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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
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Lincoln Laboratory

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NRC Research and Technical Assistance Report

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

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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.

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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

$$\log E = a_o + b_o M \qquad (1.1)$$

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

$$\log N = a - bM \qquad (1.2)$$

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

$$M_o = \mu LWD \qquad (1.3)$$

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

$$\Delta\sigma = \eta \frac{\mu D}{W} \qquad (1.4)$$

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

$$M_o \simeq 2LW^2\Delta\sigma \qquad (1.5)$$

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 coworkers, 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

$$\log N_I = A - \gamma(K - K_o) \tag{1.7}$$

(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)

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

while for Japan he found a better fit with

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

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.

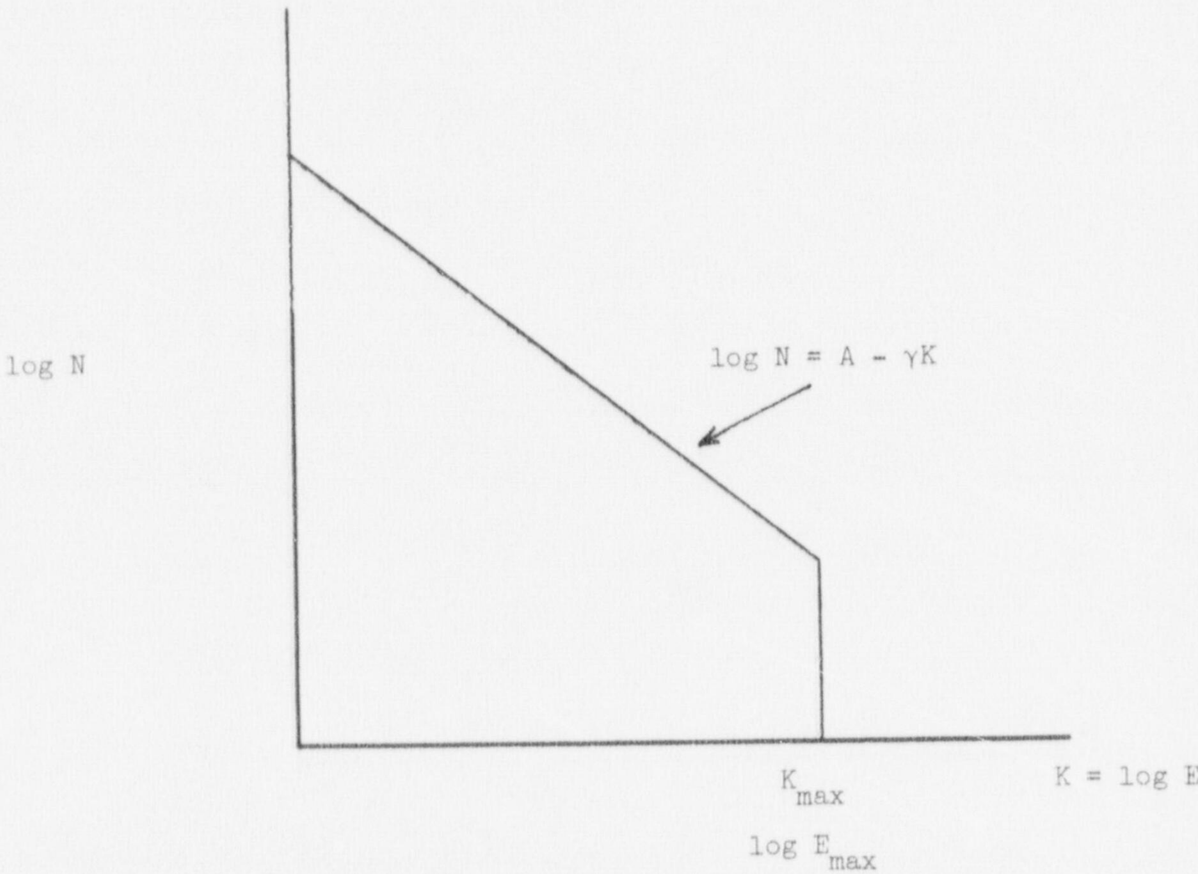

Fig. 1. Riznichenko postulates a clear-cut upper bound to total earthquake energy E, and assumes a linear frequency-energy relation for energy values below E_{max}.

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 not 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

$$P = 1 - e^{-t/T} \tag{1.14}$$

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

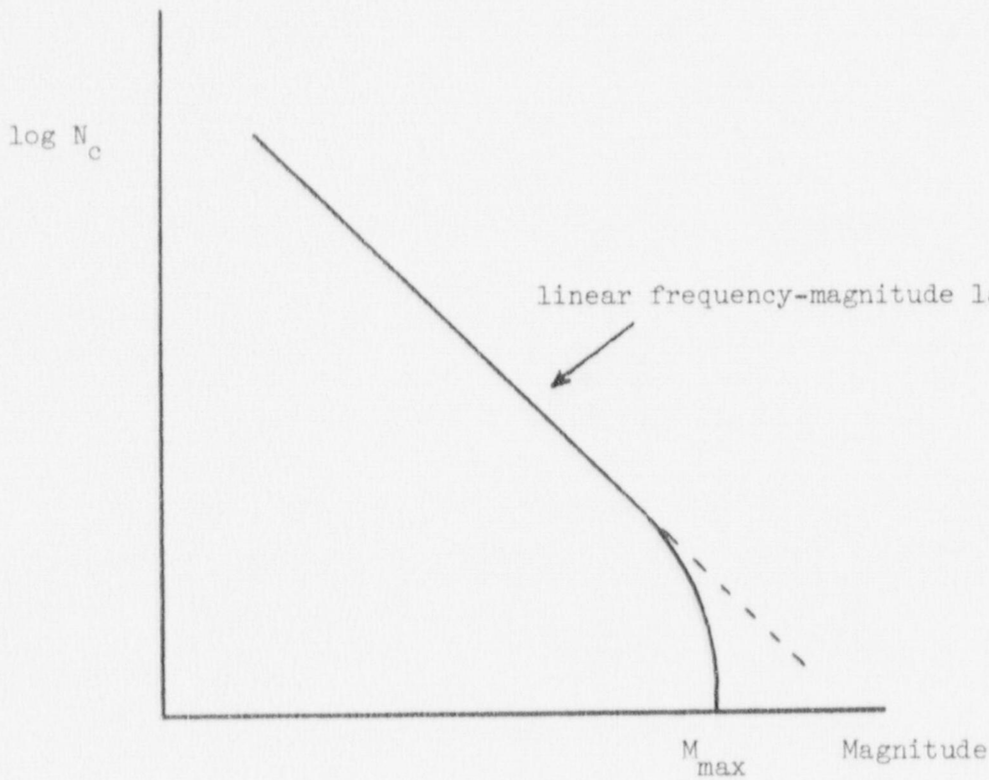

Fig 2: Ideal effect of an upper bound to earthquake magnitude, using cumulative frequency-magnitude statistics.

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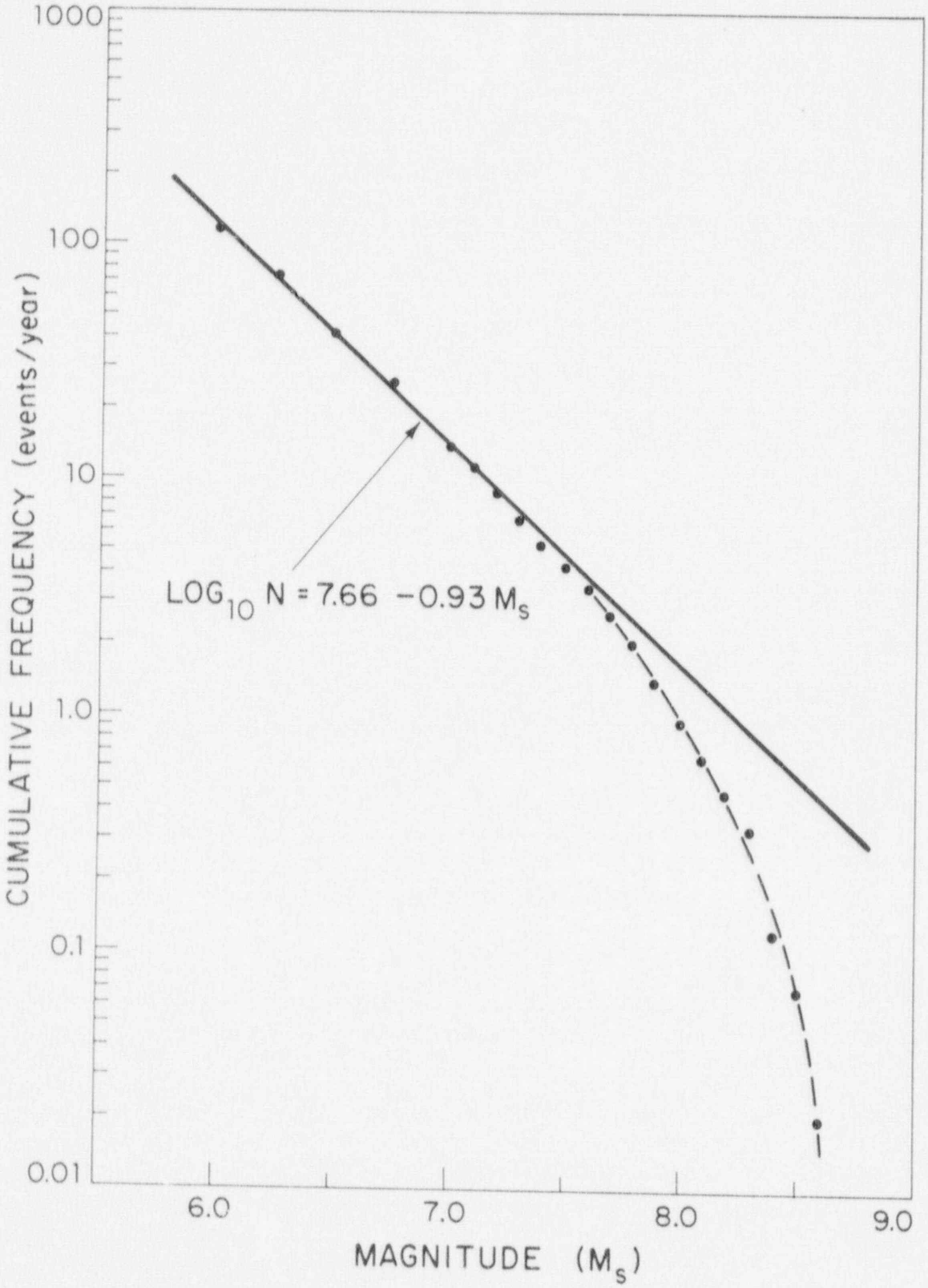

Fig. 3: Data from Gutenberg and Richter (1954).

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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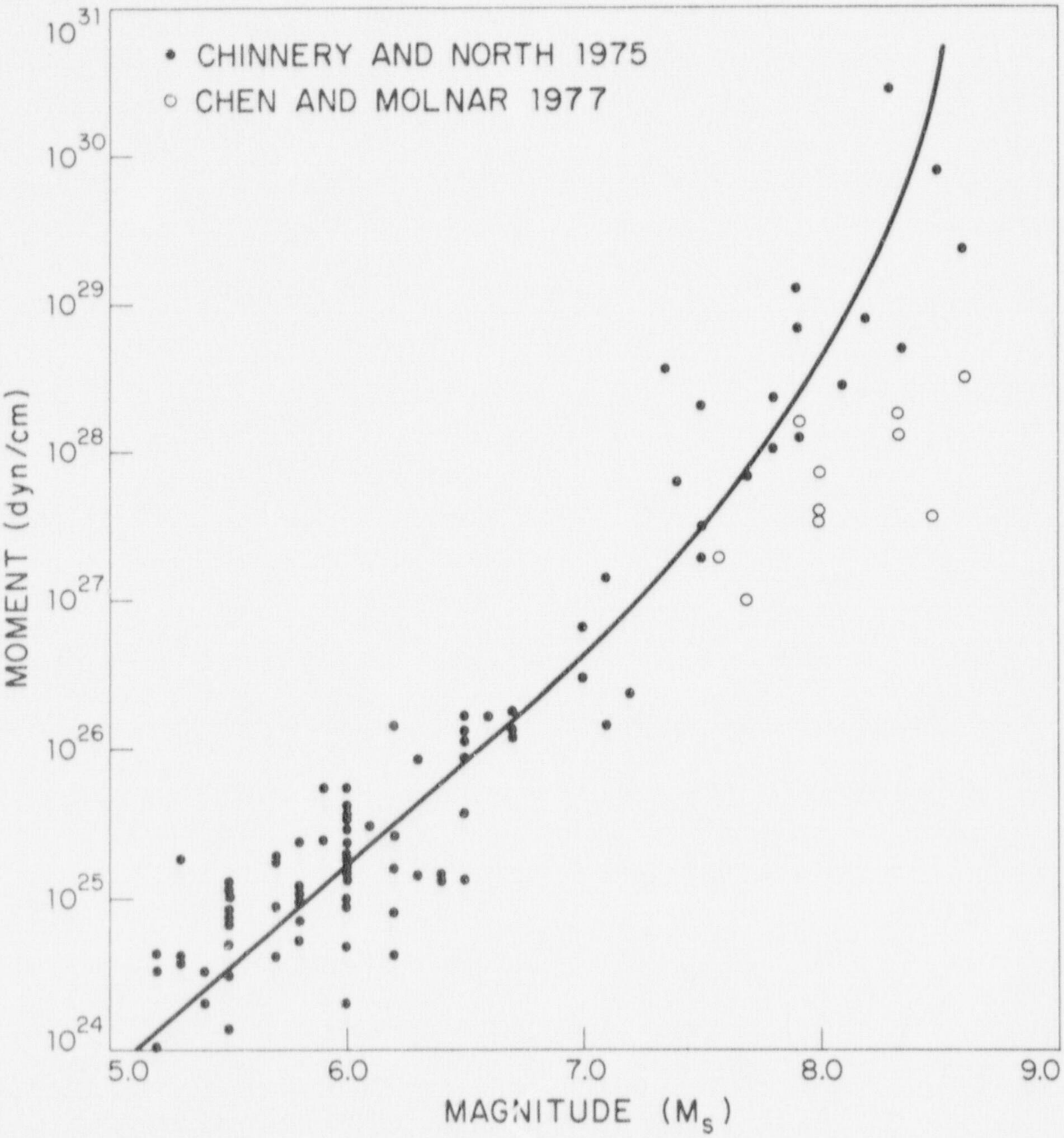

Fig. 4: Complilation of 87 published estimates of seismic moment as a function of surface wave magnitude M_s.

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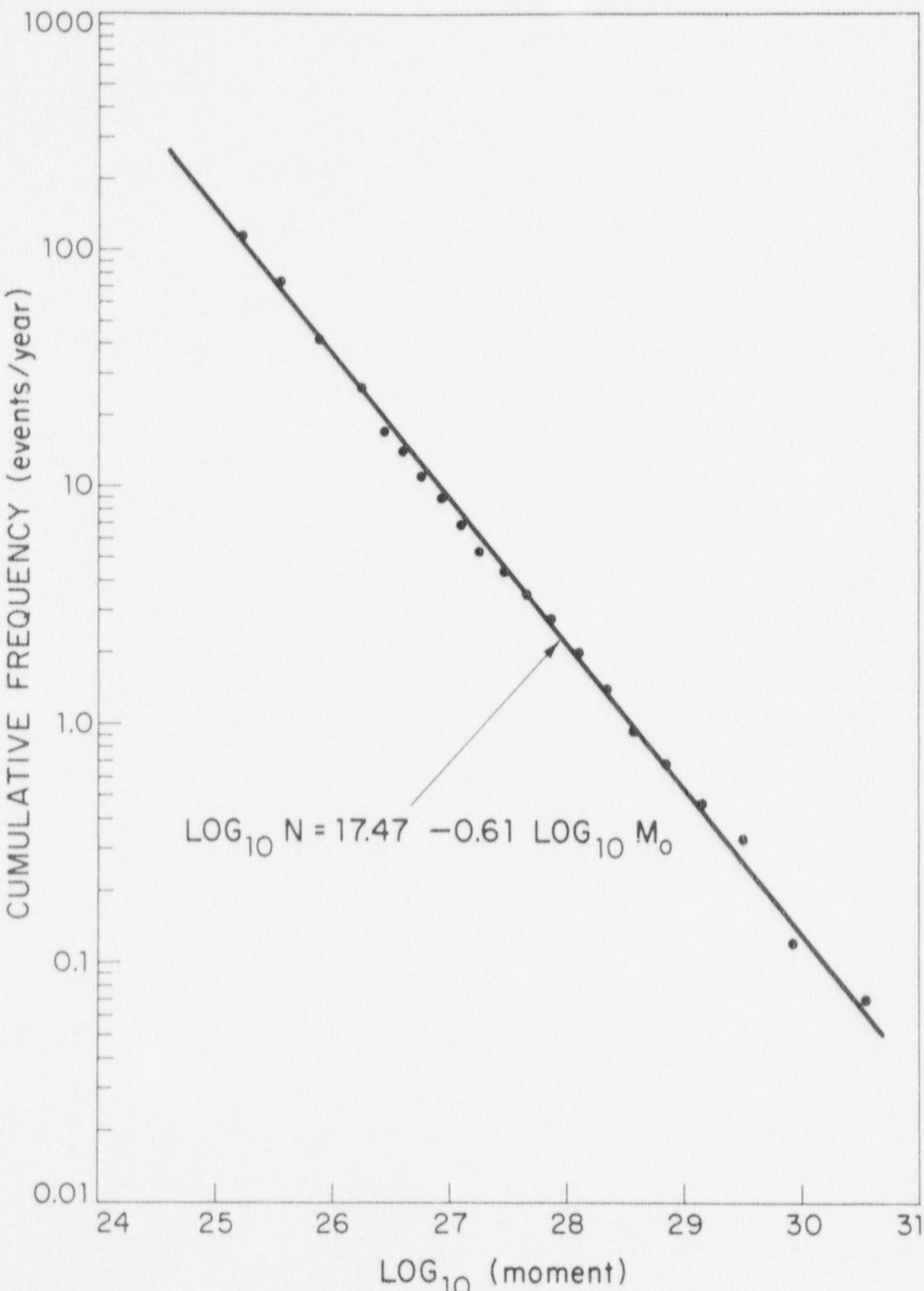

Fig. 5: Frequency - moment graph constructed from Figures 3 and 4.

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
NP-	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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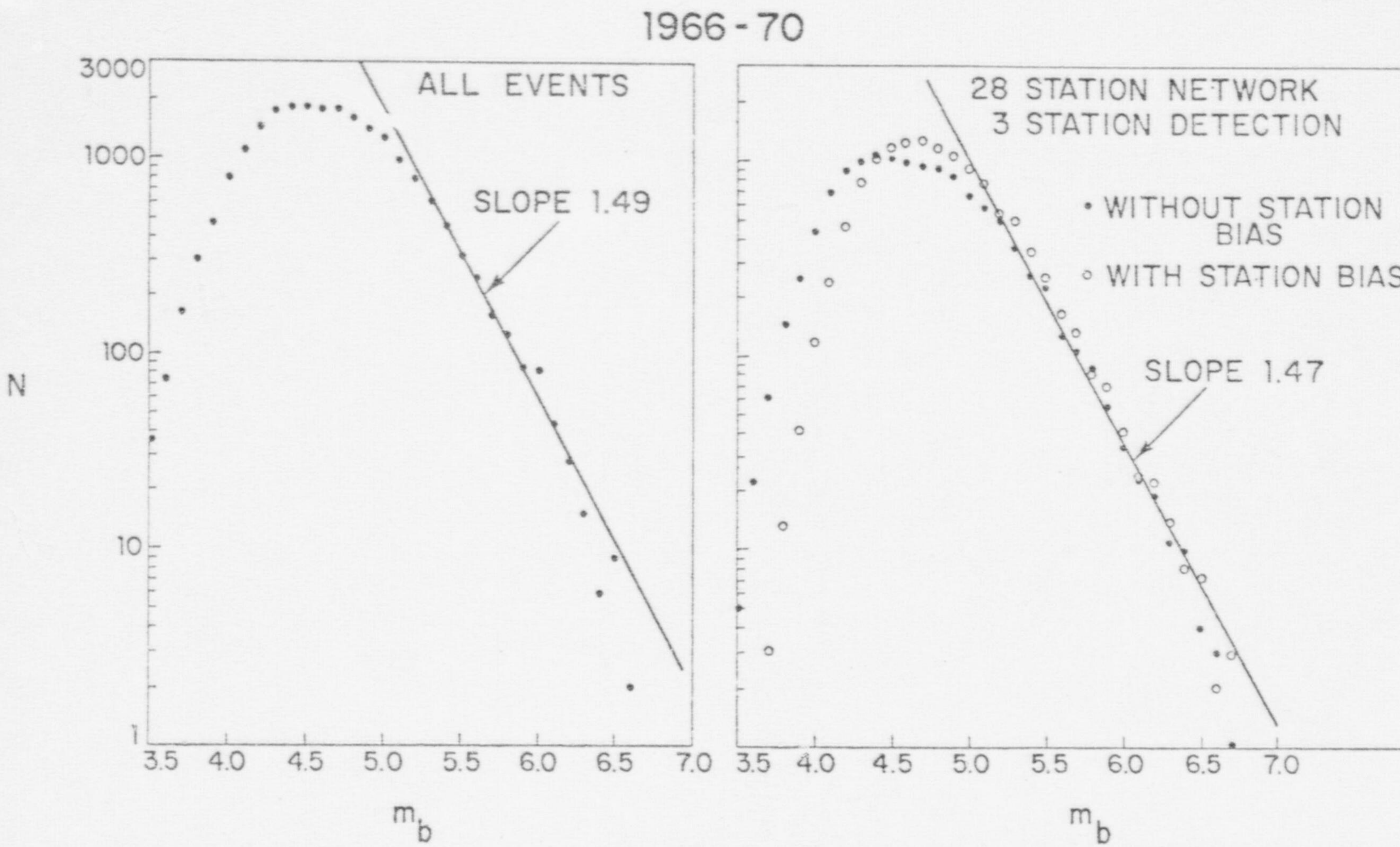

Fig. 6: Frequency magnitude data for the ISC catalog, for all listed events (left), and for a selected network of 28 stations (right). The 28 station network is listed in Table 1.

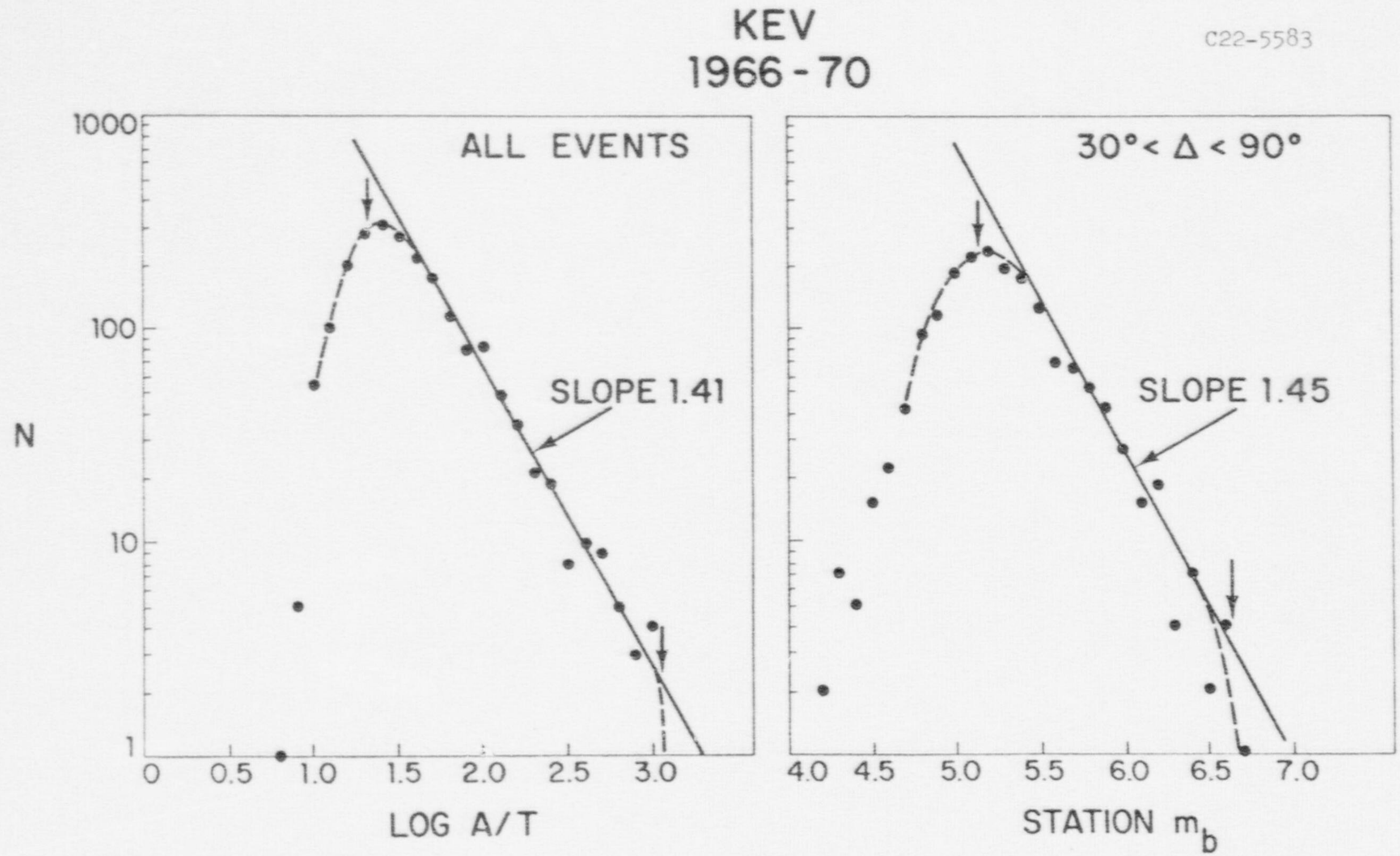

Fig. 7: Frequency magnitude data for Kevo, Finland.

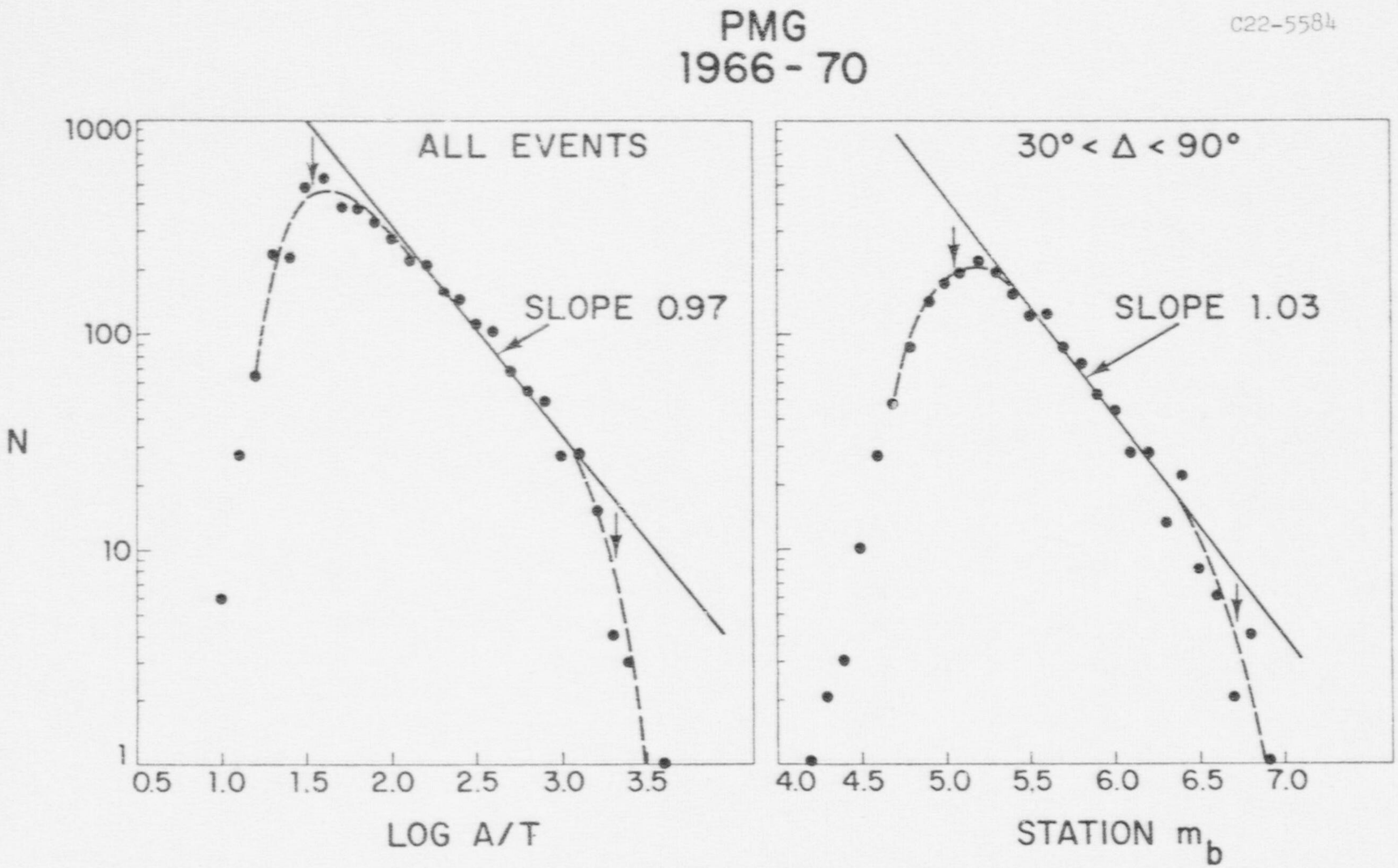

Fig. 8: Frequency magnitude data for Port Moresby, New Guinea.

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.

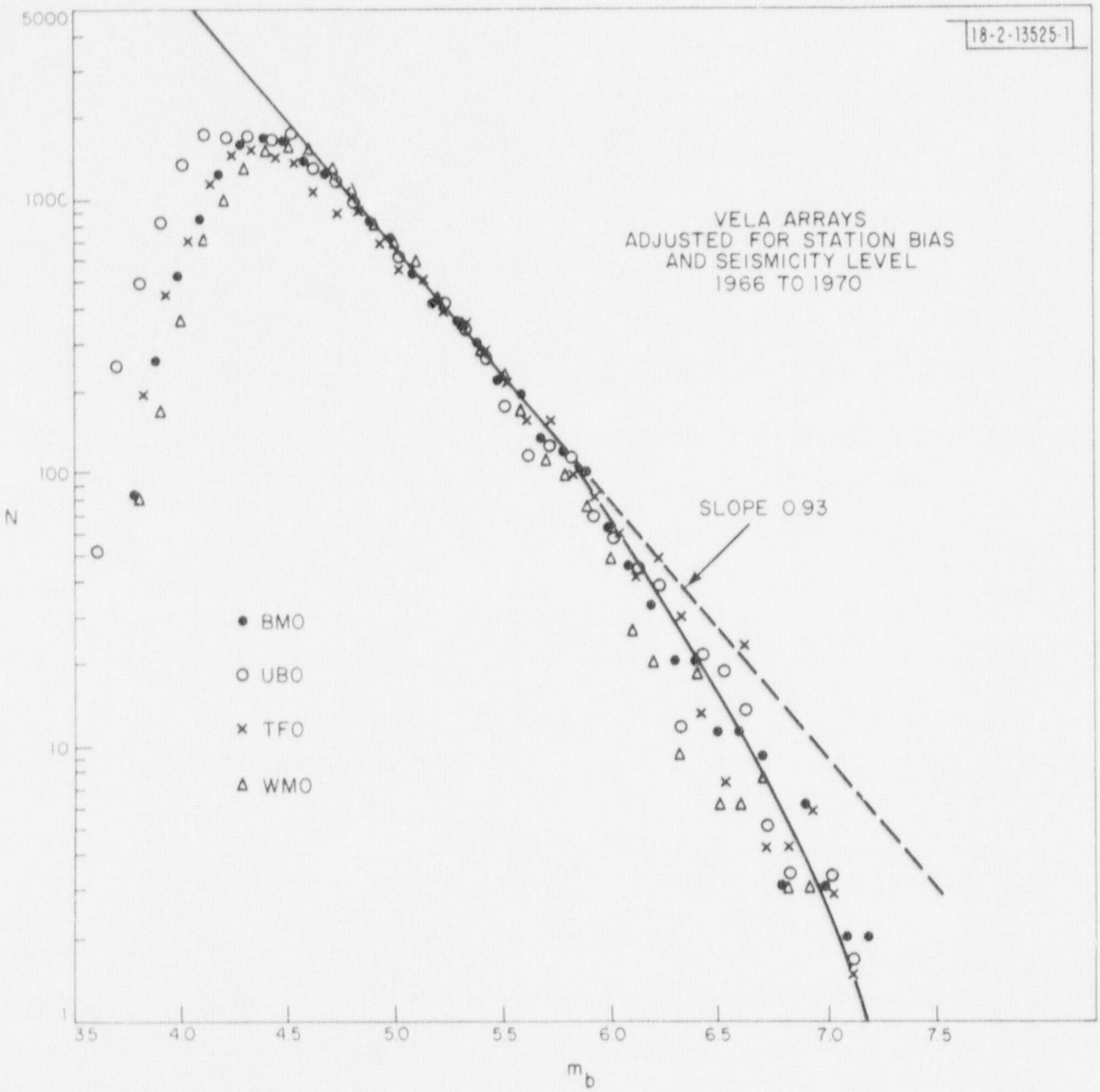

Fig. 9: Frequency-station m_b plots for four U. S. VELA observatories.

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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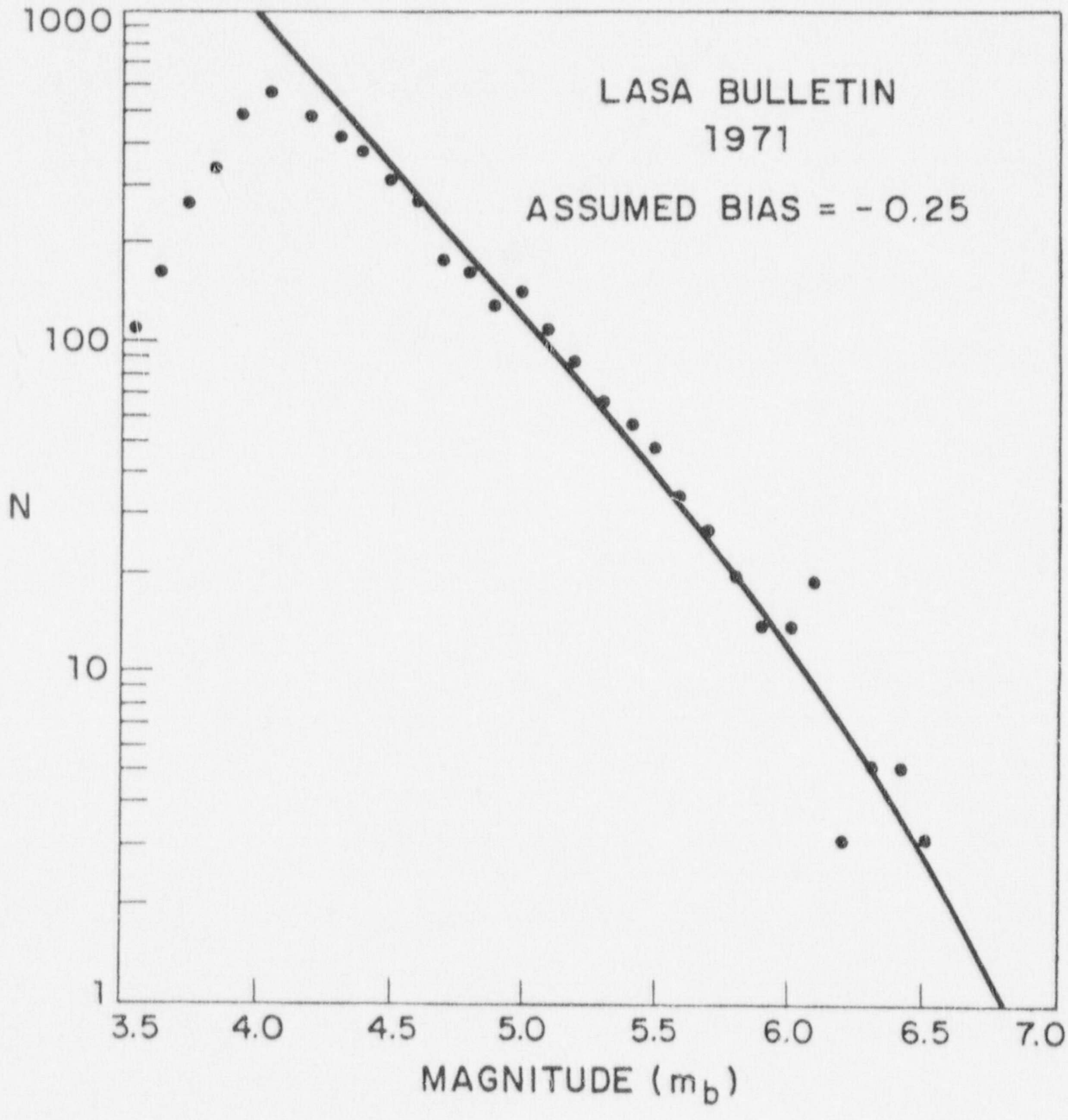

Fig. 10: Frequency-magnitude data for the Large Aperture Seismic Array (LASA) in Montana for the year 1971. The solid line is the seismicity curve shown in Figure 9.

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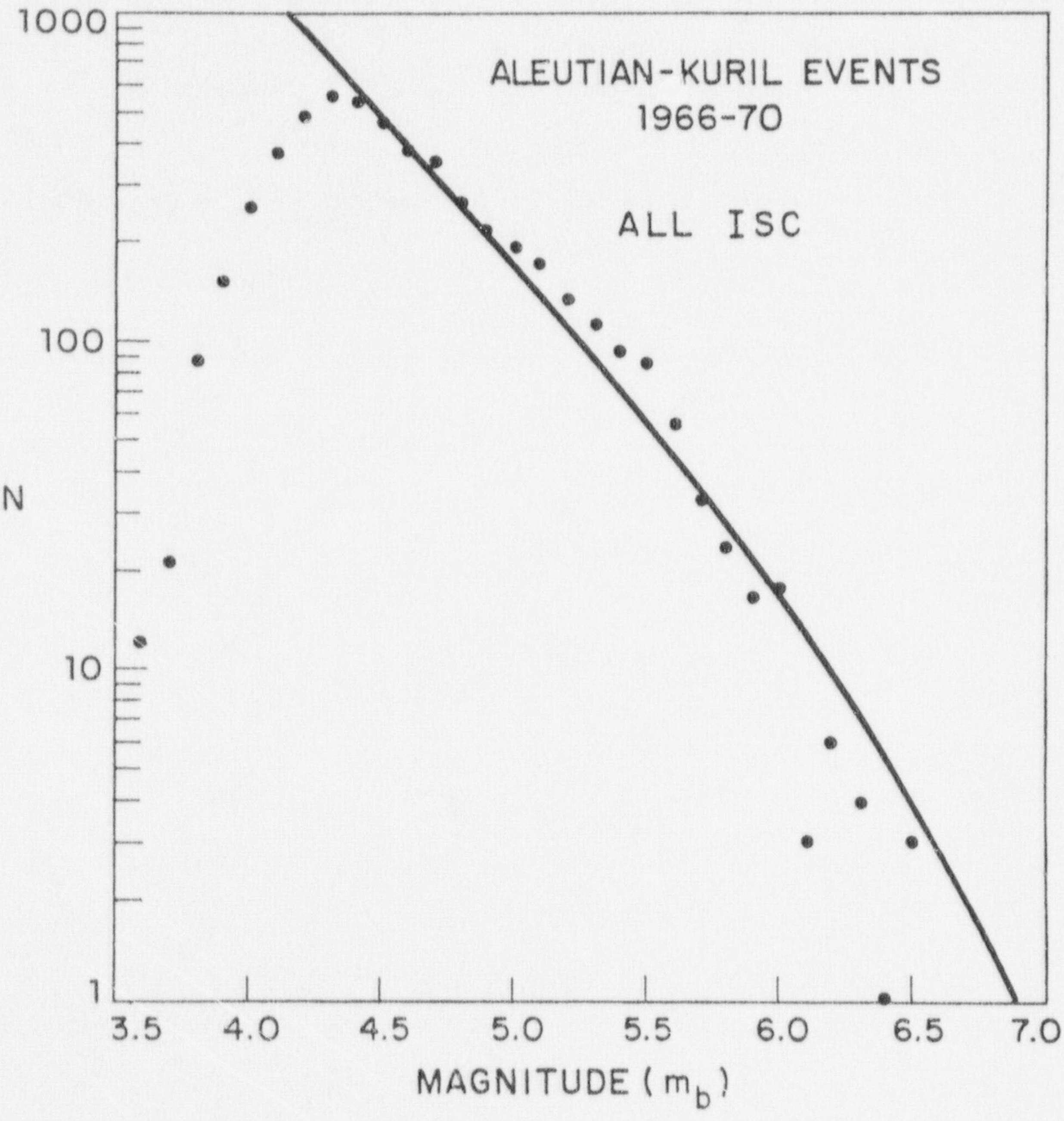

Fig. 11: Frequency-magnitude data for all events in the Aleutian-Kuril area listed in the ISC catalog, 1966-70.

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 clearl[illegible]fferent from the seismicity curve of Figure 9. In fact, it is ver[illegible]lt 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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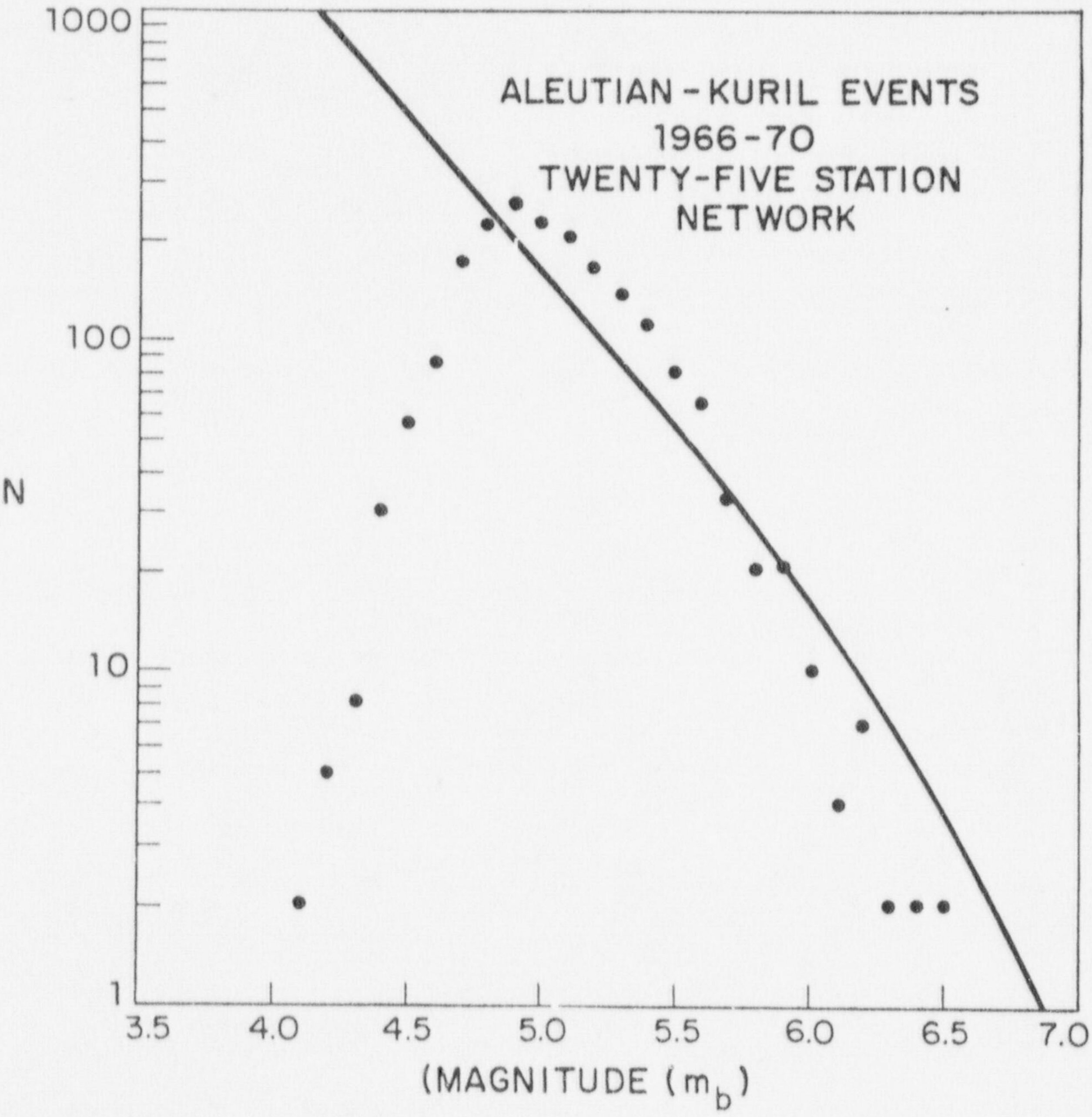

Fig. 12: Frequency-magnitude data for a 25 station network (the stations listed in Table 1, with BMO, TFO and UBO omitted).

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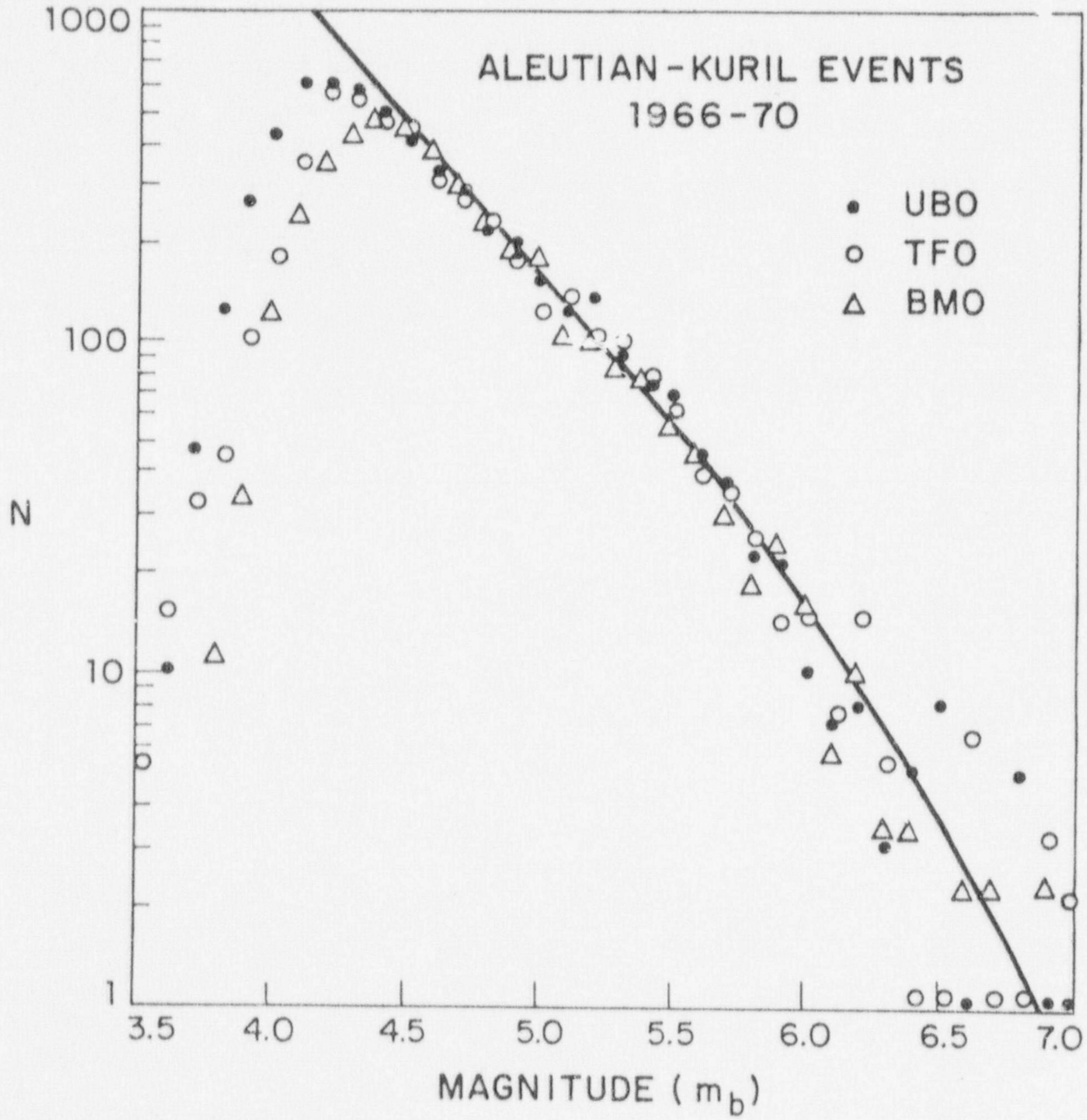

Fig. 13: Frequency-magnitude data from 3 VELA arrays for Aleutian-Kuril events. The solid curve is the same as that in Figure 9, adjusted vertically for a best fit.

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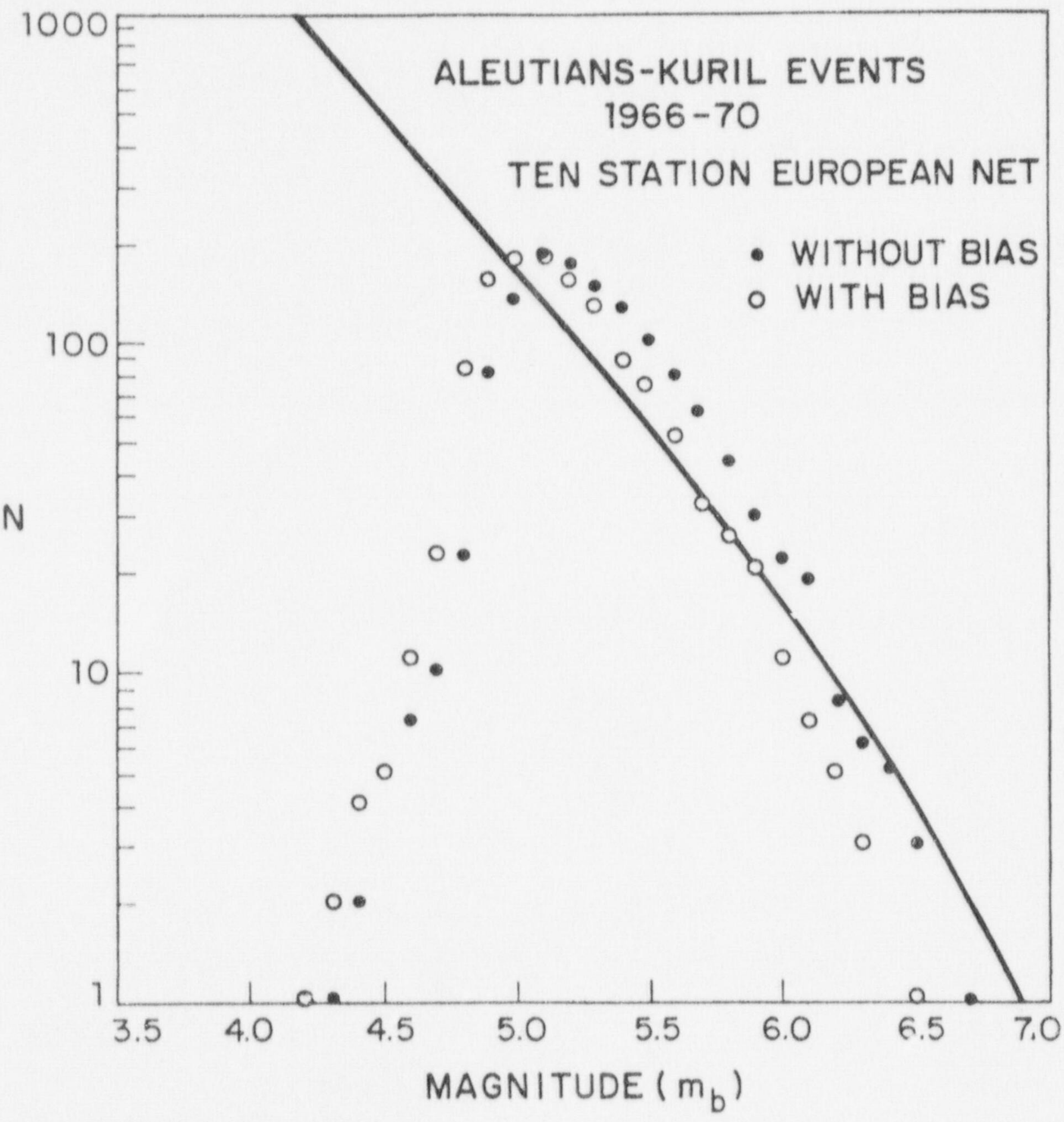

Fig. 14: Frequency-magnitude data for a 10 station European network. The stations used are listed in Table 2.

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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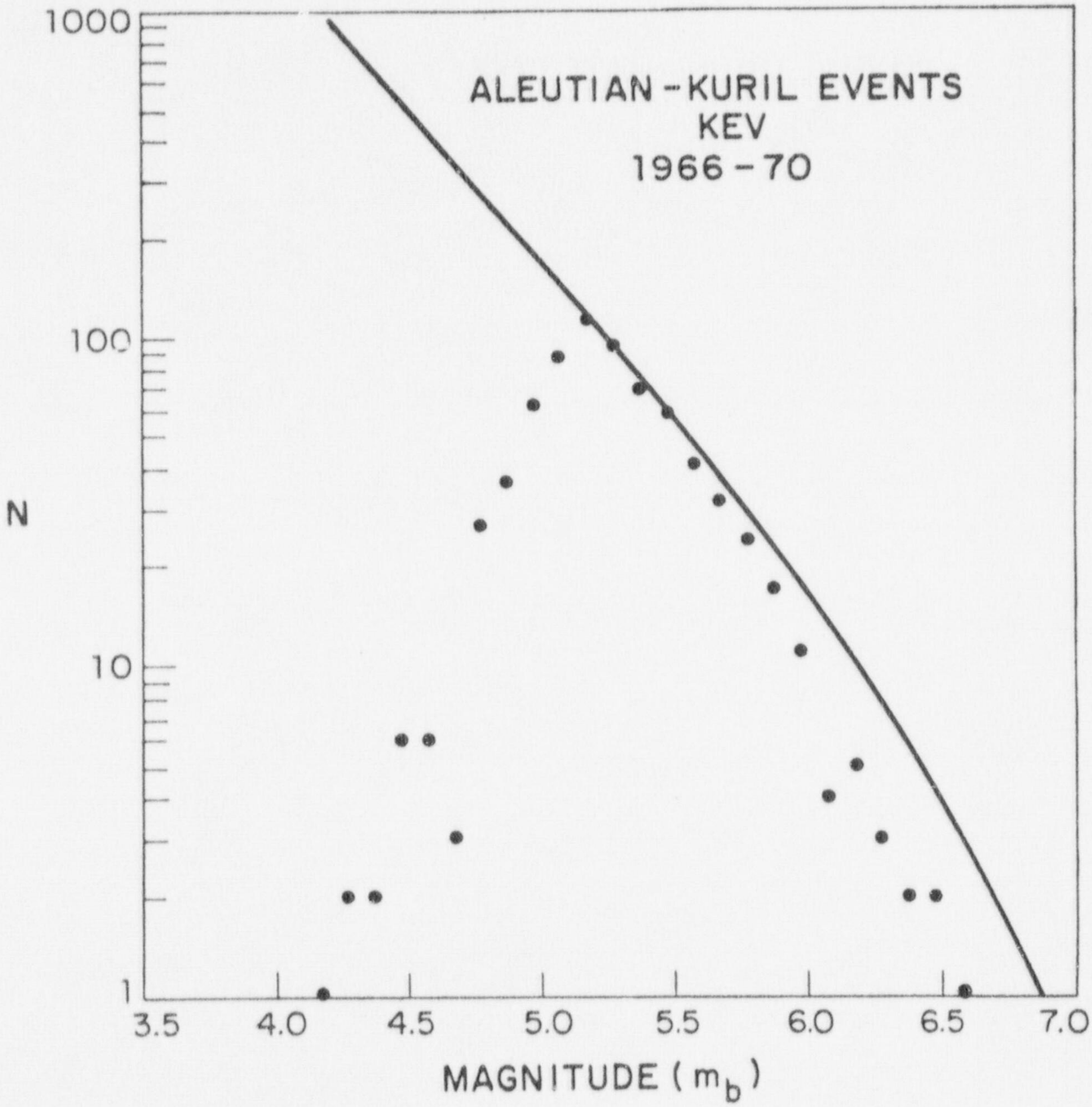


Fig. 15: Frequency-magnitude data for events in the Aleutian-Kuril area, as observed at Kevo, Finland. The solid curve is the same as those in Figures 11-14.

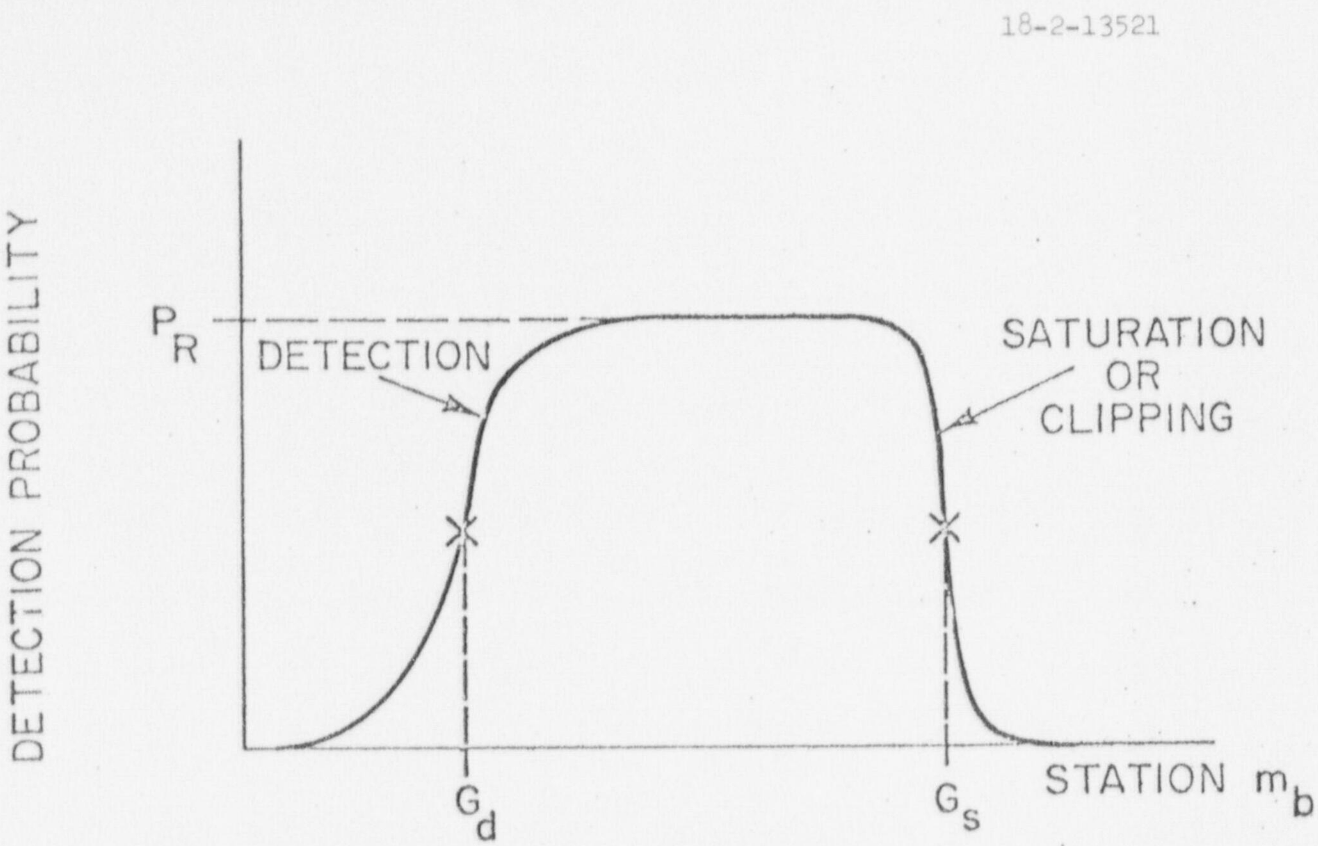
to strong biases, or the station is failing to report many large events. For the reasons discussed in the next section, the latter explanation seems most likely.

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



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.

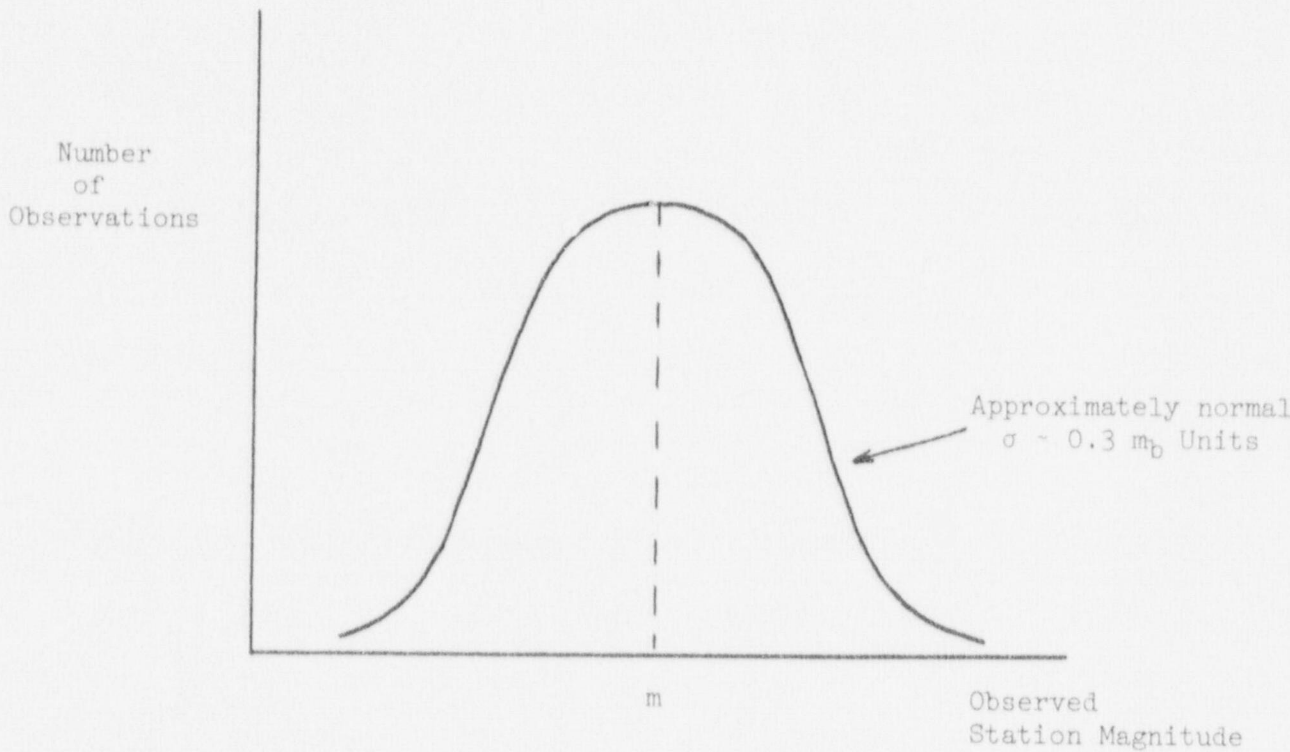


Fig. 17: The effect of scattering on observed amplitudes. An event of magnitude m will lead to a distribution of magnitudes at a network of stations. This distribution is usually approximately normal.

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.

C22-5587

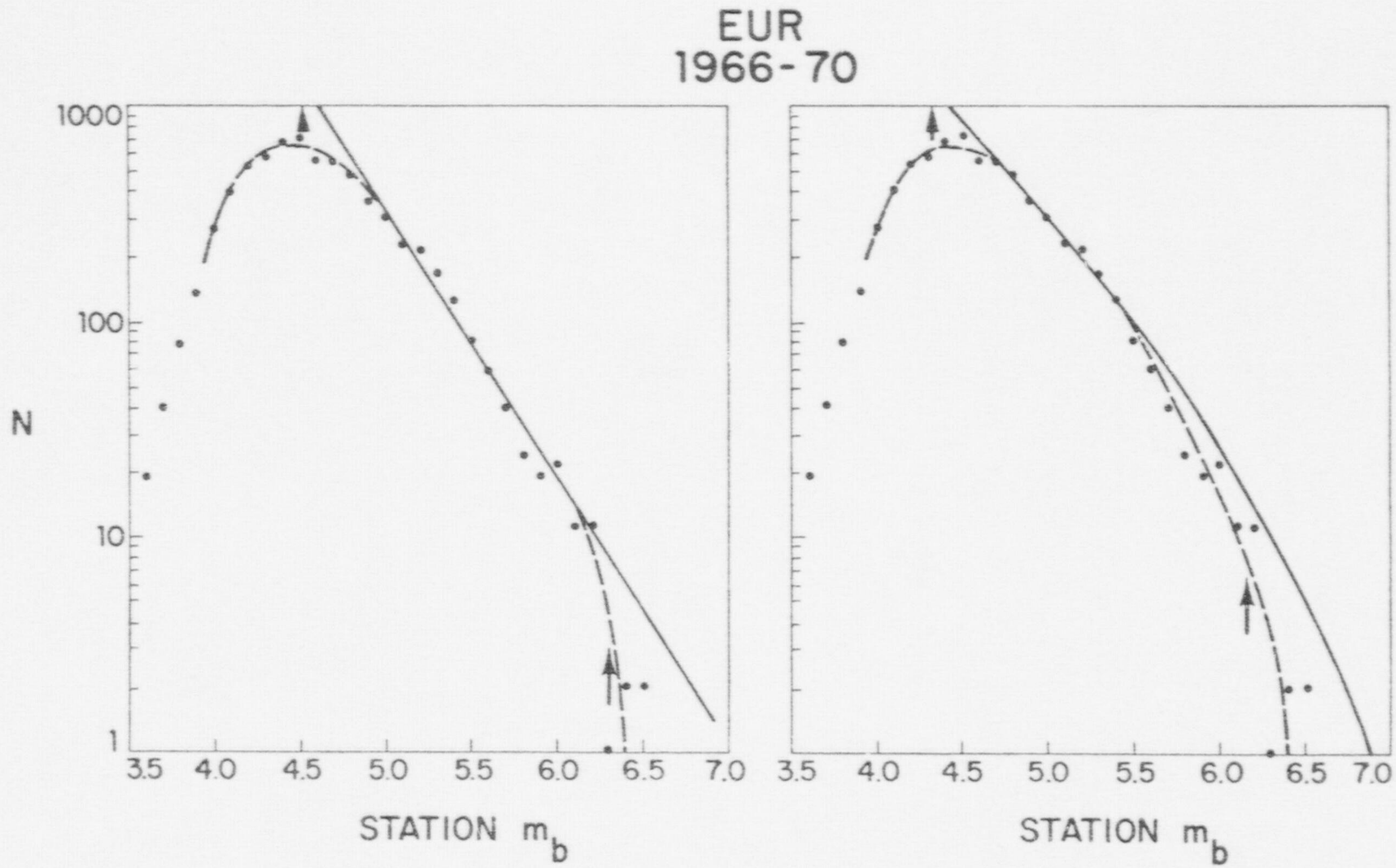


Fig. 18: Two interpretations of the reporting frequency of station EUR (Eureka, Nevada). The right hand interpretation is preferred.

C22-5461

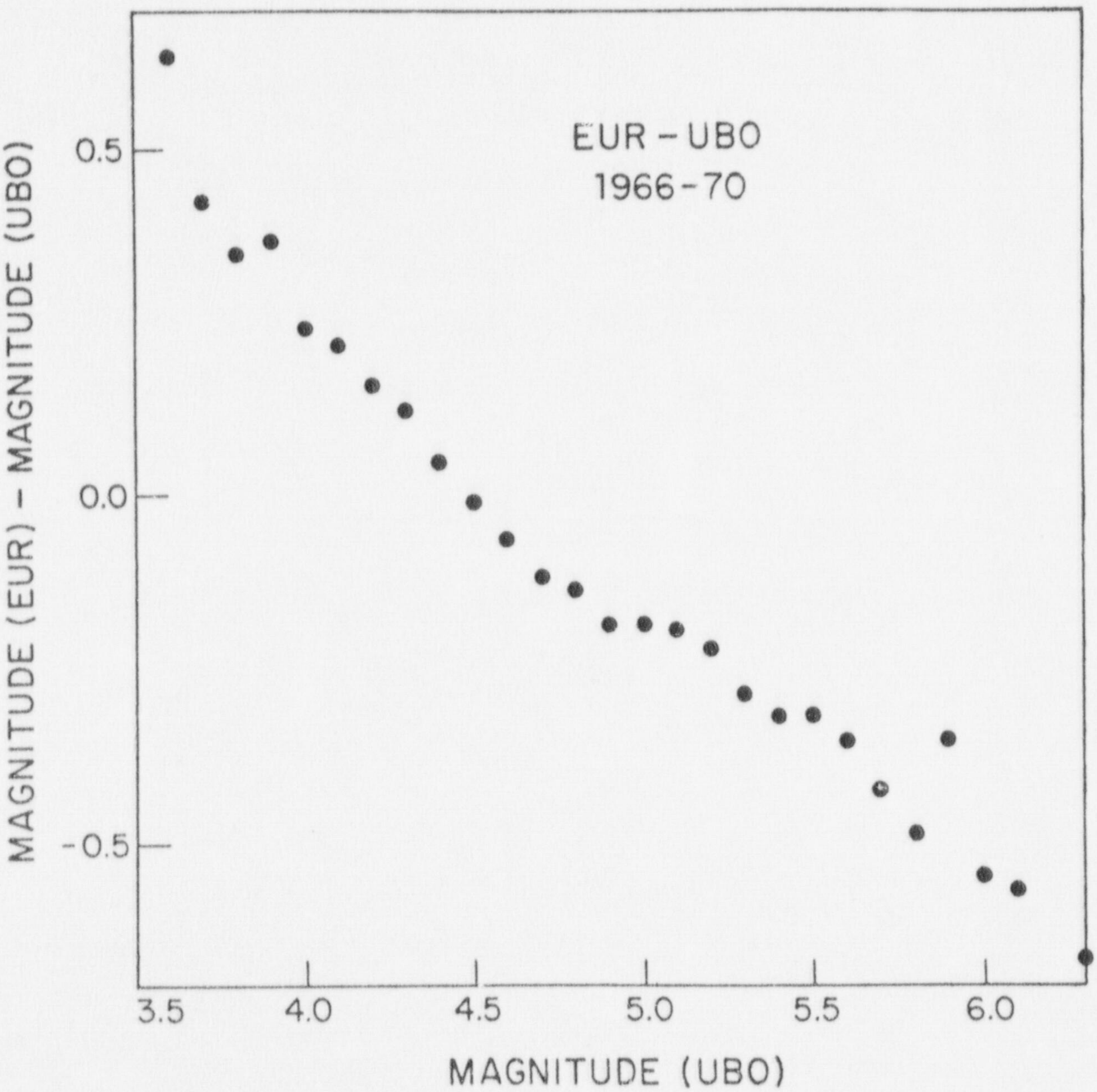


Fig. 19: Each point is the average difference between the station magnitudes of EUR and UBO for all events listed in the ISC catalog, plotted as a function of the UBO magnitude.

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.

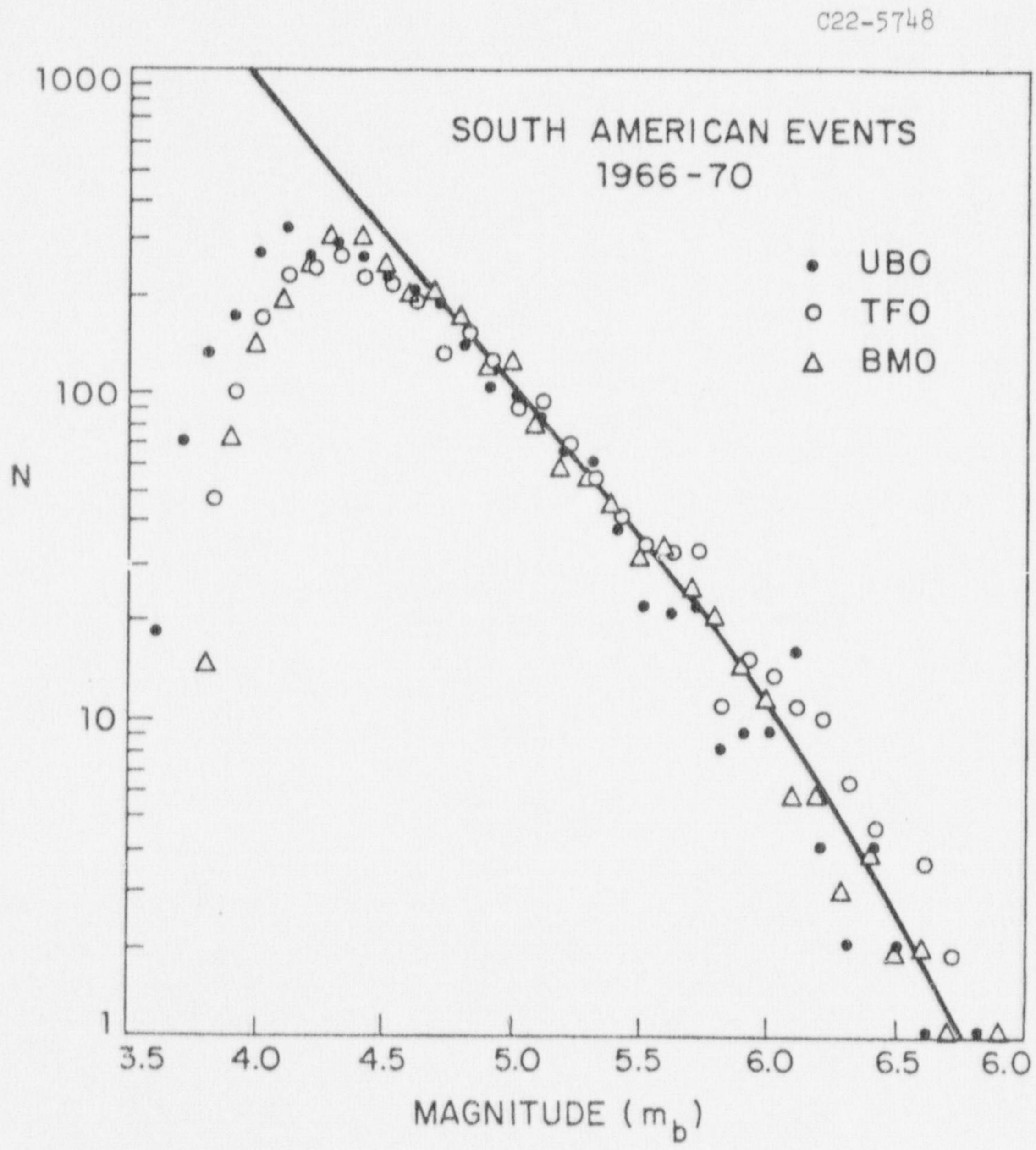


Fig. 20: Frequency-magnitude data for South American events observed at 3 VELA arrays. The solid curve is the same as that in Figure 9, adjusted vertically for a best fit.

C22-5622

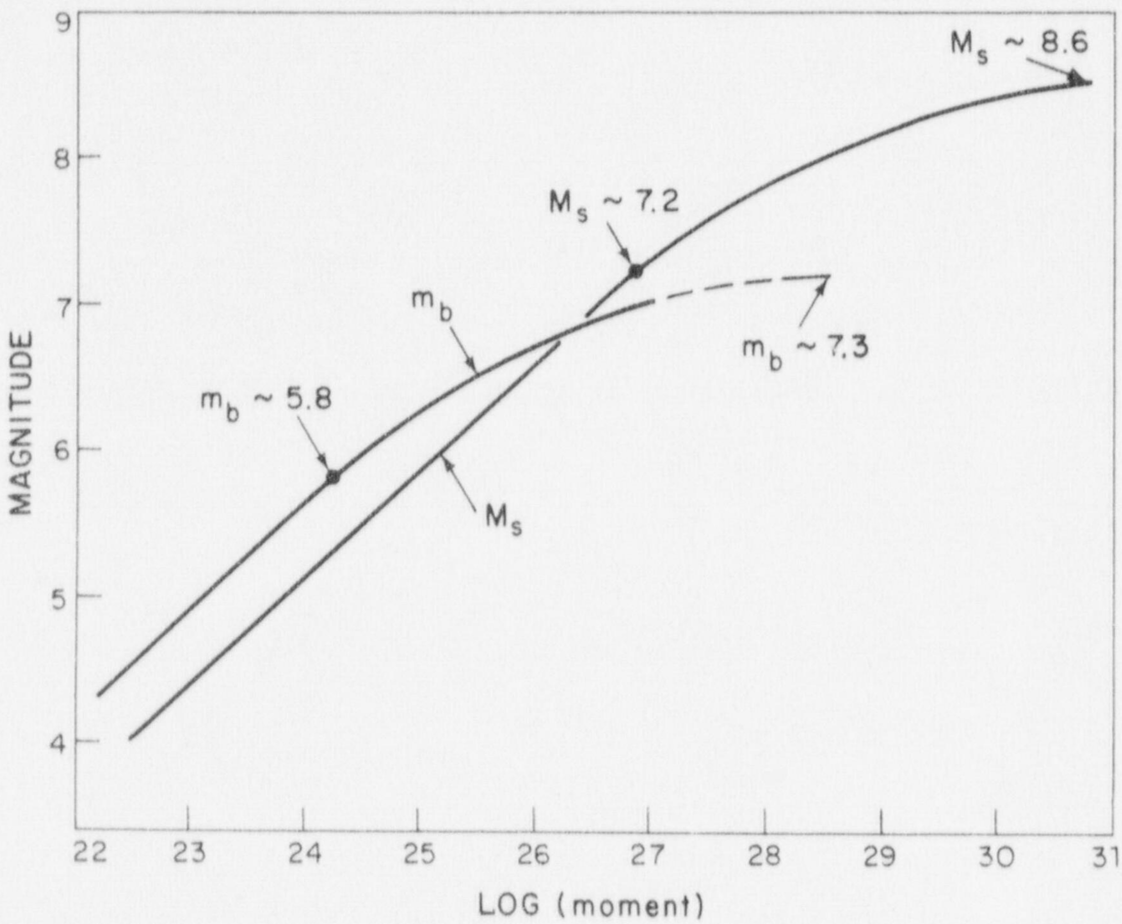


Fig. 21: An empirical m_b-moment relationship consistent with the VELA seismicity curve (Figure 9). The M_s-moment relationship from Chinnery and North (1975) is shown for comparison.

The interpretation that the curve in the VELA seismicity relationship is due entirely to saturation of the m_b scale seems reasonable. The shape of the m_b-moment curve in Figure 21 is similar to that of the M_s-moment relation, and (at least qualitatively) m_b appears to saturate at about the expected magnitude. It therefore seems unlikely that any information about upper bound magnitudes can be obtained from the existing global m_b catalogs.

2.8 Conclusions

The conclusions of this study are very negative. It does not appear that the best earthquake catalog data can shed any light on the problem of the existence or the regional variation of maximum earthquake size. This leaves only the much less comprehensive catalogs of Gutenberg and Richter (1954) and others, collected before 1960. While these older catalogs are useful for event times and locations, there are growing indications that the assigned magnitudes in these catalogs are unreliable (e.g. Chen and Molnar, 1977). At least part of this unreliability probably arises from the instrumental problems described above.

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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 he 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° 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.

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