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 (M_S or M_W), 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, r_a, 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 r_a 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 M_S, the seismic moment M_0 (or the corresponding moment magnitude $M_W = (\log M_0 - 16.1)/1.5$), the fault length L, and the fault width W.

The surface-wave magnitude, M_s, is the most widely used parameter and is available for very old events as well as recent events. Although the seismic moment, M_0, is not available for some of the old events, it directly represents the overall size of the source ($M_0 = \mu DS$, 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 aftershock activity is indeed a good approximation of the fault length [Benioff, 1962; Ben-Menahem and Toksöz, 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

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

Event	Ms	Mw	M_O (10^{27} dyne-cm)	L (km)	W (km)	S (km^2)	t (years)
Alaska, 1958	7.9	7.8	7.0	300	16	4800	60-110
Borah Peak, 1983	7.3	7.0	0.34	30	18	540	5600
Borrego Mt., 1968	6.7	6.6	0.1	40	13	520	100
Coyote Lake, 1979	5.7	5.6	0.0035	25	8	200	75
Daofu, 1981	6.8	6.7	0.13	46	10	460	100
Guatemala, 1976	7.5	7.5	2.6	250	15	3750	180-755
Haiyuan, 1920	8.6			220			700-1000
Hebgen Lake, 1959	7.5	7.3	1.0	30	15	450	2800 ± 1100
Imperial Valley, 1979	6.5	6.5	0.06	42	10	420	40
Izu, 1930	7.2	6.9	0.25	22	12	264	700-1000
Izu-Oki, 1974	6.5	6.4	0.059	20	11	220	1000
Kern County, 1952	7.7	7.3	1.0	70	20	1400	170-450
Luhuo, 1973	7.4	7.4	1.8	110	15	1650	100
Mikawa, 1945	6.8	6.6	0.087	12	11	132	$2000\text{-}4\times10^4$
Morgan Hill, 1984	6.1	6.1	0.02	30	10	300	75
N. Anatolian, 1939	7.8			350	15	5250	150-200
N. Anatolian, 1943	7.6			265	15	3975	150-200
N. Anatolian, 1944	7.4			190	15	2850	150-200
Niigata, 1964	7.5	7.6	3.0	60	25	1500	560
Parkfield, 1966	6.0	6.0	0.014	30	13	390	22
Pleasant Valley, 1915	7.7			62			5000
San Fernando, 1971	6.6	6.7	0.12	17	17	289	100-300
Tabas, 1978	7.4	7.4	1.5	65	20	1300	>1300
Tango, 1927	7.6	7.0	0.46	35	13	455	$2000\text{-}6\times10^4$
Tangshan, 1976	7.8	7.4	1.8	80	15	1200	>2000
Tottori, 1943	7.4	7.0	0.36	33	13	429	6000

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, τ, 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^{14} dates, and the amount of slip, D, in a large earthquake judged to be characteristic of the fault are used to estimate τ (= 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]; Ambraseys [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 τ 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

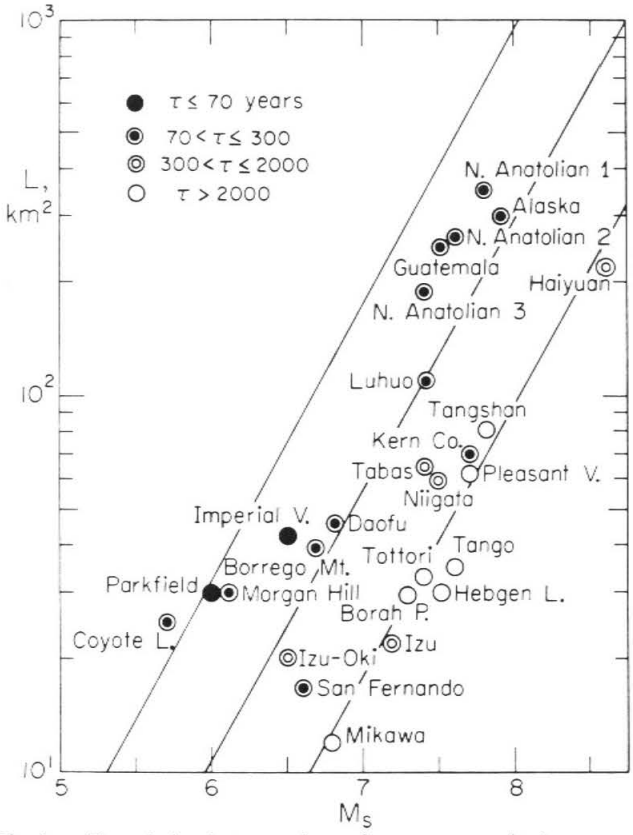

Fig. 1. The relation between the surface-wave magnitude, M_S, and the fault length, L. The solid lines indicate the trend for a constant stress drop.

$$\overline{\Delta\sigma} = \int_s \Delta\sigma D dS / \int_s D dS$$

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 $2E_S$ where E_S is the energy radiated in seismic waves. Although direct determinations of E_S are seldom available, Gutenberg and Richter's [1956] magnitude-energy relation, $\log E_S = 1.5\ M_S + 11.8$, is generally considered a good approximation for earthquakes with $M_S < 8$. Using this relation, the above relation can be written as

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

where W is the width of the fault and $\overline{D}$ is the average dislocation. If both W and $\overline{D}$ are proportional to L, $\log L \propto (1.5/3)M_S - (1/3)\log\overline{\Delta\sigma}$. 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 $\overline{D}$ are fixed, then $\log L \propto (1.5M_S) - \log\overline{\Delta\sigma}$. However, it is unlikely that both W and $\overline{D}$ stay completely constant as L increases. In fact, Scholz [1982] found that $L \propto \overline{D}$ for most crustal earthquakes. In this case, if the variation of W is small, $\log L \propto (1.5/2)M_S - (1/2)\log\overline{\Delta\sigma}$, 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 considered as the reference to which the earthquake data are compared. A use of the other relations (e.g., $\log L \propto 0.5\ M_S$, $\log L \propto 1.5\ M_S$) does not affect the conclusion of this paper qualitatively.

Figure 2 shows the relation between M_W and L, which is essentially similar to Figure 1.

4. Interpretation

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., Scholtz 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 Apsel [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

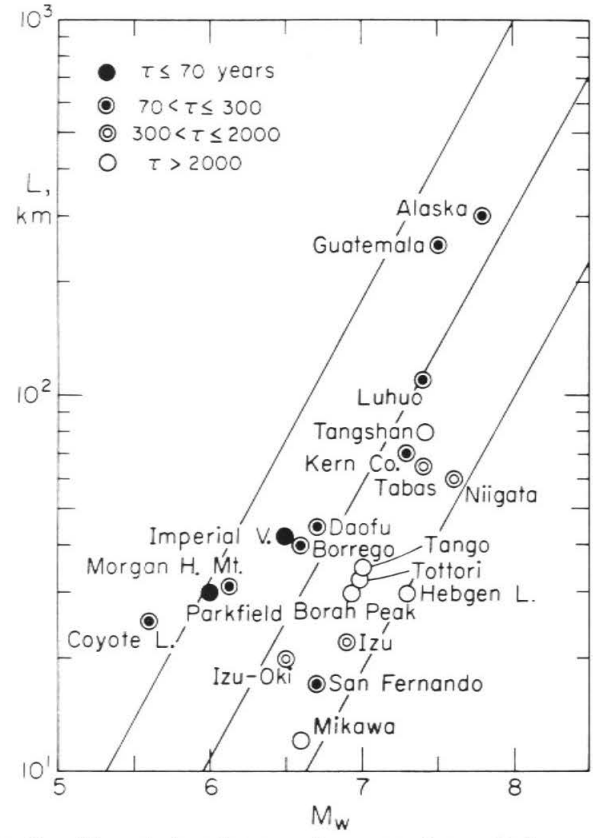

Fig. 2. The relations between the moment magnitude, M_W, and the fault length, L. The solid lines indicate the trend for a constant stress drop.

and Helmberger [1982] estimate that the average stress drop, $\overline{\Delta\sigma}$, is 5 to 10 bars, but the local stress drop, $\Delta\sigma_a$, 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 $\overline{\Delta\sigma}$ are proportional to the ratio, r_a, 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_a 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 result 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 ω-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, $\Delta\sigma_s$. 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 ω-square model, the corner frequency is proportional to $\Delta\sigma_s^{1/3}$, and the high-frequency acceleration spectral amplitude is proportional to $\Delta\sigma_s^{2/3}$. 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 ω-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 $V \to 0$ or $V \to \infty$. When V becomes very large, the repeat time decreases. However, at the same time, $\overline{\Delta\sigma}$ 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 ($\approx$ 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 ($\approx$ 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 ω-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

M_S=7.9, m_B=7.4 [Abe, 1981].

Moment: Kanamori [1977] gives 2.9 x 10^{28} 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 and W: From Kelleher and Savino [1975], L is estimated to be about 300 km. W is assumed to be 16 km.

τ: 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

M_S=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 unilateral 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 Barrientos [1985] used L=21 to 35 km, and W=18 km, and obtained a geodetic moment of 3.3 x 10^{26} dyne-cm.

τ: Scott et al. [1985] state that the last displacement occurred between 4320 ± 130 and 6800 years ago. This estimate is based on radiometric dating. Salyard [1985] estimates it to be 5600 years on the basis of scarp geomorphology.

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

Ms=6.7 [Kanamori and Jennings, 1978].

Moment: 1.1x10^{26} dyne-cm from body waves [Burdick 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].

τ: 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

Ms=5.7 (NEIS)

Moment: M_0=6 x 10^{24} dyn-cm [Uhrhammer, 1980]. M_0= 3.5 x 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 considered appropriate.

τ: Bakun et al., [1984] estimate the recurrence interval to be about 75 years.

Daofu, January 23, 1981, 21:13:51.7, 30.927, 101.098

Ms=6.8 (NEIS)

Moment: M_0=1.3 x 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].

τ: There is no paleoseismological work on the Xianshuihe 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

Ms=7.5, mb=5.8

Moment: Kanamori and Stewart [1978] obtained 2.6 x 10^{27} dyne-cm from long-period surface waves. This value is consistent with the geodetic 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, 1976]. W is estimated to be 15 km from the aftershock distribution given by Langer et al. [1976].

τ: 180 to 755 years [Schwartz et al., 1979]

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

Haiyuan, December 16, 1920, 12:05:48, 36., 105.

Ms=8.6, m_B=7.9 [Abe, 1981].

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

τ: 700 to 1000 years (personal communication, Qidong Deng, 1985).

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

Ms=7.5, m_B=7.3 [Abe, 1981].

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

L and W: Doser [1985] gives L=28 km for the unilateral rupture length. Savage and Hastie [1966] used 30 km. From the focal depth and the geodetic data, W=15 km is considered appropriate.

τ: 3250 ± 850 years [Nash, 1981], 2800 ± 1100 years [Nash, 1984].

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

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

Moment: M_0=6 x 10^{25} dyne-cm from surface waves [Kanamori and Regan, 1982], and M_0=5 x 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.

τ: 39 years assumed.

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

Ms=7.2, m_B=6.8 [Abe, 1981].

Moment: M_0=2.7 x 10^{26} dyne-cm [Abe, 1978]. Kanamori and Anderson [1975] give 2.4 x 10^{26} 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 x W=240 km^2 as the average of Kasahara [1957], Chinnery [1964], and Iida [1959].

τ: 700 to 1000 years [Tanna Fault Trenching Research Group, 1983].

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

Ms=6.5 (NEIS)

Moment: M_0=5.9 x 10^{25} dyne-cm [Abe, 1978].

L and W: L=20 km, W=11 km [Research Group for Aftershocks, 1975 , quoted in [Abe, 1978]].

τ: 1000 years [Matsuda, 1975, Matsuda, 1977].

Kern County, July 21, 1952, 11:52:14, 35.0, -119.02

Ms=7.7 [Kanamori and Jennings, 1978]. Abe [1981] gives 7.8.

Moment: Kanamori and Anderson [1975] give 2 x 10^{27} dyne-cm from Kanamori's unpublished result. Ben-Menahem [1977] gives 0.84 x 10^{27} dyne-cm. Stein and Thatcher [1981] give 0.84 x 10^{27} dyne-cm from geodetic data. M0=1 x 10^{27} dyne-cm seems appropriate.

L and W: From Benioff [1955] and Stein and Thatcher [1981], L=70 km, and W=20 km are considered appropriate.

τ: 170 to 450 years [Stein and Thatcher, 1981].

Luhuo, February 6, 1973, 10:37:10.1, 31.4, 100.6

Ms=7.4 (NEIS)

Moment: M_0=1.8 x 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].

τ: See the description for the Daofu earthquake.

Mikawa, January 12, 1945, 18:38:26, 34.75, 136.75

Ms=6.8, m_B=7.2 [Abe, 1981].

Moment: Ando [1974] gives M_o=8.7 x 10^{25} dyne-cm from geodetic data.

L and W: L=12 km and W=11 km [Ando, 1974].

τ: Matsuda [1977] gives 2 to 4 x 10^4 years; however, Matsuda [1982, written communication] states that this value is subject to large uncertainty. According to Matsuda, τ>2000 years.

Morgan Hill, April 24, 1984, 21:15:19.0, 37.32, -121.70

Ms=6.1, mb=5.7

Moment: 2.0 x 10^{25} dyne-cm [Ekstrom, 1985]. This value agrees well with that determined from long-period Rayleigh waves recorded by the IDA network, 2.3 x 10^{25} dyne-cm.

L and W: The aftershock area determined by Cockerham and Eaton [1985] indicate L=30 km, and W=10 km.

τ: Bakun et al. [1984] estimate the recurrence time to be approximately 75 years.

N. Anatolian 1, December 26, 1939, 23:57:21, 39.5, 38.5

Ms=7.8, m_B=7.7 [Abe, 1981].

L and W: L=350 km Slemmons [1977]. W=15 km assumed.

τ: 200 years [Allen, 1975]. 150 years [Ambraseys, 1970].

N. Anatolian 2, November 26, 1943, 22:20:36

Ms=7.6, m_B=7.3 [Abe, 1981].

L: L=265 km [Slemmons, 1977], W=15 km is assumed.

τ: 200 years [Allen, 1975]. t=150 years [Ambraseys, 1970].

N. Anatolian 3, February 1, 1944, 03:22:36, 41.5, 32.5

Ms=7.4, m_B=7.5

L: L=190 km [Slemmons, 1977], W=15 km is assumed.

τ: 200 years [Allen, 1975]. 150 years [Ambraseys, 1970].

Niigata, June 16, 1964, 04:01:40, 38.4, 139.3

Ms=7.5 [Abe, 1981].

Moment: M_0=3 x 10^{27} dyne-cm [Aki, 1966], Mo0=3.2 x 10^{27} dyne-cm [Abe, 1975].

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 LxW to be 80 x 30 km^2 from geodetic data.

τ: 560 years [Nakamura et al., 1964].

Parkfield, June 28, 1966, 04:26:14, 35.92, -120.53

Ms=6.0 [Kanamori and Jennings, 1978].

Moment: Purcaru and Berckhemer [1982] give 1.4 x 10^{25} dyne-cm as the average of 6 determinations.

L and W: L=30 km, W=13 km [Eaton et al., 1970].

τ 22 years [Bakun and McEvilly, 1984].

Pleasant Valley, October 3, 1915

Ms=7.7, m_B=7.3

L and W: L=62 km [Slemmons, 1977].

τ 5000 years [Wallace [1984]; and private communication, 1985].

San Fernando, February 9, 1971, 14:00:41.8, 34.41, -118.4

Ms=6.6 [Kanamori and Jennings, 1978].

Moment: Kanamori and Anderson [1975] give 1.2 x 10^{27} dyne-cm as the average of 8 determinations. Heaton and Helmberger [1979] give 1.4 x 10^{26} dyne-cm.

L and W: L=20 km, and W=20 km [Allen et al., 1971]. Kanamori and Anderson [1975] give 17 x 17 km^2 as an average.

τ: 100-300 years [Bonilla, 1973].

Tabas, September 16, 1978, 15:36:56, 33.39, 57.43

Ms=7.4 (NEIS, 12 observations)

Moment: M_0=1.5 x 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].

τ:τ>1100 years [Berberian, 1979].

Tango, March 7, 1927, 09:27:36, 35.75, 134.75

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

Moment: Kanamori and Anderson [1975] give M_0=4.6 x 10^{26} dyne-cm on the basis of Kasahara [1957] and Kanamori [1973].

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

τ: 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.

Tangshan, July 27, 1976, 19:42:54, 39.6, 117.9

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

Moment: M_0=1.8 x 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 aftershock 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 aftershocks of the largest aftershock are removed. W=15 km is assumed.

τ: 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_B=7.1 [Abe, 1981].

Moment: M_0=3.6 x 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.

τ: 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 τ>2000 years. Okada et al. [1981] give 4000 to 8000 years. Tsukuda [1984] gives 6000 years.

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