Reappraisal of north-Indian earthquakes at the turn of the 20th century

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This paper reviews the seismicity of northern India during the little-known period from 1892 to 1915 and summarizes evidence for 50 events, few of which are listed in parametric catalogues. Using instrumental data and established procedures, surface-wave magnitudes are calculated uniformly and epicentral locations are reappraised.

Introduction

AS we cannot know what will happen in the future, to estimate likely earthquake hazards we have to find out what happened in the past and extrapolate from there. Previous research has uncovered evidence of destructive earthquakes in areas where only small events have been experienced recently. This is not surprising: the time-scale of geology is vastly different from that of human history. It follows, that if we took account only of information about the last few decades of advanced instrumentation, we would have no way of knowing whether an apparently 'quite' area is in fact at risk from a damaging earthquake.

The purpose of this paper is to assess, by combining instrumental and macroseismic information, the location and to assign uniformly calculated surface-wave magnitudes to earthquakes in northern India and Pakistan for the 24-year period of instrumental recording between 1892 and 1915, both inclusive. It draws on new sources of macroseismic information, supplemented by re-examination of instrumental reports, to evaluate the position and size of significant earthquakes in the region between latitudes 25° and 35°N and longitudes 65° and 95°E, which includes much of Pakistan, northern India, Nepal and southern Tibet.

With the exception of the very large earthquakes in the region during that period, few of the 50 events we have retrieved are listed in national or global parametric catalogues. The information and procedures presented here are important for the unambiguous evaluation of tectonics and of seismic hazard in this part of the region.

Sources of information and location

The study area includes the Quetta–Chaman Fault, the Hazra Arc, the Himalayan Frontal Thrust and the area to the north in Tibet of the Karakorum Fault and the Indus-Tsangpo suture zones (Figure 1).

In preparing this work our first task was to search all existing catalogues known to us covering the region for the period 1892 to 1915 and to compare and combine entries. For each earthquake we examined all pertinent data and references cited and retrieved additional macroseismic and instrumental information. Our main sources of macroseismic information are the local and the international press, limited to the *Civil & Military Gazette*, *Paisa Akhbar*, Lahore; *Statesman*, Calcutta; *Times of India*, Bombay and *The Times*, London. These are supplemented by Oldham[1,2] (12 June 1897, Assam), Middlemiss[3], Baduwi[4], and Omori[5] (4 April 1904, Kangra); Heron[6] (21 October 1909, Baluchistan), Christiansen and Ziemendorff[7], Oddone[8], Paterson[9], Rosenthal[10], Rudolph[11], Sieberg[12], Szirtes[13], Walker[14], District Gazetteers of India, as well as unpublished consular correspondence, particularly from the less accessible regions of the North-West Frontier and Baluchistan which have been useful.

Instrumental data come from station bulletins worldwide, reporting phase data and ground amplitudes as well as from bulletins of regional networks such as *Notizie sui Terremoti Avveniti in Italia* 1897–1915, published by the *R. Ufficio Centrale di Meteorologia e Geodinamica* in Rome; *Reports, Seismological Investigations* 1895–1915, *Circulars* 1899–1913 and *Bulletins* 1913–1915 published by the Seismological Committee of the British Association for the Advancement of Science; *Bulletin de la Commission Centrale Seismologique Permanente* 1902–1908 and 1911–1913 supplemented by the *Bulletins of 1st class stations in Russia* for Irkutsk 1902–1903, 1912–1914, Pulkovo 1912–1915, Sverdlovsk 1913–1915, Tashkent 1913–1915, published by the Imperial Academy of Science in St. Petersboug; *Monatsberichte* 1899–1906 and *Wöchentlicher Erdbeben-Bericht* 1905–1915 *der Kaiserlichen Hauptstation für Erdbebenforshung* in Strassburg, as well as Gutenberg's unpublished work-sheets and station bulletins from many of the stations listed in Table 1. For the recording capabilities of these early stations, see Ambraseys and Finkel[15], material which is indispensable for the study of the seismicity of the first two decades of this century, not only for earthquakes in northern India but also in Europe, the Mediterranean region, and the Middle East.

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Into this improved data set, a separate body of information could then be incorporated, that of unassociated station onset or maximum phase readings of earthquakes in the region. This was done by examining station bulletins and identifying events with large surface-wave amplitudes recorded about the same time at regional stations, such as Bombay, Calcutta, Colombo, Irkutsk, Kodaikanal and Shimla. If arrival or maximum phase readings suggested common origin, readings from other, more distant stations were sought and the general position of the event can then be tested and occasionally confirmed by macroseismic evidence, which on its own could not have identified the event or been used to estimate magnitude. For most of the earthquakes identified in this way, readings from near stations were available from low-magnification undamped Milne instruments. The most reliable reading from these instruments is the time of the maximum amplitude, and the best available instrumental control comes from assigning this arrival to the velocity of the Airy phase maximum of surface waves. The slow velocity reduces uncertainties due to errors in reporting times[16]. In the best instances for the early period, instrumental locations appear to be accurate to a few, up to five degrees – enough to reveal gross mislocations or to establish the general area of an earthquake, but not enough to be preferred over well-determined macroseismic locations.

If the general location of the earthquake identified in this way was within or near our study region, we then searched for associated macroseismic information in newspapers and in technical reports. For the north and east parts of the region, which are rather remote, we used unpublished information from the India Office Records in London, particularly Political and Secret Correspondence (IO: L/P&S) from Baluchistan, Dir, Gilgit, Kabul, Kalat, Katmandu, Sibi, Swat, Tashkent, and from the files of the Meteorological Service (Earth Tremors 005.1914) at Quetta.

For some earthquakes in the last few years of the period, there are enough reliable reports of P phases to enable computer relocations to be carried out using the present procedures at the International Seismological Centre (ISC). The use of such procedures, however, still cannot overcome uncertainties of location arising from poor azimuthal distribution of stations. This search provided useful and sometimes detailed information revealing the occurrence of relatively large shocks which were previously unknown, allowing the instrumental determination of their surface-wave magnitudes and the identification of mislocated positions.

Epicentre locations

Instrumental epicentres for earthquakes in the first half of the period investigated are very approximate, and they must be used with caution. For this period the British Association for the Advancement of Science (BAAS)[17,18] published a considerable number of epicentres of the larger shocks for some of which it seems, macroseismic information was used but not quoted, to determine the

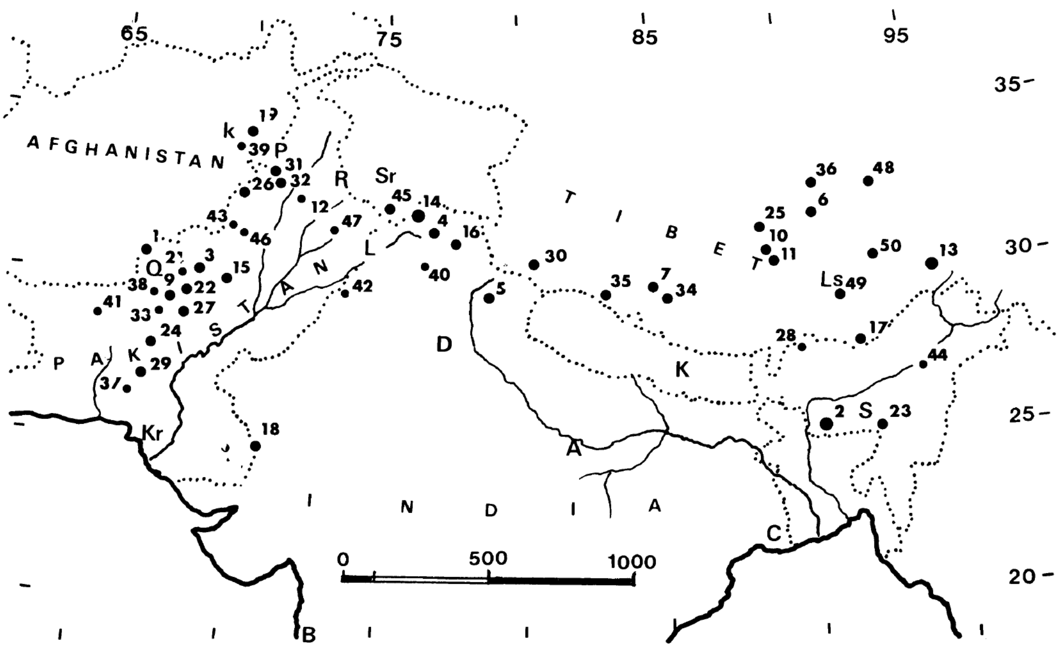

Figure 1. Location map of earthquakes in northern India during the period 1892 to 1915. Numbers refer to entries in Table 4 and size of circles denotes events of $M_S < 6.0$, 7.0 and 8.0. A, Allahabad; B, Bombay; C, Calcutta; D, Delhi; K, Katmandu; Kr, Karachi; k , Kabul; L, Lahore; Ls, Lhasa; P, Peshawar; Q, Quetta; R, Rawalpindi; S, Shillong; Sr, Srinagar.

approximate position and time of the event, whenever it was possible. This class of determinations is not really instrumental. When local observations were not available, as is the case with most of the earthquakes in remote parts of the region, epicentres are likely to be in error. The same applies to epicentral determinations made by Strassburg and by Russian stations.

Macroseismic epicentres are also an approximate indication of the location of an earthquake. For large events, such locations are adequate for the determination of their magnitude or for their association with local tectonics, considering that earthquakes of $M_S > 6.5$ will have ruptured faults tens of kilometres or more in length, in which case the definition of epicentre loses its practical meaning.

Table 1. Seismographic station used for events 1892–1915

Station	N/S°	E/W°	D°	Az°	Period	Instrument
Azores	37.75	– 25.67	83	309	1903–1915	M
Baku	40.38	49.90	27	301	1912	G
Baltimore	39.30	– 106.82	110	7	1901–1913	M
Batavia	– 06.18	106.83	44	140	1900–1915	M
Beirut	35.90	35.47	38	29	1904–1914	M
Beograd	44.82	20.45	48	305	1911–1915	W
Bidstone	55.40	– 3.10	62	320	1901–1915	M
Bombay	18.547	72.82	13	211	1900–1917	M
Calcutta	22.53	88.37	11	133	1900–1914	M
Cape	– 33.93	18.48	86	227	1900–1915	M
Cartuja	37.18	– 3.60	68	301	1905–1915	V
Cheltenham	38.73	– 76.85	108	341	1904–1909	M
Christchurch	– 43.38	172.60	112	128	1901–1913	M
Colombo	6.90	79.87	23	180	1906–1915	M
Cork	51.88	8.47	66	317	1912–1914	M
Debilt	52.10	5.18	58	315	1908–1915	W
Edinburgh	55.92	– 3.18	62	321	1901–1915	M
Gottingen	51.55	9.97	55	314	1903–1914	W
Graz	47.77	15.45	52	309	1907–1915	W
Guilford	51.25	– 0.50	61	315	1910–1915	M
Hazlemere	51.08	– 0.72	62	315	1906–1914	M
Helwan	29.85	31.35	42	283	1904–1913	M
Hohenheim	48.72	9.22	56	311	1912–1915	m
Honolulu	21.32	– 158.02	104	55	1901–1915	M
Irkutsk	52.27	104.32	29	32	1901–1914	M
Irkutsk	52.27	104.32	29	32	1912–1914	G
Kew	51.47	– 0.32	61	316	1900–1915	M
Kodaikanal	10.23	77.47	20	187	1900–1915	M
Krakow	50.05	19.93	47	312	1909–1915	B
Ksara	33.82	35.87	37	288	1910–1914	m
Leipzig	51.35	12.38	53	314	1904–1913	W
Ljubljana	46.05	14.52	52	307	1913–1914	G
Lvov	49.82	27.03	44	312	1912–1915	B
Malta	35.90	14.52	54	295	1906–1915	M
Mauritius	– 20.10	57.88	54	206	1900–1915	M
Osaka	34.70	135.5	47	69	1901–1915	O
Perth	31.95	– 115.83	71	148	1901–1915	M
Pola	44.87	13.85	53	306	1912–1915	W
Potsdam	52.38	13.07	53	315	1902–1915	W
Pulkovo	59.77	30.32	44	327	1912–1915	G
San Fernando	36.47	– 6.20	85	271	1900–1915	M
Shide	50.68	– 1.68	62	315	1900–1913	M
Stonyhurst	53.85	– 2.47	62	318	1909–1915	M
Strasburg	48.58	7.77	57	311	1906–1915	W
Sverdlovsk	56.83	60.63	30	339	1913–1915	G
Sydney	– 33.87	151.20	93	128	1906–1915	M
Tashkent	41.33	69.30	14	325	1913–1915	G
Tokyo	35.68	139.75	50	67	1900–1910	M
Toronto	43.67	– 79.40	104	345	1900–1912	M
Trieste	45.65	13.77	53	307	1911–1915	W
Uccle	50.80	4.37	58	314	1909–1913	W
Uppsala	59.85	17.63	51	325	1905–1915	W
Vieques	18.15	– 65.43	121	320	1903–1906	M
Victoria	48.52	– 123.42	99	15	1900–1912	M
Vienna	48.25	16.37	51	310	1907–1915	W

D, distance of station from centre of study area taken at 30^0N, 80^0E in degrees; Az, azimuth of station in degrees; Apparatus: B, Bosch; G, Galitzin; m, Mainka; M, Milne; O, Omori; V, Vincentini; W, Wiechert.

For a number of earthquakes, chiefly in Tibet, there is total lack of macroseismic information and instrumental data are insufficient to provide a solution. In these cases we adopted the epicentres which are given by BAAS or improved on them. Nevertheless, the macroseismic evidence is by far the most reliable, and we give it more weight.

Surface-wave magnitude

Before we proceed with the evaluation of magnitudes, it is important to review briefly the development of the surface-wave magnitude M_S. According to Abe[19], M_{GR}, the magnitude which is based on surface waves for shallow events first used by Gutenberg and Richter[20], is equivalent to the surface-wave magnitude. M_{GR} was devised in order to extend the local magnitude M_L defined in the previous year[21] to teleseismic distances and was more thoroughly developed in a subsequent paper[22]. The M_{GR} original scale was based on the maximum horizontal ground displacement A_{max}, but it was specified that measurements were to be made at periods near 20 s, although it is evident from Gutenberg's work-sheets, that quite often Gutenberg himself did not observe this rule. He used periods from 10 to 25 s, and quite often values which differ from those published in station bulletins. There are different interpretations of the structure of Gutenberg's magnitude M_{GR}, of its changes over the time of its development and of the method this author used to choose maximum phase amplitudes, a subject which is outside the purpose of this paper[19,23,24]. Table 2 lists all the earthquakes in our region for which surface-wave magnitudes have been estimated by various authors. For some of these events estimates vary widely for reasons which will shortly be seen.

Prague surface-wave magnitude

An improvement of the scale was made by Soloviev[25] who proposed a surface-wave magnitude in which the maximum ground particle velocity $(A/T)_{max}$, a physical quantity which accounts better for the seismic energy flux at a seismographic station than the ground displacement A_{max} at 20 s period, was used as the variable. Soloviev's scale is not restricted to a given period, and M_S can be calculated within a broad range of distances of 4 to 80°. The general formula for the station surface-wave magnitude M_{si} was defined later as:

$$M_{s,i} = \log(A/T)_{max} + s(D, h) + C. \tag{1}$$

In the above equation, A is the ground displacement in micrometres and T is the period in seconds associated with the maximum particle velocity $(A/T)_{max}$. $s(D, h)$ is an empirical ground velocity-distance calibration function which expresses the change of particle velocity with epicentral distance D and focal depth h, and C is a correction term which allows for the effects at the recording site, wave path, variations in depth and focal mechanism[26].

Following Soloviev, Karnik *et al.*[27] and Vanek *et al.*[28], a calibration relation is given by the following equation:

$$\mathrm{s}(D) = 1.66 \log(D) + 3.3. \tag{2}$$

This new equation was derived originally from the weighted average of 14 attenuation functions existing at the time for epicentral distances between 20 and 160° and for an wide range of surface-wave periods. These 14 attenuation functions, and subsequent functions used to control eq. (2) are given in Soloviev[29], Karnik[30], cf. Lienkaemper[24]. Later, the validity of eq. (2) was confirmed further for smaller distances of a few degrees[31,32].

The calibration relation given by eq. (2) was adopted by IASPEI in 1967 (ref. 33), specifically in order to avoid the limitations imposed by the restriction to near 20 s period waves in Gutenberg's method. Equation (1), commonly referred to as the original 'Prague formula', was then defined as:

$$M_S = \log(A/T)_{max} + 1.66 \log(D) + 3.3 + C_i. \tag{3}$$

Table 2. Published earthquake magnitudes

Date	OT	N°	E°	M_S	M_G	M_D	M_{A1}	M_{A2}	M_{A3}
1892 12 20	0020	30.9	66.5 m	6.79[1]	0	0	0	0	6.2[2]
1897 06 12	1106	25.5	91.0 m	7.98[1]	8.7[4]	8.7	0	8.2[3]	8.0[2]
1905 02 17	1142	26.0	96.0 B	7.11	0	7.3	0	7.1	6.8
1905 04 04	0050	33.0	76.0 m	7.83	8.0	8.6	8.1	8.1	7.5
1905 09 26	0128	28.8	74.1 R	6.44	0	7.1	0	0	0
1908 03 05	0220	30.2	67.7 m	6.43	0	7.5	0	7.3	6.9
1908 08 20	0953	32.0	89.0 B	7.04	0	7.0	0	7.0	0
1909 10 20	2341	28.9	68.3 m	7.12	7.2	7.0	7.1	7.1	7.0
1911 10 14	2324	31.0	80.5 G	6.42	6.8	0	0	0	0
1913 03 06	0208	30.0	85.0 B	6.41	6.2	0	0	0	0
1913 03 06	1103	30.0	83.0 B	6.50	6.4	7.3	0	0	0
1915 12 03	0239	29.5	91.5 B	6.65	0	7.1	7.0	0	0

M_G, Gutenberg and Richter[61]; M_D, Duda[60]; M_{A1}, Abe[19]; M_{A2}, Abe and Noguchi[50]; M_{A3}, Abe and Noguchi[73].
[1]Calculated in this study; [2]Abe[52]; [3]Kanamori and Abe[69]; [4]Gutenberg[69].

In this equation, C_i is a station correction term, which allows for the effects at the recording site and wave path. Recommended period ranges corresponding to maximum amplitudes of surface-waves at different epicentral distances were also given by other sources[27,33,34]. The Prague formula was devised to be used with shallow events (h < 40–50 km) and to have a depth adjustment for deeper events. Derivation of depth correction[35,36] will not be discussed here. With few exceptions, workers and agencies do not use the Prague formula according to its original definition. Since the mid-to-late 1970s, surface-wave magnitudes reported by both the National Earthquake Information Service (NEIS) and the International Seismological Centre (ISC) have been computed and published using the Prague formula, but each agency selects data using different criteria which are not consistent with the definition of the original Prague formula.

Up to 1975, NEIS published estimates of M_S from readings on horizontal components at individual stations, but from May 1975, the assessment has been made only from the vertical component of the surface-wave within the restricted period range of 18 to 22 s and for distances between 20 and 160° (e.g. ref. 24). It is theoretically more correct to use the vertical, rather than the horizontal components because the vertical component records only waves of Rayleigh type, while the horizontal records both Love and Rayleigh waves, with resulting complication in attenuation characteristics. No depth or station corrections are applied by NEIS and M_S magnitudes are not generally computed for events with focal depths greater than 50 km.

Before 1971, ISC neither reported long-period amplitudes and periods nor calculated M_S. During 1971 and 1976, ISC reported amplitudes and periods for all components, and M_S was calculated by combining vectorially, the maximum reported amplitudes of the two horizontal components at periods near 20 s for stations in the distance range 20 to 160°, and using the attenuation relationship from the Prague formula. These determinations were given with the station readings only and were not included with the epicentres. Between 1976 and 1978, magnitude determinations from the vertical components were included for very few events, the distance range was extended to between 5 and 160° and the range of allowable periods to between 10 and 60 s.

Since 1978, event magnitudes determined using these criteria have been given with epicentres, but for the whole period up to today, only those stations at distances between 20 and 160° are used in averaging them and are given as ISC M_S estimates. These are estimated only for events at depths of 60 km or less. Thus, M_S magnitudes reported by ISC are calculated with the exclusion of amplitude and period data from distances smaller than 20°. This implies, for instance, that no recordings from Indian seismographic stations can be used to calculate M_S for earthquakes with epicentres in this region, which obviously is not necessarily true.

NEIS and ISC thus use different selection criteria in choosing stations for the calculation of M_S, and for a particular event, the number of stations used and their distribution in azimuth may be different. In addition, ISC usually uses more station readings in determining event magnitudes than NEIS but neither of them reports standard deviations of their estimates.

Station magnitudes M_{si} estimated for the same event from different stations often diverge. Some such divergences represent real irregularities in wave propagation while systematic magnitude station errors may arise from other factors, and these can be corrected using station corrections. Event magnitudes are calculated from the arithmetic mean of station magnitudes and we may briefly discuss the problem that possibly escapes attention because it is so familiar, i.e. the method of averaging station magnitudes $M_{S,i}$ to calculate event magnitude M_S. In current procedures, M_S is calculated from the arithmetic mean of station magnitudes $M_{S,i}$, which involves the mean of $\log(A/T)$ terms for different stations. However, the average seismic energy density at a station is proportional to (A/T) and not to $\log(A/T)$ and therefore when $M_{S,i}$ values are averaged, the M_S is underestimated. This would not be a problem if the underestimation was constant but in reality, it varies depending on the variance or distribution of station magnitudes, $M_{S,i}$. Consider the general station magnitude equation:

$$M_{si} = \log(A/T) + b\log(D) + c, \tag{4}$$

where b and c are constants. This can be transformed to:

$$10^{M_{Si}} = 10^c D^b (A/T). \tag{5}$$

Therefore $10^{M_{S,i}}$ is proportional to the energy density at the station. Hence the new event magnitude, $M_{S,n}$, could be more correctly defined as:

$$M_{S,n} = \qquad {}^{M_{S,i}}/N]. \tag{6}$$

Note that if there is only one station magnitude then $M_S = M_{S,n}$, therefore the original station magnitude definition is correct. It can be shown, by means of an expansion, that

$$M_S \le M_{S,n}, \tag{7}$$

as long as the difference between $10^{M_{S,i}}$ and $10^{M_{S,n}}$ is not too great. The difference between $M_{S,n}$ and M_S can be estimated by R:

$$R = [(1/N \qquad {}^{M_{S,i}} - 10^{M_{S,n}})^2]/(2\ln(10)(10^{M_{S,n}})^2), \tag{8}$$

which depends on the variance of $10^{M_{S,i}}$. Although there is no exact relationship between R and the variance of $M_{S,i}$, they are roughly proportional. Thus the underestimation in magnitude by using M_S rather than $M_{S,n}$ is greatest when the station magnitudes are widely distributed about the event magnitude. R is a good estimator of the error when the underestimation is less than about 0.25 but a poor estimate for larger errors. This is due to the higher order term in the expansion becoming more important, and the errors can be significant.

Station corrections

Station correction C_i in eq. (3) for a particular station i is defined as the mean of the residual $M_S - M_{si}$ over a period of time, i.e. $C_i = \Sigma({}_jM_S - M_{si})/N$, where M_S is the event magnitude, M_{si} is the station magnitude and N is the number of events observed by the station. To the best of our knowledge, the first systematic estimation of station corrections from earthquakes in Europe was made by Karnik[30] who employed the original Prague formula to assess corrections for 170 earthquakes, mostly from Europe. He used earthquakes in Europe and adjacent regions, of all magnitudes and depths, during the period between 1904 and 1951. Station corrections were calculated from shallow earthquakes in Iran[37]; the Eurasian continent[35], Central America[39], worldwide[40] and for the Middle East up to longitude 100°E (ref. 41).

Distance correction of the original Prague formula

There is considerable discussion in the literature as to the best practice for the determination of surface-wave magnitude. Since its adoption by IASPEI in 1967, there has been much debate about the adequacy of the amplitude-distance function of M_{si} in eq. (2) discussed earlier[35,38,42–47]. It must be pointed out, however, that most of these authors examined the Prague formula at periods near 20 s in the distance range 6 or 20° to 160°, using M_S estimates made by NEIS or ISC, and not from station readings made according to the original definition of the scale.

Ambraseys and Free[35] examined the distance dependency of the residuals from station magnitudes and they found that the restriction of the data to the 18 to 22 s period range causes the original Prague formula to require a correction of its distance term for both global and regional data. The selection of data over the much broader range implicit in the original version of the Prague formula reduces this requirement. They concluded that the distance dependence $dM = M_{i,3} - M_{i,1}$, where $M_{i,1}$ and $M_{i,3}$ are the station magnitudes calculated without and with distance correction, remains statistically significant but small, when they adhere to the original definition of the Prague formula[48]. Thus,

$$dM = 0.518 - 0.282 \log(D). \tag{9}$$

However this correction becomes significant for magnitudes derived exclusively from few close stations which is usually the case with small events.

Amplitude/period data

In the calculation of M_S from damped instruments we used only ground amplitudes as reported in station bulletins. When trace amplitudes were given, because of the uncertainties in the calibration constants of early analogue seismographs, these values were not converted into ground amplitudes and were not used.

Reappraisal of magnitude in the northern Indian region

For the present study and for the period 1892 to 1915, amplitude and period data for the calculation of station magnitudes $M_{S,i}$ were taken from station bulletins. For practical purposes, our re-evaluation was divided into two broad, overlapping periods of observation, dictated chiefly by the type of instruments available. One, an early period from 1892 to 1913 in which the majority of instruments worldwide was undamped or lightly damped pendula, and two, the period from 1903 to 1915, of medium-period damped analogue recorders.

One method to calculate an equivalent surface-wave magnitude M_M for events in the early period, 1897 to 1913, is to use the maximum amplitude from the Milne pendulum, culled from the Shide Circulars[49] and station bulletins, using the formula:

$$M_M = \log(A) + 1.25 \log(D) + 4.36, \tag{10}$$

where $2A$ is the peak-to-peak trace amplitude in millimetres on the single component of the Milne pendulum and D the epicentral distance in degrees. This formula was originally derived from earthquakes $M_S \geq 5.0$ and $D \geq 4°$ in Eastern Europe, the Mediterranean region, Iran, western Asia and Africa for which trace amplitudes for the calculation of M_M as well as ground amplitude/period (A/T) data were available for the calculation of the corresponding value of M_S (ref. 37). Equation (10) is used in this study to calculate M_M earthquakes in the period 1897 to 1913. A similar relation

$$M_A = \log(A) + 1.66 \log(D) + 3.63, \tag{11}$$

was derived for large shallow earthquakes ($M \geq 7.0$) worldwide recorded at large distances[50,51]. Table 3 lists the events for which M_M and M_A were calculated, together with their standard deviation dM and the number of station magnitudes N used, (for $M \geq 6.0$ and for $M < 6.0$).

For the second period, which starts in 1903 with the operation of analogue seismographs in Europe at Potsdam, Gottingen, Uppsala and Leipzig, we used the original Prague formula and M_S estimates were corrected for both station and distance (e.g. M_{SC}) using the modified Prague formula[48]. Table 3 lists the earthquakes for which M_S and M_{SC} were calculated together with the standard deviation dM of the estimates and the number of station magnitudes N.

Summarizing, in the case of earthquakes for which amplitude and period data from damped seismographs are available, M_S may be calculated from the Prague formula[34] or from its modified version, M_{SC} with distance and station corrections[35,41,48]. In the case of earthquakes for which amplitude and period data from damped mechanical instruments are lacking, trace amplitudes from Milne recorders may be used together with eq. (10) or eq. (11) to assess equivalent magnitudes M_M. The

Table 3. List of earthquakes with $M \geq 6.0$ and $M < 6.0$ (1892–1915)

Date	OT	Epicentre	M_S	dM	M_{SC}	dM	N	M_M	dM	M_A	dM	N
$M \geq 6.0$												
1892 12 20	0020	30.9 66.5 m	0	0	0	0	0	6.79*	0	0	0	02
1897 06 12	1106	25.5 91.0 m	0	0	0	0	0	7.98*	0	0	0	09
1901 10 17	0557	30.5 68.5 m	0	0	0	0	01	6.14	0.15	6.04	0.23	07
1901 11 18	0004	32.0 77.0 B	6.48	0	6.23	0	01	6.31	0.31	6.18	0.28	10
1902 11 04	1133	32.0 91.0 B	6.67	0.05	6.47	0.01	02	6.42	0.32	6.36	0.31	13
1902 12 13	1708	30.0 85.0 B	6.92	0	6.67	0	01	6.36	0.36	6.29	0.37	14
1903 12 03	2126	19.5 95.0 m	6.56	0	6.43	0	01	6.54	0.35	6.33	0.47	03
1903 12 23	0300	29.5 67.5 m	0	0	0	0	0	5.90	0.12	5.70	0.16	05
1904 03 31	0216	31.0 89.0 B	6.97	0.07	6.91	0.08	02	6.65	0.26	6.62	0.25	20
1904 03 31	0545	31.0 89.0 B	6.22	0.01	6.17	0.14	02	6.20	0.24	6.13	0.29	12
1905 02 17	1142	30.0 95.0 B	7.11	0.19	7.11	0.14	05	6.71	0.26	6.68	0.31	18
1905 04 04	0050	32.1 76.4 m	7.83	0.18	7.83	0.05	06	7.54	0.23	7.54	0.29	20
1905 09 26	0128	30.3 69.9 m	6.44	0.21	6.44	0.08	04	6.44	0.32	6.34	0.37	14
1906 02 27	1940	31.5 77.5 m	6.45	0.11	6.40	0.22	05	6.45	0.27	6.38	0.20	16
1906 05 12	0550	28.0 92.0 B	6.49	0.13	6.44	0.17	05	6.29	0.21	6.19	0.25	12
1906 08 15	2211	25.0 71.0 m	6.05	0.35	6.08	0.25	05	5.90	0.32	5.79	0.397	05
1907 03 29	2053	35.0 70.0 m	6.20	0.39	6.15	0.29	06	6.15	0.34	6.13	0.37	12
1908 03 05	0220	30.2 67.7 m	6.58	0.30	6.43	0.30	07	6.50	0.32	6.43	0.39	12
1908 06 03	1556	28.0 67.0 B	6.24	0.29	6.18	0.30	08	6.14	0.34	6.06	0.40	09
1908 08 20	0953	32.0 89.0 B	7.12	0.26	7.04	0.20	08	6.75	0.31	6.73	0.23	17
1909 10 20	2341	28.9 68.3 m	7.16	0.22	7.12	0.25	09	7.19	0.23	7.06	0.32	16
1910 08 17	1158	27.0 67.0 m	6.33	0.34	6.33	0.27	08	6.45	0.22	6.39	0.25	17
1911 10 14	2324	31.0 80.5 G	6.43	0.30	6.42	0.28	11	0	0	0	0	0
1912 08 23	1402	33.5 71.0 m	6.35	0.41	6.32	0.26	12	6.35	0.26	6.31	0.27	20
1912 08 23	2114	33.5 71.0 m	6.30	0.31	6.27	0.14	07	6.01	0.25	5.96	0.26	16
1913 03 06	0208	30.0 85.0 B	6.42	0.38	6.41	0.26	07	6.28	0.29	6.24	0.36	14
1913 03 06	1103	30.0 83.0 G	6.55	0.23	6.50	0.26	09	6.52	0.32	6.49	0.30	18
1913 03 18	0120	33.0 91.0 R	6.18	0.22	6.25	0.32	04	5.95	0.16	5.98	0.18	09
1914 10 09	0239	32.8 75.3 m	6.39	0.33	6.23	0.28	06	0	0	0	0	0
1915 04 28	0319	33.5 93.0 R	6.16	0.38	6.05	0.23	04	0	0	0	0	0
1915 12 03	0239	31.0 93.0 R	6.74	0.29	6.65	0.28	07	0	0	0	0	0
$M < 6.0$												
1902 06 16	0136	30.0 79.0 m	6.09	0	5.93	0	01	6.04	0.20	5.88	0.28	04
1903 12 23	0300	29.5 67.5 m	0	0	0	0	0	5.90	0.12	5.70	0.16	05
1904 07 27	0520	33.0 72.0 B	5.79	0.11	5.74	0.04	02	6.03	0.23	5.89	0.30	11
1907 07 12	1720	25.0 70.0 m	5.29	0.15	5.26	0.11	04	5.60	0.36	5.43	0.42	05
1908 01 12	1019	30.2 67.7 m	5.63	0.06	5.58	0.08	03	5.88	0.33	5.67	0.39	06
1908 04 04	0618	25.3 92.6 m	5.88	0.15	5.87	0.10	06	6.00	0.32	5.92	0.39	06
1909 09 07	1528	33.0 70.0 B	5.99	0.27	5.99	0.11	06	5.91	0.21	5.81	0.22	08
1910 08 13	2119	28.0 90.0 B	5.39	0	5.47	0	01	5.53	0.22	5.41	0.36	07
1912 11 01	1900	29.0 67.0 m	5.77	0	5.45	0	01	5.75	0.24	5.75	0.24	03
1913 03 27	0913	29.5 67.5 m	5.52	0	5.63	0	01	0	0	0	0	0
1913 03 27	0815	26.5 66.5 m	5.47	0.02	5.44	0.21	02	0	0	0	0	0
1913 05 14	0850	34.5 69.2 m	4.96	0	5.07	0	01	0	0	0	0	0
1913 06 26	2330	31.0 77.0 m	4.48	0	4.59	0	01	0	0	0	0	0
1914 02 06	1142	29.0 65.0 m	5.63	0.28	5.70	0.30	14	0	0	0	0	0
1914 04 30	2200	30.0 74.0 m	5.43	0	5.19	0	01	0	0	0	0	0
1914 05 21	0826	32.0 69.5 B	5.72	0.26	5.69	0.21	01	0	0	0	0	0
1914 06 17	1700	27.0 94.0 m	5.18	0	4.94	0	01	0	0	0	0	0
1914 11 04	1106	32.0 70.0 m	5.86	0.38	5.72	0.28	06	0	0	0	0	0
1915 03 03	0145	32.0 73.0 m	5.47	0.06	5.19	0.12	02	0	0	0	0	0
1915 05 05	1512	30.0 84.0 R	6.10	0.37	5.98	0.21	04	0	0	0	0	0

B, BAAS estimates; G, Gutenberg and Richter[61]; m, macroseismic locations; R, re-assessed in this study.

comparison of M_{SC} with M_M in Table 3, shown in Figure 2, confirms the good agreement between the two methods.

Abe's[52] equivalent surface-wave magnitude M_A, eq. (11), also may be used. It has been derived from large events $M_S \geq 7.0$ recorded at large distances and for $M_M > 6.0$, as Figure 3 shows, it gives values which are almost identical with those from eq. (10). However, for smaller events, usually recorded at shorter distances ($D < 30°$), eq. (11) underestimates M_M systematically by 0.1 to 0.3 magnitude units. Note that eq. (3) used in this paper to assess surface-wave magnitudes is the relationship used by ISC/NEIC to estimate surface-wave magnitude and that eqs (10) and (11) are equivalent relations which have been calibrated against eq. (3).

Epicentral distance

Station epicentral distances D are either from macroseismic locations or from instrumental determinations adopted from various sources or, for a few events, recomputed in this study using standard ISC procedures with the available phase input data taken from bulletins. With the exception of macroseismically well-documented events, instrumental locations appear to be accurate to a few degrees, and in a few cases to more than 10°, which is enough to allow estimation of the size of an event with acceptable error. This is because majority of the observing stations are at epicentral distances of more than 40° at which an error of 5 to 10° would correspond only to an uncertainty to not more than 0.3 units in station magnitude.

Moment magnitude

There is some confusion in the literature about the definition and use of seismic energy magnitude M_W, moment magnitude **M**, and surface-wave magnitude M_S. Seismic energy magnitude M_W is defined as a linear tranformation of the logarithm of the seismic moment M_0 given by:

$$M_S \sim M_W \Leftarrow (2/3)\log(M_0) - 10.73, \qquad (12)$$

in which M_0 is in dyn.cm units (10^{-7} Nm)[53]. Kanamori[53] derived eq. (12) from the observation that in most large, $M_S \geq 7.5$, shallow earthquakes the stress-drop is about 30 bars, which he combined with the energy (E) and magnitude (M_S) relation for earthquakes in California, i.e. $\log E = 11.8 + 1.5M_S$ which in reverse form, is similar to eq. (12).

Moment magnitude **M** for shallow earthquakes in California in the range $5.0 \leq M_S \leq 7.5$, was then defined[54] as being equal to M_W from eq. (12). However, equality $\mathbf{M} = M_W = M_S$, as defined above, holds only for events that rupture the entire thickness of the seismogenic zone and its validity, therefore is regionally dependent[55]. **M** is nothing more than a definition or a transformation of M_0 through eq. (12) and for the region of our interest, $\mathbf{M} \neq M_S$ for $M_S < 6.0$. For the sake of clarity we use **M** and M_S in this work.

Relations between surface-wave magnitude M_S and seismic moment M_0, and vice versa, provide suitable functions for the correlation between one source size indicator and the other. The current relationships for assessing M_0 from the surface-wave magnitude M_S of shallow earthquakes have been derived from global or large sub-global data sets for active regions[40,55,56] and for stable continental regions[57,58].

Ekström and Dziewonski[55] derived global average relationships between M_S and $\log M_0$, in which the independ-

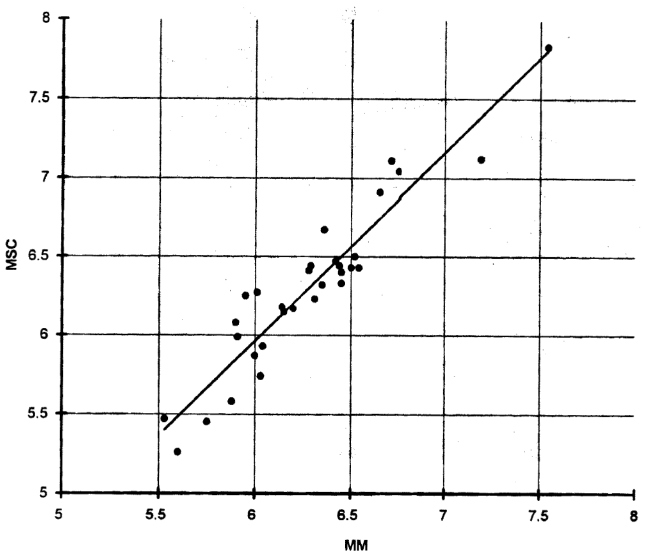

Figure 2. Comparison of magnitudes M_{SC}, from the modified Prague formula with equivalent surface-wave magnitude M_M, from Milne instruments (eq. (10)), for events recorded by both types of instruments.

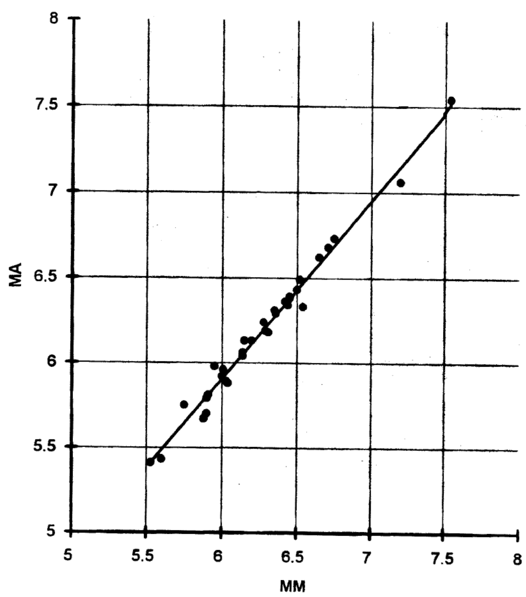

Figure 3. Comparison of equivalent surface-wave magnitudes M_M from eq. (10), with M_A, from Abe's equation (eq. (11)).

ent variable is log M_0. They used 2341 reported M_0 values from the Preliminary Determination of Epicentres (PDE), and corresponding scalar moments from the Harvard CMT catalogue. Only events up to 1987, for which both the NEIC and the CMT depths are < 50 km, in the log(M_0) range 23.5 to 28.6 were considered. A relationship was then determined in the form:

$$M_S = k - (a + b)/6 + \log M_0 \quad \text{for} \quad \log M_0 < a, \qquad (13)$$

$$M_S = k - (a + b)/6 + \log M_0 - (\log M_0 - a)^{2/6}(b - a) \quad \text{for} \quad a \le \log M_0 \le b, \qquad (14)$$

$$M_S = k + (2/3)\log M_0 \quad \text{for} \quad \log M_0 > b. \qquad (15)$$

Note that eq. (13) was derived on the assumption that the slope of the regression is one for log $M_0 < a$, and eq. (14) on the assumption that the slope is 2/3 for log $M_0 > b$. The constants in eqs (13) to (15) were determined by minimizing

$$\sum_{i=1}^{N} [M_S(\log M_{oi};\, a, b, k) - M_{Si}]^2,$$

with respect to, a, b and k.

Rather than summing over N earthquakes, a reduced data set was used in which M_S was averaged for earthquakes in narrow bins of log M_0 of 0.1 units, so that only about 40 summary data points were considered.

A good fit to the reduced data for earthquakes with moment as the independent variable in the range 2×10^{24} to 10^{28} dyn.cm was obtained with $a = 24.5$, $b = 26.4$ and $k = -10.76$, which reduces eqs (13)–(15) to:

$$M_S = -19.24 + \log M_0 \quad \text{for} \quad \log M_0 < 24.5, \qquad (16)$$

$$M_S = -19.24 + \log M_0 - 0.088\,(\log M_0 - a)^2 \quad \text{for} \quad 24.5 \le \log M_0 \le 26.4, \qquad (17)$$

$$M_S = -10.76 + (2/3)\log M_0 \quad \text{for} \quad \log M_0 > 26.4. \qquad (18)$$

These authors then rewrite eqs (16) to (18) in the form:

$$\log M_0 = 19.24 + M_S \quad \text{for} \quad M_S < 5.3, \qquad (19)$$

$$\log M_0 = 30.20 - [92.45 - 11.40 M_S]^{0.5} \quad \text{for} \quad 5.3 \le M_S \le 6.8, \qquad (20)$$

$$\log M_0 = 16.14 + 1.5 M_S \quad \text{for} \quad M_S > 6.8. \qquad (21)$$

However, since eqs (16) to (21) and eqs (16) to (18) are rewritten, formally, they are not the correct relationships for estimating log M_0 from M_S.

Regional bias in M_0 does exist and global average moment–magnitude relationships, such as eqs (16) to (21), may be inappropriate for the assessment of long-term seismic slip on faults. It also will not hold for the estimation of tectonic motion in regions, the rate of which is known from GPS measurements, and for the investigation of aseismic creep. Continental data show that the transition from a slope of unity to a larger value occurs at larger moments for continental events[55,59] for which

$$\log(M_0) = 19.24 + M_S \quad \text{for} \quad M_S < 7.16, \qquad (22)$$

$$\log(M_0) = 15.66 + 1.5 M_S \quad \text{for} \quad M_S \ge 7.16. \qquad (23)$$

To overcome the problem with eqs (19) to (21) which have been derived by fitting the data with log(M_0) as the independent variable, and at the same time to take into account the regional bias in eqs (16) to (21), we derived the following set of bi-linear relationships

$$\log(M_0) = 19.08 + M_S \quad \text{for} \quad M_S \le 6.0, \qquad (24)$$

$$\log(M_0) = 16.07 + 1.5 M_S \quad \text{for} \quad M_S > 6.0, \qquad (25)$$

for the eastern Mediterranean and the Middle East region up to 70°E, using CMT or P/SH moments and the corresponding uniformly reassessed M_S values of 577 shallow (h < 40 km) earthquakes, in the log M_0 range 22.4 to 27.3, in which M_S is the independent variable.

Discussion

The earthquakes shown in Figure 1 are widely scattered and although there is great variation in the accuracy of location, they delineate the main seismically active areas now established by modern seismological studies, as well as provide evidence for destructive earthquakes in less active regions. We may discuss a few cases, which demonstrate the procedures we used to locate or relocate events, assess their magnitude and estimate their moment magnitude.

Location

An example of rough location is provided by the earthquake of 3 December 1903. It was located by BAAS[18] north of Lhasa in Tibet at 32.0°N, 93.0°E. The earthquake was not well recorded and not even a rough instrumental solution is possible. However, we know that this shock, caused panic at Shwagu (19.5°N, 95.0°E) and was felt in the Bhano district in Burma, 1400 km south of the location given by BAAS[18]. For an adopted location in Burma the magnitude of the shock is 6.5 ± 0.35.

Occasionally, confusion of station readings or of felt reports belonging to two separate events, closely spaced in time results in mislocations. The earthquake of 17 February 1905 at 11 h 42 m GMT was well-recorded instrumentally, but macroseismic information which could help constrain the solution is totally lacking. Using maximum phase readings from Milne instruments worldwide and body phases from European and Russian stations, a rough solution puts the event at 30°N, 95°E, in south-east Tibet, and gives a magnitude of M_S 7.1. This location is supported by the large surface-wave amplitudes recorded exclusively by near Indian and Russian stations that confirm a general position close to that estimated by BAAS[18], at 26.0°N, 96.0°E, in northern Burma from where, however, we can find no felt reports. Another, very different location[13] places this event 5400 km east of our position in the Pacific, 1100 km east of the coast of Japan where a separate shock had been felt at 09 h GMT. It is probable that the epicentral location by Szirtes[13], east of Japan was influenced by the felt reports from the East Coast of Honshu[7].

For the earthquake of 26 September 1905 (M_S = 6.4) we have relatively good macroseismic and instrumental data which place the event in the Sulaiman mountains near 30.3°N, 69.9°E, but 300 and 440 km west of the locations reported by BAAS[17] and Szirtes[13] respectively. Duda[60] assigns a magnitude of 7.1 to this event for which we can find no justification.

Instrumental locations, not aided by good macroseismic information, are unreliable as the following case shows. The earthquake of 15 August 1906 at 22 h 10 m (GMT) was relatively well-recorded and BAAS[18] places its epicentre in Turkestan (sic), at 44°N, 95°E, where it is reported as felt. It is very probable that these felt reports belong to the earthquake in Kazakhstan of 13 August at 18 h 45 m (GMT), which was widely felt north-east of Alma Ata (44.5°N, 79.5°E) and that BAAS was misled in choosing a location in the Altai Mountains in Mongolia. This is because the earthquake of 15 August (M_S 6.1) occurred in the Rann of Kutchch, 3000 km south of the BAAS (1913) location, a region which showed some activity during that period. The shock was widely felt on 16 August at 02 h 45 m (local time), in the districts of Rajputana and Gujarat and around the Bay of Cambay, Jodhpur, Ahmedabad and Mirpur where it lasted several seconds. On 12 July 1907 another earthquake occurred in the same region which was felt at Ahmedabad, Jodhpur and in the Sind. The location given by BAAS[18] for this second event, most probably adopted from macroseismic information is close to these sites.

Depth

It is obvious that instrumental data of this early period are totally insufficient to allow depth determination, except indirectly when combined with good macroseismic information, in which case it is possible, sometimes to differentiate between shallow and deeper shocks. For example, the earthquake of 6 February 1914 was strong at Chagai and Kharan in Baluchistan, near 29.0°N, 65.0°E. There is no evidence that it was felt very far, not as far as Quetta 200 km away. The shock was well-recorded and its surface-wave magnitude calculated from 14 stations is 5.7. Gutenberg in his work-sheets gives an epicentral location close by the macroseismic but assign to it a depth of 100 km and a long-period body-wave magnitude of 6.8, which is rounded off to 7.0 (ref. 61). The large number of station readings of relatively small surface-wave amplitudes, and the lack of known aftershocks, suggest an event in the lower crust.

The earthquake of 29 March 1907 in Badakhshan, occurred north-east of Kabul, in the time of a deep earthquake in the Celebes, 6400 km east of Kabul, to which Gutenberg and Richter[61] assign a depth of 500 km and a body wave magnitude m_b of 7.3. For this earthquake although details are lacking, we know that the shock was widely felt in the region of Kabul and it was perceptible as far as Shimla. Maximum phases recorded at Indian and European stations suggest a subcrustal event north-east of Kabul. Its magnitude, assuming a crustal depth, is M_S 6.1.

Macroseismic and instrumental information about the earthquake of 23 August 1912 at 14 h 02 s is sufficient to locate it in the Kohat Urakzai region in north-west Pakistan near 33.5°N, 71.0°E, where it caused damage and loss of life. A second shock seven hours later, at 21 h 14 h, which was also well recorded, added to the damage. Gutenberg and Richter[61] do not mention the first shock and locate the second shock at 36.5°N, 70.5°E, to which they assign a depth of 200 km. However, well-developed surface waves at all stations, the small area over which the shock was felt compared with its magnitude (M_S 6.3) and its aftershock sequence, suggest a crustal event.

Magnitude

For the early period, the Chaman earthquake of 20 December 1892 in Baluchistan offers a good example of the use of Abe's method[52] to estimate magnitude from trace amplitudes recorded by undamped instruments other than Milne. It is also the earliest event for which we have scanty, but good instrumental data. The earthquake was recorded by Rebeur-Paschwitz undamped seismographs at Nikolaiev and Strassburg[62,63]. Using Abe's method and the trace amplitudes from the two stations, we calculate a magnitude of 6.7, which is considerably larger than the value of 6.2 assessed by Abe. As a check, we calculated M_S from the length and slip on the observed surface fault break which was associated with this event using the empirical formula[64], and got a value of M_S 6.8, same as what is calculated from instrumental data.

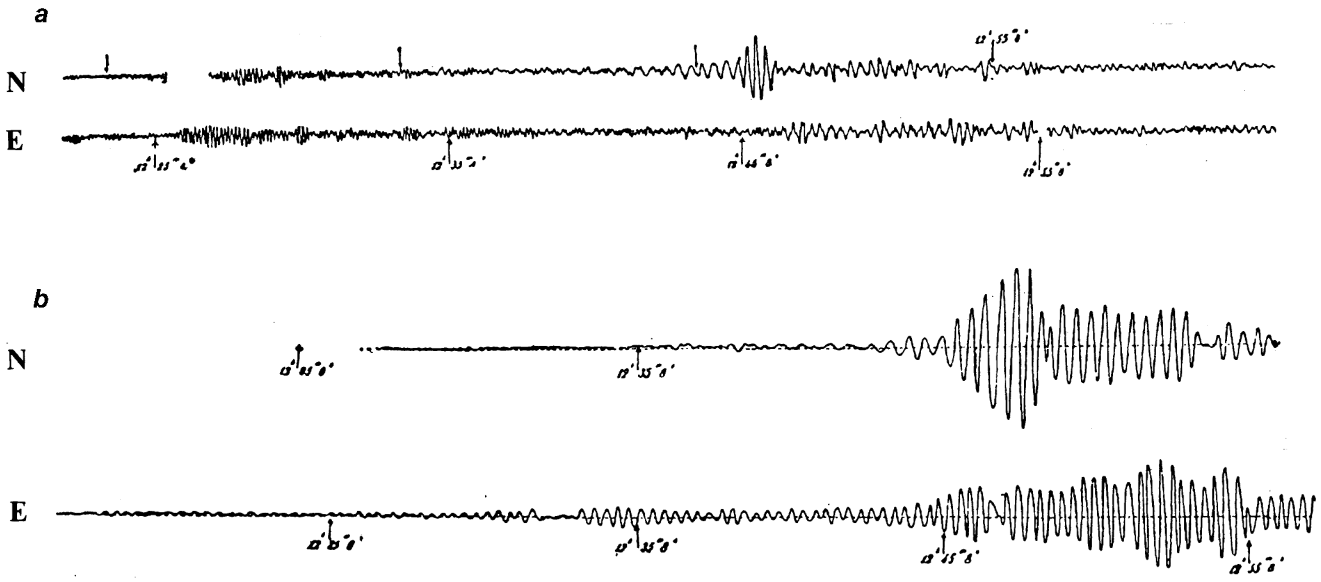

Figure 4. Seismograms of the 12 June 1897 Assam earthquake recorded at Rocca di Papa near Rome (Ä = 65°). ***a***, N-S and E-W motions obtained from a standard pendulum of period 8 s and magnification 12.5; ***b***, N-S and E-W motions recorded by a 22 s period pendulum (for details see ref. 66).

For the earthquake of 12 June 1897 in Assam there are more data from early seismographs than for the Baluchistan earthquake. These come from six instruments in Italy[65,66] and Figure 4, one from Russia[67] and two from the UK[17]. Using again Abe's method and the recorded trace amplitudes, we find that the magnitude of the event is M_S 8.0 (± 0.15). This value is very close estimates of 8.2 (ref. 68) and 8.0 (ref. 52) and to the unified magnitude m 8.0, calculated by Gutenberg[69]. However, Richter[70] inflated this value to 8.7 by converting Gutenberg's unified magnitude m into M_S through the empirical relation: $M_S = 1.59\ m - 4.0$, a relation derived for California for M_S estimates based on surface waves with periods close to 20 s. Richter's estimate 8.7, which has been adopted by later authors, places the Assam earthquake among the largest known shallow events, which in spite of its very large size left no vestiges of surface faulting.

There is also a great difference in the size of the Kangra earthquake of 4 April 1905 estimated by different authors. Its surface-wave magnitude has been assessed between 7.5 and 8.6, and a discussion of the problems associated with these estimates is given in Ambraseys and Bilham[71]. We reappraised the magnitude of this event by two different methods and sets of instrumental data. In the first method we calculated M_M from 19 Milne stations, which give M_M = 7.54 (± 0.23). In the second method we calculated M_S from 6 station operating damped seismographs, which give 7.8 (± 0.05), which we believe to be the magnitude of this event.

Moment

Magnitude-moment scaling laws are different for small events for which $\log(M_0)$ and M_S have a 1 : 1 relation. For

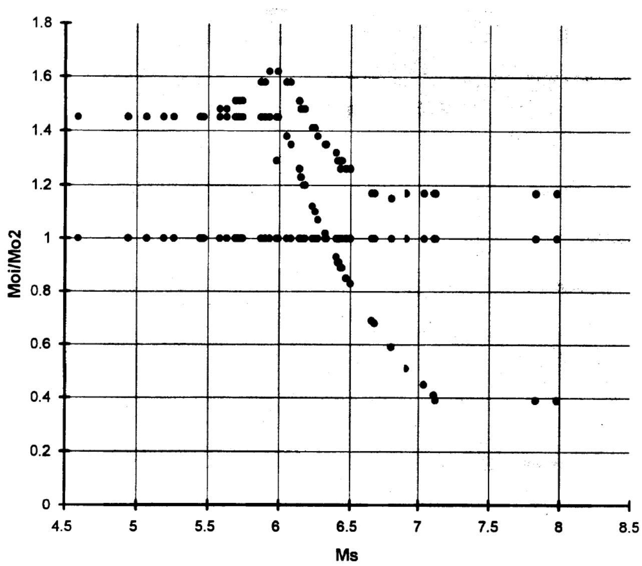

Figure 5. Ratios of the moments derived from eqs (19)–(21) M_{01} and eqs (22)–(23) M_{03} with respect to M_{02} from eqs (24) to (25) as a function of magnitude M_S for the earthquakes in Table 3. The hump of the plot for M_{01}, near M_S 6.0, is due to eq. (20).

earthquakes of intermediate size that rupture the whole depth of the seismogenic layer the ratio decreases to 1 : 1.5, perhaps reaching smaller values for very large earthquakes. For earthquakes in the eastern Mediterranean and Middle East, this transition occurs at M_S 6.0 compared to 6.6 in the case of global average and 7.2 as in the case of global continental earthquakes. For the estimation of tectonic motions, the assessment of strain release and for the scaling of attenuation laws in terms of M, regardless of these uncertainties, the fact remains that a linear log (M_0) – M_S relation with a 1 : 1.5 slope should not be used throughout the magnitude range. For this region, which includes only a small part of our study area, the transition from a slope of unity to a larger value occurs not at a larger but at a much smaller magnitude than for the average global or the continental data sets used by Ekström and Dziewonski[55].

Figure 5 shows the seismic moment ratio M_{01}/M_{02}, in which M_{01} is derived from eqs (18) to (21), and M_{02} from eqs (22) and (23), plotted against the surface wave magnitude M_S of the earthquakes in Table 4. We notice that the global relations eqs (19)–(21) overestimate M_{02} for the eastern Mediterranean and Middle East regions, by a factor between 1.4 and 1.6 for small events and by 1.2 for large earthquakes. Also, the continental relations (eqs (22)–(23)) overestimate M_{02} for small events but they underestimate it by as much as a factor of 0.4 for larger shocks.

Taking as an example the events with $M_S \geq 6.0$ in Table 4, the global relations (eqs (19)–(21)) give a total moment, which is 2.8 times greater than that from the continental relations (eqs (22)–(23)) and 1.2 times larger than the Middle East relations (eqs (24)–(25)). If we exclude the two large earthquakes of 1897 and 1905, the total moment from eqs (19) to (21) is twice that from

Table 4. Seismic moments and moment magnitudes 1892–1915

	Date	OT	Epicentre	M_{SC}	log(M_{01})	log(M_{02})	$\mathbf{M}_1$	$\mathbf{M}_2$
1	1892 12 20	0020	30.9 66.5 m	6.79	26.32	26.26	6.82	6.78
2	1897 06 12	1106	25.5 91.0 m	7.98	28.11	28.04	8.01	7.96
3	1901 10 17	0557	30.5 68.5 m	6.14	25.46	25.28	6.24	6.12
4	1901 11 18	0004	32.0 77.0 B	6.23	25.57	25.42	6.32	6.21
5	1902 06 16	0136	30.0 79.0 m	5.93	25.22	25.01	6.08	5.94
6	1902 11 04	1133	32.0 91.0 B	6.47	25.88	25.78	6.52	6.46
7	1902 12 13	1708	30.0 85.0 B	6.67	26.15	26.08	6.70	6.66
8	1903 12 03	2126	19.5 95.0 m	6.43	25.82	25.72	6.48	6.42
9	1903 12 23	0300	29.5 67.5 m	5.90	25.18	24.98	6.06	5.92
10	1904 03 31	0216	31.0 89.0 B	6.91	26.51	26.44	6.94	6.90
11	1904 03 31	0545	31.0 89.0 B	6.17	25.50	25.33	6.27	6.16
12	1904 07 27	0520	33.0 72.0 B	5.74	25.00	24.82	5.94	5.82
13	1905 02 17	1142	30.0 95.0 B	7.11	26.81	26.74	7.14	7.10
14	1905 04 04	0050	32.1 76.4 m	7.83	27.89	27.82	7.86	7.82
15	1905 09 26	0128	30.3 69.9 m	6.44	25.84	25.73	6.50	6.42
16	1906 02 27	1940	31.5 77.5 m	6.40	25.79	25.67	6.46	6.38
17	1906 05 12	0550	28.0 92.0 B	6.44	25.84	25.73	6.50	6.42
18	1906 08 15	2211	25.0 71.0 m	6.08	25.39	25.19	6.20	6.06
19	1907 03 29	2053	35.0 70.0 m	6.15	25.47	25.30	6.25	6.14
20	1907 07 12	1720	25.0 70.0 m	5.26	24.50	24.34	5.60	5.50
21	1908 01 12	1019	30.2 67.7 m	5.58	24.83	24.66	5.82	5.71
22	1908 03 05	0220	30.2 67.7 m	6.43	25.82	25.72	6.48	6.42
23	1908 04 04	0618	25.3 92.6 m	5.87	25.15	24.95	6.04	5.90
24	1908 06 03	1556	28.0 67.0 B	6.18	25.51	25.34	6.28	6.16
25	1908 08 20	0953	32.0 89.0 B	7.04	26.70	26.63	7.07	7.02
26	1909 09 07	1528	33.0 70.0 B	5.99	25.28	25.07	6.12	5.98
27	1909 10 20	2341	28.9 68.3 m	7.12	26.82	26.75	7.15	7.10
28	1910 08 13	2119	28.0 90.0 B	5.47	24.71	24.55	5.74	5.64
29	1910 08 17	1158	27.0 67.0 m	6.33	25.70	25.57	6.40	6.32
30	1911 10 14	2324	31.0 80.5 G	6.42	25.81	25.70	6.48	6.40
31	1912 08 23	1402	33.5 71.0 m	6.32	25.68	25.55	6.39	6.30
32	1912 08 23	2114	33.5 71.0 m	6.27	25.62	25.48	6.35	6.26
33	1912 11 01	1900	29.0 67.0 m	5.45	24.69	24.53	5.73	5.62
34	1913 03 06	0208	30.0 85.0 B	6.41	25.80	25.69	6.47	6.40
35	1913 03 06	1103	30.0 83.0 G	6.50	25.92	25.82	6.55	6.48
36	1913 03 18	0120	33.0 91.0 R	6.25	25.60	25.45	6.34	6.24
37	1913 03 27	0913	29.5 67.5 m	5.63	24.88	24.71	5.86	5.74
38	1913 03 27	0815	26.5 66.5 m	5.44	24.68	24.52	5.72	5.62
39	1913 05 14	0850	34.5 69.2 m	5.07	24.31	24.15	5.48	5.37
40	1913 06 26	2330	31.0 77.0 m	4.59	23.83	23.67	5.16	5.05
41	1914 02 06	1142	29.0 65.0 m	5.70	24.96	24.78	5.91	5.79
42	1914 04 30	2200	30.0 74.0 m	5.19	24.43	24.27	5.56	5.45
43	1914 05 21	0826	32.0 69.5 B	5.69	24.95	24.77	5.90	5.78
44	1914 06 17	1700	27.0 94.0 m	4.94	24.18	24.02	5.38	5.28
45	1914 10 09	0239	32.8 75.3 m	6.23	25.57	25.42	6.32	6.22
46	1914 11 04	1106	32.0 70.0 m	5.72	24.98	24.80	5.92	5.80
47	1915 03 03	0145	32.0 73.0 m	5.19	24.43	24.27	5.56	5.45
48	1915 04 28	0319	33.5 93.0 R	6.05	25.35	25.15	6.17	6.04
49	1915 05 05	1512	30.0 84.0 R	5.98	25.27	25.04	6.12	5.96
50	1915 12 03	0239	31.0 93.0 R	6.65	26.12	26.05	6.68	6.64

M_{01} from eqs (19) to (21); M_{02} from eqs (22) to (23); $\mathbf{M}_1$ moment magnitude (eq. (12)) from eqs (19) to (21); $\mathbf{M}_2$ moment magnitude (eq. (12)) from eqs (22) to (23).

eqs (22) to (23) and 1.2 times larger than that from eqs (24) to (25). It is difficult to estimate realistic errors in velocities or slip rates from seismicity and M_0, given the uncertainty in the original M_S values and the bias in the different $M_S - \log(M_0)$ relations used. We consider it likely that the global relations (eqs (19)–(21)) yield an upper bound to the seismic moment release and that the regional relations (eqs (24)–(25)) yield a more realistic estimate.

However, in estimating total moment the contribution made by the more numerous small events to the total moment release, invariably omitted, must be included in the summation of scalar seismic moments. These smaller events, because of the incompleteness of the data below a certain magnitude are usually not accounted for in the summation. How considerable their contribution can be will depend chiefly on the choice of the log $(M_0) - M_S$ relation. For bi-linear or nonlinear relations and for narrow magnitude ranges of summation, their contribution can be considerable: it can be as large as the moment contributed by the large events[72]. This increase in total moment together with long-term observations fits better with direct measurements of velocity fields derived from space-based geodetic methods and also alters the current view about regions of high aseismic deformation[71].

Conclusions

In establishing seismicity for tectonic understanding and earthquake hazard evaluation, it is important to extend the record as far back in time as possible. By combining instrumental and macroseismic information, we assessed the location and assigned uniformly calculated surface-wave magnitudes and seismic moments to 50 shallow earthquakes in northern India and Pakistan in the early period of instrumental recording from 1892 to 1915, both years inclusive. From Table 2 we can see how many earthquakes have been missed out in modern parametric catalogues and by how much individual magnitude estimates differ. Of the 31 earthquakes of $M_S \geq 6.0$ we identified, only six are listed in Gutenberg and Richter[61], which demonstrates how incomplete the earthquake record is for this early period.

The most widely accepted measure of earthquake size is magnitude, derived from instrumental measurements. Many different types of magnitudes have been developed, depending on the type of instrument used and parameter measured, serving different purposes. Of the various magnitude scales currently in use, the main one for the study of tectonics and seismic hazard from historical data is the surface-wave magnitude M_S, expressed as moment magnitude M for comparison with modern events.

The new data we present here give a small and incomplete example of the importance of the study of past earthquakes for the unambiguous evaluation of tectonics and of and of seismic hazard in this part of the world. There is a wealth of macroseismic and instrumental information for the first half of this century in various local repositories and little is known to have been done to retrieve and distil it for this part of northern India and neighbouring areas. Although not dealt with here, our investigations show that for the first half of this century the completeness of the earthquake record of India is low and needs improvement.

Any advancement of our knowledge about the assessment and mitigation of earthquake hazard should be accompanied by a growth in our accumulation of reliable observational data and field information from past earthquakes. One feels that much effort is being diverted to solving numerical problems on guessed input parameters and that more data from observations are now required.

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