USE OF SEISMIC RADIATION TO INFER SOURCE PARAMETERS

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1. BRIEF REVIEW OF SEISMIC METHODS FOR DETERMINATION OF SOURCE PARAMETERS

In this section we list earthquake source parameters relevant to the subject and briefly summarize commonly used seismological methods for the determination of these parameters.

(1) Fault Geometry

The geometry of an earthquake fault can usually be defined by three parameters: the fault strike, the angle of the fault plane and the slip angle of the fault motion. Other representations and useful relations between them are given by Jarosch and Aboodi (1970). These parameters can be determined by 1) P-wave first-motion data (Stauder, 1962; Honda, 1962), 2) S-wave polarization angles, (Stauder, 1962; Honda, 1962; Hirasawa, 1966), 3) wave forms of body waves (Langston and Helmberger, 1975; Langston, 1976; Langston and Butler, 1976), 4) radiation pattern of surface-waves (Brune, 1961; Ben-Menahem and Harkrider, 1964; Aki, 1966; Kanamori, 1970; Ben-Menahem et al., 1970), 5) the excitation of normal modes (Saito, 1967; Abe, 1970; Ben-Menahem, et al., 1971; Dziewonski and Gilbert, 1975; Gilbert and Dziewonski, 1975), 6) geodetic data (e.g., Chinnery, 1964, 1969; Savage and Hastie, 1966; Ando, 1971), and 7) field observations.

(2) Fault Dimension

The dimension of an earthquake fault can be determined from 1) the size of the aftershock area, 2) geodetic data, 3) tsunami source area, 4) directivity and asymmetry of the radiation pattern of long-period surface waves, 5) pulse width of body waves, and 5) seismic corner frequency.

1) Aftershock Area

Although there is no standard definition of aftershocks and the aftershock area, the aftershock area defined by the somewhat subjective
judgment of the investigator often provides a very good estimate of
the fault area of very large earthquakes (fault dimension \( \geq 100 \) km),
particularly for great shallow thrust earthquakes along subduction
zones (Benioff et al., 1961; Press et al., 1961; Mogi, 1968a). For
small earthquakes, errors in the epicentral location of aftershocks
and the temporal expansion of the aftershock area often cause a substantial
error in the estimate of the size of the fault plane.

2) Geodetic Data

When geodetic data (leveling and triangulation) are available
over the entire area of faulting, the size of the fault plane can
be determined very well (Savage and Hastie, 1966; Plafker, 1972; Ando,
1971; Kasahara, 1957; Chinnery, 1964; Kanamori, 1973; Jungles and
Frasier, 1973). The spatial decay rate of the displacement field
can be used to infer the vertical extent of the faulting (Knopoff,
1958), although the resolution is often limited by the quality and
quantity of the data.

3) Tsunami Source Area

When tide-gage data are available near the epicenter of a large
tsunamiogenic earthquake, the source area of tsunami can be estimated
by using the inverse refraction diagram (Miyabe, 1934; Hatori, 1966;
Abe, 1973). Usually, a good correlation between the size of tsunami
source area and the aftershock area (Hatori, 1965; Abe, 1973) is found.

4) Directivity

From the directivity of very long-period (200 to 300 sec) surface
waves, the rupture length can be estimated (Ben-Menahem, 1961). The
asymmetry of the radiation pattern can also be used (Kanamori, 1970).
For very large earthquakes such as the 1960 Chilean earthquake, the 1952 Kamchatka earthquake, the 1964 Alaskan earthquake, these methods gave a reliable estimate of the fault length, perhaps accurate to \( \pm 15\% \) (Benioff et al., 1961; Press et al., 1961; Ben-Menahem and Toksoz, 1963). However, for events whose dimension is smaller than 100 km, the resolution of these methods becomes very poor.

5) Pulse Width of Body Waves

The pulse width of body waves was used to infer the fault dimension of deep focus earthquakes (Fukao, 1970). In order to determine the source time function from observed body waves from shallow focus earthquakes, the effect of structure near the source, particularly the free surface effects, and propagation effects must be removed from the observed records. Techniques have been developed to correct for these effects (Fukao, 1971; Helmberger, 1974; Langston and Helmberger, 1975; Helmberger and Malone, 1975), and the source time function can be recovered very accurately for relatively simple events (Burdick and Mellman, 1976). For complex events, the analysis of body waves is more difficult but several attempts have been made (Fukao, 1972; Chung and Kanamori, 1976; Kanamori and Stewart, 1978; Chung and Kanamori, 1978) to recover the complex time history of the rupture process. The interpretation of the pulse width in terms of the source dimension involves assumptions on the geometry of the fault, mode of rupture and rise time of the local slip function, and is often nonunique.

6) Corner Frequency

The corner frequency of the spectrum of body-waves is a frequency domain representation of the pulse width. Brune (1970) proposed a
relation between the source dimension and the corner frequency. Although Brune's equation provides a useful average relation, the estimate for an individual event depends on the geometry of the fault, the mode of rupture and the rise time of the local slip function. When the wave form becomes complex due to source complexity and propagation effects, including reflections and refractions, interpretation of the corner frequency in terms of the source dimension becomes very difficult.

(3) Rupture Mode and Rupture Velocity

Whether the fault rupture is unilateral, bilateral or two-dimensional is usually determined from the spatial relation of the main shock to the aftershock area. For very large earthquakes, the rupture velocity can be determined from the directivity function (Ben-Menahem, 1961; Benioff et al., 1961). For multiple shocks, the apparent rupture velocity is given by the ratio of the spatial separation to the temporal separation of the individual events. The rupture velocity is sometimes determined from the wave forms of near field records (Aki, 1968; Kanamori, 1972; Abe, 1974a).

(4) Dislocation

The dislocation on the fault plane is in general a function of position and time. From geodetic data, the static value of the dislocation can be determined as a function of position on the fault. However, the details of the spatial distribution are usually very difficult to resolve (Chinnery, 1964; Kasahara, 1957; Savage and Hastie, 1966; Kanamori, 1973).

If the fault area is known, the dislocation can be estimated from the amplitude of seismic body waves, surface waves and free oscillations.
However, it is very difficult to resolve the details of the spatial distribution of the dislocation; usually, only a spatial average can be determined.

(5) Particle Velocity

The particle velocity at a point on the fault plane is directly related to the effective tectonic stress (Brune, 1970). In principle, the particle velocity can be determined from the frequency spectrum or the rise time of near field seismograms. However, it is difficult to remove the effect of rupture propagation and near source geometry from the observed seismogram. Only a few determinations of the particle velocity have been made (Kanamori, 1972; Abe, 1974a,b, 1975a).

(6) Complexity

The complexity of faulting process can be determined by the analysis of distinct arrivals on seismograms (Imamura, 1937; Miyamura et al., 1964; Wyss and Brune, 1967; Trifunac and Brune, 1970). More recently synthetic seismograms have been used to determine more details of the multiple shock sequence (Fukao, 1972; Chung and Kanamori, 1976; Kanamori and Stewart, 1978; Chung and Kanamori, 1978). Detailed study of complexity of faulting is important in understanding the stress state in the fault zone and also in predicting ground motions resulting from an earthquake.

2. SUMMARY OF RESULTS

(1) Geodetic Data

Geodetic data (both leveling and triangulation) are summarized by Rikitake (1974) and are shown in Figure 1. Figure 1 shows the
strain change in the immediate vicinity of the epicenters of various earthquakes as a function of magnitude. The strain seems to be almost constant from $2 \times 10^{-5}$ to $2 \times 10^{-4}$ regardless of the magnitude of the earthquake. Since the rigidity of crustal rocks averages $3.5 \times 10^{11}$ dyne/cm$^2$, this strain change corresponds to a stress drop of 7 to 70 bars (Chinnery, 1964).

(2) Great and Large Earthquakes

The results for large and great earthquakes are summarized by Kanamori and Anderson (1975) and Geller (1976) (Table 1).

Figure 2 shows the relation between $\log S$ ($S$: fault area) and $\log M_0$ ($M_0 = \mu \Delta S$: seismic moment) for great and large earthquakes. The remarkable linearity between $\log M_0$ and $\log S$ can be interpreted in terms of a constant average stress drop (30 to 60 bars) in earthquakes (Aki, 1972; Kanamori and Anderson, 1975; Abe, 1975b; Geller, 1976).

The effective stress $\sigma_{oe} = \sigma_0 - \sigma_f$ ($\sigma_0 = \text{initial tectonic stress on the fault plane; } \sigma_f = \text{dynamic friction during faulting}$) is the stress which drives the fault motion (Brune, 1970). The effective stress can be obtained from the frequency spectrum or the rise-time of near-field seismograms. Table 2 summarizes the results. Although these results are subject to large uncertainty, it is important that $\sigma_{oe}$ is about the same order of magnitude as the stress drop.

(3) Small Earthquakes

Figure 3 shows the relation between $\log r$ ($r$: source dimension) and $\log M_0$ (Hanks, 1977). Although the trend is similar to that for large earthquakes, the stress drop varies over a larger range (0.5 to 100 bars) than for large earthquakes. Whether this large variation
is due to real variation of the stress drop or experimental uncertainty is not clear. For very small earthquakes, the source dimension is estimated mainly from the corner frequency and the uncertainty of this measurement is very difficult to estimate.

For several earthquakes, the duration of the source time function has been determined from time-domain analyses (Figure 4, Helmberger and Johnson, 1977). These results again indicate a stress drop of 10 to 100 bars. In some cases, a very large stress drop (1 kbar or larger) has been reported (e.g., Brune et al., 1976). Although the absolute values of the stress drop are subject to large uncertainty due to the lack of information about the rupture mode and the source dimension, these results indicate a larger range of stress drop for small earthquakes than for large earthquakes.

3. MULTIPLE SHOCKS

Many seismograms indicate that earthquake fault motion is extremely complex. This complexity exists at all scales. Figure 5 shows an example of the strong-motion seismogram of the 1971 San Fernando earthquake recorded at Pacoima Dam. The "displacement" trace shows the ground displacement recorded by an instrument whose response is given by curve 3. The displacement is relatively smooth and various theoretical methods can be used to explain this trace. The stress drop has been estimated by various analyses (Mikumo, 1973; Trifunac, 1974; Hanks, 1974). The "acceleration" trace shows the ground displacement recorded by an instrument with a response shown by curve 1. The displacement in this high frequency range is extremely complex, and simple theoretical
models fail to explain this irregularity.

Figure 6 shows an irregularity at a larger scale. It is obvious that very complex source models are necessary to explain these complex events. However, estimates of the average stress drop and effective stress have been obtained using simple dislocation models or simple crack models. These models can explain the long-period component of the seismograms but fail to explain the short-period component. How meaningful is the estimate of the stress drop and other source parameters obtained by using these incomplete models? Madariaga (1977) showed that the estimate of the average stress drop depends upon the distribution of the stress drop on the fault plane. However, it is probably unlikely that the estimate of the average stress drop obtained for earthquakes using a simple model is in error by a factor of five or so.

A more detailed analysis was made for the 1976 Guatemala earthquake (Kanamori and Stewart, 1978). Figure 7 shows wave forms of P waves at seven stations which exhibit remarkable complexity. These wave forms were matched by synthetic seismograms computed for a sequence of point sources (Figure 7). The resulting source time sequence is shown by Figure 8. This result suggests that the earthquake can be represented by a sequence of approximately ten distinct events, the seismic moment of which varies by a factor of about four. The rupture can be represented by a stop-and-go sequence with an average rupture velocity of 2 km/sec. The spatial separation of the individual events is 14 to 40 km suggesting that either stress, frictional characteristics or sliding characteristics on the fault plane vary with comparable
spatial scale along the fault plane. Although the average stress drop is about 30 bars, the local stress drop for the individual events may be significantly higher than this value, perhaps by a factor of two or three. Other multiple shocks which were studied earlier include the 1923 Kanto earthquake (Imamura, 1937), the 1964 Alaskan earthquake, (Wyss and Brune, 1967), and the 1940 Imperial earthquake (Trifunac and Brune, 1970). These multiple shocks will provide important clues to the understanding of the mechanics of faulting.

4. ASPERITY

The multiple shocks can be interpreted in terms of asperities on the fault plane. Here the asperities can be geometrical asperities, heterogeneities of the frictional strength or a combination. A fault plane can probably be represented by a random distribution of stress concentrations of various scale lengths. This asperity model can be used to interpret seismicity patterns before large earthquakes.

Several investigators (e.g., Mogi, 1968b; Kelleher and Savino, et al., 1975), found that foreshocks tend to cluster near the epicenter of the main shock (Figure 9). For the 1971 San Fernando earthquake, Ishida and Kanamori (1977) found a clustering of activity for a two-year period before the main shock (Figure 10). As shown in Figure 11, these events show nearly identical wave forms at Pasadena indicating that they originated nearly at the same hypocenter. As shown by Figure 10 the distribution of small earthquakes in the epicentral area prior to 1965 was relatively random. During the period from 1965 to 1968, the seismic activity was very low in the epicentral area.
These observations may be interpreted in terms of the asperity model. The distribution of asperities is initially uniform. As the tectonic stress builds up, weak asperities break in sequence resulting in small earthquakes distributed over the fault plane. As the weak asperities break, stress concentrations occur near the stronger asperities and eventually near the strongest one. This redistribution of stress results in clustering of earthquakes near the strongest asperity, the hypocenter of the impending main shock, and relative quiescence elsewhere on the fault plane. This stage corresponds to the "foreshock" activity. When the last asperity breaks, the entire fault plane ruptures resulting in the main shock. In this case, the stress drop in the beginning of the faulting process is substantially higher than the average stress. Hanks (1974) suggested that the 1971 San Fernando earthquake was initiated by an event with a very high stress drop.

Under certain conditions, a failure of one asperity may load up the neighboring asperities and cause failure resulting in a swarm-type earthquake activity.

From the point of view of earthquake prediction, it is important to distinguish foreshock activity from swarm activity or background activity. At the present time no established method exists, but the result for the San Fernando earthquake is encouraging in that the events just before the main shock are very distinct from the earlier events in terms of both wave form and clustering characteristics. Detailed study of temporal variations of wave forms, spectra, source mechanisms, and locations of small earthquakes may be very important for identifying foreshocks.
5. LOW EARTHQUAKE STRESS DROP VERSUS HIGH FRICTIONAL STRENGTH OF ROCKS

Byerlee (1977) conclude that, under laboratory conditions, the coefficient of friction does not depend on mineralogy, pressure, temperature, and texture of the sample. The coefficient of friction was found to be $0.6 \pm 0.05$ (Figure 12). Under mid-crustal conditions, the normal stress is about 5 kbars so that a frictional strength of about 3 kbars is suggested unless the pore pressure is very large. If the pore pressure is very large, the frictional strength may be about the same order of magnitude as the earthquake stress drop. In this case, earthquakes represent a complete release of the tectonic stress.

On the other hand, if the pore pressure is small compared with the lithostatic pressure, the frictional stress is nearly two orders of magnitude larger than the stress drops in large earthquakes (see Figure 2). In this case the stress drop in earthquakes is only partial (about 1%), and the frictional stress during faulting must be very high. As a result, high heat flow may be expected along the fault zone. The lack of high heat flow along the San Andreas fault has been used as evidence against this idea (Brune et al., 1969). However, if the frictional heat is transferred by mechanisms other than conduction, the lack of a heat flow anomaly may not be compelling evidence against high frictional stress on the fault (Hanks, 1977). Since the earthquake stress drop is very uniform (about 30 to 60 bars), a mechanism which provides a uniform fractional stress drop is necessary. One such mechanism may be suggested from the results of friction experiments (J. Rudnicki, personal communication, 1977; see Figure 12). At high
pressures the shear stress (here interpreted as the frictional stress) fluctuates very little about the average value. This small fluctuation is due to difference in the mineralogy, grain size and texture of rocks. Since the fault zone is nonuniform in composition, a small amount of fluctuation in the frictional strength \( \Delta \sigma = \sigma_M - \sigma_m \) would be expected where \( \sigma_M \) and \( \sigma_m \) are the maximum and the minimum strengths respectively. In this case, when the tectonic stress exceeds the maximum frictional stress \( \sigma_M \), a sudden failure takes place and the stress on the fault plane drops. When the stress on the fault plane drops to \( \sigma_m \) then the fault is locked again. Then the average stress drop would be of the order of \( \Delta \sigma \). In this model it is the range of the frictional stress that controls the stress drop in earthquakes. Since experimental data suggest that, at pressures corresponding to the mid-crustal depth, the fluctuation of the frictional strength is very small, only a few percent of the frictional strength itself, the stress drop in earthquakes can be a very small fraction of the frictional strength. This model suggests that, if the fault zone is completely homogeneous, the stress drop is zero and stable sliding rather than earthquakes occurs. In the above discussion, dynamic loading effects and the stiffness of the crust are ignored for simplicity. If these effects are included the process would become more complex.
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Fig. 1
$S, \text{ km}^2$

$M_0 = 1.23 \times 10^{22} \frac{3}{S^2} \text{ dyne-cm.} \quad (S \text{ in km}^2)$

**Fig. 2**
Fig. 3
Fig. 4
SAN FERNANDO EARTHQUAKE
2/9/71 06:00 PST PACOIMA DAM, CALIFORNIA
DOWN COMPONENT

GROUND ACCELERATION

GROUND VELOCITY

GROUND DISPLACEMENT

SECONDS

Response

Period, sec

Fig. 5
San Fernando Earthquake $M=6.6$ STJ ($\Delta=49.9^\circ$) LPZ

Kurile Is. Earthquake $M=8.0$ OXE ($\Delta=84.5^\circ$) LPE

Fig. 6
Multiple Event Analysis

Source time function

\[ s(t) = \sum_{i=1}^{N} m_i \cdot s(t-t_i) \]

minimum

NUR (Δ=88.1°)

Obs

Syn

1.6 x 10^{26}
dyne-cm

COP (Δ=83.97°)

Obs

Syn

Source

Time Series

1 min

1.6 x 10^{26}
dyne-cm

LPB (Δ=37.9°)

Obs

Syn

1.8 x 10^{26}
dyne-cm

KEV (Δ=84.5°)

Obs

Syn

2.5 x 10^{26}

KJF (Δ=87.5°)

Obs

Syn

2.2 x 10^{26}

KRG (Δ=68.2°)

Obs

Syn

2.1 x 10^{26}

STU (Δ=84.2°)

Obs

Syn

3.0 x 10^{26}

Fig. 7
Fig. 8
Fig. 9
Fig. 10

1969, Apr. 2.8
July 2.5
Aug. 2.6
1970, Mar. 2.5
Sep. 2.6

EW, Period II (1961~1964)

ML

2.4
2.9
2.5
2.8
2.9
2.1
2.3
2.6
2.6
2.3
2.4
2.6
2.2
2.5
2.3
1.9

Fig. 11a, b
Wood-Anderson Seismogram at Pasadena

EW, L ≤ 10 km

Period IV

1969 Jul.
1969 Aug.
1970 Mar.
1970 Sep.

Depth = 10 km

Fig: 11c
FRICION MEASURED AT MAXIMUM STRESS

EXPLANATION

SYMBOL REFERENCE  ROCK TYPE

* 2F  Granite, fractured
  2G  Granite, ground surface
  3  Limestone, Gabbro, Dunite
  5  granite, ground surface
  6F  Weber Sandstone, faulted
  6S  Weber Sandstone, saw cut
  9  Granodiorite
  13 Gneiss and Mylonite
  16 Plaster in joint of Quartz Monzonite
  20 Quartz Monzonite joints
  25 Westerly Granite, Chlorite, Serpentinite,
     Illite, Kaolinite, Halloysite,
     Montmorillonite, Vermiculite
  26 Granite
  27 Kaolinite, Halloysite, Illite,
     Montmorillonite, Vermiculite

\[ \tau = 0.5 \sigma_N \]

\( \tau = 0.65 \sigma_N \)

Fig. 12