Moment-Magnitude Relations in Theory and Practice

THOMAS C. HANKS AND DAVID M. BOORE

U.S. Geological Survey

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 \( M \):

\[
\log M_0 = c M + d \quad (1)
\]

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

At the present time, the significance of such relations is 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

\[
M = \frac{1}{3} \log M_0 - 10.7 \quad (2)
\]

is identical (in reverse form) to the moment-magnitude relations of Thatcher and Hanks [1973] for southern California earthquakes (3 \( \leq \) \( M_L \) \( \leq \) 7), of Pucaru 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 [Hanks and McGuire, 1981; Boore, 1983] have indicated that ground motion scaling laws, at least in California, can be reduced to a dependence on a single parameter, \( M_0 \). Most of the empirical ground motion studies, however, relate the parameter of interest to some \( M \), and thus the transformation between \( M_0 \) and the appropriate \( M \) is necessary to equate the theoretical and empirical studies. Ideally, again, some uniformly valid moment magnitude scale would be just the thing.

Even when \( c = 1.5 \) and \( d = 16.0 \), for which (1) and (2) are nominally the same, it is worth emphasizing the important differences between (1) and (2). Equation (2) is simply a definition, as any magnitude scale is, with constants that more or less square with observational findings, if one is careful in choosing permissible ranges for the appropriate \( M \)'s; its virtue is that it is uniformly valid in \( M_0 \). Equation (1), however, will always fail for a large enough variation in \( M_0 \). This failure is due to the size-dependent frequency characteristics of the source excitation and the finite record bandwidth of any time domain amplitude-based \( M \), phenomena that lead to magnitude and peak acceleration saturation [Hanks and Kanamori, 1979].

Observationally, this failure of (1) is expressed by \( c \) values that increase with \( M_L \). A number of \( M_0 - M_L \) relations have now been reported for California earthquakes. These are summarized in Table 1, and a discussion of them forms the next section. Here it suffices to note the large difference in the \( c \) value found by Thatcher and Hanks [1973] for southern California data compared to those obtained in central California. Is this difference a function of the source region or (as we and Bakun [1984] believe) the preponderance of small \(( M_L \leq 4)\) earthquakes contributing to the central California data set? The matter is far from academic, should one be interested in estimating \( M_0 \) for an \( M_L = 6.5 \) earthquake in central California, say for the purposes of ground motion estimation mentioned two paragraphs above. The relations of Bakun and Lindh [1977], Archuleta et al. [1982] (3.5 \( \leq \) \( M_L \) \( \leq \) 6.2), and Bolt and Herrais [1983] yield 0.8, 0.4, and 1.3 \( \times \) \( 10^{25} \) dyne cm, respectively, whereas the relation of Thatcher and Hanks [1973] yields 6.3 \( \times \) \( 10^{25} \) dyne cm, 5–20 times larger.

SEISMIC MOMENT-LOCAL MAGNITUDE RELATIONS

Investigation of these relations begins with Wyss and Brune [1968], who give two. The first of these is for the "San Andreas fault." It is formed from analysis of 12 events near Hollister, Parkfield, and Brawley and an additional earthquake in the Gulf of California, the last event being the only one with \( M_L \geq 5.0 \). This relation is

\[
\log M_0 = 1.4M_L + 17.0 \quad 3 \leq M_L \leq 6
\]

Their second relation is for 259 earthquakes throughout the western United States, the \( M_L \) results being obtained from the
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 qualifies 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 Herreraiz [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 relatively small number of \( M_L \geq 5.0 \) central California 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 \( M_L \geq 5.0 \), 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 \( M_L \geq 5.0 \), 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 Herreraiz [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
source location is of no real consequence. Finally, no good explanation exists for the unusually low $M_0$ of the $M_r = 5.2$ 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 $T_d$ beginning with the direct shear-arrival; the band-width is determined at low frequencies by the earthquake corner frequency $f_0 = 1/T_d$ and at high frequencies by the source-size-independent cutoff frequency $f_{\text{max}}$ [Hanks, 1982]. The calculations of Boore [1983] generate stochastic realizations of this prescription, as constrained in addition by the related Brune [1970, 1971] spectrum and the $f_{\text{max}}$ stress drop $\Delta \sigma$ [Hanks and McGuire, 1981]. From these synthetic acceleration time histories, all manner of time domain and spectral amplitudes can be calculated [Boore, 1983], including the maximum amplitude on the Wood-Anderson seismogram that is the basis of $M_L$.

Just as in the work of Hanks and McGuire [1981], the independent variables for the Boore [1983] calculations are $\Delta \sigma$, $f_0$, and $f_{\text{max}}$. The finding of Hanks and McGuire [1981] and Boore [1983] that $\Delta \sigma$ is constant (namely, 100 bars) allows the $\Delta \sigma$ and $f_0$ dependences to be condensed to $M_0$ (e.g., equation (5) below). If, in addition, we take $f_{\text{max}}$ to be constant, the model calculations of Boore [1983] reduce to a dependence on $M_0$ alone. In particular, then, $M_0$ is the independent variable for the calculations in Figure 2, $\Delta \sigma = 100$ bars and $f_{\text{max}} = 15$ Hz being fixed parameters, and $M_L$ is the derived quantity.

In view of the many restrictive assumptions forced on the model calculations, we consider their fit to the data in Figure 2 to be surprisingly good, if any allowance at all is made for naturally arising scatter in the observations not related to variable $f_{\text{max}}$ stress drop. At the larger magnitudes, it can be improved somewhat with the finding of Luco [1982] that the $M_L$ for large earthquakes obtained from strong motion accelerograms at close distances, $M_L$ (SMA), 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

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

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 $M_0$ 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

<table>
<thead>
<tr>
<th>Location</th>
<th>Date</th>
<th>Origin Time</th>
<th>$M_L$</th>
<th>$M_0$, dyne cm</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>San Francisco</td>
<td>April 18, 1906</td>
<td>1312</td>
<td>6.3</td>
<td>7.0</td>
<td>Ben-Menahem [1978], Boore [1984], Jennings and Kanamori [1979], Thatcher [1975]</td>
</tr>
<tr>
<td>Parkfield</td>
<td>June 8, 1934</td>
<td>0447</td>
<td>5.6</td>
<td>25.064</td>
<td>Bakun and McEvilly [1984]</td>
</tr>
<tr>
<td>Parkfield</td>
<td>June 28, 1966</td>
<td>0426</td>
<td>5.6</td>
<td>25.14</td>
<td>Kanamori and Jennings [1978], Tsai and Aki [1969]</td>
</tr>
<tr>
<td>Truckee</td>
<td>Sept. 12, 1966</td>
<td>1641</td>
<td>5.8</td>
<td>24.85</td>
<td>Burdick [1977], Ryall et al. [1968], Tsai and Aki [1970]</td>
</tr>
<tr>
<td>Sargent fault</td>
<td>Dec. 18, 1967</td>
<td>1724</td>
<td>5.3</td>
<td>23.5</td>
<td>Bakun [1984]</td>
</tr>
<tr>
<td>Melendy Ranch</td>
<td>Feb. 24, 1972</td>
<td>1556</td>
<td>5.1</td>
<td>23.49</td>
<td>Helmberger and Johnson [1977]</td>
</tr>
<tr>
<td>Orroville</td>
<td>Aug. 1, 1975</td>
<td>2020</td>
<td>5.7</td>
<td>24.80</td>
<td>Langston and Butler [1976], Morrison et al. [1976], Wallace (quoted by Cohn et al. [1982])</td>
</tr>
<tr>
<td>Orroville</td>
<td>Aug. 2, 1975</td>
<td>2022</td>
<td>5.1</td>
<td>23.52</td>
<td>Fletcher et al. [1984]</td>
</tr>
<tr>
<td>Orroville</td>
<td>Aug. 2, 1975</td>
<td>2059</td>
<td>5.2</td>
<td>22.58</td>
<td>Fletcher et al. [1984]</td>
</tr>
<tr>
<td>Livermore Valley</td>
<td>Jan. 24, 1980</td>
<td>1900</td>
<td>5.8</td>
<td>24.68</td>
<td>Bolt et al. [1981], Ferguson et al. [1980]</td>
</tr>
<tr>
<td>Mammoth Lakes</td>
<td>May 25, 1980</td>
<td>1649</td>
<td>6.0</td>
<td>24.57</td>
<td>Bolt et al. [1981], Ferguson et al. [1980]</td>
</tr>
<tr>
<td>Mammoth Lakes</td>
<td>May 27, 1980</td>
<td>1450</td>
<td>6.2</td>
<td>25.03</td>
<td>Givens et al. [1982], Uhrhammer and Ferguson [1980]</td>
</tr>
<tr>
<td>Eureka</td>
<td>Nov. 8, 1980</td>
<td>1027</td>
<td>6.9</td>
<td>27.11</td>
<td>Barker and Langston [1983], Givens et al. [1982], Uhrhammer and Ferguson [1980]</td>
</tr>
<tr>
<td>Coalinga</td>
<td>May 2, 1983</td>
<td>2342</td>
<td>6.2</td>
<td>25.7</td>
<td>McKenzie et al. [1980], Lay et al. [1982]</td>
</tr>
</tbody>
</table>
LOG MOMENT - ML RELATIONS
AND DATA FOR M>5

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 Herraez [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 $a_{ms}$ 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

$$M_{0}f_{0}^{3/2} = 8.47 \Delta \sigma = 100 \text{ bars} = 10^{2.73} M_{0f_{0}}^{3/2}$$

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

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

where $A$ 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) $f_0 \ll f_s < f_{max}$: $\Delta f_0 = f_s - f_0 \approx f_s$

$$u_{WA} \sim VM_0 f_0$$

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

$$u_{WA} \sim VM_0 f_0$$

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

$$u_{WA} \sim VM_0 f_{max}$$

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) $\log M_0 \sim 1.0 M_L$

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 $f_0 \gtrsim f_{max}$ arises because the frequency dependence of $u_{WA}$ is due to $f_{max}$ 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 $M_0$ when $f_0 > f_{max}$, 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 $f_{max}$ (i.e., $f_{max} \approx f_0$), $u_{WA}$ of a Brune pulse is proportional to the product $M_0 f_0$ 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 $f_0 > f_{max}$, the $M_L$ dependence on $M_0$ is insensitive to stress drop, so that our earlier conclusion concerning the ubiquity of the arm's stress drop of 100 bars is only determined at $M_L \gtrsim 2$. Until such time as we know the underlying physical causes of $f_{max}$, there is no way, in fact, of knowing anything about earthquake stress differences when $f_0$ exceeds $f_{max}$.

For $f_0 \approx 1.25$ 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.

$M_L \approx 4\frac{1}{2}$. This asymptotic prediction is less accurate: the observations suggest a steepening above $c = 1.5$ does not occur until $M_L \geq 6$. Above $M_L \approx 6$, it is hard to define a $c$ value with much accuracy, although $c = 3$ is certainly not inappropriate for $6 \leq M_L \leq 7$. In the vicinity of $M_L \approx 7$, the "saturation point" of $M_L$ [Hanks and Kanamori, 1979], we suspect that $c$ can obtain arbitrarily large values for reasons not included in the derivation of equations (8): an arbitrarily large earthquake cannot be observed isotropically at any finite distance—any maximum record amplitude will always be associated with some smaller segment of faulting.

**SUMMARY AND CONCLUSIONS**

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 $M_0 - M_L$ 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 $M_0 - M_L$ 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 $M_0$, 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 $M_L \geq 2\frac{1}{2}$.

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.

$$\log M_0 = 1.0 m_o + 18.15 \quad m_o \leq 4.4 \quad (9a)$$

$$\log M_0 = 2.0 m_o + 13.75 \quad 4.4 \leq m_o \leq 6.9 \quad (9b)$$

With an origin shift of

$$m_o = M_L - 0.4 \quad (9c)$$

(O. Nuttli, personal communication, 1984), equations (9a) and (9b) also fit the observations in Figure 2 well and are very nearly coincident 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
HANKS AND BOORE: A MOMENT OF LOCAL MAGNITUDE

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.)