UNITED STATES OF AMERICA
NUCLEAR REGULATORY COMMISSION

BEFORE THE ATOMIC SAFETY AND LICENSING APPEAL BOARD

In the Matter of	)	
PUBLIC SERVICE COMPANY OF NEW HAMPSHIRE, *et al.*	)	Docket Nos. 50-443 50-444
(Seabrook Station, Units 1 and 2)	)	

STATEMENT OF DR. MICHAEL CHINNERY

ON REMAND TO THE ATOMIC SAFETY AND LICENSING APPEAL BOARD

SUBMITTED BY

THE NEW ENGLAND COALITION ON NUCLEAR POLLUTION

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ON THE APPLICATION OF THE "PROBABILISTIC" METHOD TO THE ESTIMATION OF SEISMIC RISK AT THE SEABROOK NUCLEAR POWER PLANT SITE

by

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Introduction

The Safe Shutdown Earthquake (SSE) is defined as "that earthquake which is based upon an evaluation of the maximum earthquake potential considering the regional and local geology and seismology and specific characteristics of local subsurface material" (NRC Rules and Regulations, Part 100, Appendix A, Section IIIc).

There are two methods which have been proposed for the estimation of the SSE:

(i) The "Deterministic" Method: In this case, the largest earthquake in the historical record in the tectonic province containing the site is taken to be the SSE. Some additional conservatism may be included by making the SSE larger than the largest historical earthquake, though this has to be based on geological evidence. In the Eastern U.S. the lack of detailed correlation between seismicity and geological structure makes it very difficult to estimate the validity and amount of this additional conservatism.

(ii) The "Probabilistic" Method: Here the historical record is taken as only a sample of the long term seismicity of the tectonic province, and an attempt is made to extrapolate this relatively short record to longer time intervals. In this case, the concept of the "maximum earthquake potential" used in

> the definition of the SSE has to be modified, and the SSE must be defined as that earthquake which will occur in the tectonic province containing the site with some fixed acceptable level of annual risk or probability. This acceptable level of risk is not defined in the NRC Rules and Regulations.

The Nuclear Regulatory Commission has ruled (Order CLI-80-33, 25 September 1980) that the second approach is not inconsistent with Appendix A, given our present understanding of earthquake science. In what follows, we explore the application of this approach to the Seabrook site.

The Historical Record

In New England the historical record of earthquake occurrence is approximately 300 years long. The only catalog of seismic events in this area that has been published in the scientific literature is that by Smith (1962, 1966). The earlier parts of this record are not very reliable. Instrumental records, again of variable quality, are available since the 1920's, but only in the last few years has a proper seismic network been installed. This network has detected relatively few events since it was created, and can contribute little to the assessment of seismic risk in the area.

We are, therefore, forced to work with the historical data set, in spite of its inadequacies. Now we have to ask two important questions: Supposing that we thoroughly understood the long term seismic characteristics of the area, how well can we predict the seismic activity during the next 50 years (the anticipated lifetime of the Seabrook plant)? And, is the 300 year historical record really representative of the long term seismic characteristics?

Both of these questions are difficult to answer. The first is most easily disposed of, since if we cannot use the past to predict the future, we have to give up any attempt to estimate seismic risk. We assume at this point that a thorough characterization of the seismicity in the past is indeed a reasonable basis on which to compute future seismic risk.

The second question cannot be disposed of so easily, and lies at the heart of all controversy concerning the estimation of seismic risk. How can we use the historical record to make the most reasonable estimate of the long term seismic characteristics? In order to tackle this question, it is convenient to consider the spatial distribution of earthquakes separately from the distribution in time and size. These two aspects are discussed in the following sections.

Selection of a Tectonic Province

The concept of a tectonic province is a legal one (as defined in Appendix A), and has no clear scientific significance. The proposition that earthquakes must in some way be related to geological structure and tectonics is inescapable, but it is not at all clear that large provinces can be defined within which the seismo-tectonic characteristics are in any sense uniform. Attempts to define such provinces usually lead to a wide range of interpretations (see, for example, McGuire 1977 and Tera Corp. Study, 1979). These difficulties certainly apply in the case of New England.

Figure 1 shows a map of the epicenters of earthquakes listed in the Smith (1962, 1966) catalog. Marked clusters of events near the Seabrook site occur in Southern New Hampshire and in Northeastern Massachusetts. In previous studies (Chinnery and Rodgers 1973, Attached as Exhibit 1, and Chinnery 1979, Attached as Exhibit 2), these two clusters have been included in one seismic zone (or tectonic province), and this is indicated by the broken line in Figure 1.

In what follows, we use this Boston-New Hampshire seismic zone as the tectonic province appropriate to the Seabrook site, recognizing that only weak arguments can be made for <u>any</u> choice of tectonic province in this region. (Tera Corp. Study, McGuire 1979) The present choice is at least a reasonable one for the historical local seismicity, since the population density has been highest in this particular area. Instrumental epicenters for 1975-79 (see Figure 2) are roughly consistent with this choice; the cluster of epicenters in Southern New Hampshire can still be seen, but recent seismicity near Cape Ann, Massachusetts, has been low (in apparent contradiction to the historical record). Certainly, neither the historical record nor the instrumental record lead to any good arguments for isolating

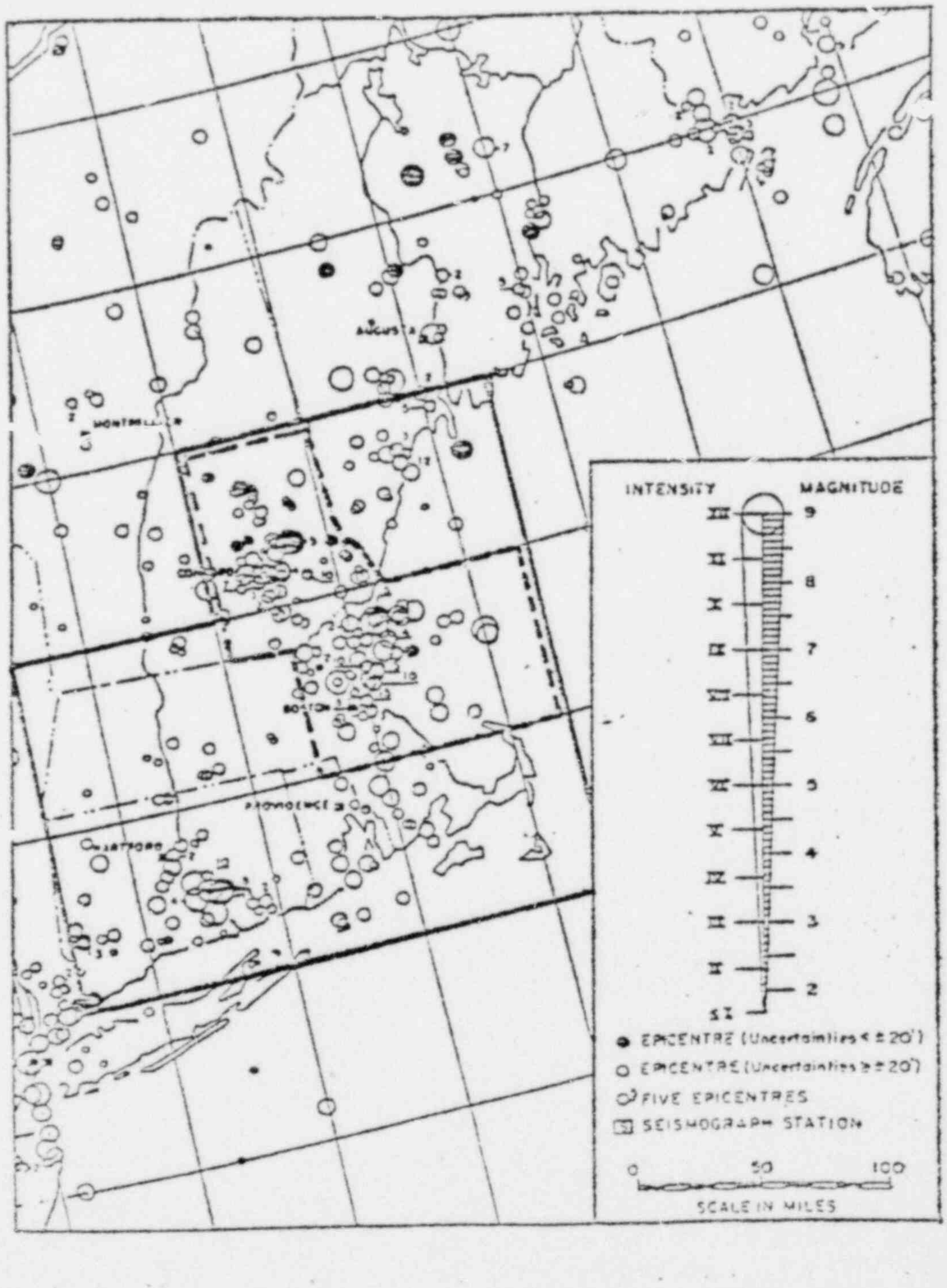

Fig. 1: Epicenters in New England from the catalog by Smith (1962, 1966). The broken line indicates the Boston-New Hampshire seismic zone (from Chinnery, 1979a).

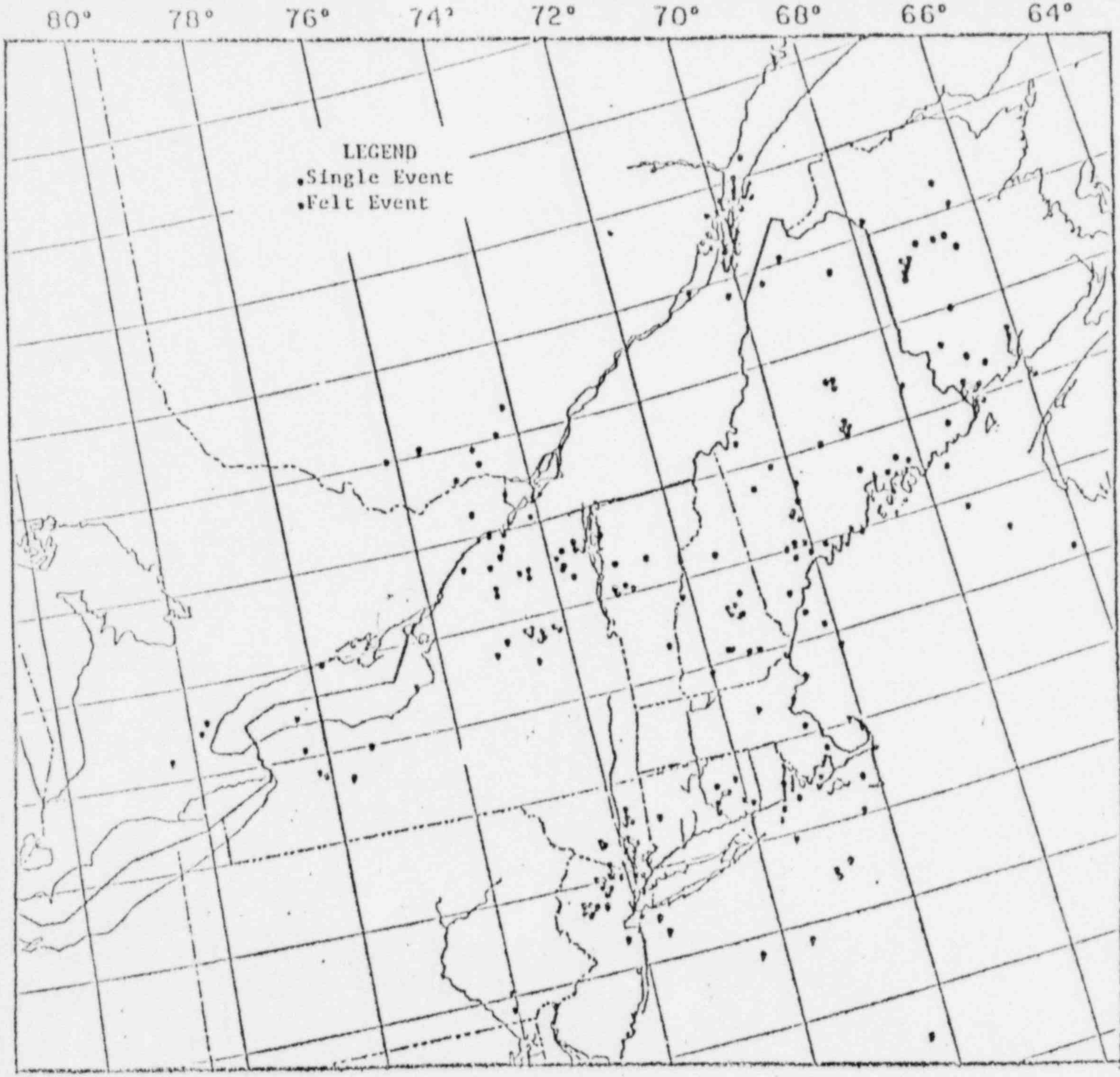

Fig. 2: Instrumentally determined epicenters during the period October 1975–December 1979 (from Bulletin No. 17, Seismicity of the Northeastern United States, NRC Publication NUREG VES-238-030, August 1980).

the Seabrook site from seismicity in Southern New Hampshire and Northeastern Massachusetts.

In view of the inadequacies of the historical record and the difficulty in selecting an appropriate tectonic province, assessment of the seismic risk at the Seabrook site can be based on a number of different assumptions. In my view, the most reasonable and most conservative assumption is that the seismicity of the Boston-New Hampshire zone is a valid basis for estimating the risk at the Seabrook site.

Frequency-Intensity Relationships

The characterization of the seismicity of a province in terms of the rates of occurrence of earthquakes of different sizes is usually accomplished using frequency-magnitude or frequency-intensity relationships. In the present case we use the latter, since only intensities are quoted in the Smith catalog. In addition, we use cumulative frequency-intensity counts, i.e., we count the number of earthquakes larger than or equal to a given intensity value during a given period.

The extraction of frequency-intensity data from a catalog such as Smith's must be carried out with care, since the completeness of the catalog at lower intensities is likely to be a strong function of population density, and therefore of time. We use the approach described in Chinnery and Rodgers 1973 (Exhibit 1) here.

Having extracted and plotted the data for the Boston-New Hampshire seismic zone, we have three important question to consider:

(i) can the data be represented by a linear frequency-intensity relationship?

(ii) if so, what is the slope of the linear relationship?

(iii) is there some upper bound to the intensity of earthquakes that can be expected in this seismic zone?

Let us consider each of these in turn.

1. Linearity of Frequency-Intensity Data

Frequency-intensity data for the Boston-New Hampshire zone are shown in Figure 3 (taken from Chinnery 1979) (Exhibit 2). Clearly, the data are sparse. For the period 1800-1959 only six data points are obtained (for intensities II to VII) and it seems likely that those for intensities II and III are unreliable due to incompleteness (even though these points are based on the very recent period 1928-1959). The remaining four data points actually lie in a relatively good straight line, but the slope of this line (about 0.50) is, as we shall see below, unusually low, and would lead to high estimates for the rate of occurrence of large earthquakes. A more reasonable interpretation is that the number of intensity VII events (3) during this period was unusually high, and that the intensity IV data set may be incomplete.

If these comments are valid, perhaps only the intensity V and VI data points are at all reliable, and we can not make any conclusions from the data themselves about the linearity of the frequency-intensity relationship. In this case, we must rely on information from elsewhere.

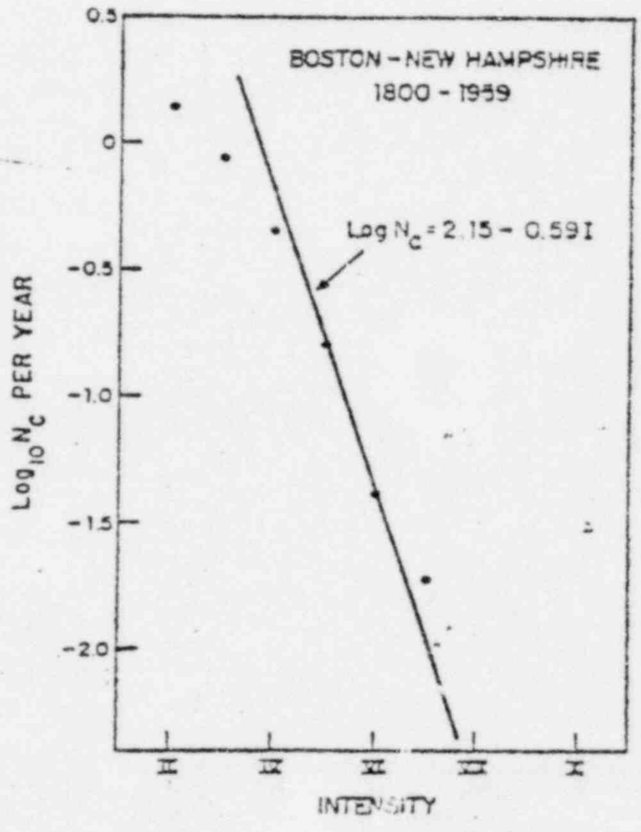

Fig. 3: Frequency intensity data from the Boston-New Hampshire seismic zone, derived from the Smith catalog using the methods given in Chinnery and Rodgers (1973) (from Chinnery, 1979a).

In my view, the current situation can be summarized as follows: The vast majority of seismologists have accepted the linearity of frequency-magnitude data as a working hypothesis. (See, for example, Evernden 1970, Veneziano 1975, and the references cited in those papers). It is, however, still a hypothesis, with no clearly developed theoretical basis. And there are a few instances where non-linearities are apparent in the data. These have led to several publications proposing non-linear relationships, though in my view these can generally be attributed to poor or inadequate data.

The linearity of frequency-intensity data has been discussed much less. Several investigators have proposed linear relationships between intensity and magnitude, (See, for example, Veneziano 1975) and, if these are valid, a linear frequency-magnitude relationship implies a linear frequency-intensity relationship. Of what scientific literature there is, the vast bulk assumes that frequency-intensity relationships are linear (see, for example, references quoted in Chinnery 1979) (Exhibit 2).

One point should be made here. Intensity (i.e., maximum epicentral intensity) is a very different scale from magnitude, and the observed linearity in the relationship between the two at commonly observed intensities has no sound theoretical basis. Certainly for very large earthquakes there must be a departure from linearity, since intensity has an inherent upper bound (intensity XII) while magnitude is an open-ended scale. Note, however, that all scales become unreliable for large events (roughly M>7), due to saturation and other effects.

In summary, the apparent linearity of much frequency-intensity data must be treated as an empirical observation. Its wide acceptance by seismologists suggests that it is useful as a working hypothesis.

2. Slope of Frequency-Intensity Data

If we accept that in any given region we can expect a linear frequency-intensity relationship, the next question must be: Does the slope of this relationship vary significantly from region to region?

The only study that has addressed this point is Chinnery 1979 (Exhibit 2). In that paper it was shown that there seems to be a remarkable uniformity in the slopes determined from various areas of the Eastern U.S. Values of this slope were typically found to lie in the range 0.54 to 0.60, and in fact, all the available data are consistent with a slope of 0.57.

This is an important point for areas such as the Boston-New Hampshire seismic zone, where some of the data points may be unreliable. If we assume that the data are to be fit with a straight line with slope about 0.57, then we can use the most reliable data points (for intensities V and VI) to define the frequency-intensity relationship (see Figure 3). In my view, more complex relationships are not justified by the data.

3. Existence of an Upper Bound Intensity

Having defined a frequency-intensity relationship, we would like to use this to extrapolate beyond the historical data points, to give an estimate of long term seismicity. The question remains: How far may we continue this extrapolation? Is there an upper limit to the size of earthquakes that can occur in an area like the Boston-New Hampshire zone? If so, what is this limit? I have examined this question in some detail (see Chinnery 1979b, Attached as Exhibit 3). My conclusion is that we do not know the answer to these questions at the present time. One aspect of the problem is worth mentioning here. All seismologists (including the author) agree that earthquake size (however measured) cannot increase indefinitely. Physical constraints arising from the earthquake source mechanism will set a limit to both source dimensions and strain release. On a global scale, this upper bound is at a rather high level, somewhat above the largest known earthquakes. On a regional level, much less is known, and there is considerable disagreement between the (guess-) estimates of different seismologists.

In a recent study (Tera Corp. Study 1979), ten experts in the seismicity of the Eastern U.S. made estimates the largest epicentral intensity that might be expected in the Cape Ann, Massachusetts region. These are listed in Table 1, and illustrate the disagreement clearly. There is little point in averaging opinions such as these. Notice, however, that 5 of the 10 experts admit the possibility that the upper bound to earthquake size may be X or greater in this region.

TABLE 1

Estimates of the Largest Earthquake Expected to Occur in the Cape Ann, Massachusetts Region (Tera Corporation, 1979)

Expert	Low Estimate	Best Estimate	High Estimate
3	IX	X	XI
4	VI	VI	X
5		XII	
10	VII	VIII	IX
13	IX	X	XI
7	6.2	6.4	6.7
8		6.0	
9	5.7	6.2	6.7
11	6.0	6.5	7.0
12	5.75		6.25

(Here, arabic numerals indicate magnitudes; as a rough conversion to intensities, 6.0 VIII and 7.0 IX or X.)

In my view, the only valid conservative interpretation of this set of opinions is that we should admit the possibility of an intensity X earthquake in the Boston-New Hampshire seismic zone, until convincing scientific evidence arises that will persuade us to revise this value.1/

1/ In its previous ruling, the Appeal Board indicated a difficulty in accepting that data from one area could be of any use in attempting to project the seismic characteristics of another area. This problem is fully (cont'd on next page)

Estimation of Seismic Risk at the Seabrook Site

In the above sections we have laid out our basis for the evaluation of seismic risk at the Seabrook site. To summarize: We have selected a "tectonic province" containing the site, which extends from Southern New Hampshire to Northeastern Massachusetts. Following Appendix A (section V, para. a.1.ii), we assume that the largest earthquakes that can occur in this province will occur at the site. Frequency-intensity data are extracted from Smith's (1962, 1966) catalog using only data after the year 1800. Through these data we will fit a linear frequency-intensity relationship, with a slope of about 0.57, and use this as a basis for extrapolating to obtai a measure of long term seismicity. Extrapolation of the line is considered valid out to an intensity of about X.

The result of applying these procedures is shown in Figure 4. The data points are the same as shown in Figure 3. The solid line has a slope of 0.57. Broken lines indicate slopes of 0.50 and 0.68; these would appear to be very wide bounds, based on other data from the Eastern U.S. (Chinnery 1979, Exhibit 2). The 1955 Cape Ann earthquake occurred a litte over 200 years ago, and has been estimated to have had an epicentral intensity of between VII and VIII.

1/ discussed in Chinnery 1979 (Exhibit 2). The empirical observation was there presented that data from three areas of the Eastern U.S. are consistent with a uniform frequency-intensity slope of about 0.57, and that the data contained no evidence for the presence of a limit to earthquake size in these areas. This is an empirical observation and is independent of the geological characteristics of the three areas.

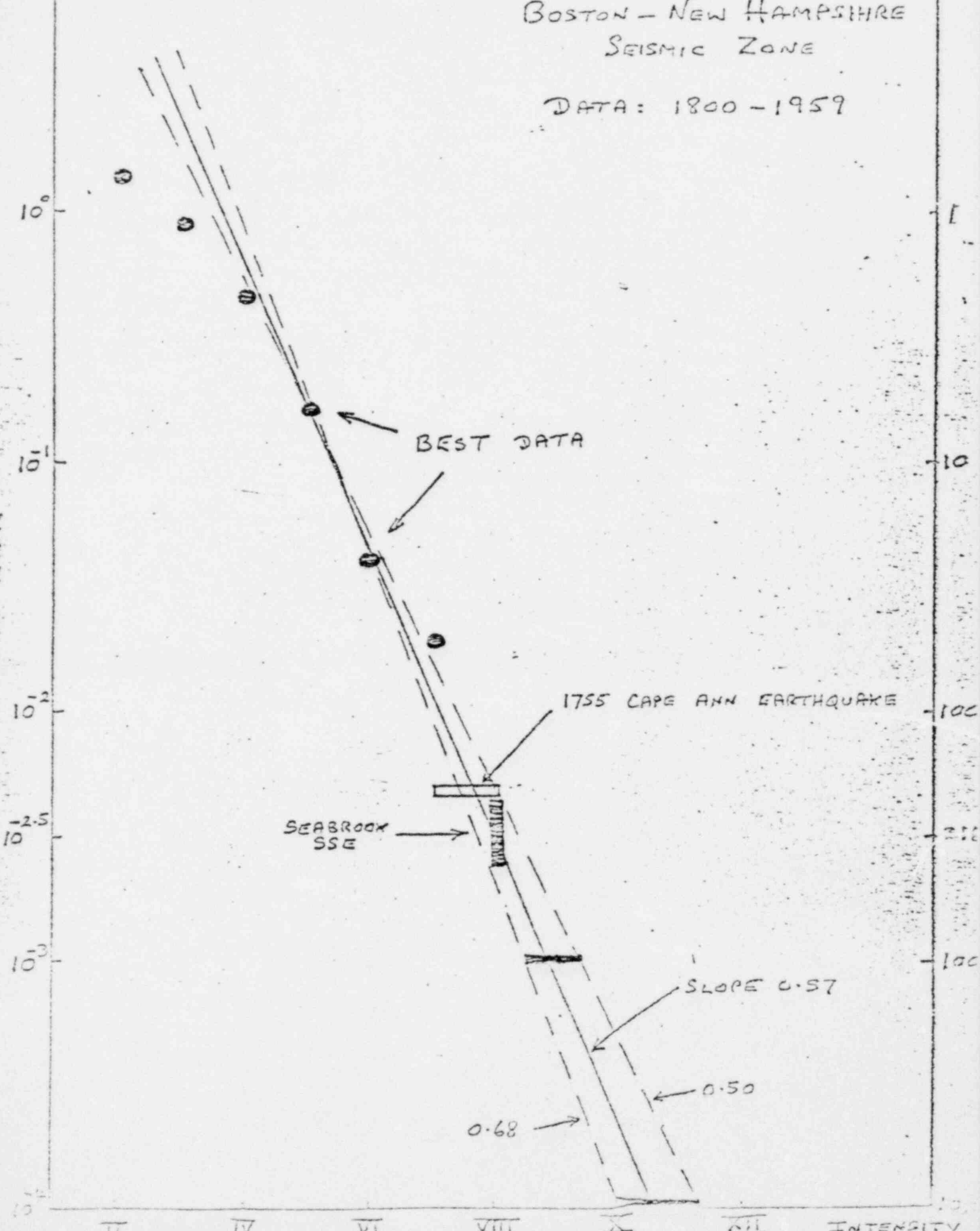

Fig. 4: Extrapolation of frequency-intensity data for the Boston-New Hampshire seismic zone (see text).

The open rectangle shows how this earthquake would plot on the present diagram. Clearly, that event is consistent with our extrapolation from later data.

The current Seabrook SSE of VIII is found to occur with an annual risk of about $10^{-2.5}$ (this corresponds to a return period of about 300 years). An annual risk of 10^{-3} (return period of 1000 years) corresponds to an intensity IX, and an annual risk of 10^{-4} corresponds to an intensity of at least X.

The problem that remains is to define the acceptable level of risk which will define the choice of the SSE. Though numbers in the range 10^{-3} to 10^{-4} per year have been mentioned in the past, I am not aware of any formal definition of this risk, which clearly involves many societal, economic and political factors.

Conclusion

This case study of the application of the "probabilistic" method brings out all the main features of the method. Most important, it indicates that the definition of the Safe Shutdown Earthquake must be accompanied by a definition of the acceptable annual risk of the occurrence of the ground motion corresponding to this size of earthquake.

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Qualifications

An updated resume is attached as Exhibit 4.

UNITED STATES OF AMERICA

NUCLEAR REGULATORY COMMISSION

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In the Matter of PUBLIC SERVICE COMPANY OF NEW HAMPSHIRE, *et al*. (Seabrook Station, Units 1 and 2)	Docket Nos. 50-443 50-444

CERTIFICATE OF SERVICE

I hereby certify that copies of the "Statement of Dr. Michael Chinnery on Remand to the Atomic Safety and Licensing Appeal Board" were mailed postage pre-paid this 17th day of February 1981 to the following:

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Exhibit 1

EARTHQUAKE NOTES, VOL. XLIV, NOS. 3-4, JULY-DECEMBER, 1973

Earthquake Statistics in Southern New England

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ABSTRACT: New England has the longest recorded history of earthquake activity in the United States. Because of the high population density and because the historical data are likely to be more complete in the Southern New England, we have examined the statistics of the earthquake data and then constructed recurrence relations in an attempt to estimate the mean return period as a function of earthquake size.

INTRODUCTION

New England has the longest recorded history of earthquake activity in the United States. Several catalogs of earthquakes in this area have been compiled, the most comprehensive of which appears to be due to Smith (1962, 1966) and covers the period from 1534 to 1959. Smith's data are used throughout this report. No attempt to include information after 1959 has been made, in order to preserve the apparent homogeneity of Smith's data set.

It seems very likely that several analyses of these data have been made in the past. However, if this is so, the results of the studies are not generally available in the scientific literature. Instead, it is common to find, in reports by insurance companies, site investigators, city planners, etc., vague statements concerning the low level of seismicity in this area, the infrequent occurrence of damaging earthquakes, and even the maximum size of earthquake that may be expected.

In view of the high population density in Southern New England, it does not seem advisable to base major planning decisions on statements such as these. Instead, we must examine the historical record in considerable detail. These data are far from perfect, but they are essentially all that we have. The level of seismicity is low enough that little information can be deduced from instrumental records, which are only available after about 1925. In addition, the historical data suggest that the seismicity observed in the last century may be unusually low.

The purpose of this paper, then, is to examine Smith's earthquake catalog in detail. We shall concentrate on the Southern New England region, because of the high population density, and because the historical data are likely to be more complete in this area. We shall examine the statistics of the earthquake data, construct recurrence relations, and attempt to estimate the mean return period as a function of earthquake size. We shall study both the area as a whole, and also several smaller subareas where much of the historical activity has been concentrated.

THE DATA

Figure 1 shows Smith's (1966) map of epicenters in the New England area, for the period 1534-1959. This map is a portion of Smith's much larger diagram covering Eastern Canada and the Northeastern United States. In all, Smith lists 729 earthquakes in the Northeastern United States.

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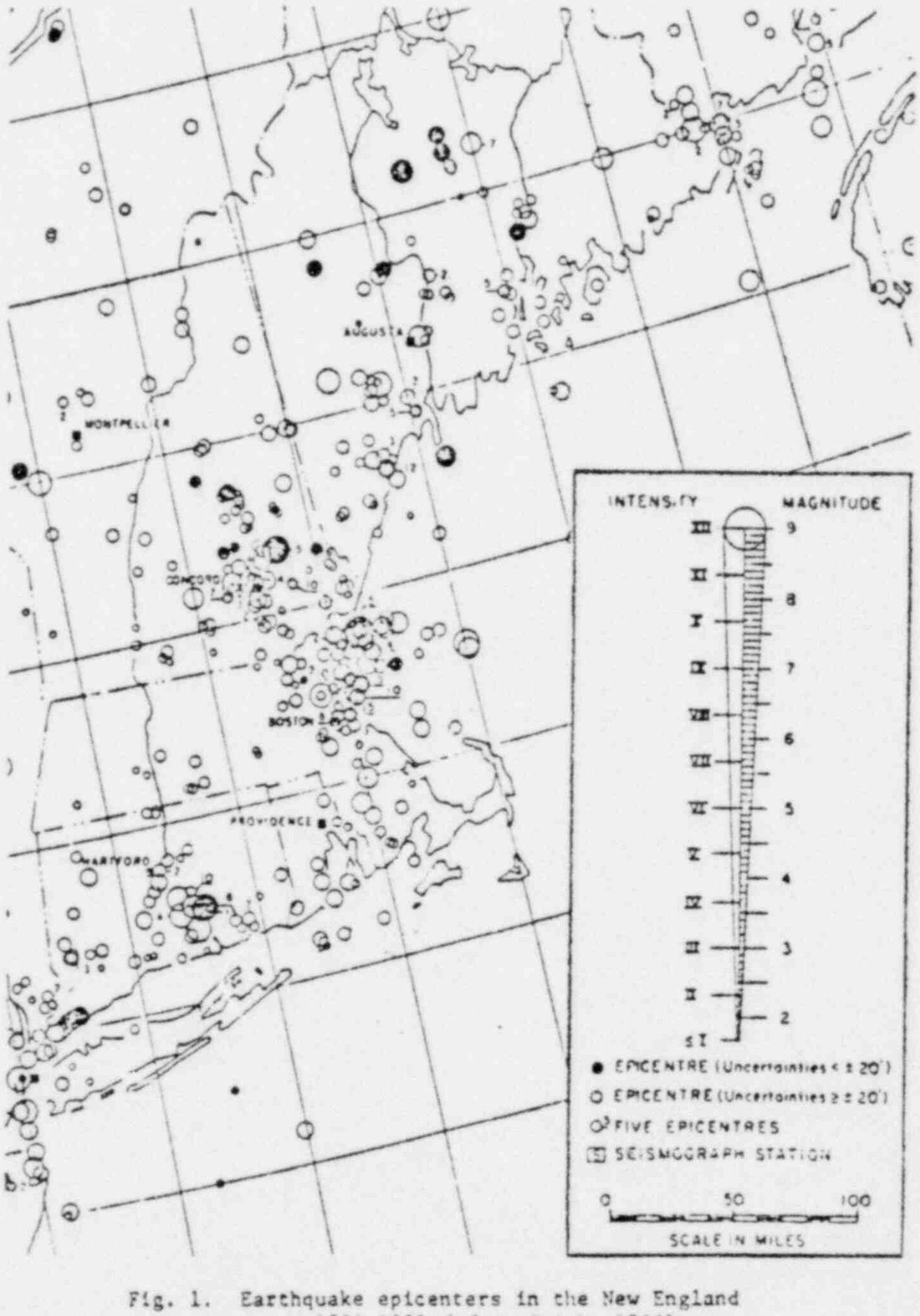

Fig. 1. Earthquake epicenters in the New England area, 1534-1959 (after Smith, 1966).

We have chosen to select a portion of the New England area for study. This portion, which we term "Southern New England," is shown in Fig. 2. Included are the states of Massachusetts, Connecticut, Rhode Island, and the southern parts of New Hampshire and Maine. The ocean area East to

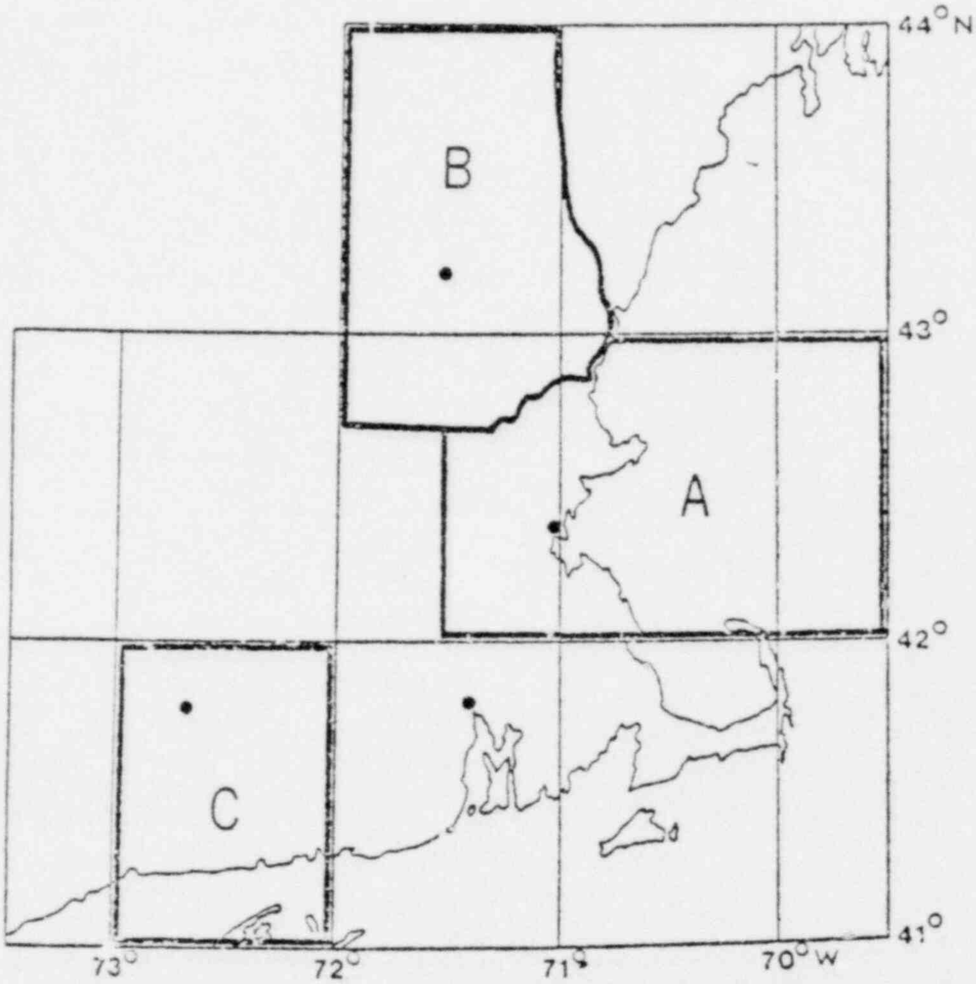

Fig. 2. The area that we have chosen to define as "Southern New England" for the purposes of this study.

69.5° West has been added to these, so that a number of poorly determined epicenters off the coast of Massachusetts may be included in the statistics. New England (defined above) is more distant than 50 miles from a major center of population (100,000 inhabitants or more). In some average sense, therefore, a random epicenter in this area is likely to be within 25 miles of a large population center. This is of some importance in attempting to estimate earthquake risk in the area as a whole.

Much of the seismic activity in this region (see Fig. 1) is concentrated into three zones, which are labelled in Fig. 2 as A, B, and C. A includes the area around Boston, B refers to the southern part of New Hampshire, and C denotes the region of Connecticut around Hartford. The division between zones A and B is rather arbitrary, and the statistics of these zones are analyzed, both separately and together, later in this report.

In all, Smith lists 353 epicenters in Southern New England. Of these 99 (28%) lie in area A, 96 (27%) lie in area B, and 55 (16%) lie in area C. The three active zones, therefore account for 70% of the data for the whole area.

There are three principal errors in this type of information. These are uncertainties in epicenter locations, problems in the determination in intensity, and incompleteness of the data set.

The uncertainties in most of the epicenter locations shown in Fig. 1 are quite large. The historical data will clearly be strongly influenced by the population distribution. Note, however, that the larger earthquakes, whose effects extended over a large area, are likely to have less accurate epicenters than small ones. More recent instrumental determinations of epicenters are also subject to error, though o' a different kind. The seismic travel time curve in New England is not well known. This is in part due to the heterogeneity of the geology, and in part due to the poor spatial distribution of epicenters in relation to observatories in the area. Even the large New Hampshire earthquakes of 1940 (M = 5.8) cannot be located to better than ± 20 km. It is doubtful whether any of the epicenters on Fig. 1 are any more accurate than this, and most are much less accurate.

The determination of the intensity of an earthquake from historical eyewitness accounts is notoriously difficult. Estimates are subject to population distribution, the personal feelings of the observer, and the interpretation of the cataloger. The influence of these factors is mixed. Observers are likely to overestimate the intensity of an earthquake shock. However, the cataloger has clearly tried to take this into account in his assignment of intensities. In addition, if the population density was sparse, it is quite possible that no report was received from the highest intensity zone close to the epicenter. In view of this, it does not appear realistic to assume that the historical reports are grossly exaggerated. In some instances they may result in underestimates of the earthquake intensity.

The worst problem of historical earthquake data is, of course, its completeness. There is no doubt that the data becomes more incomplete as one goes further into the past (at a given intensity) and to smaller intensities (at a given time). On the other hand, we need the longest time period and the largest range of intensities possible in order to arrive at meaningful statistics. Because of this, some subjectivity is necessary in selecting the portions of the data to be analyzed.

It seems likely that data regarding earthquakes that occurred before 1700 are unreliable. We have not tried to use these data. However, there is a high probability that all of the large earthquakes since then have been recorded. Similarly, it is only in the recent past that we may expect a fairly complete record of small events. Therefore, in order to try to exclude this type of deficiency from the data, we have chosen to analyze three subsets of each data set. These subsets, showing the time interval studied at each intensity, are listed in Table 1. By comparing the results for the three subsets, we will, hopefully, obtain some information about the completeness of the data set as a whole.

INTENSITIES AND MAGNITUDES

Intensities mentioned in this report refer to the Modified Mercalli Scale of 1931 (see, for example, Smith, 1962). Magnitudes, where quoted, are local magnitudes. Historical eyewitness accounts lead to estimates of the intensity of the earthquake at the observing site. Magnitudes can only be determined reliably from instrumental records. Because of the

Table 1. Subdivision of data into subsets.

Intensity	Subset 1	Subset 2	Subset 3
IX	1700-1959	1800-1959	1860-1959
VIII	1700-1959	1800-1959	1860-1959
VII	1700-1959	1800-1959	1860-1959
VI	1800-1959	1800-1959	1860-1959
V	1860-1959	1860-1959	1860-1959
IV	1900-1959	1900-1959	1900-1959
III	1928-1959	1928-1959	1928-1959
II	1928-1959	1928-1959	1928-1959

nature of the historical data, we shall use intensities throughout.

It appears in general to be possible to relate the maximum epicentral intensity I to the local magnitude M by a linear algebraic expression. Gutenberg and Richter (1956) determined the following relation for Southern California:

$$M = 1 + \frac{2}{3} I. \tag{1}$$

The number of earthquakes in the present area for which both M and I are known is small. Figure 3 shows that these data are consistent with the Gutenberg-Richter relation. A least squares fit to the data points in Fig. 3 leads to the relation:

$$M = 1.2 + 0.6 I. \tag{2}$$

In view of the uncertainties in both I and M, the difference between Eqs. 1 and 2 is negligible.

The linear relation between I and M is a useful one. Using it we may convert instrumental magnitudes into intensities (some recent earthquakes in Smith's catalog are listed with only a magnitude). More important, it enables us to compare our local statistics with global studies, which are usually quoted in terms of magnitude.

It will be convenient at this stage to define what we mean by a damaging earthquake. On the basis of the Modified Mercalli Scale of intensity, the onset of considerable destructive ability occurs at an epicentral intensity in the range VIII to IX. We have therefore chosen to define a damaging earthquake as one with an epicentral intensity of $VIII^{1/2}$ or greater. This is a very conservative choice, since even an earthquake of intensity VII is likely to cause some damage, particularly if it occurs in a heavily populated area.

An intensity of $VIII^{1/2}$ corresponds to a magnitude of about 6.5. The possible destructive effects of an earthquake of this size may be illustrated by the 1971 San Fernando earthquake in California. Considerable damage and rather high accelerations of the ground were observed in the case of this much publicized event.

In addition to the size of an earthquake, we must also consider the areal extent of the region subject to damage. This is much less well defined quantity, since it depends so much on the superficial geology. Linehan (1970) has given an earthquake intensity attenuation

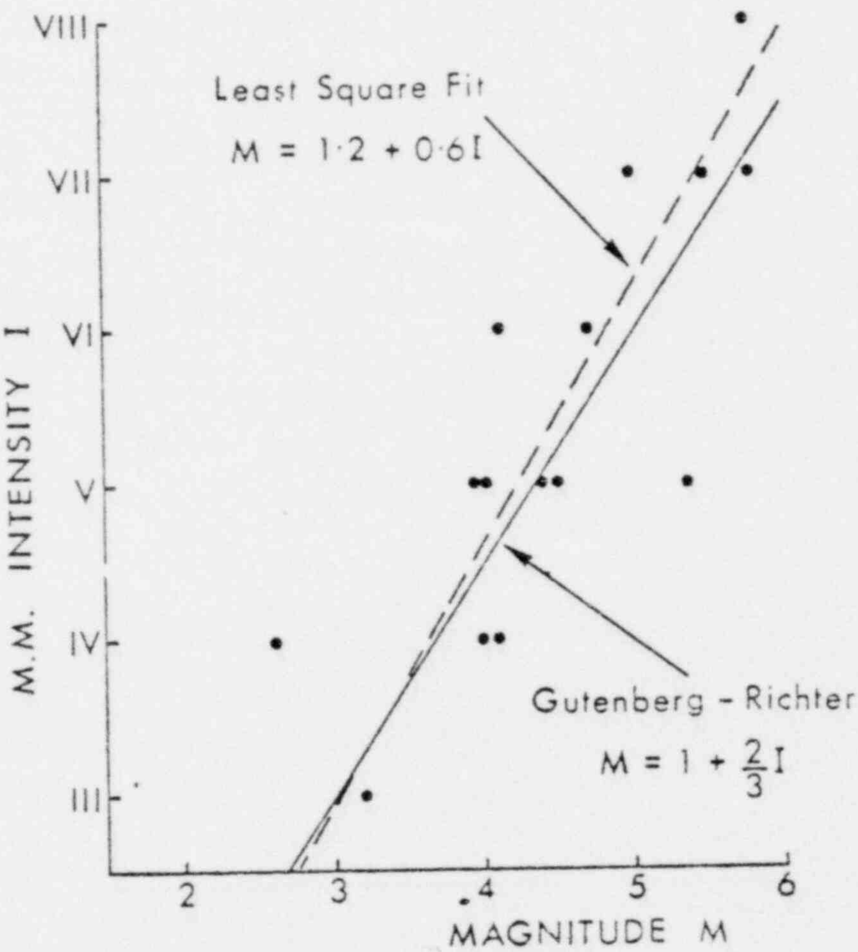

Fig. 3. The relationship between magnitude and intensity for 15 earthquakes in Northeastern United States and Eastern Canada.

scale which suggests that intensity VIII will extend out to a radius of about 15 miles from an epicentral intensity of $VIII^{1/2}$. This is also a conservative estimate, and the radius may easily be doubled in regions of unfavorable geology. In view of this, and the high population density of the area under consideration, it seems unlikely that an intensity $VIII^{1/2}$ earthquake could occur in Southern New England without causing considerable damage and loss of life.

Certainly, any earthquake with intensity greater than this can be relied upon to cause great damage. For this reason, we shall also pay some attention to the possible future occurrence of earthquakes of intensity IX and X.

FREQUENCY-INTENSITY RELATIONS

It has been clearly demonstrated in many parts of the world that there is a linear relationship between earthquake frequency and earthquake magnitude (see, for example, Evernden, 1970) of the following form:

$$\log N_c = a - bM, \tag{3}$$

where N_c is the number of earthquakes occurring within a region in a given time period with a magnitude greater than or equal to M. a and b are constants; a depends on the size of the area chosen and the length of the time period concerned, and is an overall measure of the seismicity of the area. b usually lies in the range 0.5 - 2.0, and appears to be related to the nature of the tectonic activity causing the earthquakes. Logarithms, unless ot rwise stated, are to base 10.

If a linear relation exists between magnitude and intensity, as we have discussed earlier, then clearly we may write

$$\log N_c = c - dI, \tag{4}$$

where, now, N_c is the number of earthquakes occurring within a region in a given time interval with an intensity greater than or equal to I. c and d are constants.

The Relation 4 is a very useful one. It enables us to use the data for smaller earthquakes, which are plentiful, to determine the frequency of occurrence of large earthquakes. In the later sections of this report we shall attempt to determine the constants c and d from the data in Smith's catalog.

In considering the statistics of the earthquakes, instead of using the quantity N_c, it is more convenient to define the "mean recurrence time" (MRT). MRT is simply the average time between earthquakes with a given intensity I or greater, and is equal to $1/N_c$ time periods. We shall be particularly concerned with the determination of MRT for damaging earthquakes.

Before we proceed, however, we must consider the range of validity of Eq. 4. Where complete data has been obtained, the frequency-magnitude relation (Eq. 3) has been shown to be valid over a remarkable range of magnitude (from greater than 8 down to less than 0). There is some theoretical reason to suspect that there is a limit to the possible size of earthquakes. If this limit exists, it is not well known, and may be of the order of magnitude 9. Such theoretical limits are well beyond the sizes of the earthquakes that we shall consider in this report.

We must next examine whether there is any evidence that there is an upper limit to the sizes of earthquakes to be expected in the New England area. There seems to be some confusion on this point. In fact, there is no basis for suggesting that such an upper limit exists, and as we shall see, analysis of the historical data supports this statement. The largest earthquakes that have been recorded in Southern New England are listed in Table 2. At least one, and probably two earthquakes of intensity IX (magnitude about 7) have been recorded in the past 400 years. This length of record is far too short to conclude that an event with intensity X (or greater) has not occurred in the past, or will not occur in the future. The Charleston, South Carolina, earthquake of 1886 had an intensity X, and occurred in an area that is somewhat less seismically active than New England.

Table 2. Large earthquakes in Southern New England.

Date	Location	Intensity
1568	Rhode Island	VII
1574	Rhode Island	VII
1584	Rhode Island	VII
1592	Rhode Island	VII
July 11, 1638	Off Cape Ann, Mass.	VIII
November 9, 1727	Near Newbury, Mass.	IX
June 14, 1744	Off Cape Ann, Mass.	VIII
November 18, 1755	"about 200 miles" East of Cape Ann, Mass.	IX
May 16 or 18, 1791	Near Moodus, Conn.	VIII
October 5, 1817	Northeastern Mass.	VII
December 20, 1940	Ossipee Lake, N.H.	VII
December 24, 1940	Ossipee Lake, N.H.	VII

We must therefore admit the probability that large earthquakes will occur in Southern New England, if at infrequent intervals, until some new information arises that dismisses this possibility. It should be added that the absence of a very large earthquake in the recorded history of Southern New England is not a reason for complacency. It is conceivable that a long time has elapsed since the last large earthquake in this area. If this were the case, the probability of one occurring in the near future could be quite high.

RECURRENCE RELATIONS: SOUTHERN NEW ENGLAND

We consider first the whole Southern New England region (defined in Fig. 3). Smith (1962, 1966) lists 353 events in this area during the period 1534-1959, after all obvious aftershocks are removed from the data. The distribution of these earthquakes in intensity and time is shown in Table 3. Where the intensity of an event is listed as being between two levels (e.g., IV-V), one half event has been included into each level.

Table 3. Earthquake data for Southern New England.

Intensity	Before 1700	1700-1799	1800-1859	1860-1899	1900-1927	1928-1959
IX	-	2	-	-	-	-
VIII	1	2	-	-	-	-
VII	4	-	1	-	-	2
VI	1	2	3	1/2	1	1-1/2
V	2	8	5	8-1/2	9	6-1/2
IV	5	16	24	13	22	21
III	2	16	40	23	14	26-1/2
II	3	3	-	28	5	32-1/2

It is clear from Table 3 that the data from before 1700 are very incomplete. At the lower intensity levels this incompleteness continues until late in the historical record. For this reason, we have disregarded portions of the data, and have analyzed the remainder in

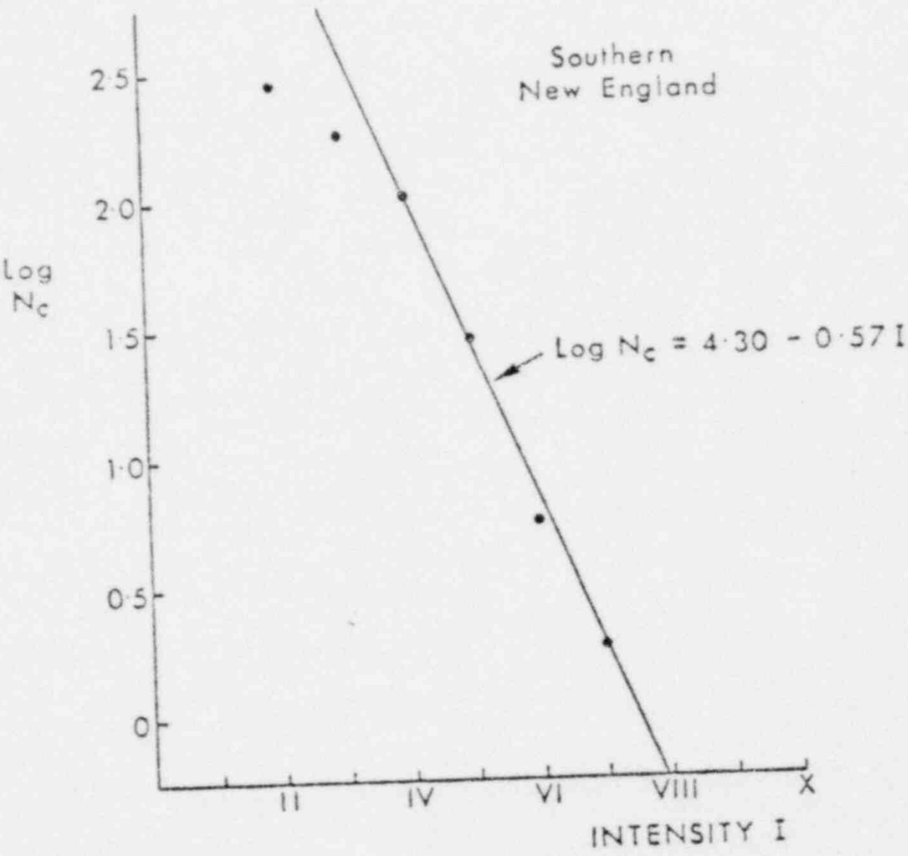

Fig. 4. Frequency-intensity plot for 135 events in Southern New England. N_c is the cumulative number of events (with intensity I or greater) per century.

the three subsets described earlier (Table 1). Note, however, that all earthquakes with intensity VIII or IX occurred before 1800. These large events will therefore only appear in subset 1 of the data.

Frequency-intensity plots for the three subsets of the data have been constructed. As may be expected, the large events in the 1700's make subset 1 very nonlinear. We conclude that subset 1 is unreliable. Subsets 2 and 3 are almost identical, and we therefore have chosen to use subset 2 (which contains more events) as our most reliable data set. The frequency-intensity graph for this subset is shown in Fig. 4. The ordinate is the logarithm (to base 10) of the cumulative number of events with intensity I or greater, per century.

The points in Fig. 4 define a fairly linear relation. The low points at intensities II and III are to be expected, since it is virtually impossible to obtain a complete record of these small events, even in the recent past. The remaining points are consistent with a slope that lies within the range of 0.54-0.60. For this reason, and because of the results given in the next section of this paper, our best estimate of the slope of the frequency-intensity relation is 0.57 (±0.03). The data then determine the following recurrence relation:

$$\text{Log } N_c = 4.30\ (\pm 0.15) - 0.57\ (\pm 0.03)\ I, \tag{5}$$

converting this into a frequency-magnitude relation; using Eq. 2, we obtain:

$$\text{Log } N_c = 5.45\ (\pm 0.20) - 0.95\ (\pm 0.05)\ M. \tag{6}$$

The "b-value" in the range 0.9 - 1.0 is very reasonable for an area such as New England. b values lying in the range 0.8 - 1.0 are found in most parts of the world (Everndən, 1970). Isacks and Oliver (1964) found a b value of 0.9 in their study of small earthquakes recorded instrumentally in New Jersey.

The errors quoted in Eqs. 5 and 6 are based only on the fit of a linear relationship to the data points. They do not include contributions from errors in the data points themselves, which are extremely hard to estimate.

RECURRENCE RELATIONS: BOSTON-NEW HAMPSHIRE REGION

Areas A and B combined (see Fig. 2) include the Boston vicinity, Northeastern Massachusetts and the associated offshore region, and the Southern half of New Hampshire. Smith lists 194 events in this active zone, which therefore accounts for just about 50% of the total activity in Southern New England. The distribution of these events in time and intensity is shown in Table 4.

Table 4. Earthquake data for Boston-Southern New Hampshire region (areas A and B combined).

Intensity	Before 1700	1700-1799	1800-1859	1860-1899	1900-1927	1928-1959
IX	-	2	-	-	-	-
VIII	1	1	-	-	-	-
VII	-	-	1	-	-	2
VI	1	-	2	1/2	1	-
V	2	6	2	4	7	1
IV	4	6	12	7-1/2	9	8-1/2
III	2	16	16	12	6	13-1/2
II	3	3	-	21	3	16

Frequency-intensity plots for these various subsets of these data show very similar features to those found in the previous section for the whole of Southern New England. Inclusion of the early data (subset 1) leads to a very nonlinear plot. Subset 3 shows some scatter due to an insufficient number of events. Subset 2 again gives the most reliable data set, and the resulting plot is shown in Fig. 5.

Linear relationships fitted to the data again have slopes in the range 0.54 to 0.60. This is an important point, for two reasons. Firstly, it substantiates the slope determined for the Southern New England region as a whole. Secondly, and more importantly, it strongly suggests that the slope (or b-value) is roughly constant throughout the area under study, within the resolution of the present data.

As before, then, we assume a slope of 0.57 ($\pm$0.03) for the frequency-intensity plot. This leads to the following recurrence relation:

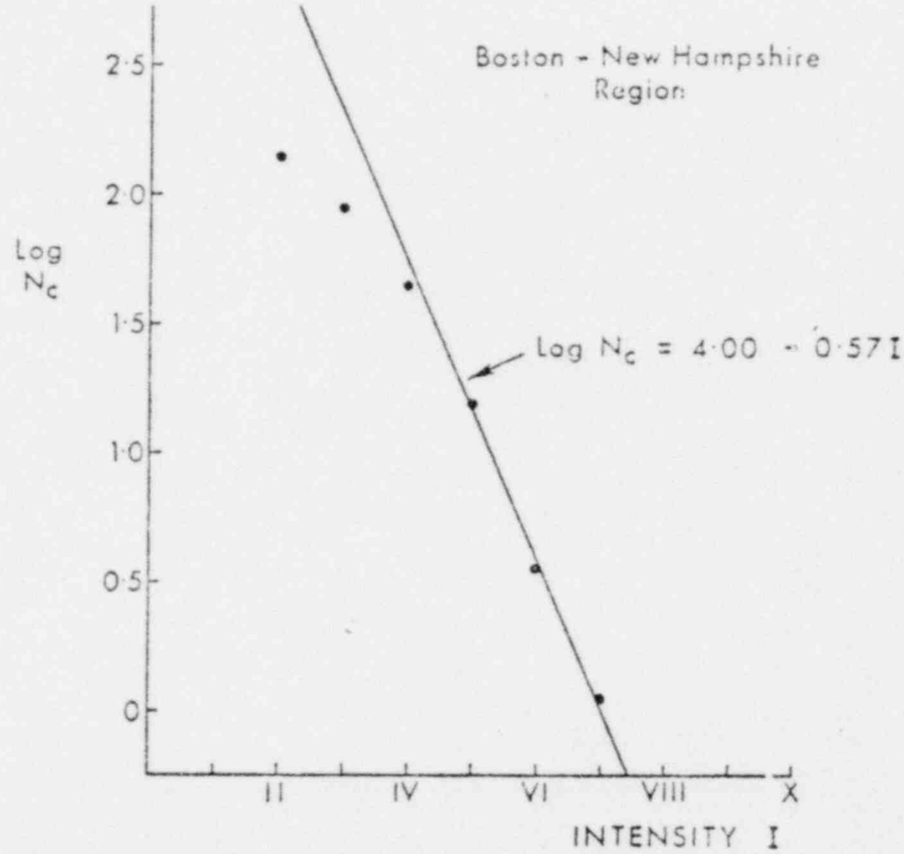

Fig. 5. Frequency-intensity plot for 65 events in areas A and B (see Fig. 2), which include the Boston vicinity and Southern New Hampshire. N_c is the cumulative number of events (with intensity I or greater) per century.

$$\text{Log } N_c = 4.00\ (\pm 0.15) - 0.57\ (\pm 0.03)\ I. \tag{7}$$

And, using Eq. 2, we find

$$\text{Log } N_c = 5.15\ (\pm 0.20) - 0.95\ (\pm 0.05)\ M. \tag{8}$$

RECURRENCY RELATIONS: AREAS A, B, C

Subdivision of the Boston-New Hampshire region into the individual subareas A and B starts to point out some of the inadequacies of the historical data set. Superficially, Smith's catalogue includes 98-1/2 events in area A, and 95-1/2 events in area B. One is tempted to ascribe one-half of the activity in the Boston-New Hampshire region to area A, and one-half to area B.

However, tabulation of the events in these two areas as functions of time and intensity shows up some marked differences. Table 5 shows this

Table 5. Earthquake data for Boston vicinity (area A).

Intensity	Before 1700	1700-1799	1800-1859	1860-1899	1900-1927	1928-1959
IX	-	2	-	-	-	-
VIII	1	1	-	-	-	-
VII	-	-	1	-	-	-
VI	1	2	2	-	-	-
V	2	5	1	1-1/2	5	1/2
IV	4	5	9	1	5	3-1/2
III	2	13	4	3	5	1-1/2
II	3	3	-	5	3	3

Table 6. Earthquake data for Southern New Hampshire (area B).

Intensity	Before 1700	1700-1799	1800-1859	1860-1899	1900-1927	1928-1959
IX	-	-	-	-	-	-
VIII	-	-	-	-	-	-
VII	-	-	-	-	-	2
VI	-	-	-	1/2	1	-
V	-	1	1	2-1/2	2	1/2
IV	-	1	3	6	4	5
III	-	3	12	9	1	12
II	-	-	-	16	-	13

tabulation for area A, and Table 6 shows the same for area B. Area A appears to have had a relatively high activity in the 1700's, which has since been steadily decreasing. On the other hand, area B appears to show a low in activity in the 1700's, which has been increasing since then.

The reality of this difference is, of course, questionable. It is likely that the New Hampshire data have been heavily influenced by the effects of population distribution, and that the earlier parts of this data set are very incomplete. This raises an interesting question. It is clear that combining the data from areas A and B leads to an estimate for the seismic activity that has been relatively uniform since the 1700's (see Table 4). Is this apparent uniformity real? It is worth while mentioning the following possibilities:

(i) The area A data may be fairly reliable, while area B may be very incomplete. Addition of the "missing" New Hampshire events will bias all of our recurrence relations in the direction of increased seismic activity. If this is the case, we have a strong indication that the seismic activity during the past 100 years or so has been unusually low. This may be the result of the statistical fluctuation, or some unknown physical process.

(ii) The early high activity in area A may be the result of exaggerated intensity estimates for some of the events in the 1700's. If this is so, it is possible that the activity of the two areas has been relatively uniform throughout the historical period.

There is no way to distinguish between these possibilities using the present data. However, the second possibility will clearly lead to the most conservative estimates for the seismic activity. Direct construction of frequency-intensity plots leads to inconclusive results because of the small number of useable events. We therefore return to our first inclination, and assume that the seismic activity of the Boston-New Hampshire region is evenly divided between areas A and B. This leads to the following estimates for the recurrence relations in areas A and B:

$$\text{Log } N_c = 3.70\ (\pm 0.15) - 0.57\ (\pm 0.03)\ I, \qquad (9)$$

$$\text{Log } N_c = 4.85\ (\pm 0.20) - 0.95\ (\pm 0.05)\ M. \qquad (10)$$

Clearly, the quoted errors are not a true reflection of the possible inaccuracies in these relations, which may be considerable. They may, however, give a more reliable estimate for the lower limit of seismic activity in the two areas.

Relatively few events have been recorded in the Hartford, Connecticut, vicinity, denoted as area C. Smith lists a total of 55 events in this area, of which only 20 fall in subset 1. This number is quite inadequate for any statistical treatment.

We may, however, get a rough idea of the activity in this region by assuming that the slope of the frequency-intensity is known (0.57 ± 0.03), and that the record of events with intensity IV is complete during the period 1900-1959.

This is sufficient to determine the following recurrence relations for area C:

$$\text{Log } N_c = 3.35\ (\pm 0.20) - 0.57\ (\pm 0.03)\ I, \qquad (11)$$

$$\text{Log } N_c = 4.50\ (\pm 0.25) - 0.95\ (\pm 0.05)\ M. \qquad (12)$$

MEAN RECURRENCE TIMES

From the recurrence relations listed in Eqs. 5 through 12, it is easy to calculate the mean recurrence times. These are listed for a variety of intensities in Table 7. It should be noted that these were determined from the cumulative event frequencies. Thus the first entry in Table 7 states that the mean interval between earthquakes with intensity VIII <u>or greater</u>, in Southern New England is about 180 years.

Table 7. Mean recurrence times (in years)

Intensity	Magnitude	Southern New England	A + B: Boston New Hampshire	Area A Boston	Area B New Hampshire	Area C Hartford
VIII	6.0-6.3	180 (±40)	360 (±80)	700 (±200)	700 (±200)	1600 (±400)
VIII-1/2	6.3-6.7	360 (±80)	700 (±200)	1400 (±400)	1400 (±400)	3000 (±1000)
IX	6.6-7.0	700 (±200)	1400 (±400)	2500 (±1000)	2500 (±1000)	6000 (±2200)
X	7.2-7.7	2500 (±1000)	5000 (±2000)	10000 (±4000)	10000 (±4000)	12000 (±6000)

It is interesting to compare these mean recurrence times with the times since the last large earthquakes in the area (Table 2). The last earthquake listed with intensity VIII occurred in 1791, just 180 years ago. Clearly, regardless of the method used to calculate future probabilities, another earthquake of this size may be expected in the near future. The subject of the determination of earthquake risk from these data will be taken up in a later paper.

It is worth emphasizing that the mean recurrence times and recurrence relations were calculated without using the large events (I > VIII) in Table 2. They are therefore independent of any errors in the intensity estimates for these large events.

CONCLUSIONS

The principal conclusions of this study may be summarized as follows:

1. The data in the Smith catalogs are consistent with a "b-value" of 0.95 ($\pm$0.05), applicable both to the Southern New England area as a whole, and also to smaller regions within this area.

2. Recurrence relations for the whole area and for certain sub-areas are listed in Eqs. 5 through 12.

3. Southern New England is likely to experience an earthquake with intensity VIII or greater in the fairly near future. The mean recurrence time for events of this size is about 180 years, while the last event of this size occurred just 180 years ago.

4. Of the total activity of Southern New England, approximately one half is concentrated in the Boston-Southern New Hampshire region (areas A and B in Fig. 2). The remainder is scattered throughout Southern New England, with a minor concentration in central Connecticut.

5. There is no evidence to suggest that there is any upper limit to the size of the earthquakes that may be expected within this area. Earthquakes of the severity of the Charleston, South Carolina, earthquake of 1886 (magnitude about 7.5, intensity about X) probably occur in Southern New England with a mean recurrence time of several thousand of years. There is no historical evidence to suggest when the last event of this size occurred.

6. Most of the large earthquakes in this area occurred during the 18th century. It is not clear if that century was unusually active, or if the last 200 years has been unusually quiet. All of the statistical conclusions in this report have been based on the data after the year 1800, and therefore do not include this earlier high activity. It is therefore possible that we have underestimated the activity of the area.

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Exhibit 2

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A COMPARISON OF THE SEISMICITY OF THREE REGIONS OF THE EASTERN U.S.*

By Michael A. Chinnery

ABSTRACT

Frequency-intensity data from the Southeastern U.S., Central Mississippi Valley, and Southern New England are compared. They are all quite parallel to one another and consistent with a slope of about 0.57. There is no evidence for the existence of upper bounds to maximum epicentral intensity in these data sets. Linear extrapolation of the frequency-intensity data to intensities of X leads to expected probabilities for the occurrence of large earthquakes. The largest events which have occurred in these three regions are consistent with these probabilities.

Introduction

Recently there have been rather detailed analyses of the seismicity of three sections of the Central and Eastern U.S. Bollinger (1973) has described an extensive set of data for the Southeastern U.S., which includes the seismically active zones of Maryland, Virginia, West Virginia, North and South Carolina, Georgia, Alabama, and Tennessee, for the period 1754 to 1970. Nuttli (1974) has listed the known events in the central Mississippi Valley seismic region for the period 1833 to 1972. And Chinnery and Rodgers (1973) have analyzed the data of Smith (1962, 1966) for the Southern New England region for the period 1534 to 1959. The purpose of this paper is to compare these three studies, and to bring out the similarities between them.

The discussion of seismic risk inevitably involves plotting frequency-intensity (i.e., maximum epicentral intensity) diagrams. In what follows we use this type of plot, since magnitude data are not available for all three regions. This raises a difficult point, since within each of these regions, the seismic activity is not uniform. The selection of the boundaries of the area to be studied is much akin to the problem of the definition of a tectonic province (which is required, for example, by the Nuclear Regulatory Commission Rules and Regulations, Part 100, Appendix A).

For the moment, we shall make the following assumptions: First, we assume that all subregions within a given region have a linear frequency-intensity relation of the form

$$\log N_i = a_i - bI$$

where N_i is the cumulative number of events in the ith subregion with intensities greater than or equal to I, and a_i is a parameter describing the level of seismic activity of the ith subregion. We assume that the slope b is common to all subregions. Second, we assume that the maximum possible intensity in each subregion, if one exists which is lower than the nominal maximum of XII, is larger than the largest event recorded within that subregion during the period of the earthquake record.

These assumptions sound very drastic, yet they are really implicit whenever we plot a frequency-magnitude or frequency-intensity curve. Furthermore, at least in

* The views and conclusions contained in this document are those of the contractor and should not be interpreted as necessarily representing the official policies, either expressed or implied, of the United States Government.

principle, they are testable. It is easy to plot frequency-intensity diagrams for portions of a region and examine both the linearity of the results and the constancy of the slope b. In practice, of course, scatter in the data often makes such a test inconclusive. However, a substantial breakdown of any of the above assumptions should be apparent in the data for the region as a whole, either by the appearance of nonlinearity in the frequency-intensity statistics, or by variations in estimates of b using different data sets. As we examine and compare the seismicity of the three areas under consideration, we shall look for information related to these assumptions.

Perhaps the most important question which we shall address is as follows: Each of these areas has had one moderately large earthquake in its recorded history (the 1755 Cape Anne, 1811-1812 New Madrid, and 1886 Charleston events). Are these large events consistent with the record of smaller earthquakes that have occurred more recently? Clearly, this question has a direct bearing on the very fundamental problem of how to extrapolate from a short record of seismicity to the occurrence of low probability events, which is particularly important in the assessment of the potential seismic hazard to critical structures such as nuclear power plants.

We shall disregard questions of the lack of stationarity of the earthquake process in these three areas, in spite of their potential importance (Shakal and Toksoz, 1977). It is very difficult to document this nonstationarity within time periods of 100 to 150 years, because of the small number of events concerned.

The Data

Southeastern U.S. Bollinger (1973) describes the seismicity of four seismic zones in the Southeastern U.S. for the period 1754 to 1970 (see Figure 1). In this study we shall restrict ourselves to the two southernmost zones, the Southern Appalachian seismic zone and the South Carolina-Georgia seismic zone. The combined area of these two zones is given by Bollinger to be 307,000 km^2. Since we would like to exclude the 1886 Charleston earthquake from consideration, we have analyzed events during the period 1900 to 1969. Even this period is probably too long for the adequate recording of intensity III events, so these have been accumulated for the period 1930 to 1969 only. Total events listed by Bollinger (1973) are shown in Table 1.

These data are easily converted into a cumulative frequency-intensity plot, and this is shown in Figure 2. The usual interpretation of such a diagram is to fit the data points with a straight line, recognizing that the data at the lower intensities is likely to be incomplete. Such a fit is shown as the solid line in Figure 2. This line corresponds to the equation

$$\log N_c = 2.31 - 0.46I. \tag{1}$$

The slope of this line is low compared to other similar regions, as we shall see below. The occurrence of three intensity VIII events during this 70-year period seems high, and in fact one of them has been shown to be an explosion (G. A. Bollinger, personal communication). Certainly a line such as the dashed line in Figure 2, which has the equation

$$\log N_c = 2.88 - 0.55I \tag{2}$$

cannot be ruled out. The slope of 0.55 in this equation is very close to the slope 0.56

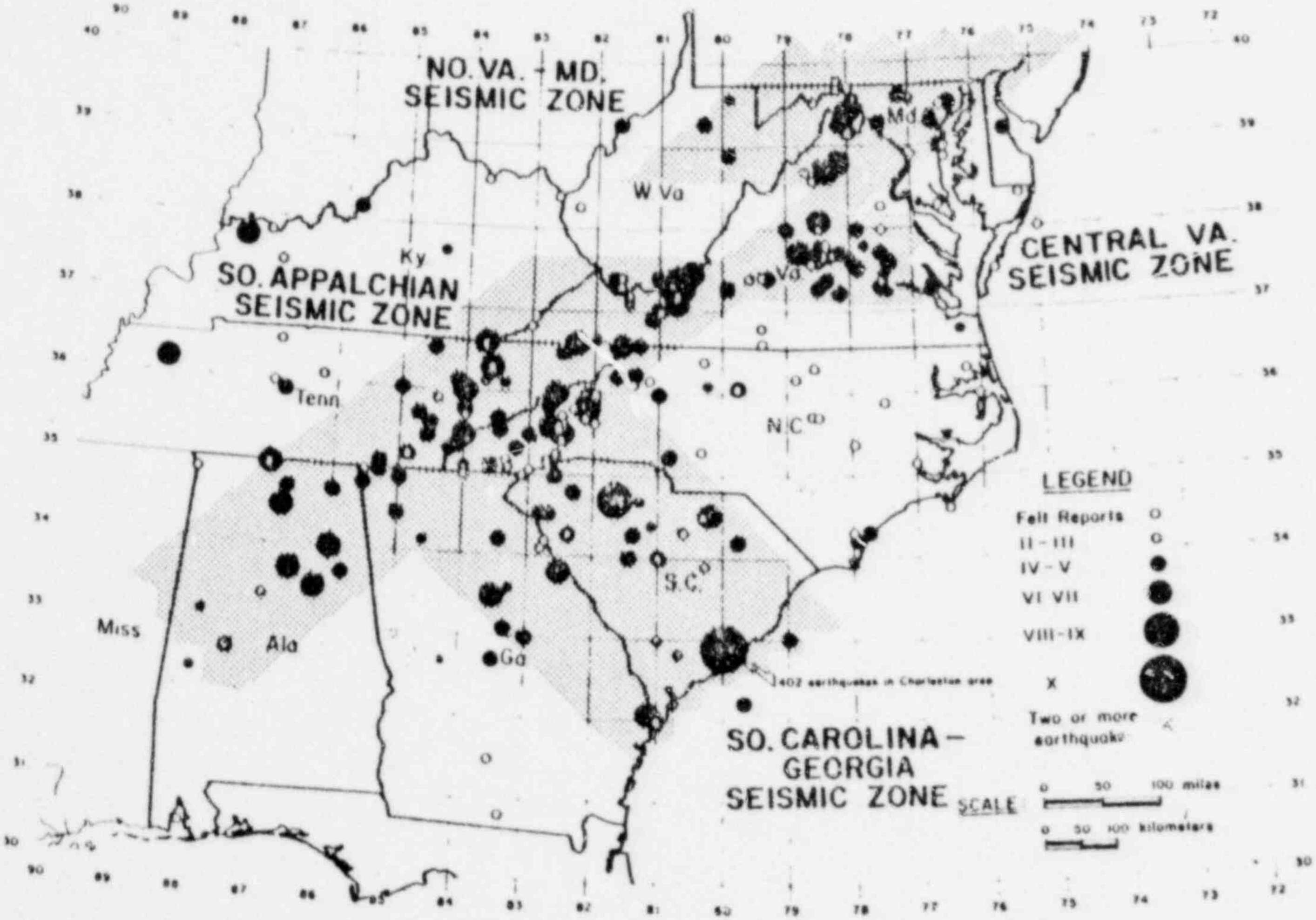

FIG. 1. Seismicity in the Southeastern U.S. for the period 1770 to 1959. Reproduced, with permission, from Bollinger (1973).

± 0.08 found by Bollinger (1973) for the whole Southeastern U.S. For the moment, we will retain both equations (1) and (2) as possible interpretations of the data.

Central Mississippi Valley. Nuttli (1974) has given a list of events in the central Mississippi Valley for the period 1833 to 1972. The epicenters of these events are shown in Figure 3. The total area of this zone is given by Nuttli to be 250,000 km^2. Since he lists few events before 1840, we have restricted ourselves to the period 1840 to 1969. Table 2 lists the events during this period as a function of intensity. As

TABLE 1
EVENTS IN SOUTHERN APPALACHIAN AND SOUTH CAROLINA-GEORGIA SEISMIC ZONES

Intensity	Period	No. of Events
III	1930–1969	10
IV	1900–1969	49
V	1900–1969	46
VI	1900–1969	17
VII	1900–1969	5
VIII	1900–1969	3

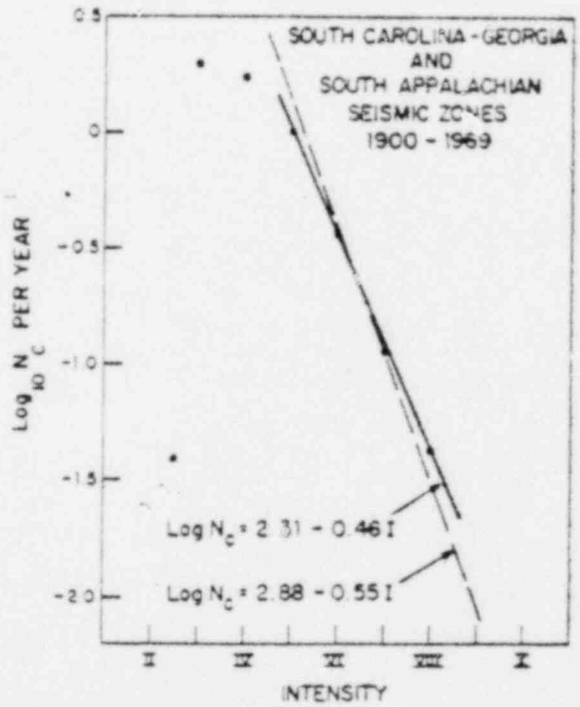

FIG. 2. Cumulative frequency-intensity plot for the data in Table 1. Two possible straight line interpretations are shown.

before, smaller events are only counted for the more recent portion of this time period. Since many events are listed with intensities intermediate between two values (such as III to IV), where this occurs one-half event has been accumulated into each value. This accounts for the fractional events listed in Table 2.

Figure 4 shows a cumulative frequency-intensity plot for the data in Table 2. A reasonable linearity is obtained, corresponding to the equation

$$\log N_c = 2.77 - 0.55I. \quad (3)$$

Southern New England. The seismicity of Southern New England has been discussed by Chinnery and Rodgers (1973), using data of Smith (1962, 1966) for the period 1534 to 1959. The region defined as Southern New England is shown by the solid line in Figure 5, which also shows the epicenters in Smith's listing. Following Chinnery and Rodgers (1973), we note that many of the listed epicenters are

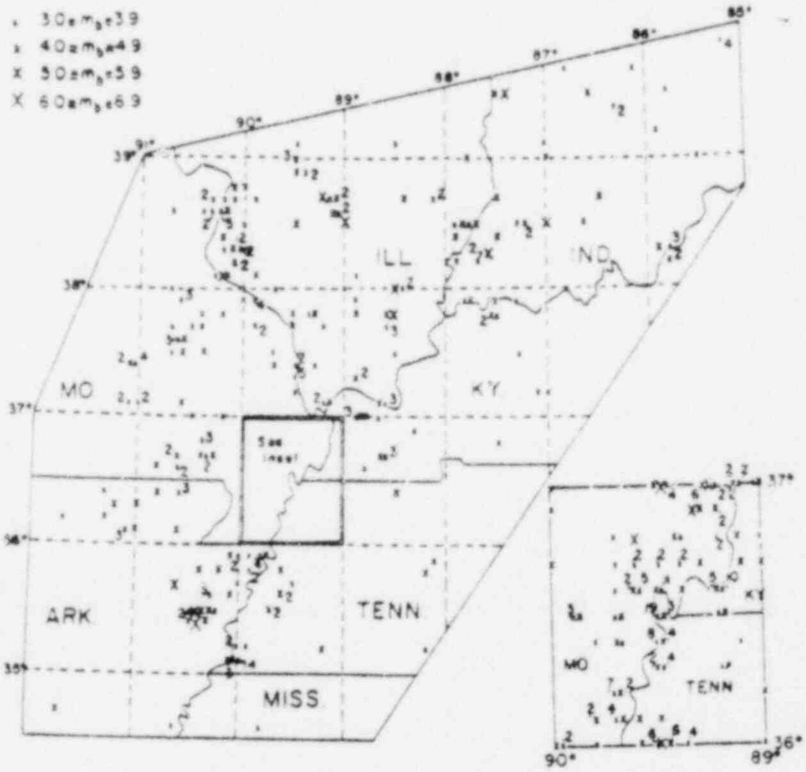

FIG. 3. Epicenters in the central Mississippi Valley region, for the period 1833 to 1972. Reproduced, with permission, from Nuttli (1974).

TABLE 2

EVENTS IN CENTRAL MISSISSIPPI VALLEY

Intensity	Period	No. of Events
II	1930–1969	22.5
III	1900–1969	94.5
IV	1870–1969	143.5
V	1870–1969	63.0
VI	1840–1969	31.5
VII	1840–1969	10.5
VIII	1840–1969	1.0
IX	1840–1969	1.0

clustered in a region extending from Boston through central New Hampshire. We have outlined this area in Figure 5, and refer to it as the Boston-New Hampshire seismic zone. The areas of the two zones in Figure 5 are approximately 100,000 km^2 (Southern New England) and 27,000 km^2 (Boston-New Hampshire zone). Since we wish to exclude the 1755 Cape Anne earthquake from the data set, events have been accumulated in both the Southern New England region and the Boston-New Hampshire zone for the period 1800 to 1959. These are listed in Tables 3 and 4, respectively. As before, small events are only accumulated for the most recent portion of the record.

The cumulative frequency-intensity plot for Southern New England is shown in Figure 6. The straight line through the data has the form

$$\text{Log } N_c = 2.36 - 0.59I. \tag{4}$$

In spite of the rather low numbers of events, this line is a reasonable fit to the data. In the case of the Boston-New Hampshire zone, however, the number of events

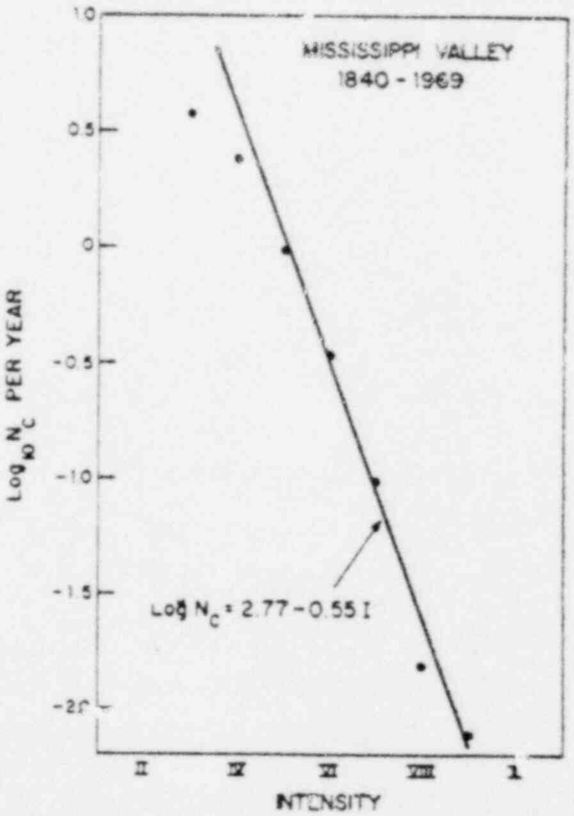

FIG. 4. Cumulative frequency-intensity plot for the data in Table 2.

becomes low enough that it becomes difficult to formulate a linear fit with any certainty. A straight line through the upper four data points has a shallow slope (about 0.50), which is significantly different from the other areas studied, and which leads to high estimates of risk for large events. We prefer to interpret these data with a line such as the one shown, which has the equation

$$\log N_c = 2.15 - 0.59I. \tag{5}$$

With this interpretation, the number of intensity VII earthquakes is anomalously high, due either to poor data or a statistical fluctuation. At least equation (5) should lead to reasonably conservative estimates for risk at high intensity levels.

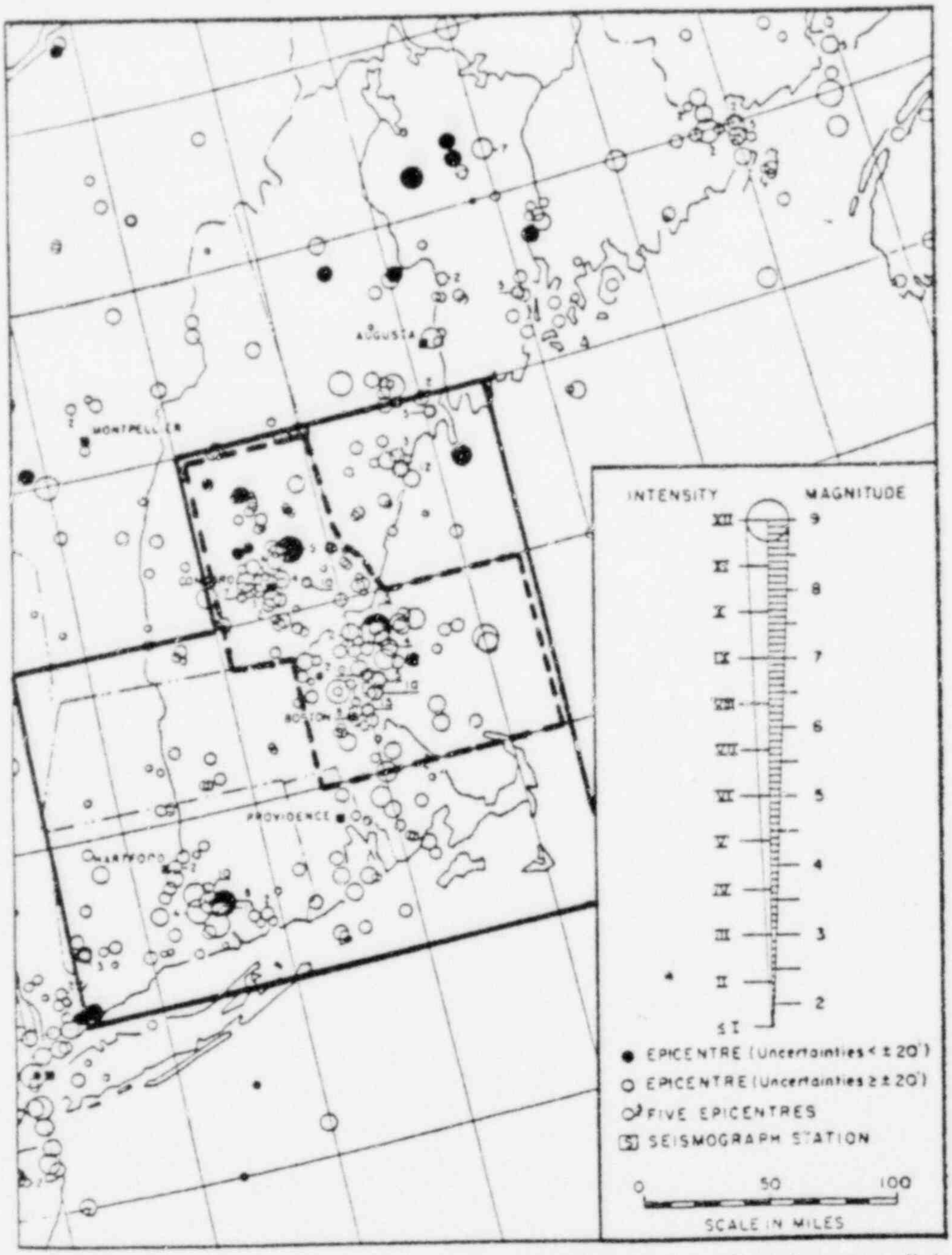

FIG. 5. Epicenters in New England, from Smith (1966). The solid line outlines the region called Southern New England in this study. The broken line indicates the Boston-New Hampshire zone (see Chinnery and Rodgers, 1973).

Comparison of Frequency-Intensity Data

The frequency-intensity data shown in Figures 2, 4, 6, and 7 are shown together in Figure 8. In this case we have omitted the individual interpretation using fitted straight lines, and show the data alone. This emphasizes the very similar character of the four recurrence curves. There is some scatter, but each of the curves is consistent with a slope somewhere in the range 0.55 to 0.60, and we show a slope of 0.57 which seems to be a reasonable average.

In view of the rather inferior quality of much historical intensity data, it is surprising how consistent the slopes of cumulative frequency-intensity data appear

TABLE 3

EVENTS IN SOUTHERN NEW ENGLAND

Intensity	Period	No. of Events
II	1928–1959	32.5
III	1928–1959	26.5
IV	1900–1959	43.0
V	1860–1959	24.0
VI	1800–1959	6.0
VII	1800–1959	3.0

TABLE 4

EVENTS IN BOSTON-NEW HAMPSHIRE ZONE

Intensity	Period	No. of Events
II	1928–1959	16.0
III	1928–1959	13.5
IV	1900–1959	17.5
V	1860–1959	12.0
VI	1800–1959	3.5
VII	1800–1959	3.0

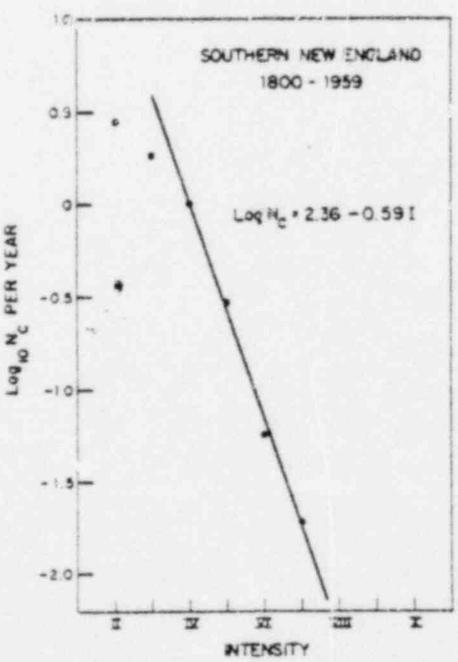

FIG. 6. Cumulative frequency-intensity plot for the data in Table 3.

to be. Both Connell and Merz (1975) and Veneziano (1975) have surveyed a number of estimates of this slope, and many of these are consistent with the present data. The mean of the 11 estimates quoted by Veneziano is 0.53, but his list contains some low values which are probably not realistic. Of particular interest are the values 0.59 for the whole U.S. (Connell and Merz, 1975) and 0.54 for California (Algermissen, 1969). A recent estimate for the area around the Ramapo fault in New York and New Jersey is 0.55 ± 0.02 (Aggarwal and Sykes, 1978).

It is interesting to compare a slope of 0.57 with the value that one would predict from known magnitude-intensity relationships. A selection of these relationships have been given by Veneziano (1975), in the form

$$M = a_1 + a_2 I. \tag{6}$$

Values of the constant a_2 have been estimated as 0.67 (Gutenberg and Richter, 1956), 0.69 (Algermissen, 1969), and 0.60 (Chinnery and Rodgers, 1973; Howell, 1973). The latter estimates of 0.60 were obtained from data in the Eastern U.S., and may be the best estimates for our present purposes.

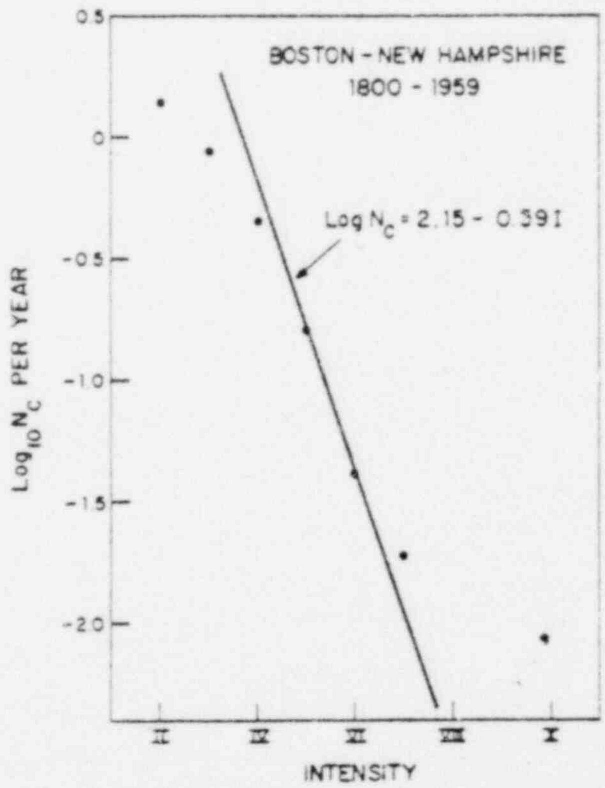

FIG. 7. Cumulative frequency-intensity plot for the data in Table 4.

There is an abdunance of frequency-magnitude data, which is usually represented by the form

$$\log N_c = a - bM \tag{7}$$

where the slope b often lies between 0.9 and 1.0 (see, for example, Chinnery and North, 1975). Combining this expression with equation (6), with $a_2 = 0.60$, would lead to a slope of the frequency-intensity relation between 0.54 and 0.60. Clearly the

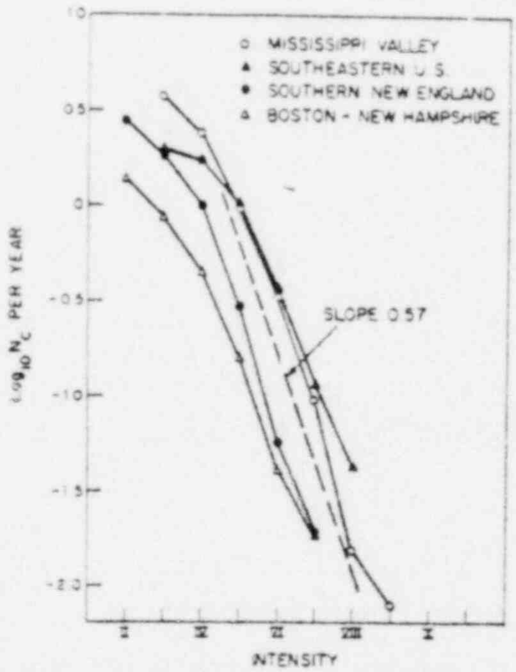

FIG. 8. Comparison of the frequency-intensity data from Figures 2, 4, and 7.

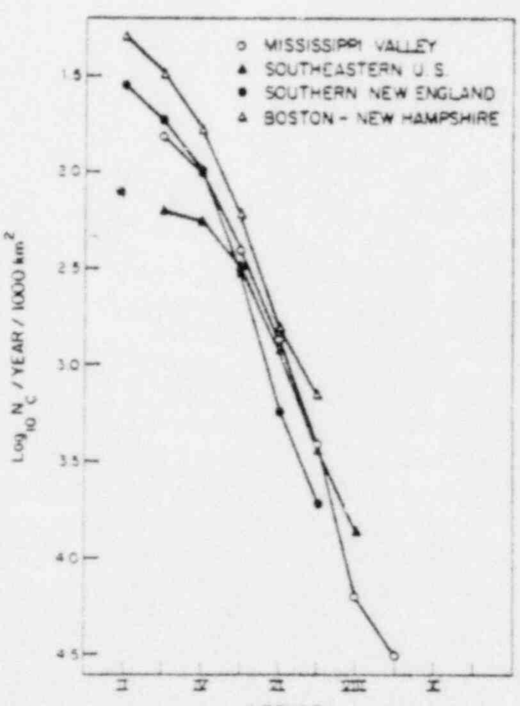

FIG. 9. The same data used in Figure 8, but normalized for the areas of the various zones.

0.57 value shown in Figure 8 is eminently reasonable and consistent with other information.

The similarity between the four sets of data shown in Figure 8 can be further emphasized by normalizing for the areas of the seismic : gions. After this normalization, Figure 9, the recurrence curves are found to lie almost on top of one another (we have chosen to normalize to 1,000 km², but this choice is completely arbitrary). The apparent similarity in seismic activity per unit area is entirely fortuitous, and is simply due to the particular regions chosen for each study. The true levels of activity in the three regions differ markedly (see, for example, the return periods calculated in Table 5). However, one is tempted to note that the activity per unit area in the Boston-New Hampshire zone is slightly larger than that in the Southeastern U.S. Is there really any good reason why an event the size of the Charleston earthquake could not occur in the Boston-New Hampshire zone?

It is interesting to search these data sets for evidence that there may be an upper bound intensity in some of these areas. Cornell and Merz (1975), for example, have proposed a frequency-intensity curve for a site in the Boston area that curves downward and becomes vertical (parallel to the ordinate axis) close to intensity VII. Since this calculation is for a single site, it is crucially dependent on our ability to predict the location of large events near Boston. Certainly, if large events could occur anywhere within the Boston-New Hampshire zone, the present data show no indications of an upper bound. Given our present knowledge concerning the mechanisms of large events in regions like the Boston-New Hampshire zone, it does not seem reasonable to propose such an upper bound.

Randomness of the Catalogs

Before attempting to calculate the risk of large events in the three areas under consideration, we should briefly address the nature of the statistical model to be used. It is usual to assume that catalogs such as these are random, i.e., described by the simple Poissonian distribution.

This problem has received ample treatment in the literature (see, for example, Lomnitz, 1966). In some cases the Poisson distribution has been shown to be a good description for large events, Epstein and Lomnitz (1966), and Gardner and Knopoff (1974) have shown that the Southern California catalog, with aftershocks carefully removed, is Poissonian. Other studies have indicated departures from Poisson statistics (e.g., Aki, 1956; Knopoff, 1964; Shlien and Toksoz, 1970). However, these departures are small, and may be disregarded for our present purposes.

One graphic method of demonstrating the approximately Poissonian character of a sequence of earthquakes is to plot the interoccurrence times (Lomnitz, 1966). In a purely Poisson process, the probability P that an interval of time T will contain at least one event is given by

$$P(T) = 1 - e^{-T/T_0}. \tag{8}$$

Here T_0 is the mean return period for events in the sample.

If the time between events in the sample is the variable t, then the frequency distribution of t is given by

$$F(t) = \frac{1}{T_0} e^{-t/T_0}. \tag{9}$$

It is easy to show that the observed interoccurrence times are quite closely represented by equation (9). Figure 10 shows a plot of these interoccurrence times for the central Mississippi Valley catalog for events with intensity greater than or equal to V during the period 1900 to 1972. Clearly, the exponential distribution is a good description of the data. The anomalously large number of events at small interoccurrence times can be attributed primarily to the presence of aftershocks in the catalog. A similar plot for Southern New England data is shown in Figure 11. Data from the Southeastern U.S. were not available in a form that would permit a similar plot to be made, but this is probably not necessary. On the basis of Figures 10 and 11, we feel justified in using the Poisson model, and in particular equation (8), to calculate probabilities.

In passing, Figures 10 and 11 make another point. It is easy to use the quantity mean return period of earthquakes in a sequence as if it has a deterministic meaning. These figures are a reminder that the mean return period is entirely a statistical

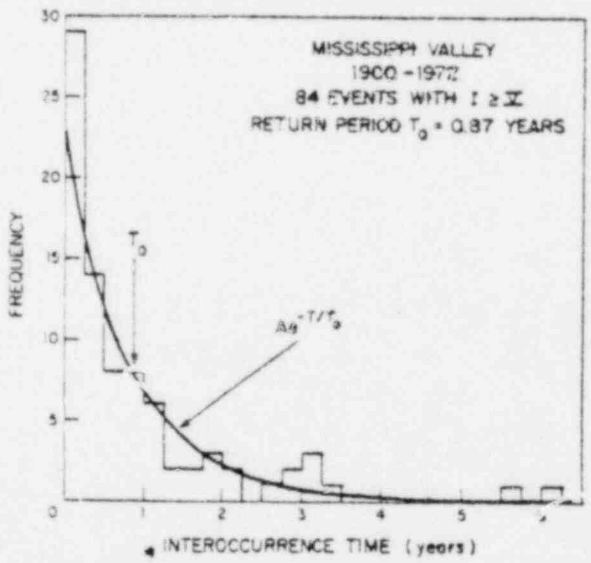

FIG. 10. Interoccurrence times using Nuttli's (1974) data for the central Mississippi Valley. The exponential curve would be expected for a Poisson distribution.

quantity, and that its only real meaning is as one of the parameters describing the probability distribution that corresponds to the catalog under consideration.

THE PROBABILITY OF LARGE EVENTS

With the above model it is now possible to address the question posed in the introduction. In each of the three areas under consideration a large earthquake occurred shortly before the periods of data that we have analyzed. Are these large earthquakes consistent with the later record of smaller events?

Our procedure is simple. We take the linear relations fitted to the frequency-intensity data, extrapolate them to larger intensities, and make estimates of the mean return periods of these larger intensities. We then use equation (8) to estimate the probability that at least one of these larger events will occur in any 200-year period, and specifically relate this to the 200-year period ending at the present time (a 300-year period was chosen for New England, since the largest event occurred in the 1700's).

The results ar shown in tabular form in Table 5. We do not pretend that these numbers are very accurate. In fact, because of the subjectivity that has to be used in obtaining the linear relations [equations (1) to (5)], there is no way to make a realistic assessment of errors. We therefore view the numbers in Table 5 as being a qualitative indication of risk, rathe than quantitative. The results for the individual areas are discussed below.

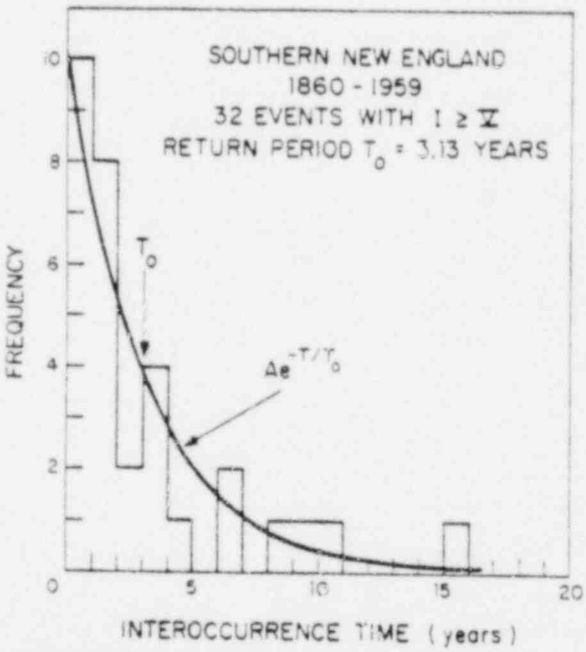

FIG. 11. Interoccurrence times for Southern New England from the data of Smith (1962, 1966).

TABLE 5

PROBABILITY OF LARGE EVENTS IN FOUR REGIONS OF THE EASTERN U.S.

Area	Equation Used (see text)	Return Period (years)			Time Before Present T Years	Probability of at Least One Event in Period T (%)		
		≥VIII	≥IX	≥X		≥VIII	≥IX	≥X
Southeastern U.S., 1900–1969	1	23	68	195	200	99	95	64
	2	33	117	417	200	99	82	38
Mississippi Valley, 1840–1969	3	43	151	537	200	99	73	31
Southern New England, 1800–1959	4	229	891	3467	300	73	29	8
Boston-New Hampshire, 1800–1959	5	371	1445	5623	300	55	19	5

The earthquake catalog for the Southeastern U.S. described by Bollinger (1973) is approximately 200 years long. Table 5 shows that, on the basis of the most recent 70 years of this catalog (which may logically be expected to be the most complete at lower intensities), there is a substantial probability of the order of 50 per cent that at least one earthquake of intensity X or greater will occur in a 200-year period. We conclude, therefore, that the Charleston earthquake of 1886 (intensity X, Bollinger, 1977) is entirely consistent with the 1900 to 1969 data.

Without any question the largest earthquakes during the past 200 years in the central Mississippi Valley were the 1811 to 1812 New Madrid events. Nuttli (1973) lists the maximum observed intensity during this sequence as X to XI, at New Madrid, Missouri; Gupta and Nuttli (1976) have recently revised this upward to XI to XII. Some question perhaps remains as to the validity of this value as a true epicentral intensity, since some amplification by the alluvium in the area might be expected. Table 5 lists the probability of an event of intensity X or greater during a 200-year period as being about one-third. The New Madrid events were therefore reasonably consistent with the data for 1840 to 1969. If it could be shown that these were the largest events in the last 300 years in this area (which is not unlikely), or that the true epicentral intensity was somewhat less than X, it would be easy to increase the calculated probability to 50 per cent or more.

The record of earthquakes for Southern New England is about 300 years long (Smith, 1962, 1966). During the period 1800 to 1959, Smith lists 3 events with intensity VII, and there are none any larger. Table 5 shows that there is a respectably high probability (about 75 per cent) that an earthquake of intensity VIII will occur somewhere in Southern New England in a 300-year period. The probability of such an event in the Boston-New Hampshire zone is about 50 per cent. The epicentral intensity of the 1755 Cape Anne earthquake is not well defined. Smith (1962) lists this event as intensity IX, which is probably somewhat high. *The Earthquake History of the United States* (NOAA publication 41-1, 1973) lists this event as intensity VIII. Other unpublished studies have deduced intensities close to VII. Whichever is correct, it cannot be said that this event is inconsistent with the subsequent seismic record.

An equally important result for the Southern New England region is that the probability of intensity IX and X events occurring within a 300-year period is quite low. The absence of these events in the historical record is therefore again consistent with the 1800 to 1959 data. Notice, too, that the return period for intensity VIII is 229 years, which is consistent with the absence of such an event during the period 1800 to 1959.

Conclusion

We can make several conclusions from this study

1. The four frequency-intensity plots that we have considered show a remarkable uniformity. All show a pronounced linearity, and have slopes which are consistent with a value of about 0.57. This, in turn, corresponds to a magnitude *b*-value in the range 0.9 to 1.0. This uniformity, and the fact that 0.57 is very close to slopes observed in other areas of both Eastern and Western U.S., suggests that frequency-intensity data can usefully be applied in seismic risk analysis. In areas where data are poor or sparse, it would appear possible to combine data from as little as one intensity value with the apparently universal slope of about 0.57 to construct a local frequency-intensity relationship. Such a procedure may be more reliable than some of those in current use.

2. The uniformity of the shape of the frequency-intensity relation over regions ranging from the Boston-New Hampshire zone and the Ramapo fault zone (Aggarwal and Sykes, 1978) to the whole of the continental U.S. suggests that the problem of nonuniformity of seismicity within a region is no impediment to the use of frequency-intensity statistics. The assumptions outlined in the introduction to this paper seem to be useful working hypotheses.

3. The question of the existence of upper bounds to maximum earthquake intensity (less than the scale maximum of XII) remains unanswered. There is no reason within the data themselves to suggest that the three large events that we have considered are the largest that could occur in these regions. Similarly, there are no statistical arguments that a very large event could not occur in other areas (such as Southern New England outside of the Boston-New Hampshire zone) that have not recorded such an event. A rational, conservative approach to the estimation of the seismic risk at a site would include the possibility of events with intensity X or more anywhere in the Eastern U.S. This topic will be discussed more fully elsewhere.

4. The validity of linear extrapolation of the frequency-intensity data has been tested by predicting the probability of occurrence of large earthquakes in the historical record, and comparing this probability with the known occurrence of large earthquakes in each of the three areas. The Charleston and Cape Anne earthquakes are both consistent with more recent data from small events (calculated probabilities of these events are 50 per cent ore more). The New Madrid sequence is only slightly anomalous. The chance that such an event would occur during the past 200 years is about 30 per cent, but the chance that it would occur in a 300-year record approaches 50 percent. Thus, it appears that linear extrapolation of frequency-intensity data to intensities of IX and X is a valid procedure in these areas.

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Manuscript received October 17, 1978

Exhibit 3

Annual Report

M. A. Chinnery

Investigations of the Seismological Input to the Safety Design of Nuclear Power Reactors in New England

15 August 1978

Prepared for the Nuclear Regulatory Commission under Contract NRC-04-77-019

Lincoln Laboratory

MASSACHUSETTS INSTITUTE OF TECHNOLOGY

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Massachusetts Institute of Technology

Lincoln Laboratory

AN INVESTIGATION OF MAXIMUM POSSIBLE EARTHQUAKES

Annual Report

Project Title: Investigations of the Seismological Input to the Safety Design of Nuclear Power Reactors in New England.

NRC Contract: NRC-04-77-019

Principal Investigator: Michael A. Chinnery, Group Leader
Applied Seismology Group
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Period of Contract: 1 January 1977 – 31 December 1977

15 August 1978

Abstract

This report describes research carried out under NRC Contract NRC-04-77-019 during the period 1 January 1977 to 31 December 1977. A detailed study of available scientific literature concerning the estimation of maximum possible earthquakes shows that all available methods are empirical and lack a sound physical basis. Evidence that even the empirical methods are valid is very weak, primarily because of the short length of the earthquake record in most areas. An attempt to use global earthquake catalogs to examine the regional variation of maximum possible earthquakes is unsuccessful. It is demonstrated that saturation of the magnitude scale and biases introduced by instrumental clipping combine to make m_b values for large earthquakes very unreliable, and to obscure the presence or absence of maximum possible earthquakes. A progress report on a study of New England crust and upper mantle structure is included.

Table of Contents

Introduction

This report describes research carried out under NRC Contract NRC-04-77-019 during the period 1 January, 1977 to 31 December, 1977. The major effort during this period consisted of two studies aimed at evaluating the possibility of estimating the maximum possible earthquake that might be expected within a given region.

The first study consisted of a review and assessment of available scientific literature on this topic. Since much of the research in this area has been carried out in the Soviet Union, this review provides a reasonably comprehensive set of references, and a discussion of the various approaches which have been tried.

The second study was an attempt to look for evidence of upper bounds to earthquake size within global body wave magnitude catalogs, and in particular in the ISC catalog. This study soon turned into an attempt to understand the sources of bias in the magnitudes listed in this catalog, since until these are understood it is impossible to search for maximum possible events. It transpires that these biases, together with saturation of the m_b scale, make m_b catalogs essentially useless for this type of study.

A third area of research, into the crust and upper mantle structure of New England, got underway during the period covered by this report, and a progress report is included in the Appendix.

1. MAXIMUM POSSIBLE EARTHQUAKES: CURRENT STATUS

1.1 Introduction

We would like to know whether or not there is a limit or "upper bound" to the size of earthquakes for a variety of reasons. First, earthquake size is usually intended to be a measure of energy release. However, energy usually varies strongly with size. For example, the standard relation between magnitude M and energy E (in ergs) is

$$\log E = a_o + b_o M \tag{1.1}$$

Bath (1966) reviews several estimates for the constants a_o and b_o, and shows that b_o appears to lie in the range 1.4 to 2.0. Since the number N of earthquakes is usually described by the relation

$$\log N = a - bM \tag{1.2}$$

where b is about 1 (see, for example, Richter 1958), the total seismic energy release is dominated by the largest events. We shall have reason to question both equations 1.1 and 1.2 later in this report, but the conclusion appears to remain valid. Analysis of the energy budget of the earth requires knowledge of the rate of occurrence and energy release in the largest events that occur.

Second, Brune (1968) has shown how the relative slip of tectonic plates can be estimated from earthquake size, and showed that the total slip is dominated by the largest events that occur. The fundamental question of how much tectonic motion is released in seismic slip (Davies and Brune, 1971) can only be answered clearly once we understand these large events.

And, thirdly, the estimation of maximum earthquake size is important in the estimation of seismic risk. The possibility that large events may occur, even infrequently, in an area can lead to a seismic hazard that is unacceptable for certain critical facilities such as nuclear power plants. The NRC Rules and Regulations, Part 100, Appendix A, set out the seismic safety standards for these structures, and define the Safe Shutdown Earthquake to be based on an evaluation of the "maximum earthquake potential" of an area (Hofmann, 1974). The purpose of the present study is to assess our ability to estimate this quantity.

We can usefully divide the overall problem into two parts. First, what is the evidence that earthquakes considered as a global phenomenon have a maximum possible size? And second, how does this maximum possible size vary from region to region? The first question ought to be much simpler to answer than the second, and it is logical to examine it first. However, as we shall see, it is difficult to give convincing answers to either of these questions.

1.2 Definitions

There are two important definitions that we must explore before we continue. The first is the definition of "maximum", and the second is the definition of "size".

The term "maximum" is not, unfortunately, always used with the same meaning. One definition is the obvious one, which refers to the largest possible event that can occur given the physical conditions of the source area. A second definition, sometimes used, includes the concept of probability. A certain probability level may be accepted as being "negligible", according to engineering design standards or other arguments, and the "maximum" earthquake defined as one which will occur with this probability level (or less) during the projected lifetime of a structure.

These two definitions are very different, and it is essential that they be clearly distinguished from one another. We shall use the terminology

M_{max} for the "true" maximum possible magnitude (E_{max} for the maximum possible energy, etc), and M^P_{max} for the magnitude that occurs with probability P, which defines the accepted probability of "negligibility". As we shall see in the next section, very different methods must be used in the estimation of M_{max} and M^P_{max}.

The definition of earthquake "size" is even more difficult. There are a large number of quantities which attempt to measure this size. A partial list includes:

a) Body wave magnitude (m_b)

b) Surface wave magnitude (M_s)

c) 100 second period magnitude

d) seismic moment (M_o)

e) radiated seismic energy

f) elastic potential energy release

g) maximum epicentral intensity (I)

h) maximum epicentral acceleration

i) local magnitude (M_L)

The basic problems here are not only to decide which of these measures of size are the most appropriate for a given situation, but to recognize that the relationships between these measures are in general poorly understood and in some cases demonstrably very non-linear. In particular, some of these quantities have built-in upper bounds which can obscure the search for a fundamental upper limit to earthquake size. We shall examine this problem in more detail in section 2.

An additional complication, which arises in the literature very frequently, is that the term magnitude is so often used without proper definition. All practical measures of magnitude are restricted to some limited portion of the seismic spectrum, and are closely tied to the method of measurement employed. There is so much variability in both of these factors that the term magnitude alone is almost meaningless, particularly when the characteristics of large earthquakes are concerned.

Quite often, in reference to the local seismicity of area, the term magnitude refers to local magnitude M_L. Of all measures of magnitude this is one of the hardest to quantify. It was introduced originally by Richter, and designed for local shocks in California. Its definition is very arbitrary, and refers to the logarithm of the maximum recorded trace amplitude of a specific instrument (Wood-Anderson seismograph) at a specific distance (100 km). Because the instrument will record a wide range of frequencies in the short period band, and because there is no seismic phase identification, the significance of the maximum trace amplitude is not clear. For small earthquakes, the maximum trace amplitude will often refer to body wave arrivals at short distances. For large earthquakes, the maximum trace amplitude will usually be associated with fundamental mode or higher mode (L_g phase) surface waves.

The principal usefulness of M_L is, of course, that it is a measure of ground motion in the near field at a range of frequencies that are relevant to engineering considerations. Improvements in the estimation of M_L (Kanamori and Jennings, 1978) may lead to a more consistent scale, but its relation to far field magnitude determinations is still unclear.

1.3 Approaches to the Problem

The number of papers in the literature that attempt to get to the heart of the problem of the estimation of the maximum possible earthquake is quite small. The majority of these are the work of scientists in the USSR, where there has been a long term interest in this topic.

Unfortunately some of these papers are hard to obtain and difficult to read.

A number of approaches to the problem have been proposed (see, for example, Shenkova and Karnik, 1974). First, there are a number of broad arguments that attempt to limit the upper size of earthquakes on the basis of physical principles, including fault geometry and slip, and the strength of earth materials. Generally speaking, these arguments make a convincing case in favor of a global upper bound, but give little indication where this might be. A second approach uses earthquake statistics, either in the form of frequency-magnitude data or modelled by the theory of extremes. These two analytical techniques generally lead to similar results, but both turn out to be severely limited by the definitions of magnitude used. A third approach, which seems very logical yet which lacks any convincing physical basis, attempts to relate the size of the maximum possible earthquake to the level of seismic activity in a region. It would be very nice if such a relationship were to exist, but there is no clear evidence that it does. More recent approaches have tended to focus on information from non-seismic sources, such as geological and geomorphological data. Some of these approaches are statistical, using pattern recognition techniques. Others are more deterministic, and attempt to link long term geological fault movement to short term earthquake slip.

In virtually all of these approaches one problem predominates. The record of earthquakes is relatively short in most parts of the world. Data before about 1900 are generally qualitative and hard to interpret. Adequate seismic networks have only been available since the early 1960's, and (as we shall see in section 2) there are still problems in defining the size of large earthquakes. It therefore becomes very difficult to establish empirical data for maximum possible earthquakes in specific regions, since these largest events may have return periods of 1000 years or more. Without these empirical estimates, it is virtually impossible to examine the validity of many proposed approaches.

1.4 Physical Arguments

There seems to be universal argreement that any measure of size of an earthquake must have an upper bound. This argument is often intuitive, but it can be refined to some extent. Certainly equations 1.1 and 1.2 cannot both be valid for indefinitely large M, since this would imply an infinite release of seismic energy per unit time (Newmark and Rosenblueth, 1971). However, both of these equations are poorly defined at large magnitudes, so the argument is not too helpful.

Intuition is often carried into the discussion of regional upper bounds. Newmark and Rosenblueth (1971) remark that earthquakes with $M > 9$ in the continents and $M > 7$ under the deep oceans are unlikely, though they admit there is no real basis for these estimates. In fact, if M is surface wave magnitude M_s, we shall see that M probably does not exceed about 8.6 anywhere, but this is an artifact of the magnitude scale and not a true upper bound (section 2.3). Earthquakes of $M_s > 7$ have been observed several times on the mid-ocean ridges, where the activity is low.

Sometimes intuition is quantified by the use of Bayesian statistics Connell and Merz (1974, 1975) propose an upper bound to earthquake epicentral intensities in the Boston area on the basis of a presumption that such an upper bound exists, and "conversations with seismologists". The resulting seismicity curve is used to estimate seismic risk in this area (see also Esteva, 1969; Veneziano, 1975). It seems likely that this study reflects a general belief that areas of low seismicity should have low upper bounds to earthquake size (see section 1.6).

It is possible to go somewhat beyond intuition. Tsuboi (1956) has proposed an upper bound to earthquake energy. He first relates earthquake energy to the volume V the strained region around the source, then assumes that the strain is uniform throughout this volume, and then uses field evidence for the maximum strain which the earth's crust can withstand (about 10^{-4}). Then, if V is limited by the thickness of the crust, an upper bound to energy of about 5×10^{24} ergs is obtained. It is hard to assess the validity of the assumptions used in obtaining this result.

A very similar approach has been given by Shebalin (1970), though it is less convincing. He quotes linear relations between earthquake magnitude and both mean length of focus and vertical extent of focus, from an earlier paper (Shebalin, 1971). He then uses limitations on both length and depth to set an upper bound to magnitude. The validity of his starting relations is very much open to question.

Similar procedures have been outlined by Hofmann (1974), who describes how magnitude fault-length relationships (e.g. Bonilla and and Buchanan, 1970) may be used to assign maximum magnitudes. Obviously this type of approach presupposes that we can clearly define the location and length of all active faults in an area, that breakage beyond the present fault length is impossible, and that the magnitude-fault length relation is single valued (this is equivalent to proposing that all earthquakes have the same stress drop). Each of these assumptions is difficult to justify.

Shenkova and Karnik (1974) raise the possibility that the rate of strain accumulation may set limits on the maximum energy released in an earthquake. They indicate, for example, that if upper and lower bounds can be placed on a Benioff strain release graph, the maximum possible earthquake will be specified. This approach is meaningless unless the record of earthquakes already contains at least one maximum possible event.

These studies are typical of those attempting to use physical arguments. The strength of rock, under various physical conditions, is not well known. However, we know even less about the limitations on the size of the zone of slip, and it is this variable which probably limits the usefulness of physical arguments. The largest known fault area is probably the 1960 Chile earthquake, which was about 1000 km long and perhaps 200 km wide on a shallow dipping fault plane (Kanamori and Cipar, 1974). There do not seem to be any convincing arguments why fault breaks could not be larger than this on occasion. Could the entire Aleutian arc system break at once, for example?

The effect of strength of rock is related to stress drop. The basic problem can then be formulated as follows: Seismic moment M_o is defined by

$$M_o = \mu LWD \tag{1.3}$$

where μ is the rigidity, L is the length (long horizontal dimension), W is the width (shorter vertical or down dip dimension), and D is the average fault offset.

The stress drop $\Delta\sigma$ can be written

$$\Delta\sigma = \eta \frac{\mu D}{W} \tag{1.4}$$

where η is a geometrical factor which typically ranges from 0.25 (for long strike slip faults) to 0.75 (for long dip slip faults), as is shown by Chinnery (1967).

So we may generally write

$$M_o \simeq 2LW^2 \Delta\sigma \tag{1.5}$$

If stress drops are roughly the same (about 50 bars) for all earthquakes, as has been suggested (Kanamori and Anderson, 1975), then limitations to seismic moment M_o depend only on limitations to the dimensions of the fault area.

However, questions about the constancy of $\Delta\sigma$ remain. Some studies appear to indicate local stress drops as high as several kilobars (Archambeau, 1978). In the eastern US, the occurrence of moderate sized earthquakes in the lower crust with no surface expression of movement would appear to require rather small fault dimensions and correspondingly large stress drops. To take an example, if a fault area 20 x 20 km were possible in an area of stress concentration in the Eastern US, with a stress drop of one kilobar, equation 1.5 gives a seismic moment of over 10^{28} dyne-cm (equivalent to an M_s of over 7.5, see Figure 4). This is probably larger than any earthquakes so far observed in this area.

We conclude, then, that while physical arguments support the idea that there must be an upper bound to earthquake size, and suggest that there may be a substantial regional variation of this upper bound, we cannot yet constrain the appropriate parameters enough to estimate the sizes of these upper bounds.

1.5 Arguments Using Earthquake Statistics

A variety of authors have attempted to use the statistical characteristics of the earthquake record to estimate maximum possible earthquakes. It is not at all clear that existing earthquake catalogs are good enough for this type of study. Certainly, in the example discussed in detail in section 2 of this report, it is clear that problems of saturation of the magnitude scale and individual station detection completely obscure the presence or absence of upper bounds.

There are two possible approaches to the analysis of earthquake catalogs. The first involves the use of the frequency-magnitude curve, which is discussed extensively in section 2. The other is based on Gumbel's (1958) Theory of Extremes. Gumbel described three asymptotic distributions which may be used to model the distribution of largest events occurring in a sequence of equal time periods through the earthquake record. The Type I asymptotic distribution of largest values corresponds to a linear frequency-magnitude relation, with no upper bound. The Type II asymptotic distribution includes the case where large events are less frequent than would be expected on the basis of smaller events, i.e. a non-linear frequency-magnitude curve. The Type III asymptotic distribution specifically includes an upper bound. Algebraic details can be found, for example, in Yegulalp and Kuo (1974).

Applications of the Type I distribution generally accomplish no more than the use of linear frequency magnitude statistics, and no upper bound is included. Papers using this distribution include Epstein and Lomnitz (1966), Gayskiy and Katok (1965), Milne and Davenport (1968), Connell (1968), Karnik and Hubnerova (1968, 1970), Yegulalp and Kuo (1974), Shenkova and Karnik (1974) and Shakal and Toksoz (1977). Though some of these papers mention maximum magnitude earthquakes, it is clear that what is discussed is the quality M^{P}_{max}, the magnitude which has a probability of occurrence (during some fixed period) that is less than P.

Studies that attempt to use the Type III asymptotic distribution are potentially more interesting. These include P'ei-shan and Lin

(1973), and Yegulalp and Kuo (1974). The first of these studies does not define the magnitude used, while the second is based on Gutenberg and Richter's (1954) data. They can both be shown to be formally equivalent to trying to fit the frequency-magnitude curve with a truncated distribution (Cosentino et al., 1976, 1977). We note that Knopoff and Kagan (1977) have argued that frequency-magnitude statistics are to be preferred over extremal statistics since the first uses all of the available data.

To anticipate section 2, there is no doubt that saturation of the M_s scale begins in the range 7-7.5. It is interesting to note that most of the estimates of M_{max} from these studies are greater than $M_s = 7.5$, and the vast majority are greater than $M_s = 8.0$. As long as saturation of the magnitude scale is not considered, there is no way that the results can be unambiguously interpreted as indicating the presence of an upper bound with regional variations.

1.6 Use of the Level of Seismic Activity

Perhaps the most persistent attempts to study the nature of earthquake upper bounds have been made in the USSR by Riznichenko and his co-workers, beginning with Riznichenko (1962, 1964a, 1964b). Many associated references are listed by Riznichenko and Bagdasarova (1975).

Riznichenko's basic postulate is that there is a clear cut upper bound to the energy released in an earthquake. Setting the total energy release $E = 10^k$ joules, he discusses the problem in terms of E_{max} and K_{max}. He uses an implied relationship between energy and the observed quantity, magnitude, of the form

$$\log E = a + bM \qquad (1.6)$$

The particular values of a and b used are not quoted (and are still open to question), and the particular definition of magnitude M is not given.

He recognized from the beginning that it was difficult or impossible to determine K_{max} directly from the observed earthquake catalog of an area. He has therefore focussed on the possibility of establishing a relationship between K_{max} and the level of seismic activity A in the frequency-energy relation

$$\log N_I = A - \gamma(K - K_o) \qquad 1.7)$$

(A is therefore the activity at the reference energy level K_o). He has discussed the form of the relationship $A(K_{max})$ in several papers (Riznichenko 1964a, Rznichenko and Bagdasarova 1976 and others). Briefly, his argument is to relate the energy K of an earthquake to a volume radius R (for Central Asia he obtained $R^3 = 0.315\ 10^{K-10}$), to average the activity A over a circular region of radius R to obtain $\bar{A}$, and then determine an empirical relation between $\bar{A}$ and K_{max}. For Central Asia he determined (Riznichenko and Bagdasarova, 1976)

$$\log \bar{A} = 2.84 + 0.21\ (K_{max}-15) \qquad (1.8)$$

while for Japan he found a better fit with

$$\log \bar{A} = 2.84 + 0.39\ (K_{max}-15) \qquad (1.9)$$

These equations are intended to be valid for $15<K<19$, or $10^{22}<E<10^{26}$ ergs.

The form of these equations was derived very artificially (Riznichenko, 1964a). K_{max} was simply chosen as the largest event for a given region (often using a short time sample), and $\bar{A}$ determined for the region. The plot of $\bar{A}$ against K_{max} had considerable scatter, and a linear relation was fitted to the largest K_{max} values (Riznichenko and Zakharova, 1971). In 1964 the constants estimated in equation 1.8 were 2.80 and 0.20, so there has been little change in the relation in the subsequent 12 years. The difference in the slope found for Japan (0.39 instead of 0.21) is disturbing.

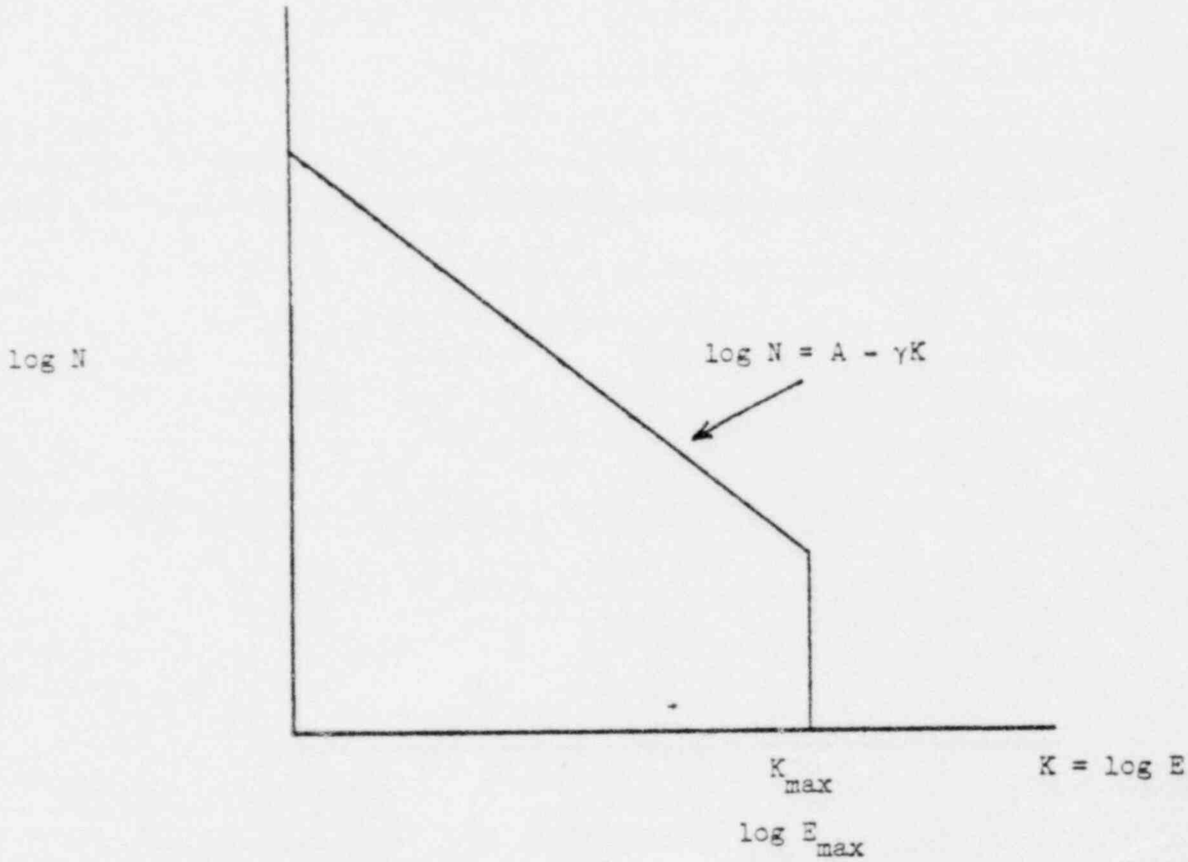

Fig. 1. Riznichenko postulates a clear-cut upper bound to total earthquake energy E, and assumes a linear frequency-energy relation for energy values below E_{max}.

Obviously, the problem in this approach is that K_{max} needs to be determined in some regions before the general law can be established. We must allow, however, the possibility that successive application of the equation in various regions (e.g. Gorbunova, 1969: Drumya and Stepanenko, 1972) may improve the constants by an iterative or "boot-strapping" method. The logical basis for the expression 1.8 is not established. Whether or not it works in practice is less clear. The authors compare 31 large earthquakes in Japan with the predictions of equation 1.9. Twenty-one are found to be in agreement, 10 are found to be larger than the predicted K_{max}, though the authors note that uncertainties in many of the epicenters make it hard to make a firm conclusion from this result.

The situation is far from satisfactory. The existence of a relation between K_{max} and A is not proven, and appears to be more of a hope than a scientific fact.

We should note, in passing, that if the maximum value is defined using a probability P (E^{p}_{max}), then there is a very clear relation between the maximum value and the rate of seismic activity. This has been described, in a most obscure way, by Housner (1970). His argument may be restated as follows: Let us assume a linear unbounded frequency-magnitude law of the form

$$\log N = a - bM \qquad (1.10)$$

where N is the cumulative number of events, with magnitude $\geq M$, per unit area, during a unit time period (per year, say). Suppose that N_n is the number of events/year that can be considered negligible for risk purposes.

Then $$\log N_n = a - b\, M^p_{max} \tag{1.11}$$

For two different regions, with different a and b values, we have

$$\log N_n = a_1 b_1 M^p_{max}(1) = a_2 b_2 M^p_{max}(2)$$

so,

$$M^p_{max}(2) = \frac{b_1}{b_2} M^p_{max}(1) + \frac{a_2 - a_1}{b_2} \tag{1.12}$$

It is reasonable to set $b_1 \simeq b_2 \simeq 1$, and then

$$M^p_{max}(2) = M^p_{max}(1) + (a_2 - a_1)$$

or

$$M^p_{max}(2) = M^p_{max}(1) + \log \frac{N_o^{\ 2}}{N_o^{\ 1}} \tag{1.13}$$

where N_o is the number of events with magnitude 0, which may be taken as an indication of the level of activity. In a simple example, if area 2 has a seismicity of one-hundredth of area 1, then the M^p_{max} value for area 2 will be two units smaller than the M^p_{max} for area 1.

The reason that Housner's (1970) argument is obscure is that he tries to associate the above with a true M_{max} value, as shown in Figure 1. Clearly the analysis really refers to our unbounded frequency-magnitude law.

In summary, existing literature sometimes attempts to postulate a relationship between seismic activity and the upper bound to earthquake size, but success in establishing the nature and even the validity of this relationship has been essentially non-existent.

1.7 Pattern Recognition Approaches

Recognizing the fundamental difficulties involved in trying to relate the size of maximum possible earthquakes to the level of seismic activity alone there have been several attempts to include a variety of other geophysical and geological information.

Riznichenko and Dzhibladze (1974) have compared and correlated the estimation of K_{max} using the level of seismic activity, the gradient of the Bouguer gravity anomaly (suggested by Tsuboi, 1940, and Berg et al., 1964), and the velocity of vertical movements determined by geodetic and geomorphological methods. The three estimates were combined together to obtain a single estimate using weights of 1.0 for the seismic data, and 0.5 for each of the other methods. The results are no more convincing than those based on seismic activity alone. This paper is notable, however, for its extensive collection of references.

Shenkova and Karnik (1974) state frequency-energy data are not reliable enough for the estimation of K_{max}, and urge the inclusion of data on "environmental properties and the rate of energy accumulation" (i.e. Benioff graphs). However they give little indication how these pieces of information should be tied together.

In view of the interest of several Russian geophysicists in pattern recognition problems (see, for example, Gelfand et al., 1976), it is not surprising that attempts have been made to apply these methods to the determination of M_{max}. This topic is addressed by Bune et al. (1975), and an application to the Carpathian region is described by Borisov and Reysner (1976). The general idea is to look for those combinations of observable features that appear to be indicative of the observed M_{max} values. The features selected include such items as rates of recent vertical motion, nearby volcanism, presence of fractures and fracture intersections, seismic activity, gravity anomaly etc. The data analysis follows the usual procedures. Most of the features chosen were found to vary strongly with M_{max}.

The basic problem of this analysis is, however, not addressed by the authors. In order to deduce the appropriate relationship, values of known M_{max} are needed in a substantial number of regions. Since these are not readily available, the authors used "estimates made by experts". This introduces such a strongly subjective element into the analysis that it must be regarded as meaningless.

1.8 Other Studies

Two recent studies should be mentioned, the first for completeness and the second because it has an interesting approach to the problem.

Caputo (1977) has proposed a complex model which purports not only to account for the linearity of the frequency-magnitude relation, but to predict the maximum seismic magnitude and moment. The assumptions on which the author bases his analysis appear to be completely unreasonable, and the paper is meaningless.

Smith (1976), on the other hand, has proposed using geological data to obtain a mean rate of slip for a fault zone over the past 10's of thousands of years or longer. Then, if the frequency-moment relationship for the area is linear, and can be defined (see Chinnery and North, 1975; Smith's argument here is less rigorous), then there must be an upper bound moment that is consistent with observed slip (Brune, 1969). Smith uses geological data of Hamilton (1975) to obtain these upper bound moments (which he converts back to upper bound magnitudes).

This approach is one of the most reasonable that we have seen, but problems still remain. There are considerable difficulties in the definition of the frequency-moment relationship for a limited zone. Even if this can be estimated, however, there must still be difficulties in the interpretation of geological slip data. Slip on the San Andreas fault system has clearly been distributed over a rather wide zone on a geological time scale. It is likely that individual faults could carry much of this slip for a period of time, and then it could be transferred to other neighboring faults. To put this another way, Smith's (1976) approach requires that the earthquake process be stationary over the period of the geological data on each fault considered. This is a questionable assumption for the fault zone as a whole, and may be invalid for individual faults within the system. And, of course, there appears to be no way to apply Smith's method to regions such as the Eastern US, where geological information on fault slip is available.

1.9 Discussion and Conclusions

The basic problem in attempting to determine the maximum possible earthquake in a region can be stated quite simply. If the earthquake record for the region has a length T years, then evidence is available that bears on the earthquakes that have mean return periods of up to T years, or a probability of occurrence down to 1/T per year. This evidence is not necessarily good evidence, for the largest earthquakes in the sample.

The occurrence of large earthquakes appears to be described quite well by a Poisson distribution (Epstein and Lomnitz, 1966; Lomnitz, 1966). The probability that at least one event with an annual probability of 1/T will occur within a period of t years is

$$P = 1 - e^{-t/T} \qquad (1.14)$$

So, if t = T, the probability is 63%. This suggests that in more than ...ird of all regions studied there is likely to be an apparent ...ency of large events.

To phrase this another way, a 100 year record of earthquakes will only give reliable information (at the 90% level) for those earthquakes with a mean return period of about 40 years or less, or an annual probability of .025 or more. In practice, of course, the length of the earthquake record is often considerably less than 100 years, and this applies to most of the regions of the USSR studied in the quoted literature, and to California and other active zones. Clearly, then, a 100 year record of seismicity is only adequate for the determination of maximum possible earthquakes if the mean return periods of these earthquakes are significantly less than 50 years. This implies that the maximum possible earthquake must have occurred several times during the period of observation.

In all of the literature that has been surveyed, there is no case of a specific region where a maximum possible earthquake can be clearly defined. Even when all regions are considered together in a global earthquake record, the apparent upper bound to surface wave magnitude M_s can easily be accounted for on the basis of saturation of the magnitude scale (Chinnery and North, 1975). Perhaps the most useful contribution to this area that could be made at the present time would be the clear and unambiguous demonstration of the existence of an upper bound to earthquake size in just one region, anywhere on the globe.

It is necessary to add, here, that we have not attempted to define the term "region". This is a thorny topic (see, for example, Hadley and Devine, 1974) which has been emphasized by the term "tectonic province" which appears in the NRC Rules and Regulations, Part 100, Appendix A. We shall not discuss it further here, except to note that given a map of epicenters for the earthquakes in a seismic zone it is always possible to select a region that contains no large events. The validity of such a selection is very questionable.

It appears, then, that existing seismic data are unable to throw any light on the questions of the existence and size of maximum possible earthquakes. In spite of the deep seated belief of many seismologists and earthquake engineers that upper bounds must exist, the only reasonable approach, given our current state of knowledge, is to assume that these upper bounds are at rather high levels in all areas.

We are therefore forced into the classic method of simple extrapolation of linear frequency-magnitude or frequency-intensity relationships. This raises an additional problem which deserves discussion.

In the context of the evaluation of the seismic risk to critical structures such as nuclear power plants, we would like to establish a way to determine the size of the earthquake that occurs with some fixed risk probability within a given region. Following McGuire (1976) and others, we may usefully set this fixed probability at 10^{-4} per year. If the earthquake process is stationary over long periods of time, such an earthquake will have a mean return period of 10,000 years. If the process is non-stationary, this statement is meaningless. However, in practice we have very little alternative but to assume that the available record of earthquakes is representative of the rates of occurrence of both small and large earthquakes in the immediate past and the immediate future.

The problem of stationarity is not easily set aside. Evidence from very long compilations of earthquakes in the Mediterranean area and China (the latter was discussed by Lee and Brillinger, 1978) show disturbing changes in seismicity on time-scales of a few hundred years. The seismic record in New England shows similar changes during its 300 year length (Chinnery and Rodgers, 1973; Shakal and Toksoz, 1977). Clearly this raises the possibility that large earthquakes may be associated with some long term average level of seismicity which is very different from the recent short record of smaller events. It is important that research into the stationarity of earthquake processes in various tectonic environments continue.

The most promising avenues for future investigations into maximum possible earthquakes would appear to lie in three areas. First, we need more information on the nature of the strain and stress fields in seismic zones. Second, we need to improve our understanding of the ultimate strength of crustal materials in a vareity of tectonic settings. It seems likely that the true upper bound is controlled by the size of the region of accumulating stress, and the ability of the crustal rock to withstand that stress. Thirdly, the information from geological and geomorphological data on long term fault slip, where surface faulting is visible, must place some constraints on the largest possible earthquakes (Smith, 1976). This approach needs further development, though the question of stationarity may limit its usefulness.

2. ANALYSIS OF GLOBAL CATALOGS

2.1 Characteristics of Global Catalogs

A logical place to seek for information on the existence of upper bounds to earthquake size, and the variation of these upper bounds with tectonic region, is within earthquake catalogs. There are basically two kinds of catalogs, those compiled for a limited region using data from a local network, and those compiled for the whole world using a global network of stations. We have chosen to begin this study by analyzing the global earthquake catalog, since this seems most likely to contain evidence for regional variations, if they exist.

In order to be useful for this study, a global catalog must have two important characteristics. First, it must be complete, particularly for large earthquakes, and preferably for medium-sized events as well. Second, it must use a clearly defined measure of earthquake magnitude which is uniformly applied to all events. As we shall see, this turns out to be a much more restrictive condition than it appears to be at first sight.

Several global catalogs are available. Those including events since the early 1900's include Gutenberg and Richter (1954), Duda (1967) and Rothe (1969). Unfortunately, the global distribution of seismic stations was very poor until 1960, and these catalogs all suffer from a high degree of non-homogeneity. With the establishment of the World Wide Standard Seismograph Network (WWSSN) in the early 1960's, a much more homogeneous data set became available. Data from this network, together with a variety of data from other stations were analyzed by two organisations. The U.S. Coast and Geodetic Survey, and its successors the National Ocean Survey and the U.S. Geological Survey, have produced a fairly rapid bulletin (the PDE, or Preliminary Determination of Epicenters) issued on the average about 6 months after an event occurred. The International Seismological Center (ISC) has chosen to collect all the available data, including the PDE bulletin, and issue a more comprehensive catalog. Typical delays in the publication of the ISC catalog ranged from two to three years. Both the PDE and ISC catalog began consistent routine bulletin production at the beginning of 1964, and since then have maintained the production of very uniform catalogs.

Both catalogs, since 1964, have recorded a body wave magnitude m_b for essentially all events. This magnitude is based on the maximum peak to peak amplitude in the first few seconds of the P-wave arrival on short period instruments (operating in a rather narrow frequency band centered at about 1 hz). Surface wave magnitudes M_s (at a period of about 20 seconds) were recorded very irregularly, and only in the last year or two have attempts been made to measure M_s on a routine basis. The requirement that the catalog be complete forces us to focus on the body wave magnitude m_b. For reasons which are outlined in the next sections, this is not desirable, but there is little that can be done about it. Attempts to relate M_s to m_b have shown a large scatter (see, for example, Aki, 1972).

In the sections that follow we shall concentrate on the ISC catalog for a very practical reason - it is available in detail on magnetic tape (the detailed PDE listing is not). This facilitates a variety of computer analyses of the very large amount of data concerned.

2.2 Earthquake Statistics

There are two basic representations of the statistical characteristics of an earthquake catalog. One deals with the relationship between earthquake frequency and earthquake magnitude. The other utilizes Gumbel's (1958) theory of extremes, and is concerned only with the largest event within a given time period. Though these two approaches appear to be very different, they give very similar results when applied to the same data set (see, for example, Chinnery and Rodgers, 1973, and Shakal and Toksoz, 1977). Because of this, and because the frequency-magnitude approach uses all of the data in a catalog, it is to be preferred. Knopoff and Kagan (1977) have specifically shown that extremal statistics are much inferior in some cases. For this reason, we shall use the frequency-magnitude approach throughout.

Gutenberg and Richter (see Richter, 1958) demonstrated that local earthquakes in California obeyed a frequency-magnitude relation of the form:

$$\log N_i = a_o - bM \tag{2.1}$$

where N_i is the number of earthquakes with magnitudes in a small range centered on M, and a_o and b are constants. This form of the equation is necessarily discrete (the constant a_o depends on the size of the magnitude intervals in which the earthquakes are accumulated). In many cases, it is more convenient to use the cumulative form:

$$\log N_c = a - bM \tag{2.2}$$

where, now, N_c is the number of events with magnitude M and greater. This equation may be regarded as being continuous, and is more amenable to analysis. It is easy to show that if equation 2.1 is valid, then equation 2.2 is also linear and has the same slope or b-value. Values for the constant b typically lie close to 1.0.

Unfortunately, there is no sound theoretical basis for a linear frequency-magnitude curve, and it must be regarded as empirical. Even

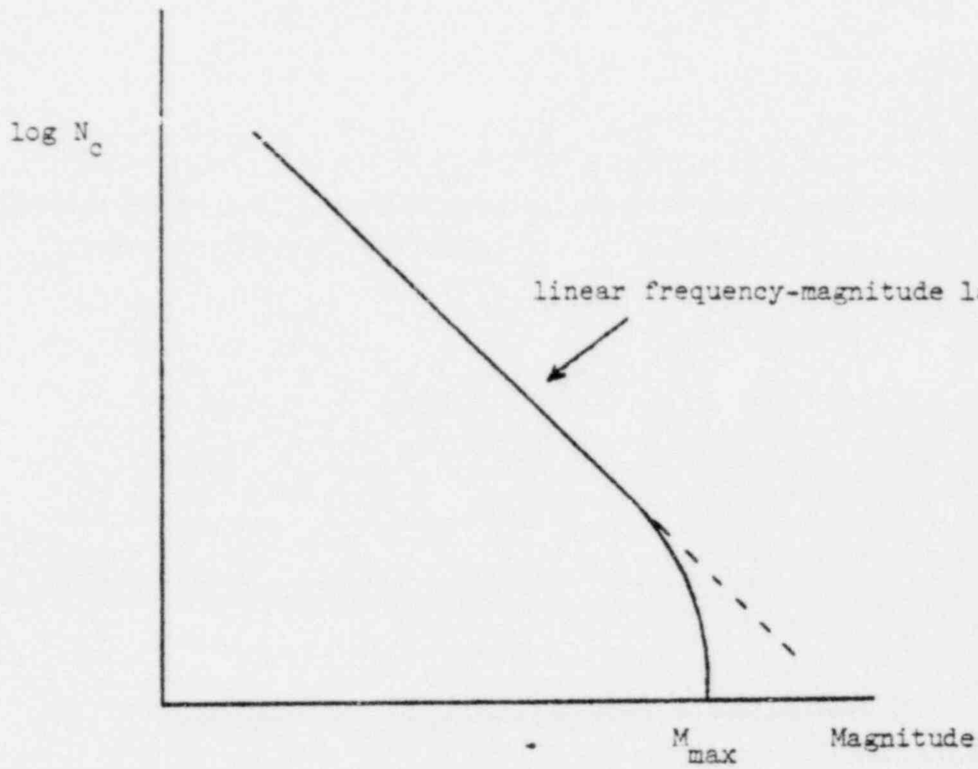

Fig 2: Ideal effect of an upper bound to earthquake magnitude, using cumulative frequency-magnitude statistics.

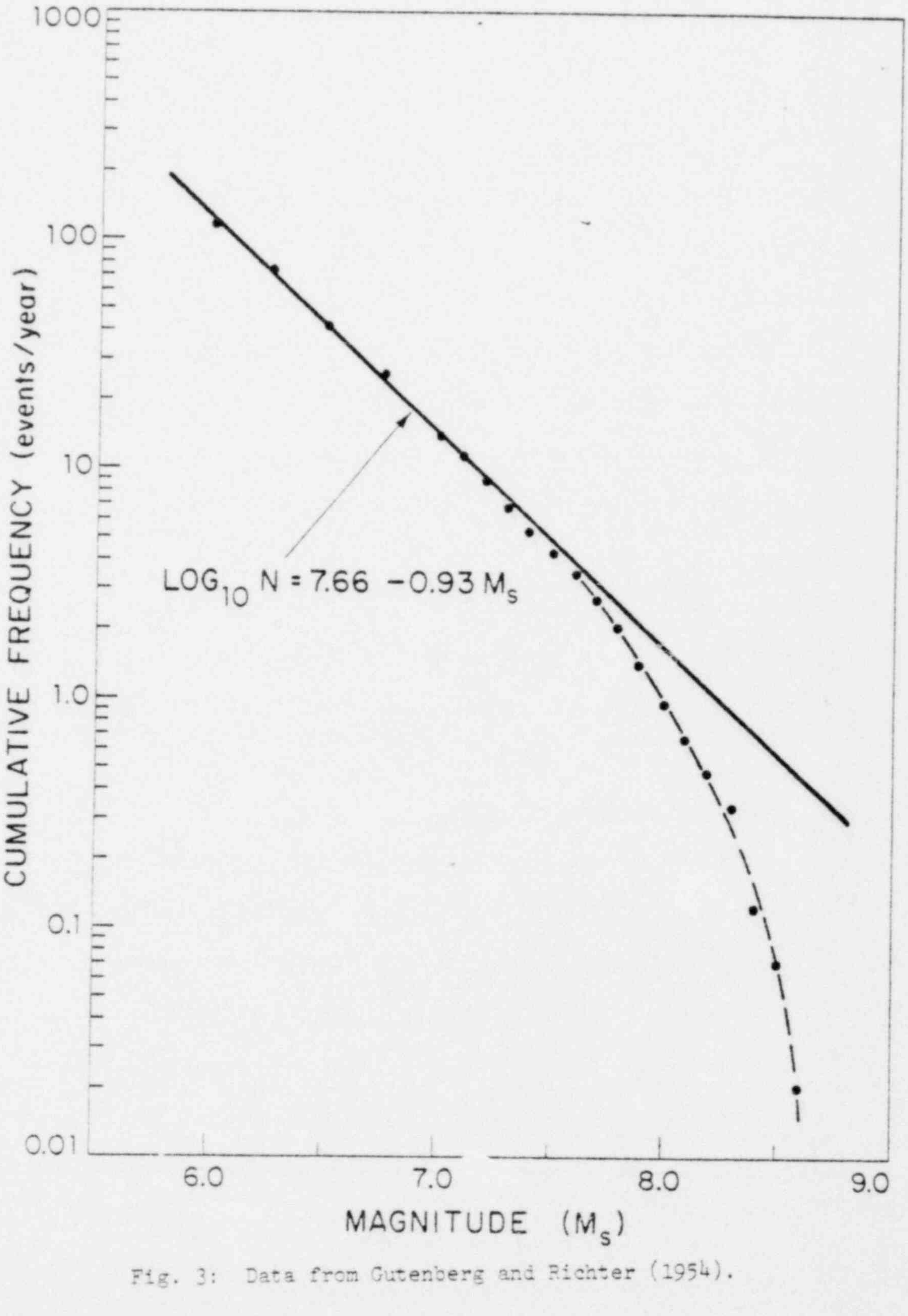

Fig. 3: Data from Gutenberg and Richter (1954).

using observational data, the universality of a linear relation is not clear. Many of the reasons for this will be discussed in the sections that follow.

In an ideal world, the presence of an upper bound to earthquake magnitude will reveal itself by a departure from linearity at the upper end. Figure 2 shows an idealised representation of this non-linearity. Unfortunately, there are two other effects that can also lead to a curve similar to Figure 2. First, any measure of magnitude based on a limited spectral band has a built-in saturation property. This is discussed in the next section. And second, seismic instruments frequently have a limited dynamic range, and the magnification is often set to record medium sized earthquakes. In this case, large earthquakes will cause the instrument to go off-scale, and a measure of magnitude is impossible. As a result, there may be a purely instrumental upper-bound to measureable magnitude for a given instrument. The effect of this on network determinations of event magnitude is discussed in later sections.

2.3 Saturation of the Magnitude Scale

Several authors (Chinnery and North, 1975; Kanamori and Anderson, 1975, etc) have recently pointed out that because of the shape of the spectrum of the radiation emitted by an earthquake source, any measurement of magnitude based on a limited spectral band of frequency must saturate. For example, M_s is usually measured at about 20 seconds period. When the source is large enough that fracture propagation lasts for longer than 20 seconds, the amplitude of the 20 second radiation will not change with increasing size, though its duration in general will.

An example of this effect was discussed by Chinnery and North (1975). Figure 3 shows the cumulative frequency magnitude curve for large events listed in the classic study of Gutenberg and Richter (1954). It appears that the listed magnitudes are very close to present day M_s values (Evernden, 1970).

This diagram has often been used as a basis for discussing the existence of an upper bound to earthquake magnitude (see, for example, Housner, 1970). It is, however, possible to interpret this curve in another way. Figure 4 shows a compilation of recent data relating surface wave magnitude M_s to the seismic moment M_o. The highest two points correspond to the 1960 Chile and 1964 Alaska earthquakes. Both have been extensively studied and seem reasonably reliable. The observational data clearly indicate a saturation of the M_s scale which seems to begin at about M_s=7.5, and be complete at about M_s=8.5. The solid line in Figure 4 is a rough form of the M_s-M_o relation.

At this point we can legitimately ask if the fall-off in Figure 3 can be wholly attributed to this saturation. We can say this much: if the data in Figure 3 are translated into a frequency-moment graph, the result is very linear (see Figure 5).

Kanamori and Anderson (1975) have argued that the frequency-moment graph should be linear, with a slope of 0.67, if all earthquakes have the same stress drop. It therefore seems reasonable to postulate that this is the case, and to conclude that the Gutenberg and Richter result (Figure 3) can be explained as saturation of the M_s scale.

There are two important points that arise from this study. First, on a global scale, there is no direct evidence for an upper bound to seismic moment, though McGarr (1976) has argued on geometrical grounds that such an upper bound must exist fairly near the highest moment data point on Figure 5.

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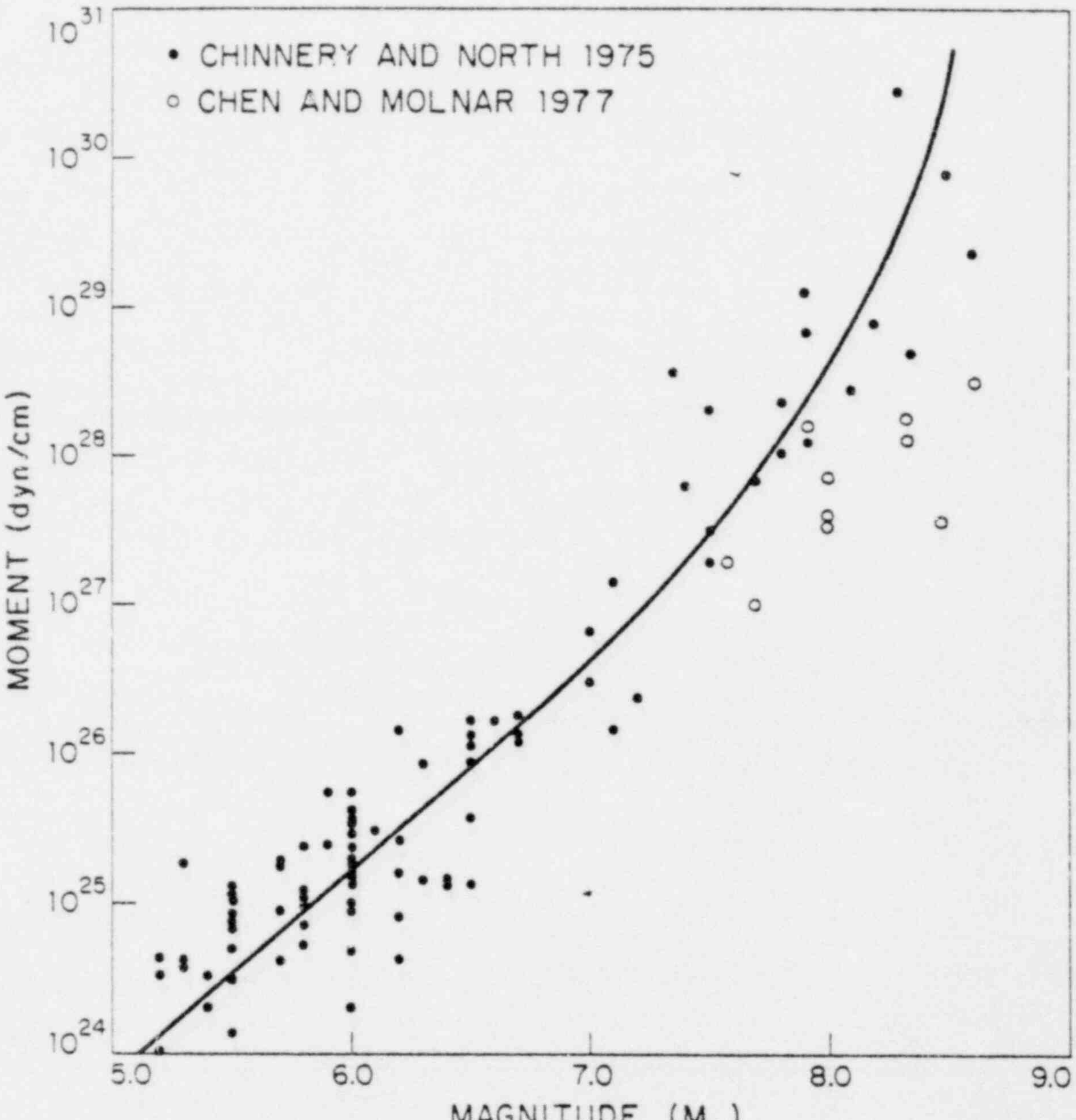

Fig. 4: Complilation of 87 published estimates of seismic moment as a function of surface wave magnitude M_S.

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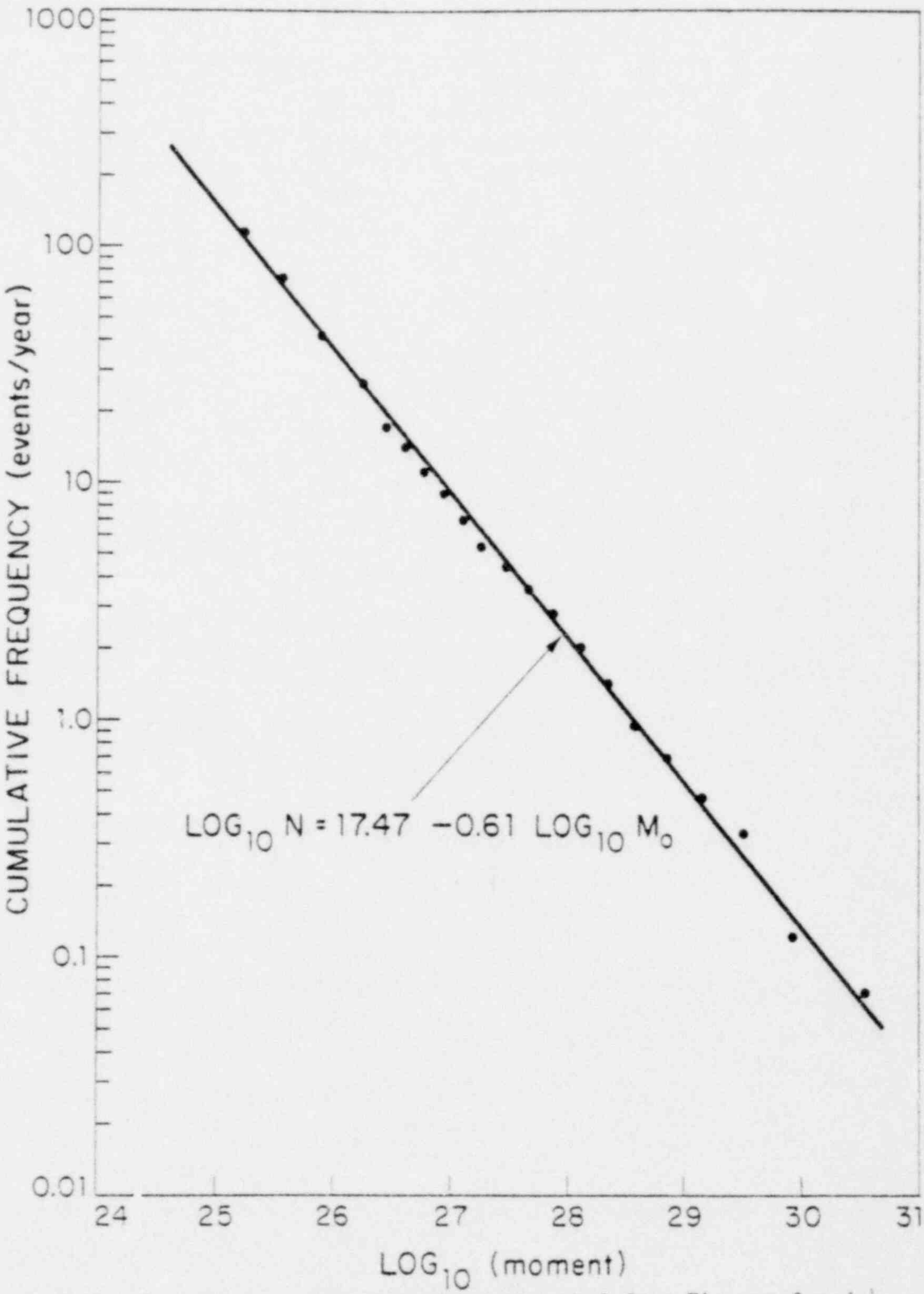

Fig. 5: Frequency - moment graph constructed from Figures 3 and 4.

Second, the importance of magnitude saturation is demonstrated. When we come to examine global catalogs using the 1 hz m_b scale, we must expect saturation to occur at lower magnitudes. This will clearly make the problem of trying to estimate regional variations in maximum earthquakes very difficult.

2.4 The ISC Catalog

An incremental frequency magnitude plot of data in the ISC catalog for the period 1966-70 is shown in the lefthand portion of Figure 6. Although ISC data are available for a longer period, we have chosen to limit ourselves to this 5-year span in order, as we shall see, to compare the overall catalog with certain special stations that were only operating during this time.

The resulting plot is typical of all frequency-m_b data currently available (e.g. Brazee and Stover, 1969, Brazee, 1969). There is no clear linear portion to the graph, and this has led some authors to propose a non-linear relation (e.g. Shlien and Toksoz, 1970; Merz and Cornell, 1973; Stewart, 1974). It is therefore very difficult to determine a unique b-value, though typical attempts to do this lead to high values of up to 1.5 or more (see Figure 6). At low magnitudes many events are not reported, and the plot curves downwards. At the high end, of particular interest to us, the graph appears to steepen, and end near m_b=6.5 or 6.6. No events larger than 6.6 appear in the catalog during this time period.

It seems reasonable to ask if these catalog characteristics are in any way the result of the stations used in the analysis. As many as 500 or more stations feed data in to the ISC, many of them very irregularly. To examine this question, we selected a subset of 28 stations which operated continuously throughout 1966-70, and which report regularly to the ISC. The stations used are listed in Table 1. Magnitudes were recomputed as the average of those reported by the 28 stations, and a requirement that at least 3 of the stations must have reported the event was superimposed. The resulting frequency-magnitude graph is shown in the righthand portion of Figure 6 (the solid points). A second data set was formed by applying the station magnitude biases determined by North (1977) to the 28 station network. The results are shown as open circles.

The 28 station network shows very similar characteristics to the catalog as a whole. In particular, the general curvature of the graph and the fall-off at high magnitudes are preserved. This is convenient since it allows us to study the 28 station network instead of the whole catalog.

There are reasons to suspect that biases may be introduced into the network magnitudes by the process of averaging the reported station magnitudes. This problem will be discussed in more detail in later sections of this report. It suggests, however, that it may be worthwhile looking at the frequency-m_b characteristics of the events reported by individual stations.

Figure 7 shows plots of the events reported by Kevo, Finland, for 1966-70. On the left are counts of log A/T values (A is the observed amplitude of ground motion, and T is the observed dominant period), which are independent of source location. The values are converted into station m_b by the application of a standard amplitude distance correction. This correction is best known in the distance range 30 to 90 degrees, and the righthand side of Figure 7 shows events in this distance range. Similar data for Port Moresby, New Guinea, are shown in Figure 8.

TABLE 1: 28 STATION NETWORK

STATION CODE	LOCATION	BIAS (North, 1977)
ALQ	Albuquerque, N.M.	-0.20
BHA	Broken Hill, Zambia	-0.28
BMO	Blue Mtns., Oregon	-0.29
BNS	Bensberg, Germany	+0.20
BUL	Bulawayo, Rhodesia	-0.07
CAN	Canberra, Australia	-0.02
CLK	Chileka, Malawi	-0.27
COL	College, Alaska	+0.01
COP	Copenhagen, Denmark	+0.36
EUR	Eureka, Nevada	-0.24
KEV	Kevo, Finland	+0.02
KHC	Czechoslovakia	+0.10
KJN	Kajaani, Finland	+0.14
LJU	Ljubljana, Yugoslavia	+0.29
MBC	Mould Bay, Canada	+0.14
MOX	Moxa, Germany	+0.02
NOR	Nord, Greenland	-0.14
NF-	Northwest Territories, Canada	0.00
NUR	Nurmijarvi, Finland	+0.19
PMG	Port Moresby, New Guinea	+0.10
PRE	Pretoria, South Africa	-0.07
PRU	Czechoslovakia	+0.04
RES	Resolute, Canada	+0.13
SJG	San Juan, Puerto Rico	+0.24
TFO	Tonto Forest, Arizona	-0.32
TUC	Tucson, Arizona	-0.14
UBO	Uinta Basin, Utah	-0.11
WIN	Windhoek, South Africa	-0.09

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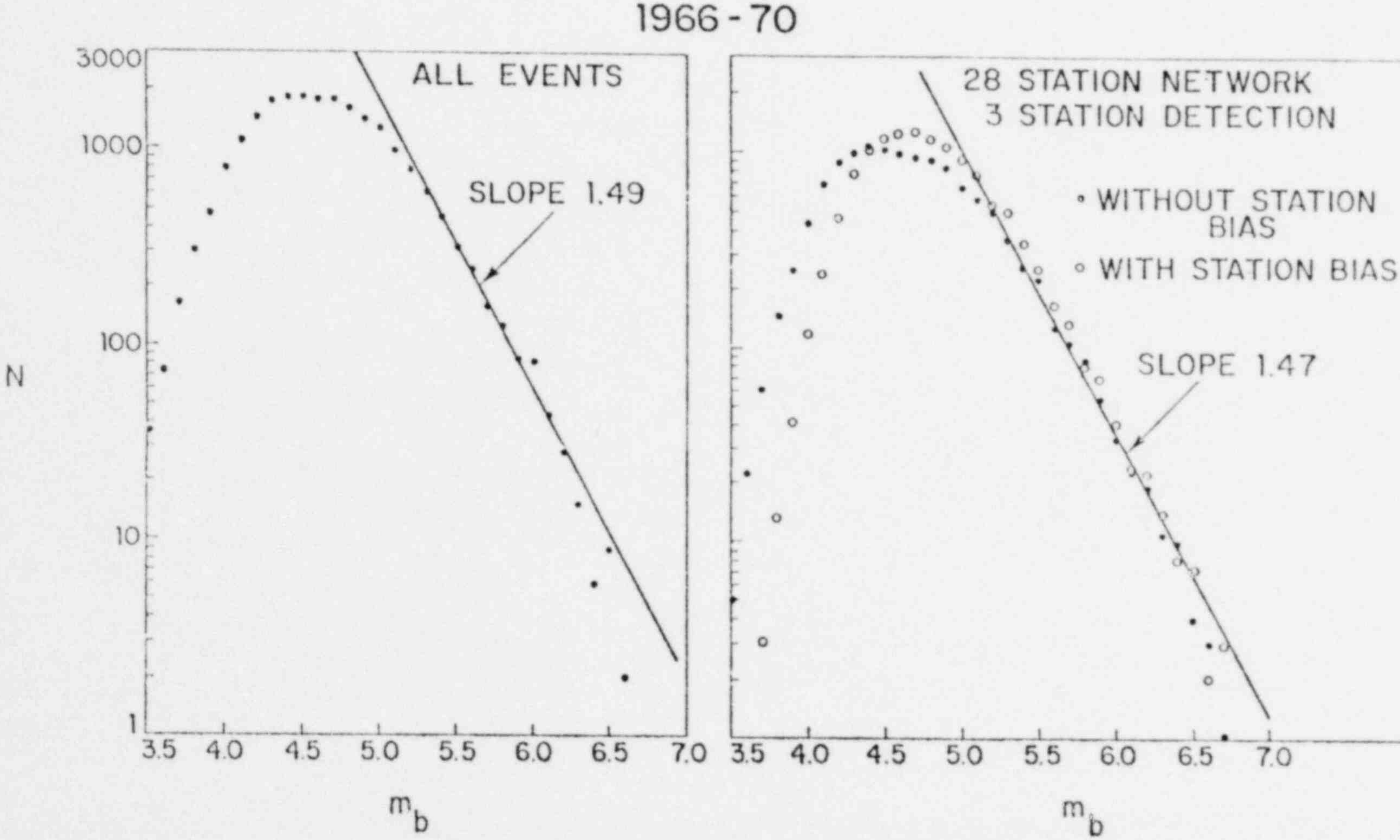

Fig. 6: Frequency magnitude data for the ISC catalog, for all listed events (left), and for a selected network of 28 stations (right). The 28 station network is listed in Table 1.

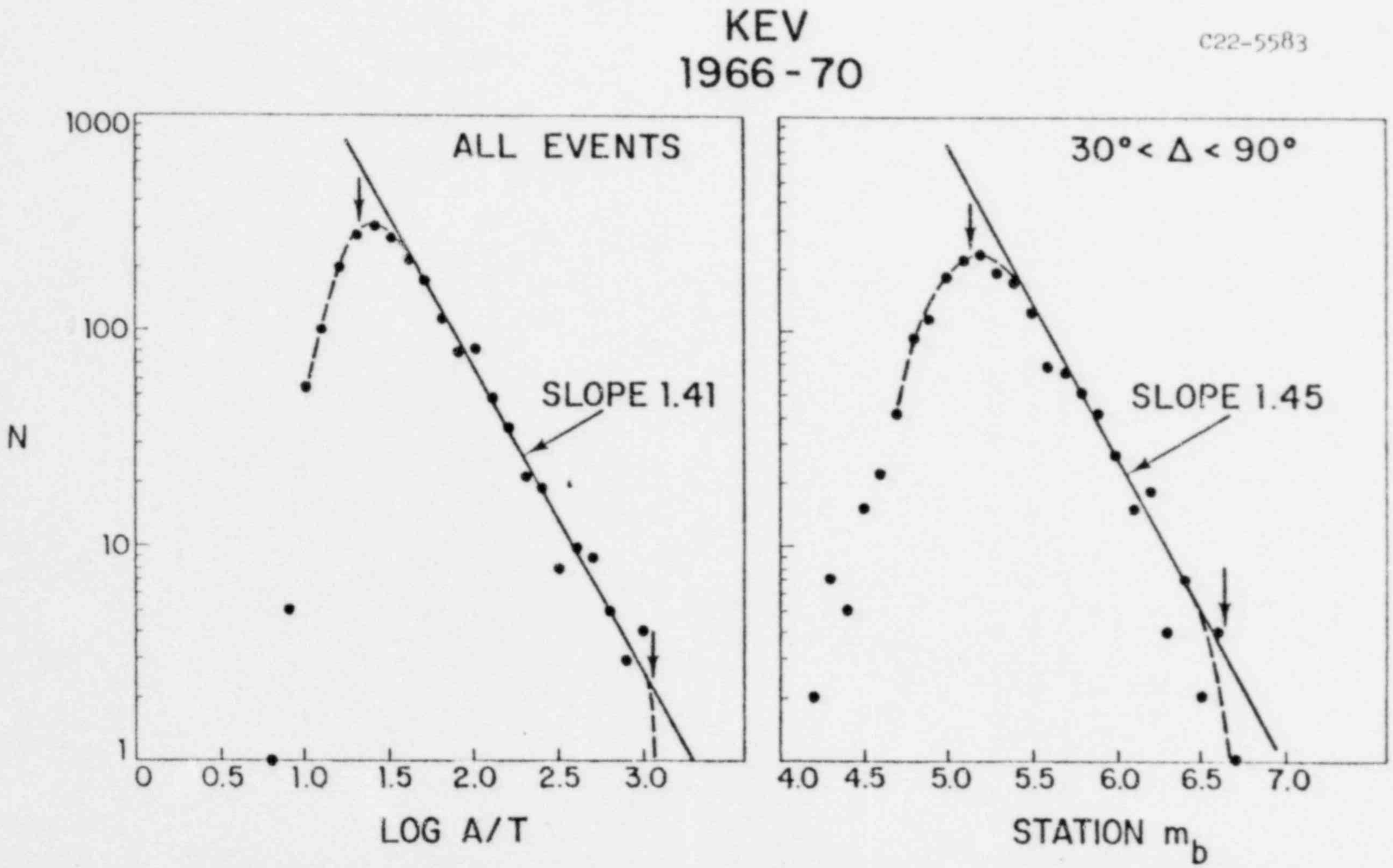

Fig. 7: Frequency magnitude data for Kevo, Finland.

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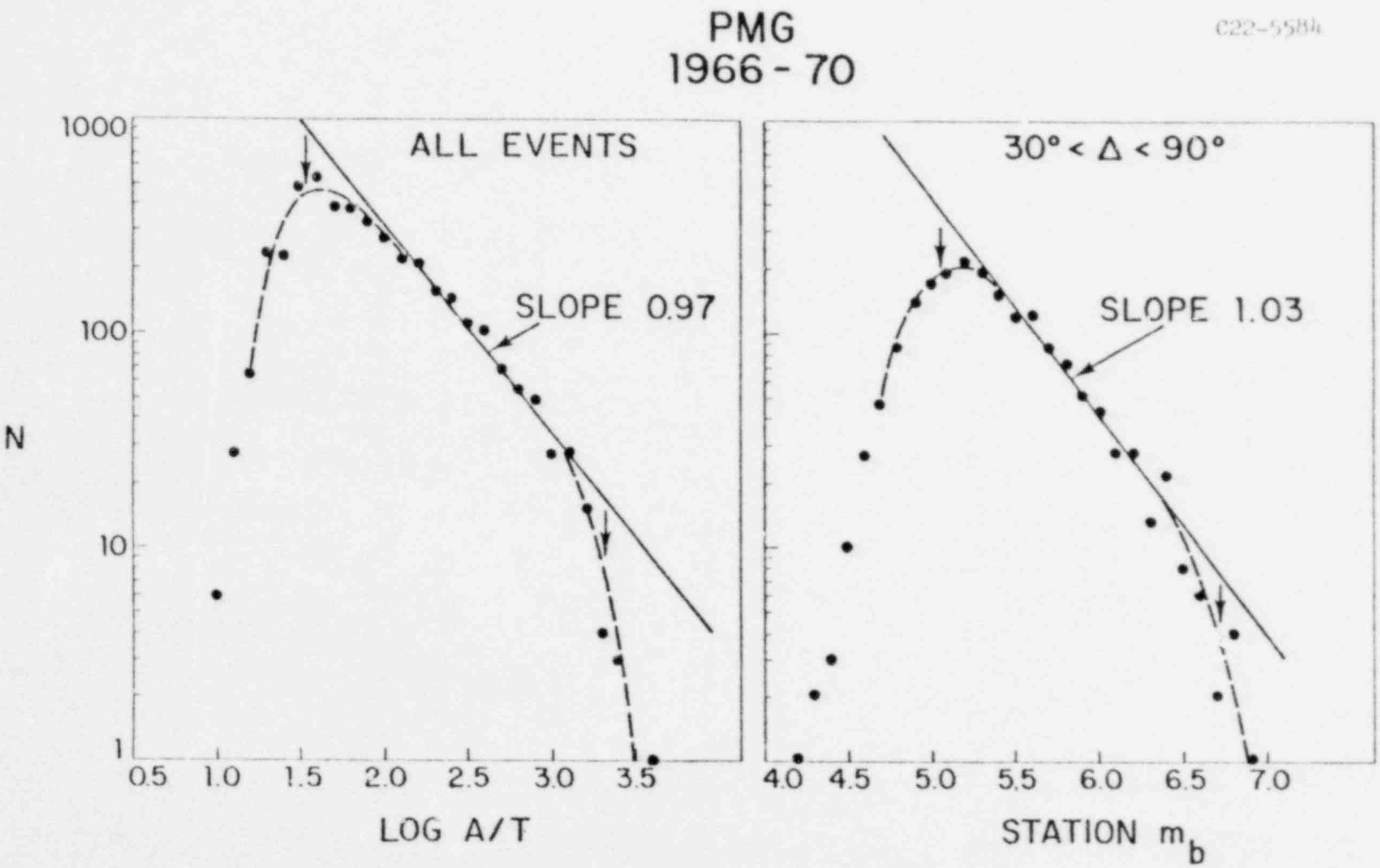

Fig. 8: Frequency magnitude data for Port Moresby, New Guinea.

We have compiled similar plots for all of the stations in the 28 station network. A wide variety of behavior is seen. If attempts are made to fit the frequency-m_b plots with a straight line, slopes are found to lie anywhere within the range 0.9 to 1.5. Figures 7 and 8 show clearly the differences that are observed.

There are two possible interpretations of these data. If the differences in b-value are real, this could indicate an important regional variation in seismicity characteristics (clearly PMG and KEV sample different portions of global seismicity). The second alternative is that station reporting characteristics vary considerably, and the data are not good enough to define a true b-value.

Perhaps the most surprising result is obtained when frequency-station m_b plots are made for the U.S. VELA observatories. These are BMO (Blue Mountains, Oregon), UBO (Uinta Basin, Utah), TFO (Tonto Forest, Arizona) and WMO (Wichita Mountains, Oklahoma). The four plots are superimposed in Figure 9. Each station has been adjusted horizontally according to the station biases of North (1977), and small vertical adjustments have been made to improve coincidence, recognizing that there are small differences in the seismicity sampled by each station. Again, only events in the distance range 30° to 90° are included.

Remarkably, these data are all consistent with a seismicity curve that is linear, with a slope of about 0.9, up to m_b=5.8, and then the curve bends downwards and approaches the vertical in the range m_b=7.0 to 7.5. This relation, indicated as a solid line on Figure 9, is remarkably similar to the Gutenberg-Richter M_s curve (Figure 3) in shape. However, it differs dramatically from those observed by normal stations. Notice, for example, that these observatories record many events in the range m_b=6.7 to 7.2, whereas none are listed in the ISC catalog.

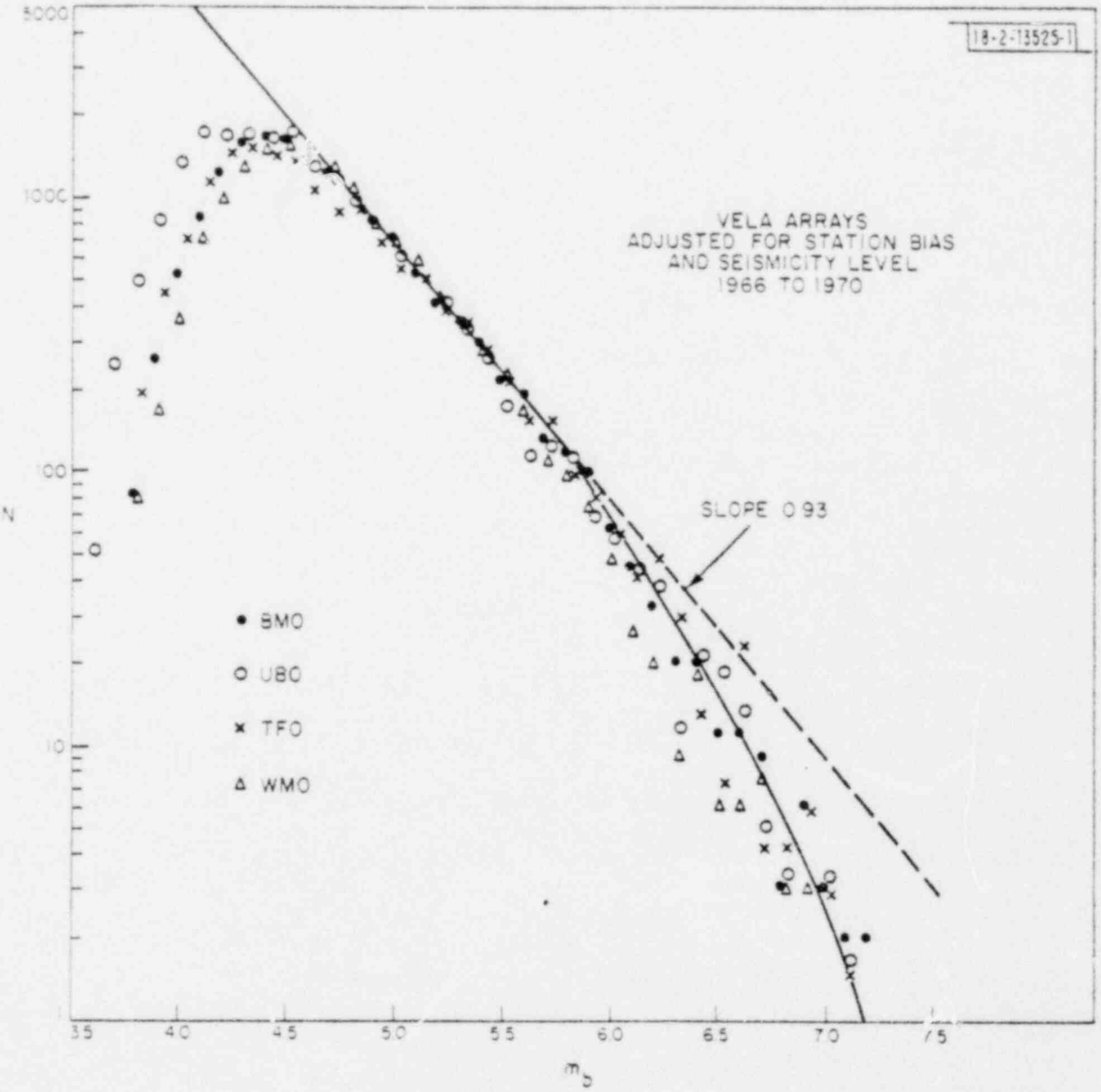

Fig. 9: Frequency-station m_b plots for four U. S. VELA observatories.

There are a number of important differences between the VELA arrays and the average analog seismic station. The operators of the VELA arrays were highly trained specialists, who made an unusual attempt to measure magnitudes carefully and consistently. More important, each of the arrays was equipped with a low gain channel, which gave the arrays a much larger dynamic range than the average station. These points strongly suggest that the VELA data may be more reliable than regular station reports. An additional suggestion that this is the case is obtained from the Large Aperture Seismic Array (LASA) in Billings, Montana. Figure 10 shows data from this array for a completely different time period (1971). The seismicity curve shown in Figure 9 is an excellent fit to this data set (in Figure 10 this seismicity curve has been adjusted vertically for a best fit).

In order to investigate this problem in more detail, it would clearly be advantageous to limit the geographical region within which the events are located. In this case we may expect a well defined seismicity curve, and we can test the ability of various networks to detect this curve. This is done in the next section.

2.5 Events in the Aleutian-Kuriles Region

The analysis of the previous section was repeated for events in the Aleutian-Kurile Island area (defined by longitudes 135°E to 140°W, and latitudes 30°-90°). The important seismicity of this area lies within the 30° to 90° range of stations in both Europe and the U.S.

Figure 11 shows the total ISC data base for this area for 1966-70. The frequency-magnitude data do not disagree strongly with the seismicity curve shown, which is that shown in Figure 9 adjusted vertically for a best fit. Upon closer examination, it transpires that the catalog for

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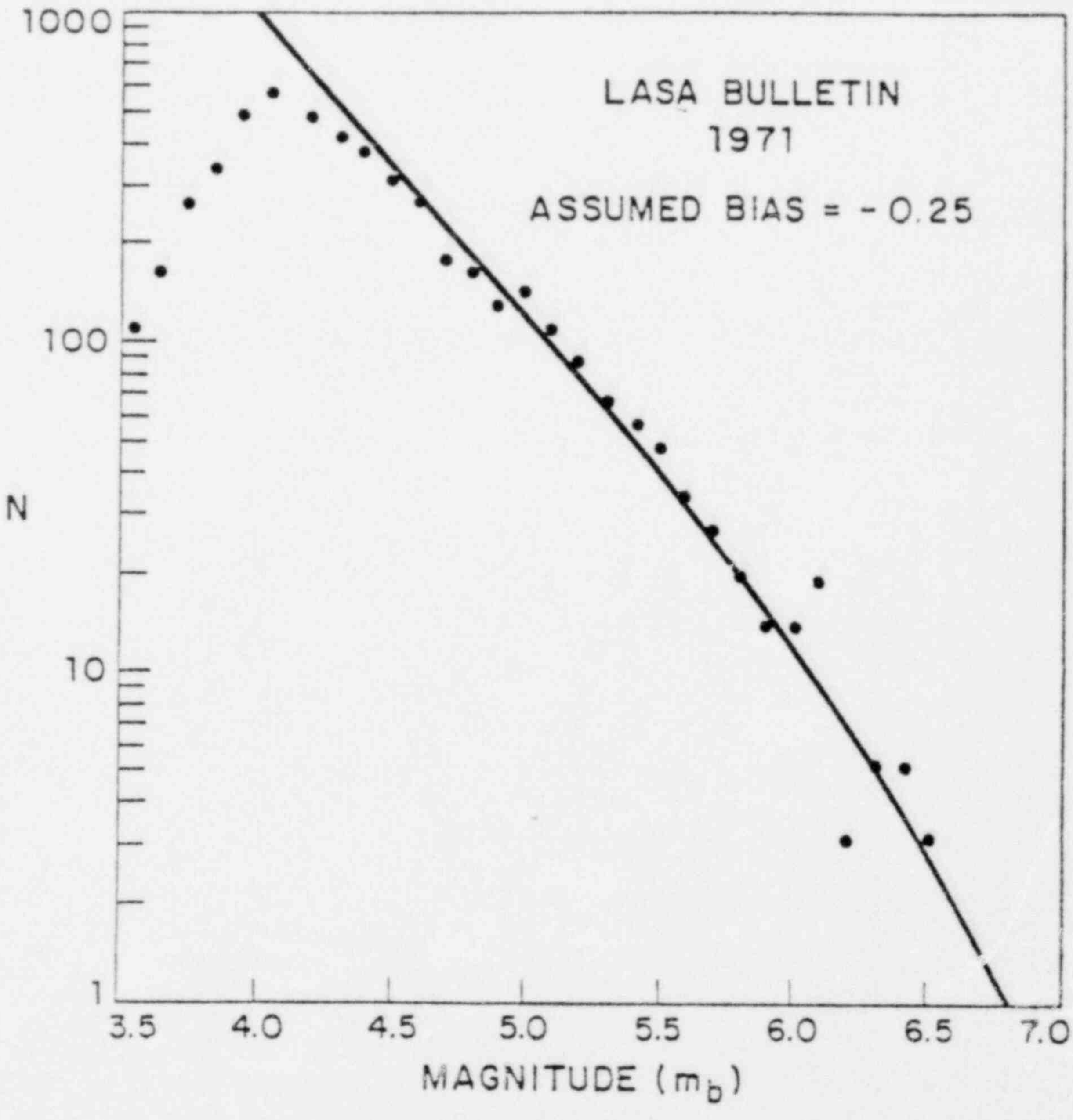

Fig. 10: Frequency-magnitude data for the Large Aperture Seismic Array (LASA) in Montana for the year 1971. The solid line is the seismicity curve shown in Figure 9.

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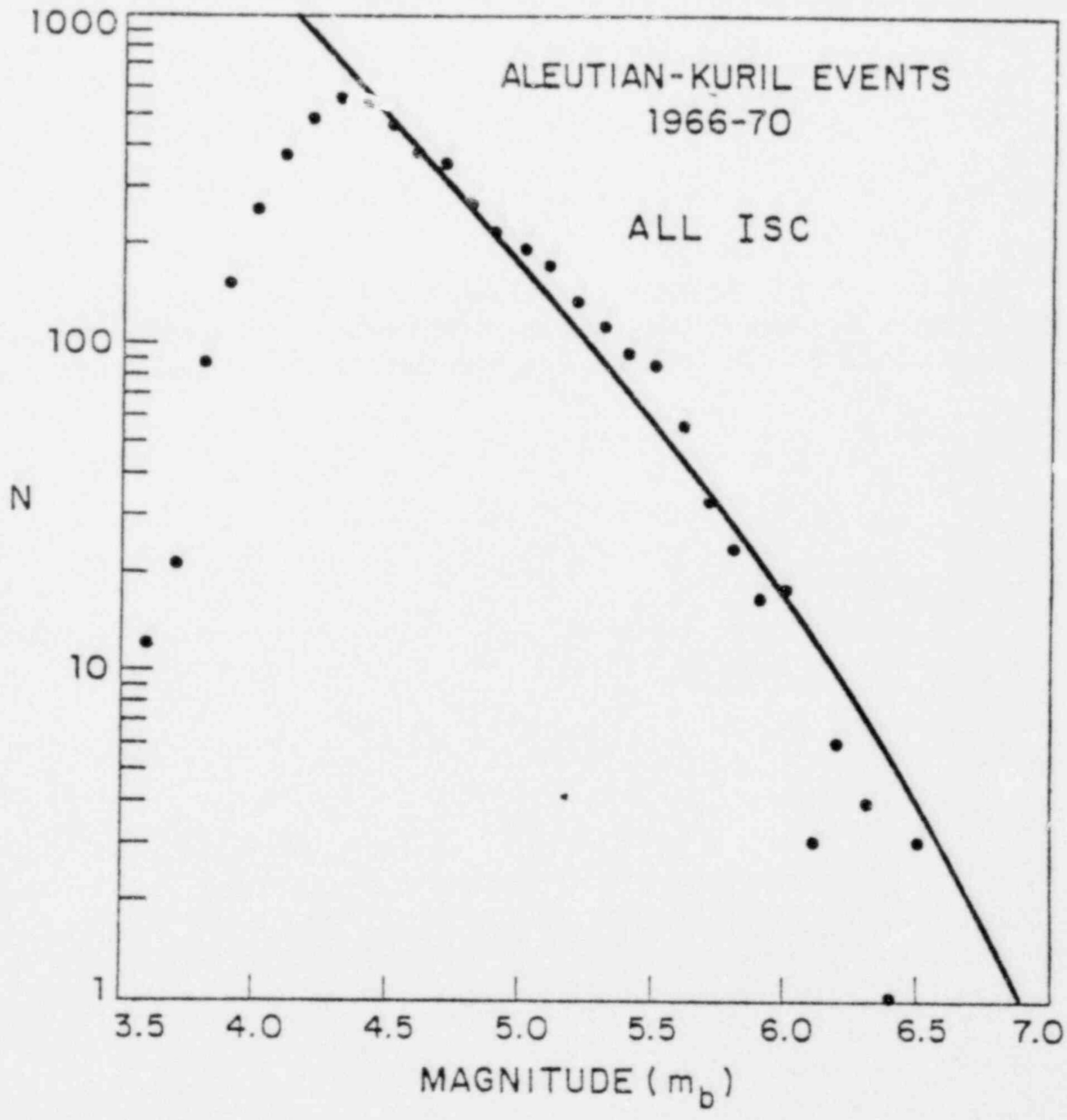

Fig. 11: Frequency-magnitude data for all events in the Aleutian-Kuril area listed in the ISC catalog, 1966-70.

this area is heavily biased by the reports from the VELA observatories, particularly for low and moderate events.

The situation is clarified in Figure 12, which shows the data for a twenty-five station network (this is the same network as that listed in Table 1, with the VELA sites BMO, TFO and UBO removed). As before, three station detection is required before an event is included. Now the shape of the network curve is clearly very different from the seismicity curve of Figure 9. In fact, it is very difficult to locate the seismicity curve in any "best fit" position by vertical movement.

On the other hand, data from the VELA arrays for this area show excellent agreement with the global seismicity curve, as shown in Figure 13. Notice again that the VELA arrays record many events with magnitudes between 6.5 and 7.0, while the 25 station network shows none (Figure 12). It is not possible to attribute this effect to the geographical location of the stations used, since there are 6 North American stations included in the 25 station network.

We can accentuate the problem further by considering only stations in Europe. Figure 14 shows the same data for a 10 station European network, which is listed in Table 2. The addition of the biases of North (1977) do not change the disagreement in shape with the VELA stations, but they do reduce many of the network magnitudes. This results from the generally positive bias of European stations (Table 2).

If the postulated seismicity curve (Figures 9 and 13) is real, there are clearly problems with the magnitudes reported by the individual stations in the network. As an example, Figure 15 shows the observations of Aleutian-Kurile events by station KEV (Kevo, Finland), which was discussed earlier (Figure 7). Either the reported magnitudes are subject

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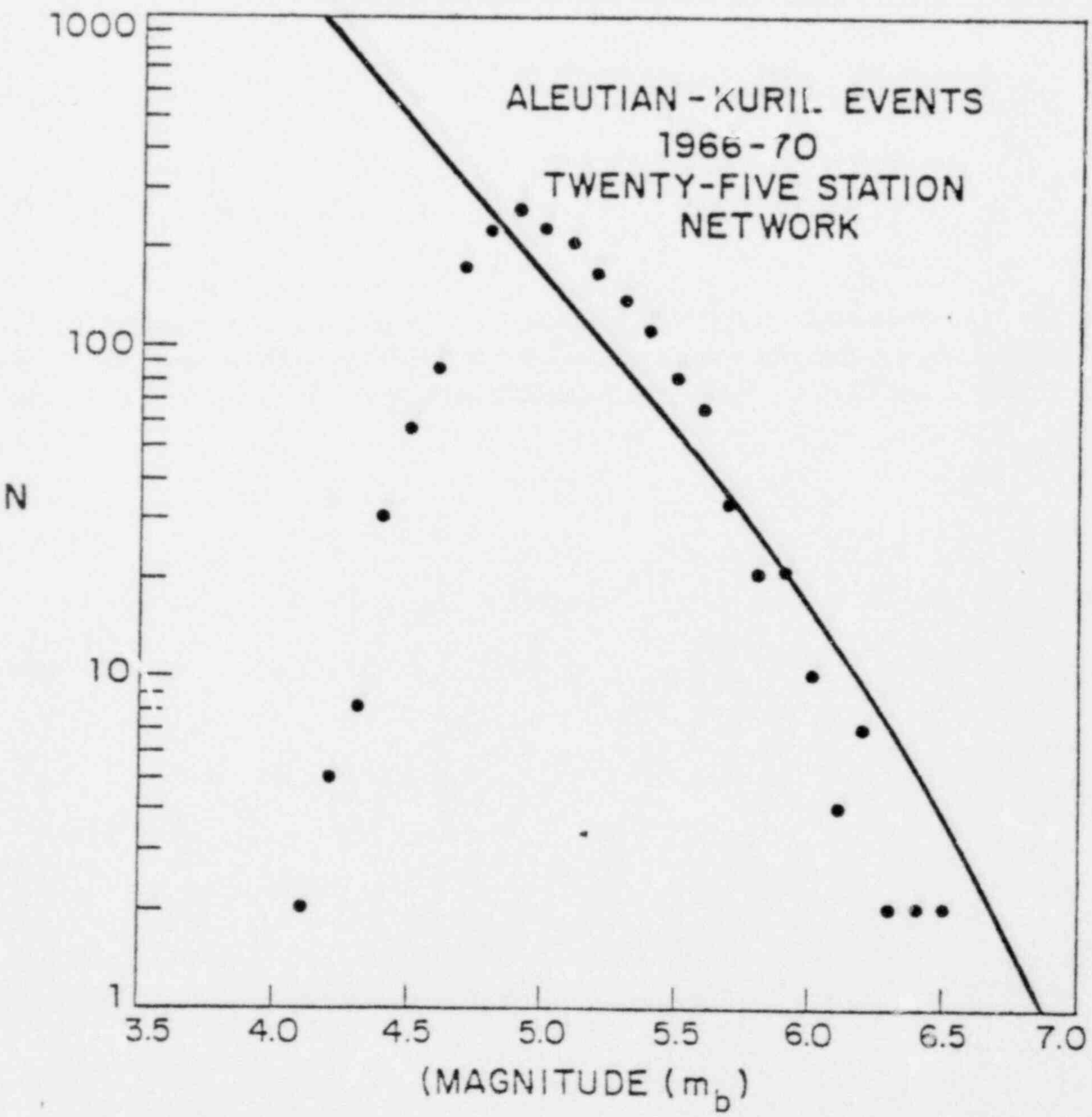

Fig. 12: Frequency-magnitude data for a 25 station network (the stations listed in Table 1, with BMO, TFO and UBO omitted).

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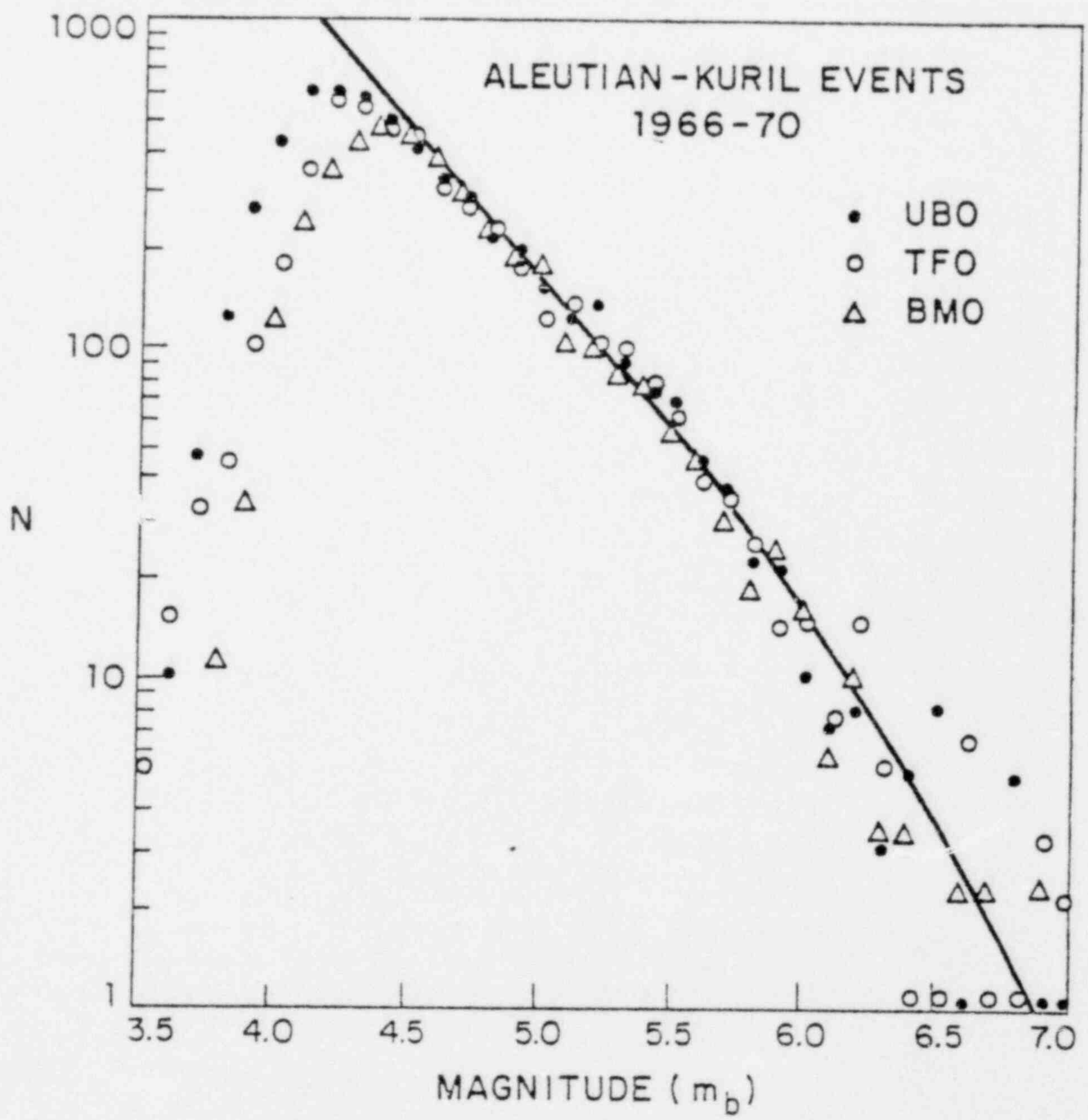

Fig. 13: Frequency-magnitude data from 3 VELA arrays for Aleutian-Kuril events. The solid curve is the same as that in Figure 9, adjusted vertically for a best fit.

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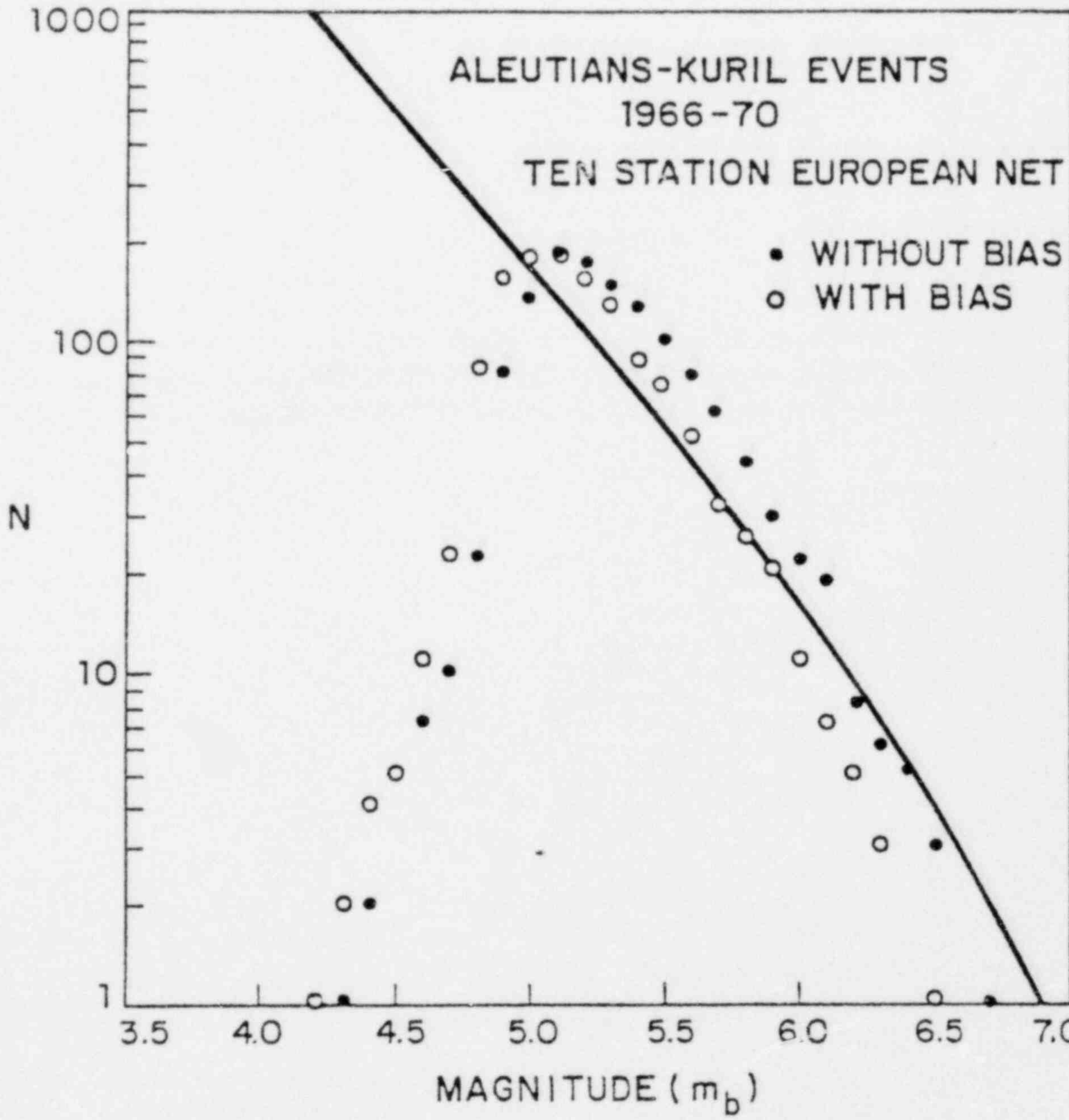

Fig. 14: Frequency-magnitude data for a 10 station European network. The stations used are listed in Table 2.

TABLE 2: 10 STATION EUROPEAN NETWORK

STATION CODE	LOCATION	BIAS (North, 1977)
BNS	Bensberg, Germany	+0.20
COP	Copenhagen, Denmark	+0.36
KEV	Kevo, Finland	+0.02
KHC	Czechoslovakia	+0.10
KJN	Kajaani, Finland	+0.14
LJU	Ljubljana, Yugoslavia	+0.29
MOX	Moxa, East Germany	+0.02
NUR	Nurmijarvi, Finland	+0.19
PRU	Czechoslovakia	+0.04
STU	Stuttgart, Germany	+0.29

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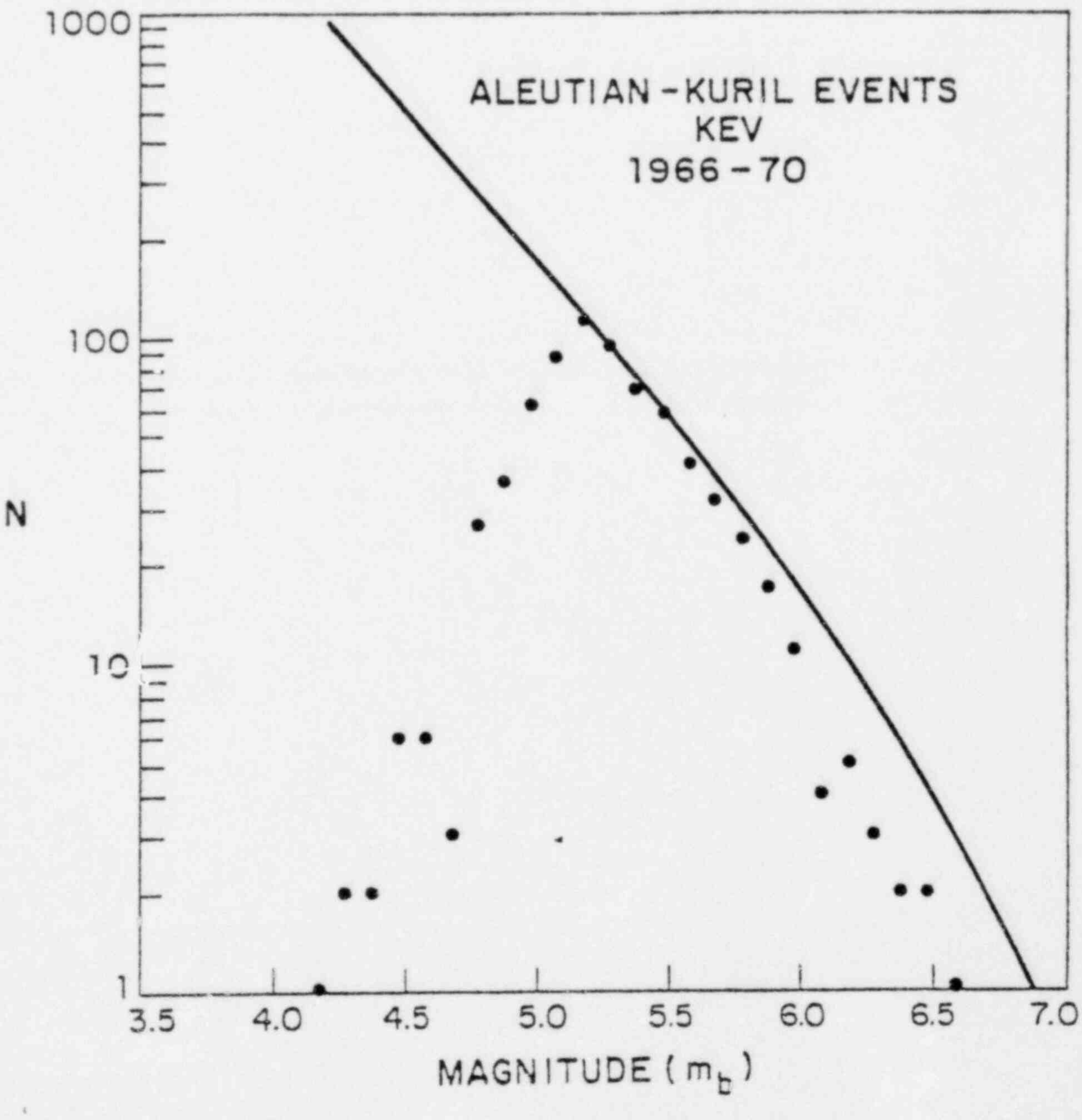

Fig. 15: Frequency-magnitude data for events in the Aleutian-Kuril area, as observed at Kevo, Finland. The solid curve is the same as those in Figures 11-14.

to strong biases, or the station is failing to report many large events. For the reasons discussed in the next section, the latter explanation seems most likely.

2.6 Interpretation

At this point we are faced with two possibilities. Either the U.S. VELA arrays (and perhaps LASA, too) have a poorly calibrated low gain channel, which leads to the systematic overestimation of the magnitudes of large events, or the magnitudes of these large events is systematically underestimated by the global network of analog seismic stations. We have been unable to find any independent evidence for the first of these alternatives, and it must be considered unlikely. It is possible, however, to suggest an explanation for the second of these alternatives, based on the dynamic range of typical analog stations, and the process of averaging which is used to obtain a network magnitude.

Any seismic station can be described by a detection probability curve. The general form of this curve, and the parameters necessary to define it, are shown in Figure 16. For our present purposes, since we are examining an earthquake catalog, we should regard this as the curve describing the probability that the station will report an amplitude of an earthquake to the analysis center (e.g. the ISC). If, for example, the station does not operate for a portion of a given time period, the maximum probability P_R will be less than 1.0.

The probability curve falls off at both low magnitudes (where the signal is not measureable) and at high magnitudes (when the instrument is off-scale). The 50% detection levels can conveniently be used to define the dynamic range of a given station. Notice that in practice the location of these points will depend to some degree on the diligence

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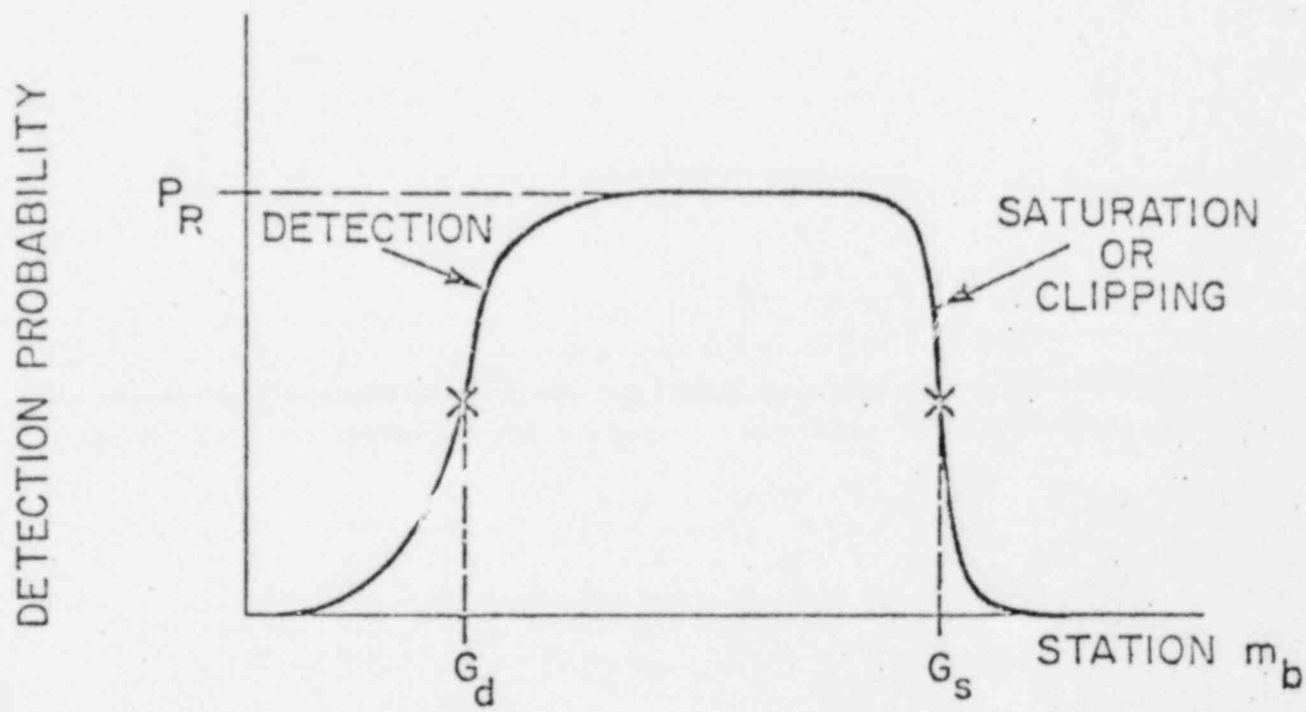

STATION DETECTION PARAMETERS

- G_d 50% DETECTION THRESHOLD
- γ_d SPREAD OF DETECTION CURVE
- G_s 50% SATURATION THRESHOLD
- γ_s SPREAD OF SATURATION CURVE
- B STATION MAGNITUDE BIAS
- P_R PROBABILITY OF REPORTING

Fig. 16: Form of the Detection Probability Curve for a seismic station.

of the operator. This is an additional complication which is hard to model; though it may be one of the most important effects in determining the dynamic range for amplitude reporting.

Amplitudes are generally measured with a rule on the seismogram, which is traced by a beam of light on photographic paper. The smallest amplitude measurable depends on the line thickness, which is typically about 1 mm. One would expect amplitudes of a few millimeters to be easily measurable. With larger events, however, problems arise. Most operators record the amplitude, zero to peak, of the first swing of the trace. When this intersects the edge of the paper, most operators will not report an amplitude. Also, when the trace amplitude becomes more than a few cm, the ability of an operator to locate the tip of the peak (or trough) will depend on the quality of the photographic recording, which is usually quite variable. And very large events, even if they do not go off-scale, are usually difficult to measure.

On purely geometrical grounds, one would expect the dynamic range of amplitude reporting to be between 2 and 3 orders of magnitude (i.e. between 2 and 3 m_b units). As we shall see, however, it seems to be between 1 and 2 orders of magnitude in practice, and "complete" recordings of amplitudes (the flat part of the detection probability curve) is usually limited to less than 1 order of magnitude (sometimes much less).

Where the station probability curve will sit on an absolute m_b scale will depend on the station magnification and the station bias B (the latter are seldom more than a few tenths of a magnitude unit: see Table 1).

The station detection probability curve has then to be considered in the light of scattering processes in the earth. These are illustrated in Figure 17. Because of scattering, an event of magnitude m will lead to a distribution of observed magnitudes at a network of stations. This distribution is often roughly normal with standard deviation about 0.3 m_b units (Von Seggern, 1973), and its mean (in the absence of station bias) will be an estimate of m. However, when the magnitude of the event approaches either the detection threshold or the clipping threshold of the stations, the distribution becomes skewed.

Effects near the detection threshold have been discussed by Ringdahl (1976) and by Christoffersson et al (1975). Those stations where scattering produces a low amplitude will not report, whereas those where a large amplitude occurs will report. This leads to a net positive bias when the station reports are averaged to produce a network magnitude. Methods can be devised for including the fact that some stations did not report an event (the maximum likelihood method) but these methods are cumbersome, and require a detailed knowledge of the detection probability curves. It does not appear possible to apply them to a data set such as the ISC catalog.

An equivalent bias arises at the clipping threshold of stations, although this has not been discussed in the literature. It is, of course, reversed in sign. When a large event occurs, those stations where scattering produces a large amplitude will usually not report, while those stations that receive a low amplitude will report. The result is a negative bias to the network magnitudes reported for large events. This negative bias will be quite substantial, up to 0.5 or 1 magnitude unit, and can adequately account for the difference between the VELA seismicity curve and the ISC catalog seismicity curve.

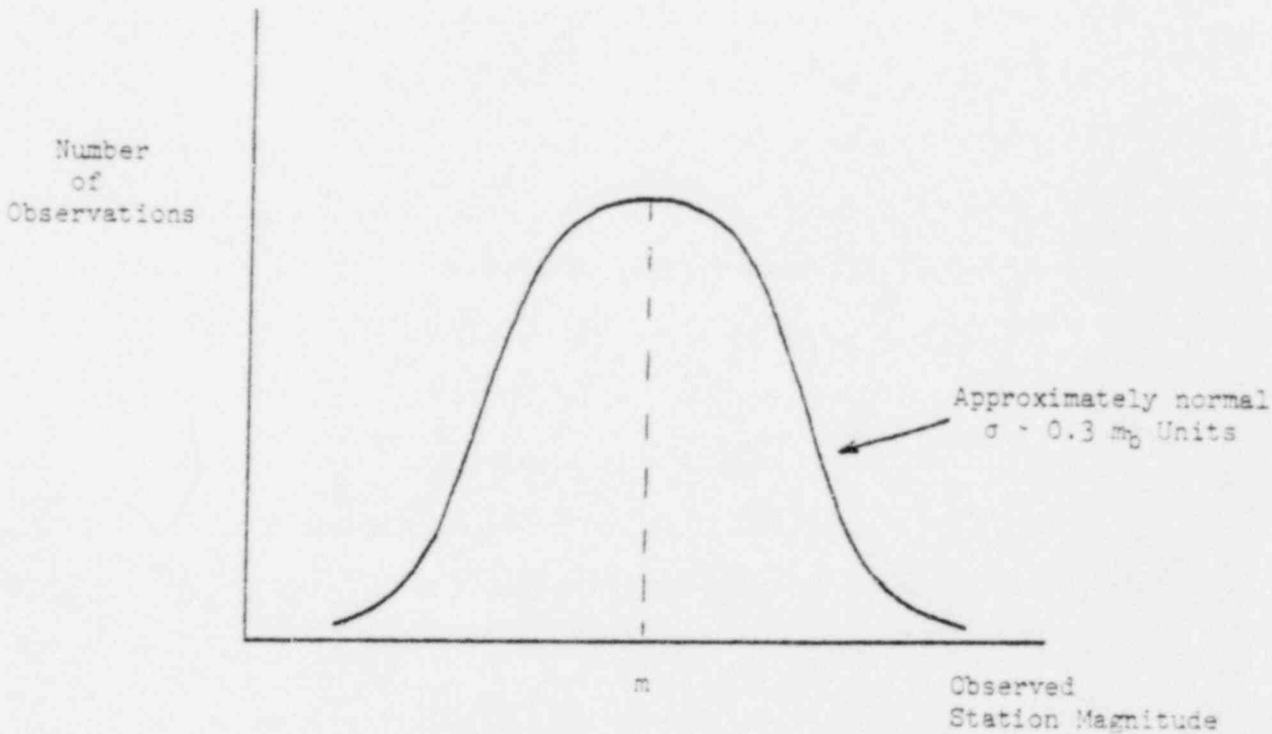

Fig. 17: The effect of scattering on observed amplitudes. An event of magnitude m will lead to a distribution of magnitudes at a network of stations. This distribution is usually approximately normal.

We can illustrate our argument by using data from a single station. Figure 18 shows the data for EUR (Eureka, Nevada). The left hand portion of this figure shows a conventional interpretation of the reporting characteristics of this station. An arbitrary straight line is fitted to the data, and detection and clipping thresholds (indicated by arrows) are determined at m_b=4.5 and 6.3 respectively. In the right hand portion of the figure, the VELA seismicity curve is used (EUR is quite close to the observatory UBO). In this interpretation the station fails to report many events for m_b greater than 5.5. The thresholds are now 4.3 and 6.1, and "complete" reporting is limited to the range 4.7 to 5.5. A similar interpretation for staton KEV using Figure 15 suggest that this station carries out "complete" reporting over an even smaller range, perhaps as little as 0.3 m_b units (from 5.2 to 5.5).

A different representation of the same phenomenon for station EUR is shown in Figure 19. Here, for each interval of 0.1 m_b units of UBO reported magnitudes, we have averaged the difference in reported magnitude between EUR and UBO for events in the ISC catalog during the period 1966-70. The theoretical interpretation of such a data set has been discussed in detail by Chinnery and Lacoss (1976). If the detection probability curve for EUR were horizontal (Figure 16) then this plot should be horizontal too. The presence of a detection threshold shows as pronounced positive biases as low magnitudes. There is a hint of a flat portion of the curve in the vicinity of 5.0-5.5, and then the data continue becoming more negative. This must be interpreted as being due to a clipping threshold. In general terms, Figure 19 is entirely consistent with the right hand preferred interpretation of Figure 18.

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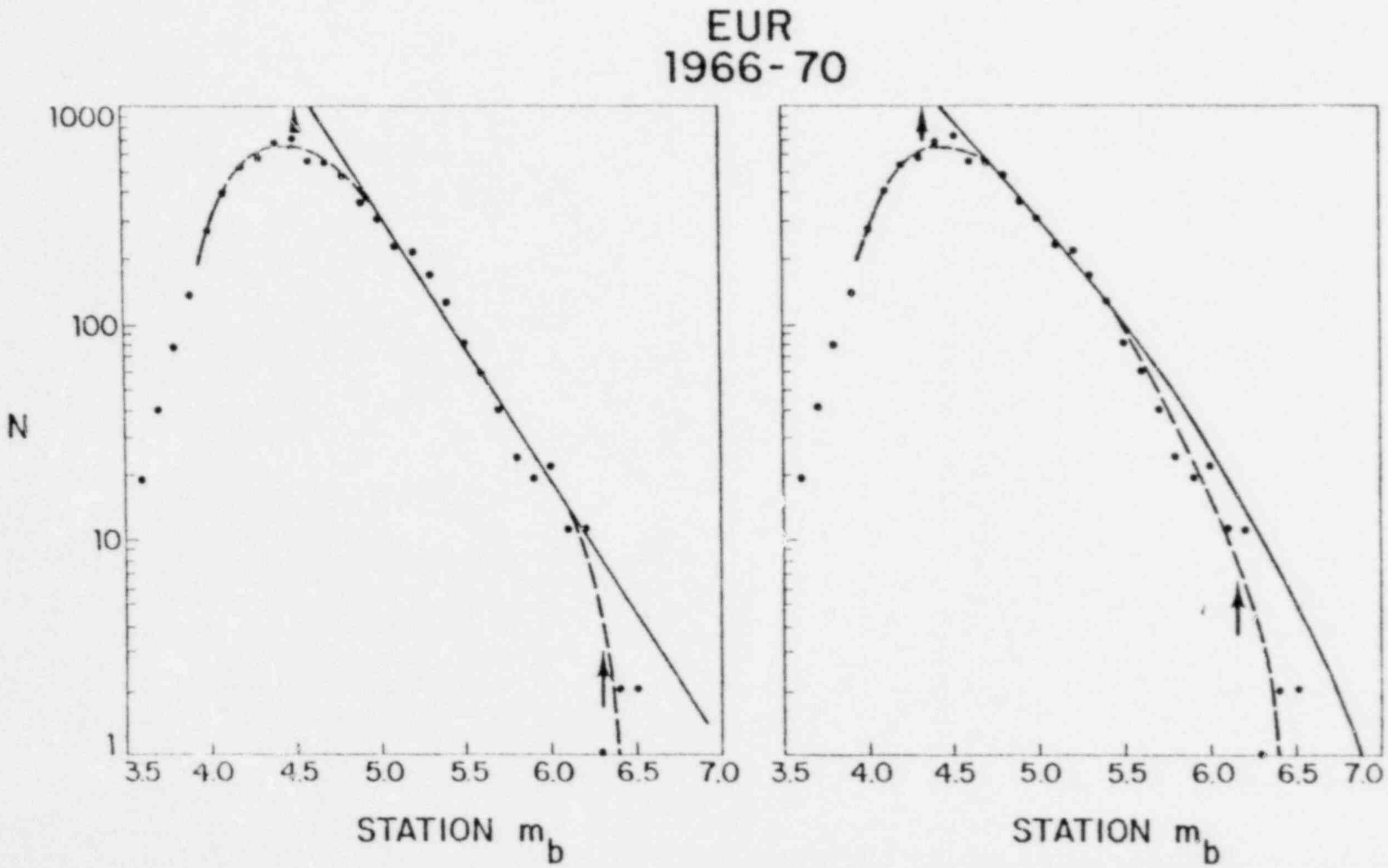
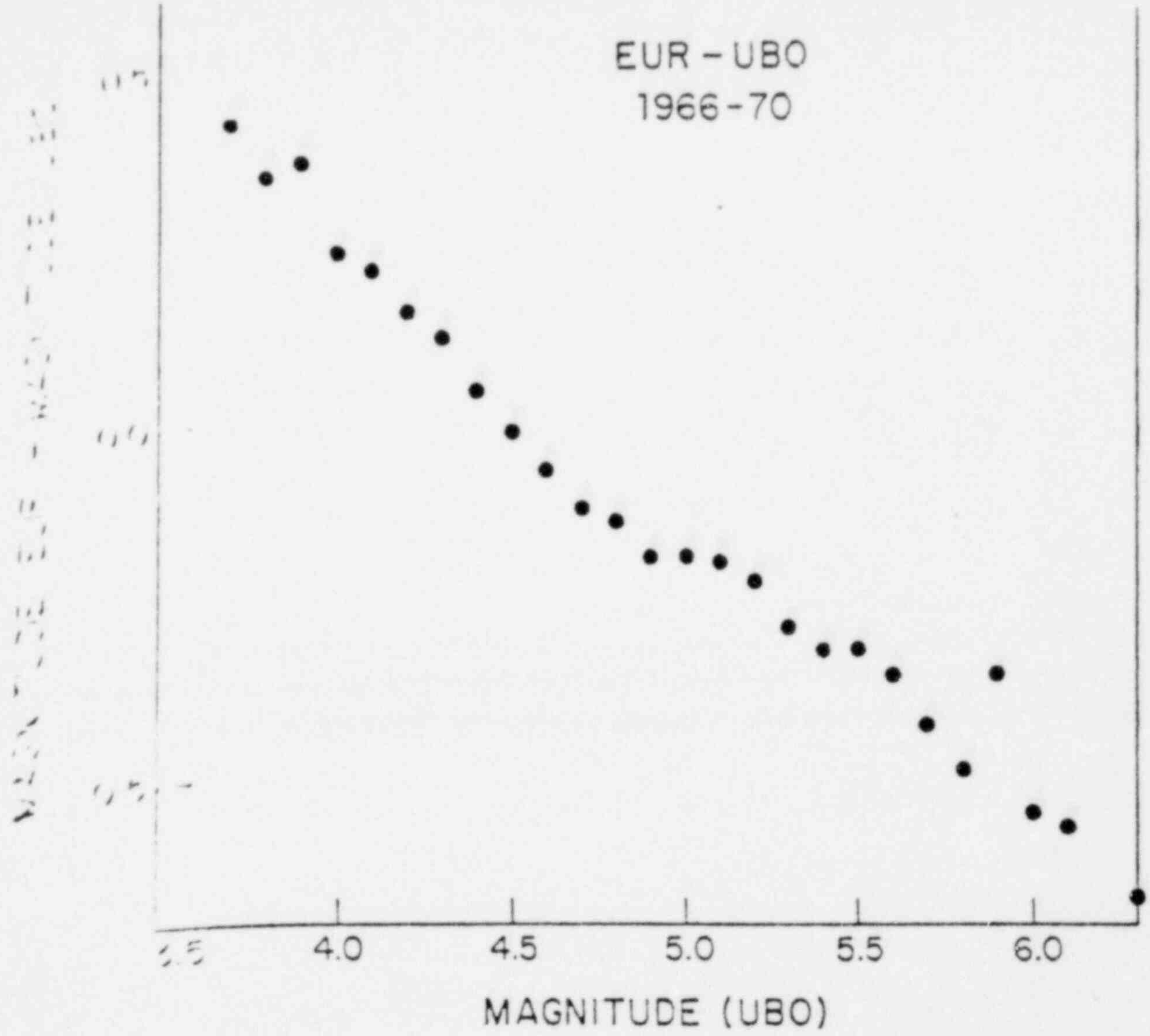


Fig. 18: Two interpretations of the reporting frequency of station EUR (Eureka, Nevada). The right hand interpretation is preferred.



Each point is the average difference between the station magnitudes of EUR and UBO for all events listed in the ISC catalog, plotted as a function of the UBO magnitude.

2.7 Discussion

The results described above provide convincing evidence that instrumental clipping of analog stations is an important problem, and that the magnitudes of larger events published in the ISC catalog are biased low and unreliable. A corollary to this conclusion is that it is virtually impossible to study the seismicity characteristics of different regions using this (or similar) catalogs, since each region is "monitored" by a different set of stations, with different operating and reporting characteristics.

The VELA arrays appear to be unique in their wide dynamic range, and, until a global network of digital stations becomes available and has accumulated a substantial data set, the VELA data is the only reliable source of information on upper bounds. So far, we have not discovered any evidence for regional variations in seismicity using these arrays. As an example, Figure 20 shows data for shallow seismicity along the South American subduction zone. The global curve (Figure 9) is again an excellent fit.

If we assume that the VELA seismicity curve is valid and represents saturation of the m_b scale, we can use similar arguments to those of Chinnery and North (1975) to construct an m_b-moment relationship. Assuming that the relationship between m_b and M_s at low magnitudes is

$$m_b = M_s + 0.5 \qquad (2.3)$$

(see, for example, Lambert et al, 1974), then the form of the m_b-moment curve is as shown in Figure 21. Some doubt about the constant in equation 2.3 remains, so the horizontal location of the m_b-moment curve is not well defined.

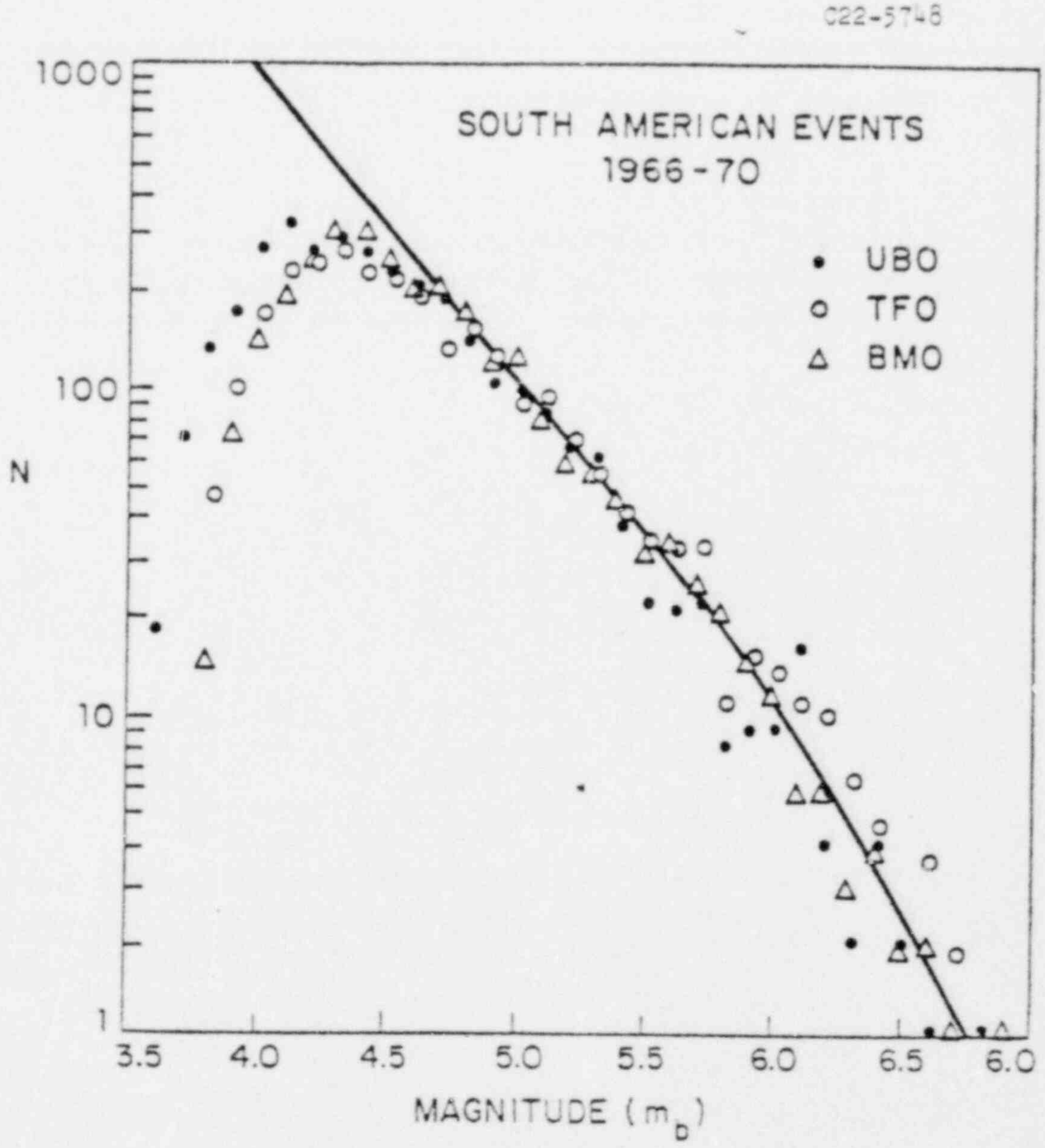


Fig. 20: Frequency-magnitude data for South American events observed at 3 VELA arrays. The solid curve is the same as that in Figure 9, adjusted vertically for a best fit.

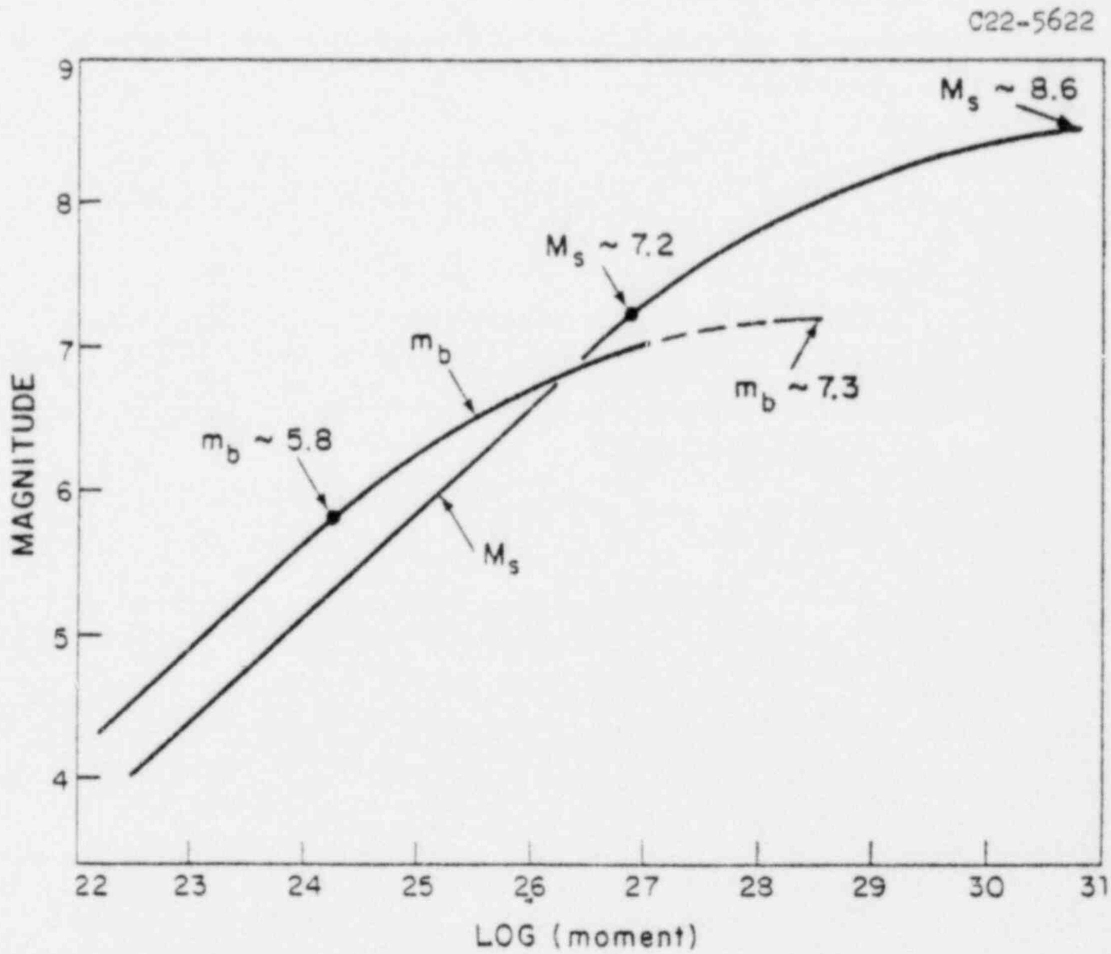


Fig. 21: An empirical m_b-moment relationship consistent with the VELA seismicity curve (Figure 9). The M_s-moment relationship from Chinnery and North (1975) is shown for comparison.

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APPENDIX

Progress Report: New England Crust and Upper Mantle Structure

The recent establishment of the northeastern seismic array has allowed us to construct a preliminary model of the crust and upper mantle structure beneath New England. Because the array has only been in full operation for approximately 2 years, the dataset is limited, and we have analyzed the data using a variety of techniques including:

1. observations of relative JB residuals
2. a time term analysis using P_n arrivals
3. three-dimensional modeling using teleseismic P-waves
4. analysis of array diagrams
5. refraction studies

Preliminary results indicate a crustal thickening under central New Hampshire coupled with a slight crustal thickening westward towards the North American craton. There is also some suggestion of a region of relatively low velocity in the upper mantle beneath central New Hampshire and southern Maine.

Methods of Analysis and Results

The relative arrival times of teleseismic P waves were read from enlarged copies of 16 mm develocorder film. In general, the first few cycles exhibit coherence across the array so relative arrival measurements were taken from a prominent peak or trough early in the signal. This procedure was required for a number of weakly recorded teleseisms in which the first break was too emergent or obscured by noise. In this way, arrival times could be measured to 0.1 sec. Elevation corrections were applied to the data by assuming a vertical phase velocity of 6.0 km/sec and dividing this into the station elevations.

Absolute travel time residuals were calculated with respect to JB tables and are defined to be

$$R_{ij}^{JB} = T_{ij}^{obs} - T_{ij}^{JB}$$

where R_{ij}^{JB} is the absolute residual with respect to JB tables for station i, event j; T_{ij}^{obs} is the observed travel time using origin times from PDE bulletins; T_{ij}^{JB} is the theoretical travel time through a JB earth.

The residuals were reduced by calculating relative residuals with respect to a mean residual computed for each event;

$$R_{ij} = R_{ij}^{JB} - \frac{1}{N} \sum_{i=1}^{N} R_{ij}^{JB}$$

where N is the number of stations reporting P arrivals for a given event. The utilization of relative residuals reduces source effects and mislocation errors, removes errors in origin time, and reduces effects of travel path through an inhomogeneous mantle. In this way, positive residuals represent late arrivals where the waves have been slowed in the crust or upper mantle beneath the array.

There are several consistent trends in the teleseismic P wave residuals which suggest the presence of large scale regional structures in the crust and upper mantle beneath the array. The data show both azimuthal variations in residual values, and variations in average station residuals across the array.

The data were inverted to a depth of 350 km using the three dimensional modeling technique of Aki et al. (1977). Perhaps the most interesting result is the presence of a regional zone of relatively low velocity in the upper mantle beneath central New Hampshire and southern Maine. This zone of relatively low velocity correlates spatially with the Mesozoic White Mountain plutonic series. It is thought that the source of these intrusive complexes is deep-seated (Chapman, 1976), and it is possible that this anomaly is related to the formation of these plutons.

S time term analysis using P_n arrivals indicates that the variations in average station residuals may be due to variations in crustal thickness and/or velocity. This is in contrast to the observed azimuthal distribution of residuals for each station which is probably due to deeper effects. It was assumed that the distribution of average residuals is caused by crustal thickness variations, and the data were inverted to find a crustal thickness map of New England. The resulting map suggests a crustal thickening beneath central New Hampshire, with more normal thicknesses in Massachusetts and Maine. The contours of the map parallel the northeasterly trend of the Appalachians.

The variations in crustal thickness observed across the network are also supported by analysis of array diagrams. These are stereographic projections of slowness and azimuth anomalies observed from a plane wave fit to the wavefront traversing the network. These studies indicate a Moho which dips 2^o or less to the northwest. This is not surprising because it is expected that the crust would thicken from the continental margin towards the North American craton.

In addition to the above mentioned studies, an average crustal velocity model has been compiled for eastern Massachusetts and southern New Hampshire by combining results from timed quarry blasts with the time term analysis. The model is currently being used in earthquake location programs at M.I.T. and is as follows:

layer (km)	P velocity (km/sec)
0 - 7.3	5.68
7.3-26.1	6.26
26.1-38.0	7.33
Moho	8.13

Future Studies

Studies for the next year will be aimed at improving the preliminary crust and upper mantle model for New England. This will be achieved by using additional teleseismic P and PkP data. The database is currently being expanded to include readings from short period stations in Connecticut and eastern New York.

The structural models derived from the residual studies will be compared to those from long period surface wave dispersion studies. Phase velocities are presently being computed as a function of azimuth from the Quebec-Maine border event of June 15, 1973, and simple crustal models will be developed. Phase velocities will also be measured using the two station technique.

More elaborate models will be generated by performing a simultaneous inversion of phase velocity and attenuation following the techniques of Lee and Solomon (1975).

A study of the Lg phase, a short period higher mode Love wave, will be initiated to compare the effect of regional geologic structure on Lg propagation. The data will be collected using three component, digital recording event detectors developed at MIT.

References

Aki, K., A. Christoffersson, and E. S. Husebye, Determination of the three-dimensional seismic structure of the lithosphere, J. Geophys. Res, 82, 277-296, 1977.

Chapman, C. A., Structural evolution of the White Mountain magma series, Geol. Soc. Am, Mem., 146, 281-300, 1976.

Lee, W. B. and S. C. Solomon, Inversion schemes for surface wave attenuation and Q in the crust and the mantle, Geophys. J. R. Astron. Soc., 43, 47-71, 1975.

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References

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Chapman, C. A., Structural evolution of the White Mountain magma series, Geol. Soc. Am, Mem., 146, 281-300, 1976.

Lee, W. B. and S. C. Solomon, Inversion schemes for surface wave attenuation and Q in the crust and the mantle, Geophys. J. R. Astron. Soc., 43, 47-71, 1975.

Exhibit 4

RESUME

PERSONAL DATA

Name: Michael A. Chinnery

Date of Birth: 27 September, 1933

Place of Birth: London, England

Citizenship: U.S.

Marital Status: Married, two daughters

Wife's Name: Thora Elizabeth (nee Hawkey)

Wife's Place of Birth: Peterborough, Ontario, Canada

Wife's Nationality: Canadian

Military Service: Royal Air Force, 1952-54

Rank: Pilot Officer (Flying Officer, Reserve)

Branch: Fighter Control

Area: England and Egypt

Present Home Address: 110 Gray Street
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Present Business Address:

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Security Clearance: Secret

DATE.................................... SIGNED....................................

EDUCATION

High School: Brentwood School, Brentwood, Essex, England 1944-51

Praepostor 1949
School Praepostor 1951
Head of House 1951
County Major Scholarship 1950
State Scholarship 1951

Undergraduate: Corpus Christi College, Cambridge University 1954-57

Mawson Open Scholarship 1954
Caldwell Open Scholarship 1955
Foundation Scholarship 1956
Natural Sciences Tripos Part I 1956 (1st Class)
Natural Sciences Tripos Part II (Physics) 1957 (upper 2nd Class)

Graduate: Geophysics Laboratory, Department of Physics, University of Toronto, Canada 1958-62

Computation Center Fellowship 1959
Canadian Kodak Fellowship 1960
Imperial Oil Fellowship 1960
National Research Council Fellowship 1960

M.A. Thesis: "The Application of Dislocation Theory to Geodynamics" (Advisor: J. A. Steketee)

Ph.D. Thesis: "The Dynamics of the Strike-Slip Fault" (Advisor: F. S. Grant)

Degrees:

A.V.C.M.	1945	Victoria College of Music
B.A.	1957	Cambridge University
M.A.	1959	University of Toronto
M.A.	1961	Cambridge University
Ph.D.	1962	University of Toronto
M.A. (ad eundem)	1967	Brown University
Sc.D.	1977	Cambridge University

EMPLOYMENT

Trainee Engineer (computer construction and development) Plessey Company, Ilford, Essex, England	1954 (summer)
Geophysicist (seismic exploration and interpretation) Seismograph Service (England) Ltd, Keston, Kent	1957-58
Research Assistant (operate mass spectrometer) Dept. of Physics, University of Toronto, Canada	1958 (summer)
Geophysicist (field party leader; seismic, magnetic, and electromagnetic exploration) Huntec Ltd, Toronto, Canada	1959 (summer)
Lecturer (part-time) Dept. of Physics, University of Toronto, Canada	1961-62
Instructor II Dept. of Geophysics, University of British Columbia	1962-63
Assistant Professor Dept. of Geophysics, University of British Columbia	1963-65
Research Associate Dept. of Earth and Planetary Sciences, M.I.T.	1965-66
Associate Professor Dept. of Geological Sciences, Brown University Providence, Rhode Island	1966-71
Professor Dept. of Geological Sciences, Brown University	1971-73
Senior Research Associate Dept. of Earth and Planetary Sciences, M.I.T.	1973-present
Group Leader Applied Seismology Group, Lincoln Laboratory, M.I.T. Cambridge, Massachusetts	1973-present

CONSULTANT TO

Huntec Ltd, Toronto	1958-65
Arthur D. Little, Inc, Cambridge, Massachusetts	1966-73
Earth Sciences Research, Inc, Cambridge, Mass.	1969-73
Lincoln Laboratory, M.I.T.	1971-73
National Aeronautics and Space Administration	1976-present

(Lincoln Laboratory does not allow its employees to consult for industry)

FIELD WORK

Seismic exploration, Milford Haven, England	1957
Electromagnetic, resistivity, magnetic and seismic exploration in Alaska, Northwest Territories, and Alberta	1959
Gravity survey, British Columbia	1960
Shallow seismic exploration, Northern Quebec	1961
Gravity survey, Northern Ontario	1962
Shallow seismic exploration, British Columbia	1964

COURSES TAUGHT

Applied Geophysics (physics undergraduates)

Applied Geophysics (geology undergraduates)

Elasticity Theory (graduate)

Dislocation Theory (graduate)

Introduction to Geophysics (undergraduate/graduate)

Introduction to Seismology (graduate)

Earthquakes (introductory undergraduate)

Planetary Physics (undergraduate)

Data Analysis (graduate)

Tectonophysics (graduate)

plus various seminars and portions of courses

PROFESSIONAL SOCIETIES AND OFFICES HELD

American Geophysical Union
- Secretary, Tectonophysics Section, 1968-70
- Program Chairman, Tectonophysics Section, 1969 Annual Meeting
- Program Chairman, Tectonophysics Section, 1970 Annual Meeting
- Associate Editor, Journal of Geophysical Research, 1969-72
- Associate Editor, Geophysical Research Letters, 1974-76
- Member, Committee on Education and Human Resources, 1979-present
- Secretary, Seismology Section, 1980-present

Seismological Society of America
- Nominations Committee, 1974

Seismological Society of America (Eastern Section)
- Resolutions Committee, 1973
- Chairman, Executive Committee, 1973-75
- Member, Executive Committee, 1975-77

*Society of Exploration Geophysicists
- Membership Committee, 1963-65

*Royal Astronomical Society of Canada
- Secretary, Vancouver Center, 1963
- President, Vancouver Center, 1964

American Association for the Advancement of Science

*Society of the Sigma Xi
- Member, 1966-73
- Treasurer, Brown University Chapter, 1968-72

Royal Astronomical Society
- Fellow, 1973-present

*presently inactive

COMMITTEES AND MISCELLANEOUS ACTIVITIES

Resident Faculty Advisor, Acadia Residence, University of British Columbia, 1962-64

Member, Gravity Sub-committee, National Research Council of Canada, 1964-65

Associate Resident Fellow, Mead House, Brown University, 1967-69

Member, Dining Services Committee, Brown University, 1969-71

Member, Graduate Council, Brown University, 1969-71

Chairman, University Lectureships Committee, Brown University, 1971-73

Department of Geological Sciences, Brown University; committee memberships during the period 1966-73:

- Foreign language committee (chairman)
- Geology Club (chairman)
- Graduate examinations committee (chairman)
- Lecture series (chairman)
- Geophysics committee (chairman)
- Undergraduate program committee (member)
- Graduate admissions and awards committee (chairman)

Testified before the Advisory Committee on Reactor Safeguards, Nuclear Regulatory Commission, concerning seismic risk at the Seabrook nuclear power plant site, 1974

Appeared as expert witness at the licensing hearings for the Seabrook nuclear power plant, 1975

Member, Panel on Seismograph Networks, Committee on Seismology, National Academy of Sciences, 1975-77

Participant, Conference on earthquake prediction on the global scale, U.S. Geological Survey, Denver, 1976

Chairman, Advisory Committee on Earth Dynamics, N.A.S.A., 1976-77

Meeting Chairman, Summer workshop on the application of space techniques to geodynamics, N.A.S.A., Denver, 1977

Member, Working Group on Upgrading WWSSN Stations, National Academy of Sciences, 1977

Gave special invited lecture on the application of space techniques to geodynamics, International Association of Seismology and Physics of the Earth's Interior, Durham, England, 1977

Member, Panel on Storage of Digital Seismic Data, Committee on Seismology, National Academy of Sciences, 1977-78

Member, Working Group on Solid Earth Data, Committee on Data Interchange and Data Centers, National Academy of Sciences, 1977-78

Member, Proposal review panel, N.A.S.A., 1978

Chairman, Advisory Committee on Geology and Geophysics, N.A.S.A., 1978-present

Member, Space and Terrestrial Applications Advisory Committee, N.A.S.A., 1978-present

Participant, Conference on Seismic Gaps, U.S. Geological Survey, Boston, 1978

I.A.S.P.E.I. representative to joint I.U.G.G./I.U.G.S. working group to formulate a post-geodynamics program for the 1980's, Washington, 1978

Member, Group of Experts study of seismicity in the Eastern U.S., Nuclear Regulatory Commission, 1978-present

Member, Proposal review panel, N.A.S.A., 1979

Participant, Conference on the Determination of Earthquake Parameters, U.S. Geological Survey, Denver, 1979

Member, Seismic Research Review Panel, Vela Seismological Center, U.S. Air Force, 1979

Chairman, Study on Geophysical Data Policy, Geophysics Research Board, National Academy of Sciences, 1979-present

I.C.G. delegate to Symposium on Quantitative Methods of Assessing Plate Motions, I.U.G.G., Canberra, Australia, 1979

Gave technical presentation to the Nuclear Regulatory Commission on the application of probabalistic methods to the estimation of seismic risk, Washington, 1980

Member, Panel on Data Problems in Seismology, Committee on Seismology, National Academy of Sciences, Woods Hole, 1980

PUBLICATIONS

The following list includes a variety of different kinds of publications. Papers in scientific journals are indicated by an asterisk (*).

1. Chinnery, M.A., The application of dislocation theory to geodynamics, M.A. thesis, University of Toronto, 88pp., 1959.

2. *Chinnery, M.A., Some physical aspects of earthquake mechanism, J. Geophys. Res., 65, 3852-54, 1960.

3. *Chinnery, M.A., Terrain corrections for airborne gravity gradient measurements, Geophysics, 26 480-89, 1961.

4. *Chinnery, M.A., The deformation of the ground around surface faults, Bull. Seism. Soc. Am., 51, 355-72, 1961.

5. Chinnery, M.A., The dynamics of the strike-slip fault, Ph.D. thesis, University of Toronto, 138pp., 1962.

6. *Chinnery, M.A., The stress changes that accompany strike-slip faulting, Bull. Seism. Soc. Am., 53, 921-32, 1963.

7. *Chinnery, M.A., The strength of the earth's crust under horizontal shear stress, J. Geophys. Res., 69, 2085-89, 1964.

8. *Chinnery, M.A., The vertical displacements associated with transcurrent faulting, J. Geophys. Res., 70, 4627-32, 1965.

9. *Grant, F.S., Gross, W.H., and Chinnery, M.A., The shape and thickness of an Archean greenstone belt by gravity methods, Can. J. Earth Sci., 2, 418-24, 1965.

10. *Chinnery, M.A., Secondary faulting I: Theoretical aspects, Can. J. Earth Sci., 3, 163-74, 1966.

11. *Chinnery, M.A., Secondary faulting II: Geological aspects, Can. J. Earth Sci., 3, 175-90, 1966.

12. Toksoz, M.N., and Chinnery, M.A., Seismic travel times from Longshot and structure of the mantle (abstract), Trans. Am. Geophys. Union, 47, 164, 1966.

13. Chinnery, M.A., The dislocation fault model with a variable discontinuity (abstract), Trans. Am. Geophys. Union, 47, 166, 1966.

14. *Chinnery, M.A., and Toksoz, M.N., P-wave velocities in the mantle below 700 km, Bull. Seism. Soc. Am., 57, 199-226, 1967.

15 *Chinnery, M.A., Leombruno, W., McConnell, R.K., and O'Brien, J., Observations of fatigue deformation in contact loading, ASME publication 67-DE-53, 12pp., 1967.

16. Chinnery, M. S., The vertical displacements associated with transcurrent faulting, in Proceedings of the VESIAC conference on the current status and future prognosis for understanding the source mechanism of shallow seismic events in the 3 to 5 magnitude range, VESIAC report 7885-1-X, 299-306, 1967.

17. *oksoz, M. N., Chinnery, M. A., and Anderson, D. L., Inhomogeneities in the earth's mantle, Geophys. J., 13, 31-59, 1967.

18. Chinnery, M. A., Evidence for lateral variations in the lower mantle (abstract), Trans. Am. Geophys. Union, 48, 194, 1967.

19. Chinnery, M. A., Source time function for a wrench fault movement (abstract), Trans. Am. Geophys. Union, 48, 203, 1967.

20. Chinnery, M. A., Theoretical investigations of the mechanism of faulting, in U.S. Upper Mantle Project Progress Report, 126-7, 1967.

21. *Chinnery, M. A., and Petrak, J. A., The dislocation fault model with a variable discontinuity, Tectonophysics, 5, 513-29, 1968.

22. Chinnery, M. A., and Rodgers, D. A., The stressed zone at the lower edge of a strike-slip fault (abstract), Trans. Am. Geophys. Union, 49, 299, 1968.

23. Chinnery, M. A., Earthquake magnitude and source parameters (abstract), Earthquake Notes, 39, 13, 1968.

24. Chinnery, M. A., Measurement of the first and second derivatives of the travel time curve using LASA (abstract), Geol. Soc. Am. Special Paper 101, 294, 1968.

25. Chinnery, M. A., Direct measurement of the second derivative of the travel time curve (abstract), Geol. Soc. Am. Special Paper 115, 215, 1968.

26. *Chinnery, M. A., Velocity anomalies in the lower mantle, Phys. Earth & Plan. Int., 2, 1-10, 1969.

27. *Chinnery, M. A. Theoretical fault models, Pubs. Dom. Obs. Ottawa, 37, 211-23, 1969.

28. Chinnery, M. A., Review of "Non-elastic Processes in the Mantle", edited by D. C. Tozer, Trans. Am. Geophys. Union, 50, 497, 1969.

29. Rodgers, D. A., and Chinnery, M. A., The displacements and strains associated with a curved strike-skip fault (abstract), Trans. Am. Geophys. Union, 50, 238, 1969.

30. Chinnery, M. A., The velocity anomaly at 2000 km depth (abstract), Trans. Am. Geophys. Union, 50, 244, 1969.

31. *Chinnery, M. A., Earthquake magnitude and source parameters, Bull. Seism. Soc. Am., 59, 1969-82, 1969.

32. *Chinnery, M. A., Earthquakes and the Chandler wobble, Comments on Earth Sci.: Geophys., 1, 1-7, 1970.

33. *Chinnery, M. A., Earthquake displacement fields, in Earthquake Displacement Fields and the Rotation of the Earth, edited by Mansinha and others, Reidel Press, The Netherlands, 17-38, 1970.

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35. Rodgers, D. A., Chinnery, M. A., and McConnell, R. K., An assessment of the glacial rebound mechanism for earthquakes in the Eastern U. S. (abstract), Trans. Am. Geophys. Union, 52, 277, 1971.

36. Chinnery, M. A., and Jovanovich, D. B., The effect of the lithosphere-asthenosphere boundary on earthquake displacement fields (abstract), Trans. Am. Geophys. Union, 52, 343, 1971.

37. Chinnery, M. A., P-wave arrivals at the Large Aperture Seismic Array, in Upper Mantle Project U.S. Final Report, 56, 1971.

38. Chinnery, M. A., Theoretical and field investigations of the mechanics of faulting, in Upper Mantle Project U.S. Final Report, 164, 1971.

39. Chinnery, M. A., Investigations of P-wave arrivals at LASA, Final Technical Report, ARPA Contract F44620-68-C-0082, 85pp, 1971.

40. Wells, F. J., and Chinnery, M. A., Variations in the annual component of polar motion at individual observatories (abstract), Trans. Am. Geophys. Union, 53, 345, 1972.

41. Jovanovich, D. B., and Chinnery, M. A., Evidence for a cusp in the travel time curve at 35° (abstract), Trans. Am. Geophys. Union, 53, 452, 1972.

42. *Chinnery, M. A., and Wells, F. J., On the correlation between earthquake occurrence and disturbances in the path of the rotation pole, in Rotation of the Earth, edited by Melchior and Yumi, Reidel Publishing Co., The Netherlands, 215-220, 1972.

43. Rodgers, D. A., and Chinnery, M. A., A revised velocity structure for New England (abstract), Trans. Am. Geophys. Union, 53, 452, 1972.

44. *Rice, J. R., and Chinnery, M. A., On the calculation of changes in the earth's inertial tensor due to faulting, Geophys. J., 29, 29-90, 1972.

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57. Chinnery, M. A., A Statement before the AEC Advisory Committee on Reactor Safeguards, October 31, 1974, 11pp.

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62. Landers, T. E., and Chinnery, M. A., Spectral analysis of earthquake occurrence rates, in Seismic Discrimination, Semi-Annual Technical Summary, Lincoln Laboratory, M.I.T., 31 December 1974, 49-51.

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65. Chinnery, M. A. and North, R. G., Global frequency-magnitude relationships and their implications (abstract), Earthquake Notes, 46, 55, 1975.

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67. Chinnery, M. A., (Editor), Seismic Discrimination 23rd Semi-Annual Technical Summary, Lincoln Laboratory, M.I.T., 30 June 1975.

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69. Chinnery, M. A., and Landers, T. E., Correlations between seismic activity in widely separated regions, in Seismic Discrimination, Semi-Annual Technical Summary, Lincoln Laboratory, M.I.T., 30 June 1975, 56.

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75. Chinnery, M. A., Problems in magnitude estimation, in Seismic Discrimination, Semi-Annual Technical Summary, Lincoln Laboratory, M.I.T., 31 December 1975, 1-2.

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77. Christoffersson, L. A., Lacoss, R. T., and Chinnery, M. A., Estimation of network magnitude and station detection parameters, in Seismic Discrimination, Semi-Annual Technical Summary, Lincoln Laboratory, M.I.T., 31 December 1975, 6-8.

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79. Chinnery, M. A., Damaging earthquake probability studies in the Eastern U.S. and their potential application to nuclear power plant siting (abstract), Geol. Soc. Am. Abstracts with Programs, 8, 150-1, 1976.

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86. Chinnery, M. A., and Lacoss, R. T., Magnitude differences between station pairs, *in* Seismic Discrimination, Semi-Annual Technical Summary, Lincoln Laboratory, M.I.T., 30 June 1976, 14-15.

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88. Chinnery, M. A., (Editor), Seismic Discrimination, 26th Semi-Annual Technical Summary, Lincoln Laboratory, M.I.T., 31 March 1977.

89. Chinnery, M. A., and Johnston, J. C., Estimation of station detection characteristics from bulletin data, *in* Seismic Discrimination, Semi-Annual Technical Summary, Lincoln Laboratory, M.I.T., 31 March 1977, 44-6.

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91. Chinnery, M. A., Correlation of seismic activity with changes in the rate of rotation of the earth, *in* Seismic Discrimination, Semi-Annual Technical Summary, Lincoln Laboratory, M.I.T., 31 March 1977, 99.

92. Chinnery, M. A., Major problems of geodynamics, Paper presented to NASA Earth Dynamics Summer Workshop, Boulder, Colorado, July 18-23, 1977.

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102. Chinnery, M. A., Scatter in observed m_b values, in Seismic Discrimination, Semiannual Technical Summary, Lincoln Laboratory, M.I.T., 31 March 1979, 48-9.

103.*Chinnery, M. A., A comparision of the seismicity of three regions of the Eastern U.S., Bull. Seism. Soc. Am., 69, 757-772, 1979.

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111. Chinnery, M. A., Generation of a synthetic arrivals list, in Seismic Discrimination, Semiannual Technical Summary, Lincoln Laboratory, M.I.T., 31 March 1980.

112. Chinnery, M. A. and Tiberio, M. A., Investigation of the properties of CD Network III using synthetic data, U. S. Delegation paper US/GSE/7 presented to the Committee on Disarmament, Geneva, 7 July 1980. Also presented to NATO Advanced Study Institute on Identification of Seismic Sources, Oslo, Norway, 8-18 September, 1980.

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