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Crustal Attenuation of Shearwaves in Pithoragarh Region

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ABSTRACT

The region of Kumaon Himalaya is one of the seismically active regions of Himalaya. Frequent seismic activities in this region demonstrate the seismotectonic nature of the region. In this paper, necessary efforts are being made to study the attenuation properties of shear wave in Pithoragarh region using the data from strong motion network of eight stations installed in Kumaon Himalaya. In the present work, data of eight events recorded at Dharchula station have been utilized for inversion. The input to the algorithm used in the present work is simply the S phase of acceleration record and the hypocentral distance. The raw data have been processed using different steps like instrumental scaling, baseline correction and filtering. In the present work all records have been filtered using band passed Butter worth filter within ranges. Since the strong motion sites are located in weathered rock terrains, necessary efforts have been made to apply site amplification estimated by H/V spectral ratio techniques of Nakamura (1988) to noise data. Site amplification curves have been prepared which show less amplification at low frequencies compared to high frequencies.

Using the inversion algorithm developed by Joshi (2006), the frequency dependent Quality factor(Q) has been calculated which gives $Q_\beta(f) = 63\, f^{1.25}$ in frequency range. Stress drop for different events has been calculated using corner frequencies obtained from inversion. The stress drop of events ranges from very low to high value indicating highly unstable tectonic activity. Low value of coefficient (< 200) and high frequency dependence (>0. 8) in the $Q_\beta(f)$ relationship suggest that the region is seismically and tectonically active and is characterized by large number of heterogeneities.

INTRODUCTION

The region of Kumaon Himalaya is among the seismically active regions. This region falls in the highest zone of the seismic zoning map of India. During last 100 years this region has been visited by 14 earthquakes of magnitude greater than 6.0. The devastation caused by earthquake in any region is directly related to the attenuation characteristics of the medium and the amount of energy released during an earthquake. Attenuation characteristics of the medium control the decay of the seismic energy in the lithosphere and source characteristic of an earthquake controls the amount of energy released in an earthquake. Different rock materials have different property of attenuation of seismic energy. Attenuation of wave is a reduction in the amplitude or energy caused by heterogeneity and anelasticity in the earth. The attenuation can be defined by the inverse of the quality factor Q. The quality factor Q is a dimensionless parameter used to measure the tendency of material to dissipate energy during deformation. Although different fundamental definitions have been proposed for Q, the common idea has been to term a ratio of potential energy to the dissipated energy over one period of harmonic deformation (Pelton 2005). The attenuation is a petrophysical parameter that is sensitive to the lithology and physical properties like pressure, temperature, saturation with fluid and gas etc (Toksöz, Johnston & Timur 1978).

Very little work has been done in this part of Higher Himalaya to estimate the attenuation properties of medium. Paul, Gupta & Pant. (2003) have estimated a frequency dependent coda Q relationship using single back scattering model (Aki & Chouet 1975). They found $Q(f)= (92 \pm 4.73)f^{1.0\pm .023}$ using data from five stations installed in Kumaon Himalaya in frequency range. Based on the study of the aftershocks data of the Chamoli earthquake, Mandal et al. (2001) have estimated $Q_c(f)$ as $30\pm0.8f^{1.21\pm0.03}$ for the region surrounding the main shock, while the estimate of $Q_c(f)$ given by Gupta, Singh & Kumar. (1995) is $126f^{0.95}$ in frequency range. Gupta, Singh & Kumar. (1995) analyzed seven local earthquakes ($2.4\ d \leq M_L\ d \leq 4.9$) recorded at five stations in the Garhwal Himalaya. In all of the studies related to the estimation of coda Q(f) relation for Himalayan region, the single backscattering model proposed by Aki & Chouet (1975) has been used. Although attenuation property of any medium between earthquake source and observation point is explained by $Q_c(f)$, $Q_\beta(f)$ is required for successful prediction of strong ground motion in any region using Stochastic modelling technique (Boore 1983), Semi empirical modelling technique (Joshi, Singh & Kavita Garoti 2001 and Joshi & Midorikawa 2004) and composite source modelling technique (Zeng, Anderson & Su, 1994). Therefore estimate of $Q_\beta(f)$ not only serve purpose for attenuation studies of the region but also it is among useful parameters needed for successful prediction of strong ground motion for engineering use in any region. Joshi (2006) has developed a technique, which uses S phase of accelerogram as input in an inversion algorithm to give $Q_\beta(f)$ and corner frequency of input events. In the approach by Joshi & Midorikawa (2004) two-step inversion algorithm was applied to remove near site effects in the accelerogram due to unavailability of sufficient data for site studies. As we have enough data for the Dharchula region of Kumaon Himalaya we have used this data for computation of site effect at Dharchula station and hence accordingly the algorithm developed earlier by Joshi (2006) has been modified to estimate $Q_\beta(f)$ with prior knowledge of site effects at the recorded site.

Geology of the region surrounding Dharchula

Figure 1 shows the seismotectonic feature of the region. The Kumaon sector manifests strong deformation and reactivation of some of the faults and thrusts during Quarternary times. This is amply evident by the recurrent seismicity patterns, geomorphic developments and by geodetic surveys (Valdiya 1999). This region shows development of all the four morphotectonic zones, which are demarcated by intracrustal boundary thrust of regional dimension. These zones from south to north are: Siwalik or Sub Himalaya, Lesser Himalaya, Great Himalaya and Tehys Himalaya (Paul, Gupta & Pant 2003). Under a major Seismicity project funded by Department of Science and Technology, Govt. of India, a network of eight strong motion stations has been installed in the highly

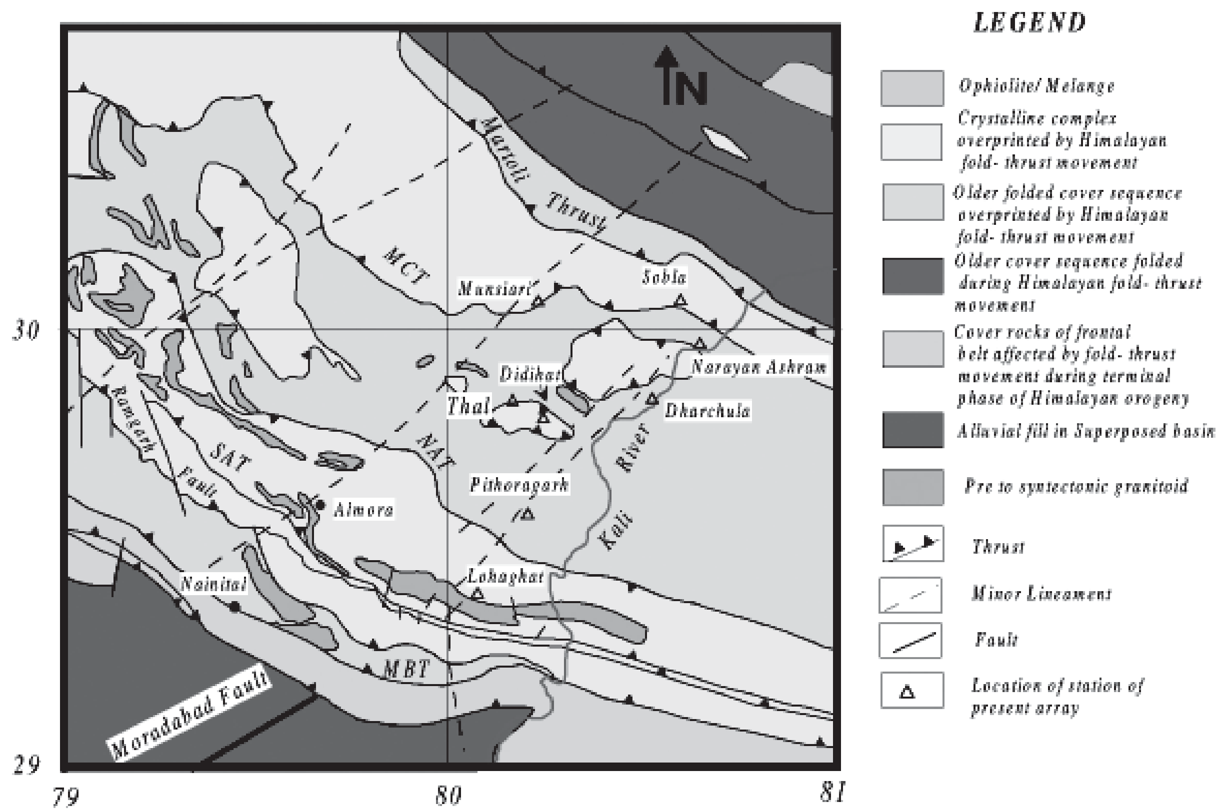


Figure 1. Location of strong motion recording stations installed in the Kumaon Himalaya. The triangle shows the location of strong motion stations. The geology and tectonics is after GSI (2000).

mountainous terrain of Kumaon Himalaya. Location of these eight stations along with the geology of the region is shown in Fig 1. Among these stations, Munsiari and Sobla lie in Greater Himalaya and rest lie in Lesser Himalaya. The Lesser Himalaya comprises of various thrust sheets and nappes being sandwiched between the Main Boundary Thrust (MBT) and Main Central Thrust (MCT) at the base of Great Himalaya. The great Himalayas comprises mainly Kyanite-Sillimanite bearing high grade psammatic Gneiss and Schist intruded by anatectic Tertiary leucogranite (Paul, Gupta & Pant 2003).

DATA

A strong motion network of eight stations having three-component force balance accelerometer has been installed in the Kumaon Himalayas. The natural frequencies of sensors in three directions are 206,202 and 202 Hz, respectively. The instruments are aligned in north south and east west direction at each station. In order to have nearly continuous digital recording mode, the threshold level of instrument was set at 0.005% of amplification at free surface(FS). The sampling interval of digital data is kept at .01 sec. The purpose of such low threshold level is to get data of maximum events in the duration of project.

The minimum inter station distance between these stations is approximately 11km. The stations are installed in lesser as well as Higher Himalayan Crystalline zone. The maximum numbers of stations are installed in the border district of Pithoragarh. This network has recorded eight events from March 2006 to July 2006. The arrival time of primary and secondary phases from recorded events has been selected for localization of events. For localization of events only those events, which were recorded at more than three stations, were used. Six events were

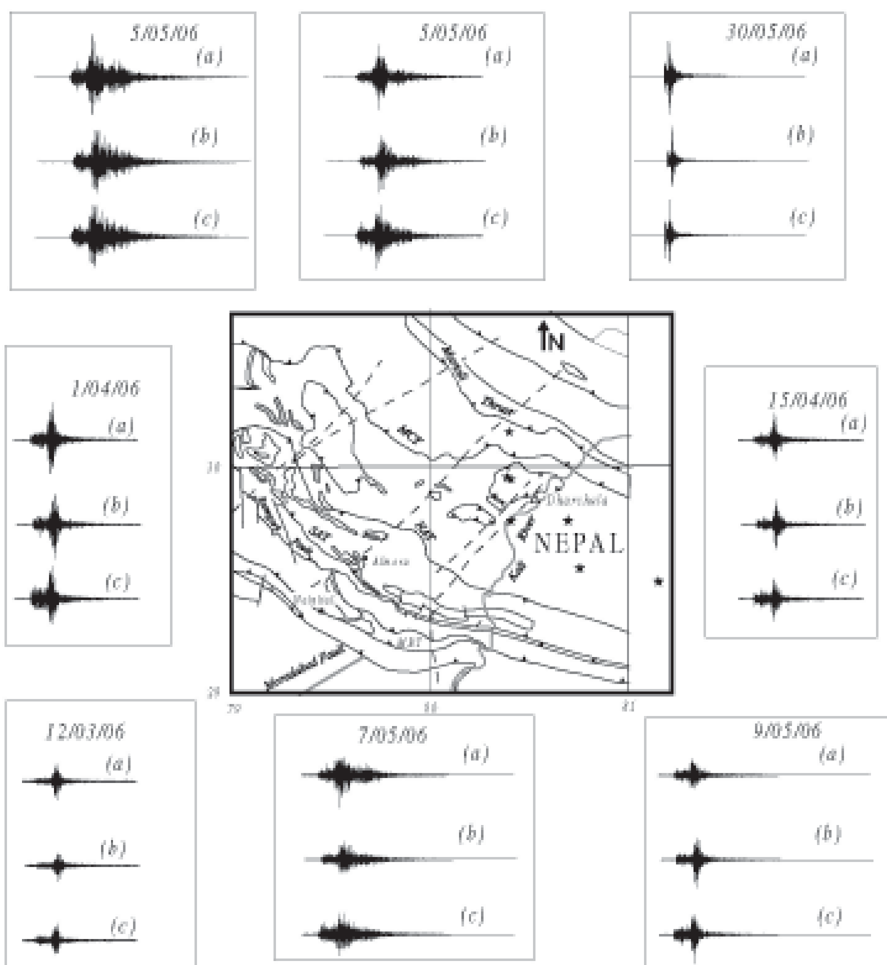


Figure 2. Processed (a) NS, (b) EW and (c) vertical component of accelerograms of different events recorded at the Dharchula station. Star denotes the epicenter of events. Triangle shows the location of recording station. The tectonics of the region is taken after GSI (2000).

Table 1: Locations of events recorded by strong motion network.

Date	Origin time	Epicenter	Depth(km)	ERH	RMS	No. ofstations
05/05/06	8:00:28.72	29º38.65′,80º 42.16′	30	2.7	.25	05
05/05/06	8:49:46.48	29º40.43′,80º 45.98′	25	4.2	.33	05
30/05/06	18:25:17.8	29º 53.81′,80º 26.9′	2.2	.9	.31	04
01/04/06	19:42:52.1	30º10.14′,80º 24.63′	10	6.5	.16	03
12/03/06	8:09:32.31	30º9.36′,80º 23.0′	34	4.1	.33	03
07/05/06	06:46:03.72	29º57.57′,80º 47.89′	35	14.1	.23	03

Table 2. Velocity model (after Sarkar et al, 2001)

Depth (km)	Vp (km/s)
2.0	4.0
17.0	5.2
37.0	6.0

localized using HYPO71 program originally developed by Lee & Lahr (1972). The parameters of these events obtained using HYPO71 are given in Table 1. In the present work, we have used data of eight events recorded at Dharchula station. The input to the algorithm used in the present work is simply the S phase of acceleration record and the hypocentral distance.

These accelerograms have been processed using the procedure suggested by Boore and Bommer (2005). The processing steps involve instrumental scaling, baseline correction and filtering. In the present work all records have been filtered using band passed Butter worth filter within ranges defined in Table 3. Different frequency ranges for different records have been assigned in order to remove low frequency noise below corner frequencies of source spectra. The processed acclerograms recorded at this station are shown in Fig 2.

Inversion

The acceleration spectra of shear waves at a distance R due to an earthquake of seismic moment M_o can be given as (Boore 1983 and Atkinson & Boore 1998):

$$A(f) = C\, S(f)\, D(f) \qquad \text{....(1)}$$

where the C term is constant at a particular station for a given earthquake, S(f) represents the source acceleration spectra and D(f) denotes a frequency-dependent diminution function which modifies the spectral shape and is given as (Boore and Atkinson, 1987):

$$D(f) = [e^{-\pi f R/\ Q(f)\beta}/R\ ]P(f,f_m)$$

In the above equation $P(f,f_m)$ is a high-cut filter and $e^{-\pi f R/Q(f)\beta}/R$ is the propagation filter. The term $Q_\beta(f)$ is the frequency-dependent shear wave quality factor. In this work we have used f_m as 50.0 Hz, which is the Nyquist frequency of the processed records at a sampling interval of 0.01 sec. This expression serves as the basis for our inversion. For a double-couple seismic source embedded in an elastic medium, considering only S waves, C is given as:

$$C = M_o R_{\theta\phi}\ .\ FS\ .\ PRTITN/(4\ \pi\rho\beta^3\ ) \qquad \text{.....(2)}$$

In the above expression, M_o is the seismic moment, $R_{\theta\phi}$ is the radiation pattern, FS is the amplification due to the free surface, PRTITN is the reduction factor that accounts for partitioning of energy into two horizontal components, and ρ and β are the density and the shear- wave velocity, respectively. $S(f,f_c)$ defines the source spectrum of the earthquake. Using the spectral shape based on ω^{-2} decay at high frequency proposed by Aki & Chouet (1975) and Brune (1970), $S(f,f_c)$ is defined as:

$$S(f,f_c) = (2\pi f)^2/\ (1+(f/f_c)^2) \qquad \text{.....(3)}$$

Eq (1) is linearised by taking its natural logarithm.

This modifies eq (1) as:

$$\ln A(f) = \ln C + \ln (S(f, f_c)) - \pi f R/ Q_\beta(f)\beta - \ln (R) + \ln P(f,f_m) \quad \ldots..(4)$$

This is now in linearised form with unknown $Q_\beta(f)$ and f_c. The term representing the source filter $S(f,f_c)$ is replaced with eq (3). With an assumption of known values of f_c, we are left with only unknown $Q_\beta(f)$, which can be obtained from inversion by minimizing it in a least- squares sense. The least-squares inversion minimizes:

$$\chi^2 = \sum [A_s(f) - S(f, f_c)]^2 \quad \ldots.(5)$$

where $S(f,f_c)$ is the theoretical source acceleration spectrum and $A_s(f)$ is the source spectrum obtained from the record after substituting parameters $Q_2(f)$ obtained from the inversion of eq (4). Rearranging known and unknown quantities on different sides, we obtain the following form from the equation (5):

$$-\pi f R/ Q_\beta(f)\beta = \ln A(f) - \ln C - \ln (S(f, f_c)) + \ln (R) - \ln P(f,f_m) \quad \ldots..(6)$$

Substituting the term related to the source spectrum $S(f,f_c)$ as $(2\grave{A}f)^2/(1+(f/f_c)^2)$, we obtain:

$$-\pi f R/ Q_\beta(f)\beta = \ln A(f) - \ln C - \ln (2\pi f)^2 + \ln(1+(f/f_c)^2) + \ln (R) - \ln P(f,f_m) \ldots..(7)$$

In eq (7), the dependence on the corner frequency (or change in the corner frequency) has been linearised by expanding $\ln(1+(f/f_c)^2)$ in a Taylor series around f_c. Accordingly we obtain the following expression:

$$-\pi f R/ Q_\beta(f)\beta = \ln A(f) - \ln C - \ln [(2\pi f)^2/(1+(f/f_c)^2)] - [2/(1+(f/f_c)^2)](f/f_c)^2(\Delta f/f_c) + \ln (R) - \ln P(f,f_m) \ldots ..(8)$$

Here Δf_c is the small change in the corner frequency and is an unknown quantity that is obtained from the inversion. We obtained following set of equations at a particular station for the ith earthquake for frequencies f_1, f_2, f_3 ……….. f_n, where n denotes total number of digitized samples in the acceleration record:

$$\begin{aligned}
-\pi f_1 R_{11}/ Q_\beta(f_1)\beta + F(f_1,f_{c1})\Delta f_c &= D_{11}(f_1) \\
-\pi f_2 R_{11}/ Q_\beta(f_2)\beta + F(f_2,f_{c1})\Delta f_c &= D_{11}(f_2) \\
\vdots \qquad\qquad \vdots \qquad &\qquad \vdots \\
-\pi f_n R_{11}/ Q_\beta(f_n)\beta + F(f_n,f_{c1})\Delta f_c &= D_{11}(f_n)
\end{aligned}$$

where $F(f,f_c) = 2/(1+(f_1/f_c)^2)](f_1/f_c)^2(1/f_c)$ is the term obtained from the expansion of $\ln(1+(f/f_c)^2)$ in terms of Taylor series around f_c.

In the matrix form, the above set of equations can be written as:

$$\left|\begin{array}{cccccc}
-\pi f_1R_1/\beta & 0 & 0 & \ldots\ldots\ldots & 0 & F(f_1,f_{c1}) \\
0 & -\pi f_2R_1/\beta & 0 & \ldots\ldots\ldots & 0 & F(f_2,f_{c1}) \\
0 & 0 & -\pi f_3R_1/\beta & \ldots\ldots\ldots & 0 & F(f_3,f_{c1}) \\
\vdots & \vdots & \vdots & & \vdots & \vdots \\
0 & 0 & 0 & \ldots\ldots & -\pi f_nR_1/\beta & F(f_n,f_{c1}) \\
 & & & \vdots & & \\
 & & \multicolumn{2}{c}{\text{for } m^{th} \text{ earthquake}} & & \\
-\pi f_1R_m/\beta & 0 & 0 & \ldots\ldots\ldots & 0 & F(f_1,f_{cm}) \\
0 & -\pi f_2R_m/\beta & 0 & \ldots\ldots\ldots & 0 & F(f_2,f_{cm}) \\
0 & 0 & -\pi f_3R_m/\beta & \ldots\ldots\ldots & 0 & F(f_3,f_{cm}) \\
\vdots & \vdots & \vdots & & \vdots & \vdots \\
0 & 0 & 0 & \ldots\ldots & -\pi f_nR_m/\beta & F(f_n,f_{cm})
\end{array}\right|
\left|\begin{array}{c}
1/Q_\beta(f_1) \\ 1/Q_\beta(f2) \\ 1/Q_\beta(f3) \\ \vdots \\ \vdots \\ \vdots \\ 1/Q_\beta(f_n) \\ \Delta f_c
\end{array}\right|
=
\left|\begin{array}{c}
D_1(f_1) \\ D_1(f_2) \\ D_1(f_3) \\ \vdots \\ D_1(f_n) \\ \vdots \\ D_m(f_1) \\ D_m(f_2) \\ D_m(f_3) \\ \vdots \\ D_m(f_n)
\end{array}\right|$$

This can be represented in the following form:

$$G\,m = d$$

Model parameters are contained in the model matrix 'm' and the spectral component in the data matrix 'd'. Inversion of the 'G' matrix gives the model matrix 'm' using the Newton's method as below:

$$m = (G^TG)^{-1}G^T\, d$$

In the present inversion we have performed a grid search for f_c. We use the initial values of f_c as 0.01 Hz and than it is increased in Δf_c increments of 0.01 Hz up to 10.0 Hz. We get different solutions for different possibilities of f_c. The final solution is that which minimizes eq (5). The above inversion is prone to the problems if G^TG is even close to singular and for such a case, we prefer the singular value decomposition (SVD) to solve m (Press et al., 1992). Formulation for the SVD is followed after Lancose (1961). In this formulation, the G matrix is decomposed into U_p, V_p and Λ_p matrices as given by Fletcher (1995):

$$G^{-1} = V_p\Lambda_pU_p^T \quad \ldots\ldots.(9)$$

Where, V_p, U_p and Λ_p are matrices having nonzero eigenvectors and eigenvalues. One of the important assumptions in this formulation is that the acceleration spectra should be free from site amplifications due to local heterogeneities. For this purpose the entire acceleration record has been corrected for site amplifications. The site amplifications have been computed using the data available at the station. The software QINV developed by Joshi (2006) has been modified and used for present work. The obtained $Q_\beta(f)$

relationship is used to compute the source displacement spectrum at each stations.

One of the important parameters of an earthquake source is the stress drop '$\Delta\sigma$' which is defined as the difference of preexisting tectonic stress and the dynamical frictional stress. For computing this we need to relate the average slip with the stress drop. The solution of this problem is available in terms of a shear crack model under uniform stress drop for various source geometries e.g., circular (Eshelby 1957; Keilis Borok 1959), two-dimensional in plane (Starr 1928) or antiplane (Knopoff 1958). For a circular crack of radius r_o, the stress drop $\Delta\sigma$ is given as (Papageorgiou & Aki 1983):

$$\Delta\sigma = 7M_o/16\ r_o^3 \qquad \text{......(10)}$$

This stress drop is remarkably constant falling in the range of 1 to 10 MPa for wide range of magnitudes ($5 \leq M_s \leq 8$) . The other parameter representing the source is its size, which is defined by the radius for circular rupture. The corner frequency f_c of the source spectra is related to the radius r_o of the equivalent circular crack, which is used to model the earthquake. Such relations have been given by Brune (1970, 1971):

$$r_o = 2.34\beta\ /2\pi f_c \qquad \text{......(11)}$$

RESULTS AND DISCUSSION

The first step in the inversion scheme is the identification of S phases from the available strong motion data. One of the major requirement of using eq.(1) for inversion is that the acceleration spectrum shown in the left hand side of equation should be free from site amplifications. This can be achieved in two ways (1) by using records from those stations where site amplifications are not effective within the range of frequencies required for the inversion (i.e., rock site) or (2) by removing the site amplifications directly from the acceleration spectra. For this a micro tremor study has been conducted at this station be keeping the threshold level as low as .005% of maximum value. This study is conducted for two months and during this period several noise data was collected. Using this noise data and the H/V technique of Nakamura (1988), site amplification curves has been prepared at Dharchula station and is shown in Fig 3. The average value of site amplification shows that at low frequencies the amplification is less pronounced than at high frequencies.

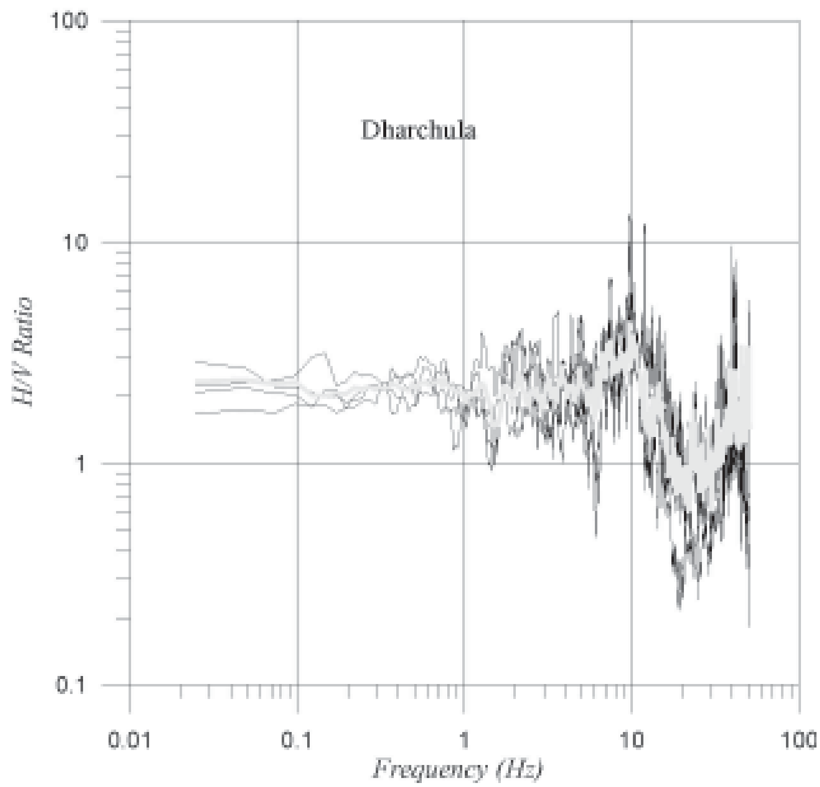
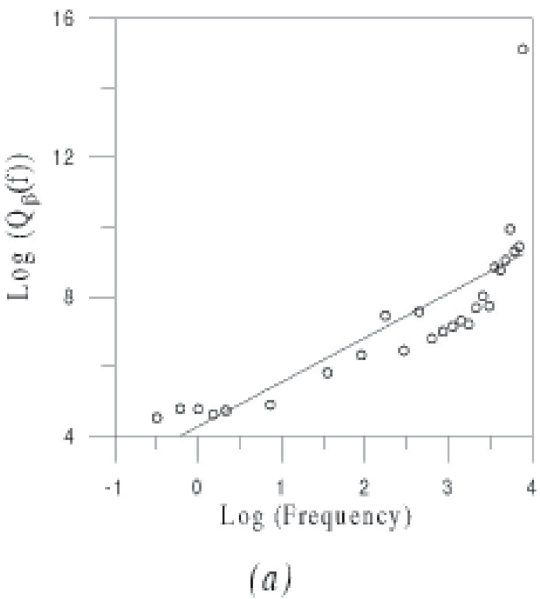


Figure 3. Site amplification at Dharchula station. The thick line shows the average value of site amplification at this station.



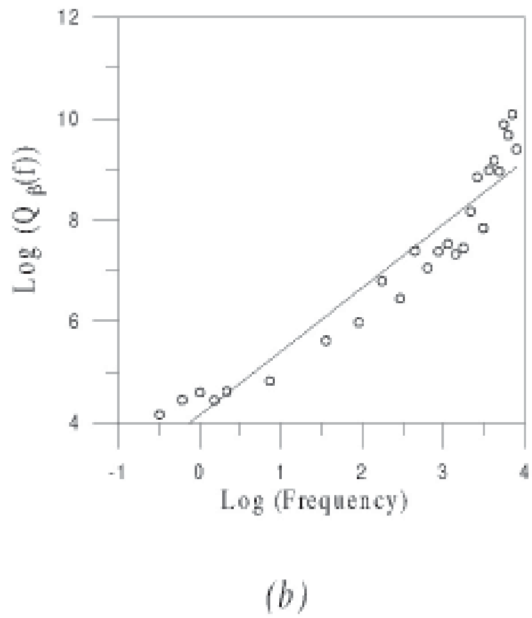


Figure 4. The Q_β (f) relations obtained after using accelrograms (a) uncorrected and (b) corrected for near site amplifications.

Table 3: Filtering parameters used in the processing of strong motion records.

Event No.	Date	Filtering ranges	
		Low (Hz)	High (Hz)
1	05/05/06	.4	45.0
2	05/05/06	.5	45.0
3	30/05/06	.4	45.0
4	01/04/06	1.3	45.0
5	12/03/06	2.0	45.0
6	07/05/06	2.0	45.0
7	15/04/06	1.0	45.0
8	09/05/06	2.0	45.0

Due to limitation in computational work selected frequencies have been considered for solving the matrix form of eq. (3). The selections of frequencies are made in such a way to have equal weightage in each quadrant of logarithmic scale for deriving $Q_\beta(f)$ relationship in entire frequency band. In order to test the dependency of developed algorithm on the site amplifications, the inversion is performed using accelerograms with and without corrected for site amplifications. Using the obtained values of Q at different frequencies two $Q_\beta(f)$ relations are obtained as shown in Fig 4.

It is seen that use of accelerogram without corrected for site amplification gives $Q_\beta(f) = 71\ f^{\,1.27}$ which has rms error in model and data matrix as.45 and .63, respectively. The functional form of $Q_\beta(f)$ has been obtained from regression analysis. It is seen that $Q_\beta(f)$ can be represented in terms of $Q_o f^n$ and in the present regression study Q_o and n varies as 70±46 and .94±.82, respectively. The rms error is reduced significantly for the case when accelerograms corrected for site amplifications have been used as input in the inversion algorithm. For this case, rms error in data and model matrix is obtained as .14 and .03, respectively, for $Q_\beta(f) = 63f^{\,1.25}$. The obtained regression study for this case gives Q_o and n which varies between 75±45 and 1.06± .614, respectively. This shows that introduction of accelerogram corrected for site amplification gives regression relation in which less rms error and comparatively less variation in Q_o and n is obtained. Therefore site correction plays an important role in determining the attenuation characteristics of the medium. The inversion also gives the corner frequencies corresponding to each input event and this has been reported in Table 3. Using the estimates of corner frequency, source parameters like crack radius of input events have been estimated using eq (10&11)and are given in Table 3.

The obtained $Q_\beta(f)$ relation is compared with other relations available for tectonically active Indian and worldwide regions and is shown in Fig 5. It is seen that the present relation falls in a range which is characterized by the tectonically active region. In order to check obtained $Q_\beta(f)$, simple derivations have been made. From eq (1) we know that acceleration spectra $A_1(f)$ and $A_2(f)$ due to event 1 and 2, respectively are given as:

$$A_1(f) = C_1\ S_1(f)\ [e^{-\pi f R_1/\ Q(f)\beta}/R_1\ ]$$

$$A_2(f) = C_2\ S_2(f)\ [e^{-\pi f R_2/\ Q(f)\beta}/R_2\ ]$$

This can be rewritten as:

$$R_1A_1(f) = C_1\ S_1(f)\ [e^{-\pi f R_1/\ Q(f)\beta}] \quad ....(12)$$

$$R_2A_2(f) = C_2\ S_2(f)\ [e^{-\pi f R_2/\ Q(f)\beta}] \quad ....(13)$$

Dividing equation 12 and 13 we get

$$R_1A_1(f)/\ R_2A_2(f) = (C_1/C_2)\ (S_1(f)/\ S_2(f)\ )\ [e^{-\pi f(R1-R2)/Q(f)\beta}] \quad ......(14)$$

Taking natural logarithm of eq. (14) on both sides we get:

$$\ln[R_1A_1(f)/\ R_2A_2(f)] = \ln(C_1/C_2) + \ln\ (S_1(f)/\ S_2(f)\ - \pi f(R_2 - R_1)/Q(f)\beta$$

This can be further rearranged as:

$$\ln[R_1A_1(f)/\ R_2A_2(f)] - \ln(C_1/C_2) + \ln\ (S_1(f)/\ S_2(f) = \pi f(R_2 - R_1)/Q(f)\beta \quad ...(15)$$

The term on the left hand side of eq.(15) can be computed from the observed data and known values of C_1, C_2, $S_1(f)$ and $S_2(f)$, respectively. It is seen that if we consider the spectra of observed records above the maximum corner frequency of two events , then

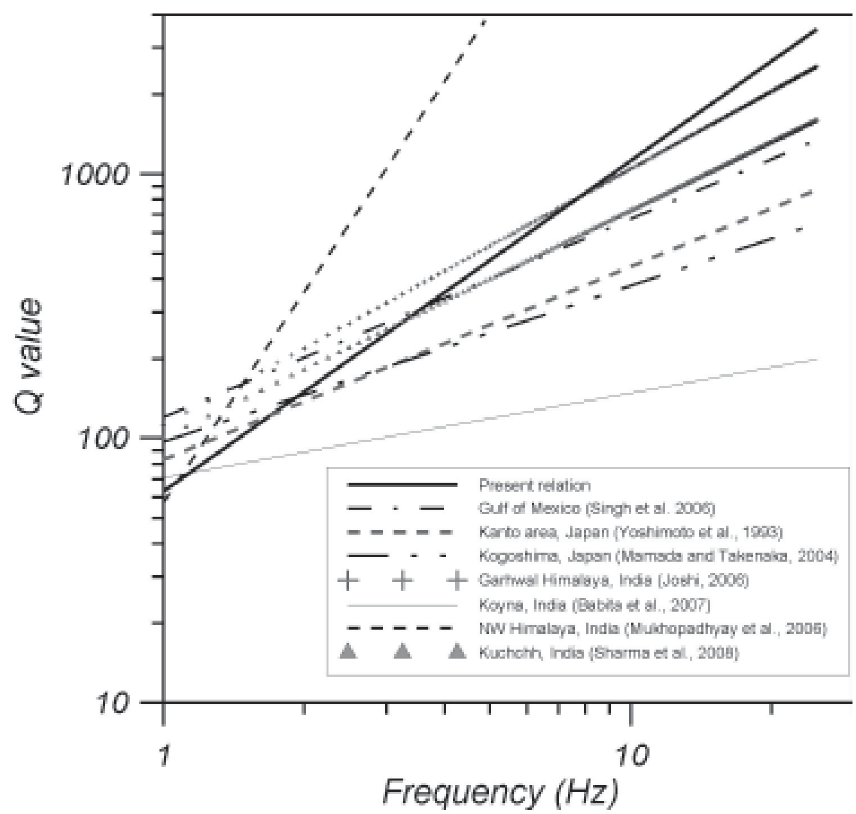


Figure 5. Comparison of $Q_\beta(f)$ obtained in the present study with that obtained for various active regions.

the left hand side and right hand side of eq (15) show almost same dependency on frequency. This is because the source acceleration spectrum is almost flat above the corner frequency. There will be a constant shift in the left hand side which depends on the ratio of seismic moments and radiation patterns of two events used in this computation. Using the obtained relation of Q(f), we have computed theoretical value $\pi f(R_2 - R_1)/Q(f)\beta$ and compared it with the value obtained form theoretical data using expression in the left hand side of equations. Using several combinations of events used in the present study this term is computed from observed data and is compared with the theoretical value. This comparison is shown in Fig 6. This shows that the theoretical value of term $\pi f(R_2 - R_1)/Q(f)\beta$ matches closely with that obtained from observed data for most of the cases.

Using the obtained value of $Q_\beta(f) = 63 f^{1.25}$ accelerograms are corrected for frequency-dependent diminution function and source spectrum has been obtained. This is shown in Fig 6. Fig 7 shows that realistic match is seen between the theoretical Brunes spectra and the source spectra obtained from acceleration records, thereby establishing the efficacy

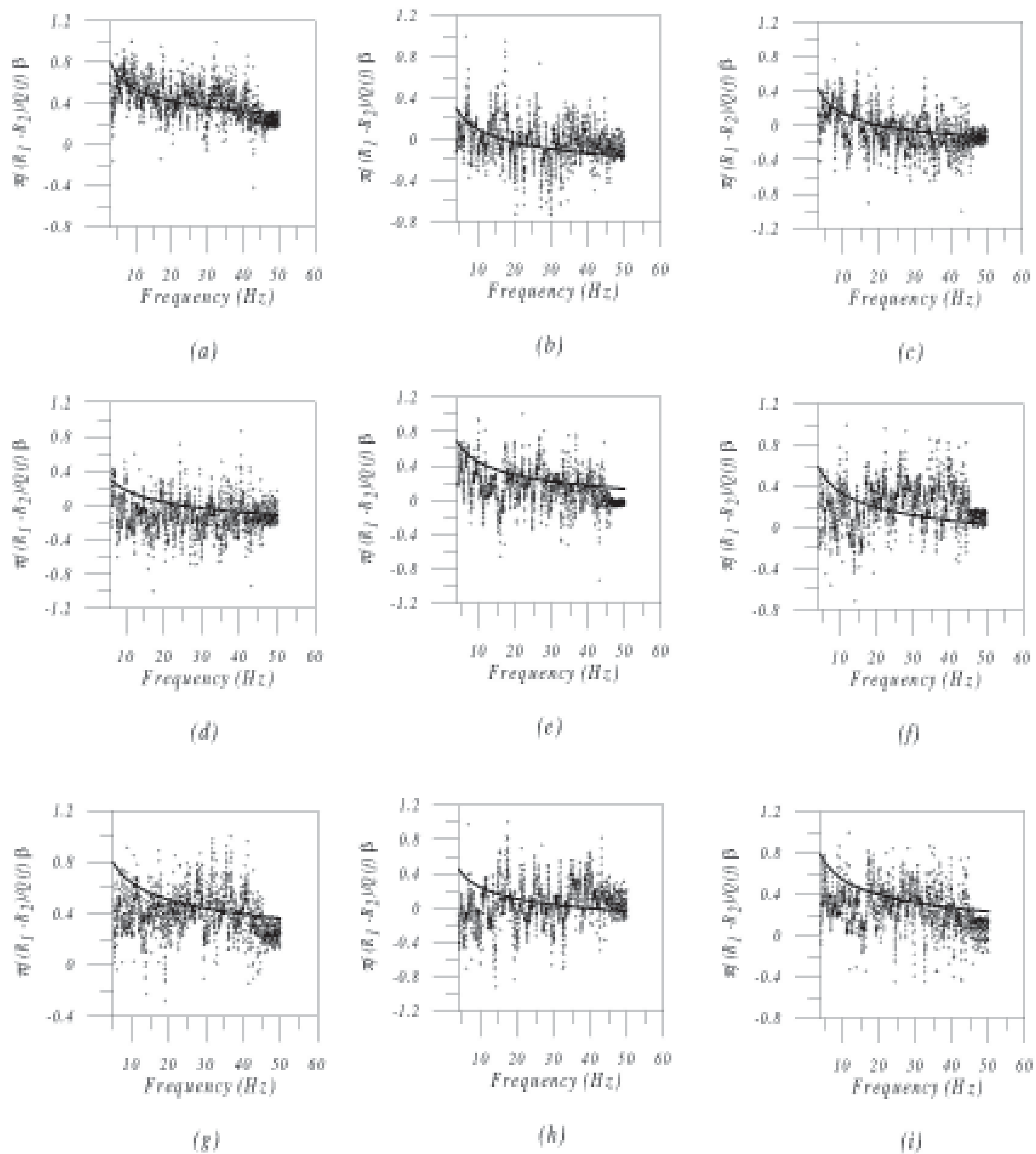


Figure 6. comparison of theoretical and observed $\pi f(R_2 - R_1)/Q(f)\beta$ obtained from (a) event 1 and 2, (b) event 1 and 6; (c) event 1 and 2; (d) event 1 and 4; (e) event 1 and 5; (f) event 2 and 5; (g) event 2 and 8; (h) event 2 and 6 and (i) event 4 and 5. The nomenclature of events is given in Table 2.

of the obtained $Q_{\beta}(f)$ relationship from inversion. As the present inversion is strongly dependent on single station data therefore present relation are valid only for limited region and can be applied to larger area by considering more data set. Besides estimating the nature of the material between source of an earthquake and the observation point, the developed $Q_{\beta}(f)$ relation can be used for realistic simulation of strong ground motion using stochastic simulation technique.

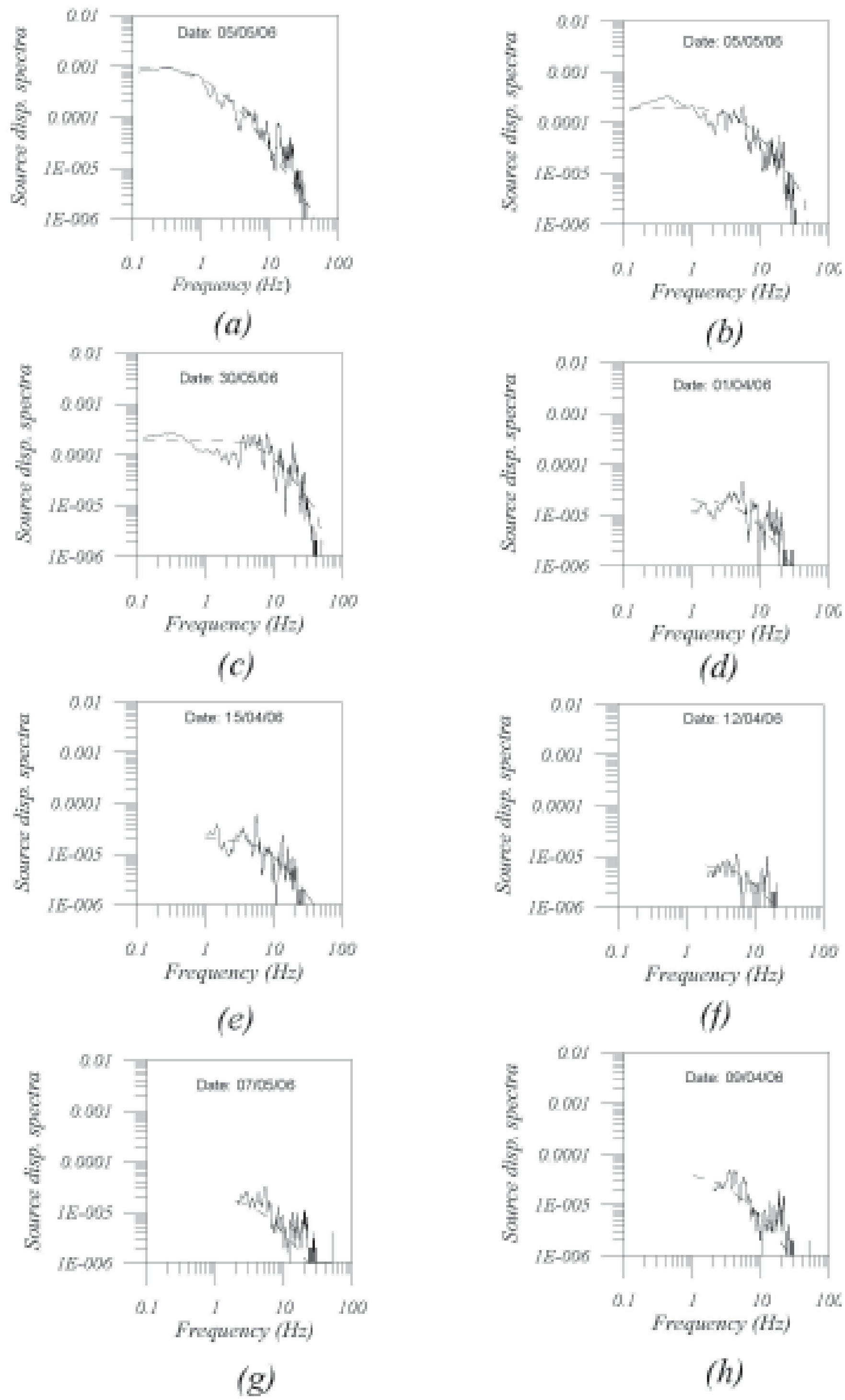


Figure 7. Comparison of theoretical and observed source spectra of the event recorded at Dharchula station on (a) 5/05/06, (b) 5/05/06 (08:00, (c) 30/05/06, (d) 01/04/06, (e) 15/04/06, (f)12/04/06, (g) 07/05/06 and (h) 15/ The theoretical source spectra are shown by thick black line.

CONCLUSIONS

Present work gives an effective algorithm for obtaining $Q_\beta(f)$ relation from single station acceleration data. Data of eight events recorded in a local strong motion network of eight stations in Kumaon Himalaya have been used in this study. Using records of eight events recorded at Dharchula stations, $Q_\beta(f) = 63 f^{1.25}$ is obtained as frequency dependent relation. The corner frequency obtained from inversion is used to compute source parameters of studied events. The obtained $Q_\beta(f)$ relation shows that the region is seismically active and characterized by local heterogeneities. Further large difference in the value of stress drop for various studied events also confirms that region is seismically unstable.

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