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SEISMIC MOMENT - MAGNITUDE RELATIONSHIP FOR THE GARHWAL HIMALAYA REGION

M. L. SHARMA[1] AND H. R. WASON [2]

1. Earthquake Engineering Department, University of Roorkee, INDIA

2. Earth Sciences Department, University of Roorkee, INDIA

ABSTRACT

The digital seismic data recorded by a telemetered seismic array have been used to compute seismic moment-magnitude relationship in the Garhwal Himalaya region. The seismic moments for 18 shallow focus earthquakes are computed using Brune's model (Brune, 1970) of earthquake source. The seismic moments computed range from 7.0 x 10^{18} dyne-cm to 6.23 x 10^{21} dyne-cm. The earthquakes are observed to have very low stress drops ; less than 1 bar in case of seven events and between 1 bar and 10 bars for another seven events. The maximum stress drop is found to be 38 bars for an earthquake of magnitude 4.2 with focal depth 15 km. The empirical relationship between seismic moment and local magnitude is computed using regression analysis. This relationship for Garhwal Himalaya region is obtained as $\log(M_0) = (0.89 \pm 0.43)\, M_L + (18.18 \pm 0.21)$. The shallow focus events having low stress drops in this region point to the low strength of the crust to withstand the accumulated stresses.

INTRODUCTION

Earthquakes are believed to originate from spontaneous slippage along planes of weakness, i.e., faults after elastic strain accumulation over a long period of time. Faults may be considered as the slip surfaces across which discontinuous displacement occurs, and the faulting process may be modeled mathematically as a shear dislocation in an elastic medium which is equivalent to a double-couple body force. The scaling parameter of each component couple of a double couple body force is its moment. Hanks and Wyss (1972) gave the procedure to determine the seismic moment of an earthquake using body wave spectra based on dislocation theory . Brune(1970) developed a powerful theory describing the nature of seismic spectrum radiated from the seismic source by considering the physical process of the energy release. The source model relates the corner frequency and low frequency asymptote to seismic moment along with other source parameters. In addition to the studies of major and moderate earthquakes several observational studies have used the corner frequency and low frequency asymptote of microearthquakes to determine their seismic moments, faulting dimensions and stress drops (Archuleta et al., 1982; Fletcher, 1980 ;Tucker and Brune, 1977 and O'Neill, 1984). A general result of these investigations is that the scaling behaviour of small events ($M \leq 3$) differs somewhat from that of larger events, signifying a dissimilarity in the rupture processes (Frankel and Wennerberg, 1989). Specifically, the corner frequencies of the smaller events are observed either to increase very slowly or remain unchanged with decrease in seismic moment. This is interpreted to signify that the faulting dimensions of small events are relatively uniform as seismic moment falls below 10^{20} dyne-cm so that stress drop decreases with decreasing seismic moment (Archuleta et al., 1982 ; Aki, 1984 and Papageorgiou and Aki, 1983). In contrast, studies of the spectra of larger events with 3.5 $\leq$ M $\leq$ 6 find no variations in the stress drop with decrease in seismic moment (Thatcher and Hanks, 1973). Observations of spectra consistent with Brune's theory have been reported for microearthquakes with very low stress drops (Douglas and Ryall, 1972) as well as high stress drops in addition to large earthquakes (Hanks and Wyss, 1972 and Wyss and Hanks, 1972), thus covering a very wide range of computed values for source parameters based on this theory.

In the present study, the digital data pertaining to 18 local seismic events acquired during the period Feb. 13, 1989 to May 7, 1989 by a Telemetered Digital Seismic Array (TDSA) is used to compute the source parameters and to obtain their empirical relationships with local magnitude. TDSA have five vertical component and two 3-component stations (Fig. 1). The digital bit stream from each seismometer station is multiplexed into a computer digital interface by a control module which allows direct entry of digital data into the data acquisition system (Wason et al., 1986). Omega timing signals from La Reunion transmitting station are added to provide a common time base for all the seismometer stations. The triggered event data are acquired by use of an event detection technique (Sharma, 1992a). The triggered events are checked for seismic events through visual display and also by using automatic phase pickers (Sharma, 1992b).

All the stations of the array are situated in the Lesser Himalaya region except the station at Roorkee, which is located in the alluvial plains of the Ganges. The Lesser Himalaya zone consists of the sediments of the Precambrian-Paleozoic and locally Mesozoic age, metamorphosed and subdivided by thrusts with progressively older rocks towards the north. The Lesser Himalaya has been thrusted southwards over the Siwaliks of the Sub-Himalaya along the Main Boundary Thrust. The northern boundary of the Lesser Himalaya is defined by the Main Central Thrust which separates it from the Higher Himalaya.

DATA ANALYSIS

The hypocentral locations of the seismic events are computed using the HYPO71 computer program developed by Lee and Lahr (1975). The velocity of P-wave is taken to be 5.2 km/s for an upper 15 km thick layer and 6 km/s for the underlying half space (Kumar et al., 1987). The hypocentral parameters determined using the P- and S- wave arrival times from three or more stations are tabulated in Table I. In order to determine the magnitudes of earthquakes the recorded amplitudes are adjusted to match the response for a Standard Wood-Anderson seismograph and attenuation laws given by Richter (1935) have been used. The local magnitudes so computed are tabulated in Table I.

A time window of 2.84 s (sampling rate 90 samples/s) is applied from the onset of the P-wave to compute P wave spectra (Sharma and Wason, 1994). This has been done so that the energy from the P waves is only considered for the spectra and the S-wave effect is excluded. After removal of the DC bias, the sample window is cosine tapered with 10 percent tapering at both ends. The FFT algorithm by Cooley and Tukey has been used to compute Fourier spectra (Claerbout, 1985). After conversion of velocity to displacement

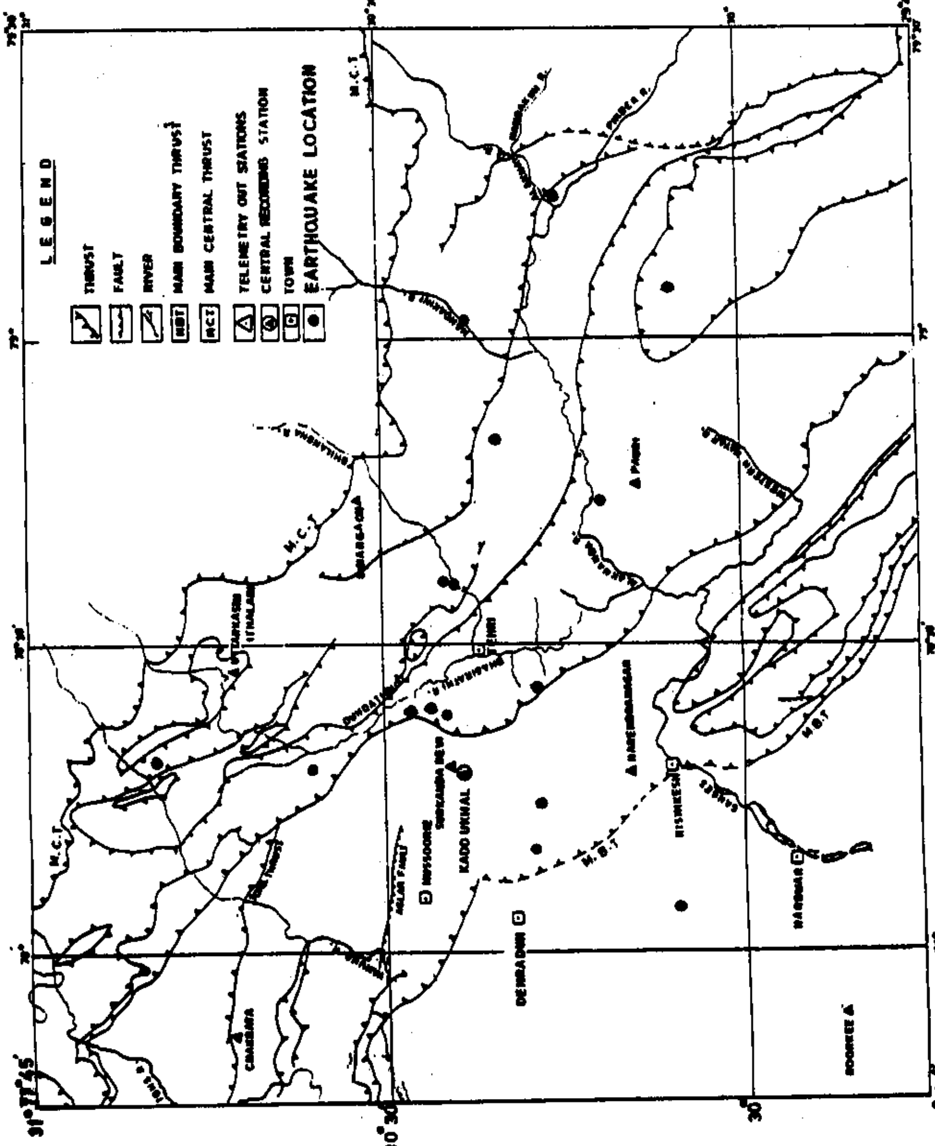

Fig. 1 Tectonic map (After Fuchs and Sinha, 1978) of the area showing locations of recording stations

in the Fourier domain, the spectra is corrected for the attenuation by taking an average value of 300 for the quality factor, Q. This value is an approximation to the attenuation characteristics of this region as there has been no reliable estimation of Q. The geometrical spreading correction is applied to take into account the epicentral distance of the station to the source. Fourier spectra thus calculated are indicated in Fig. 2.

TABLE I HYPOCENTRAL PARAMETERS AND MAGNITUDE OF THE SEISMIC EVENTS

Sl. No.	Date	Origin Time H M Sec	Latitude Deg	Latitude Min	Longitude Deg	Longitude Min	Focal depth (Km)	Magnitude M_L
1	25 11 88	05:36:56.5	31	58.0	78	53.8	15.4	4.2
2	08 03 89	11:36:39.7	30	26.	78	23.9	6.2	2.01
3	16 03 89	22:06:35.0	30	28.0	78	23.8	10.1	3.4
4	29 03 89	10:59:15.1	30	25.0	78	35.9	10.0	2.4
5	30 03 89	03:04:44.7	30	25.0	78	23.7	2.9	2.7
6	02 04 89	10:38:57.9	30	16.6	79	14.3	10.0	3.1
7	04 04 89	01:12:40.8	30	16.1	78	35.3	24.0	2.9
8	09 04 89	08:03:41.3	30	24.0	79	01.5	20.0	2.5
9	11 04 89	09:36:35.9	30	25.0	78	35.3	10.8	1.8
10	11 04 89	10:55:42.8	30	18.8	78	19.6	13.6	3.5
11	12 04 89	17:39:20.2	30	49.3	78	18.9	16.0	3.0
12	16 04 89	02:07:59.0	30	19.7	78	44.9	11.4	2.8
13	19 04 89	04:38:54.3	30	06.7	79	05.1	4.1	2.9
14	22 04 89	21:40:37.4	31	05.2	78	04.6	16.1	3.0
15	23 04 89	11:42:32.5	30	33.9	78	25.8	14.9	1.4
16	03 05 89	11:26:52.8	30	21.8	79	44.6	10.0	1.9
17	03 05 89	21:15:23.8	30	20.6	78	49.6	10.0	1.6
18	04 05 89	01:33:27.8	30	36.4	78	18.0	1.0	1.6

The frequency response of the Butterworth low pass filter has been modified to include the effects of attenuation for estimating the spectral parameters, i.e., f_0 and Ω_0, for the analytical model (Johnson and McEvilly, 1974). This has the form

$$\Omega_o[1+f/f_o]\exp^{(-\pi f \tau)} \qquad (1)$$

where f_0 is the frequency, τ is the travel time divided by the quality factor Q, and γ is the slope of the curve trend beyond the corner frequency f . In finding the corner frequency the effect of Q is not taken into account while taking the Fourier transform using this formulation. The curves with particular value of travel time of the P-wave are used as master curves to be overlain on the observed spectra to estimate f_0 and Ω_0. The method used here has the advantage of offering a uniform and systematic approach to the problem of determining the spectral estimates.

The two source parameters, i.e., seismic moment and stress drop, are then computed from the spectral analysis of the recorded data using Brune's model (Brune, 1970 and Keilis-Borok, 1959). The formulations which are widely used to compute the source parameters (Archuleta et al., 1982 ; Fletcher, 1980 ; Tucker and Brune, 1977 ;

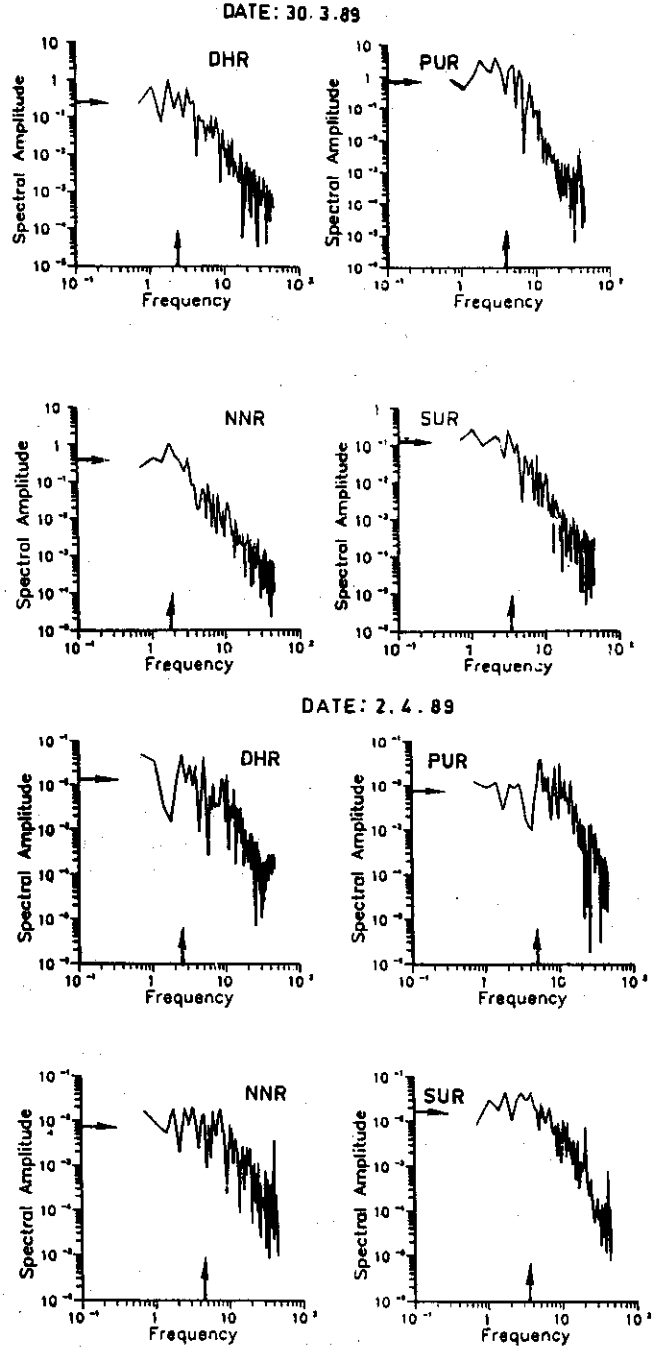

Fig. 2 Examples of Fourier Spectra of P-wave (2.84 sec window) for local earthquakes used for the analysis. The ↑ and → mark shows the corner frequency and Ω_0 value

O'Neill, 1984 ; Frankel and Wennerberg, 1989 ; Papageorgiou and Aki, 1983 and Thatcher and Hanks 1973) are as follows :

$$M_0 = 4\pi\rho v^3 D \Omega_0 / R_{\Theta\Phi} \tag{2}$$

$$r = v / (\pi f_0) \tag{3}$$

$$\Delta\sigma = 7 M_0 / (16 r^3) \tag{4}$$

where M_0, r, $\Delta\sigma$, D and $R_{\Theta\Phi}$ denote seismic moment, source radius, stress drop, epicentral distance and radiation pattern respectively. As fault plane solutions of the individual events could not be determined due to small number of recording stations, the radiation pattern $R_{\Theta\Phi}$ is approximately taken to be 85 percent (Prochazkova, 1980 and Sharma and Wason, 1994). The average value of seismic moments computed from different recording stations for an event is taken for M_0. The estimated values for the seismic moment and the stress drop for these earthquakes are tabulated in Table II.

TABLE II SOURCE PARAMETERS OF THE SEISMIC EVENTS

Sr. No.	Date	M_0 (dyne-cm)$\times10^{20}$	$\Delta\sigma$ (bars)	r (m)
1	25 11 88	62.32	38.49	545
2	08 03 89	0,81	0.62	363
3	16 03 89	34.66	19.83	424
4	29 03 89	22.84	19.25	373
5	30 03 89	15.26	6.15	477
6	02 04 89	2.55	0.92	493
7	04 04 89	10.07	4.05	477
8	09 04 89	7.29	2.57	498
9	11 04 89	0.63	0.33	437
10	11 04 89	3.73	1.15	521
11	12 04 89	4.18	1.38	509
12	16 04 89	4.79	1.47	521
13	19 04 89	6.81	2.74	477
14	22.04.89	53.32	21.49	477
15	23 04 89	0.40	0.41	347
16	03 05 89	0.36	0.24	402
17	03 05 89	0.07	0.04	395
18	04 05 89	0.28	0.21	381

The computed seismic moments for the seismic events range from 7 x 10^{18} dyne-cm to 6.2 x 10^{21} dyne-cm. The stress drops for four of the earthquakes are found to be more than 10 bars, with a maximum of 38.4 bars for an earthquake of local magnitude 4.2 having a focal depth of 15.4 km. For seven earthquakes the stress drops are computed to be between 10 bars and 1 bar and for another seven earthquakes the stress drops estimated to be very low, i.e. less than 1 bar.

SEISMIC MOMENT-MAGNITUDE RELATIONSHIP

Based on such studies several seismic moment relationships are proposed for various regions. In order to determine the empirical relationships between seismic moment-Vs-magnitude and stress drop-Vs-magnitude the linear fits are obtained by least squares regression analysis. The values of seismic moment are plotted versus M_L on a semi-log scale in Fig 3.

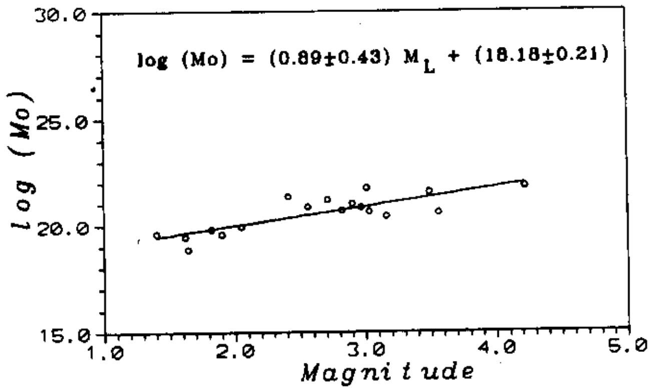

Fig. 3. Plot of seismic moment (M_0) Vs. Magnitude (M_L)

The relationships obtained with coefficients of the linear fits computed at the 90% confidence level using the t-distribution (Chatfield, 1989), are as follows :

$$\log(M_0) = (0.89 \pm 0.43)M_L + (18.18 \pm 0.21) \quad (5)$$

$$\log(\Delta\sigma) = (0.77 \pm 0.23)M_L - (1.78 \pm 0.46) \quad (6)$$

Similar relationships other seismic regions have been proposed by different authors are listed in Table III.

TABLE III a AND b VALUES COMPUTED FOR SEISMIC MOMENT MAGNITUDE RELATIONSHIP (LOG M_0 = a + b M_L) FOR DIFFERENT AREAS BY VARIOUS AUTHORS.

Sr No.	Author	Region	a	b
1	Present study	Garhwal Himalaya	18.18 ± 0.21	0.89 ± 0.43
2	Wyss and Brune (1968)	San Andreas	17.0	1.4
3	Thatcher and Hanks (1973)	Southern California	16.0	1.5
4	Johnson and McEvilly(1974)	California	17.60 ± 0.28	1.16 ± 0.06
5	Oncescu (1983)	Romania	18.0 ± 0.5	1.1 ± 0.1
6	Hanks and kanamori (1979)	(given for moment magnitude scale)	16.05	1.5

In Fig. 4 these relationships obtained between magnitude and seismic moment for different areas are sketched with the relationship worked out in the present study for the purpose of comparison. Hanks and Kanamori (1979) proposed a moment-magnitude scale given by the relation

$$M = \frac{2}{3}\log(M_0) - 10.7 \qquad (7)$$

or

$$\log(M_0) = 1.5M + 16.05 \qquad (8)$$

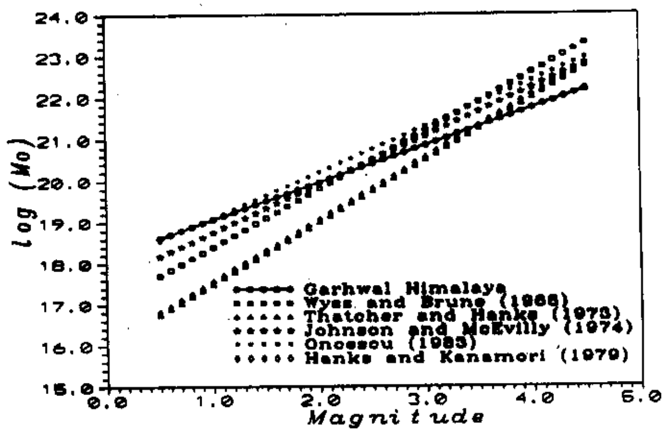

Fig. 4. Comparison of various seismic moment -magnitude relationships for different areas.

This moment magnitude scale is consistent with M_L in the range 3-6, with M_S in the range 5-8, and with M_W introduced by Kanamori (1977) for great earthquakes (M 8). This relationship is also shown in Fig. 4. This moment-magnitude relationship (equation 7) give higher values for seismic moment compared to the present relationship proposed for Garhwal Himalaya region in the present study for magnitudes higher than 3.5. For magnitudes below 3.5 the relationship proposed here give higher seismic moment. The moment-magnitude relationship proposed by Hanks and Kanamori (1979) is based on the results obtained by Thacther and Hanks (1973) for M_L in the range 3-6. Wells and Coppersmith (1994) computed the moment magnitude using the relation given by Hanks and Kanamori (1979) and found that the estimated seismic moments match with the otherwise known seismic moments for the earthquakes greater than magnitude 5.7 but for magnitude ranging between 4.7 to 5.7, the seismic moments estimates are on the lower side which is also evident from Fig. 4. Since in the present study the local earthquake data (upto magnitude 4.2) is used and the computations of the seismic moment are based on the body wave spectra (primary waves only), the relationship proposed here holds for magnitude less than 5. This relationship may not be extended to magnitudes greater than 5.

CONCLUSIONS

The source parameters are computed for shallow focus earthquakes occurring in Garhwal Himalaya region. The linear regression curves given in Fig. 4 show the trends of source parameters with respect to the local magnitude as given in equation 5. The seismic moment-magnitude relationship as given in equation 5 is proposed for the region based on local earthquake data. As the moment -magnitude relationship as proposed by Hanks and Kanamori (1979) yields low seismic moments in comparison to seismic moments estimated for the same events otherwise for magnitudes less than 5, the proposed relationship is preferred in this range. The rate of occurrence of earthquakes in this area is high and the relatively low stress drops at shallow depths show that the rock mass constituting the upper crust in the region has low strength for accumulation of strain and the rocks undergo brittle fracturing. The interpretation suggested above shall remain speculative until more data is obtained and possible sources of distortion and contamination of the spectra are better understood, as the present spectra are not corrected for site response.

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