

EARTHQUAKE REPEAT TIME AND AVERAGE STRESS DROP
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Abstract. Existing data on source parameters of large crustal earthquakes (subduction events are not considered here) over a wide range of repeat times indicate that, for a given magnitude (M₅ or Mₑ), earthquakes with long repeat times have shorter fault lengths than those with short repeat times. A shorter fault length for a given magnitude indicates a larger average stress drop which reflects the average strength of the fault zone. Our result therefore suggests that faults with longer repeat times are stronger than those with shorter repeat times. In terms of an asperity model in which the average strength of a fault zone is determined by the ratio, rₛ, of the total area of the asperities (strong spots on a fault plane) to the total area of the fault zone, the above result suggests that rₛ is proportional to the repeat time. Our result provides a method to estimate seismic source spectra from the fault length and the repeat time of a potential causative fault.

1. Introduction

The repeat time of earthquakes on a given fault segment is controlled by the rate of tectonic loading (long-term slip rate) and the stress accumulation and release mechanism on the fault. Most large earthquakes at active plate margins have relatively short (30 to 200 years) repeat times, while some intraplate events have repeat times as long as several thousand years, even if they are relatively close to a plate boundary.

In this paper, we examine published source parameters of large crustal earthquakes for which repeat times have been estimated, in an attempt to see whether events with grossly different repeat times have different source characteristics. In particular, we examine average stress drops associated with faulting. Although stress drops may be controlled by many parameters other than the repeat time, the large range of the repeat times (i.e., 20 to several thousand years) among different earthquakes would help isolate the factor that determines the earthquake stress drop.

Earthquakes on subduction thrust boundaries are not considered here, because they involve fault geometries and depths very different from the crustal earthquakes considered here.

2. Data

The data on the source parameters and the repeat times are summarized in Table 1. Among the various source parameters, we use the surface-wave magnitude Mₛ, the seismic moment M₀ (or the corresponding moment magnitude Mₑ = (log M₀ - 16.1)/1.5), the fault length L, and the fault width W.

The surface-wave magnitude, Mₛ, is the most widely used parameter and is available for very old events as well as recent events. Although the seismic moment, M₀, is not available for some of the old events, it directly represents the overall size of the source (M₀ = µDS, where µ = rigidity, D = fault offset, S = fault area), and allows more quantitative interpretations of the data than does the magnitude.

The fault length, L, can be determined from various data such as the extent of the surface break, geodetic data, aftershock area, macro-seismic data, and the spectrum of radiated seismic waves. However, surface breaks do not always represent the entire extent of the fault, particularly for small events. The size of the aftershock area is often used to estimate the fault length. Since the aftershock area is not always defined rigorously, and because it varies as a function of time, some ambiguity exists in this method, too. However, several studies have demonstrated that the size of the aftershock area at a relatively early stage of the activity is indeed a good approximation of the fault length [Benioff, 1962; Ben-Menahem and Toksoz, 1963; Mogi, 1968; Wyss, 1979; Kanamori and Given, 1981]. The fault length estimated from the size of the aftershock area can be checked against macro-seismic data, such as the intensity distribution, the tsunami source area, and the surface rupture. Geodetic data can also be used to crosscheck the result.

Since the aftershock data are available for most events of Table 1, we primarily use this method in this paper. In several cases, some subjective judgment is necessary, as discussed in the Appendix. Despite these inevitable ambiguities, we believe that the fault length estimated from the initial aftershock area is a reasonably good approximation of the length of the seismic rupture zone, at least for the events examined here.

Many recent studies suggest that the slip is not uniform on the fault plane defined by the aftershock area, but is concentrated in a much smaller area (e.g., 1968 Borrego Mountain earthquake, Burdick and Mellman, 1976; 1979 Imperial Valley earthquake, Hartzell and Helmberger, 1982; 1979 Coyote Lake earthquake, Liu and Helmberger, 1983). Although we consider this nonuniformity to be an important feature of seismic faulting, we first use the fault length defined by the extent of the aftershock area to establish the scaling relations.

The width of the fault, W, is even more difficult to determine than the fault length, L. The vertical extent of the aftershock area can be used, but the lack of aftershocks at a large depth does not necessarily mean that no seismic slip occurs there. Because of the increased temperature at large depths, the fault zone there may not be capable of generating aftershocks, even if it can slip coseismically.

In principle, the vertical extent of the fault can be estimated from geodetic data, but the available data are seldom complete enough to resolve it. Furthermore, geodetic observations usually include afterslip as well as the coseismic slip. In this paper, the
width estimated from the aftershocks is used for most events.

We present two diagrams: (1) $M_S$ versus $L$, (2) $M_W$ versus $L$. The $M_S$ versus $L$ diagram (Figure 1) involves the parameters which are directly determined from the data without much interpretation. The $M_W$ versus $L$ diagram (Figure 2) is similar to the $M_S$ versus $L$ diagram, but it involves the seismic moment, $M_0$. Since the method and the type of the data used for the determination of $M_0$ vary among the investigators and the events, some ambiguity exists concerning the value of $M_0$. However, $M_0$ represents the physical size of the source more directly than $M_S$ and is easier to interpret.

In this paper, the values of $M_0$ determined from the amplitude of seismic waves and the geodetic data are used, and the values obtained by different methods for each event are crosschecked for consistency (for details, see the Appendix).

The repeat time, $t$, of earthquakes has been estimated by various methods. For many Japanese events, the slip rate, $V$, along a fault estimated from geomorphological data supplemented by $C^4$ dates, and the amount of slip, $D$, in a large earthquake judged to be characteristic of the fault are used to estimate $t = D/V$ [Matsuda, 1975a].

When historical data are available for a very long period of time, repeat times can be estimated from such data (e.g., earthquakes along the North Anatolian fault: Allen [1975]; Ambresays [1970]. More recently, the offset patterns in fault zones exposed by trenching are used to determine the time history of the activity of the fault [e.g., Clark et al., 1972; Sieh, 1978].

In this study, we use the values of $t$ determined by various methods, and the references are given in the Appendix. A number of the published recurrence intervals, such as those for the 1952 Kern County and 1971 San Fernando earthquakes, are admittedly based on very scanty and debatable evidence; nevertheless, we have felt obligated to use such numbers when no other data are available.

3. Results

3.1 $M_S$ Versus $L$ and $M_W$ Versus $L$ Diagram. Figure 1 shows the relation between the surface-wave magnitude, $M_S$, and the fault length, $L$, for earthquakes having different repeat times.

In general, for a given $M_S$, earthquakes having a longer repeat time have a shorter fault length (see the events with $7 < M_S < 8$). On the other hand, for a given $L$, earthquakes with a longer repeat time tend to have a larger $M_S$ (see the events with $30 < L < 50$ km), although the total number of events is relatively small. This situation is best illustrated by comparing three representative earthquakes: 1927 Tango, 1966 Parkfield, and the 1976 Guatemala earthquakes. Both the Guatemala and Tango earthquakes have about the same $M_S$, yet $L = 250$ and $35$ km for the Guatemala and Tango earthquakes, respectively. The repeat time is about 180 to 755 years for the Guatemala event [Schwartz et al., 1979] and several thousand years for the Tango earthquake [Matsuda, 1975a]. Both the Parkfield and Tango earthquakes have about the same $L$, but they have a very different $M_S$: $M_S = 7.6$ for Tango and $M_S = 6.0$ for Parkfield. The Parkfield earthquake has a very short repeat time (about 22 years; Bakun and McEvilly [1984] compared with that for the Tango earthquake (at least 2000 years).

Since $M_S$ is a purely empirical parameter, and since only one spatial dimension, $L$, is given, we cannot directly interpret the $M_S$ versus log $L$ diagram in terms of the stress drop. Here we define the average stress drop by

### TABLE 1. Earthquake Source Parameters (for details, see the Appendix)

<table>
<thead>
<tr>
<th>Event</th>
<th>$M_S$</th>
<th>$M_W$</th>
<th>$M_D$</th>
<th>$L$</th>
<th>$W$</th>
<th>$S$</th>
<th>$t$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(10^{27} dyne-cm)</td>
<td>(km)</td>
<td>(km)</td>
<td>(km²)</td>
<td>(years)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Alaka, 1958</td>
<td>7.9</td>
<td>7.8</td>
<td>0.1</td>
<td>300</td>
<td>16</td>
<td>4800</td>
<td>60-110</td>
</tr>
<tr>
<td>Borah Peak, 1983</td>
<td>7.3</td>
<td>7.0</td>
<td>0.1</td>
<td>40</td>
<td>13</td>
<td>520</td>
<td>100</td>
</tr>
<tr>
<td>Borrego Mt., 1968</td>
<td>6.7</td>
<td>6.6</td>
<td>0.0035</td>
<td>25</td>
<td>8</td>
<td>200</td>
<td>75</td>
</tr>
<tr>
<td>Coyote Lake, 1979</td>
<td>5.7</td>
<td>5.6</td>
<td>0.13</td>
<td>46</td>
<td>10</td>
<td>460</td>
<td>100</td>
</tr>
<tr>
<td>Daofu, 1981</td>
<td>6.8</td>
<td>6.7</td>
<td>2.6</td>
<td>250</td>
<td>15</td>
<td>3750</td>
<td>180-755</td>
</tr>
<tr>
<td>Guatemala, 1976</td>
<td>7.5</td>
<td>7.5</td>
<td>3.0</td>
<td>1300</td>
<td>390</td>
<td>22</td>
<td></td>
</tr>
<tr>
<td>Haiyuan, 1920</td>
<td>8.6</td>
<td></td>
<td></td>
<td>220</td>
<td>700-1000</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hehegen Lake, 1959</td>
<td>7.5</td>
<td>7.3</td>
<td>3.0</td>
<td>1300</td>
<td>390</td>
<td>22</td>
<td></td>
</tr>
<tr>
<td>Imperial Valley, 1979</td>
<td>6.5</td>
<td>6.5</td>
<td>0.06</td>
<td>42</td>
<td>10</td>
<td>420</td>
<td>40</td>
</tr>
<tr>
<td>Izu, 1930</td>
<td>7.2</td>
<td>6.9</td>
<td>0.25</td>
<td>22</td>
<td>12</td>
<td>264</td>
<td>700-1000</td>
</tr>
<tr>
<td>Izu-Oki, 1974</td>
<td>6.5</td>
<td>6.4</td>
<td>0.059</td>
<td>20</td>
<td>11</td>
<td>220</td>
<td>100</td>
</tr>
<tr>
<td>Kern County, 1952</td>
<td>7.7</td>
<td>7.3</td>
<td>1.0</td>
<td>70</td>
<td>20</td>
<td>1400</td>
<td>170-450</td>
</tr>
<tr>
<td>Luhuo, 1973</td>
<td>7.4</td>
<td>7.4</td>
<td>1.8</td>
<td>110</td>
<td>15</td>
<td>1650</td>
<td>100</td>
</tr>
<tr>
<td>Mikawa, 1945</td>
<td>6.8</td>
<td>6.6</td>
<td>0.087</td>
<td>12</td>
<td>11</td>
<td>132</td>
<td>2000-4x10^4</td>
</tr>
<tr>
<td>Morgan Hill, 1984</td>
<td>6.1</td>
<td>6.1</td>
<td>0.02</td>
<td>30</td>
<td>10</td>
<td>300</td>
<td>75</td>
</tr>
<tr>
<td>N. Anatolian, 1939</td>
<td>7.8</td>
<td></td>
<td></td>
<td>350</td>
<td>15</td>
<td>5250</td>
<td>150-200</td>
</tr>
<tr>
<td>N. Anatolian, 1943</td>
<td>7.6</td>
<td></td>
<td></td>
<td>265</td>
<td>15</td>
<td>3975</td>
<td>150-200</td>
</tr>
<tr>
<td>N. Anatolian, 1944</td>
<td>7.4</td>
<td></td>
<td></td>
<td>190</td>
<td>15</td>
<td>2850</td>
<td>150-200</td>
</tr>
<tr>
<td>Niigata, 1964</td>
<td>7.5</td>
<td>7.6</td>
<td>3.0</td>
<td>60</td>
<td>25</td>
<td>1500</td>
<td>560</td>
</tr>
<tr>
<td>Parkfield, 1966</td>
<td>6.0</td>
<td>6.0</td>
<td>0.014</td>
<td>30</td>
<td>13</td>
<td>390</td>
<td>22</td>
</tr>
<tr>
<td>Pleasant Valley, 1915</td>
<td>7.7</td>
<td></td>
<td></td>
<td>62</td>
<td></td>
<td>5000</td>
<td></td>
</tr>
<tr>
<td>San Fernando, 1971</td>
<td>6.6</td>
<td>6.7</td>
<td>0.12</td>
<td>17</td>
<td>17</td>
<td>289</td>
<td>100-300</td>
</tr>
<tr>
<td>Tabas, 1978</td>
<td>7.4</td>
<td>7.4</td>
<td>1.5</td>
<td>65</td>
<td>20</td>
<td>1300</td>
<td>&gt;1300</td>
</tr>
<tr>
<td>Tango, 1927</td>
<td>7.6</td>
<td>7.0</td>
<td>0.46</td>
<td>35</td>
<td>13</td>
<td>455</td>
<td>2000-6x10^4</td>
</tr>
<tr>
<td>Tangshan, 1976</td>
<td>7.8</td>
<td>7.4</td>
<td>1.8</td>
<td>80</td>
<td>15</td>
<td>1200</td>
<td>&gt;2000</td>
</tr>
<tr>
<td>Tottori, 1943</td>
<td>7.4</td>
<td>7.0</td>
<td>0.36</td>
<td>33</td>
<td>13</td>
<td>429</td>
<td>6000</td>
</tr>
</tbody>
</table>
However, it is unlikely that both 

Since the details of the scaling relations are unknown, these lines should not be given too much significance; they should be con-

Experimental studies on rocks and other materials demonstrate that the static friction between two surfaces generally increases as the time of stationary contact increases (e.g., Scholz and Engelder, 1976; Shimamoto and Logan, 1984; Richardson and Nolle, 1976). Our results are consistent with these laboratory results, although the time scales involved are very different between laboratory and in-situ conditions.

Many recent studies indicate that the displacement and the stress change on a fault plane are very nonuniform in space. One of the best documented cases is the 1979 Imperial Valley, California, earthquake for which a large number of near-field strong-motion records are available for detailed modeling. Hartzell and Helmberger [1982], Olson and Aspel [1982], Hartzell and Heaton [1983] and Archuleta [1984] made extensive analyses of this data set to determine the distribution of slip on the fault. Although the results obtained in these studies differ in detail, an important conclusion is that a major proportion of the slip is concentrated in an area which is much smaller than the total aftershock area. Hartzell

Fig. 1. The relation between the surface-wave magnitude, $M_s$, and the fault length, $L$. The solid lines indicate the trend for a constant stress drop.

$$\overline{\Delta \sigma} = \int_0^L \Delta \sigma D S / \int_0^L D S$$

where $\Delta \sigma$ and $D$ are the stress drop and the dislocation on the fault plane, $S$, respectively. Following Kanamori [1977], the numerator can be written as $2E_5$ where $E_5$ is the energy radiated in seismic waves. Although direct determinations of $E_5$ are seldom available, Gutenberg and Richter's [1956] magnitude-energy relation, $\log E_5 = 1.5M_s + 11.8$, is generally considered a good approximation for earthquakes with $M_s < 8$. Using this relation, the above relation can be written as

$$\overline{\Delta \sigma} = 2\mu E_5 / M_0 = 2 \times 10^{-1.5M_s + 11.8/L} W D$$

where $W$ is the width of the fault and $D$ is the average dislocation. If both $W$ and $D$ are proportional to $L$, $\log L \propto (1.5/3)M_s - (1/3) \log \Delta \sigma$. However, for large crustal earthquakes, $W$ is more or less bounded by the thickness of the seismogenic zone, and would not increase as fast as $L$. If $W$ and $D$ are fixed, then $\log L \propto (1.5M_s - \log \Delta \sigma$. However, it is unlikely that both $W$ and $D$ stay completely constant as $L$ increases. In fact, Scholz [1982] found that $L \propto D$ for most crustal earthquakes. In this case, if the variation of $W$ is small, $\log L \propto (1.5/2)M_s - (1/2) \log \Delta \sigma$, which is intermediate of the above two extreme cases. The solid lines in Figure 1 indicate the lines of constant stress drop for this intermediate case. The trend in Figure 1 indicates that the earthquakes having a longer repeat time have a higher average stress drop. Since the details of the scaling relations are unknown, these lines should not be given too much significance; they should be con-

Fig. 2. The relations between the moment magnitude, $M_w$, and the fault length, $L$. The solid lines indicate the trend for a constant stress drop.
and Helmberger [1982] estimate that the average stress drop, $\sigma$, is 5 to 10 bars, but the local stress drop, $\sigma_l$, is about 200 bars.

We assume that the area where a large amount of slip occurred is a strong spot on the fault plane, and will call it the (fault) asperity. Then the average strength of the fault and the average stress drop $\sigma$ are proportional to the ratio, $r_f$, of the total area of the asperity (or asperities, if more than one asperity exists) to the total area of the fault plane. In terms of this asperity model, our results can be interpreted that $r_f$ increases as the repeat time increases.

5. Discussion

Kanamori and Anderson [1975] demonstrate that the average stress drop is higher for intraplate than interplate earthquakes. Since intraplate events have generally longer repeat times than interplate events, our results is essentially the same as that of Kanamori and Anderson [1975]. However, the distinction between "intraplate" and "interplate" is often ambiguous. The use of repeat time, or slip rate, as a parameter in the scaling relation provides a clearer physical basis.

The present result suggests a scheme to estimate strong ground motions of intraplate events with very long repeat times such as those in the eastern United States. Boore [1983] used an $\alpha$-square source model and successfully explained most essential features of strong ground motions of earthquakes in the western United States by scaling the spectrum with an appropriate stress scaling parameter, $\Delta\sigma$. To apply this method to regions such as the eastern United States is difficult because no seismological data to estimate the stress parameter are available. In such a case, geological estimates of the length of a potential causative fault and the repeat time are the key parameters. Given the fault length and the repeat time, we can estimate, from Figure 1, the magnitude and the average stress drop of the expected event. Although the average stress drop and the stress scaling parameter are not necessarily the same, we may assume that they are proportional to each other, since both of them are related to the strength of the fault zone. Once the scaling parameter is estimated, Boore's [1983] method can be applied. For the $\alpha$-square model, the corner frequency is proportional to $\Delta\sigma^{1/3}$, and the high-frequency acceleration spectral amplitude is proportional to $\Delta\sigma^{2/3}$. Hence, for a given seismic moment, a factor of 5 difference in the stress drop suggests a factor of 3 difference in the acceleration spectral amplitude. A factor of 5 difference in the average stress drop is commonly seen between intraplate and interplate earthquakes. [Kanamori and Anderson, 1975; Scholz, 1982].

However, the details of the source spectrum at high frequencies are still unknown. It is possible that the spectral shape is better represented by a model other than the $\alpha$-square model. In that case the above scaling is not appropriate.

In general, faults with short repeat times have large slip rates and vice versa. Our results therefore can be restated that faster moving faults have smaller ratios of asperity area to the total fault area. Although we do not have a direct evidence for this, it is instructive to consider the limiting cases where $V = 0$ or $V \rightarrow \infty$. When $V$ becomes very large, the repeat time decreases. However, at the same time, $\Delta\sigma$ decreases, and the earthquake would have creeplike character and the seismicity may be characterized by frequent small events without any large events. This situation may be compared to that along the transform faults in the East Pacific Rise which represent the fastest moving plate boundary ($= 20$ cm/year). The seismicity there is characterized by the absence of large earthquakes, although this may be partly due to their proximity to the spreading center and to the relatively high-temperature lithosphere there. Along the transform faults in the Mid-Atlantic ridge where the slip rate is very low ($= 2$ cm/year), relatively large ($M_S > 6$) earthquakes occur occasionally. When $V$ becomes very small, the repeat time increases indefinitely and, at a certain point, the fault zone would cease to be seismogenic.

The data set used in this study includes events with strike-slip, thrust and normal-fault mechanisms. It is possible that the average stress drop varies depending upon the fault type: most probably it is greatest for thrust events, least for normal-fault events, and intermediate for strike-slip events [e.g., Sibson, 1974]. No obvious trend is found, however, in the data set used here.

Ruff and Kanamori [1980] show that the strength of plate coupling at subduction zones generally increases with the convergence rate. This conclusion may appear to contradict the conclusion of the present paper. However, the conclusion of Ruff and Kanamori [1980] was for subduction-zone events for which the repeat time is within a very small range, 30 to 200 years, while the present model intends to explain the variation of the stress drop over a much wider range of the repeat time. 30 to several thousand years. The variation of stress drops within a narrow range of repeat time may be controlled by other factors, such as the fault geometry.

6. Conclusion

Existing data on source parameters of crustal earthquakes over a wide range of repeat times indicate that the earthquakes with long repeat times have higher average stress drops than those with short repeat times. The repeat time is therefore a useful parameter to scale seismic spectra.

Recent studies [e.g., Hanks and McGuire, 1981; Boore, 1983] have shown that high-frequency strong motions in the western United States can be explained by an $\alpha$-square source model if the spectrum is scaled by an appropriate stress scaling parameter. To apply this method to regions where no seismological data are available, an estimate of the stress parameter is required. Our results (Figures 1 and 2) may be used to estimate it, if the fault length and the repeat time of a potential causative fault are estimated by geological methods.

Appendix

Source parameters of the events used in this paper

Alaska, July 10, 1958, 06:15:56, 59.3, -136.5
Ms=7.9, $m_p=7.4$ [Abe, 1981].

Moment: Kanamori [1977] gives 2.9 x 10^25 dyne-cm, but this value is estimated from the rupture area, and is not reliable. Ando [1977] gives 4 x 10^27 dyne-cm, and Ben-Menahem [1977] gives 7 x 10^27 dyne-cm.

L: From Kelleher and Savino [1975], L is estimated to be about 300 km. W is assumed to be 16 km.

$\tau$: Plafker et al. [1978] give two recurrence intervals: 110 years or less, and 60 years.

Borah Peak, October 28, 1983, 14:06:22.5, 44.03, -113.91
Ms=7.3 (NEIS).

Moment: Doser and Smith [1985] give 2.1 x 10^26 dyne-cm from body waves. Tanimoto and Kanamori [1985] give 3.4 x 10^26 dyne-cm from long-period surface waves.

L and W: Doser and Smith [1985] give L=21 km for the unilaterial rupture length. L=30 km is inferred from the aftershock area.
W is estimated to be 18 km from the
depth of the main shock. Stein and Bar-
rrientos [1985] used \( L = 21 \) to 35 km, and
\( W = 18 \) km, and obtained a geodetic
moment of \( 3.3 \times 10^{26} \) dyne-cm.

Scott et al. [1985] state that the last dis-
placement occurred between \( 4320 \pm 130 \)
and 6800 years ago. This estimate is based
on radiometric dating. Salyard [1985] esti-
mates it to be 5600 years on the basis of scar
geomorphology.

Borrego Mountain, April 9, 1968, 02:28:59.1, 33.19,
116.13

\( M_s = 6.7 \) [Kanamori and Jennings, 1978].

Moment: \( 1.1 \times 10^{26} \) dyne-cm from body waves [Bur-
dick and Mellman, 1976] and surface
waves [Butler, 1983].

L and W: \( L = 38 \) km from the aftershocks during
the first 22 hours [Allen and Nordquist,
1972]. \( L = 40 \) km from the aftershocks
during the period from April 12 to April
18 [Hamilton, 1972]. \( W \) is estimated to be
13 km from Hamilton [1972].

\( \tau \): Sharp [1981] gives a range 30 to 860
years, but states that, if the magnitude
and displacement of the 1968 event are
typical, approximately one such event per
century at a given point is predicted.

Coyote Lake, August 6, 1979, 17:05:22.3, 37.11, -121.53

\( M_s = 5.7 \) (NEIS)

Moment: \( M_0 = 6 \times 10^{24} \) dyn-cm [Uhrhammer,
1980]. \( M_0 = 3.5 \times 10^{24} \) dyne-cm [Liu and
Helmberger, 1983].

L and W: \( L = 25 \) km [Lee et al., 1979]. \( L = 23 \) km
[Uhrhammer, 1980]. \( W = 8 \) km is con-
sidered appropriate.

\( \tau \): Bakun et al., [1984] estimate the recur-
cence interval to be about 75 years.


\( M_s = 6.8 \) (NEIS)

Moment: \( M_0 = 1.3 \times 10^{26} \) dyne-cm [Zhou et al.,
1983a].

L and W: From the aftershock data, \( L = 46 \) km,
\( W = 10 \) km [Zhou et al., 1983a].

\( \tau \): There is no paleoseismological work on
the Xianshuiku fault. The estimate of slip
rate (5 to 10 mm/year , Tang et al.
[1984] and historical seismicity indicate a
recurrence interval of about 100 years.

Guatemala, February 4, 1976, 09:01:42.2, 15.27, -89.25

\( M_s = 7.5, \ mb = 5.8 \)

Moment: Kanamori and Stewart [1978] obtained
\( 2.6 \times 10^{27} \) dyne-cm from long-period sur-
face waves. This value is consistent with the
geonetic data [Lisowski and
Thatcher, 1981].

L and W: \( L = 250 \) km from the aftershock area
[Langer et al., 1976]. The extent of the
surface break is about 190 km [Plafker.

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1976]. \( W \) is estimated to be 15 km from the
aftershock distribution given by
Langer et al. [1976].

\( \tau \): 180 to 755 years [Schwartz et al., 1979].

Schwartz (1985) suggests an interval of
425 to 725 years between the 1976 earth-
quake and the previous event along the
Motagua fault.


\( M_s = 8.6, \ mb = 7.9 \) [Abe, 1981].

L and W: \( L = 220 \) km is given by Deng et al. [1985].
\( W \) is not given.

\( \tau \): 700 to 1000 years (personal communica-
tion, Qidong Deng, 1985).

Hebgen Lake, August 18, 1959, 06:37:15, 44.7, -110.8

\( M_s = 7.5, \ mb = 7.3 \) [Abe, 1981].

Moment: Doser [1985] gives \( 1 \times 10^{27} \) dyne-cm
from body waves. Geodetic data indicate
\( 1.35 \times 10^{27} \) dyne-cm [Savage and Hastie,
1966].

L and W: Doser [1985] gives \( L = 28 \) km for the uni-
lateral rupture length. Savage and Hastie
[1966] used 30 km. From the focal depth
and the geodetic data, \( W = 15 \) km is con-
sidered appropriate.

\( \tau \): 3250 ± 850 years [Nash, 1981], 2800 ±
1100 years [Nash, 1984].

Imperial Valley, October 15, 1979, 23:16:53.4, 32.61,
-115.32

\( M_s = 6.9 \) is given by NEIS, but this value is strongly
influenced by European data. If a proper azimuthal
average is taken \( M_s = 6.5 \), which is considered to be
more appropriate.

Moment: \( M_0 = 6 \times 10^{25} \) dyne-cm from sur-
face waves [Kanamori and Regan, 1982], and
\( M_0 = 5 \times 10^{25} \) dyne-cm from strong-
motion data [Hartzell and Helmberger,
1982].

L and W: \( L = 42 \) km determined by the distance
between the epicenter and the cluster
near Brawley [Johnson and Hutton,
1982]. \( W \) is assumed to be 10 km.

\( \tau \): 39 years assumed.

Izu, November 25, 1930, 19:02:47, 35.0, 139.0

\( M_s = 7.2, \ mb = 6.8 \) [Abe, 1981].

Moment: \( M_0 = 2.7 \times 10^{25} \) dyne-cm [Abe, 1978].
Kanamori and Anderson [1975] give \( 2.4 \times
10^{25} \) dyne-cm as the average of
Kasahara [1957] and Chinnery [1964].

L and W: \( L = 22 \) km and \( W = 12 \) km [Abe, 1978].
Kanamori and Anderson [1975] give \( L \times
W = 240 \) km² as the average of Kasahara
[1957], Chinnery [1964], and Iida
[1959].

\( \tau \): 700 to 1000 years [Tanna Fault Trench-
ing Research Group, 1983].

Izu-Oki, May 08, 1974, 23:33:25.2, 34.6, 138.8

Luhuo, February 6, 1973, 10:37:10.1, 31.4, 100.6
Ms=7.4 (NEIS)
Moment: \( M_0 = 5.9 \times 10^{25} \) dyne-cm [Abe, 1978].
L and W: L=20 km, W=11 km [Research Group for Aftershocks, 1975, quoted in [Abe, 1978]].
\( \tau \): 1000 years [Matsuda, 1975, Matsuda, 1977].

Morgan Hill, April 24, 1984, 21:15:19.0, 37.32, -121.70
Ms=6.1, mb=5.7

Mikawa, January 12, 1945, 18:38:26, 34.75, 136.75
Ms=6.8, mb=7.2 [Abe, 1981].
Moment: \( M_0 = 1.8 \times 10^{27} \) dyne-cm [Zhou et al., 1983b].
L and W: From the aftershocks L=110 km, and W=15 km [Zhou et al., 1983b].
\( \tau \): See the description for the Daofu earthquake.

Niigata, June 16, 1964, 04:01:40, 38.4, 139.3
Ms=7.5 [Abe, 1981].

Parkfield, June 28, 1966, 04:26:14, 35.92, -120.53
Ms=6.0 [Kanamori and Jennings, 1978].

San Fernando, February 9, 1971, 14:00:41.8, 34.41, -118.4
Ms=6.6 [Kanamori and Jennings, 1978].

Tabas, September 16, 1978, 15:36:56, 33.39, 57.43
Ms=7.4 (NEIS, 12 observations)
Moment: \( M_0 = 1.5 \times 10^{27} \) dyne-cm [Niazi and Kanamori, 1981].
L and W: L=65 km from the aftershock area [Berberian, 1979]. W=20 km is used from figure 2 of Berberian [1979].
\( \tau \): 1100 years [Berberian, 1979].

\( L \times W \times 10^{21} \) dyne-cm as the average of 6 determinations.
L and W: L=60 km [Kayano, 1968], W=25 km (estimated from the vertical extent of the aftershock area). Kayano [1968]. Abe [1975] estimated L x W to be 80 x 30 km² from geodetic data.
\( \tau \): 560 years [Nakamura et al., 1964].

\( L \times W \times 10^{21} \) dyne-cm as the average of 6 determinations.
L and W: L=30 km, W=13 km [Eaton et al., 1970].
\( \tau \): 22 years [Bakun and McEvilly, 1984].

\( L \times W \times 10^{21} \) dyne-cm as the average of 6 determinations.
\( \tau \): 100-300 years [Bonilla, 1973].
orth American plate bound-

by Afa. K .. A fault model for the Niigata earthquake of Division Tottori. September 7, 1927. 09:27:36, 35.75, 134.75

Ms=7.6, m=7.6 [Abe, 1981].

Moment: Kanamori and Anderson [1975] give $M_o=4.6 \times 10^{26}$ dyne-cm on the basis of Kasahara [1957] and Kanamori [1973].

L and W: L=35 km is from the athershock area determined by Nasu [1935]. $W=13$ to 15 km is used by Kasahara [1957] and Kanamori [1973] to interpret the geodetic data.

\[ \tau: \] Matsuda [1977] gives 3 to 6 x 10$^4$ years. However, Matsuda (written communication, 1982) states that this value is very uncertain. Matsuda believes that it is longer than 2000 years.


Ms=7.8 [Abe, 1981]. NEIS gives 7.9

Moment: $M_o=1.8 \times 10^{27}$ dyne-cm [Butler et al., 1979].

L and W: Butler et al. [1979] give L=140 km on the basis of the extent of the athershock area determined from teleseismic data. However, this is probably an overestimate because of the errors involved in teleseismic data. Chinese local data indicate $L=80$ km, if the athershocks of the largest athershock are removed. $W=15$ km is assumed.

\[ \tau: \] Since there seems to have been no historical earthquake on the same fault, the recurrence interval is probably more than 2000 years.

Tottori, September 10, 1943, 08:36:53, 35.25, 134.00

Ms=7,4. m=7.1 [Abe, 1981].

Moment: $M_o=3.6 \times 10^{26}$ dyne-cm [Kanamori, 1972].

L and W: L=33 km is estimated from the aftershocks located by Omote [1955]. Kanamori [1972] used L=33 km and $W=13$ km to interpret the geodetic data.

\[ \tau: \] Matsuda [1977] gives 2 to 6 x 10$^4$ years. However, Matsuda (written communication, 1982) states that this value is very uncertain. Matsuda believes that $\tau>2000$ years. Okada et al. [1981] give 4000 to 8000 years. Tsukuda [1984] gives 6000 years.

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References


