique, DODGE *et al.* (1996) found that foreshocks before mainshocks ($4.7 < M < 7.3$) in California were located approximately 1 km or less from their mainshock's epicenter. Returning to the present study, the median epicentral error in the Harvard CMT catalog was estimated by SMITH and EKSTRÖM (1997) to be 25 km or greater for $M_w = 6.5$ events. Relative locations between foreshocks and mainshocks are probably better determined than this owing to common unmodeled velocity structure and station site effects. While we have neither relocated nor estimated uncertainty in the relative locations of these event pairs, it is reasonable to assume that at least some of the apparent interevent range of 15–30 km seen here reflects location errors. Clearly, the 75-km window used here is larger than necessary, and was adopted to insure that badly mislocated events would not be excluded from the analysis; it does not imply that the actual event separations are this large. To correct the resulting assays for this oversized window, a "background" rate, corresponding to chance pairings, must be estimated and removed from the pair counts.

Correction for "Background" Events

A certain number of earthquakes are expected to fall within our interevent distance and time windows due to the "background" seismicity in the regions of the mainshocks. The effect of the background seismicity is seen in the linear trends in Figure 3 at large interevent times and distances. To correct for this effect, background rates were estimated from the seismicity by counting earthquakes located within 75 km of each mainshock and occurring during the time period 300 to 100 days before each mainshock. Based on these rate estimates, the numbers of background earthquakes expected to fall into the actual foreshock windows were subtracted from the counts of possible foreshocks. This method may slightly overestimate the background rate because $M \geq 5$ aftershocks, which were not removed from the catalog, could be included in the background count, while foreshocks were limited to one per mainshock. However, inspection of the background events revealed none to be $M \geq 5$ aftershocks.

Focal Mechanisms

I categorized the mainshocks as either thrust, strike-slip or normal according to the plunge of the tension axis (p1) and plunge of the null axis (p2), as in TRIEP and SYKES (1997) (Table 1). These definitions include some oblique thrust events as "thrust," and some oblique normal events as "normal." The use of the plunges of principal axes to classify the focal mechanism leads to clear misclassifications. These arise because knowledge of which focal plane is the slip plane is not included in the CMT solution.

Table 1

Definition of focal mechanism types in Harvard CMT catalog

Focal mechanism	p1	p2
Thrust	$>45°$	—
Normal	$<45°$	$<45°$
Strike-slip	—	$>45°$

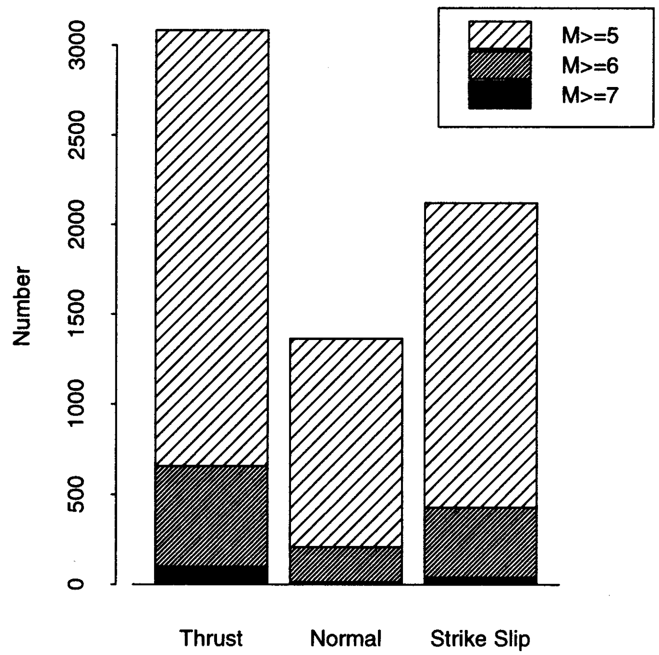


Figure 5
Earthquakes listed with depths of 50 km or less in the Harvard CMT catalog (1977–1996) and used in this study, classified according to focal mechanism and magnitude.

Figure 5 shows the distribution of earthquake focal mechanisms among the mainshocks used in this study. To characterize the *fs-ms* pairs, I used the mainshock's focal mechanism rather than the foreshock's, assuming it to be more representative of the regional faulting style. Thus, in interpreting the results, this classification may be applied to the region in which a potential foreshock is located, but may not be used to condition the probability of a mainshock based on the potential foreshock's focal mechanism.

Retrospective Foreshock Frequencies

The retrospective foreshock frequency associated with a set of mainshocks is defined as the fraction of the mainshocks that are preceded by a foreshock. To estimate this rate, I used time and distance windows of 0–10 days and 0–75 km, set the foreshock magnitude threshold $M_f^{\min} = 5$, and considered mainshock magnitude thresholds of $M_m^{\min} = 5$, $M_m^{\min} = 6$, and $M_m^{\min} = 7$. The resulting sets of foreshock-mainshock pairs have magnitude differences distributed between zero and an upper limit approximately given by $\Delta M = M_m^{\min} - M_f^{\min}$. I call ΔM the magnitude difference aperture. The larger the aperture is, the greater the number of foreshocks that will be counted. This is true for any assumed distribution of magnitude difference. Therefore, to quantitatively compare different foreshock assays, it is necessary to normalize for the magnitude apertures used. Other differences among studies, including the choice of time and spatial windows used to define the foreshocks, are of less importance, because foreshock-mainshock pairs are tightly clustered in space and time, and thus self-defining. Oversized windows include too many background events, but this error can be approximately corrected for. However, no such clustering of magnitude difference apparently exists among foreshocks and mainshocks, as shown below, and consequently variations in the observable range of magnitude difference (aperture) can be expected to have a first-order effect on the resulting assay.

To compare the present results to other studies, I assume that the magnitude differences among the foreshock-mainshock pairs have a uniform distribution over their range of observation (or aperture), and calculate a "unit foreshock rate" by dividing the apparent foreshock rate by the magnitude difference aperture.

The $M \geq 6$ mainshocks in the Harvard CMT catalog between 1978 and 1996 ($N = 1108$) have a foreshock rate ($M \geq 5$ foreshocks, corrected for estimated background seismicity) of 13.2% (Table 1). The thrust earthquakes among these ($N = 533$) have a significantly higher rate (17.5%) and the strike-slip earthquakes ($N = 397$) have a significantly lower rate (8.0%). Among the $M \geq 7$ mainshocks in the Harvard CMT catalog ($N = 149$), both the overall frequency of foreshocks (16.5%) and the departures from this rate observed among subsets defined according to focal mechanism are similar to the rates in the $M \geq 6$ set. However, because of the smaller numbers of *fs-ms* pairs, the differences among the $M \geq 7$ subsets are not statistically significant.

The overall foreshock rate of 13.2% found for $M \geq 6$ mainshocks preceded by $M \geq 5$ foreshocks (13.2% per magnitude unit) is comparable to results obtained in other studies. Jones (1984) found that 7 out of 20 ($M \geq 5$) earthquakes which occurred in California between 1966 and 1980 were preceded by ($M \geq 2$) foreshocks (a foreshock rate density of 12% per magnitude unit). Jones and Molnar's (1979) study of worldwide ($M \geq 5$) foreshock activity before 161 $M \geq 7$ mainshocks (1914–1973) detected a rate of 24.8%, or approximately 12% per magnitude unit. More recently, Abercrombie and Mori (1996) examined 59 ($M \geq 5$) mainshocks in California and Nevada, and found that 26 of them were preceded by ($M \geq 2$) foreshocks (15% per magnitude unit). Michael and Jones (1998) looked at 33 $M \geq 5$ strike-slip earthquakes along the San Andreas fault physiographic province and found that 17 were preceded by ($M \geq 2$) foreshocks, corresponding to a rate density of 17% per magnitude unit. A similar result was also reported by Lindh and Lim (1995) for the same region, time period and magnitude ranges. Bowman and Kisslinger (1984) determined that 15% of $m_b \geq 5.0$ earthquakes in the Adak thrust zone were preceded by foreshocks. While they estimated a completeness threshold of $M_f^{\min} = 2.5$ (m_b), inspection of their Figure 2 suggests an alternate threshold of, perhaps, $m_b = 3.8$, which would correspond to a foreshock rate density of about 12% per magnitude unit. Variations among these studies involving the region studied and the choice of spatial and temporal windows make it difficult to compare these results. Given these differences, the range of results is surprisingly narrow, from 12% to 17% per magnitude unit. This range coincides closely with the 95% confidence range estimated here for the foreshock rate among $M \geq 6$ mainshocks in the Harvard catalog) and may be considered a robust, worldwide meta-result.

The present results for $M_m^{\min} = 7$ do not follow this trend, however, and instead indicate a foreshock rate density of between 5.5% and 11.7% per magnitude unit (95% confidence range) (Table 2). However, because of the small number of $M \geq 7$ mainshocks ($N = 149$), these rates are not significantly different from the corresponding rates determined with the more numerous $M \geq 6$ events.

Comparison of the present results of those of VON SEGGERN's (1981) is made with the recognition that Von Seggern's analysis did not adequately correct for background seismicity. While Von Seggern's assay was known to include background events, he did not attempt to quantify their presence. Instead, he interpreted the relative rates, but cautioned against assigning significance to the absolute rates. He concluded that "the rate . . . for shallow foreshocks is significant but still small," and that "the true rate of foreshock occurrence was less than 20%." Von Seggern's Table 4 reports a foreshock rate of 18% (9 out of 51) among $M_s \geq 7$ mainshocks of all kinds (mostly 0–100 km depth). This result may be compared most directly to the apparent rate of foreshocks before $M \geq 7$ mainshocks (19 out of 95, or 20%) observed in the present study before the estimated rate of background activity is removed (Table 2). The spatial and temporal distribution of foreshocks relative to their mainshocks observed by Von Seggern is similar to those seen in this study. Most of Von Seggern's foreshocks occurred within 10 days of their mainshocks, and most fell within 0.5 degree of the mainshocks.

Sensitivity to Parameter Choices

The above estimates of foreshocks rate depend on several choices made in the analysis, perhaps the most important being the minimum magnitude of foreshocks and the duration and spatial size of the foreshock window. I argued above that the results should be fairly insensitive to increases in the space and time window parameters as long as the window exceeds the natural scale of clustering. In addition, I suggested that the use of events with magnitudes below the completeness level of the catalog would produce foreshock assays approximately 28% low. I explore here the actual sensitivity of key results to these choices.

For both the $M \geq 6$ and $M \geq 7$ mainshock sets, I increased the time window to 20 days and increased the distance window to 100 km, separately. The apparent

Table 2

Summary of retrospective foreshock rates

Mainshock category	Number of mainshocks	Number of possible foreshocks (1)	Expected number of background events	Fraction of mainshocks preceded by a foreshock (2) (%)	95% confidence range (3) for foreshock frequency (%)	95% confidence range (3) for foreshock rate per unit magnitude difference (%)
Harvard CMT catalog ($M \geq 6$) 1978–1996						
All	1108	161	14.4	13.2	11.3–15.4	11.3–15.4
Thrust	533	103	9.7	17.5*	14.4–21.0	14.4–21.0
Normal	172	23	2.1	12.2	7.7–18.0	7.7–18.0
Strike-slip	397	34	2.4	8.0**	5.5–11.1	5.5–11.1
Harvard CMT catalog ($M \geq 7$) (1978–1996)						
All	149	26	1.5	16.5	10.9–23.4	5.5–11.7
Thrust	95	19	1.2	18.7	11.5–28.0	5.8–14.0
Normal	13	3	0.1	22.3	—	—
Strike-slip	41	4	0.3	9.0	—	—

* Foreshock rate among thrusts exceeds foreshock rate among all $M > 6$ earthquakes in the Harvard CMT catalog at the 95% confidence level.
** Foreshock rate among strike-slip events is lower than foreshock rate among all $M > 6$ earthquakes in the Harvard CMT catalog at the 99% confidence level.
(1) A spatial window of 75 km and a temporal window of 10 days were used.
(2) Corrected for estimated background rate.
(3) Range represents sampling uncertainty.

foreshock rates decreased in all cases (suggesting that I may be overcorrecting for the background). These changes are minor and fall within the given 95%-confidence ranges (reflecting sampling uncertainty) listed in Table 2.

Increasing the minimum magnitude from 5.0 to 5.5 resulted in higher estimates for (normalized) foreshock rates before $M \geq 6$ and $M \geq 7$ mainshocks. These increases correspond to apparent deficits in the ($M \geq 5.0$) catalog of 13% and 33%, respectively, which are consistent with the predicted deficit of 28%.

Dependence on Focal Mechanism

Foreshock rate varied significantly among subsets of mainshocks defined by their focal mechanisms. Among the ($M \geq 6$) thrust earthquakes in the Harvard CMT catalog, 17.5% were preceded by a ($M \geq 5$) foreshock, this rate being significantly higher (95% confidence) than the corresponding rate of 13.2% observed before all ($M \geq 6$) earthquakes. Also, strike-slip mainshocks were preceded by foreshocks only 8.0% of the time, a rate significantly lower than that for all ($M \geq 6$) events at the 99% confidence level. Thus, foreshocks occurred before shallow $M \geq 6$ thrust events at approximately twice the rate they did before shallow $M \geq 6$ strike-slip events.

The relatively high foreshock rates observed before thrust earthquakes appear to contradict a result of ABERCROMBIE and MORI (1996) (their Fig. 2b), who reported a lower frequency of foreshocks before reverse mainshocks than before strike-slip mainshocks ($M \geq 5$) in California and western Nevada. There are at least two possible ways to resolve these apparent differences. First, ABERCROMBIE and MORI (1996) considered small numbers of earthquakes: 14 reverse mainshocks (2 with foreshocks) were compared to 38 strike-slip mainshocks (19 with foreshocks). Statistical error in these samples may partially mask the underlying frequencies. In addition, ABERCROMBIE and MORI (1996) were able to use the well-determined depths in the California and Nevada catalogs to establish a clear inverse dependence of foreshock rate on the depth of the mainshock, over a depth range from 0 to about 20 km. Because reverse events tend to be deeper than strike-slip events in California, at least some of the apparent dependence on focal mechanism they report might actually be a coupled depth (via normal stress) effect, as suggested by ABERCROMBIE and MORI (1996).

The present result appears to contradict JONES (1984), who found no foreshocks preceding four ($M_L \geq 5$) thrust mainshocks in southern California, and who also suggested that the foreshock rate for thrust earthquakes in California might be lower than the rate for strike-slip events (15% per magnitude unit, based on 16 mainshocks), although they cautioned that their data set was too small to definitively distinguish between these rates.

Also, my result appears at first to conflict with JONES and MOLNAR (1979), who concluded that there was a higher foreshock rate before large earthquakes in non-subduction zones than in subduction zones. They reported a foreshock rate of 20% (10% per magnitude unit) among 120 $M \geq 7$ earthquakes in subduction zones, and a rate of 29.3% (15% per magnitude unit) among 41 $M \geq 7$ events in non-subduction zones. (Here, I assumed a magnitude aperture of 2 in Jones and Molnar's study.) However, there are too few data in this study to infer a significant rate difference with confidence; the 90% confidence ranges are, respectively, 7%–14% per magnitude unit (subduction zone events) and 9%–22% per magnitude unit (non-subduction zone events).

In spite of the difficulties in comparing these studies, the apparent discrepancies between them and the present worldwide result suggests that California foreshocks may be a special case, owing to the particular regional tectonics there, and are not representative of worldwide foreshock rates. This possibility is explored below.

Subduction Zones and Continental Thrusts

In the Harvard ($M \geq 6$) assay, 17.5% of the thrust-type mainshocks were preceded by foreshocks, and most of these are located in the shallow portions of the major circum-Pacific subduction zones (Fig. 7a). These events involve the subduction of water-saturated oceanic crust in the uppermost portion of descending slabs and have a median depth of 30 km. In contrast, only 3 foreshocks were found among the 35 thrust earthquakes in the Himalayan collision belt involving the Indio-Australia and Eurasian plates, and none were found before the 7 thrust (and oblique thrust) earthquakes in California. These events are mountain-building earthquakes involving the transpression of continental, crystalline rock and have a median depth of 15 km. The contrast suggests that the foreshock rate may be higher in shallow subduction zones than in continental thrust belts, although the number of events in these regional subsets are generally too small to prove it. This could explain both the high foreshock rate for thrust earthquakes observed in this study and the low rates observed among California thrust earthquakes (JONES, 1984; ABERCROMBIE and MORI, 1996).

Foreshocks preceding normal mechanism earthquakes occurred along the east African rift, on the spreading centers in the Indian Ocean, and along the western Pacific subduction zones from New Zealand to Japan (Fig. 7b).

Most of the foreshocks that preceded ($M \geq 6$) strike-slip mainshocks lie along the convergent boundary between the Pacific and Indio-Australian plates, from the New Guinea Thrust to the New Hebrides Thrust (Fig. 7c). Three were located in California. In other regions, few strike-slip earthquakes were preceded by foreshocks.

Prospective Foreshock Frequencies

From the perspective of real-time, short-term earthquake hazard assessment, it is the probability of earthquakes following earthquakes, not preceding them, that is

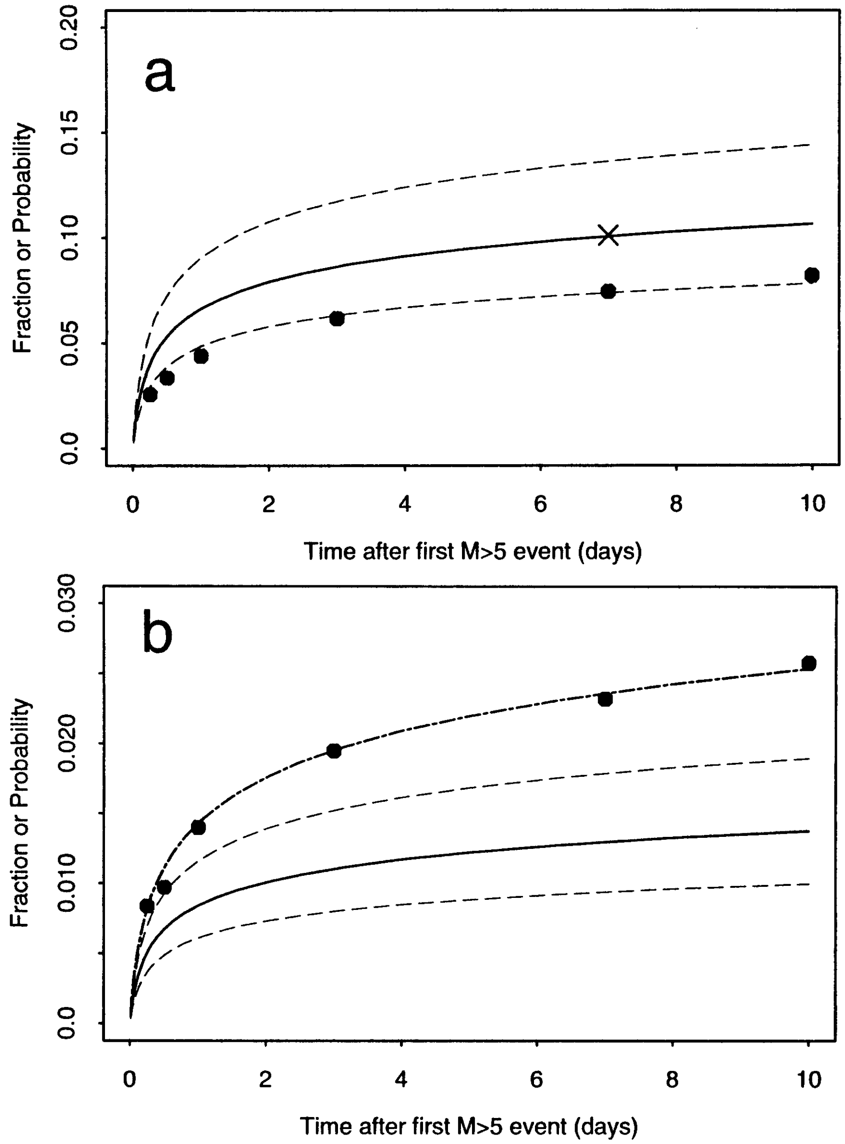


Fig. 6.

of interest. Here I use the retrospective foreshock frequencies, such as those given in Table 1, to calculate "prospective foreshock frequencies." I calculate the fraction of shallow $M \geq 5$ earthquakes in the Harvard catalog that were followed within time periods up to 10 days and within distances up to 75 km by a larger earthquake, and compare these frequencies to the corresponding prospective probabilities calculated with a generic model of aftershock activity in California

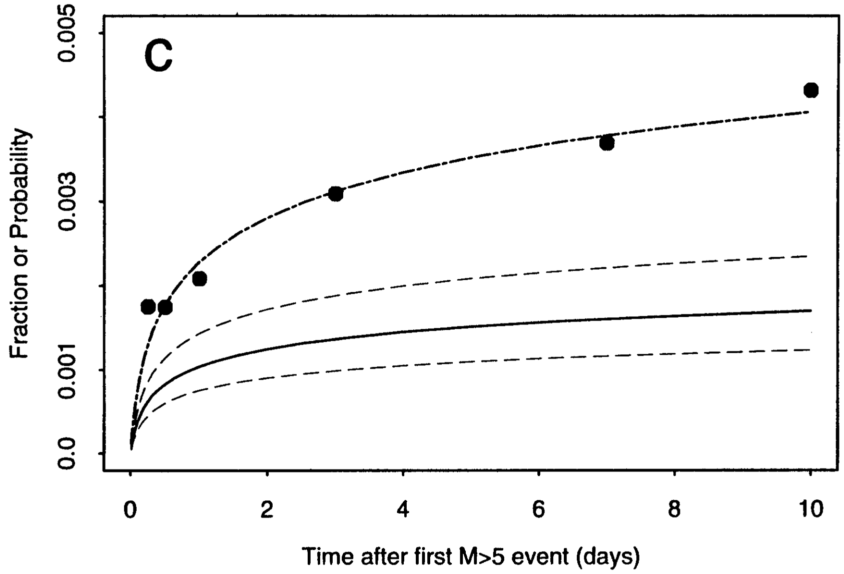


Figure 6
Observed cumulative frequencies (solid circles) and predicted probabilities based on the California generic clustering model (solid lines) of larger earthquakes which occur after an $M \geq 5$ earthquake, as a function of the time after the $M \geq 5$ earthquake. Broken lines indicate the 95% confidence range of California generic model. Observed frequencies are calculated at times 0.25, 0.5, 1, 3, 7 and 10 days after the first ($M \geq 5$) event. (a) Case of any larger earthquake following a $M \geq 5$ event. (b) Case of an $M \geq 6$ earthquake following an $M \geq 5$ event. (c) Case of an $M \geq 7$ earthquake following an $M \geq 5$ event. " × " in (a) marks a generic forecast often announced by the USGS after significant California earthquakes of "a 10% chance of a larger event in the next week." In (b) and (c), dash-dot lines indicate the world generic model (defined by the model parameters $a = -1.5$, $b = 0.8$, $p = 1.0$ and $c = 0.05$), which was foward-fitted to the data points shown.

(REASENBERG and JONES, 1989, 1994). Earthquakes that otherwise qualify as $M \geq 5$ mainshocks, but which are themselves aftershocks, are not included in the survey. I look separately at the cases in which $M \geq 5$ earthquakes are followed by any larger ($M \geq 5$) event, by a $M \geq 6$ event, and by a $M \geq 7$ event.

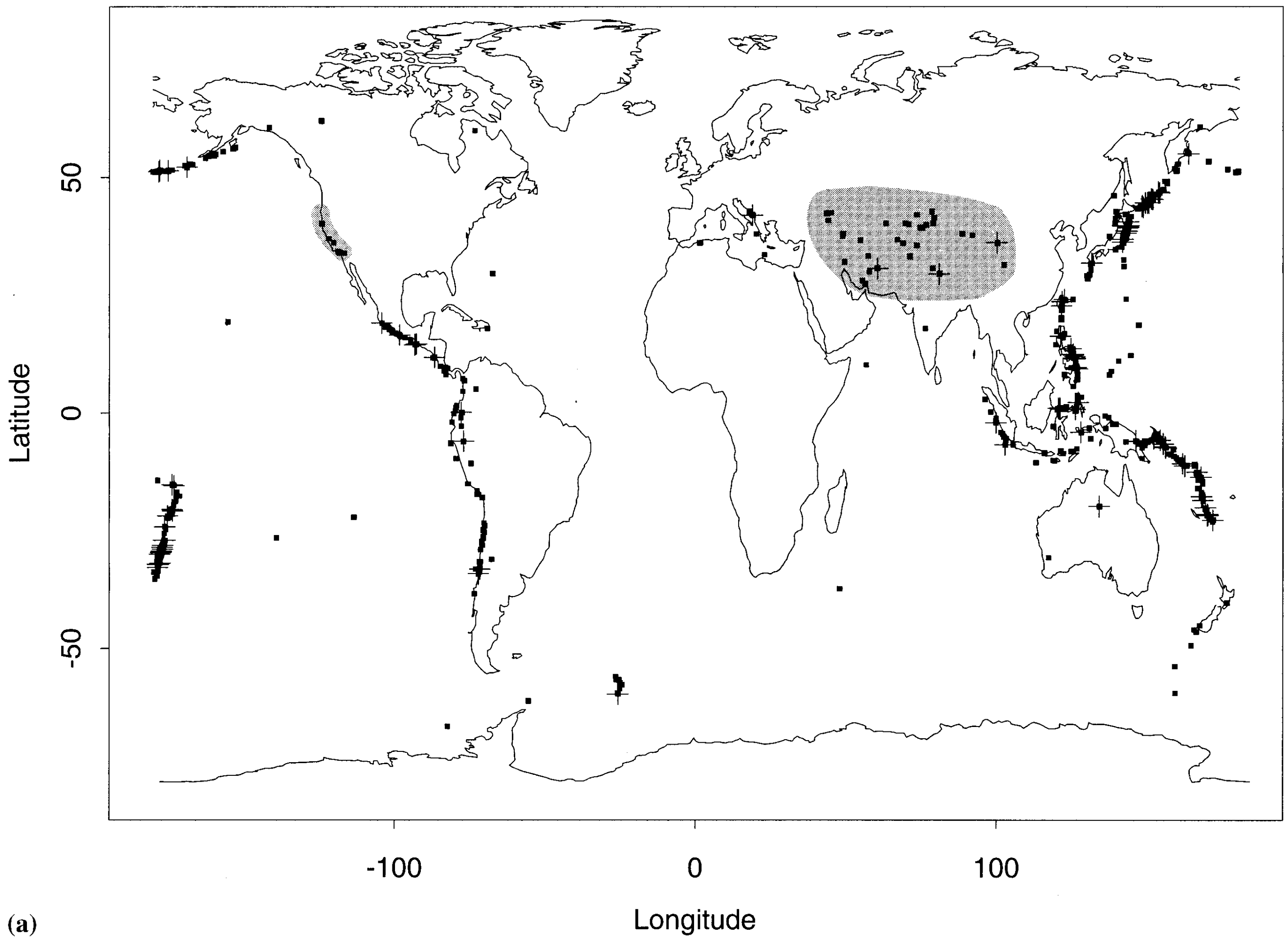
The observed frequency of larger ($M \geq 5$) earthquakes following a $M \geq 5$ earthquake is 7.5% in a week, approximately 25% lower than the California generic model (Fig. 6a). The observed frequency of $M \geq 6$ earthquakes following a $M \geq 5$ earthquake is 2.3% in a week, approximately twice the generic model and outside is 95% confidence range (Fig. 6b). The observed frequency of $M \geq 7$ earthquakes following a $M \geq 5$ earthquake is approximately 0.4%, again approximately twice the generic model and outside its 95% confidence range (Fig. 6c). An explanation for this highly non-self-similar result involves the catalog incompleteness, which has caused the observed frequencies of $M > 5$ mainshocks to be artificially low. Each observation point in Figure 6 represents a ratio between an observed number of foreshock-mainshock pairs and an observed number of potential ($M \geq 5$) foreshocks. In the cases of pairs involving $M \geq 6$ or $M \geq 7$ earthquakes, the catalog deficiency equally affects both quantities in the ratio, since both rely directly on the ability to observe $M \geq 5$ earthquakes; the combined effect of the missing events cancels out in the ratio. But in the case of pairs involving two $M \geq 5$ events, the error associated with incompleteness enters twice (and is, hence, squared), while the error in the number of potential foreshocks is still only directly affected by the incompleteness. Consequently, the ratio (and points in Fig. 6a) will be artificially lowered. From this line of reasoning, I conclude that the observed frequencies of $M \geq 6$ and $M \geq 7$ earthquakes following $M \geq 5$ events (points in Figs. 5b and 5c) are unbiased estimates of the true frequencies, while the $M \geq 5$ observations (points in Fig. 6a) are artificially low. In Figure 6a, the " $\times$ " indicates the probability usually stated in USGS forecasts in California after a $M \geq 5$ mainshock ("10% probability of an equal or larger event in the next 7 days"). While the corresponding frequency observed in the Harvard catalog is 7.5%, extrapolation from the $M > 6$ and $M > 7$ results suggests that the actual worldwide rate may be as high as 15%.

The observed frequencies of $M > 6$ and $M > 7$ earthquakes in the Harvard catalog were used to forward fit a new model (which I call the "world generic model") having the same form as in REASENBERG and JONES (1989) but defined by new parameter values ($a = -1.5$, $b = 0.8$, $p = 1.0$ and $c = 0.05$). Numerically, the biggest difference between these models is in the a-value. But there are other important differences as well. The world generic model may be better determined than the California generic model because it is based on 171 foreshock-mainshock pairs, compared to only 62 aftershock sequences in the REASENBERG and JONES (1989) compilation. In addition, the world generic model directly reflects the behavior of large earthquakes involved in foreshock-mainshock sequences, while the California generic model is an extrapolation from a smaller aftershock activity.

Summary and Discussion

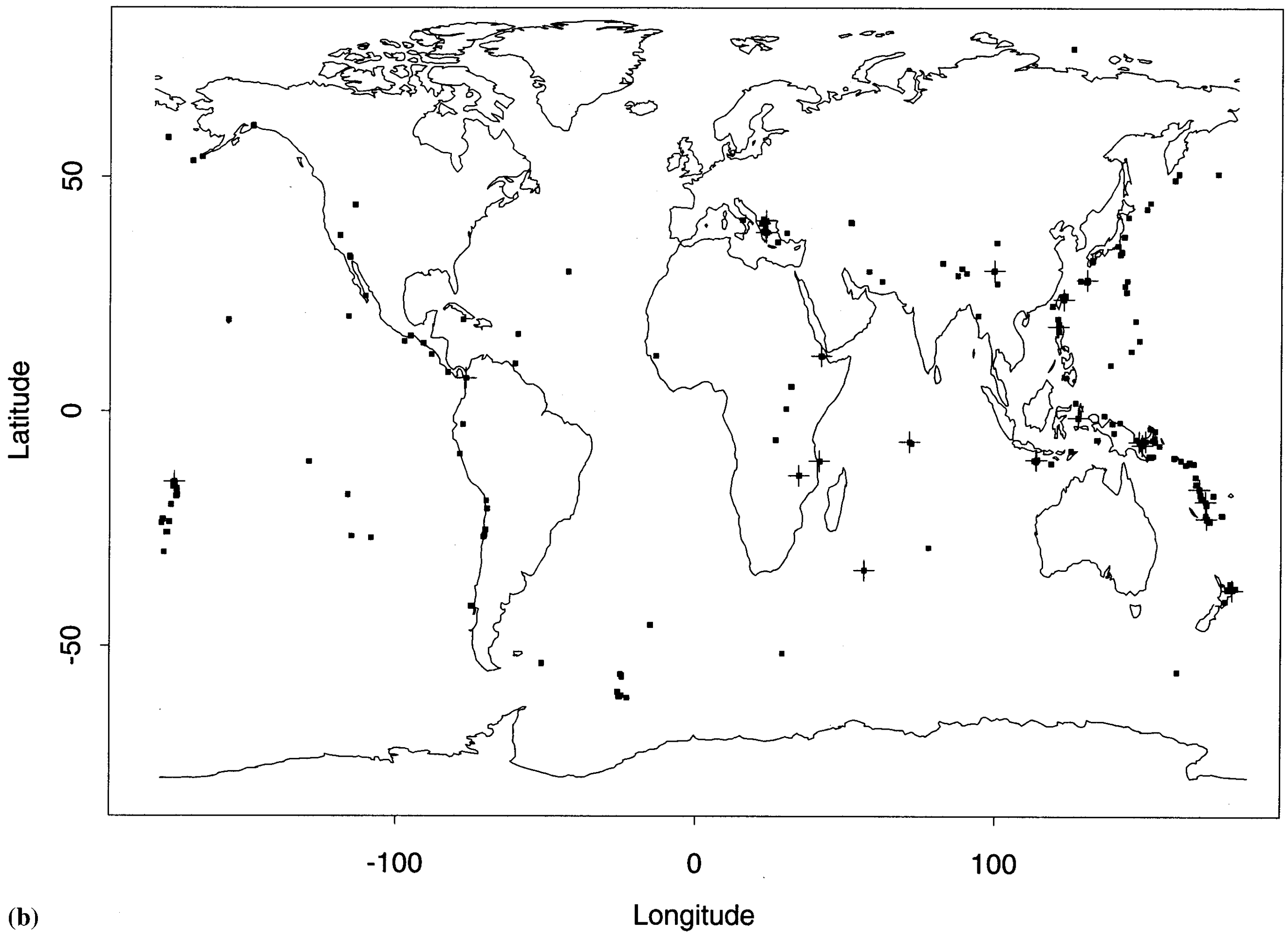
The overall foreshock rate of 13.2% per magnitude unit found here for all $M > 6$ earthquakes in the Harvard CMT catalog (1978–1996) lies within the range of estimates of foreshock rate density (12%–17% per magnitude unit) obtained by other studies using regional and worldwide catalogs. The higher rate estimates within this range (for example, those reported in MICHAEL and JONES, 1998 and LINDH and LIM, 1995) were obtained with relatively small numbers of $M \geq 5$ mainshocks (33 and 30, respectively) and were limited to strike-slip earthquakes in the San Adreas Fault zone. Larger and broader studies (JONES and MOLNAR, 1979; JONES, 1984; AGNEW and JONES, 1991; and the present one) find slightly lower overall rates in the range 12%–15%.

This study found with high statistical confidence that the foreshock rate before $M \geq 6$ thrust mainshocks in the Harvard catalog is about twice the corresponding



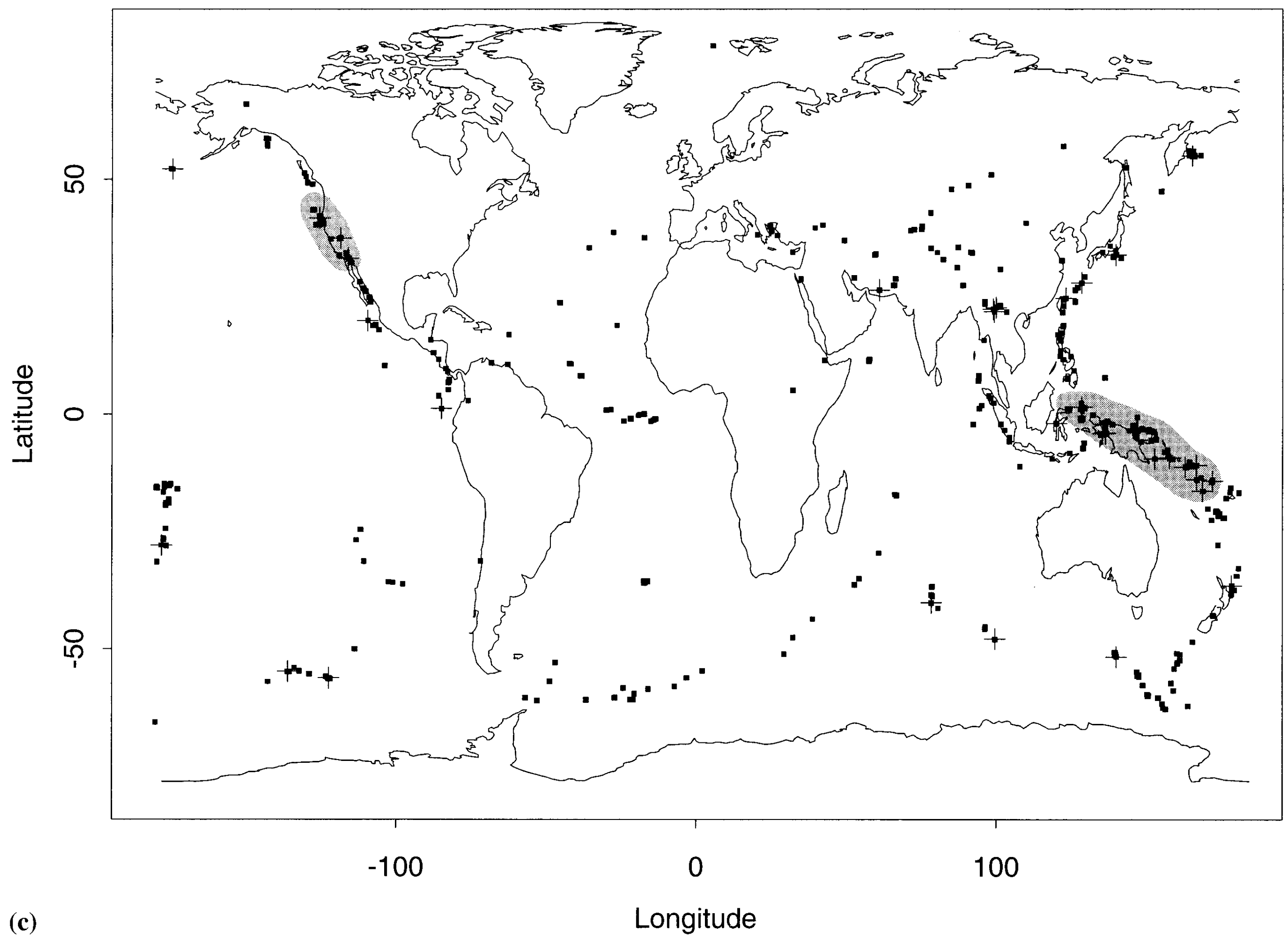
(a)

Fig. 7.



(b)

Fig. 7.



(c)

Fig. 7.

rate before strike-slip mainshocks. The apparent contradiction between the resolvable trend in the worldwide data of higher rates among thrusts than strike-slips and the opposite trend inferred in studies of California $M \geq 5$ earthquakes (JONES, 1984; ABERCROMBIE and MORI, 1996) may be resolved by hypothesising that most California thrusts (with the exception of off-shore events associated with the subduction of the Juan de Fuca plate) are typical of continental thrust belts which involve unsaturated, crystalline continental rock (such as the Himalayan collision belt) but not typical of subduction zone thrust events which occur in young, water-saturated ocean sediments. This would imply that the generic California model may be representative of most earthquakes in California, but may significantly underestimate the conditional probabilities following potential foreshocks in Cascadia, a region whose clusters may be more typical of shallow subduction zones.

The worldwide occurrence of $M \geq 6$ and $M \geq 7$ earthquakes in the 10-day periods following (and 75-km distance range from) $M > 5$ earthquakes in the Harvard catalog exceeds by a factor of about 2 the rate predicted by the California generic model (Figs. 5b and 5c). This difference may be understood, in part, by the dominance of shallow subduction thrusts in the Harvard data set (which were found to have a relatively high foreshock rate), the presence of an active strike-slip plate boundary in California (associated with a low foreshock rate worldwide), and the absence of an active subduction zone in most of California (an important exception being the Cascadia megathrust). In this sense, the California generic model can be considered "correct" for most of California, while the world generic model would better represent worldwide foreshock-related conditional probabilities. In certain other regions where sufficient seismological data are available, regionally determined generic clustering models (see, for example, REASENBERG *et al.*, 1990; JONES *et al.*, 1995) may provide a better basis than the worldwide generic model for estimating short-term earthquake probabilities.

Figure 7
Locations of Harvard catalog $M > 6$ mainshocks (1978–1996) and their ($M > 5$) foreshocks, separated according to the focal mechanism of the mainshock. Solid squares indicate mainshocks; plus signs indicate foreshocks. (a) Thrust events. Shaded areas, which mark continental thrust zones, have a lower foreshock rate (3 out of 35 mainshocks) than is found along the shallow portions of the major circum-Pacific subduction zones. (b) Normal events. (c) Strike-slip events. Shaded areas along the New Hebrides and New Guinea thrusts (presumably including strike-slip events on transform faults associated with the subduction zone) and the San Andreas Fault zone have average or higher foreshock rates, compared to that for strike-slip events in the rest of the world.

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