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The seismicity data base for the Global Seismic Hazard Assessment Program

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Abstract
A successful Global Seismic Hazard Assessment Program (GSHAP) must incorporate an input seismicity data base of 1) unprecedented uniformity in magnitude, 2) large time span from historical to modern times, and 3) true global scope. Data analysis requirements will vary not only by geographic region but will differ among the *modern instrumental era* (1964-present), the *early instrumental era* (~1900-1963), and the *pre-instrumental era* (pre-1900). The basic objective is a robust determination of the frequency-magnitude relation for all the land and coastal areas of the Earth. This will include constraining the minimum magnitude of completeness at the low magnitude end and the maximum credible earthquake at the high end. To accomplish this, the data base for plate boundary zones should require instrumental data spanning only ~3 decades (1964-1993), except for $M \geq 7$. Stable continental regions, however, and maximum magnitude assessment elsewhere, will require comprehensive evaluation of historical seismicity. Hence the composition of the global seismicity data base will vary both geographically and temporally. This means that catalog completeness needs to be specified both in space and time, a requirement that is virtually nonexistent in existing seismicity compilations.

An axiom of scientific investigation holds that if research is not published it is as if it were never done. Much the same applies to earthquakes. If an earthquake is not in an existing catalog, it might as well never have happened, no matter what its actual seismic moment release. Therefore, the accurate record of seismic moment release on planet Earth very much depends on a comprehensive assessment of the existing seismicity data base — the regional and global catalogs of seismicity that have been compiled throughout the world. And a comprehensive global seismicity data base is one of the essential inputs to a comprehensive global seismic hazard assessment.

1. The existing record of Earth's seismicity

Our knowledge of Earth's seismicity may conveniently be divided into three time periods with bounds defined by advances in scientific instrumentation. The development of the first seismograph systems (as opposed to seismoscopes) occurred in the late nineteenth century; but more importantly for our purposes, the compiling of earthquake parameters based on instrumental data in permament catalogs dates to the very end of the nineteenth century (*e.g.*, Milne, 1911; Gutenberg and Richter, 1954; Abe and Noguchi, 1983a,b; Pacheco and Sykes, 1992). Hence the *pre-instrumental era* may be taken as pre-twentieth century. This historical time period has a tremendous geographical variation: it stretches back for millennia in some regions such as the Middle East and China and is essentially nonexistent in others such as Antarctica or unpopulated areas of other continents such as the Americas, Africa and Asia.

The instrumental time period, then, is the twentieth century. The *early instrumental era* extends from 1900 through 1963. It ends there because beginning in 1964 global catalogs incorporated phase and amplitude data from a significant number of the standardized and calibrated instruments of the WorldWide Standardized Seismograph Network (WWSSN) operated by the U.S. Geological Survey. The year 1964, the first full year of operation of the WWSSN and the year in which the International Seismological Summary (ISS) became the International Seismological Centre (ISC), is the beginning of the *modern instrumental era*, which continues to the present.

A possible fourth time period of use in seismic hazard assessment, which will not be covered in this report, is the *pre-historic era*, the domain of paleoseismology. Because repeat times of very large earthquakes, even at plate boundaries, often exceed the length of the historic record, paleoseismology offers the prospect of gaining recurrence information — previously inaccessible — on these critical events for seismic hazard analysis.

1.1. *The modern instrumental era, 1964-1992*

The modern instrumental era is the time span covered by the International Seismological Centre (ISC, 1964-1990) and the Preliminary Determination of Epicenters (PDE) (U.S. Geological Survey, 1964-1992). Standard magnitudes (m_b and/or M_s) are assigned by these agencies for all major events that are recorded teleseismically, *i.e.*, geocentric distance $\Delta \gtrsim 20°$, representing approximately ray paths that enter the lower mantle. Originally only the short-period m_b was reported; calculation of M_s, for example, was not begun by the USGS until 1968, and M_s was not adopted by the ISC until 1978 (and not with depth restrictions on its calculation until mid-1981).

In a significant analytical advance, routine centroid-moment tensor (CMT) determinations of earthquake faulting mechanisms have been computed for nearly all earthquakes of $m_b \gtrsim 5.5$ or $M_s \gtrsim 5.0$ for 1977 to the present (*e.g.*, Dziewonski *et al.*, 1981, Dziewonski and Woodhouse, 1983; Harvard University, 1977-1992). As part of the inversion the static seismic moment M_o is determined for the best double-couple solution; hence, moment magnitude $\boldsymbol{M}$ (Hanks and Kanamori, 1979) is available for all CMT events. This capability essentially divides the modern instrumental era into two sub-eras: 1964-1976 for standardized global locations and magnitudes and 1977-1992 for those capabilities plus CMT solutions for all larger events. One caution, however, is that the CMT catalog's geographic coverage is not uniform: smaller, southern hemisphere events are not so well reported as those in the northern hemisphere.

The seismicity of the modern instrumental era is most completely reported by the International Seismological Centre. The ISC Bulletin (ISC, 1964-1990) has a lag time of about two years. Although other organizations, such as the U.S. Geological Survey and the Moscow and Beijing seismological centres, are more timely and report locations and magnitude globally, they utilize fewer stations and report fewer events than does the ISC.

The annual number of earthquakes occurring worldwide through 1987 as reported by the ISC and its predecessor, the International Seismological Summary (ISS) (ISS, 1913-1963), is shown in fig. 1. The beginning of the modern instrumental era is marked by a dramatic, order-of-magnitude increase in reported and calculated events in 1963-64. Between 1964 and 1987 a further increase by a factor of 3-4 has occurred. The difference between number of events with ISC-calculated hypocenters and total number arises form publication of contributed solutions for smaller events from local and regional agencies. The difference between total calculated hypocenters and the total with m_b or M_s calculated arises because many events are too small or do not satisfy the ISC restriction criteria for magnitude computation.

Earthquake size estimation in the modern era depends mainly on the teleseismic m_b and M_s magnitudes provided by the ISC or occasionally by other agencies mentioned above. Figure 1 shows that only 25-50% on the total reported ISC events are assigned an m_b value. And M_s — begun only in 1978 by the ISC — is calculated for only 4-10% of all reported events. It is clear from this that even in the

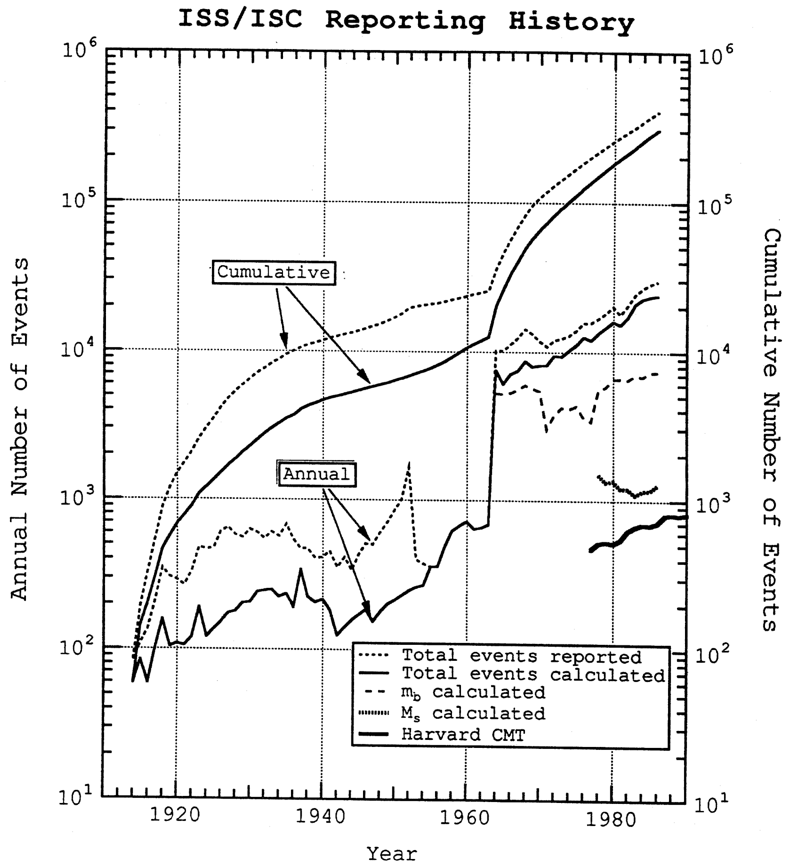

Fig. 1. The earthquake location and magnitude reporting history of the International Seismological Summary and the International Seismological Centre. The major increase in number of events occurs in 1963-64 with the advent of the ISC and the WWSSN. Data for the ISS (1913-1963) from Ambraseys and Melville (1982). Also shown is the annual reporting history of the centroid-moment tensor catalog (Harvard University, 1977-1992).

modern era a major problem for a comprehensive global seismicity data base will be uniform magnitude assignment, *i.e.*, a single size descriptor for each earthquake.

For reasons argued in a following section, the size descriptor of choice is moment magnitude $\boldsymbol{M}$. (For the purposes of this report $\boldsymbol{M} = M_w$, where $\boldsymbol{M}$ is moment magnitude as defined by Hanks and Kanamori (1979) and M_w is moment magnitude as originally defined by Kanamori (1977)). Since $\boldsymbol{M}$ is recovered directly from seismic moment M_o, the Harvard CMT catalog constitutes an important subset of events for which M_o, hence $\boldsymbol{M}$, is directly determined from seismic waveform data of the Global Digital Seismic Network. The number of events for which CMT solutions are available has been included on fig. 1; it varies from 471 to 807 events/yr and totals over 9000 events for 1977-1990. Even with adding a number of events with computed M_o from the literature, less than 5% of the modern era data will have directly determined M_o. The percentage will plummet for the early instrumental era and of course is zero for pre-instrumental times.

An additional characteristic of modern era data that requires quantitative evaluation is the minimum magnitude threshold for complete reporting, M_c. Specification of M_c is a difficult problem, even in the modern era, for the threshold will vary in space and time. Habermann (1982) has examined this problem and finds for 1963-1979 that $M_c \simeq m_b$ 4.5-5.0 globally. Ringdal (1986) reports a 4-station 90% detection threshold capability of m_b 3.9-4.5 in the northern hemisphere, increasing to m_b4.2-4.8 in the southern hemisphere. The issue is complicated by the fact that at low values m_b is biased toward higher values by the ISC (and PDE) practice of ignoring contributions from nondetecting stations. Nevertheless, for the modern instrumental era, M_c probably lies between m_b~4.5 and m_b~5.0. We can examine this question of data base completeness more closely by analyzing cumulative seismicity rates through the modern instrumental era.

Global seismicity is frequently characterized in terms of a Gutenberg-Richter frequency-magnitude relation, often called a recurrence curve. Two recent studies (Kanamori, 1988; Pacheco and Sykes, 1992; hereinafter called K88 and PS92) have derived global recurrence curves in terms of moment magnitude M_w (or $\boldsymbol{M}$). The curves are virtually identical for $M_w \gtrsim 7.5$, but for smaller events, PS92 identify a break to a smaller b-value (0.90 as opposed to 1.21-1.31), leading to a smaller number of «predicted» earthquakes for M_w<7.5. According to Pacheco and Sykes, the change in slope occurs because above M_w7.5, earthquake fault zones are limited by the thickness of the brittle seismogenic zone, while below M_w7.5 they are unbounded.

Cumulative event curves for the ISC period 1964-1986 in four magnitude bands are compared in fig. 2, with activity rates projected from both the PS92 and K88 recurrence curves. The comparison is only approximate because the expected rates are in terms of M_w while the ISC data are in terms of m_b or M_s, as indicated. Nevertheless in the magnitude 5 to 7 range, M_w is close enough to m_b/M_s that such a comparison is useful, given the caveats noted below.

It is clear from fig. 2 that the expected seismicity rates of K88 are considerably higher than observed ISC seismicity in all magnitude bands except $M_w \geq 7.0$. The PS92 curve fits the data well for the magnitude 4 and 7 bands. For the M_w 6.0-6.9 band the PS92 predicted rate is slightly higher than the observed ISC m_b 6.0-6.9 rate but lower than the M_s 6.0-6.9 rate. Since M_s is very close to M_w in this band, this behavior is expected because the PS92 curve is for shallow events only, while the ISC data contain both deep and shallow events. The effect of removing the deep (h>70 km) events from the ISC m_b cumulative curve in shown in the M_w 5.0-5.9 band in fig. 2. The actual ISC data curve is moved closer to the PS92 curve but still exceeds it, which is probably due to conversion of m_b to M_w. Published $m_b - M_w$ regressions (*e.g.*, Dziewonski and Woodhouse, 1983; Giardini, 1988) have considerable scatter but project that m_b overestimates M_w in the low m_b5 range but underestimates M_w above ~m_b5.6-5.8. The net effect is that many events in the ISC m_b5.0-5.9 band really belong in the M_w4.0-4.9 or M_w6.0-6.9 band.

This data-comparison exercise was performed in order to specify global expected levels of seismic activity and thereby establish a basis for estimating the completeness of the global seismicity data set. The comparison suggests that when the differences between m_b/M_s and $M_w/\boldsymbol{M}$ are accounted for, the PS92 global frequency magnitude relation does a good job of predicting the observed levels of seismic activity during the time period of our best and most complete data, the modern instrumental era. In contrast, the K88 relation, without the change in b-value slope, projects rates of $\boldsymbol{M}$<7 earthquakes that far exceed the observed ISC levels.

In addition to the modern instrumental era, the PS92 model should be suitable for the early instrumental and historic eras if we accept the important assumption that worldwide earthquake production is a stochastic or time-invariant process. Given that the ultimate energy source for earthquakes is the earth's internal heat and this heat flow varies spatially but not temporally, this assumption should be a sound one, although there is limited support for a possible long-period coupling of global

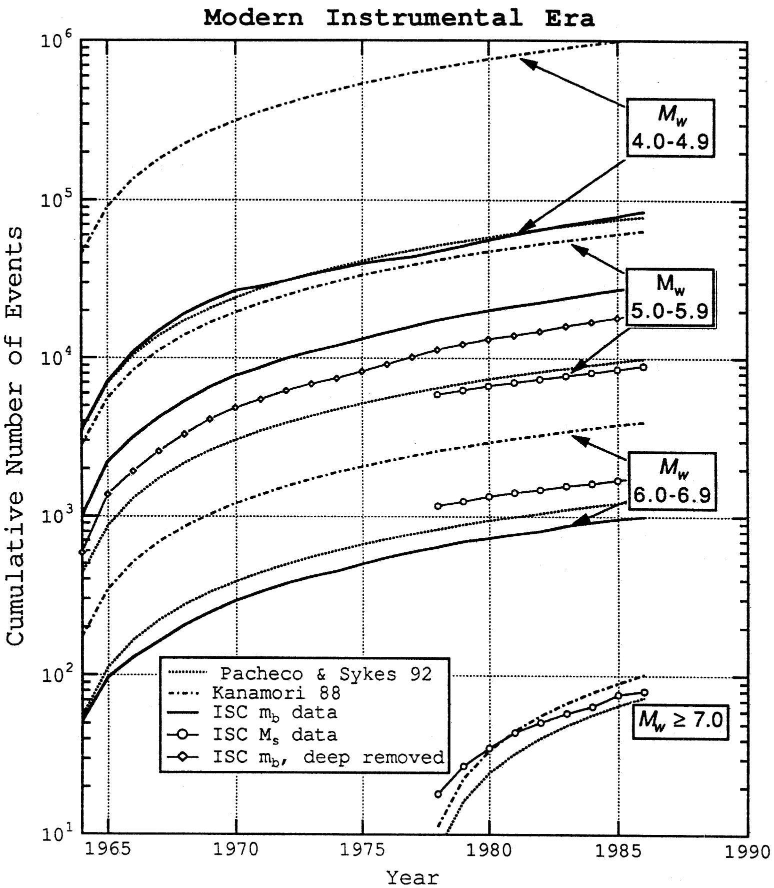

Fig. 2. A comparison of ISC seismicity data with the predicted levels derived from frequency-magnitude formulas in Kanamori (1988) and Pacheco and Sykes (1992). Note the good agreement with both curves at $M_w \geq 7.0$ (the ISC computed M_s with no depth restriction from 1978 to mid-1981, leading to unreliable numbers in those years). Good agreement is observed with the Pacheco and Sykes but not the Kanamori curves for the m_b4.0-4.9 range, the M_s5.0-5.9 range and m_b/M_s6.0-6.9 range.

seismic moment release with the amplitude of the Chandler wobble (Kanamori, 1977, 1978; Abe and Kanamori, 1979). Global time-invariance does not imply, however, regional time-invariance; there are numerous examples of significant regional variation of seismic moment release on time scales of decades to centuries. Indeed, the seismic cycle, modeled as a gradual elastic strain accumulation and abrupt release, requires such regional variation.

To summarize this section, if the PS92 global seismicity model and the approximation of a globally stochastic seismic moment release are accepted, then we can anticipate that in an average year the Earth will produce ~4000 $\boldsymbol{M} \geq 4.0$, ~500 $\boldsymbol{M} \geq 5.0$, ~63 $\boldsymbol{M} \geq 6.0$, ~8 $\boldsymbol{M} \geq 7.0$, and ~3 $\boldsymbol{M} \geq 7.5$ shallow earthquakes. These shallow rates match well with the existing ISC data base for the modern instrumental era and will therefore be used to assess completeness of reporting in the early instrumental and pre-instrumental eras. (Although a hazard is associated with intermediate and deep events (h > 70 km) in some locales (for example, Romania), this paper will focus on shallow seismicity with its much more pervasive hazard. See, however the concluding remarks concerning deep events in the GSHAP data base).

1.2. *The early-instrumental era, ~1900-1963*

For half a century (1913-1963) the International Seismological Summary provided the only global seismicity catalog that approaches comprehensiveness. Other sources, such as the *Bureau Central International Sismologique* (BCIS), covered more restricted time periods and/or geographic regions (for example, BCIS coverage extends back only to 1952 and within $\Delta \sim 50^{\circ}$ of Central Europe). The ISS evolved from its predecessor, the British Association for the Advancement of Science (BAAS), which published circulars of time and amplitude data for the global network of Milne seismographs for the time period 1899-1913. Formal ISS bulletins began in 1918 as a consequence of an IUGG resolution in 1922. Epicenters of the early time period (~1913-1917) were adopted from BAAS bulletins.

For those accustomed to working with modern data, ISS data must seem woefully inadequate. No amplitude data are published; no magnitudes are computed or provided; the only size indications are the number of recording stations and their distance and the notation of registered «long» waves at particular stations. In the early years station distribution and accurate timing were major problems and epicentral accuracy suffered accordingly. Working with a subset of ISS data for Persia, Ambraseys and Melville (1982) estimate average location error of ~30-80 km for 1940-1960 data, increasing to ~40-250 km for 1918-1940 data. Focal depths, if computed at all, are quite uncertain.

Despite these deficiencies, the ISS represents a unique and irreplaceable record of Earth's seismicity for half of the twentieth century. It is the primary data source for subsequent studies, the most important being the massive study of Gutenberg and Richter (1954) (hereafter GR54) and more recent work (*e.g.*, Abe, 1981, 1984; Pacheco and Sykes, 1992), which is concerned mainly with correction factors for the Gutenberg-Richter magnitude M_{GR} to make it equivalent to modern teleseismic M_s. The GR54 catalog was extended through 1965 and some earlier events added by Duda (1965) and Rothé (1969). In fig. 3 the cumulative ISS and GR54 data sets are compared to the expected PS92 cumulative rates developed in the previous section. All three sets of curves have been set to the same starting year for ease of comparison.

The format of fig. 3 enables us to characterize the seismicity data of the early instrumental era. Even in its first decade the ISS calculated event locations that in numbers were equivalent to the global $M_w \geq 6$ rate. Given the concentration of seismic stations in the northern hemisphere — a situation that prevails up to the present — this almost certainly represents a combination of $M_w < 6$ detections where stations were concentrated and $M_w > 6$ detections where stations were sparse. ISS coverage and reporting gradually improved until by the mid-1950s it was calculating event locations at a global $M_w \sim 5.5$ global rate and reporting events at an $M_w \sim 5.0$ global rate. Without earthquake size and depth data on these events, however, these comparisons should be considered only rough approximations.

The GR54 catalog reported shallow events at a global $M_w \geq 7$ rate until about 1910-1915. Thereafter, the number of reported shallow events gradually increased until by the 1940s, it contains events at numbers equivalent to a global $M_w \geq 6$ rate. These totals include the GR54 class «d» and «e» earthquakes, which respectively are M_{GR} 5.3-5.8 and $M_{GR} < 5.3$ events so, again, it appears that the catalog is a mix of coverage better than $M_w 6$ in many, perhaps most, northern hemisphere regions and worse than $M_w 6$ in other regions, particularly in the southern hemisphere. For $M_w \geq 7$, GR54 is not complete for 1904-1918 for class «a» and «b» events (fig. 3); thereafter its rate exceeds that of PS92 because the magnitude corrections of Abe (1981; 1984), Abe and Noguchi (1983a,b), and PS92 drop many GR54 $M_{GR} 7.0$-7.2 events to $M_w < 7.0$.

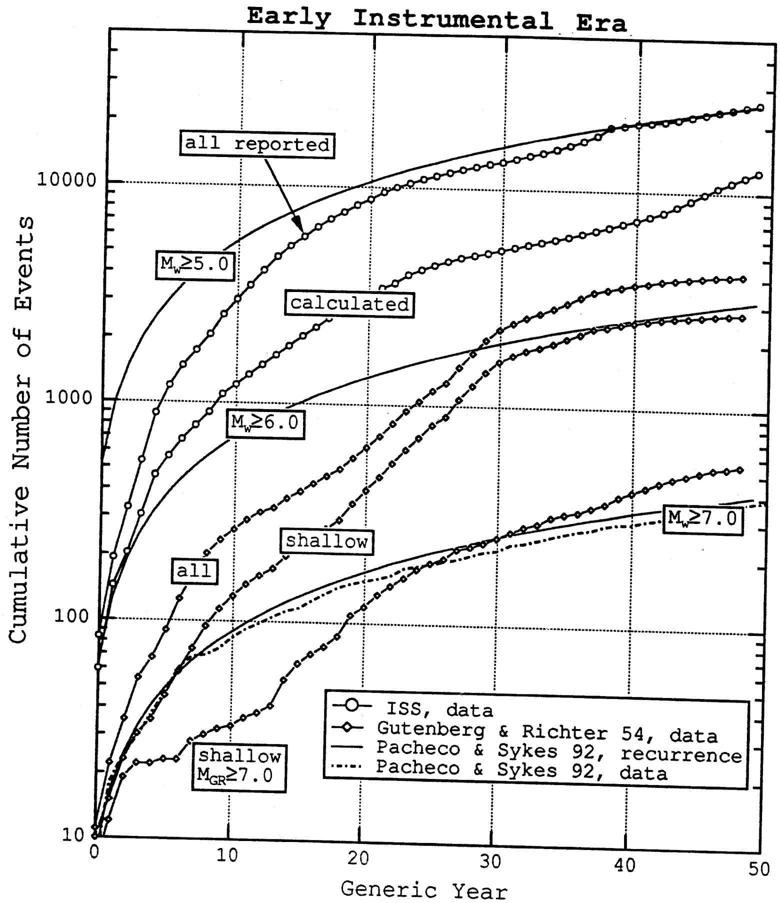

Fig. 3. The reporting history of the two principal data sources for the early instrumental era, the ISS (International Seismological Summary, 1913-1963) and Gutenberg and Richter (1954), which includes the Addenda through 1952. Also shown is the first 50 years of data of the Pacheco and Sykes (1992) (PS92) $M_w \geq 7.0$ catalog. The cumulative catalogs over a 50-year period are compared to predicted global shallow seismicity rates for various magnitude levels using the M_w frequency-magnitude relation of PS92. Generic years are used so that all curves might start at year zero; actual beginning dates are 1913 for ISS, 1904 for Gutenberg and Richter, and 1900 for PS92. The GR54 $M_{GR} \geq 7.0$ curve is for class «a» and «b» events only.

In summary, the catalogs of GR54 and PS92 give a good indication of global seismicity in the early instrumental era. Most events of $M_w \gtrsim 5.5$-6.0 will at least be listed. A careful compilation and analysis of regional catalogs will be necessary to try to achieve completeness at the M_w5.0-5.5 level for the ~1940-1963 time frame, at the M_w5.5-6.0 level for the ~1918-1940 period, and at the M_w6.0-6.5 level for the ~1900-1918 period. Relatively few events will have M_{GR} from GR54 or M_s from Abe (1981,1984), Abe and Noguchi (1983a,b), or PS92. Therefore a primary problem will be uniform and internally consistent magnitude assignment for the large number of early instrumental era events that a) have instrumental data but either no magnitudes or non-standard magnitudes assigned,

b) have no instrumental data but do have isoseismal intensity information, or c) have only an epicentral or maximum intensity reported. A strategy do deal with such events will be outlined in a following section.

1.3. *The pre-instrumental (historical) era, pre-1900*

The historical, pre-1900 era is unique in that there are no instrumental data for detecting, locating and estimating earthquake size. A valid question is whether this lack of quantitative data should preclude historical seismicity from being used at all in a global seismic hazard data base. Yet to ignore it would mean ignoring the largest known earthquakes in a majority of the Earth's continental regions. This is simply a consequence of the fact that for most civilized areas of the Earth, the historical time period is much longer than the twentieth century instrumental era. The largest earthquake that a seismic zone is considered capable of producing, designated M_{max}, is an essential, sometimes controlling, element in seismic hazard analysis (*e.g.*, Panel on Seismic Hazard Analysis, 1988; Reiter, 1990). Away from plate boundaries the historical era is often the most comprehensive and sometimes the only source of information about a region's largest earthquakes.

For data from the historical era to be usefully incorporated into the global seismicity data base, an estimate of event size and the uncertainty associated with that estimate is essential. Except in the rare instances of surface faulting, tsunami generation, or preserved liquefaction effects, the only source of size information is from intensity reports, that is, the effect of the earthquake's ground motion on people, the land, or structures. In addition, intensity data may be all that are available for a large subset of events in the early instrumental era.

There are a number of earthquake intensity scales in common use today. This discussion will be in terms of the Modified Mercalli Intensity (MMI) scale, but could equally well apply to the Rossi-Forel (RF), the Medvedev-Sponheuer-Kárník (MSK), the Mercalli-Cancani-Seiberg (MCS), or the Japanese Intensity scales, or any other for which the conversion to one of these basic scales is known.

No global catalog of earthquakes of the historical era exists in which the known intensity data of each event is tabulated or compiled. The existence and quality of national or regional historical seismicity catalogs varies greatly country by country and region by region. A listing of some example catalogs, selected for their comprehensive listing of intensity data, is contained in table I. The table is not a complete listing by any means, but is intended to convey the type of data that is available for analysis in the historical era. Those data sources that are listed provide sufficient information so that if a relation between isoseismal area or radius and a uniform size measure such as $\boldsymbol{M}$ or M_w exists for the region, then the earthquake can be categorized with respect to size, albeit with perhaps a large uncertainty.

It is evident that the historical era presents difficult problems for a uniform global seismicity data base. Perhaps the two most difficult are 1) specification of completeness levels, specific to time period and region, and 2) a uniform size estimate with uncertainty bounds. One of the more important tasks will be to determine just what data exist. There are many seismically active countries — perhaps a majority — for which no comprehensive historical seismicity catalog exists. In the United States, for example, a tremendous amount of work has been done evaluating historical earthquakes, but the information is scattered in literature papers, project reports, regional catalogs, and so on. Also, an unfortunate dichotomy has arisen between analysis of Western and Eastern U.S. earthquakes so that the data and analysis techniques are unnecessarily separated (*e.g.*, Hanks and Johnston, 1992). A global project that unifies within one data set with a uniform format the information available for important historical worldwide seismicity will be a highly significant contribution for seismic hazard and research efforts.

2. A uniform global magnitude scale

A major portion of the effort in compiling the global seismicity data base will be the assignment of a uniform size indicator or mag-

Table I. Selected catalogs of historical earthquake data.

Country/Region	Reference	Remarks
Australia	«Atlas of Isoseismal Maps of Australian Earthquakes, parts 1 and 2» (Everingham *et al.*, 1982; Rynn *et al*, 1987)	Standardized MMI maps; data points shown
Russia	«New Catalog of Strong Earthquakes in the USSR from Ancient Times through 1977» (Kondorskaya and Shebalin, 1982)	Isoseismal radii given; no maps or data points
West Africa	«Seismicity of West Africa» (Ambraseys and Adams, 1986)	Most comprehensive source for the region; numerous MMI maps
Italy	«Atlas of Isoseismal Maps of Italian Earthquakes» (Postpischl (Editor), 1985)	From 990-1980 A.D.; $I_o \geq$IX; comprehensive maps; MCS scale
China	«Catalog of Chinese Earthquakes» (Gu *et al.*, 1983a,b)	Comprehensive source; extends over 2000 years; ~10% of events have isoseismal maps
Europe	«Seismicity of the European Area, part 1 and 2» (Kárník, 1969, 1971)	Period of coverage, 1800-1955; isoseismal radii, no maps; MSK scale
NW Europe	«Magnitude Assessment of Northwestern European Earthquakes» (Ambraseys, 1985a)	Many MSK isoseismal maps; primary data source for NW Europe; companion to Ambraseys (1985b)
East. Europe	«Seismicity of Central Europe» (Procházková, 1990)	MSK isoseismal maps; data points and descriptions
Japan	«Descriptive Catalogue of Disastrous Earthquakes in Japan» (Usami, 1975)	Japanese intensity scale; thorough coverage of larger events
Britain	«An Analysis of British Earthquakes» (Principia Mechanica Ltd, 1982)	Many MMI maps; lengthy descriptive material
Spain Portugal	«General Catalog of Isoseismals of the Iberian Peninsula» (Mezcua, 1982)	Many MSK maps and descriptions; data points often shown
Brazil	«Seismic Activity in Brazil in the Period 1560-1980» (Berrocal *et al.*, 1983)	A source for historical era; contains a few MMI maps
Iran and region	«A History of Persian Earthquakes» (Ambraseys and Melville, 1982)	Definitive source for Persia; covers historical-to-modern eras

nitude to each earthquake. This will not be straightforward because a broad range of tectonic environments, reporting practices and instrumentation contribute to a melange of different magnitude estimation practices. Earthquake size can be variously reported as teleseismic m_b and M_s, regional M_L and m_{bLg}, M_{coda}, $M_{Richter}$, M_w, M_o, I_o, felt area, area within a given intensity contour, a magnitude inferred from various intensity measures or some combination of the above. Or, as with ISS data, no magnitude or intensity data at all are given.

The most logical uniform size estimator for use in the global seismicity data base is the seismic moment magnitude scale M_w, originally developed by Kanamori (1977, 1978) and formally defined and designated by bold $\boldsymbol{M}$ by Hanks and Kanamori (1979). Application entails estimating the seismic moment M_o of each event, then applying the formula $\boldsymbol{M} = 2/3 log(M_o)$-10.7, where M_o is in dyne-cm. $\boldsymbol{M}$ is the magnitude scale of choice because of a number of inherent advantages that apply to the use of seismic moment as a measure of earthquake source strength. The greatest of these is that the static moment is simply given by $M_o = \mu UA$, where μ is the rigidity modulus of the source zone and U is the average slip on a planar fault of area A. Thus M_o, which can be obtained from remotely recorded seismic waves via spectral inversion or waveform matching methods, relates directly to physical parameters: fault dimensions, slip and rigidity. Such is not the case with amplitude-based magnitudes.

Another significant advantage of moment magnitude is that a given earthquake can have only one M_o, hence $\boldsymbol{M}$. If the M_o of one earthquake is larger than another either it involved a greater fault area, greater slip, occurred in more rigid material, or some combination of these parameters. The same earthquake may have a variety of amplitude-based magnitudes, however, depending on the frequency band of measurement. For large (or «slow») earthquakes for which the corner period is longer than the period band for magnitude measurement, saturation will occur, a problem that does not arise with a moment magnitude scale.

An M_o-based scale entails other advantages. Cumulative moment sums are indicative of either average slip rates on single faults or average seismic strain rates over a region (volume of crust). These quantities are valuable for assessing the seismic potential of a region and are easily obtained if the seismicity is in terms of $\boldsymbol{M}$. Finally the $\boldsymbol{M}$-scale is convenient to use as it agrees well with m_b for $m_b \lesssim 5.5$, with M_L for $3 \lesssim M_L \lesssim 7$, with M_s for $5 \lesssim M_s \lesssim 7.5$, and is synonymous with M_w at larger magnitudes (Hanks and Kanamori, 1979; Hanks and Boore, 1984).

Even if we accept that M_o or $\boldsymbol{M}$ is the best single parameter measure of earthquake size, this does not mean it is a complete measure. In the ideal case one would like to combine the moment tensor, which yields M_o, the faulting mechanism and degree of adherence to double-couple shear rupture, with the source time function $\Delta\tau$ (or its approximate equivalent in the frequency domain, the corner frequency). We could then determine if a given event was brief but powerful (relatively low M_o, but short $\Delta\tau$, large slip and high stress drop) or prolonged but energetic (large M_o, long $\Delta\tau$, low stress drop, and small slip). When M_o or $\boldsymbol{M}$ is estimated indirectly from amplitude magnitudes or intensity data, it is possible to obtain too high a value for M_o for above average stress drop earthquakes and too low a value for those with below average stress drops. Such detailed source parameter information will be available for just a tiny fraction of the global data base, however. In lieu of the complete description, the static seismic moment M_o serves as the best-available single-parameter descriptor of earthquake size.

2.1. *Regression analysis*

We have seen from fig. 1 that less than 5% of modern instrumental era data has a directly determined seismic moment. For the early instrumental era the percentage is surely $\ll 1\%$ and for historical times it is zero. Therefore a uniform global data base requires some method by which amplitude magnitudes and intensity data can be converted to $\boldsymbol{M}$. One simple, straightforward, and widely used technique is regression analysis. We will outline an approach that depends on earthquakes that have M_o determined by inversion, waveform matching or spectral levels and also have amplitude magnitude and/or isoseismal data. These important events allow one to obtain a relation between $log(M_o)$ and other size measures — M_s, m_b, M_L, MMI areas, or other such measures. This relation, once established, can then be used to estimate $log(M_o)$ when only the secondary parameters are available.

An important question in regression analysis concerns the direction in which the regression should be performed. When both $log(M_o)$ and magnitude or isoseismal area are known for an event, should $log(M_o)$ be considered the independent variable and other size measures derived from it or vice versa? One could argue that because M_o is the fundamental measure of earthquake size, other size measures should be regressed on it. This would be true if *correlation* of two parameters were all that were sought. Our purpose here, however, is *prediction*. In this case $log(M_o)$ is the unknown (dependent) quantity and regression is performed on the known (independent) earthquake size parameters. Bonilla *et al.* (1984) have a particularly clear discussion of this point.

2.2. *Regression on instrumental data*

An example of an instrumental data regression analysis is shown in fig. 4, adapted from Johnston (1993). The data set consists of 48 earthquakes from stable continental regions (SCRs), which as opposed to tectonically active intraplate zones, have not experienced significant tectonic deformation since the Mesozoic. In performing regressions it is an important general rule to strive for a homogeneous data set, which in this case means selecting earthquakes that occur in crust with a similar geologic and tectonic history. It does not necessarily mean that the quakes must all occur in the same region, although regionally specific regressions are a common and useful procedure (*e.g.*, Ekström and Dziewonski, 1988).

The data points of fig. 4 are earthquakes that have a teleseismic M_s determination and have M_o determined independently, *e.g.*, by centroid-moment tensor inversion (Harvard University, 1977-1992) or special literature studies. The results of the second-order regression are shown on the figure and compared to two linear regressions that used much larger global, mainly plate boundary, data sets. The SCR $log(M_o)$–M_s relation does not differ significantly from the global relations for $4.5 \leq M_s \leq 7.5$. This will not always be the case, however; Ekström and England (1989) demonstrate that use of a global rather than regionally specific M_o-M_s relation in the Aegean Sea region results in an overestimate of the seismic moment release rate by a factor of three.

The value of regression analysis resides in its predictive capabilities. Once the $log(M_o)$–M_s relation is established, the seismic moment of events without M_o but with M_s can be estimated from the relation. The uncertainty of the estimate will depend on the statistical robustness of the regression. M_s is probably the most important instrumental data regression, but others are, of course, just as feasible. Regressions between $log(M_o)$ and m_b typically exhibit considerably more scatter than with M_s (*e.g.*, Dziewonski and Woodhouse, 1983), reflecting the problem of estimating M_o from short-period data. Regressions on other magnitudes, such as M_L, m_{bLg}, or non-standard m_b and M_s, are certainly possible but often are limited by lack of an adequate number of events with M_o determinations.

2.3. *Regression on intensity data*

This section will highlight the value of continuing to systematically collect and analyze intensity data even in this modern instrumental era. This is because modern earthquakes with both an instrumentally determined M_o and well determined isoseismal contours are especially valuable for regressions between $log(M_o)$ and isoseismal areas (or radii), which may then be used to estimate the moment magnitude of historical earthquakes. Even though intensity data are subjective and may be subject to large uncertainties that arise from unknown variations in focal depths, regional attenuation, site response, type of building construction, and bias in reporting practices, useful magnitude estimates can be obtained within $\sim\pm0.5$ magnitude units or better, depending on the quality and abundance of the macroseismic data (*e.g.*, Ambraseys, 1985b; Ambraseys and Adams, 1986). Johnston (1993) estimated that moment magnitude could be recovered to $\mathbf{M}\pm0.35$ units if

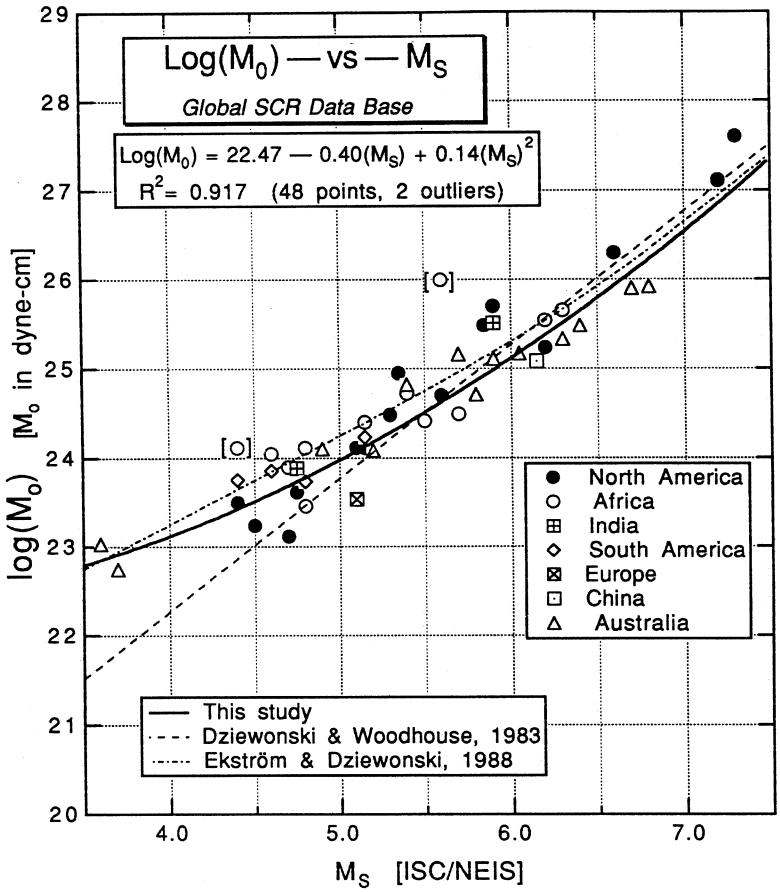

Fig. 4. An example of a regression analysis of $log(M_o)$ on instrumental data, in this case teleseismic M_s. The data are drawn only from stable continental plate interiors (SCRs), a restrictive subset of intraplate seismicity. A quadratic curve fit is shown, which yielded better residuals than a linear regression. Shown for comparison are two global $log(M_o)$–M_s regressions, illustrating that in this case for $m_b \gtrsim 5$ there is no significant difference between global and region-specific curves. Note: some M_{GR} magnitudes corrected to M_s are included. Adapted from Johnston (1993).

isoseismal areas for several different intensity levels for the same event were available for averaging.

An example of a regression of $log(M_o)$ on MMI felt area is given in fig. 5, again for a set of stable continental earthquakes from Johnston (1993). A quadratic curve fit to the data yields considerably lower residuals than a linear fit. Note that although data from a number of continents are combined, no systematic offsets of the data of any continent are observed. This suggests that a similar tectonic crustal environment (in which anelastic attenuation in similar) is more important than geographical proximity. The largest event in the data set, the 1934 ***M***8.1 Bihar, India earthquake, was a Himalayan front rather than a SCR event, but its isoseismal pattern was virtually entirely within the Indian craton.

Surprisingly, the lower intensities (MMI I-VII) are more valuable for moment magnitude estimation than the higher intensities for two

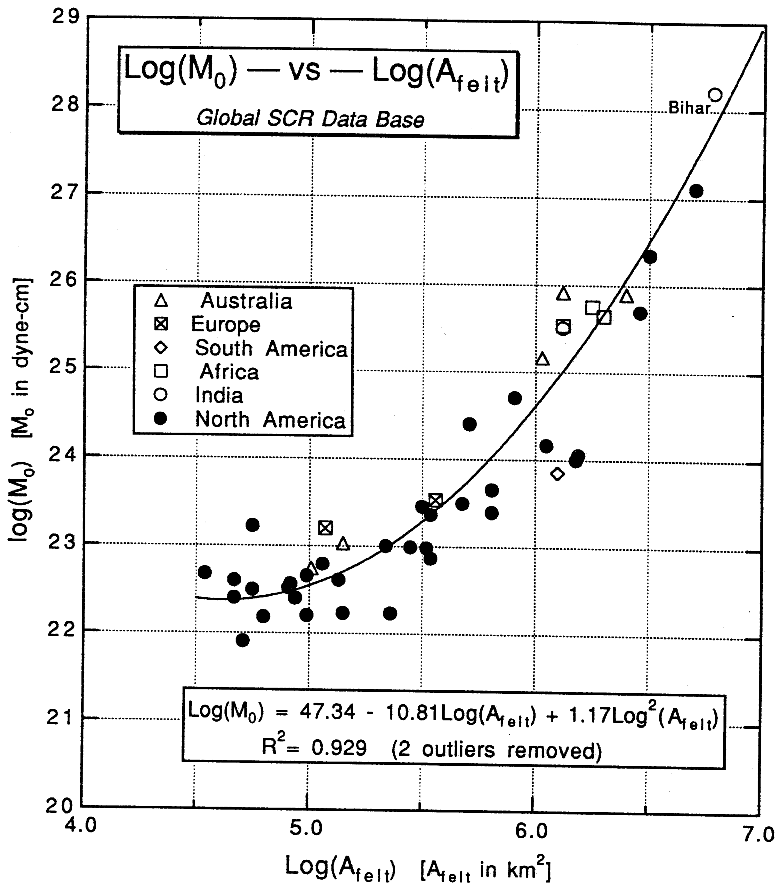

Fig. 5. An example of a regression analysis of $log(M_o)$ on macroseismic or intensity data, in this case MMI felt area for stable continental plate interiors. A quadratic curve fit again was superior to a linear regression. Most of the data scatter probably results from differences in hypocentral depth and regional attenuation, parameters not considered in the regression. Nevertheless, the data are sufficient to constrain the moment magnitude associated with a particular A_{felt} within useful limits. Adapted from Johnston (1993).

reasons. First, the data set for MMI VIII-XI is much smaller than for MMI I-VII, leading to less robust regressions. Second, the upper intensities are more subject to extremes of soil conditions, type of construction and the presence or absence of population centres, so data scatter is increased. I_o (epicentral) or I_{max} (maximum) intensity is an especially poor indicator of earthquake size and should be used only in last resort when no other data are available. (Such is unfortunately the case in many historical earthquake catalogs).

2.4. *A hierarchy of estimation methods for moment magnitude*

Construction of a global seismicity catalog that spans the three data eras described above will require a ranking strategy for assigning event size. All available size indicators should be reported for each event, but for hazard analysis, each event needs one assigned magnitude. In this section we will briefly outline a proposed hierarchy for assigning ***M***.

Let us imagine a hypothetical earthquake

Table II. Hierarchy of estimation methods for $\boldsymbol{M}$.

I. *General categories*

1) $\boldsymbol{M}$ derived directly from M_o
2) $\boldsymbol{M}$ estimated from standard teleseismic magnitudes
3) $\boldsymbol{M}$ estimated from measured isoseismal areas
4) $\boldsymbol{M}$ estimated from regional or non-standard instrumental magnitudes
5) $\boldsymbol{M}$ estimated from quoted intensity areas, radii or magnitudes
6) $\boldsymbol{M}$ estimated from number of recording stations
7) $\boldsymbol{M}$ estimated from epicentral intensity, I_o
8) $\boldsymbol{M}$ assigned by judgement

II. *Detailed categories*

	Method	Assigned quality	Estimated uncertainty of $\boldsymbol{M}$
1a)	Instrumental M_o (spectra, waveform matching, inversion)	A	±0.20
1b)	M_o from field observations	A1	±0.25
2a)	Weighted average of m_b, M_s	B	±0.25
2b)	Teleseismic (20 sec) M_s	B	±0.25
2c)	Teleseismic (1 sec) m_b	B1	±0.30
2d)	Instrumental m_{bLg}	B1	±0.30
2e)	M_{GR} (Gutenberg and Richter, 1954) $\simeq M_s$ (not «class d»)	B1	±0.30
3a)	Isoseismal areas: average of 3-6 contour areas	C	±0.35
3b)	Isoseismal areas: average of 1-2 contour areas	C	±0.35
4a)	Direct M_L-$log(M_o)$ regression	D	±0.40
4b)	Regional m_b, M_s (non-ISC/NEIS)	D	±0.40
4c)	M_L regressed to m_b/M_s then to M_o	D	±0.40
4d)	M_{GR}, «class d» (Gutenberg and Richter, 1954)	D	±0.40
4e)	$\boldsymbol{M}$ from number of recording stations	D1	±0.45
5a)	Magnitudes based on intensity equated to m_b, M_s, m_{bLg}	E	±0.50
5b)	$\boldsymbol{M}$ estimated from quoted isoseismal areas or radii	E	±0.50
7)	$\boldsymbol{M}$ estimated from I_o only	X	±1.0
8)	$\boldsymbol{M}$ assigned by judgement	Z	±1.2

for which a full suite of size indicators are available, from M_o to I_o. There may well be over a dozen different size measures: how do we choose to represent this event in the GSHAP data base? In table II a hierarchy is listed in which various methods are ranked. Any given method takes precedence over all those below it. If, as argued above, $\boldsymbol{M}$ is preferred size measure, then M_o, if available, determines $\boldsymbol{M}$. If M_o is not available, teleseismic or well calibrated regional magnitudes determine $\boldsymbol{M}$ by application of appropriate regressions. Next in priority is regressions on isoseismal areas to estimate $\boldsymbol{M}$, followed by magnitude assignments that are not well calibrated against standard magnitudes or M_o. These are followed in turn by $\boldsymbol{M}$-estimates from number of recording stations (see below), from quoted not measured isoseismal data, and finally estimates from I_o/I_{max}.

The order of the table II hierarchy is somewhat a matter of judgement and may vary regionally. For example, we have ranked $\boldsymbol{M}$ determined from good macroscopic intensity data above $\boldsymbol{M}$ determined from poor (*i.e.*, non-standard magnitudes, indirect or multiple regressions to obtain M_o) instrumental data; each technique has its own set of uncertainties that are difficult to compare. Within a region of uniform geology and type of construction, I_o/I_{max} might be a more reliable size estimator than indicated in table II.

Another arguable ranking in table II is 4-e, $\boldsymbol{M}$ estimated from number of recording stations, N_{st}. We have found this to be a very useful size measure, especially in the early instrumental era when few calibrated magnitudes are available. The problem is that N_{st} varies tremendously over the course of the twentieth century; therefore, N_{st} must be normalized. A particularly good normalizing factor is N_{max}, the maximum number of stations reported by the ISS or ISC for a single earthquake in a given year, because this is an index of the number of active, not just existing, seismic stations (Ambraseys and Melville, 1982). Figure 6 shows N_{max} as a function of time from 1913 to 1990. There is considerable

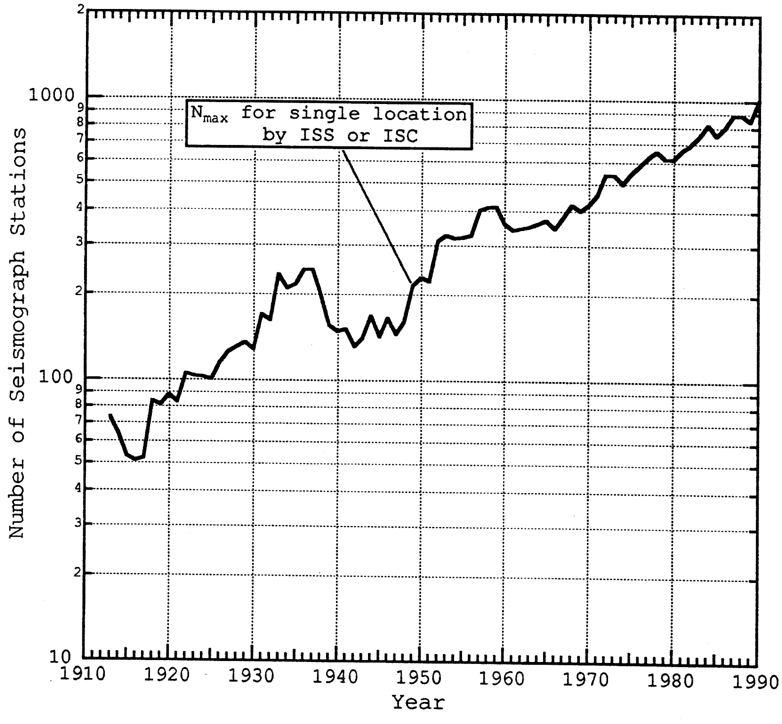

Fig. 6. The maximum number of seismic stations used for a single earthquake location by the ISS and ISC in any given year. This number, rather than the total number of operating stations per year, is used to normalize a regression of number of reporting stations ($\Delta \geq 20°$) versus seismic moment (see text). Data for 1913-1969 from Ambraseys and Melville (1982, fig. 4.1).

variation, but the overall increasing trend is evident, as are the marked decreases in reporting during world wars. In 1990 over 1000 stations reported the deep (h = 603 km) Sakhalin Island earthquake (***M***7.2) of 12 May 1990 to the ISC.

When N_{st} is used to estimate ***M***, geographic location is important. For example, a relatively small event occurring near a concentration of seismographs will have an abnormally high N_{st}, hence ***M***, for its true size. In order to minimize this effect, a good practice is to total only stations beyond $\Delta = 20^\circ$ (designated N_{20st}). The variable for regression on $log(M_o)$ or ***M*** is then $log(N_{20st}/N_{max})$. An example of an application to an Australian earthquake is given below. Ambraseys and Melville (1982) discuss a similar application by using N_{st} to determine M_s in Persia in which they also include maximum distance to a recording station as a variable. They do not normalize N_{st}, however, but instead derive a series of regressions for different time periods.

Events with multiple size measures can be used to evaluate the internal consistency of the table II hierarchy. As an example with particularly abundant data, we will apply the ranking of table II to the largest of the 22 January 1988 Tennant Creek earthquakes, Northern Territory, Australia.

1a) instrumental M_o	five independent M_o determinations, including Harvard CMT; average $log(M_o)$ = 25.88, which yields ***M*** = 6.56	***M*** = 6.56
1b) field observations	fault length from surface scarp of 13-16 km; down-dip fault width from aftershocks~14 km; average fault slip U measured at surface~1 m; using $M_o = \mu U$ *(fault area)*, and $\mu = 3.3 \times 10^{11}$ dyn/cm^2, yields $M_o = 6.0\text{-}7.4 \times 10^{25}$ dyn-cm	***M*** = 6.49-6.55
2b) teleseismic M_s	M_s(ISC) = 6.5; M_s(PDE) = 6.7; using fig. 4 regression on av. M_s = 6.6 yields $log(M_o)$ = 25.93	***M*** = 6.59
2c) teleseismic m_b	m_b(ISC) = 6.3; m_b(PDE) = 6.5; using regression from Johnston (1993) on av. m_b yields $log(M_o)$ = 25.79	***M*** = 6.49
3a) isoseismal areas	from McCue (1990) obtain MMI A_{felt}, A_{IV}, A_V, and A_{VI} in km^2. Using regressions from Johnston (1993) yields = 26.34 (***M***6.8), 26.36 (***M***6.9), 26.31 (***M***6.8), and 25.34 (***M***6.2) for average ***M*** = 6.67	***M*** = 6.67
4e) N_{st}	N_{St}(ISC) = 591; N_{20st} = 564. From fig. 6, N_{max} = 887; from Johnston (1993), $\boldsymbol{M} = 7.49 + 2.94 log(N_{20st}/N_{max}) + 0.71 log^2(N_{20st}/N_{max})$	***M*** = 6.94
7) I_o	I_o was not available; I_{max}=MMI VII ~35 km from the epicenter; most regressions on I_o/I_{max} yield ***M*** in the range 4.8-5.8	***M*** = 4.8-5.8

Each of these estimates of the moment magnitude are within the stated uncertainties of table II of the true ***M*** if it is taken as the category 1-a value. Usually M_o will not be available, either from seismic wave analysis or field data. Then intercomparison of a number of estimation techniques can be valuable for estimating the reliability of individual determinations of ***M***.

In this Tennant Creek example the data were high quality, and the results reflect an above-average internal consistency among the various table II approaches. A counterexample to Tennant Creek is the 25 November 1988 Saguenay, Quebec earthquake for which an average of six independent M_o determinations yields a moment magnitude of ***M***5.9. The short-period magnitude m_{bLg}, however, was 6.5, yielding an estimated ***M*** = 6.5-6.8. Even worse, if instrumental data were not available and the Saguenay moment magnitude were estimated from the isoseismal areas of Drysdale and Cajka (1989), the SCR regressions of Johnston (1993) yield ***M***~6.9. Detailed analysis of this event (*e.g.*, Boore and Atkinson, 1992) show it to be abnormally deep (h=27 km) with an abnormally high stress drop ($\Delta\sigma$>500 bars). It is therefore an outlier in the SCR data set, but had it been an historical earthquake we would not have known that and assigned ***M*** = 6.9±0.35.

The above Tennant Creek and Saguenay applications illustrate best- and worse-case examples of indirect estimation of moment magnitude. Regressions will only predict «average» behavior; earthquakes in the outlying tails of frequency distributions will be poorly handled by regression. Fortunately, by definition, truly abnormal events are rare. The advantages of having a single uniform size measure with uncertainty estimates outweigh the disadvantage of inadequate treatment of abnormal events.

3. Concluding remarks: the scope of the GSHAP data base

It is clear that an exhaustive reexamination the entire global seismicity of data base from all three eras as discussed above is a formidable task, one beyond the scope of GSHAP. Decisions are necessary as to how to best curtail the compilation of the data base without unduly adding to the uncertainty of subsequent hazard analysis. We will conclude this report by examining some of the issues involved with establishing the optimum scope of the project to assemble the GSHAP data base.

The first and primary issue is the size of the data base. If the GSHAP data base were to set a minimum magnitude goal of ***M***~4.0, the sheer number of events would be prohibitive. In the 29 years of the modern instrumental era alone, ~115 000 shallow events of ***M***≥4.0 are expected from the PS92 recurrence formula. if the early instrumental era is included, there would be >350 000 expected instrumental events. Moreover, numerous events in the magnitude 3 range would require analysis as to whether their size was underestimated.

The number of historical events requiring analysis is unknown, but an ***M*** ~4.0 threshold would require analysis of $\mathrm{MMI_o}$VI-level events, probably at least doubling the effort needed to compile the GSHAP historical data base. Theoretically, then, an ***M***~4.0 threshold could lead to a data base approaching one-half million events (an ***M***5.0 threshold yields ~50 000-100 000 events by the same reasoning). Since it is extremely rare for an earthquake in the ***M***4 range to be destructive, why should such a threshold be considered for a hazard analysis data base?

M4-range earthquakes contribute negligibly to deterministic hazard assessment, but they are important for probabilistic hazard analysis. This is in part because they can affect determination of the b-value slope in recurrence curves, but also because at high frequencies the low probability that ***M***4-range earthquakes will produce high ground accelerations is partially counterbalanced by their relative abundance (Reiter, 1990). In many stable continental regions, large earthquakes are so infrequent that the major portion of the hazard may be contributed by ***M***4-range and low ***M***5 events.

The most practical approach for GSHAP may be to take a regional approach to the minimum magnitude threshold question. In plate boundary zones with high seismicity, the M_{min} threshold could be set much higher than in many midplate regions with little loss to the rigor of the probabilistic assessment of the region's seismic hazard.

A second issue concerns scope of the data base project but not in terms of number of events. To what extent should GSHAP undertake reanalysis or new analysis of existing data? There are numerous areas in which global data could be markedly improved with modern techniques. A non-exhaustive listing includes: 1) reanalyze m_b and M_s assignments for PDE data from 1964-1968 and ISC from 1964-1981, applying current frequency and depth restrictions; 2) electronically scan the ISS bulletins and relocate the ISS earthquakes using modern location techniques; 3) compile a global compendium of standardized isoseismal maps and descriptions for important earthquakes in all three eras and construct new maps for unanalyzed quakes for which data are available; and 4) compile a global bibliography of all primary and secondary data sources, cross-referenced to events in the GSHAP data base.

Each of these projects would be of immense value and result in a significantly improved data base. But are any of them feasible and/or realistic for GSHAP to undertake? The alternative is to limit the data base project strictly to material already in existing data bases and the literature and try to account for its shortcomings with increased uncertainty bounds on earthquake locations and magnitudes, completeness levels, and maximum credible magnitudes.

A third issue also relates to scope in terms of numbers of events but has a different, almost philosophical, aspect. Should deep earthquakes and oceanic crust earthquakes be included in the GSHAP data base? Clearly the vast majority of oceanic ridge and transform events contribute nothing to the seismic hazard of continental regions. The same holds true for most deep oceanic subduction events, although large events at depths exceeding 500 km can be felt at the surface. The hazardous intermediate-depth earthquake zones within continents, such as in Romania and Columbia or the Hindu Kush zone of South-Central Asia, can be easily isolated and included in the shallow event analysis. Why, then, should the GSHAP consider including deep and oceanic events?

The answer to this question relates to the overall goal that the GSHAP defines for the seismicity data base. If it is to be narrowly focused on supplying the input data that are necessary to quantify seismic hazard worldwide, then the deep and oceanic events should be omitted because of the attendant savings in time and effort. But a broader view is also possible. The GSHAP project is unique in that for the first time in history its intent is to compile a comprehensive global seismicity data base that stretches from historical times to the present. The value of such a product will go far beyond the immediate and focused GSHAP hazard objectives. Such a truly global data base will be of immense value to earth science research, but only if it is truly global.

We believe that for the relatively small increase in effort involved, GSHAP should commit to producing the most complete catalog possible of Earth's seismicity. It has never before been attempted. It will ensure that the value of the GSHAP will continue long after the global analysis of hazard has been completed.

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