

EARTHQUAKE RECURRENCE INTERVALS AND PLATE TECTONICS

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ABSTRACT

The frequency of occurrence of earthquakes with different seismic moments is expressed in terms of the rate of slip on a fault and to the largest seismic moment likely to occur in the region. Beginning from the Gutenberg-Richter empirical expression relating the relative recurrence of events with different magnitudes and using another empirical relation between magnitude and seismic moment, the relative number of events with seismic moment greater than or equal to M_0 is given by $N(M_0) = \alpha M_0^{-\beta}$. β can be determined from parameters in these empirical expressions. From average rates of slip on faults, this expression can be used to give the recurrence rates for events of different seismic moment: $\alpha = (1 - \beta) \dot{M}_0^\Sigma / M_0^{\max 1 - \beta}$ where M_0^{\max} is the maximum possible seismic moment in a region, and \dot{M}_0^Σ is the average rate of occurrence of the seismic moment. On a fault of area A , with shear modulus μ , and long-term average rate of slip \bar{v} , $\dot{M}_0^\Sigma = \mu A \bar{v}$. A similar expression can be given for α for regions where deformation is distributed over a broad area without a major throughgoing fault. Using rates of convergence at island arcs determined from plate motions for the last 5 m.y., the calculated frequency of occurrence of earthquakes with large seismic moments agrees well with the historic record. At present, uncertainties in the requisite parameters and in assumptions on which the recurrence relation is based, however, make such an approach only marginally better than reliance on the historic record alone. With more data constraining β , the largest possible seismic moments, the role of fault creep, and long-term rates of slip or deformation, however, this approach to seismic risk should be a more reliable predictor of recurrence rates than the existing historic record. As an example, the formulas are applied to the southeastern Caribbean.

INTRODUCTION

Most studies of seismic risk begin with an estimate of the frequency of occurrence of earthquakes with different magnitudes in the region of interest. This estimate is often based on the historic record of seismicity. Most seismologists agree, however, that in general the seismic record is too short to evaluate the frequency of occurrence with any confidence. The variability in seismic activity in regions with long historic records, such as Turkey (Ambraseys, 1971) and Northeast China (Lee *et al.*, 1976) implies that recurrence intervals are often several hundred years. Moreover, in some regions, such as Turkey and China, large earthquakes often occur in clusters instead of being more uniformly spaced in time as they do, for instance, in southwest Japan (Imamura, 1937). To establish the length of such recurrence intervals with confidence requires data spanning longer times: 10^3 yr for most regions and for some, 10^4 yr. If this is so, then methods of estimating seismic risk from seismicity alone will not be reliable for a long time in the future. Or, from another point of view, we must wait a long time to test any proposed method.

It logically follows that other methods for evaluating seismicity must be sought. Careful studies of active faulting during the Quaternary can crudely extend the historic record thousands of years. Although such data are limited by inaccuracies in dating and difficulties in estimating offsets or lengths of faults during prehistoric events, several such studies have been used to constrain the frequency of occurrence

of events in selected regions (e.g., Solonenko *et al.*, 1966, 1968; Plafker and Rubin, 1967; Plafker, 1969; Clark *et al.* 1972; Allen, 1975; Sieh, 1978). Plate tectonics offers another constraint, the average rate of slip on faults. We can be quite certain that slip will continue on plate boundaries and therefore that earthquakes will continue to occur on the major faults. By associating amounts of displacements with earthquakes of different sizes, this rate of slip can be used as a constraint on the frequency of earthquakes of different sizes. Wallace (1970) discussed such a constraint, but his formula implicitly assumes that each point on the fault slips only during earthquakes of one magnitude. It does not include the relative frequency of occurrence of earthquakes with different magnitudes. The purpose of this study is to develop a formalism for determining the frequency of occurrence of earthquakes that uses information about the relative frequency of occurrence and the average rate of slip as well as more general geometries than simply a single fault.

The formulation given here will not reliably predict the times of occurrence of earthquakes, if only because of the clustering that occurs in some regions. It is possible that this formulation could be used in conjunction with Rikitake's (1974, 1975, 1976) approach of parameterizing earthquake occurrence in terms of ultimate strain. To exploit such an approach one must estimate the ultimate strain, which surely varies regionally. Moreover, some knowledge of the relative frequency of occurrence of smaller events and of the amount of aseismic deformation must be included. Although the present paper considers only the mean recurrence intervals and does not deal with this problem, probably it could be extended to do so.

THE SEISMIC MOMENT

It is useful to describe the "size" of earthquakes in terms of the seismic moment instead of magnitude. The seismic moment: $M_0 = \mu A \bar{u}$, where μ is the shear modulus, A is the fault area, and \bar{u} is the spatial average slip on the fault during the earthquake, is the fundamental parameter that describes the static aspects of earthquakes. Aki (1966, 1967) showed that the long-period spectral density of body and surface waves is proportional to M_0 , and subsequently Ben Menahem *et al.* (1969) showed the far-field static-strain field also is proportional to M_0 . Because the amplitude of ground motion is physically and directly related to the seismic moment, quantitative estimates of seismic risk that include a deterministic prediction of ground motion will include the seismic moment as a parameter (e.g., Hanks, 1976; Heaton and Helmberger, 1978).

For seismicity, an important aspect of the seismic moment is that a summation of the moments for individual earthquakes occurring in a finite time interval can be used to estimate the rate of slip on a fault (Brune, 1968). Consider a fault with length L , and width W , on which slip occurs by earthquakes. After a sufficiently long period of time earthquakes will have occurred along all segments of the fault, and will have caused an average displacement \bar{U} . Brune (1968) showed that because we can estimate quite well the value of the shear modulus and if we can estimate the area of the fault $A = LW$ that slips during earthquakes, we can calculate the average slip during an interval from the sum of the seismic moments of the earthquakes that occur during the interval

$$\bar{U} = \sum_{x=1}^N \bar{u}(x) = \sum_{x=1}^N M_0(x) / \mu A. \quad (1)$$

In reality, some slip occurs by fault creep and not by earthquakes and the relative proportion of fault creep and seismic slip varies with depth. Therefore the seismic slip will not equal the total slip, and the depth (or width) of the fault cannot be chosen without some arbitrariness. These limitations are discussed below. Nevertheless, if the relative number of earthquakes with different seismic moments is known and if the rate of slip is known, then the frequency of occurrence of earthquakes with various seismic moments can be derived from equation (1). Moreover, equation (1) can be generalized for broader zones of deformation, in which there are many faults of different orientations, by expressing the moment as a tensor instead of as a scalar (see Appendix A).

RECURRENCE RELATION FOR THE SEISMIC MOMENT

Gutenberg and Richter (1954) and numerous subsequent studies showed that the number, $N(M)$, of earthquakes with magnitudes greater than or equal to a given value, M , obeys a relation of the form

$$\log N(M) = a - bM. \quad (2)$$

A similar relation for seismic moments can be obtained by using another empirical expression

$$\log M_0 = cM + d \quad (3)$$

(e.g., Wyss and Brune, 1968; Thatcher and Hanks, 1973; Kanamori and Anderson, 1975; Ríznichenko, 1976; Kanamori, 1977a, 1978). Combining these, yields

$$\log N(M_0) = a + \frac{bd}{c} - \frac{b}{c} \log M_0$$

where $N(M_0)$ is now the number of events with seismic moment greater than or equal to M_0 . This can be written as

$$N(M_0) = \alpha M_0^{-\beta} \quad (4)$$

where

$$\alpha = 10^{\left(a + \frac{bd}{c}\right)} \quad \text{and} \quad \beta = b/c.$$

If $N(M_0)$ is the number of events/year, then α has units of $(\text{dyn-cm})^\beta/\text{yr}$.

For each region in the Earth, there seems to be a maximum magnitude, or maximum seismic moment above which there are no earthquakes. Therefore, for $M_0 > M_0^{\max}$, $N(M_0) = 0$, and equation (4) can be written more precisely

$$N(M_0) = \alpha M_0^{-\beta} [1 - H(M_0 - M_0^{\max})] \quad (5)$$

where $H(x)$ is a Heaviside function.

In equations (2) and (3), the constants b , c , and d are quite well determined and uncertainties can be attached to them. Although there is some regional variation in

each of them, regional variations in seismicity are described primarily by the value of α . Because of the short time sample for establishing the level of seismicity, the uncertainty in the value of α is probably large and difficult to estimate. Therefore, α is uncertain also. Our purpose here is to use assumed or measured rates of slip to estimate α .

To do this it helps to describe the recurrence relation [equation (5)] alternatively in terms of a density function $n(M_0)$, defined by

$$N(M_0) = \int_{M_0}^{\infty} n(M_0) dM_0$$

or

$$n(M_0) = -\frac{dN}{dM_0},$$

where $n(M_0)$ has dimensions of $(\text{dyn-cm} \times \text{year})^{-1}$. From equation (5),

$$n(M_0) = \alpha \beta M_0^{-\beta-1} [1 - H(M_0 - M_0^{\max})] + \alpha M_0^{\max-\beta} \delta(M_0^{\max} - M_0) \quad (6)$$

where $\delta(x)$ is the Dirac δ function. The function $n(M_0)$ is used to integrate the seismic moments of all events to obtain the total seismic moment.

THE SEISMIC MOMENT RATE AS A CONSTRAINT

When a sufficiently long period of time τ is considered, the total accumulated seismic moment M_0^{Σ} on a fault can be estimated simply from $M_0^{\Sigma} = \dot{M}_0^{\Sigma} \tau = \bar{\mu} A \bar{v} \tau$, where \dot{M}_0^{Σ} is the rate of occurrence of seismic moments and \bar{v} is the average slip rate on the fault excluding fault creep (Brune, 1968; Davies and Brune, 1971; Hanks *et al.*, 1975). The rate of occurrence of seismic moments should also be equal to

$$\dot{M}_0^{\Sigma} = \int_0^{\infty} M_0 n(M_0) dM_0.$$

From equation (6), provided $\beta < 1$, we have

$$\dot{M}_0^{\Sigma} = \frac{\alpha}{1-\beta} M_0^{\max 1-\beta}. \quad (7)$$

This expression is wrong if $\beta = 1$. Discussion of this case is given in Appendix C. Empirical data, as well as theoretical arguments indicate that $\beta \approx \frac{2}{3}$ (Wyss and Brune, 1968; Thatcher and Hanks, 1973; Kanamori and Anderson, 1975; Riznichenko, 1976; Kanamori, 1977a) except possibly for very small events (Kanamori and Anderson, 1975).

We can write equation (7) as

$$\alpha = \frac{(1-\beta)\dot{M}_0^{\Sigma}}{M_0^{\max 1-\beta}} \quad (8)$$

and combining with equation (4),

$$N(M_0) = (1 - \beta) \frac{\dot{M}_0^\Sigma}{M_0^{\max^{1-\beta}}} M_0^{-\beta}. \quad (9)$$

Let $T(M_0)$ be the average return period, or recurrence interval for events with moments greater than or equal to M_0 (but not greater than M_0^{\max}). Then $T(M_0)$ is the reciprocal of $N(M_0)$

$$T(M_0) = \frac{M_0^{\max^{1-\beta}}}{(1 - \beta) \dot{M}_0^\Sigma} M_0^\beta. \quad (10)$$

Equations (9) and (10) express the recurrence relationships for earthquakes occurring on major fault zones, such as plate boundaries. Similar expressions can be derived for earthquakes in regions with diffuse seismicity and more complicated deformation. Appendix B gives an example for a particular geometry and style of deformation.

Finally, it is possible to convert these expressions from moment to magnitude using Kanamori's (1977a, 1978) formula

$$M = \frac{\log M_0}{1.5} - 10.7. \quad (11)$$

SOME ILLUSTRATIVE EXAMPLES

There probably is no region on the Earth where the historic record of seismicity is known well enough that it can be used as a conclusive test of any method for predicting long-term seismicity. At the same time, the length of time needed to obtain a meaningful measure of the average seismicity is probably less for a long zone of relatively uniform seismicity than that for a short one. With this in mind, two simple examples are considered: the slip on the main underthrust zone at island arcs and the normal faulting presumably due to flexure of the downgoing slab of lithosphere at the trenches in the Pacific and Indian Oceans (Figure 1).

Subduction occurs along trenches in the Pacific and Indian oceans that are approximately 33,000 km long, at an average rate of about 7.5 cm/yr (Forsyth and Uyeda, 1975). Because most of these earthquakes occur in the mantle, let us assume that $\mu = 7 \times 10^{11}$ dynes/cm². If the fault width is 100 km, (Davies and Brune, 1978; Kanamori, 1977b), then $\dot{M}_0^\Sigma = 1.7 \times 10^{29}$ dyne = cm/yr. The largest moment that has been measured is 2×10^{30} dyne-cm, obtained by Kanamori and Cipar (1974) for the 1960 Chilean earthquake. From Kanamori's (1977a) summary of the seismic moments for the largest events in this century, which is essentially complete for the last 50 yr, the Chilean earthquake is the only one with $M_0 > 10^{30}$ dynes-cm. Five earthquakes with $M_0 \geq 2 \times 10^{29}$ dyne-cm, including the Chilean earthquake, occurred during this period at subduction zones in the Pacific and Indian Oceans, and there were 17 events with $M_0 > 2 \times 10^{28}$ dynes-cm. Assuming that $M_0^{\max} = 2 \times 10^{30}$ dyne-cm and that $\beta = \frac{2}{3}$, equations (9) and (10) predict recurrence intervals of 35 yr (for $M_0 \sim 2 \times 10^{30}$ dyne-cm), 7.6 yr (for $M_0 \geq 2 \times 10^{29}$ dyne-cm), and 1.6 yr (for $M_0 \geq 2 \times 10^{28}$ dyne-cm) or 1.4, 6.6, and 30.8 events (with moments greater than 2×10^{30} , 2×10^{29} , and 2×10^{28} dyne-cm, respectively) per 50-yr period. The uncertainties in

the parameters used to estimate the frequency of occurrence are much larger than is implied by the agreement with the observed values tabulated in Kanamori's (1977a) summary. Nevertheless, the agreement is reassuring that a gross error has not been made. A discussion of these uncertainties is given below, after considering the case of the flexure of the plate.

Before being subducted, the oceanic lithosphere is bent down at island arcs to form a trench (Figure 1). Earthquakes often occur in the trench, and fault-plane solutions typically indicate normal faulting (Stauder, 1968a and b). Simple bending of the plate implies a strain of about 1 to 2 per cent at the outer surface. Forsyth and Chapple (1978) inferred that earthquakes with normal faulting at deep sea trenches occur at depths shallower than 25 km. Let us assume this value for the thickness of the region that is strained. If the lithosphere over a distance of 150 km seaward of the trench is stretched due to the flexure, then for an average convergence rate of 7.5 cm/yr, each part of the lithosphere is strained for 2 m.y. If a strain of 1

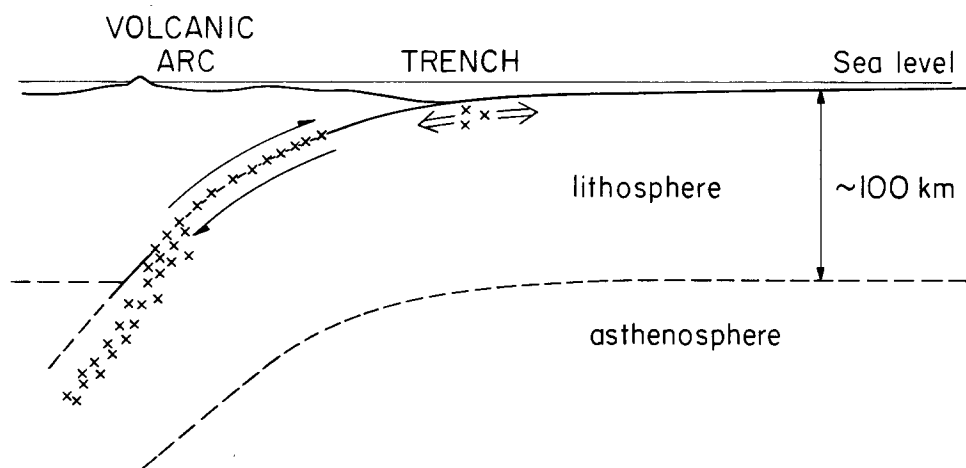


FIG. 1. Shallow cross section of subduction zone. Oceanic lithosphere bends down at trench and plunges into asthenosphere beneath volcanic arc. Underthrusting of one plate beneath the other occurs on the shallow part of inclined seismic zone. Most large shallow events occur at such zones. The zone continues to greater depths, with generally smaller events occurring within the downgoing slab of lithosphere. The transition between shallow interplate and intermediate depth intraplate events is not well defined. At very shallow depths, the main underthrust zone is usually aseismic, probably because of weak sediments of the overlying plate. Beneath the trench another zone of shallow earthquakes results from a stretching of the upper part of the lithosphere. Normal faulting prevails for these events.

per cent eventually accumulates during this time, the strain rate is 0.5 per cent/m.y. and the rate of stretching is 7.5×10^{-2} cm/yr. From equation (B1) in Appendix B, $\dot{M}_0^S = 2 \times 7 \times 10^{11} \text{ dyne-cm}^{-2} \times 33,000 \times 10^5 \text{ cm} \times 25 \times 10^5 \text{ cm} \times 7.5 \times 10^{-2} \text{ cm/yr} = 8.8 \times 10^{26} \text{ dyne-cm}$. The largest such event known is the August 1977 Indonesian earthquake, for which $M_0 \approx 1 \times 10^{29} \text{ dyne-cm}$ (Stewart, 1978). From equation (10), we would expect one such event every 341 yr. Only 44 yr earlier, in 1933, another event, the Sanriku earthquake, occurred with a comparable seismic moment, ($M_0 \approx 4.3 \times 10^{28} \text{ dyne-cm}$) (Kanamori, 1971).

This poor agreement between the predicted and observed frequency of occurrence could be the result of statistical variation in a short time sample, it could mean that the assumptions used to derive equation (B1) are not applicable to this case, or it could indicate that the values of the parameters used in equations (10) and (B2) are erroneous. Without waiting another 300 yr, it is difficult to decide whether or not the historic record is too short. As the fault-plane solutions of these two large events

are similar to what is assumed for equation (B1), the form of equation (B1) probably is reasonable. It is possible that the parameters in equation (B1) are in error. Both Kanamori (1971) and Stewart (1978) inferred that the ruptures during these two events extended to depths of 75 km or more. If we apply the same formula, assuming that a layer of thickness 75 km instead of 25 km were strained, calculated recurrence intervals would be three times less. In fact, Kanamori inferred that such earthquakes do not arise from bending of the plates but, instead, because of a pull applied by the downgoing slab of lithosphere. Therefore, the rate of stretching of the lithosphere might be greater than that assumed for flexure of the plate alone. If such large events are as common in trenches as the two in the last 50 yr suggest, then they cannot be due to flexure of the lithosphere alone (see also Hanks, 1978). Two large events are not enough to establish recurrence intervals with confidence, and our ignorance both of the historic record and the processes taking place make it difficult to account for the disagreement between the calculated and observed recurrence intervals.

A DISCUSSION OF UNCERTAINTIES

The uncertainties can be divided into two groups: those that reflect inaccuracies in measured parameters and therefore can be assigned quantitative values and those arising from unproven assumptions about processes or phenomena and therefore in some cases are difficult to quantify.

In principle, the value of β in equation (4) can be determined experimentally, although in fact this has not been done yet. As $\beta = b/c$, defined in equations (2) and (3), we can estimate its uncertainty. Measurements from a few regions and for very large events yield $c \approx 1.5 - 1.8$ (Wyss and Brune, 1968; Thatcher and Hanks, 1973; Kanamori and Anderson, 1975; Riznichenko, 1976) and simple physical arguments have been used to rationalize $c = 1.5$, (Kanamori and Anderson, 1975; Hanks, 1977; Kanamori, 1977a, 1978). Most studies give b in the range of $0.8 < b < 1.2$, with the majority ≤ 1 (e.g., Page, 1968; Wyss, 1973). Values of b as small as 0.5 are sometimes observed but rarely for large areas or for long periods of time. Thus, $\beta \approx \frac{2}{3}$ but could be as small as 0.5 or, less likely, as large as 0.75.

From equation (9), the frequency of occurrence of the largest events will be 50 per cent greater if $\beta = 0.5$ instead of $\frac{2}{3}$. For smaller events such as with $M_0 \geq 10^{-3} M_0^{\max}$ (and therefore with $M \geq M^{\max} - 2$); however, if $\beta = 0.67$ there will be 33 events in a given time interval, but if $\beta = 0.5$, only 16 events will occur in the same interval. Differences in the predicted levels of seismicity of this amount are quite small with the short historic record. The uncertainty in β is probably not our most serious problem, especially since β ultimately can be measured directly.

A much more serious problem is the uncertainty in α in equation (4) which is a result of ignorance of the parameters in equation (8), from which it is determined. Let us consider the parts individually. The moment rate, \dot{M}_0^2 , depends upon the shear modulus, the fault length, the fault width, and the rate of slip. Not only is the shear modulus known relatively well, but because it will be assumed to be the same for all events, it cancels out of equations (9) and (10) by entering into both numerator and denominator.

If we are concerned only with a finite length of a fault, or a finite area of a fault system, either of which exceeds the dimensions of the largest possible event, this length is chosen and known as accurately as desired.

The rates of slip for major plate boundaries are known quite well (± 10 to 20 per cent) (e.g., Minster *et al.*, 1974; Minster and Jordan, 1978), assuming that the rates

for the last few million years apply to the present. For broader zones, the average strain rate or displacement rate across them is less well known, but one of the motivations for this study stems from the conviction that accumulating data from neotectonic studies will eventually constrain the rates better.

It is much more difficult to estimate the fault width because, at sufficiently great depth, slip occurs purely by fault creep, and the transition from purely fault creep to largely seismic slip and brittle failure is not sharp. Nevertheless, the uncertainty in the width of the fault zone, although large, may not be very important. Because large events rupture through the entire lithosphere, the fault width, like the shear modulus, will cancel in the numerators and denominators of equations (9) and (10).

To use equations (9) and (10) one must assume a value for the largest possible seismic moment. Although making a choice may seem rather arbitrary, in practice it probably can be done with an uncertainty less than a factor of five. As noted above, the value of the shear modulus and the fault width will cancel with those chosen for the moment rate. Thus, the important parameters are the fault length and average displacement. From the known rupture lengths in different tectonic settings, the maximum fault lengths probably can be estimated with an uncertainty less than a factor of two. At most island arcs, maximum rupture lengths are between 300 and 1000 km (Kelleher *et al.*, 1973). Central America and parts of South America apparently are exceptions, where rupture lengths rarely exceed 200 km (Kelleher *et al.*, 1973). For major strike-slip faults, fault lengths rarely exceed 300 km (e.g., Richter, 1958). In more complicated regions, where long faults do not exist but there are many short faults (such as within Japan or the basin and range province), the maximum fault length will be shorter yet but probably can be estimated from maps of Quaternary faults or from data for historic events.

The maximum average displacement also probably can be estimated within a factor of two. Data from a wide range of moments (and magnitudes) imply that average stress drops on faults during earthquakes are nearly always between 1 and 100 bars (Hanks, 1977). For the largest events, stress drops are typically about 30 bars (Abe, 1975; Kanamori, 1977a, 1978). This observation can be used to estimate the average slip on faults (Kanamori, 1977a; 1978).

For instance, for strike-slip faults, the stress drops are given by (Knopoff, 1958)

$$\Delta\sigma = \frac{1}{2} \mu \frac{u}{W}.$$

If $\Delta\sigma = 30$ bars $= 3 \times 10^7$ dyne cm^{-2} , $\mu = 3.3 \times 10^{11}$ dyne/ cm^2 (appropriate for crustal rock), and $W = 20$ km, then $u \sim 4$ m. For thrust faults, common at island arcs, from Starr (1928)

$$\Delta\sigma = \frac{4}{3} \mu \frac{u}{d}.$$

Such events probably occur in the mantle where $\mu = 7 \times 10^{11}$ dyne cm^{-2} . For $\Delta\sigma = 30$ bars and $W = 120$ km, $u \sim 4$ m. Larger displacements seem to occur where fault widths are greater (Kelleher *et al.*, 1974). Thus, the maximum average displacement probably can be predicted well.

It is more difficult to quantify our ignorance of the tectonic processes that are acting. It has been assumed implicitly, above, that the tectonics of the region under consideration are well known. Yet many phenomena, about which we are unaware or unsure, may occur. The example of the bending of the lithosphere, discussed

above, illustrates this problem. Unrecognized active faults and unrecognized styles of deformation are likely to exist in all regions. Plate boundaries are not single, infinitesimally narrow fault zones. Even near plate boundaries, some intra-plate earthquakes occur without contributing significantly to the long-term slip rate, and as yet without an obvious pattern. It is difficult to assign quantitative values to such uncertainties.

There are other uncertainties which can be crudely quantified. One concerns the applicability of equations (2) or (4) which describe the frequency of occurrence of events. It has been implicitly assumed above that the Gutenberg-Richter frequency magnitude relation [equation (2)] applies to every segment of the fault, so that slip at every point on a fault occurs during earthquakes of all magnitudes. It is possible that at least along some segments of a fault, slip occurs only by very large events and along neighboring segments it occurs during smaller events. For this case, equation (2) would not apply locally to each segment of the fault, but only as an average over the entire fault. In terms of equations (9) and (10), if M_0^{\max} is the maximum moment for earthquakes along the fault zone and if $\beta = \frac{2}{3}$, events with $M_0 \sim M_0^{\max}$ will account for $\frac{1}{3}$ of the slip, but these events will occur only in certain segments of the fault. They will be three times more common in these segments than they would be if smaller events occurred in these segments also. Such a situation might exist for the San Andreas fault (Allen, 1968; Sieh, 1978), where slip seems to occur during major earthquakes along some segments but possibly occurs only by smaller events and fault creep along others. In principle, if one knows that different segments fail with different maximum moments, the fault could be divided up into segments with different maximum moments assigned to each. Unfortunately we do not know enough about most, if any, fault zones to do this with confidence, and an erroneous assumption leads to an error of a factor of about three in the return period for some segments.

Another major uncertainty concerns the role of fault creep. What fraction of slip on faults occurs by fault creep is difficult to establish, but along some segments of faults the seismic slip has been only a very small fraction of the total slip since measurements of fault creep and seismicity began (e.g., Wesson *et al.*, 1973). Once again, presumably with careful study, it is possible to establish what fractions of slip occur either by fault creep or by earthquakes, but at the present time our ignorance leads to an uncertainty that can be quantified only crudely.

Finally, there exists the possibility that during the brief instant of geological time while we live, major changes in the tectonics are occurring and that average rates of plate motion for the last 3 to 5 m.y. will not apply to the next thousand years. The likelihood that such a change is occurring in a particular region is surely small, but difficult to estimate. Presumably, careful geological (neotectonic) work in regions of active faulting can establish the history of faulting for the last few hundreds to tens of thousands of years. If rates of motion inferred from plate motions and from geological studies are applicable to the last 10^4 yr, then the likelihood of a change in rate during our lifetime surely is negligible. Therefore, careful neotectonic studies offer the hope of resolving uncertainties concerning this question.

This discussion of uncertainties is intended to give the reader two impressions. The first is pessimistic: at the present time we cannot know the long-term average seismicity of a region with much confidence. The uncertainty in the frequency of occurrence of events with $M > 6$ in a region is probably at least a factor of three and probably five. Strict application of the formulas discussed above (or especially of the historic record) should not ignore the large uncertainties.

The second intended impression, however, is optimistic. Careful seismic and

geological studies should be able to reduce the uncertainties substantially without the need for waiting for several cycles of recurrence of major earthquakes in any region. Seismic studies should allow β to be estimated and further careful analysis of major earthquakes will help constrain M_0^{\max} in various regions. Geological work on active faults has demonstrated that approximate rates of movement and frequencies of earthquakes can be established for the last few hundreds to thousands of years. Further work should help resolve how the Earth deforms and hopefully this will eventually be expressible in terms of a tensorial strain rate.

AN APPLICATION TO A PARTICULAR REGION: THE SOUTHEASTERN CARIBBEAN

The Caribbean plate moves east relative to South America at about 2 cm/yr, sliding along a transform boundary in or north of Venezuela and overthrusting the Atlantic at the Lesser Antilles arc (Figure 2) (Molnar and Sykes, 1969). The Atlantic lithosphere bends down in front of the arc before being subducted by it, and there is also a possible continuation of this down-flexed lithosphere beneath Trinidad and northern Venezuela.

For the transform boundary, which may be the El Pilar fault or might be just north of the Venezuelan coast, let us assume that the fault width is 20 km. Let us consider only the easternmost 600 km of it and assume that $\mu = 3.3 \times 10^{11}$ dynes cm^{-2} , appropriate for crustal rock. Therefore, $\dot{M}_0^{\Sigma} = 3.3 \times 10^{11}$ dynes $\text{cm}^{-2} \times 2 \times 10^6$ cm $\times 6 \times 10^7$ cm $\times 2$ cm/yr $= 7.9 \times 10^{25}$ dyne-cm/yr. For a maximum moment, a reasonable fault length is 300 km and a reasonable average displacement is 4 m, values appropriate for the 1906 San Francisco earthquake. Thus, $M_0^{\max} = 7.9 \times 10^{27}$ dyne-cm, or from equation (11), $M^{\max} = 7.9$. If $\beta = \frac{2}{3}$, then from equation (10) the repeat time for such an event is 300 yr. Indeed, one very large event in 1726 did occur in this region in the last 300 yr (Fiedler, 1970; Kelleher *et al.*, 1973), and the area of maximum intensity (IX), was about 300 km long and aligned parallel to the El Pilar fault (J. Grases, personal communication, 1978). In the 300-km segment in the easternmost part of the fault, the repeat time would then be 600 yr. For events with $M_0 \geq 1.1 \times 10^{25}$ dyne-cm ($M \gtrsim 6$), the recurrence interval is calculated to be about 4 yr. Between 1955 and 1967, there seem to have been three such events, all of which occurred at the east end of the zone.

The uncertainty in the repeat times is probably a factor of about five. The greatest uncertainty, which allows a longer but not shorter repeat time, is our ignorance of what fraction of slip occurs by fault creep. None is assumed above.

For the underthrust zone, the length is about 800 km. Let us assume a width of 100 km, and $\mu = 7 \times 10^{11}$ dyne/cm². Thus, $\dot{M}_0^{\Sigma} = 1.1 \times 10^{27}$ dyne-cm/yr. For a maximum moment, let the length be 300 km and the average displacement be 4 m, thus $M_0^{\max} = 8.4 \times 10^{28}$ dyne-cm ($M = 8.6$). Then for such an event, $T = 230$ yr. For earthquakes with $M_0 \geq 1.1 \times 10^{25}$ dyne-cm ($M \gtrsim 6$), the calculated return period is about 0.6 yr. In the last 15 yr there have been only a few such events, and the region seems more quiescent than one would expect. The discussion of uncertainties for the transform segment applies here also.

The length of the flexed zone in front of the arc is also about 800 km. Following the discussion in Appendix B and above, but with a subduction rate of 2 cm/yr, $\dot{M}_0^{\Sigma} = 2 \times 7 \times 10^{11}$ dynes-cm⁻² $\times 25 \times 10^5 \times 8 \times 10^7$ cm² $\times 2 \times 10^{-2}$ cm/yr $= 5.6 \times 10^{24}$ dyne cm/yr. If the maximum moment were to correspond to a fault width of 60 km, a fault length of 200 km and an average displacement of 5 m, then $M_0^{\max} = 4.2 \times 10^{28}$ dyne cm ($M^{\max} = 8.4$), essentially that for the 1933 Sanriku earthquake (Kanamori, 1971). The return period for $\beta = \frac{2}{3}$, is $T = 22,500$ yr. If we consider only

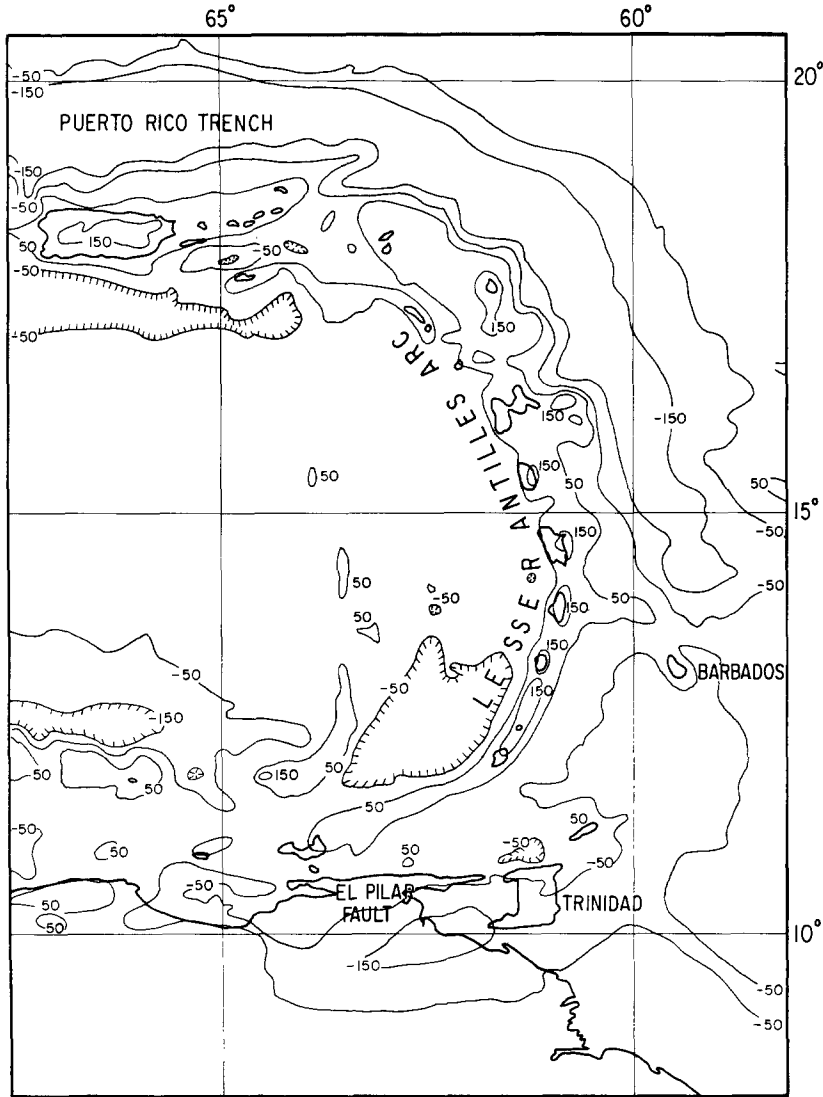


FIG. 2. Free air gravity anomalies of the Caribbean region (Bowin, 1976); contours in mgal. The Atlantic Ocean floor bends down in front of the lesser Antilles arc, causing the large negative gravity anomaly, and underthrusts the arc. The cross section in Figure 1 applies to such a region, except that here the trench is filled with sediments. The Caribbean slides eastward past South America along a right lateral strike-slip fault system (probably the El Pilar fault) and past the west central Atlantic along a left lateral strike-slip fault that dips south beneath Puerto Rico (see Molnar and Sykes, 1969). The continuity of the negative gravity anomaly north of Puerto Rico is presumably due to a downward flexure of the Atlantic lithosphere (Molnar, 1977). The similar gravity anomaly in northeast Venezuela and Trinidad suggests a similar flexure of the Atlantic-South America lithosphere, continuous with that in front of the Lesser Antilles arc.

the 400-km portion that lies within 200 km of Barbados, we would expect one such earthquake to occur every 45,000 yr.

Notice that if the maximum possible moment were smaller by a factor of 100, $M_0^{\max} = 4.2 \times 10^{26}$ dyne-cm ($M_0^{\max} \approx 7.0$), the calculated repeat time for such an event would be 225 yr for the entire arc and 450 yr for the segment within 200 km of Barbados. If, however, the maximum moment were 4.2×10^{28} dyne-cm, then the repeat time for events with $M_0 \geq 4.2 \times 10^{26}$ ($M \geq 7.0$) is 1,040 yr for the whole arc

and 2,080 yr for within 200 km of Barbados. Again the return periods are clearly very uncertain.

In his catalog of events in the Eastern Caribbean for 1953 to 1960, Robson (1964) lists no earthquakes that appear to have occurred beneath the trench axis and that would have been assigned a magnitude greater than 7. Since that time, however, one such event, on December 25, 1969 ($M \approx 7.2$) did occur. Clearly there is too little activity to estimate what the largest possible moment could be or to constrain the frequency of occurrence.

The basement tectonics of northeastern Venezuela and much of Trinidad may be a continuation of those that characterize the Lesser Antilles trench. The large negative free air gravity anomaly associated with the buried trench east of the Lesser Antilles continues beneath Trinidad and into Venezuela for about 200 to 300 km (Follinsbee, 1972; Bowin, 1976). A simple interpretation is that the lithosphere is flexed down and that the basin is filled with young sediments. The magnitude of the gravity anomaly (up to 200 mgal) corresponds to a maximum depression of the crust of up to about 4 to 5 km, which is comparable to that of deep sea trenches elsewhere. Therefore, it seems reasonable to assume that the tectonics are similar to those at deep sea trenches.

TABLE 1
RETURN PERIODS (years) FOR EARTHQUAKES WITH SEISMIC MOMENTS GREATER THAN OR
EQUAL TO M_0^*

M_0^{\max}	M_0				
	1.1×10^{28} ($M \sim 8$)	3.5×10^{26} ($M \sim 7$)	1.1×10^{26} ($M \sim 6$)	3.5×10^{25} ($M \sim 5$)	1.1×10^{24} ($M \sim 4$)
1.1×10^{28}	42,000 yr	4,200	420	42	4.2
3.5×10^{26}	—	1,300	130	13	1.3
1.1×10^{26}	—	—	42	4.2	0.42

* Calculated for different values of M_0^{\max} . $\dot{M}_0^{\Sigma} = 7.9 \times 10^{23}$ dyne-cm/yr. $\beta = \frac{2}{3}$.

Let us assume that earthquakes occur only in the upper 20 km of the crust, where $\mu = 3.3 \times 10^{11}$ dynes cm^{-2} , and that the strain rate is comparable to that for the flexure of the Lesser Antilles trench, above. Therefore, from equation (B1) $\dot{M}_0^{\Sigma} = 2 \times 3.3 \times 10^{11}$ dynes/ $\text{cm}^2 \times 20 \times 10^5$ cm $\times 300 \times 10^5$ cm $\times 1.5 \times 10^5$ cm/7.5 m.y. = 7.9×10^{23} dynes-cm/yr. Table 1 lists return periods calculated for values of $M_0^{\max} = 1.1 \times 10^{28}$ dynes-cm ($M^{\max} \approx 8.0$), 3.5×10^{26} dyne-cm ($M^{\max} \approx 7.0$), and 1.1×10^{25} dyne-cm ($M^{\max} \approx 6.0$). The historic record is clearly very short for such a region. Gutenberg and Richter (1954) list one event with $M = 6\frac{1}{2}$, at a depth of 100 km (December 23, 1945; 10N, 62W) and another, shallow event with $M = 6$ (February 27, 1940; $8\frac{1}{2}$ N, 62W), and the United States Geological Survey and International Seismological Center routinely report one or two events with $M > 4$ each year. Given the quality of its location, it is suspected that the event with $M = 6\frac{1}{2}$ occurred further north, where intermediate depth events are common, and not in the region under consideration. In any case, the uncertainties in the frequency of occurrence are very large and difficult to quantify. The region is probably straining more slowly than is assumed here, because flexure is oblique to the direction of plate motion. Consequently, the real value of \dot{M}_0^{Σ} is probably less than that assumed here. Moreover, because very large events ($M_0 > 10^{27}$ dynes-cm) do not seem to occur in intraplate regions, except at the bottoms of trenches where ocean floor is actively being

subducted, $M_0^{\max} \simeq 3.5 \times 10^{26}$ dynes-cm is considered to be a more reasonable value than 1.1×10^{28} dyne cm. This yields an estimate for the repeat time of an event with $M \simeq 7$ of 1300 yr, which is more likely to be too short than too long, and for an event of $M \geq 6$ of about 130 yr.

SUMMARY

This study is motivated by the conviction that the historic record of seismicity from most regions is too short to allow estimation of the frequency of occurrence of major earthquakes with much confidence. Accordingly, the purpose is to develop a formalism that relates the frequency of occurrence of earthquakes to the rate of slip on a major fault, or to the rate of deformation of a region with many faults. Because earthquakes cause slip on faults and finite strain in the Earth, their relative sizes are best expressed in terms of a parameter proportional to the amount of slip and to the areal extent of faulting, i.e., the seismic moment, $M_0 = \text{shear modulus} \times \text{rupture area} \times \text{average slip}$ (Aki, 1966, 1967). With plate tectonics we can now measure the long-term average rates of slip on major faults and plate boundaries, and neotectonic studies offer the possibility of establishing rates of deformation not only in regions of more complicated faulting but also for shorter periods of time (tens of thousands of years). The frequency of occurrence of earthquakes with different seismic moments is therefore related to and can be estimated from assumed (or measured independently) rates of slip in the region of interest.

First, the relative frequency of occurrence of earthquakes with different seismic moments [equation (4)] is deduced from the Gutenberg-Richter empirical formula relating frequency of occurrence as a function of magnitude [equation (2)] and from another empirical formula relating magnitude and seismic moment [equation (3)]. Presumably there is a maximum possible seismic moment that can occur in any region, and the frequency of occurrence relation includes the maximum moment for the region of interest as a parameter.

The total seismic moment that occurs on a fault zone in a given amount of time is the sum (or integral) of the various moments weighted according to their relative frequencies of occurrence. This total moment is expressed as a function of the maximum possible moment [equation (7)].

Given the rate of slip on a fault of known area, we obtain the rate at which the total seismic moment increases. Equating this moment rate, determined from plate motions or other methods, to the total (integrated) moment from earthquakes leads to an estimate of the frequency of occurrence of events with various seismic moments [equation (9)]. The frequency of occurrence can alternatively be expressed as a mean repeat time for events with different seismic moments [equation (10)].

Examples of the use of these formulas are given for underthrusting of oceanic lithosphere and for flexing of the downgoing slab in the trench at island arcs. Then, the formulas are used to predict the frequency of occurrence of earthquakes with different seismic moments in the Southeastern Caribbean.

The repeat times [equation (10)] and recurrence relations [equation (9)] are expressed in terms of three quantities: (a) the exponent, β , in the relative recurrence formula [equation (4)], which corresponds to the value of b in the Gutenberg-Richter frequency magnitude formula [equation (2)]; (b) the maximum possible seismic moment in the particular region; and (c) the average rate of change of seismic moments in the region. The uncertainties in estimating the values of each of these are sufficiently large, that the use of equations (9) and (10) probably allows only a modest improvement in the reliability with which the seismicity can be predicted

over that based on the historic record alone. However, with further seismic and geological work, it should be possible to reduce the uncertainties substantially. In particular, it should be possible to measure β directly and reliably for a relatively short period of seismic history. With further studies of major earthquakes, it should be possible to estimate the maximum possible seismic moments with greater certainty. Geological studies of Quaternary faulting and tectonics offer the possibility of determining rates of slip and frequencies of occurrence of earthquakes on active faults and of deciding what fraction of slip occurs by fault creep and by seismic slip. With such work and the resulting reduction in the uncertainties, the formalism given here should allow a more reliable assessment of the average seismic risks over long periods of time than if based only on the historic record, especially in regions where return periods are relatively long.

ACKNOWLEDGMENTS

This work was stimulated by talks by and discussions with L. Taylor and J. F. Tomblin at the First Caribbean Earthquake Engineering Conference in Trinidad, 1978, and by interest of Instituto Tecnológico Venezolano de Petróleo in Caracas. I thank K. Aki, W. P. Chen, T. C. Hanks, E. Schumacher, and an anonymous reviewer for reading the manuscript and making helpful suggestions. Preparation of the manuscript and publication expenses were supported by the U.S. Geological Survey Contract 14-08-001-16758.

APPENDIX A

The moment tensor

Gilbert (1970) showed that the seismic moment could be written as a tensor so that the orientation and sense of faulting were expressed by it

$$M_{0ij} = \mu A (\bar{u}_i n_j + \bar{u}_j n_i) \quad (A1)$$

where n_i are the direction cosines of the normal to the fault plane, and \bar{u}_i is the vector representation of the average slip. For slip on a planar fault, the principal values of M_{0ij} are M_0 , 0, and $-M_0$ and are oriented in the directions of the T , B , and P axes of fault-plane solutions (Gilbert, 1970).

Kostrov (1974) showed that the average irrotational strain in a volume, V , due to N discrete slips on different faults in the volume is

$$\bar{\epsilon}_{ij} = \frac{1}{2\mu V} \sum_{x=1}^N M_{0ij}^{(x)}. \quad (A2)$$

Again, if the average strain rate in the volume and the relative distribution of earthquakes with different moment tensors are known, equation (A2) can be used to deduce the frequency of occurrence of events with different components of the moment tensors.

Often what is known (from plate motion) is the relative velocity of two blocks separated by a wide zone of deformation but not the total strain field in the zone. The average change in length, l , across the deforming zone is given by

$$\Delta l = \bar{\epsilon}_{ij} \frac{l_j l_i}{l} = \frac{1}{2\mu A} \sum_{x=1}^N M_{0ij}^{(x)} l_i l_j / l^2 \quad (A3)$$

where l_i is a vector describing the distance and orientation across the deforming zone, $l = \sqrt{l_i l_i}$ is the scalar value of the distance, and $A = V/l$ is the cross-sectional area through V and normal to the l_i direction (Chen and Molnar, 1977). If we know the rate of shortening or extension across a region, the orientation of the fault planes, and the relative moment distribution, then equation (A3) can be used to deduce the recurrence relation of events with different moments.

Note that Kostrov's (1974) formulation includes only irrotational strain. If the method were applied to the case where one main fault cut through the volume under consideration, the calculated displacements from equation (A3) would be in error (Chen and Molnar, 1977). The following conceptual comparison is intended to illustrate this (Figure A-1). A region is stressed as in Figure A-1a. In one case, the region deforms by slip on one fault, and in the other by slip on a pair of orthogonal faults. Notice that the moment tensors will be the same for equal average displacements on faults of equal area. In both cases the average strain given by equation

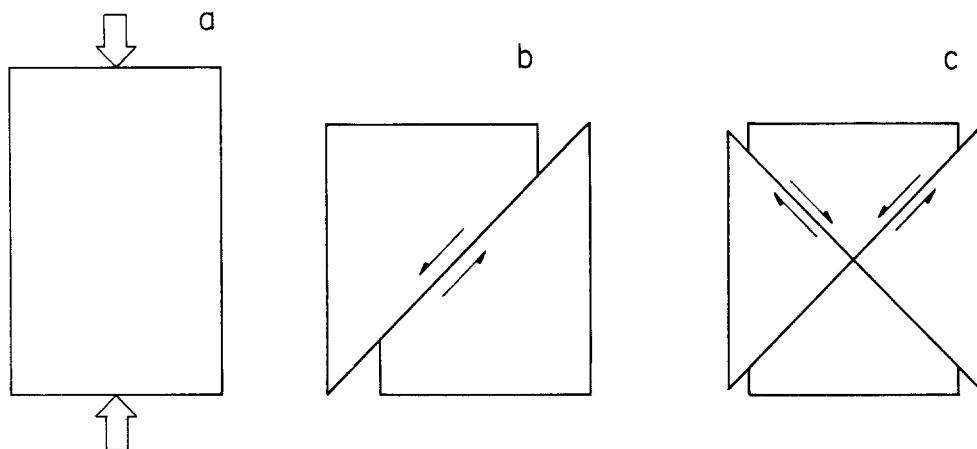


FIG. A-1. Differences in strain and displacement for cases with only one fault or conjugate faults. (a) Block is loaded in vertical direction. (b) Slip can occur on one main fault. (c) Conjugate faults. Strain calculated from equation (A1) is the same for either, but displacements are different. Brune's (1968) method works best for (b) and Kostrov's (1974), for (c).

(A3) will be that for pure shear—contracting along one of the axes bisecting the fault planes and expanding along the other such axis. This strain field will accurately describe the latter case above (Figure A-1c), but will be in error for the former case (Figure A-1b) because the strain is not irrotational.

Thus, whether to use Brune's (1968) or Kostrov's (1974) formulation depends upon the tectonic style of the region. Where there is a single major fault, or system of parallel faults with consistent slip vectors, such as along a transform fault or at a subduction zone, Brune's formulation is applicable. Where, instead, faulting occurs on orthogonal or more randomly distributed planes, such as intracontinental regions of Asia or the basin and range province, Kostrov's is more applicable.

APPENDIX B

Strain rate and frequency of earthquakes in a region of extension or compression

From equation (A2), if we know how a region is deforming and can express it in terms of a tensorial strain rate, then we can calculate the frequency of occurrence

of the elements of the moment tensor, as was done for a single fault. It is unlikely that this will be useful until the distribution and relative importance of faults in a particular region are better known. Nevertheless, there are regions where we do know the rate of shortening (or extension) across them and if we know the orientations of faults we can use equation (A3) as a constraint on the summed moments. As an example, consider the case of shortening in a north-south direction on planes striking east-west and dipping 45° north and south, then for each earthquake

$$M_{0_{ij}} = \mu A' \bar{u} \begin{pmatrix} -1 & 0 & 0 \\ 0 & 0 & 0 \\ 0 & 0 & 1 \end{pmatrix} = \begin{pmatrix} -M_0 & 0 & 0 \\ 0 & 0 & 0 \\ 0 & 0 & M_0 \end{pmatrix}$$

where $i, j = 1, 2, 3$ correspond to the north, east and upward directions, A' is the rupture area, and \bar{u} is the average displacement. From equation (A3) the shortening in the north-south direction, defined by $l_i = (l, 0, 0)$, due to many earthquakes is given by

$$\Delta l = \frac{1}{2\mu A} \sum_{x=1}^N M_0^{(x)}$$

where A is the cross-sectional area with east-west strike. Therefore, the summed seismic moment is given by

$$M_0^\Sigma = 2\mu A \Delta l.$$

If the rate of shortening (or extension) \bar{v} is known, then we have

$$\dot{M}_0^\Sigma = 2\mu A \bar{v}. \quad (\text{B1})$$

The factor of two arises because the faulting occurs on planes oblique to the direction of shortening. Substituting equation (B1) into equation (7) yields an estimate of α to be used in equations (9) and (10). These recurrence relations will not include earthquakes with slip on planes for which no north-south shortening occurs, so that they will not necessarily include all of the seismicity. It is necessary to know the entire strain field in order to include all of the seismicity.

APPENDIX C

For large magnitudes, it appears that $\beta \approx \frac{2}{3}$, but for smaller magnitudes simple theoretical arguments (Kanamori and Anderson, 1975) imply that c in equation (3) might be approximately equal to 1. If b in equation (2) is also equal to 1, then $\beta = 1$. Let us consider the effect of this on the relative contribution of small earthquakes to the summed or integrated moment. Let

$$N(M_0) = \alpha_1 M_0^{-\beta} [1 - H(M_0 - M_0^{\max})] \quad \bar{M}_0 \leq M_0 \quad (\text{C1})$$

$$N(M_0) = \alpha_2 M_0^{-1} \quad M_0 < \bar{M}_0. \quad (\text{C2})$$

As M_0 approaches \bar{M}_0 , $N(M_0)$ calculated from equation (C2) should approach $N(\bar{M}_0)$ given by equation (C1). Therefore

$$\alpha_1 \bar{M}_0^{-\beta} = \alpha_2 \bar{M}_0^{-1}$$

or

$$\alpha_2 = \alpha_1 \bar{M}_0^{1-\beta}.$$

From equations (C1) and (C2), we have the following expressions for $n(M)$

$$n(M) = \alpha_1 \beta M_0^{-\beta-1} + \alpha_1 M_0^{-\beta} \delta(M_0 - M_0^{\max}) \quad \bar{M}_0 \leq M_0 \quad (C3)$$

$$n(M) = \alpha_2 M_0^{-2} = \alpha_1 \bar{M}_0^{1-\beta} M_0^{-2} \quad M_0 < \bar{M}_0. \quad (C4)$$

Let us calculate $\dot{M}_0^\Sigma = \int_{M_0^{\min}}^\infty M_0 n(M_0) dM_0$, where we consider only events with seismic moments greater than M_0^{\min}

$$\dot{M}_0^\Sigma = \frac{\alpha_1 M_0^{\max 1-\beta}}{1-\beta} - \frac{\alpha_1 \beta \bar{M}_0^{1-\beta}}{1-\beta} + \alpha_1 \bar{M}_0^{1-\beta} \ln (\bar{M}_0 / M_0^{\min}). \quad (C5)$$

Notice that if we let M_0^{\min} approach zero, the effect of the smallest earthquakes allows the accumulated moment to increase indefinitely, something we know not to occur. For very small events, β must be small.

Let us evaluate equation (C5) with some reasonable parameters. Let $\beta = \frac{2}{3}$. If $\bar{M}_0 = 10^{-3} M_0^{\max}$ ($\bar{M} = M^{\max} - 2$), then the second term is 6.7 per cent of the first term. If $\bar{M}_0 = 10^{-6} M_0^{\max}$, then the second term is only 0.67 per cent of the first. If $\bar{M}_0 = 10^{21}$ dyne-cm ($\bar{M} \sim 4.3$) and $M_0^{\min} = 10^{15}$ dyne-cm ($M^{\min} \sim -0.7$), then for $M_0^{\max} = 10^{27}$ dyne ($M^{\max} \sim 7.3$), which is quite small, the third term is less than 5 per cent of the first term. Given the uncertainties in the parameters in equations (9) and (10), errors of several per cent are not important. Only if equations (2) and (3) could be extended to earthquakes much smaller than can be recorded would these smaller events contribute substantially to total slip on the faults, and then only if β in equation (4) were equal to one.

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Manuscript received July 24, 1978