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SCALING LAWS FOR LARGE EARTHQUAKES: CONSEQUENCES FOR PHYSICAL MODELS

BY CHRISTOPHER H. SCHOLZ

ABSTRACT

It is observed that the mean slip in large earthquakes is linearly proportional to fault length and does not correlate with fault width. This observation is interpreted in the light of the two possible classes of models for large earthquakes: *W* models, in which stress drop and slip are determined by fault width, and *L* models, in which these parameters are fundamentally determined by fault length. In the *W* model interpretation, stress drop systematically increases with *L*/*W*, the aspect ratio, and, as a consequence, seismic moment. The correlation of slip with length means that the rupture length is determined by the dynamic stress drop. This conflicts with the observation that the length of large earthquakes is often controlled by adjacent rupture zones of previous earthquakes or by tectonic obstacles. It also conflicts with the observations for small earthquakes that stress drop is nearly constant and does not correlate with source radius over a broad range. In the *L* model interpretation, the correlation between slip and length means that stress drop is constant, namely about 7.5, 12, and 60 bars for interplate strike-slip, thrust, and Japanese intraplate earthquakes, respectively. *L* models require that the fault be mechanically unconstrained at the base. *W* models predict that mean particle velocity increases with fault length, but rise time is constant. *L* models predict the opposite.

INTRODUCTION

A central problem in earthquake seismology has been to find scaling laws that relate the static parameters such as slip and static stress drop to the dimensions of the rupture and to understand these relationships in terms of the dynamics parameters, the most fundamental of which are rupture velocity and dynamic stress drop.

In doing so, it is essential to distinguish between small earthquakes and large earthquakes. Tectonic earthquakes nucleate and are bounded within a region of the earth between the surface and a depth h_0, the seismogenic layer. The seismogenic depth, h_0, depends on the tectonic environment but in a given region the maximum width of an earthquake occurring on a fault of dip δ is $W_0 = h_0/\sin\delta$. We will define a small earthquake as one with a source radius $r \leqq W_0/2$ and a large earthquake as one in which $r > W_0/2$. Thus a small earthquake can be represented as a circular source in an elastic medium, whereas a large earthquake is more suitably treated as a rectangular rupture with one edge at the free surface.

It has been repeatedly demonstrated (e.g., Aki, 1972; Thatcher and Hanks, 1973; Hanks, 1977) that the stress drops of small earthquakes are nearly constant and independent of source dimensions over a broad range of source radius. This result, when interpreted with dynamic models of finite circular ruptures (Madariaga, 1976; Archuleta, 1976; Das, 1980), simply means that the dynamic stress drop is constant.

Since this parameter is related to a material property of the fault, this result is one of great physical significance.

If dynamic stress drop is also constant for large earthquakes, the dynamic models of rectangular faulting in an elastic medium (Day, 1979; Archuleta and Day, 1980; Das, 1981) would predict that the mean slip is a linear function of fault width. In the next section we will show that this prediction is not borne out by the observations. What is observed instead is that slip correlates linearly with fault length. The principal point of this paper is to discuss the consequences of that observation on the physics of large earthquakes.

The Observations

We wish to discover whether or not there is any correlation between the mean slip, $\bar{u}$, in a large earthquake and the length, L, or width, W, of the rupture. The data set we will use is that assembled by Sykes and Quittmeyer (1981) and is given in Table 1. In the table, the observed parameters are L, W, and either $\bar{u}$ or M_0. The stress drop, $\Delta\sigma$, has been calculated from a model and will be discussed in a later section.

The principal observation that can be made about these data is shown in Figures 1 and 2, where we plot, in linear coordinates, $\bar{u}$ versus L for large interplate strike-slip and thrust earthquakes. A strong correlation is evident, indicating simple proportionality of the form

$$\bar{u} = \alpha L \tag{1}$$

where the proportionality constant $\alpha \approx 1.25 \times 10^{-5}$ for strike-slip and 2×10^{-5} for thrust earthquakes. A correlation between u and L, has been suggested before (Bonilla and Buchanan, 1970; Slemmons, 1977) and Sykes and Quittmeyer (1981) have argued that it is linear for strike-slip earthquakes. This linear correlation applies equally to intraplate earthquakes: large intraplate earthquakes in Japan exhibit this property, with a larger proportionality constant, $\alpha = 1 \times 10^{-4}$ (Matsuda *et al.*, 1980).

The intraplate Japanese earthquakes and the strike-slip earthquakes all have essentially the same width, $W = 15$ km, therefore there appears to be no relationship between $\bar{u}$ and W. In the case of the large subduction zone thrust events, there appears to be a weak correlation between W and L in the data set, such that $W \propto L^{1/2}$. Therefore we should expect to find a correlation between $\bar{u}$ and W^2 for these events. However, since that correlation is not observed for the strike-slip or intraplate Japanese events, we will restrict ourselves to discussing only the correlation between L and $\bar{u}$ and not address the question as to whether or not there is a causal basis for the correlation between L and W for the thrust events.

The observations are very clear and simple: slip increases linearly with fault length and is insensitive to fault width. For interplate events, the proportionality constant is greater for thrust than for strike-slip events, and it is higher still for intraplate events. The property is illustrated in Figure 3, where we show the measured slip as a function of distance along the fault for two of the strike-slip earthquakes in our data set. These two earthquakes have essentially the same width, but different lengths, and the plots are at the same scale. In the next section, we will discuss the consequences of this property on the physical mechanism of earthquakes.

TABLE 1

PARAMETERS OF LARGE INTERPLATE EARTHQUAKES
[AVERAGED FROM SYKES AND QUITTMEYER (1981)]

No.	Date	Location	M_0 (10^{27} dyne-cm)	L (km)	W (km)	L/W	$\bar{u}$ (cm)	$\Delta\sigma$ (bars)
		Strike-Slip Earthquakes						
1	10 Jul. 1958	SE Alaska	4.3	350	12	29	325	26
2	9 Jan. 1857	S. California	7	380	12	32	465	36
3	18 Apr. 1906	San Francisco	4	450	10	45	450	44
4	19 May 1940	Imperial Valley, Calif.	0.23	60	10	6	125	13
5	27 Jun. 1966	Parkfield, Calif.	0.03	37	10	4	30	4
6	9 Apr. 1968	Borrego Mountain, Calif.	0.08	37	12	3	25	3
7	15 Oct. 1979	Imperial Valley, Calif.	0.03	30	10	3	30	4
8	4 Feb. 1976	Guatemala	2.6	270	15	18	150	9
9	16 Oct. 1974	Gibbs Fault Zone	0.45	75	12	6	170	14
10	26 Dec. 1939	Ercincan, Turkey	4.5	350	15	23	285	18
11	20 Dec. 1942	Erbaa Niksar, Turkey	0.35	70	15	5	112	8
12	1 Feb. 1944	Gerede-Bolu, Turkey	2.4	190	15	13	275	18
13	18 Mar. 1953	Gönen-Yenice, Turkey	0.73	58	15	4	280	21
14	22 Jul. 1967	Mudurnu, Turkey	0.36	80	15	5	100	7
		Thrust Earthquakes						
15	6 Nov. 1958	Etorofu, Kuriles	44	150	70	2.1	840	37
16	13 Oct. 1963	Eruppu, Kuriles	67	275	110	2.5	445	12
17	16 May 1968	Tokachi-oki, Japan	28	150	105	1.4	355	10
18	11 Aug. 1969	Shikotan, Kuriles	22	230	105	2.2	180	5
19	17 Jun. 1973	Nemuro-oki, Japan	6.7	90	105	0.86	140	5
20	4 Nov. 1952	Kamchatka	350	450	175	2.6	890	14
21	28 Mar. 1964	Prince Wm. Sound, Alaska	820	750	180	4.2	1215	18
22	4 Feb. 1965	Rat Island, Aleutians	125	650	80	8.1	480	10
23	10 Jan. 1973	Colima, Mexico	3	85	65	1.3	110	5
24	29 Nov. 1978	Oaxaca, Mexico	3	80	70	1.1	110	5
25	22 May 1960	S. Chile	2000	1000	210	4.8	1900	21
26	17 Oct. 1966	C. Peru	20	80	140	0.6	360	12

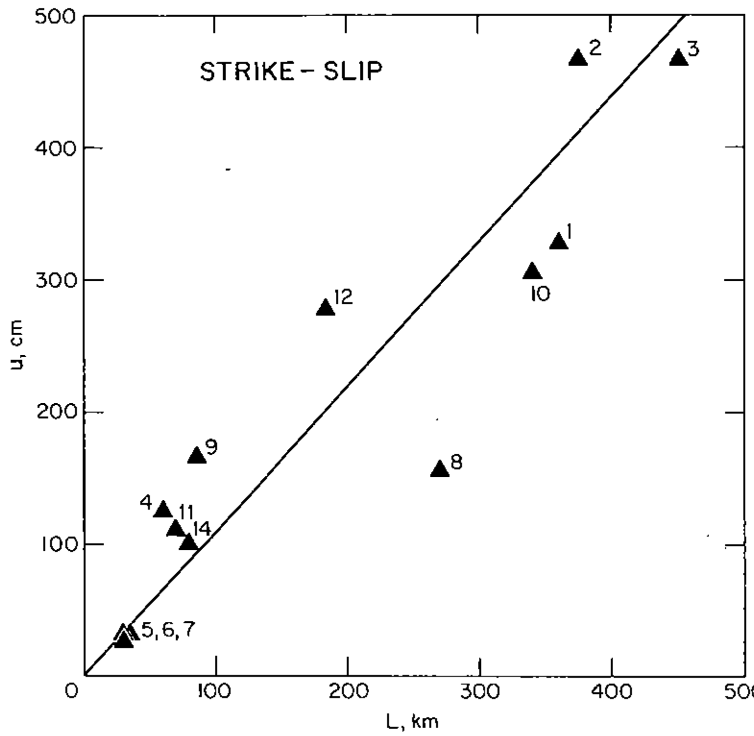

FIG. 1. A plot of mean slip, u versus fault length for the strike-slip events. The line drawn through the data has a slope of 1.25×10^{-5}. Numbers are references in Table 1.

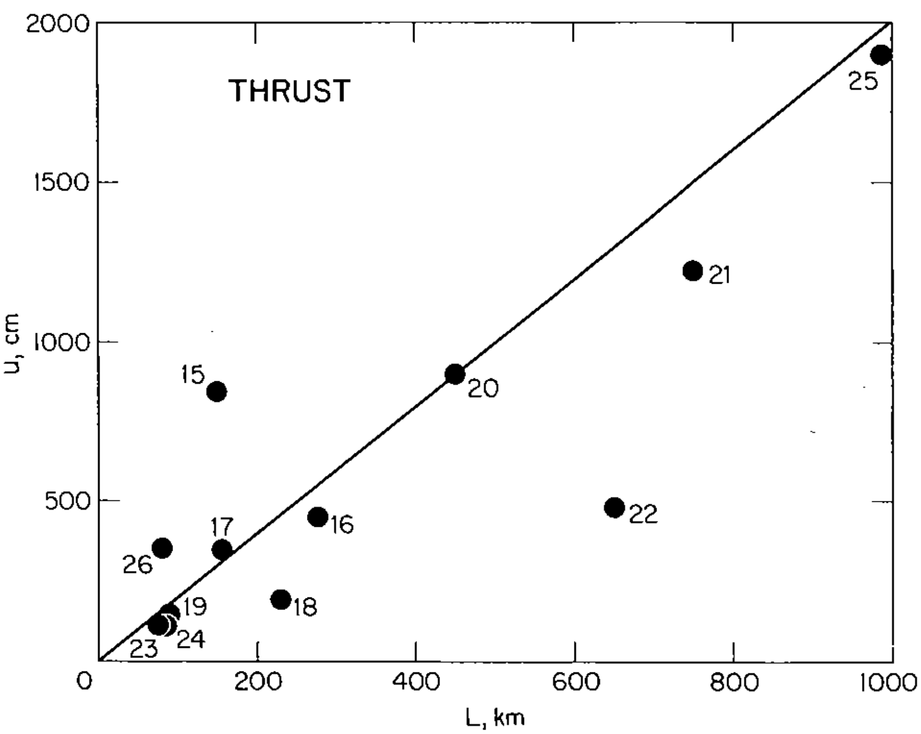

FIG. 2. The same as Figure 1, for the thrust events. The slope of the line is 2×10^{-5}.

The Physical Consequences

In this section we will discuss some very general physical consequences of $\bar{u}$ scaling with L. We will do so in terms of the two possible classes of models of rupture in a rectangular region. The first of these, which we will call the class of W models, contains the models in which the slip and stress drop are determined by the small dimension of the rupture, i.e., by W. The second class is that of L models, in which $\Delta\sigma$ and $\bar{u}$ are determined by L. Although specific earthquake models may vary greatly in detail, any three-dimensional model of rupture on a rectangular fault must be either a W model or an L model. More specifically, any purely elastic model will be a W model, hence all three-dimensional rupture models previously described in the literature, whether they be kinematic or dynamic, are W models.

Interpretation in terms of W models. In a W model, $\Delta\sigma$ will be related to fault width by

$$\Delta\sigma = C\mu \frac{\bar{u}}{W} \tag{2}$$

where C is a geometrical term that should be nearly constant for $L > 2W$ (see, e.g., Sykes and Quittmeyer, 1981). Seismic moment is therefore given by

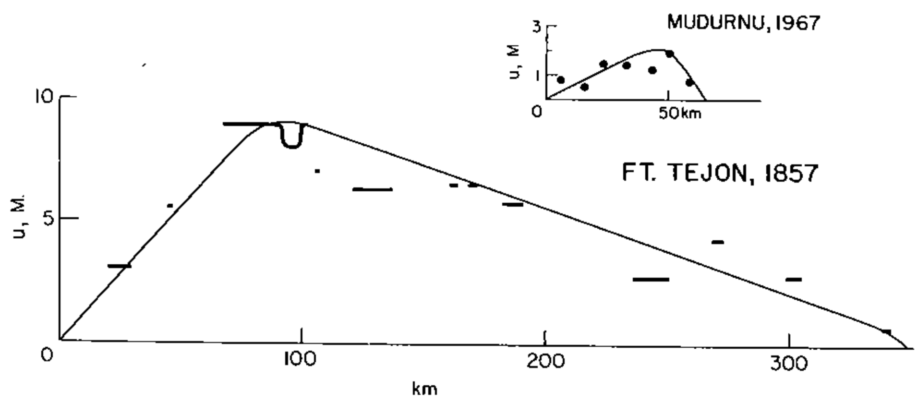

FIG. 3. Surface slip as a function of distance along the fault plane for two representative strike-slip earthquakes of similar width but different length. Data for the Mudurnu earthquake are from Ambraseys (1970) and for the Ft. Tejon earthquake from Sieh (1978).

$$M_0 = \frac{\Delta\sigma}{C} L W^2. \tag{3}$$

In Figure 4 we show a plot of log M_0 versus log LW^2. The data for each type of earthquake define a line, but with a slope of less than one, indicating that with this class of models we must interpret these data as meaning that stress drop systematically increases with moment.

This arises because, combining (1) with (2) we obtain

$$\Delta\sigma = C\mu\alpha \frac{L}{W}. \tag{4}$$

Since the widths of large earthquakes are restricted by W_0, they tend to grow principally in the L dimension, and L/W, and hence $\Delta\sigma$, tends to systematically increase with M_0. The offset between the strike-slip and thrust data in Figure 4 is due to the fact that W_0 is much smaller for strike-slip than for subduction zone thrust earthquakes, so that a strike-slip earthquake of the same moment as a thrust event must have a much greater aspect ratio, and hence $\Delta\sigma$.

Sykes and Quittmeyer (1981) calculated stress drops for these events using a W model (specifically, a dislocation model). These are listed in Table 1 and plotted against L/W in Figures 5 and 6. Without arguing whether or not these represent "true" stress drops, they are at least internally consistent with a W model and illustrate the type of scaling that the observations demand when interpreted in the light of a W model.

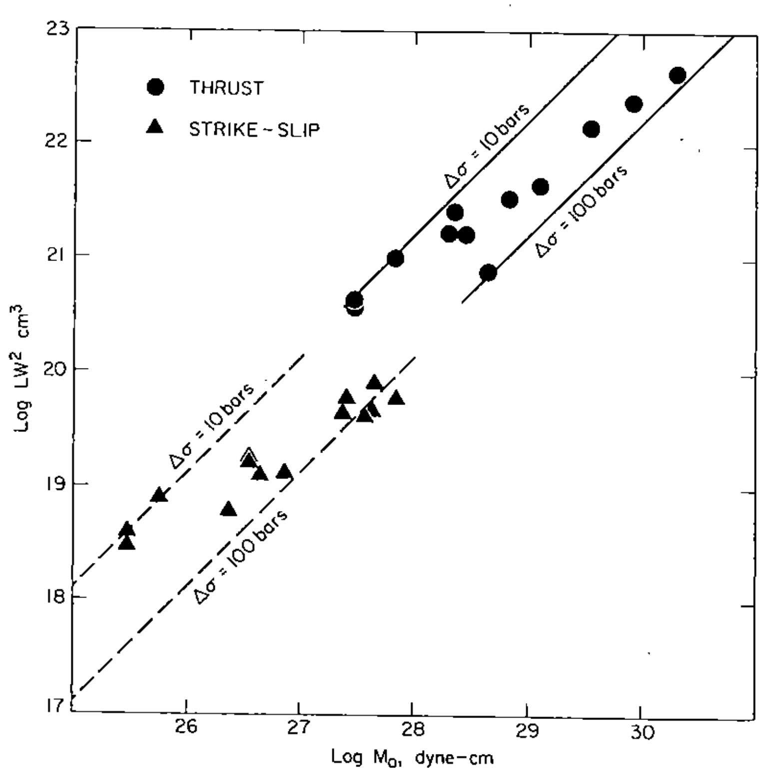

FIG. 4. Plot of log LW^2 versus log M_0 for the large interplate earthquakes from the data set of Sykes and Quittmeyer (1981). The lines of slope 1 are constant stress drop lines, assuming $C = 0.6$ for the thrust events, and 0.3 for the strike-slip events.

Because of our basic observation that $\bar{u} = \alpha L$, we expect that M_0 is proportional to $L^2 W$. Since

$$L^2 W = A^{3/2}\left(\frac{L}{W}\right)^{1/2} \tag{5}$$

and since the aspect ratio varies only by a factor of about 20 in the data set, it is not surprising that Aki (1972) and Kanamori and Anderson (1975) found a good correlation between M_0 and $A^{3/2}$ for large earthquakes (since the observations were plotted in logarithmic coordinates). However, in the context of their models (W models), their interpretation of this as meaning that stress drop is constant is incorrect since L/W is not constant.

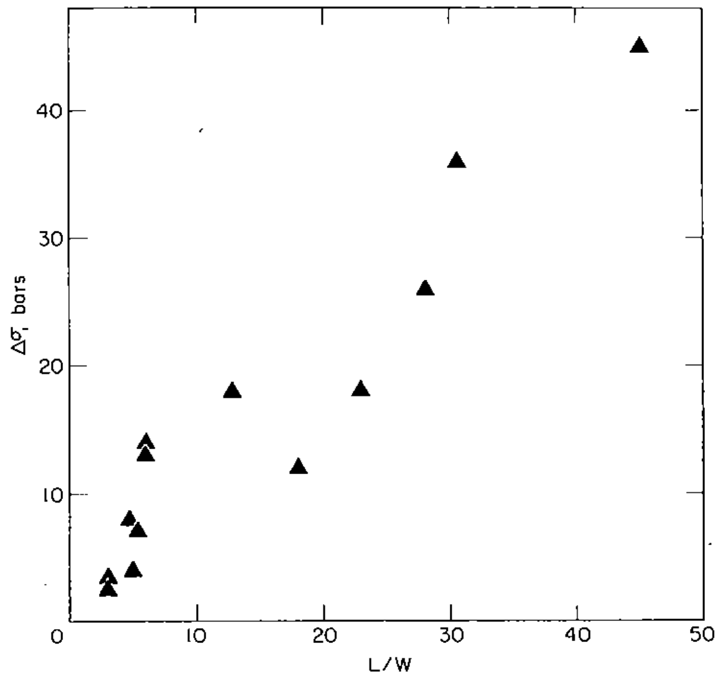

Fig. 5. Stress drop plotted versus aspect ratio for the strike-slip earthquakes.

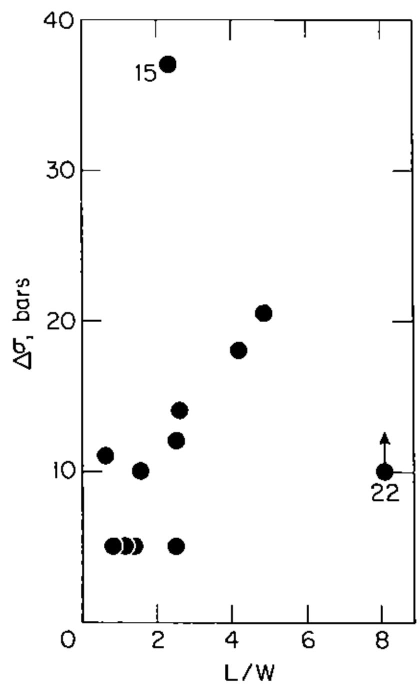

Fig. 6. Stress drop versus aspect ratio for the thrust earthquakes. Event 22 is an oblique slip event for which stress drop was calculated based only on the dip slip component and is hence underestimated. Event 15 is an anomalously deep event in the Kuriles (Sykes and Quittmeyer, 1981).

Interpretation in terms of L models. The essential ingredient of a W model is that the fault is pinned at the base. This is why all elastic models are W models. The base of a large earthquake, however, is at the base of the seismogenic layer, below which the deformation must be inelastic.

In an L model, the fault must be unrestrained at both top and bottom. One way this might occur is shown in Figure 7. Figure 7A shows the conventional W model of an earthquake: slip occurs above some depth W_0 and is constrained to be zero at W_0. In Figure 7B, we illustrate another model of the situation just prior to a large earthquake. Slip has occurred aseismically below W_0, completely through the

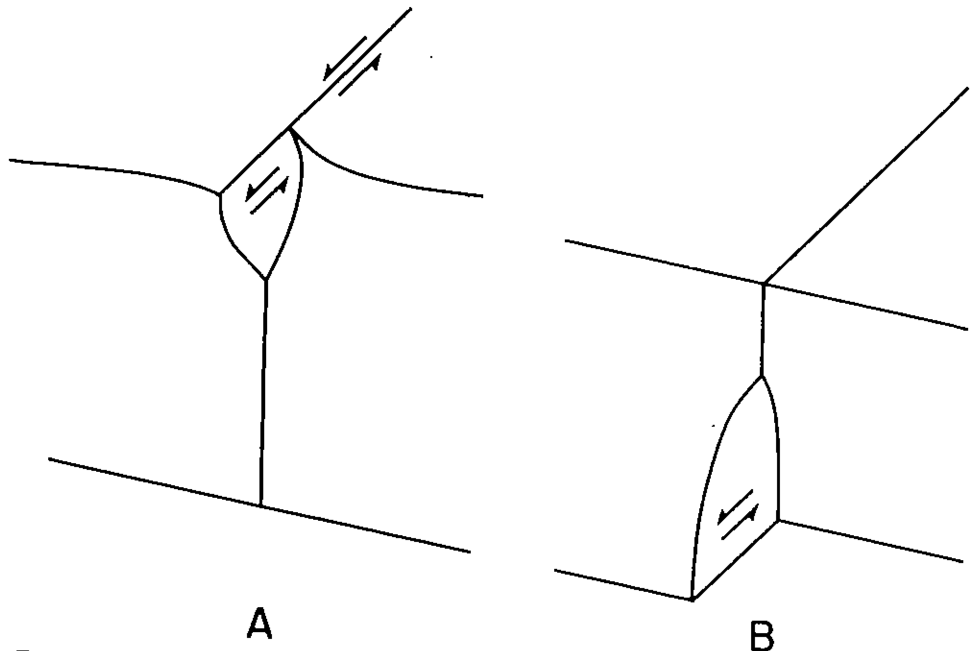

FIG. 7. In 7A is shown schematically a W model of a large earthquake, in which slip is constrained to be zero at the base of the fault. In 7B, we show a possible mechanism that may result in an L model. The figure shows the situation just prior to a large earthquake. Aseismic slip has occurred beneath the seismogenic layer. If this preslip is larger than the slip in the earthquake, the earthquake may not be constrained at the base.

lithosphere. With this model, the large earthquake can be viewed as the dynamic extension to the surface of a fault with an effectively infinite width. We may suppose that in this case the base of the fault does not restrain the slip so long as the slip in the earthquake is less than the preslip at depth. If this latter statement is true, an L model will result.

Any such discussion of how L models may physically operate is conjectural at present. We simply offer it as a hypothetical case for the purpose of the present discussion. We calculated the final slip for a quasi-dynamic L model, the details of which may be found in the Appendix. The principal result is that

$$\bar{u} = \left(\frac{1}{2}\frac{\Delta\sigma_d}{\mu}\right) L \tag{6}$$

where $\Delta\sigma_d$ is the dynamic stress drop. So according to this type of model, the observation that slip scales with L must be interpreted as meaning that stress drop is constant. In Figure 8 we show a plot of log M_0 versus log L^2W. The line drawn through the data is the prediction of the L model for $\Delta\sigma_d = 10$ bars. Equating (1) with (6) we find that $\Delta\sigma_d \approx 7.5$, 12, and 60 bars for the interplate strike-slip, thrust, and Japanese intraplate earthquakes, respectively.

Some Further Problems with the W Model Interpretation

It was pointed out in the previous section that if we interpret the observations in terms of a W model, then we must conclude that the stress drop increases with the

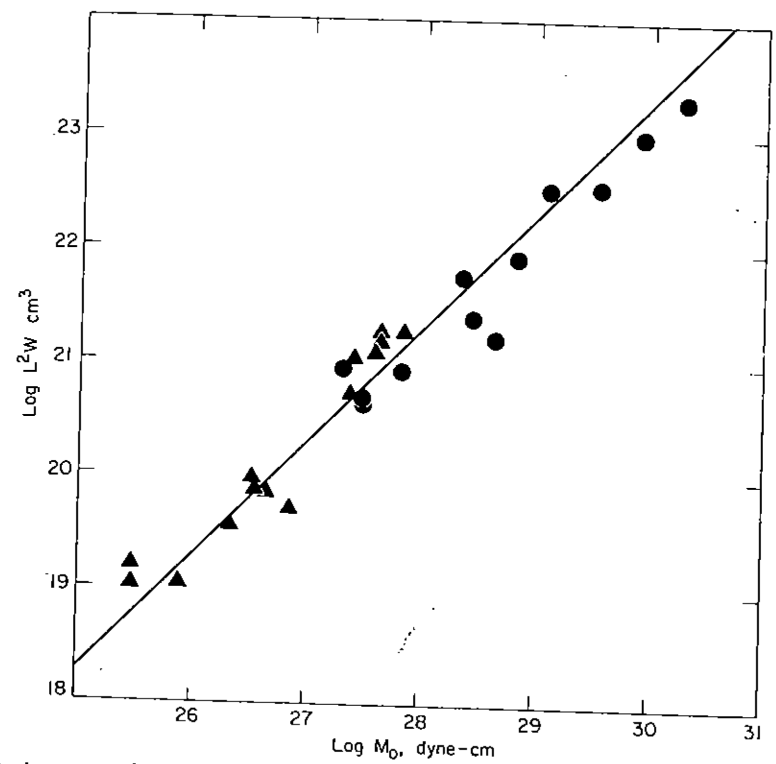

Fig. 8. A plot of log L^2W versus log M_0. The line drawn through the data has a slope of 1, and is the prediction of the L model for a constant dynamic stress drop of 10 bars.

aspect ratio of the earthquake. The stress drop discussed was the final (or static) stress drop calculated from (2). The results of dynamic models show that the static stress drop is closely related to the dynamic stress drop, $\Delta\sigma_d$, i.e., the difference between the applied tectonic stress and the dynamic frictional strength of the fault, and hence is essentially a material property (Madariaga, 1976; Day, 1979; Archuleta and Day, 1980; Das, 1981). The conclusion that this varies with aspect ratio poses additional problems with the W model interpretation.

Since fault width is constant for the strike-slip events, and varies much less than length for the thrust events, we can explain the correlation between $\Delta\sigma$ and L/W if there is a linear correlation between $\Delta\sigma_d$ and L. The only way this could arise without violating causality is if the dynamic stress drop determines the length. This is not physically unreasonable, since the dynamic stress drop determines the stress intensity factor at the edge of the rupture, which is important in rupture growth. We can offer no explanation, however, as to why the relationship should be linear.

This interpretation, however, leads to some conflicts with other observations. The first is that it conflicts with the principal observations that led to the concept of seismic gaps: that the lengths of large earthquakes are often controlled by the rupture zones of previous earthquakes or by structural features transverse to the fault zone. Of course, one could soften the original assumption to: $\Delta\sigma_d$ determines the length unless the rupture encounters a rupture zone of a previous earthquake or a transverse feature. The rejoinder is that if the latter were as common as is thought, it would have the effect of destroying the correlation between $\bar{u}$ and L that is observed.

It is worth giving a specific example. If we compare the 1966 Parkfield earthquake (L = 30, km, $\bar{u}$ = 30 cm, W = 15 km) and the 1906 San Francisco earthquake (L = 450 km, $\bar{u}$ = 450 cm, W = 10 km), we need to explain the difference in u by a difference in $\Delta\sigma_d$ of about a factor of 15. Since the correlation between $\bar{u}$ and L is also good in these examples, we also need to argue that $\Delta\sigma_d$ determined L in these cases. On the other hand, it can be argued that the length of the 1966 earthquake was determined by the length of the gap between the rupture zone of the 1857 earthquake (or the fault offset near Cholame) and the southern end of the creeping section of the San Andreas fault. Similarly, the 1906 earthquake filled the gap between the northern end of the creeping section at San Juan Bautista and the end of the fault at Cape Mendocino. If our argument that $\Delta\sigma_d$ determines L is true, then these latter observations are coincidences. Almost identical arguments can be made for many of the other earthquakes in our data set.

The second problem is that our interpretation that for large earthquakes $\Delta\sigma_d$ determines the rupture length directly contradicts the observations for small earthquakes. Although stress drop can be interpreted to increase with source radius over a limited range in some data sets (Aki, 1980), it shows no obvious variation with source radius over a very broad range (Hanks, 1977). We can offer no reasonable explanation for why large earthquakes should behave differently than small earthquakes in this important respect.

The third point is less a conflict than a surprising consequence of this interpretation. The Hoei earthquake of 1707 ruptured about 500 km of the Nankai trough in Japan (Ando, 1975; Shimazaki and Nakata, 1980). The same plate boundary was ruptured twice subsequently, in two sets of delayed multiple events, and Ansei I and II events of 1854, and the Tonankai and Nankaido events of 1944 and 1946. In support of a time-predictable model of earthquake recurrence, Shimazaki and Nakata argued that the greater recurrence time between the first two sequences (147 yr) and the second (91 yr) is because the slip (and stress drop) were greater in 1707 than in either 1854 or 1946, the greater uplift at Muroto Point in 1707 (1.8 m) than in 1856 (1.2 m) or 1946 (1.15 m) being the evidence. The reason why this should happen is readily explained by the correlation between $\bar{u}$ and L. Thus, the ratio of fault length of the Hoei and Ansei II earthquakes, 500 km/300 km = 1.7 can explain the ratio of uplift at Muroto Point, 1.8/1.2 = 1.5 and recurrence time, 147/91 = 1.6. However, if this is interpreted as being due to a difference in stress drop, then one has to argue that a significant change in stress drop (50 per cent) can occur on the same fault zone between successive earthquakes. One could argue that this could occur because the slip in one earthquake might change the relative position of asperities on the fault. However, since the slip in an earthquake is about 10^{-5} L, this would mean that the gross frictional properties of the fault are controlled by asperities of dimensions on the order of 10^{-5} or less of the rupture dimensions. Since there will be a very large number of such small features, the average change between successive earthquakes would more likely be expected to be negligible.

None of the above-mentioned problems exist for the L model interpretation, since the observations would then be interpreted as meaning that stress drop is essentially constant for each of the different classes of earthquakes discussed.

Differences in the Dynamics of L and W Models

The difference between the dynamic behavior of W and L models arises because of the difference in what controls the healing process. In a W model, healing originates from the base of the fault, so that the part of the fault that is slipping at any one time is a patch of length $L \approx W$ (see, e.g., Day, 1979; Archuleta and Day, 1980; Das, 1981). In an L model, where the base of the fault is unconstrained, healing does not occur at that edge. In that type of model, the entire fault continues to slip until the final length is reached, since healing is initiated only from the ends of the fault (see Appendix).

Insofar as the L and W models represent opposite extremes concerning the mechanism of large earthquakes, it is useful to discuss the contrasting way in which they scale. For earthquakes in which $L < 2W$, the models are indistinguishable in their gross manifestations. In Figure 9 we schematically show a comparison between

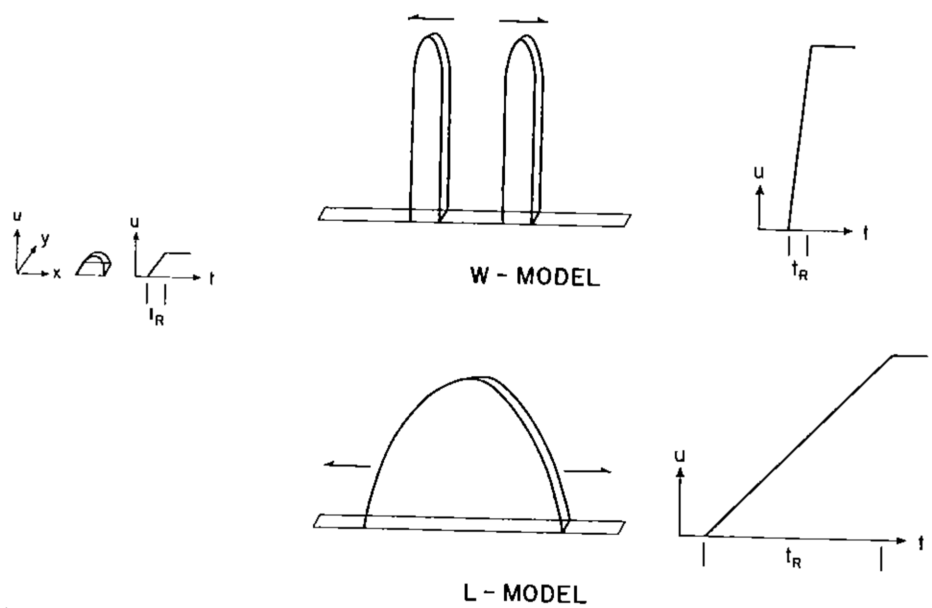

Fig. 9. A schematic diagram to illustrate the contrasting way in which a model in which width determines the slip (W model) scales with length as compared to a length-dependent model (L model).

an earthquake of dimensions about $L = 2W$ and one of the same width but about 15 times longer. Specifically, this might be a comparison of the 1966 Parkfield earthquake, say, and the 1906 San Francisco earthquake.

On the *left* of the figure we show a snapshot of slip on the fault during the smaller earthquake. We only show the part that is actually slipping during the snapshot. We also show the time history of slip at some representative point. For simplicity, it is simply shown as a ramp with a rise time, t_R. On the *right* is shown the predictions of the two models for the longer earthquake.

In a bilateral case, as shown, the W model predicts that the slipping portion of the fault splits into two patches of length $\approx W$ that propagate away from each other at twice the rupture velocity as they sweep over the fault surface. Since the rise time, $t_R \approx W/2\beta$, remains the same but the slip is 15 times greater, the stress drop, and hence particle velocity, must be 15 times greater.

In the L model, the rupture sweeps out over the fault as an expanding patch, with slip continuing within its boundaries until after the final dimensions are reached. In that model, the stress drop and particle velocities are the same as in the smaller event, but the rise time, $t_R \approx L/2\beta$ is much longer.

In terms of predicting the strong ground motions for a 1906 size earthquake, say, from observed ground motions for a 1966 size earthquake, the difference between the W and L models is critical. The W model would predict that the average particle velocities would be much higher and the duration would be about the same. The L model would predict nearly the opposite.

It is interesting to recall an account of the shaking in 1906, made by Professor T. J. J. See at Mare Island (Reid, 1910, p. 4). This location is 40 km from the fault and is just opposite the most likely position of the epicenter (Boore, 1977) and so the rupture must have principally been propagating away (in both directions) from this observer. The account is: ". . .then the tremors grew steadily, but somewhat slowly, becoming gradually stronger and stronger, until the powerful shocks began, which became so violent as to excite alarm. Their duration was unexpectedly long, about 40 seconds. . .and the subsiding tremors then began." Although one cannot prove anything with such an account, it is certainly consistent with the notion that slip may have continued locally for 40 sec or so, which would be more consistent with an L than with W model.

In conclusion, it appears that at present the most elementary feature of the mechanism of large earthquakes, the physical dimension that controls the process, has not been identified. If it is the width, then serious problems arise in understanding why stress drop should correlate with length. If, on the other hand, it is the length, other serious problems arise in understanding the physical process at the base of the fault that allows this to be the case.

Appendix: A Quasi-Dynamic L Model

We take that the slip that occurs prior to the initiation of healing within a rupture propagating at a constant rupture velocity v is, from Kostrov (1964),

$$u(x, y, t) = \dot{u}_0\left(t^2 - \frac{(x^2 + y^2)}{v^2}\right)^{1/2} \quad \text{for} \quad \frac{(x^2 + y^2)^{1/2}}{v} \leqq t \leqq t_h \tag{A1}$$

where x and y are measured relative to the point of rupture initiation and t_h is the time of initiation of healing. The asymptotic particle velocity, $\dot{u}_0$, which scales the slip is

$$\dot{u}_0 = K\frac{\Delta\sigma_d}{\mu}\beta \tag{A2}$$

where K is a function of rupture velocity.

The results of dynamic models verify equation (A1) and also show that initiation of healing propagates inward from the final rupture perimeter at a velocity close to a wave velocity of the medium (Madariaga, 1976; Day, 1979; Das, 1980, 1981). Those results also show that the slip that occurs after healing initiates is small compared to the total slip, so that we can closely approximate the final slip by assuming that healing is abrupt.

Hence the final slip is

$$u(x, y) = \dot{u}_0\left(t_h{}^2 - \frac{(x^2 + y^2)}{v^2}\right)^{1/2}. \tag{A3}$$

If we assume that healing originates from all edges of a rectangular fault in calculating (A3), we would generate a W model (see Day, 1979, pp. 23–26). In order to generate an L model, we assume that no healing originates at the base of the fault. In either case, no healing originates at the top, since it is at the free surface.

In the particular case studied here, we assumed that $v = 0.9\beta$, for which $K = 0.81$ (Dahlen, 1964), and that healing propagates at $\sqrt{3}\beta$. An example is shown in Figure A1 for a bilateral case with an aspect ratio of 4. The scaling resulting from this modeling was reported as equation (6) in the main body of the text.

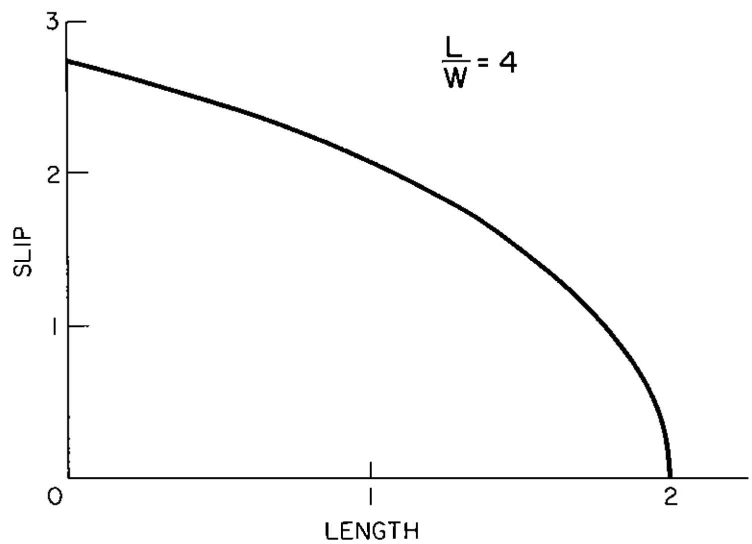

FIG. A1. Dimensionless slip, u' versus length, L', at the free surface from the center to the end of the fault. The model is a quasi-dynamic L model that simulates a dynamic model with boundary conditions similar to those shown in Figure 7B, as described in the text. The normalization relations are $u = \Delta\sigma_d/\mu\ Wu'$ and $L = WL'$. The case shown is bilateral with aspect ratio 4.

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