Separation of intrinsic and scattering seismic wave attenuation in Northeast India

Simanchal Padhy¹,² and N. Subhadra¹

¹National Geophysical Research Institute (Council of Scientific and Industrial Research), Uppal Road, Hyderabad-500007, India.
E-mail: padhy@rediffmail.com
²Earthquake Research Institute, The University of Tokyo, 1–1–1 Yayoi, Bunkyo-ku, 113–0032 Tokyo, Japan

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SUMMARY
We have analysed the local earthquakes (2.0 ≤ Ml ≤ 5.5) occurred in northeast (NE) India recorded by a temporary seismic network of 10 stations operated by National Geophysical Research Institute (NGRI), Hyderabad to evaluate the relative contributions of scattering loss (Qsc⁻¹) and intrinsic absorption (Qi⁻¹) to total attenuation (Qt⁻¹) using the multiple lapse time window analysis assuming multiple isotropic scattering in a medium of uniformly distributed scatterers. The results show that Qt⁻¹ is greater than Qsc⁻¹ at high frequencies (f > 3 Hz), while the opposite is observed at low frequencies (f < 3 Hz). The observed frequency dependence of Qsc⁻¹ corresponds to the scale length of lithospheric heterogeneity beneath NE India, at least comparable with the wavelength for the lowest frequencies analysed, of about 1 km. The observed Qi⁻¹ for the study region obtained with single scattering theory is close to Qt⁻¹ at high frequencies, in agreement with theoretical prediction for an idealized case of uniform distribution of scatterers. However, a discrepancy exists between the two at low frequencies, which can be explained by a depth-dependent velocity and attenuation structure. High value of Qi⁻¹ is correlated with the geology and tectonic settings of the region characterized by Himalayan and Burman collision zones with a strong lateral heterogeneity. The Qi⁻¹ estimates obtained in this study can be used to infer the average temperature of the lower crust with an upper limit estimate of ~800 °C assuming a lower crustal gabbroic lithology.

Key words: Seismic attenuation; Wave scattering and diffraction; Wave propagation.

1 INTRODUCTION
The attenuation of seismic waves is a reduction in amplitude or energy caused by scattering from heterogeneity or intrinsic absorption due to anelasticity or both. Scattering of high-frequency seismic waves has been an important component in ground-motion prediction used in seismic hazard analyses and earthquake engineering. Estimates of scattering attenuation may be useful in accurate prediction of ground motion. For example, Galluzzo et al. (2008) show that in a highly scattering medium (low scattering Qi) the duration of ground motion is much larger than that in a medium characterized by high scattering Q, regardless of the value of intrinsic Q. The attenuation of seismic wave can be characterized by the inverse seismic quality factor, Qi⁻¹, defined as the fractional loss of energy per cycle. In general, the quality factor is a function of frequency, Q = Q(f). The total attenuation can be expressed as Qt⁻¹ = Qsc⁻¹ + Qi⁻¹ (Dainty 1981), where Qi⁻¹ represents the intrinsic absorption caused by the conversion of elastic energy to heat or other forms of energy, and Qsc⁻¹ is the scattering attenuation caused by the redistribution of energy that occurs when seismic waves interact with the heterogeneities randomly distributed in the lithosphere. Measurement of Qsc⁻¹ is important to characterize the medium heterogeneity and that of Qi⁻¹ is used to infer the temperature variations and other physical parameters, such as compositional variations, the presence of partial melts and/or fluids of the lithosphere (Kampfmann & Berckhemer 1985).

The effects of Qi⁻¹ and Qsc⁻¹ cannot be directly separated from the estimate of Qt⁻¹ from direct waves due to the trade-off in the estimate (Dainty 1981). As the energy lost by direct waves is recuperated in the coda, which in turn is formed by scattering processes inside the Earth, a scattering model is necessary. Several authors have separated Qi⁻¹ and Qsc⁻¹ by using the multiple lapse time window analysis (MLTWA) method applicable to energy transport (Shang & Gao 1988; Zeng et al. 1991) and energy flux (Frankel & Wennerberg 1987) models of the coda energy envelopes as a function of distance and lapse time. Of these methods, the MLTWA method has been widely applied to several areas in the world to resolve both the mechanisms of Qi⁻¹ (Sato et al. 2012 and references therein). Some of the recent findings based on this method include: Ugalde et al. (2007) in Kachchh, western India based on the assumption of both uniform half-space and flat layered structure; Badi et al. (2009) in the Nuevo Cuyo region of southern central Andes,
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Figure 1. Map of NE India showing main structural features, station locations (triangles), event locations (circles) and major earthquakes (stars). The seismic stations are Rupa (RUPA), Bhairabkund (BDK), Bhalkupung (BPG), Seisojsha (SJA), Tezpur (TZR), Dwookmukh (DMK), Jogigopha (JPA), Nangalbibra (NGL), Hameren (HMN) and Jagirroad (JGR). Small square in the inset represents the study area. The ‘Assam gap’ is shown by the closed box. The tectonic feature KL stands for Kopili lineament.

Chung et al. (2009) in Mt Fuji of Japan, Carcole & Sato (2010) in Japan, Mukhopadhyay et al. (2010) in Garwhal-Kumaun Himalayas, India, Ugalde et al. (2010) in the south western Colombia, all based on a uniform half-space; and Del Pezzo et al. (2011) in north central Italy based on a flat layered structure. Results show that the attenuation coefficients obtained using the MLTWA for a uniform model are overestimated (Hoshiba et al. 2001; Bianco et al. 2005; Del Pezzo & Bianco 2010). Del Pezzo & Bianco (2010) applied the corrections to the attenuation coefficients for a uniform model to get their true estimates for a realistic two-layer earth model.

The northeast (NE) India is a tectonically complex area including a subduction zone. The region is characterized by an active tectonics and a high level of seismicity (Fig. 1) posing the highest seismic hazard of India. Over a dozen major earthquakes of $M \geq 7$ and the 1897 June 12 ($M$ 8.5; Oldham 1899) and 1950 August 15 ($M$ 8.7; Tandon 1955) great hazardous earthquakes have rocked the region over the past 110 yr. This region comes under maximum rating zone (zone V) in the seismic hazard zonation map for Indian subcontinent (Bhatia et al. 1999) prepared under the Global Seismic Hazard Assessment Program (GSHAP). The peak ground acceleration (PGA) value for the region is found to be $\sim 1$ m s$^{-2}$ at 100 km, which decays to $\sim 0.1$ m s$^{-2}$ at 400 km distance (Padhy et al. 2010). To better understand the damage distribution and hence to accurately obtain the level of seismic shaking expected from future large earthquakes, it is very important to study attenuation properties of the medium and the underlying causes of attenuation in this tectonically complex region.

Earthquakes occurring in this area are monitored by a local seismic network operated by National Geophysical Research Institute (NGRI), Hyderabad. There are very few studies on attenuation of high-frequency seismic waves in this region (Hazarika et al. 2009; Padhy & Subhadra 2010a,b). Hazarika et al. (2009) interpreted coda waves of local earthquakes based on the single isotropic scattering model of Sato (1977). Padhy & Subhadra (2010a) studied the attenuation of high-frequency $S$ waves and coda waves, its frequency and lapse time dependence in 1.5–24 Hz, by adopting theories of both single scattering (Aki & Chouet 1975) and multiple scattering (energy flux and diffusion) to bandpass-filtered seismogram envelopes. They discussed the effect of multiple scattering at larger distances on $Q^{-1}$ estimates. Padhy & Subhadra (2010b) estimated the total attenuation of $P$ and $S$ waves in the lithosphere beneath this region using the extended coda-normalization method (Yoshimoto et al. 1993). Their results show that both $P$ and $S$ waves undergo a strong attenuation along ray paths. However, the nature of total attenuation in the region is not resolved yet in the aforementioned studies. The purpose of this study is, therefore, to resolve both the mechanisms of attenuation, that is, intrinsic and scattering attenuation using the MLTWA, under the assumption of multiple isotropic scattering and a uniform distribution of scatterers in the medium. Finally, we compare our results with other regions investigated using the similar methodology, as used in this study.

2 GEOLOGY AND SEISMOTECTONICS

The NE India and its surroundings constitute a tectonically complex region, where the eastern Indian Plate underthrusts beneath the Eurasian Plate resulting in a complex deformation in the region that is manifested by the region’s high seismicity (Fig. 1), and neotectonic geological expressions (Nandy 2001). Based on detailed seismotectonic evaluation of NE India, the region can be characterized by three major tectonic units (Fig. 1). First, the Himalaya...
3 METHOD OF ANALYSIS

We briefly introduce the theoretical background for separation of scattering ($Q_{sc}^{-1}$) and intrinsic absorption ($Q_{in}^{-1}$) from the total attenuation ($Q_{t}^{-1}$).

3.1 Estimation of $Q_{t}^{-1}$ and $Q_{sc}^{-1}$ using MLTWA for a uniform half-space model

The MLTWA technique consists of comparing the theoretical energy integrated in three successive time windows on the seismogram, as a function of hypocentral distance, with the correspondent observed values measured from filtered (in successive frequency bands) seismograms. The fit yields the values of the seismic albedo, $B_0$ and the inverse extinction length, $L_e^{-1}$ for any frequency band. The parameter $B_0$ is defined as the dimensionless ratio of the scattering attenuation to the total attenuation, and $L_e^{-1}$ is the inverse of the distance over which the S-wave energy is decreased by $e^{-1}$. The inverse quality factors $Q_t^{-1}$, $Q_{sc}^{-1}$, and $Q_{in}^{-1}$ are then calculated through the expressions $Q_t^{-1} = L_e^{-1} + 2$, $Q_{sc}^{-1} = B_0Q_t^{-1}$, and $Q_{in}^{-1} = (1 - B_0)Q_t^{-1}$, where $V_S$ is S-wave velocity (Hoshiba et al. 1991).

We calculated, for each seismogram, the observed energy envelopes by squaring the amplitudes of bandpassed traces, as a function of time $t$, to obtain $E_{obs}(f|t|)$. Then, the function $E_{obs}(f|t|)$ is integrated over three successive time windows of 15 s duration starting from the S-wave onset, $r/V_S$ to obtain $E_{c,obs}(f|r|)$, ($i = 1, 2, 3$), where $t_i$ is a fixed reference lapse time that satisfies the condition $t_i \geq \frac{r}{V_S}$ for all $r$ chosen for coda normalization. This coda window is selected in accordance with typical coda studies (Rautian & Khalturin 1978) to ensure that it does not include significant direct waves and the signal amplitude was large for $t_i \geq \frac{r}{V_S}$ (i.e. the signal reaches about twice the noise level). A correction for the geometrical spreading effect is made by multiplying $E_{c,obs}(f|r|)$ by $4\pi r^2$.

The theoretical energy density at a given lapse time and a hypocentral distance, in case of isotropic scattering and uniform half-space, is calculated by using the approximate analytical solution of the radiative transfer equation in three dimensions (Paaschens 1997). The form of solution can be written as

$$E_{0}(r, t) \approx \frac{E_0 e^{-\gamma t}}{4\pi r^2 V_S} \frac{1}{\delta} \left[ t - \frac{r}{V_S} \right]$$

$$+ E_0 H \left[ t - \frac{r}{V_S} - \left( \frac{1 - \frac{r}{V_S}}{\frac{4\pi r^2 V_S}{\gamma}} \right)^{1/8} - e^{-\gamma t} \right] G \left[ V_S t_0 \left( 1 - \frac{r^2}{V^2 S t^2} \right)^{3/4} \right]$$

$$G(x) = \frac{8}{3}\left( \frac{x}{\pi} \right)^{-3}\sum_{N=1}^{\infty} \frac{\Gamma \left( \frac{3}{2} + \frac{2}{N} \right)}{\Gamma \left( \frac{3}{2} \right)} x^{N-1}$$

where $E_0$ is the total incident wave energy at $t = 0$ and is equal to $1$, $\delta$ is the Dirac function, $H(x)$ is the Heaviside unit step function, which is zero for $x < 0$ and one for $x > 0$, $t$ is the travel-time, $f$ is frequency, $r$ is the hypocentral distance, $\eta = \frac{2\pi f}{\gamma V_S}$, and $\eta_0 = \frac{2\pi f}{\gamma V_S}$ are the absorption and scattering coefficients, respectively. $G(x)$ can be approximated within a few percent of error as $G(x) \approx e^{\sqrt{1 + 2.0256x}}$. We adopted this approximation to model absorption and scattering attenuation in the region. Recently, several studies (Carcole & Sato 2010; Del Pezzo & Bianco 2010; Ugalde et al. 2010; Del Pezzo et al. 2011) used the Paaschens (1997) model to separate scattering and intrinsic absorption. Ugalde & Carcole (2009) compared the performance of the exact and approximate solutions of the 3-D radiative transfer equation. Based on the Paaschens (1997) model, eq. (1) is integrated over three successive time windows to obtain the theoretical energy as a function of distance, $r$ for given values of $\eta_i$ and $\eta_s$, as follows:

$$E_{i,th}(r) = \int_{t_i(r)}^{t_i+1(r)} E_0(r, t) dt_i, \quad i = 1, 2, 3,$$

where $t_i(r)$ represents the three successive time windows of 15 s duration starting from the S wave. The energies can also be parametrized in terms of $B_0$ and $L_e^{-1}$, defined as

$$L_e^{-1} = \eta_i + \eta_s, \quad \text{and} \quad B_0 = \frac{\eta_i}{\eta_i + \eta_s}.$$

We calculate $E_{i,th}(r, L_e^{-1}, B_0)$ for $i = 1, 2, 3$ after setting the appropriate range of variability for the inverse extinction lengths $L_e^{-1}$ and seismic albedos, that is, $2 \times 10^{-3} \leq L_e^{-1} (\text{km}^{-1}) \leq 0.1$ and $0.05 \leq B_0 \leq 0.999$. Finally, we compare the observed energy with the theoretical one for a given model to obtain $Q_t^{-1}$ and $Q_{sc}^{-1}$ following Bianco et al. (2002).

4 DATA ANALYSIS AND RESULTS

Waveform data used in this study are selected from local earthquakes which occurred between 2001 January and 2006 October. After a careful inspection of the data, we selected 320 high quality...
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4.1 MLTWA analysis

Each seismogram was bandpass filtered in the frequency bands with central frequencies of 1.5, 3, 6, 12 and 24 Hz. Then, the time integral of observed energy as a function of distance and frequency was calculated for three successive time windows, defined at 0–15, 15–30 and 30–45 s, starting from the S-wave onset. Fig. 2 shows an example of vertical-component seismogram recorded at station TEZ with the three successive time windows over which the energies are measured. The first time window (0–15 s) shows a relatively large data scatter than the latter two time windows. This effect has also been observed in other areas (Sato et al. 2012, section 8.1), which can be attributed to the effects of source radiation. Direct S waves are more affected by the source radiation in comparison to lately arriving coda waves, which sample greater volumes of the Earth and represent an average property over a large number of scatterers (Pujades et al. 1997).

In this study, we averaged the data in a moving window of 2 km in length, sliding 50 per cent of its length each step along the distance range. The averaged data consisted of 70 points, 2 km wide evenly spaced between 25 and 165 km distance. We estimated the best-fitting \( L_e^{-1} \) and \( B_0 \) values by minimizing the L2-norm-based misfit function between the theoretical energy for a given model and the
Figure 3. Observed integrated seismic energies from the vertical component seismograms for the 0–15 s (circles), 15–30 s (squares) and 30–45 s (plus symbols) lapse time windows calculated for the hypocentral distance range 25–165 km. The continuous lines represent the theoretical curves obtained from the Paaschens (1997) model. Frequency and corresponding best-fitting model parameters $L_e^{-1}$ and $B_0$ are shown in each subfigure.

The averaged energy over each time window following Hoshiba et al. (1991) as

$$M(L_e^{-1}, B_0) = \sum_{i=1}^{N} \sum_{k=1}^{3} [E_{k,obs}(r) - E_{k,th}(r)]^2,$$

where $E_{k,obs}(r)$ ($k = 1, 2, 3$) is the integrated energy measured from the observed data, $E_{k,th}(r)$ is the theoretical energy integrated in the lapse time interval, obtained by using eq. (2), $N$ is the number of data points in each energy window. The minimum of the misfit function $M(L_e^{-1}, B_0)$ corresponds to the best-fitting $L_e^{-1}$ and $B_0$ values. The total residual is obtained by summing the relative residual over each time window. Fig. 4 shows the total residuals normalized to their minimum value, for each frequency band. The results showed that the best fits to the data are obtained when the $L_e^{-1}$ and $B_0$ values are between 0.016 and 0.028 km$^{-1}$ and 0.32 and 0.55, respectively.

To estimate the errors in $L_e^{-1}$ and $B_0$, we used $F$ distribution for 68 degrees of freedom for each model estimate at the 90 per cent confidence level. Using the $F$-distribution at the 90 per cent confidence level, the ratio of two random variables each having 68 degrees of freedom is 1.3. The parameters $Q_c^{-1}$, $Q_s^{-1}$ and $Q_i^{-1}$ were obtained from the best-fitting $L_e^{-1}$ and $B_0$ values by using the following equations:

$$Q_i^{-1} = \frac{V_S}{2\pi f L_e}, \quad Q_s^{-1} = \frac{B_0}{Q_i}, \quad \text{and} \quad Q_i^{-1} = 1 - \frac{B_0}{Q_i}.$$

The expected coda-Q ($Q_{cexp}$) is calculated by using the following expression (Hoshiba 1991; Mayeda et al. 1992):

$$Q_{cexp}^{-1} = Q_i^{-1} \left[ 1 - \frac{C_2 + 2C_3(\eta_s V_s t) + 3C_4(\eta_s V_s t)^2 + \cdots}{1 + C_2(\eta_s V_s t) + C_3(\eta_s V_s t)^2 + \cdots} \right] + Q_i^{-1},$$

where $C_0$ is the coefficient for the 0th order scattering and $\eta_s$ is the scattering coefficient expressed in per kilometre. We cut-off at the 10th order for our analysis. Table 1 lists the values of
Figure 4. Residual map for each frequency band for the study area. Each point represents the ratio between the sum of squared residuals and the minimum solution (residual/minimum residual). The crosses denote the estimates within 90 per cent level of confidence using the $F$ distribution for 68 degrees of freedom, while the big star represents the minimum residual.

$L_e^{-1}$, $B_0$ and their uncertainties, $Q_i^{-1}$, $Q_s^{-1}$ and $Q_c^{-1}$ for five frequency ranges. Fig. 5 shows their frequency dependencies. The frequency dependent power-law forms of $Q^{-1}$ obtained are $Q_i^{-1}(f) = (0.007 \pm 0.001)f^{-1.16 \pm 0.05}$, $Q_s^{-1}(f) = (0.01 \pm 0.004)f^{-1.46 \pm 0.1}$ and $Q_c^{-1}(f) = (0.017 \pm 0.006)f^{-1.2 \pm 0.2}$ in 1.5–24 Hz for the study area.

5 DISCUSSIONS

In this work, we extend our earlier studies on attenuation of high-frequency seismic waves (Padhy & Subhadra 2010a,b) in NE India by resolving the relative contributions of scattering and intrinsic absorptions to the total attenuation using an extended data set. We applied the MLTWA for a spatially uniform medium to the earthquakes with hypocentral distance in the range 25–165 km. Using a grid search approach, we found the attenuation parameters that best fit the data to the model. The results generally support that $Q_i^{-1}$ and $Q_s^{-1}$ of this study with $Q_c^{-1}$ of the previous study (Padhy & Subhadra 2010a), which was calculated on the basis of single scattering theory. The $Q_c^{-1}$ obtained for the region with single scattering theory (Padhy & Subhadra 2010a) is similar to $Q_i^{-1}$ at high frequencies. Both $Q_i^{-1}$ and $Q_s^{-1}$ are higher than $Q_c^{-1}$ at low frequencies ($f < 4$ Hz), which might be attributed to use of different lapse times and hence different volumes sampled in both studies. For example, in this study we compute $Q_i^{-1}$ and $Q_s^{-1}$ in a time
Table 1. Values of inverse extinction length \( (L_e^{-1}) \), seismic albedo \( (B_0) \), absorption coefficient \( (\eta_i) \), scattering coefficient \( (\eta_s) \), total attenuation \( (Q_t^{-1}) \), scattering attenuation \( (Q_{sc}^{-1}) \) and intrinsic absorption \( (Q_i^{-1}) \) for each frequency band. The signed plus/minus entries for \( L_e^{-1} \) and \( B_0 \) are their error estimates within 90 per cent confidence interval obtained using the \( F \) distribution.

<table>
<thead>
<tr>
<th>( f (\text{Hz}) )</th>
<th>( L_e^{-1} ) (km)</th>
<th>( B_0 )</th>
<th>( \eta_i ) (km(^{-1}))</th>
<th>( \eta_s ) (km(^{-1}))</th>
<th>( Q_t^{-1} )</th>
<th>( Q_{sc}^{-1} )</th>
<th>( Q_i^{-1} )</th>
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<td>0.55</td>
<td>0.0126</td>
<td>0.0154</td>
<td>10.4E–03</td>
<td>5.7E–03</td>
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<tr>
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Figure 5. Variation of \( Q_i^{-1} \) and \( Q_{sc}^{-1} \), expected \( Q_C^{-1}(Q_{C,exp}) \) and measured \( Q_C^{-1} \) for different lapse times for the region obtained with a single scattering theory (Padhy & Subhadra 2010a) with frequency. The regression lines to the estimated \( Q_i^{-1}, Q_{sc}^{-1} \) and \( Q_{C,exp}^{-1} \) values for each frequency band are also shown.

window from \( r/V_S \) to \( r/V_S + 45 \) s, while Padhy & Subhadra (2010a) obtain \( Q_C^{-1} \) from \( 2t_S \) up to \( 2t_S + 60 \) s (\( t_S \) is S-wave traveltime). However, \( Q_{sc}^{-1} \) is lower than \( Q_C^{-1} \) at all lapse times analysed at high frequencies \( (f > 4 \text{ Hz}) \), the cause is not yet clearly understood.

The value of seismic albedo, \( B_0 \), is more than 0.5 at low frequencies \( (f < 3 \text{ Hz}) \) and is less than 0.5 at high frequencies \( (f > 3 \text{ Hz}) \). The high albedos found in the region suggest a high degree of lithospheric heterogeneity at the scale length corresponding to this frequency range. \( Q_{sc}^{-1} \) is found to be more frequency dependent than \( Q_i^{-1} \), the value of frequency dependent coefficient, \( n \) being 1.1, 1.4 and 1.2 for \( Q_i^{-1}, Q_{sc}^{-1} \) and \( Q_t^{-1} \), respectively. \( Q_{sc}^{-1}(f) \) decays very fast (faster than \( f^{-1} \)) as frequency increases, which suggests that the medium could be characterized by a Gaussian autocorrelation function. The size of heterogeneities is deduced from the frequency dependence of \( Q_{sc}^{-1}(f) \). Here, \( Q_{sc}^{-1}(f) \) dominates over \( Q_{C,exp}^{-1}(f) \) for the scale length corresponding to the 1–3 Hz frequency range. Supposing a Rayleigh scattering regime based on the assumption of isotropic scattering \( (a < \lambda, ka \text{ is small, where } k \text{ is the wavenumber, } a \text{ is the scale length of heterogeneity and } \lambda \text{ is the seismic wavelength}) \), the strong frequency dependence of \( Q_{sc}^{-1} \) occurs when the heterogeneities responsible for the scattering are, at least, comparable with the wavelength for the lowest frequencies analysed (around 1 km; Wu & Aki 1988).
The expected $Q_c^{−1}$ is nearly identical to the observed $Q_i^{−1}$ at high frequencies, in agreement with the simulation results (Frankel & Wennerberg 1987) and laboratory experiments (Matsunami 1991). However, the discrepancy between the two at low frequencies is relatively more than that at high frequencies. The discrepancies between the (i) observed and expected $Q_c^{−1}$ and (ii) $Q_s^{−1}$ and $Q_i^{−1}$ both at low frequencies have also been observed in earlier studies using the MLTWA for a uniform medium (Sato et al. 2012, section 8.1), which can be explained by a depth-dependent $Q_i^{−1}$ (Hoshiba et al. 1991; Mayeda et al. 1992; Bianco et al. 2005).

### 5.1 Influence of model assumptions on observed attenuation

We discuss the influence of model assumptions on the estimated attenuation parameters. First, the present model assumes a spherically symmetric source radiation. There exists a fluctuation in the normalized amplitudes from different sources due to different radiation patterns. Fig. 6 shows the variation of absorption, $\eta_i$ (Fig. 6a) and scattering $\eta_s$ (Fig. 6b) coefficients with frequency, where the coefficients $\eta_i$ and $\eta_s$ are obtained from the estimates of $L_s^{−1}$ and $B_0$ within the 90 per cent confidence interval using the $F$ distribution, as shown in Fig. 4, according to eq. (3). A wide distribution around the expected value of absorption, $\eta_i$ (Fig. 6a) and scattering $\eta_s$ (Fig. 6b) coefficients, especially at low frequencies could be related to the effect of source radiation and the complexities in the shallower part of the crust. As already mentioned, a relatively large scatter in data points, as observed in the first time window (Fig. 3), can be due to the effect of source radiation.

Second, we assume isotropic scattering in the model, which assumes that the wavelength is much larger than the scale length of the heterogeneity. At high frequencies, where the wavelength is much smaller than the dominant scale length of heterogeneity, more energy is scattered into the forward direction and less energy is scattered into the backward direction (Aki & Richards 1980). Thus, forward scattering ($ka < 1$) dominates at high frequencies. Consequently, the scattering coefficients obtained with an isotropic assumption, in a medium with strong forward scattering, are underestimated. Conversely, the scattering coefficients obtained with an isotropic assumption, in a medium with strong backscattering, are overestimated (Mayeda et al. 1992). This effect may possibly explain the observed high scattering coefficient at low frequencies in NE India.

Third, we assume a spatially uniform velocity and attenuation structure. Margerin et al. (1998) interpreted coda in terms of energy leakage into the mantle for a heterogeneous crust overlying a rather homogeneous mantle by numerically solving the radiative transfer equation. Results obtained using the MLTWA for a layered structure show that the attenuation coefficients obtained with the MLTWA for a uniform model are overestimated (Hoshiba et al. 2001; Bianco et al. 2005; Del Pezzo & Bianco 2010).

The results of this study are based on the MLTWA for a uniform half-space. The model used is relatively simple, the results of this study are found to be similar to that obtained using the same method for several seismically active regions of the world. The results based on the simple model gives an overall nature of lithospheric heterogeneities in NE India. The results represent an average and first-order estimate of the attenuation in the region. An accurate description of attenuation parameters and their frequency dependencies, obtained with a more realistic layered earth model for the study area, such as the number of layers and their thicknesses, velocity and attenuation for each layer is necessary for a correct interpretation of attenuation results, which is a subject of future study.

### 5.2 Correlation with the geological and tectonic features

Here we try to correlate the observed high attenuation of the region with the geological features and other geophysical results, such as receiver function analysis, seismic tomography, seismic reflection and resistivity studies.

As already mentioned, the NE India region belongs to an active subduction zone characterized by Himalayan and Burman collision zones along the India–Eurasia–Burma Plate boundaries with a strong lateral heterogeneity, and the presence of major tectonic elements such as the Himalayan frontal arc, Kopili fault, the SP, and Indo-Burma ranges (Fig. 1). Seismically, the region is characterized by high-$b$ value and high fractal dimension in the SP and along the Kopili fault (Bhattacharya et al. 2002) that accounts for high degree of medium heterogeneity. A strong attenuation of seismic waves is, therefore, expected for the region similar to other subduction zones worldwide. Our low-$Q_s$ (59 at 1 Hz for S-waves) estimates are highly correlated with low-$Q_s$ observed in our previous study (71 at 1 Hz for S-waves) in the region (Padhy & Subhadra 2010b) obtained with coda-normalization.

Several authors studied the structure of the lithosphere beneath the region from the controlled-source seismic (Hirn & Sapin 1984;
Lepine et al. (1984) and receiver function analysis (Kumar et al. 2004; Mitra et al. 2005). They found a strong lateral variation in crustal structure with a thin (<35–38 km) crust near the Plateau, a thicker crust beneath the Bengal basin (<44 km) and near the Assam Valley (<41 km) with a dipping Moho. Such a large variation in crustal structure is also supported from gravity variations suggesting a dipping Moho (~150 mGal at the MBT to ~300 mGal beneath the Great Himalaya (Kono 1974). Bilham & England (2001) proposed the pop-up structure for the plateau to explain its thin crust and high elevation. The above structural heterogeneities may likely give rise to the observed high scattering at low frequencies.

As mentioned, intrinsic absorption could be caused by several factors, such as thermal, compositional anomaly, the presence of melts/fluids, etc. Here we suggest possible factors causing increased intrinsic absorption based on other geophysical evidences. First, intrinsic absorption is temperature dependent. Assuming the attenuation to be entirely due to thermally induced intrinsic absorption, and the lower crustal gabbro lithology, the $Q_i$ estimates can be used to obtain a first order estimate for temperature of the lower crust (Menke et al. 1995), either by using the empirical relationship between temperature and attenuation or by comparing our $Q_i$ estimates with the experimentally determined $Q_i$ values for different temperature gabbros (Kampfmann & Berckhemer 1985). The results of comparison show an upper limit of lower crustal temperature at about 400 °C below the gabbro solidus of the 1200 °C isotherm (i.e. ~800 °C). Near solidus temperatures are ruled out because the attenuation of gabbro at its solidus (~1200 °C) is very high ($Q_i$ ~ 20) compared to our $Q_i$ estimates (Kampfmann & Berckhemer 1985). Also, because factors other than temperature (e.g. scattering from heterogeneities) also contribute to the observed attenuation, our temperature estimate represents an upper limit. This temperature estimate for the study region is consistent with that for continental lithosphere at about 50 km depth, obtained on the basis of laboratory studies of rocks (Artemieva et al. 2004). However, a more accurate estimate of temperature requires information on mineralogical composition of the underlying rock (Kampfmann & Berckhemer 1985).

Second, tomographic imaging of $P$ velocity ($V_P$) in the region shows prominent low-$V_P$ zones both in the crust and uppermost mantle (Kayal & Zhao 1998; Bhattacharya et al. 2008). These velocity anomalies include a low-$V_P$ zone between the SP and Mikir hills at the lower crust down to 35 km depth, which corresponds to the Kopili fault, and a low-$V_P$ zone in the Bengal basin, south of the SP, extending down to a depth of ~20 km, attributed to its thick alluvium sediments (Kayal & Zhao 1998; Bhattacharya et al. 2008). These low velocity anomalies are supported by gravity and deep drilling studies (Verma & Mukhopadhyay 1977; Metivier et al. 1999) and are correlated with anomalously low resistivity values (reduced by 40–50 per cent) near the SP obtained from deep resistivity measurements (Kayal & Banerjee 1988). Tiwari et al. (2004) showed the presence of subsurface fluids in faults and fracture zones from the fluctuations in pore pressure along active fault zones in the region, based on nonlinear dynamics analysis. Based on geochemical studies in the region, Walia et al. (2010) showed an increase in Radon emanation rate in the SP. All these characteristic features may suggest the likely presence of fluids causing an increased intrinsic absorption in the region.

The results of comparison show that NE India is more heterogeneous than southern India (Tripathi & Ugalde 2004), south central Alaska (Dutta et al. 2004), NE Italy (Bianco et al. 2005) at all the studied frequencies and Kanto-Tokai, Japan (Felher et al. 1992), Canary Islands (Canas et al. 1998), NW Turkey (Bindi et al. 2006), and Garhwal Himalaya (Mukhopadhyay et al. 2010) at frequencies below 5 Hz (Fig. 7a). Similarly, in view of $Q_{isci}^{-1}$, the region seems to be more attenuative than southern India (Tripathi & Ugalde 2004), south central Alaska (Dutta et al. 2004), NE Italy (Bianco et al. 2005), and Garhwal Himalaya (Mukhopadhyay et al. 2010) at all analysed frequencies; and less attenuative than Kanto-Tokai, Japan (Felher et al. 1992) and Nuevo Cuyo, south central Andes (Badi et al. 2009) at frequencies above 3 Hz (Fig. 7b). Thus the true frequency dependences of $Q_{isci}^{-1}$ and $Q_{isci}^{-1}$ strongly depend on geology and tectonic setup of the region, although in general both decrease with frequency.

An interesting comparison can be made with $Q_{isci}^{-1}$ and $Q_{isci}^{-1}$ obtained for different parts of India using the same method at similar distance range. A few studies separated $Q_{isci}^{-1}$ and $Q_{isci}^{-1}$ in India using MLTWA. For example, Tripathi & Ugalde (2004) showed a predominant $Q_{isci}^{-1}$ in 1–10 Hz for southern India, $Q_{isci}^{-1}$ in 1–2 Hz and $Q_{isci}^{-1}$ in 2–30 Hz in Kachchh, western India (Ugalde et al. 2007), $Q_{isci}^{-1}$ in 1–16 Hz in Garhwal Himalaya, northern India (Mukhopadhyay et al. 2010), while this study shows a dominant $Q_{isci}^{-1}$ below about 3 Hz and $Q_{isci}^{-1}$ above 3 Hz. A common feature to all these aforementioned studies is that scattering dominates at frequencies below about 3 Hz. This suggests a more heterogeneous crust, consistent with the tectonic settings of the area with a characteristic heterogeneity scale length of about 1 km. At frequencies more than 3 Hz, however, their attenuation characteristics vary, that may be related to the diverse tectonic settings across the northern (seismically active) and southern (seismically stable) parts of India. Although this study uses a simple model, the results obtained underlying this model has improved our knowledge on attenuation mechanisms in the region. However, it would be necessary to consider a more realistic model including depth-dependent velocity and attenuation parameters to have a more accurate description of attenuation relations and their frequency dependences. This is attempted in a future ongoing study.

6 CONCLUSIONS

We separated intrinsic absorption and scattering attenuation using the MLTWA for a spatially uniform medium. The MLTWA results show that scattering dominates at low frequencies ($f < 3$ Hz), while intrinsic absorption dominates at high frequencies ($f > 3$ Hz). The frequency dependence of $Q_{isci}^{-1}$ corresponds to the scale length of lithospheric heterogeneity of about 1 km. The observed $Q_i$ for the study region obtained with a single scattering theory is close to $Q_i$ in agreement with theoretical prediction for a spatially uniform medium. However, a discrepancy exists between the two at low frequencies, which suggests that it would be necessary to consider a model with non-isotropic scattering and a depth-dependent velocity structure to obtain an accurate description of attenuation in NE India. The observed attenuation implies a broad region of sub-solidus with an upper limit of lower crustal temperature of ~800 °C assuming a gabbroic lithology.

5.3 Comparison with previous results

In Fig. 7, we compare our results with previous ones in different areas investigated with the same technique at similar distance range.

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Figure 7. Comparison of (a) $Q_s^{-1}$ and (b) $Q_c^{-1}$ estimated for various regions in the world. Regions are south central Alaska (0–250 km; Dutta et al. 2004); Garhwal Himalayas (10–175 km; Mukhopadhyay et al. 2010); southern India (0–255 km; Tripathi & Ugalde 2004); Nuevo Cuyo, southern central Andes (90–350 km; Badi et al. 2009); NE Italy (0–200 km; Bianco et al. 2005); Kanto-Tokai, Japan (0–230 km; Felner et al. 1992); NW Turkey (10–140 km; Bindi et al. 2006); Canary Islands (0–140 km; Canas et al. 1998), NW Colombia (0–255 km; Vargas et al. 2004), Kachchh, western India (Ugalde et al. 2007; 0–110 km) and NE India (25–165 km; this study).

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