INTRODUCTION

Moment-magnitude relations have played an important role in earthquake mechanism studies since seismic source parameter determinations became routine in the early 1970's. These empirically defined relations have always been written as a linear relation between log $M_0$ and $M_L$:

$$\log M_0 = c M_L + d \tag{1}$$

where $M_0$ is seismic moment and $M_L$, in general, can be any magnitude but in practice is usually $M_I$ (local magnitude) or $M_s$ (surface-wave magnitude).

At the present time, the significance of such relations are twofold. First, any definition of a moment magnitude scale (that is, some moment magnitude $M$ determined from log $M_0$) would, ideally, have coefficients (of the inverse relation) not too different from those in (1) for whatever $M$ is actually in use for the region of interest. Happily enough, this can be arranged. Hanks and Kanamori [1979] noted that the moment magnitude

$$M = \frac{1}{3} \log M_0 - 10.7 \tag{2}$$

is identical (in inverse form) to the moment-magnitude relations of Thatcher and Hanks [1973] for southern California earthquakes ($3 \leq M_L \leq 7$), of Purcaru and Berckhemer [1978] for a set of global earthquakes ($5 \leq M_L \leq 7$), and of that implicit in the work of Kanamori [1977] for great earthquakes ($M_L \geq 7$).

Second, recent studies of high-frequency strong ground motion as finite-duration, band-limited, white Gaussian noise in acceleration, we can estimate $M_0$ as a function of $M_L$ alone, by fixing the $a_{\text{rms}}$ stress drop at 100 bars and $f_{\text{max}}$ at 15 Hz. These model calculations fit available California moment-magnitude data for $0 \leq M_L \leq 7$, $10^{17} \leq M_0 \leq 10^{28}$ dyne cm with striking accuracy. This range in source strength is entire: earthquakes with $M_0 \geq 10^{28}$ dyne cm are unlikely to occur in California, and earthquakes with $M_L < 0$ cannot be recorded in California, at least under ordinary conditions of recording earthquakes at ordinary hypocentral depths. More fundamentally, the remarkably good fit of model to data implies that the $a_{\text{rms}}$ stress drop of 100 bars (to a factor of 2 or so) is a stable and pervasive feature of all ($M_L \geq 7$) California earthquakes whose spectral corner frequency lies in the "visible" bandwidth, $f_0 \leq f_{\text{max}}$.

This paper is not subject to U.S. copyright. Published in 1984 by the American Geophysical Union.

Paper number 4B0432.
AR technique [Brune et al., 1963] calibrated by the 13 earthquakes above,
\[
\log M_o = 1.7M_L + 15.1 \quad 3 \leq M_L \leq 6
\]

We have not included these relations in Table 1, since neither of them fits neatly as “central California” or “southern California,” even though we shall shortly conclude that this distinction is immaterial.

There are two noteworthy features of the central California relations in Table 1. The first of these is the small number of \( M_L \geq 5.0 \) earthquakes that contribute to the data set, either in part or in sum. The situation is even worse than indicated in Table 1, since three \( M_L \geq 5.0 \) earthquakes have been used more than once. Only nine different \( M_L \geq 5.0 \) earthquakes are involved in all of these studies, whereas Thatcher and Hanks [1973] included 43 \( M_L \geq 5.0 \) southern California earthquakes. As we shall see shortly, this is the reason for the differences between the central and southern California relations.

The second feature is that none of the central California studies of Table 1 qualify as a systematic, regional study of the \( M_o - M_L \) relationships of central California earthquakes. Four of these five studies are for very localized source regions, and the study of Bolt and Herraiiz [1983] is, in effect, one as well, since 10 of the 16 events are taken from Johnson and McEvilly [1974]. This feature, however, is of no real consequence, since the very similar relations for four different relations certainly indicates a regional norm—for \( M_L < 5.0 \).

The recent study of Bakun [1984] does qualify as a systematic, regional study, at least for \( M_L < 5.0 \). He determined seismic moments for 118 events in five separate source regions: Parkfield, San Juan Bautista, Sargent fault, Coyote Lake, and Livermore Valley. Even though his study, as well, works with only five \( M_L \geq 5.0 \) events (and 14 \( M_L \geq 4.0 \) events), Bakun [1984] detected a clear change in the \( c \) value at \( M_L \approx 3 \). In summary form, with ranges of validity, Bakun [1984] finds
\[
\log M_o = 1.2M_L + 17 \quad \frac{1}{2} \leq M_L \leq \frac{3}{2} \quad (3a)
\]
and
\[
\log M_o = 1.5M_L + 16 \quad 3 \leq M_L \leq 6 \frac{1}{2} \quad (3b)
\]

Bakun’s study underscores the necessity of working with a wide range of \( M_L \). Thatcher and Hanks [1973] did not detect a change in the \( c \) value at \( M_L \approx 3 \), but only 12 of their 138 earthquakes had \( M_L < 3 \) and only four with \( M_L < 3 \). The studies in central California suffer from the opposite problem: the relatively small number of \( M_L \geq 5.0 \) earthquakes analyzed to date.

Table 2 gives \( M_o - M_L \) data for 18 \( M_L \geq 5.0 \) earthquakes in central California, for which we know of a quantitative estimate of \( M_o \). These include all of the \( M_L \geq 5.0 \) earthquakes used in the central California studies of Table 1, as well as nine additional events culled from the literature. These are plotted in Figure 1, together with the \( M_o - M_L \) relations of Table 1 for the given ranges of validity. With the exception of the relation of Fletcher et al. [1984] for \( 4.3 \leq M_L \leq 5.7 \), none of the central California relations is a close approximation to the data of Table 2 above \( M_L \approx 5 \), stated ranges of validity notwithstanding. The “southern California” relation of Thatcher and Hanks [1973], however, is a close approximation to the “central California” earthquakes with \( M_L \geq 5 \), excepting the two largest. Clearly, as Bakun [1984] has found, the log \( M_o - M_L \) data have positive curvature (see also Figure 2).

Straight-line fits of equation (1) to various ranges of the data will result in \( c \) values increasing with \( M_L \). As straight-line approximations, we can concur with the findings of Bakun [1984], equation (3) above with their given ranges of validity. Above \( M_L \approx 6 \), however, the \( c \) value must be even larger. In the next section, we describe model calculations that recover in detail both the continuous curvature of the log \( M_o - M_L \) observations and their absolute values, allowing us to forego altogether straight-line fits to log \( M_o - M_L \) data, across limited magnitude ranges chosen more or less arbitrarily.

Before proceeding to these calculations, however, several brief statements on the \( M_o \) estimates in Table 2 and Figure 1 are appropriate. First, we have preferred the teleseismic estimates of \( M_o \) for the Mammoth Lakes earthquakes (Table 2) over the locally determined values of Uhrhammer and Ferguson [1980] and Archuleta et al. [1982]. The teleseismic estimates are typically 2 to 4 times larger than the local determinations, the latter having strongly conditioned the \( M_o - M_L \) relations of Archuleta et al. [1982] and Bolt and Herraiiz [1983]. Second, the Eureka earthquake is hardly a “central California” earthquake, but it is our point of view, on the basis of the “southern California” fit [Thatcher and Hanks, 1973] to “central California” earthquakes (Figure 1), that

<table>
<thead>
<tr>
<th>Study</th>
<th>Relation</th>
<th>Range of Validity</th>
<th>Number of Events</th>
<th>Number of Events</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Central California</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Johnson and McEvilly [1974]</td>
<td>( \log M_o = (1.16 \pm 0.06)M_L + 17.60 \pm 0.28 )</td>
<td>( 2.4 \leq M_L \leq 5.1 )</td>
<td>13</td>
<td>1</td>
</tr>
<tr>
<td>Bakun and Lindh [1977]</td>
<td>( \log M_o = (1.21 \pm 0.03)M_L + 17.02 \pm 0.07 )</td>
<td>( 10^{13} \leq M_o \leq 10^{13} )</td>
<td>28</td>
<td>2</td>
</tr>
<tr>
<td>Archuleta et al. [1982]</td>
<td>( \log M_o = (0.96 \pm 0.06)M_L + 18.14 \pm 0.23 )</td>
<td>( 2.9 \leq M_L \leq 6.2 )</td>
<td>40</td>
<td>1</td>
</tr>
<tr>
<td><strong>Southern California</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Thatcher and Hanks [1973]</td>
<td>( \log M_o = (1.05 \pm 0.08)M_L + 17.76 \pm 0.33 )</td>
<td>( 3.5 \leq M_L \leq 6.2 )</td>
<td>29</td>
<td>1</td>
</tr>
<tr>
<td>Fletcher et al. [1984]</td>
<td>( \log M_o = (1.11 \pm 0.15)M_L + 17.92 \pm 1.02 )</td>
<td>( 3 \leq M_L \leq 6.2 )</td>
<td>16</td>
<td>6</td>
</tr>
<tr>
<td>Archuleta et al. [1982]</td>
<td>( \log M_o = (1.36 \pm 0.22)M_L + 16.78 \pm 1.07 )</td>
<td>( 4.3 \leq M_L \leq 5.7 )</td>
<td>7</td>
<td>2</td>
</tr>
<tr>
<td>( \log M_o = (1.08 \pm 0.14)M_L + 18.00 \pm 0.51 )</td>
<td>( 2.8 \leq M_L \leq 4.1 )</td>
<td>7</td>
<td>N/A</td>
<td></td>
</tr>
</tbody>
</table>
source location is of no real consequence. Finally, no good explanation exists for the unusually low $M_0$ of the $M_L = 5.2$ Oroville aftershock [Fletcher et al., 1984].

**MODELING OF THE MOMENT-MAGNITUDE DATA**

Figure 2 presents a large number of moment-magnitude data for central California earthquakes. The preponderance comes from Bakun ([1984], 118 events), although the results from a number of other studies have also been used (Figure 2 caption). The entries in Table 2 are also included in Figure 2, these being our preferred estimates for these earthquakes.

Model calculations (large solid circles in Figure 2) are obtained from the numerical simulations of Boore [1983] according to the prescription of Hanks and McGuire [1981] that the far-field shear-wave acceleration be finite-duration, band-limited white Gaussian noise. The duration is the faulting duration $T_a$ beginning with the direct shear-arrival; the band-limited corner frequency $f_0 = 1/T_a$ and at high frequencies by the source-size-independent cutoff frequency $f_{\text{max}}$. At the larger magnitudes, it can be observed that the model calculations reproduce the continuous log moment $M_L$ for large earthquakes obtained from strong motion accelerograms at close distances, $M_L$ (SM), is slightly biased with respect to the $M_L$ obtained from standard Wood-Anderson seismograms necessarily located at much greater distances (hundreds of kilometers), $M_L$ (WA). For $M_L > 5.3$ the Luco [1982] correction is

$$M_L(\text{SM}) = 0.7M_L(\text{WA}) + 1.5$$

and this correction to our model calculations is given by the dashed line in Figure 2.

Observationally, Figure 2 does not leave much to the imagination. There is good reason to believe that earthquakes significantly larger than the 1906 earthquake cannot occur in California [Hanks and Kanamori, 1979]. Figure 2, however, also contains data for the smallest earthquakes ($M_L \approx 0$) that can be recorded in California, at least under ordinary conditions of recording earthquakes at ordinary crustal depths. While there is yet a data gap for $10.25 < M_0 < 10.27$ dyne cm in central California (Figure 2), we are confident that there are no latent surprises here, if available southern California data for this $M_L$ range and our model calculations mean anything at all. The model calculations reproduce the continuous log $M_0 - M_L$ curvature in Figure 2 very well (or, if one prefers, c-values that increase with $M_L$), but a more fundamental result is at hand: consistent with the findings of Hanks and McGuire [1981] and Boore [1983]—but here expressed for the entire range of earthquakes that can be recorded locally in the far-field shear-wave acceleration be finite-duration, band-limited white Gaussian noise. The duration is the faulting duration $T_a$ beginning with the direct shear-arrival; the band-limited corner frequency $f_0 = 1/T_a$ and at high frequencies by the source-size-independent cutoff frequency $f_{\text{max}}$. At the larger magnitudes, it can be observed that the model calculations reproduce the continuous log moment $M_L$ for large earthquakes obtained from strong motion accelerograms at close distances, $M_L$ (SM), is slightly biased with respect to the $M_L$ obtained from standard Wood-Anderson seismograms necessarily located at much greater distances (hundreds of kilometers), $M_L$ (WA). For $M_L > 5.3$ the Luco [1982] correction is

$$M_L(\text{SM}) = 0.7M_L(\text{WA}) + 1.5$$

and this correction to our model calculations is given by the dashed line in Figure 2.

Observationally, Figure 2 does not leave much to the imagination. There is good reason to believe that earthquakes significantly larger than the 1906 earthquake cannot occur in California [Hanks and Kanamori, 1979]. Figure 2, however, also contains data for the smallest earthquakes ($M_L \approx 0$) that can be recorded in California, at least under ordinary conditions of recording earthquakes at ordinary crustal depths. While there is yet a data gap for $10.25 < M_0 < 10.27$ dyne cm in central California (Figure 2), we are confident that there are no latent surprises here, if available southern California data for this $M_L$ range and our model calculations mean anything at all. The model calculations reproduce the continuous log $M_0 - M_L$ curvature in Figure 2 very well (or, if one prefers, c-values that increase with $M_L$), but a more fundamental result is at hand: consistent with the findings of Hanks and McGuire [1981] and Boore [1983]—but here expressed for the entire range of earthquakes that can be recorded locally in the far-field shear-wave acceleration be finite-duration, band-limited white Gaussian noise. The duration is the faulting duration $T_a$ beginning with the direct shear-arrival; the band-limited corner frequency $f_0 = 1/T_a$ and at high frequencies by the source-size-independent cutoff frequency $f_{\text{max}}$. At the larger magnitudes, it can be observed that the model calculations reproduce the continuous log moment $M_L$ for large earthquakes obtained from strong motion accelerograms at close distances, $M_L$ (SM), is slightly biased with respect to the $M_L$ obtained from standard Wood-Anderson seismograms necessarily located at much greater distances (hundreds of kilometers), $M_L$ (WA). For $M_L > 5.3$ the Luco [1982] correction is

$$M_L(\text{SM}) = 0.7M_L(\text{WA}) + 1.5$$

and this correction to our model calculations is given by the dashed line in Figure 2.

Observationally, Figure 2 does not leave much to the imagination. There is good reason to believe that earthquakes significantly larger than the 1906 earthquake cannot occur in California [Hanks and Kanamori, 1979]. Figure 2, however, also contains data for the smallest earthquakes ($M_L \approx 0$) that can be recorded in California, at least under ordinary conditions of recording earthquakes at ordinary crustal depths. While there is yet a data gap for $10.25 < M_0 < 10.27$ dyne cm in central California (Figure 2), we are confident that there are no latent surprises here, if available southern California data for this $M_L$ range and our model calculations mean anything at all. The model calculations reproduce the continuous log $M_0 - M_L$ curvature in Figure 2 very well (or, if one prefers, c-values that increase with $M_L$), but a more fundamental result is at hand: consistent with the findings of Hanks and McGuire [1981] and Boore [1983]—but here expressed for the entire range of earthquakes that can be recorded locally in the far-field shear-wave acceleration be finite-duration, band-limited white Gaussian noise. The duration is the faulting duration $T_a$ beginning with the direct shear-arrival; the band-limited corner frequency $f_0 = 1/T_a$ and at high frequencies by the source-size-independent cutoff frequency $f_{\text{max}}$. At the larger magnitudes, it can be observed that the model calculations reproduce the continuous log moment $M_L$ for large earthquakes obtained from strong motion accelerograms at close distances, $M_L$ (SM), is slightly biased with respect to the $M_L$ obtained from standard Wood-Anderson seismograms necessarily located at much greater distances (hundreds of kilometers), $M_L$ (WA). For $M_L > 5.3$ the Luco [1982] correction is

$$M_L(\text{SM}) = 0.7M_L(\text{WA}) + 1.5$$

and this correction to our model calculations is given by the dashed line in Figure 2.

Observationally, Figure 2 does not leave much to the imagination. There is good reason to believe that earthquakes significantly larger than the 1906 earthquake cannot occur in California [Hanks and Kanamori, 1979]. Figure 2, however, also contains data for the smallest earthquakes ($M_L \approx 0$) that can be recorded in California, at least under ordinary conditions of recording earthquakes at ordinary crustal depths. While there is yet a data gap for $10.25 < M_0 < 10.27$ dyne cm in central California (Figure 2), we are confident that there are no latent surprises here, if available southern California data for this $M_L$ range and our model calculations mean anything at all. The model calculations reproduce the continuous log $M_0 - M_L$ curvature in Figure 2 very well (or, if one prefers, c-values that increase with $M_L$), but a more fundamental result is at hand: consistent with the findings of Hanks and McGuire [1981] and Boore [1983]—but here expressed for the entire range of earthquakes that can be recorded locally in the far-field shear-wave acceleration be finite-duration, band-limited white Gaussian noise. The duration is the faulting duration $T_a$ beginning with the direct shear-arrival; the band-limited corner frequency $f_0 = 1/T_a$ and at high frequencies by the source-size-independent cutoff frequency $f_{\text{max}}$. At the larger magnitudes, it can be observed that the model calculations reproduce the continuous log moment $M_L$ for large earthquakes obtained from strong motion accelerograms at close distances, $M_L$ (SM), is slightly biased with respect to the $M_L$ obtained from standard Wood-Anderson seismograms necessarily located at much greater distances (hundreds of kilometers), $M_L$ (WA). For $M_L > 5.3$ the Luco [1982] correction is

$$M_L(\text{SM}) = 0.7M_L(\text{WA}) + 1.5$$

and this correction to our model calculations is given by the dashed line in Figure 2.

Observationally, Figure 2 does not leave much to the imagination. There is good reason to believe that earthquakes significantly larger than the 1906 earthquake cannot occur in California [Hanks and Kanamori, 1979]. Figure 2, however, also contains data for the smallest earthquakes ($M_L \approx 0$) that can be recorded in California, at least under ordinary conditions of recording earthquakes at ordinary crustal depths. While there is yet a data gap for $10.25 < M_0 < 10.27$ dyne cm in central California (Figure 2), we are confident that there are no latent surprises here, if available southern California data for this $M_L$ range and our model calculations mean anything at all. The model calculations reproduce the continuous log $M_0 - M_L$ curvature in Figure 2 very well (or, if one prefers, c-values that increase with $M_L$), but a more fundamental result is at hand: consistent with the findings of Hanks and McGuire [1981] and Boore [1983]—but here expressed for the entire range of earthquakes that can be recorded locally in
Fig. 1. Moment-magnitude relations and data for 18 central California earthquakes, M_L > 5. The moment-magnitude relations are given in Table 1: 1, Johnson and McEvilly [1974]; 2, Bakun and Lindh [1977]; 3, Archuleta et al. [1982], a, 2.9 < M_L < 6.2, b, 3.5 < M_L < 6.2; 4, Bolt and Herriz [1983]; 5, Fletcher et al. [1984], a, 4.3 < M < 5.7, b, 2.8 < M < 4.1. The dashed line is the relation of Thatcher and Hanks [1973]. The M > 5 data are from Table 2, the box representing the range of estimates for the 1906 earthquake.

California—the a_mw stress drop of 100 bars is a stable (to a factor of 2 or so) and entire feature of California earthquakes.

ASYMPTOTIC APPROXIMATIONS TO THE NUMERICAL CALCULATIONS

The curvature of the synthetic moment-magnitude relation in Figure 2 results from a complex interaction in the frequency domain, resulting in three essential bandwidths determined by f_s, the natural frequency of the standard Wood-Anderson torsion seismograph (1.25 Hz); f_max (15 Hz); and f_0, the earthquake corner frequency fixed by the constant stress drop relation

$$M_0f_0^{3/2}$$

where \( \beta \) is the shear-wave velocity [Hanks and McGuire, 1981]. Straight-line approximations of the Wood-Anderson instrumental response at gain V to acceleration and the square" source acceleration spectrum [Aki, 1967; Brune, 1970] with the f_max cutoff [Hanks, 1982] are shown at the top of Figure 3. The Wood-Anderson record spectrum is formed from the product of the two, which reduces to addition in the logarithmic space of Figure 3. In passing from the largest earthquakes to the smallest, \( f_0 \) sweeps from a value <<f_s to a value >>f_max. This results in three essential bandwidths for the record spectrum, approximated by the three sketches in the lower part of Figure 3. For each of these, we can extract a linear relation between log \( M_0 \) and \( M_L \) that approximates the curvature of the calculations (and the observations) with connected straight-line segments, in the following manner.

For each of these boxcar-like spectrums, the maximum Wood-Anderson record amplitude can be estimated with the relation

$$u_{WA} \sim \Delta f_0$$

where \( \Delta \) is the constant value of spectral amplitudes across the bandwidth \( \Delta f_0 \) (Figure 3). For the three cases in Figure 3, \( \Delta f_0 \) and \( u_{WA} \) are

(i) \( f_0 \ll f_s \ll f_{\text{max}} \): \( \Delta f_0 = f_s - f_0 \approx f_s \)

$$u_{WA} \sim VM_0f_0^{3/2}$$

(ii) \( f_s \ll f_0 < f_{\text{max}} \): \( \Delta f_0 = f_0 - f_s \approx f_0 \)

$$u_{WA} \sim VM_0f_0$$

(iii) \( f_s < f_{\text{max}} < f_0 \): \( \Delta f_0 = f_{\text{max}} - f_s \approx f_{\text{max}} \)

$$u_{WA} \sim VM_0f_{\text{max}}$$

Taking \( M_L \sim \log u_{WA} \) and \( f_0 \sim M_0^{-1/3} \) for constant stress drop (e.g., equation (5)), we can write equations (7) as

(i) log \( M_0 \sim 3.0 M_L \)

(ii) log \( M_0 \sim 1.5 M_L \)

(iii) log \( M_0 \sim 1.0 M_L \)

Taking \( M_L \sim \log u_{WA} \) and \( f_0 \sim M_0^{-1/3} \) for constant stress drop (e.g., equation (5)), we can write equations (7) as

In view of the highly idealized assumptions leading to equation (8), it would be unwise to make too much of these asymptotic, linear relations. Indeed, our principal interest in them is to illustrate, in a qualitative sense, the nature of the intrinsically complicated model calculations in Figure 2. Even so, the approximations (8) agree with the observational and model results of Figure 2 reasonably well, as described in the paragraphs below.

We estimate with equations (5) and (2) that \( f_0 \) should begin to exceed f_max at \( M_L \approx 2\). At this magnitude and smaller, then, we infer a c value of 1 (equation (8c)), and this seems to be appropriate (e.g., Figure 2 and Bakun [1984]). The one-to-one correspondence between \( M_L \) and log \( M_0 \) when \( f_0 \gg f_{\text{max}} \) arises because the frequency dependence of \( u_{WA} \) is due to \( f_{\text{max}} \) alone, a fixed parameter (equation (7c)). It is incorrect, although commonly held [Randall, 1973; Archuleta et al., 1982], that the maximum Wood-Anderson displacement amplitude to a Brune pulse is linearly proportional to \( M_0 \) alone when \( f_0 > f_s \), yielding a one-to-one relationship between log \( M_0 \) and \( M_L \) for small events. In fact, in the absence of the effect of \( f_{\text{max}} \) i.e., \( f_{\text{max}} \ll f_0 \), \( u_{WA} \) of a Brune pulse is proportional to the product \( M_0f_0 \) when \( f_0 > f_s \). We have captured this result in equation (7b), although it may be obtained directly by evaluating the Brune [1970, 1971] displacement pulse at the time of maximum displacement. Equations (7c) and (8c) moreover tell us that when \( f_0 > f_{\text{max}} \) the \( M_L \) dependence on \( M_0 \) is insensitive to stress drop, so that our earlier conclusion concerning the ubiquity of the a_mw stress drop of 100 bars is only determined at \( M_L \approx 2\). Until such time as we know the underlying physical causes of \( f_{\text{max}} \), there is no way, in fact, of knowing anything about earthquake stress differences when \( f_0 \) exceeds \( f_{\text{max}} \).

For \( f_0 \approx 1.25 \text{ Hz} (=f_s) \) at \( M_L \approx 4.5 \), as suggested by the Oroville aftershocks [Fletcher et al., 1984], the approximations suggest the c value should increase from 1.5 to 3 at
Fig. 2. Moment-magnitude data for central California earthquakes (crosses, box for the 1906 earthquake) and model calculations after Boore [1983] (solid circles, dashed line for Luco [1982] correction (equation (4)). Data sources: D. J. Andrews (personal communication, 1983), Archuleta et al. [1982], Bakun [1984], Bakun and Lindh [1977], Fletcher et al. [1984], Followill et al. [1982], Helmberger and Johnson [1977], Helmberger and Malone [1975], Uhrhammer [1981], and Table 2. The model calculations are described in the text.

The difference in $c$ values for the moment-magnitude relations of central and southern California (Table 1) is a geographic appearance, not a geographic reality; it results from the preponderance of small ($M_L < 5$) earthquakes that form the bulk of the central California data set. The continuous, positive curvature of the log $M_o - M_L$ observations can be approximated by the linear relations of Bakun [1984], reproduced here as equation (3), for $M_L \leq 6$. The appropriate linear approximation above $M \approx 6$ has not been defined, nor will it ever be well-defined empirically, if we correctly anticipate that $c$ will become very large in the neighborhood of $M_L = 7$. Neither can it be expected that our asymptotic approximation (8) will be very helpful in this range.

Our full numerical calculations, however, match the continuous curvature of log $M_o - M_L$ data very well, a data set that represents the entire range of earthquakes that can be locally recorded in California. In view of this fit (Figure 2), we simply need not be concerned about linear approximations: the calculations of Boore [1983] allow one to calculate $M_L$ for any $M_o$, given chosen values of $\Delta \sigma$ and $f_{\text{max}}$. The remarkably good fit of model to data in Figure 2 must mean that the $a_{\text{rms}}$ stress drop of 100 bars is a well-conditioned and pervasive property of California earthquakes in the "visible" bandwidth, $f_0 \leq f_{\text{max}} \approx 15$ Hz, corresponding to $M_L \gtrsim 2\frac{1}{2}$.

Finally, it is worth emphasizing that just as the results in Figure 2 (either theoretical or observational) do not depend on the adjectives "central" or "southern" when speaking of California earthquakes, neither do they depend on the modifier "California" when speaking of "plate margin" earthquakes. In a summary of average source-parameter relations for plate
margin earthquakes on a world-wide basis, Nuttli [1984] has determined the relations.

\[ \log M_0 = 1.0 m_b + 18.15 \]

\[ \log M_0 = 2.0 m_b + 13.75 \]

With an origin shift of

\[ m_b = M_L - 0.4 \]

(O. Nuttli, personal communication, 1984), equations (9a) and (9b) also fit the observations in Figure 2 well and are very nearly concordant with the model calculations presented there.

Acknowledgments. We have enjoyed the critical commentary of W. H. Bakun, R. B. Herrmann, and A. McGarr in preparing this manuscript for publication. O. Nuttli provided us with a preprint of Nuttli [1984] and his thoughts on this study that allowed us to write the paragraph just above at a late date.

REFERENCES


Ferguson, R., B. Schechter, and R. A. Uhrhammer, Bulletin of the Seismographic Stations of the University of California, vol. 50, number 1, 87 pp., Univ. of Calif., Berkeley, 1980.


McKenzie, M. R., R. D. Miller, and R. A. Uhrhammer, Bulletin of the
Seismographic Stations of the University of California, vol. 50, number 2, 156 pp., Univ. of Calif., Berkeley, 1980.


(Received August 22, 1983; revised January 27, 1984; accepted March 13, 1984.)