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Estimating maximum expectable magnitude of earthquakes from fault dimensions

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Notes

Estimating maximum expectable magnitude of earthquakes from fault dimensions

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ABSTRACT

The evaluation of seismic risk at locations where sensitive man-made structures are planned depends critically on a correct estimate of the maximum expectable earthquake magnitude, M_{max}, in that region. By assuming that the longest fault (or fault unit) with length L_{max} could break in a single earthquake, one estimates M_{max} from L_{max} on the basis of a magnitude versus source-length relation, which is derived empirically. The maximum expectable ground accelerations are then estimated from M_{max}. I propose that a more accurate estimate of M_{max} can be obtained by determining the maximum expected rupture area, A_{max}, and using the magnitude-area relation $M = \log A + 4.15$ (valid for $M > 5.6$). A_{max} can be obtained from the product of L_{max} times the expected fault width. The latter can probably be estimated more accurately than L_{max} on the basis of tectonic analysis and microearthquakes studies. The M_{max} estimates derived from rupture *area* give more accurate results than the estimates based on rupture *length* alone, because narrow faults produce less powerful earthquakes than do wide faults of the same length.

INTRODUCTION

Recently the discussion was reopened in *Geology* (Mark, 1977; Bolt, 1978) on how to estimate the most likely value for the maximum magnitude, M_{max}, of an earthquake from the lengths of geologically mapped active faults. This question is fundamental in assessing the earthquake hazard to nuclear power plants and dams. The aim is to estimate the maximum accelerations that could result at the proposed construction site owing to the largest likely earthquake rupture along faults mapped near the site. The expected accelerations are usually estimated from M_{max} using empirical relations involving magnitude, source distance, and attenuation. The question addressed here is how to obtain the best estimate of M_{max} if the dimensions of active faults in the area are mapped and the length of the longest fault segment, L_{max}, that could rupture in one earthquake is identified (this could be the length of an entire fault).

The following question arises here: Should L_{max} be assumed to equal the *source length* of the next maximum earthquake along this fault, or is it possible that L_{max} will be only the *surface rupture*, with the source length deeper in the crust extending for an order of magnitude beyond L_{max}? I believe we must assume that L_{max} equals the source length of the earthquake with M_{max}, because the mapped surface trace is the result of many ruptures over geologic time. No fault can remain hidden at shallow depth in the crust with repeated displacements along it without reaching the surface, where it can be mapped.

M_{max} is estimated empirically from L_{max}, from data on a set of earthquakes from which a linear fit for the magnitude-source dimension relation of the form

$$M = a \log L + b \qquad (1)$$

was derived. Here it is assumed that the energy radiated from crustal volumes of dimensions L_{max} will be about the same as in past earthquakes with the same dimensions. This is a good first-order assumption for large earthquakes, because in large volumes the average energy density in the Earth's crust is approximately constant, as evidenced by fairly constant stress drops of large earthquakes (see, for example, Kanamori and Anderson, 1975).

Some of the data sets that were compiled to establish the constants in equation 1 contain surface rupture lengths for L that are ten times smaller than the source dimensions (for example, Bonilla and Buchanan, 1970). Surface rupture lengths that are not equal to the source lengths are interesting for some geological and engineering problems. For example, many shallow earthquakes in the range of $5 < M < 6$ do not produce surface ruptures, which is important information for estimating the probability of a dam failure due to a surface dislocation at the base of the dam. However, for estimating the energy (of M) radiated from an earthquake source with dimension L_{max} by equation 1, L_{max} has been defined above as source length, and all partial surface ruptures must be excluded from the data set, as was done in a recent summary of rupture lengths by Slemmons (1977).

The articles in *Geology* so far discussing the estimate of $M_{\max}$ by equation 1 have concentrated on the mathematical methods of fitting a curve through length versus magnitude data graphs. Mark (1977) pointed out that the curves fitted through most length-magnitude data were regressions of L on M, which do not allow the correct estimate of M if L is given. Bolt (1978) further showed that using the method of least-squares regression and treating L as error free may produce a biased estimate.

In addition to arguing that L should be defined as source length, not surface rupture, I propose that one can get a better estimate of $M_{\max}$ from the *rupture area, A,* by an equation of the form

$$M = \alpha \log A + \beta, \qquad (2)$$

rather than from *rupture length* alone.

DATA

The surface rupture length for some shallow earthquakes furnishes a good estimate of the source length (for example, San Francisco, 1906; Borrego Mountain, 1968; San Fernando, 1971). However, it is also well known that some shallow earthquakes do not rupture through to the surface or rupture only along parts of the source length. Therefore, other data on source length must be used to check the validity of the assumption that the surface rupture extended over the entire source length. Methods that allow such checks include static dislocation modeling to fit geodetic displacement measurements, distribution of aftershocks, and seismic signal analysis of surface or body waves.

Table 1 lists seven earthquakes for which the surface rupture as quoted by Bonilla and Buchanan (1970) underestimates the source dimension by factors ranging from 5 to 45. For the sake of brevity I will not discuss the data on these source lengths in detail, especially because Slemmons (1977) quoted essentially the same source lengths as those given in Table 1, which in most cases are averages of the estimates of several investigators who used different methods to estimate source length. In addition, the Kagi (Taiwan, 1907) and the Manix (California, 1947) earthquakes, which had magnitudes of M_S = 7.1 and 6.4, respectively, must be omitted from the calibration set because doubt exists that the surface rupture equaled the source length in these cases (Richter, 1958). Finally, I will make the minor adjustment to double the length quoted by Bonilla and Buchanan (1970) for the Tango earthquake (M_W = 7.0, 1927), on the basis of detailed investigations (Utsu, 1969; Kanamori, 1973). Figure 1 shows the data set with the partial surface-rupture values (crosses) replaced by the improved source-length data presented here (solid dots). The scatter in Figure 1 is reduced by these adjustments; however, it is still large. Many authors have noted that differences exist in equation 1 depending on region and fault type, but because of insufficient data these differences cannot always be well defined. World average curves are usually used to estimate $M_{\max}$.

I suggest that the most valid average curve is obtained for *rupture area* versus magnitude. An estimate of expected magnitude based on area will be better than estimates based on length because a long, thin fault can release less elastic strain energy than a long, wide fault. Because we now understand tectonic processes much better than we did 10 yr ago, we are able to estimate the maximum expected fault width, $W_{\max}$, with the same or less uncertainty as we can estimate $L_{\max}$ of a future earthquake. $W_{\max}$ can probably be estimated correctly to within less than a factor of 2. Thus, from $A = W \cdot L$ the magnitude of the long, thin 1906 San Francisco earthquake would correctly be estimated as smaller than the magnitude of the shorter but wider 1897 Assam earthquake.

Data on area versus magnitude are shown in Figure 2. Some of the best data were summarized by Kanamori and Anderson (1975). In these cases the original data sources were not listed for the

TABLE 1. SOURCE LENGTH COMPARED TO SURFACE RUPTURE LENGTH

Location	Date	Magnitude M_S	Source Length	Source Width	Surface rupture* (km)	Factor by which source length was longer than surface break
Alaska, United States	1964	9.2†	700 ± 100	150	63	11
Inangahua, New Zealand	1968	7.1	45	25	1	45
Assam, India	1897	8.7	240	120	19	13
Niigata, Japan	1964	7.4	100	20	20	5
Tottori, Japan	1943	7.4	37	12	8	5
Tesikaga, Japan	1959	6.2	10 to 13	(A = 100)	2	5
Hawke's Bay, New Zealand	1931	7.9	100	30	10	10

*From Bonilla and Buchanan (1970).

†M_w instead of M_s.

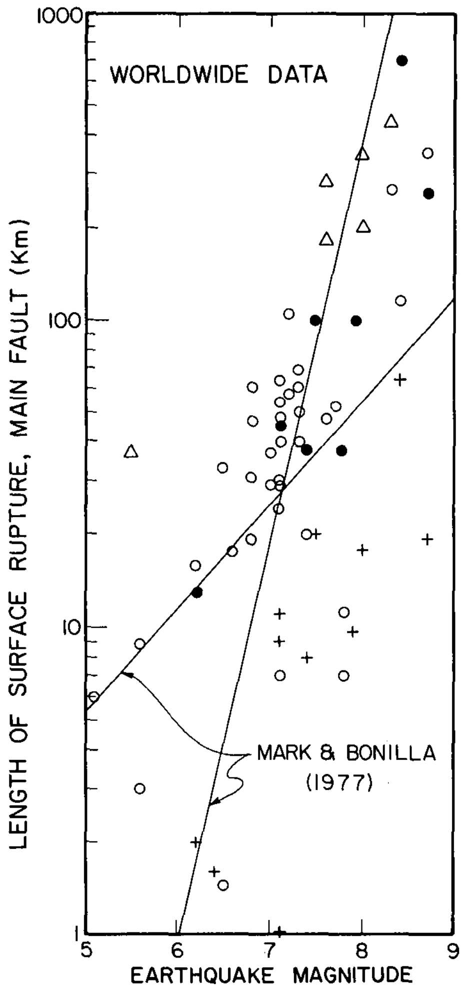

Figure 1. Earthquake length estimated from surface rupture as a function of surface-wave magnitude. Open circles and open triangles (long, thin ruptures) are unchanged from Bonilla and Buchanan (1970), with regression curves fitted by Mark and Bonilla (1977); crosses are data from Bonilla and Buchanan that do not represent source length. To estimate expectable magnitude, crosses must be omitted (three values) or replaced by true source length (eight values replotted as solid dots).

sake of brevity. As the definition of aftershock areas by Utsu and Seki (1954) and Utsu (1969) appears to overestimate the source dimensions, only their best defined data were included (Table 2; Fig. 2). The rupture area was calculated for Turkish and Asian earthquakes from the data of Ambraseys and Zatopek (1968) and Chen and Molnar (1977), respectively. Other high-quality data were also added (Table 2). Note that in Figure 2 the long, thin faults of the transform type (for example, San Andreas, Anatolia) now fit the average curve much better than in Figure 1 (triangles).

The shortcomings of the surface-wave-magnitude scale introduce some problems: For $M < 6.5$ the magnitudes used are generally local magnitudes, M_L, and for great earthquakes ($M > 8$) the M_S scale becomes saturated (for example, Kanamori, 1977). At the lower end of the data, M_L was taken if M_S was not available, because the difference between the two scales is not large in this range. On the other hand, for great earthquakes, M_S commonly underestimates the earthquake strength considerably (for example, Kanamori and Anderson, 1975). Therefore, the extension of the M_S scale, called the M_W scale (Kanamori, 1977), was used for all earthquakes where M_W was available. The use of the M_W scale brings the data of the greatest earthquakes nicely in line with the curve defined by most of the large earthquakes in Figure 2.

It seems that the data set of 90 area estimates shown in Figure 2 has eliminated most of the gross errors and provides good coverage for the range $5.6 < M < 9.6$. This set, therefore, can be used for deciding on a relation of the form of equation 2 for estimating expected magnitude. This brings us back to the question discussed by Mark (1977) and Bolt (1978) of how to determine the constants α and β. First, we note that the least-squares fits by Utsu and Seki (1954) and Utsu (1969) resulted in slopes of approximately 1.0. Wyss and Brune (1968) also found that for small earthquakes in California $\alpha \approx 1.0$.

Next, one can show that the theoretically expected slope equals 1.0. Wyss and Brune (1968) and Thatcher and Hanks (1973) had found empirically that the moment magnitude relation is

$$\log M_0 = 1.5\,M + \text{constant}, \qquad (3)$$

where M can be M_L or M_S (Kanamori and Anderson, 1975). By substituting the following expression for the moment,

$$M_0 = k_1 r^3\, \Delta\sigma$$

(r = source radius, $\Delta\sigma$ = stress drop) in equation 3, we obtain

$$M \sim \tfrac{2}{3} \log M_0 = \tfrac{2}{3} \log (k_1\,(\Delta\sigma)\, r^3) \sim k_2\,(\Delta\sigma)^{2/3}\, r^2 = k_3\, \Delta\sigma^{2/3} A. \qquad (4)$$

The symbols k_i denote constants that are of no consequence, and we assume the area $A = \pi r^2$. By equation 4 we see that magnitude will be directly proportional to A if the stress drop is constant. The evidence for constant stress drop in large earthquakes and the theoretical relation between moment and magnitude was dis-

TABLE 2. EARTHQUAKE RUPTURE AREAS

Date	Location	M_s	Length (km)	Width (km)	Area (km^2)	Reference
09 Jul 05	Mongolia	8.4*	200	40†	8,000	Okal (1977)
23 Jul 05	Mongolia	8.4*	300	40†	12,000	Okal (1977
18 Apr 06	San Francisco	8.3	450	10	4,500	Thatcher (1975)
03 Jan 11	Kebin, Kirgizia	7.7*	180	40	7,200	Chen and Molnar (1977)
16 Dec 20	Haiyuan, Kansu	7.8*	200	50	10,000	Chen and Molnar (1977)
01 Sep 23	Kanto	7.9*			6,900	Kanamori and Anderson (1975)
22 May 27	Ku-Long, Kansu	7.6*	150	40	6,000	Chen and Molnar (1977)
07 Jul 29	Aleutians	7.3	100	90	900	Sykes (1971)
25 Nov 30	North Izu	7.1			240	Kanamori and Anderson (1975)
10 Aug 31	Fuh-Yun, Sinkiang	8.0*	300	50	15,000	Chen and Molnar (1977)
21 Sep 31	Saitama	6.75			200	Kanamori and Anderson (1975)
02 Mar 33	Sanriku	8.4*			18,500	Kanamori and Anderson (1975)
11 Mar 33	Long Beach	6.25			450	Kanamori and Anderson (1975)
15 Jan 34	Bihar-Nepal	8.1*	130	50	6,500	Chen and Molnar (1977)
21 Feb 36	Kawachi-Yamato	6.5			230	Utsu and Seki (1954)
19 Apr 38	Turkey	6.75	15	14	210	Ambraseys and Zatopek (1968)
05 Nov 38	Fukushima-oki	7.7			10,000	Utsu and Seki (1954)
01 May 39	Oga Peninsula	6.9			520	Utsu and Seki (1954)
26 Dec 39	Anatolia	8.0	350	20	7,000	Ambraseys and Zatopek (1968)
19 May 40	Imperial Valley	7.1			780	Kanamori and Anderson (1975)
20 Dec 42	Anatolia	7.25	70	10	700	Ambraseys and Zatopek (1968)
26 Nov 43	Anatolia	7.6	280	14	3,920	Ambraseys and Zatopek (1968)
01 Feb 44	Anatolia	7.6	180	20	3,600	Ambraseys and Zatopek (1968)
07 Dec 44	Tonankai	8.1*			9,600	Kanamori and Anderson (1975)
20 Dec 46	Nankaido	8.1*			9,600	Kanamori and Anderson (1975)
14 May 48	Alaska Peninsula	7.5	50	50	2,500	Sykes (1971)
28 Jun 48	Fukui	7.3			390	Kanamori and Anderson (1975)
10 Jul 49	Khait, Tajikistan	7.6	70	30	2,100	Chen and Molnar (1977)
22 Aug 49	Queen Charlotte, Canada	8.1	380	10†	3,800	Sykes (1971)
26 Dec 49	Near Imaichi	6.5			120	Utsu and Seki (1954)
15 Aug 50	Assam, India	8.6*	250	80	20,000	Chen and Molnar (1977)
04 Mar 52	Tokachi-oki	8.1*			19,000	Kanamori and Anderson (1975)
07 Mar 52	Daishoji-oki	6.6			600	Utsu and Seki (1954)
21 Jul 52	Kern County	7.7			1,400	Kanamori and Anderson (1975)
18 Mar 53	Turkey	7.4	50	18	900	Ambraseys and Zatopek (1968)
16 Dec 54	Fairview Peak	7.1			220	Kanamori and Anderson (1975)
21 Dec 56	Miyakejima	6.0			550	Utsu (1969)
26 May 57	Anatolia	7.1	50	8	400	Ambraseys and Zatopek (1968)
27 Jun 57	Muya, Siberia	7.4*	35	30	1,050	Chen and Molnar (1977)
04 Dec 57	Gobi-Altai, Mongolia	8.1*	270	50	13,500	Chen and Molnar (1977)
07 Apr 58	Central Alaska	7.3	85	10	850	Sykes (1971)
10 Jul 58	Fairweather, Alaska	7.9	350	15+	5,250	Tocher (1960); Slemmons (1977)
22 May 60	Chile	9.5*			200,000	Kanamori and Anderson (1975)
13 Nov 60	Aleutian	7.0			300	Sykes (1971)
07 May 61	Hyogo Prefecture	5.9			570	Utsu (1969)
19 Aug 61	Kitamino	7.0			570	Kanamori and Anderson (1975)
30 Apr 62	Miyagi Prefecture	6.5			330	Utsu (1969)
27 Mar 63	Wakasa Bay	6.9			160	Kanamori and Anderson (1975)
03 Aug 63	North Atlantic	6.7			350	Kanamori and Anderson (1975)
13 Oct 63	Kuril Islands	8.5*			44,000	Kanamori and Anderson (1975)
17 Nov 63	North Atlantic	6.5			240	Kanamori and Anderson (1975)
15 Mar 64	Spain	7.1			1,000	Kanamori and Anderson (1975)
04 Feb 65	Rat Island	8.7*	650	120	78,000	Kanamori and Anderson (1975)
30 Mar 65	Rat Island	7.5			4,000	Kanamori and Anderson (1975)
28 Jun 66	Parkfield	6.4			260	Kanamori and Anderson (1975)
04 Jul 66	Aleutian Islands	7.2			420	Kanamori and Anderson (1975)
19 Aug 66	Anatolia	6.8	30	10	300	Ambraseys and Zatopek (1968)
12 Sep 66	Truckee	5.9			100	Kanamori and Anderson (1975)
17 Oct 66	Peru	8.1*	130	70	9,000	Abe (1972); Dewey and Spence (1979)
12 Nov 66	South of Hokkaido	5.9			210	Utsu (1969)
05 Jan 67	Mogod, Mongolia	7.4	40	30	1,200	Chen and Molnar (1977)
22 Jul 67	Mudurnu, Turkey	7.1	80	20	1,600	Ambraseys and Zatopek (1969)
21 Feb 68	Ebino	6.1			85	Utsu (1969)
09 Apr 68	Borrego Mountain	6.7			450	Kanamori and Anderson (1975)
16 May 68	Tokachi-oki	8.2*			15,000	Kanamori and Anderson (1975)
01 Jul 68	Saitama	5.8			60	Kanamori and Anderson (1975)
05 Aug 68	Ehime Prefecture	6.6			380	Utsu (1969)
31 Aug 68	Dasht-e-Bayaz, Iran	7.3	80	20	1,600	Ambraseys and Tchalenko (1969)
28 Feb 69	Portugal	7.8*			4,000	Kanamori and Anderson (1975)
28 Apr 69	Coyote Mountain	5.8**	10	3	30	Thatcher and Hamilton (1973)
11 Aug 69	Kurile Islands	8.2*			15,300	Kanamori and Anderson (1975)
09 Sep 69	Gifu	6.6			190	Kanamori and Anderson (1975)
31 May 70	Peru	7.9*	140	55	7,700	Abe (1972; Dewey and Spence (1969)
09 Feb 71	San Fernando	6.6			280	Kanamori and Anderson (1975)
30 Jul 72	Sitka, Alaska	7.1	170	10	1,700	R. Page, W. Gawthrop (personal commun.)
23 Dec 72	Managua	6.2	15	8	120	Brown and others (1973)
16 Jun 73	Nemuro-oki	7.7			6,000	Kanamori and Anderson (1975)
03 Oct 74	Peru	8.1*	200	50	10,000	Dewey and Spence (1979)
01 Aug 75	Oroville, California	5.7**	15	7	105	Lahr and others (1976)
20 Nov 75	Hawaii	7.2	40	15	600	M. Ando (in prep.)
04 Feb 76	Motagua, Guatemala	7.5	300	20	6,000	Plafker (1976)
24 Nov 76	Turkey	7.3	55	18	1,000	Toksöz and others (1977)
04 Nov 77	Aleutian Islands	6.7	31	10	310	Billington and Engdahl (1978)

*M_W instead of M_S.

†Assumed by similarity to other quakes in same area.

**M_L instead of M_S.

cussed explicitly by Kanamori and Anderson (1975). On the basis of the empirical and theoretical considerations above, I propose that we settle on $\alpha = 1$, regardless of fitting method.

Next comes the question of β. Bolt (1978) pointed out that all L (or area) and M values are affected by errors. For some data points, approximate errors can be estimated; for many less well-documented events, errors are hard to estimate. Also, it is especially important to avoid strong influence on the results by the data at the low- and high-magnitude ends; for these the conventional M_S definition cannot be used. Therefore, the solution may be to treat the data set as one in which individual points may have large but unknown errors. Therefore, I fit a median with slope 1 through the data set. The resulting relation between magnitude and area is

$$M = \log A + 4.15, \qquad M > 5.6, \quad (5)$$

with A in square kilometres. If the slope of the curve is not fixed at 1, the least-squares regression gives

$$M = 0.93 \log A + 4.38, \qquad M > 5.6, \quad (6)$$

with $\sigma_\alpha = \pm 0.04$, $\sigma_\beta = \pm 0.13$, and $\sigma_M = \pm 0.30$, assuming the parameter A to be error free. For the reasons given above, I propose that equation 5 be used for estimating the most likely magnitude expected from rupture of an area A. In practice, A is the largest rupture area expected on a given fault A_{max}. The resulting estimate of M is therefore called the maximum expectable magnitude, even though M_{max} is an average estimate.

The root-mean-square error of M for the data is ±0.3 units in equation 5, but this variance is not all due to errors. Some of it is due to differences in the elastic energy density per unit source volume; that is, high-stress earthquakes produce a larger magnitude for the same source area compared to low-stress earthquakes according to equation 4. The high-stress earthquakes plot, therefore, below the average curve, which represents stress drops of about 70 bar.

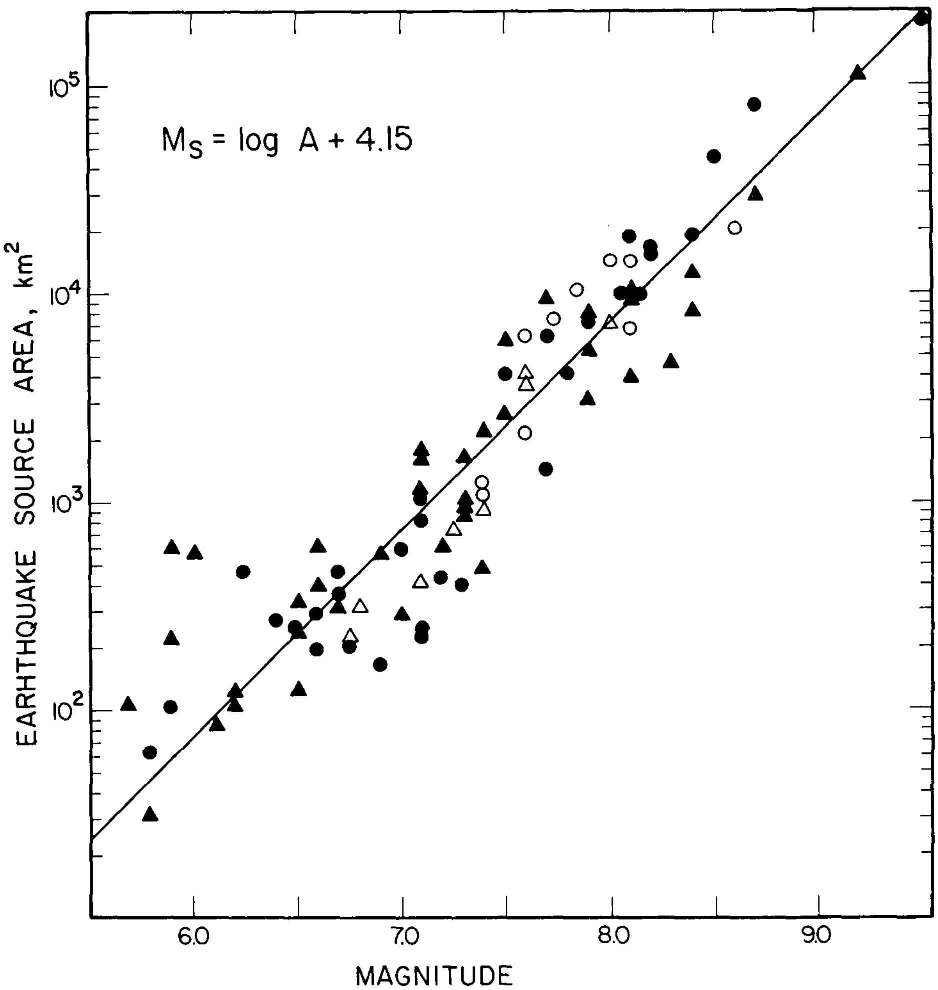

Figure 2. Earthquake rupture area as a function of magnitude (M_L, M_S, M_W). Data from Tables 1 and 2. For facilitating identification, data from Kanamori and Anderson (1975), Chen and Molnar (1977), and Ambraseys and Zatopek (1968) are differentiated by solid dots, open circles, and open triangles, respectively. Large earthquakes for which M_W was estimated from rupture area were not used.

On theoretical grounds, M is expected to depend on LD^2 (King and Knopoff, 1968); therefore, this function is sometimes plotted (Bonilla and Buchanan, 1970; Slemmons, 1977). I believe it is very difficult to estimate the average source dislocation, D, before the earthquake occurs. However, if we actually can estimate D in addition to L and W, we can calculate the stress drop and the entire spectrum of the radiated waves (assuming an f^{-2} source, where f is frequency). From this we can calculate the expected ground motion (for details, see Hanks, 1976) at any frequency and distance (if attenuation is approximately known), and therefore we could assess the seismic risk much more accurately than from an estimate of the much less informative parameter M_S. Therefore, if D is known, M_S is not needed, but if D (or stress drop) is not known in advance, then M_S can be estimated from A using the data in Figure 2.

Two practical examples of estimating M on the basis of A from equation 5 are those of the Sitka (1972) and the Tokachi-Oki (1968) earthquakes. These events filled seismic gaps identified by Sykes (1971) and Fedotov (1965), respectively. The lengths of both seismic gaps were about 170 km. The Sitka gap was located on a strike-slip plate boundary, whereas the Tokachi-Oki earthquake was located in a subduction zone. The most reasonable values for the widths would therefore have been W(SIT) = 10 km (in analogy to the San Andreas fault), and W(TOK) = ⅔L = 113 km. The corresponding estimates for the rupture areas would thus have been calculated as A(SIT) = $1.7 \cdot 10^3$ km^2 and A(TOK) = $1.9 \cdot 10^4$ km^2, and by equation 5 the expectable maximum magnitudes would have been estimated as M(SIT) = 7.4 (observed was M_S = 7.1), and M(TOK) = 8.4 (observed was M_W = 8.2). Even with somewhat different L or W estimates, the respective magnitudes would have been estimated correctly to within 0.3 units. Even though the two gaps had about the same length, the estimated magnitudes differed by a full magnitude unit, with the Sitka earthquake estimated correctly to be the less powerful event.

The ultimate quality of maximum expectable magnitude estimates depends strongly on the correct identification of L_{max} or A_{max} that may break in one earthquake. If either of the gaps discussed above had been filled in by two consecutive and abutting ruptures, instead of by one, our estimate of M_{max} would have been too large.

The estimate of L_{max} is obtained by mapping active faults, by estimating which transverse feature of such faults may segment it enough to stop a rupture there, and from the maximum dimensions of historic large and great earthquakes in the area. The maximum width can also be estimated from the largest historic earthquakes in the area and by microseismic surveys in which the width of the active fault can be mapped (for example, San Andreas fault). Generally, I propose that for transform faults (vertical strike slip) and for normal faults $5 \leq W \leq 20$ km should be used unless evidence exists for an especially thick brittle crust, as in Tibet (Chen and Molnar, 1977). For thrust earthquakes, $70 \leq W \leq 150$ km is appropriate. However, in all cases one should assume that $W \leq \frac{2}{3}L$. I believe that it is very unlikely that W_{max} will be misjudged in excess of a factor of two; however, it can be a difficult problem to estimate L_{max} correctly to within that accuracy.

CONCLUSIONS

The maximum expectable magnitude of earthquakes can be estimated from the length of geologically mapped faults under the assumptions that the entire fault will rupture in one event and that one knows the correct magnitude-source dimension relation from past earthquakes. Earthquakes commonly rupture only partially through to the surface, and generally they produce subsidiary ruptures. In these cases, the surface rupture does not represent an estimate of the source length. Such data must not be used in establishing the magnitude-length relation.

The expected rupture area can be estimated with about the same accuracy as the expected fault length. It is proposed that magnitude be estimated from *rupture area* data instead of from surface rupture-length data alone, for two reasons. There exists a theoretical relation between magnitude and area, and better results are obtained because a wide fault will radiate more energy than a narrow fault of the same length. The equation relating magnitude to rupture area is $M = \log A + 4.15$, with A in square kilometres. This relation is valid for $M > 5.6$.

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