CHAPTER 4
STUDY OF 3-D VELOCITY STRUCTURE IN NORTHEASTERN INDIA BY SEISMIC TOMOGRAPHY

4.1 Introduction

Seismic tomography is the technique of mapping or imaging the three-dimensional variation of the earth’s heterogeneous structure. It actually measures the heterogeneity and/or attenuation structure of the earth by using seismic waves from natural sources (occurrence of earthquakes) and artificial sources (explosions and vibroseis) and in the process brings out the 3-D image of the heterogeneity of the earth crust or the entire lithosphere. In tomographic techniques seismic rays are back projected to its source. An image of an object which is the outcome of inversion of data generated by an experiment may be defined as tomography. Using this technique two or three dimensional earth’s structure can be imaged from a set of observations at the periphery of a targeted Earth volume. Properties of seismic rays such as arrival times of a dense set of rays criss-crossing and sampling a volume of a body are used to locate anomalies in space (which could be anomalies in seismic velocity or attenuation) and to find their extent. Therefore, the size and homogeneity of the ray set is unique to this technique. Since it is a complex non-linear problem, as such the problem is linearized in some manner, especially by adding a-priori constraints on the solution to make it solvable, and the velocity is computed by a matrix inversion.

Seismic tomography is an extensively used method for exploring the lateral variation of seismic wave velocities in the Earth’s crust. In this method a set of significant earthquakes which are locally recorded are characterized. Then these earthquakes are considered to enlighten the earth’s interior with seismic waves. The
arrival time of the waves at seismic stations can then be used to calculate the wave velocities through the Earth’s lithosphere. Seismic tomography can roughly be divided into two categories namely the travel time tomography and the full–waveform tomography. Travel time tomography uses the arrival times of the seismic data while the full-waveform tomography uses the waveform data including amplitudes and phases.

Local earthquake tomography (LET) is a technique to get a clear picture about the tectonic processes, which may have occurred in the past within a crustal volume. LET may offer a much higher spatial resolution of the structure imaged because the seismic sources are located within the modelling volume and so the seismic ray density sampling the anomalous body is much higher. Apart from this higher wave frequency and closer station spacing also contributes towards the high resolution. It is a powerful technique to obtain insights into the tectonic processes within a crustal volume.

4.2 Empirical Formulation of Local Earthquake Tomography (LET)

Using the ray theory, the following path integral indicates the body-wave travel time $T_{ij}$ from an earthquake $i$ to a seismic station $j$ (Thurber, 1983);

$$T_{ij} = \int_{sources}^{receiver} u ds$$  \hspace{1cm} (4.1)

Where $u$ is the slowness field (reciprocal of velocity) and $ds$ is an element of path length. The actual observations in local earthquake tomography are the arrival times $t_{ij}$, where

$$t_{ij} = \tau_i + T_{ij}$$  \hspace{1cm} (4.2)
and \( \tau_i \) indicates origin time. The receiver locations and the observed arrival times are the only known parameters in the LET problem. In order to initiate the LET programme a forward problem is first solved. In this arrival times \( t_{ij} \) to different stations from an earthquake are calculated using an assumed velocity model and assumed hypocentral location and origin time. The difference in the observed and calculated arrival time \( r_{ij}^{cal} \) is called travel time residual (Equation 4.3). In successive iterations the assumed velocity model, location and origin time of earthquake are systematically changed so that this residual is minimized. Arrival times can suffer from significant uncertainty. The unknown model parameters are the source coordinates \((x_1, x_2, x_3)\), origin times, ray paths and slowness field. Equations (4.4) and (4.5) determine the calculated arrival times \( t_{ij}^{cal} \) from using trial hypocenters and origin times, and an initial model of the seismic velocity structure (a priori information), in a given a set of arrival times \( t_{ij}^{obs} \) measured at network of stations (times of first P and or S waves). The nonconformity between observed and ‘predicted’ (calculated) arrival times are the residuals \( r_{ij} \) where

\[
r_{ij} = t_{ij}^{obs} - t_{ij}^{cal}
\]  

(4.3)

The linear approximation below indicates the relation of the residuals to perturbations to the hypocenter and velocity structure parameters.

\[
r_{ij} = \sum_{k=1}^{3} \frac{\partial T_{ij}}{\partial x_k} \Delta x_k + \Delta \tau_i + \int_{\text{source}}^{\text{receiver}} \partial u ds
\]  

(4.4)

where the hypocenter partial derivatives are proportional to the components of the ray vector times the seismic slowness at the source. Simultaneous inversion relating the travel time residual ‘r’ to model parameter, using a finite parameterization changes, can be represented by the below linearized equation
\[
\begin{align*}
  r &= \Delta t_e + \frac{\partial t}{\partial x_e} \Delta x_e + \frac{\partial t}{\partial y_e} \Delta y_e + \frac{\partial t}{\partial z_e} \Delta z_e + \sum_{n=1}^{N} \frac{\partial t}{\partial v_n} \Delta v_n \\
  \text{(4.5)}
\end{align*}
\]

where $\Delta t_e$, $\Delta x_e$, $\Delta y_e$, $\Delta z_e$, and $\Delta v_n$ are perturbations to the hypocentral parameters (earthquake origin time and location) and the velocity parameters, and $\partial t / \partial x_e$, $\partial t / \partial y_e$, and $\partial r / \partial v_n$ are partial derivatives of the arrival time with respect to the earthquake coordinates and velocity parameters respectively. $N$ is the total number of velocity parameters. There is one such equation for each observed arrival.

An approximate ray tracing method is used to calculate the required travel times (equation 4.5) As represented below, a simple interpolation function is used to calculate the velocity at a given point $(x,y,z)$ (Thurber, 1983).

\[
V(x, y, z) = \sum_{i=1}^{2} \sum_{j=1}^{2} \sum_{k=1}^{2} V(x_i, y_j, z_k) \left[ \left( 1 - \left| \frac{x - x_i}{x_2 - x_1} \right| \right) \left( 1 - \left| \frac{y - y_j}{y_2 - y_1} \right| \right) \left( 1 - \left| \frac{z - z_k}{z_2 - z_1} \right| \right) \right]
\]

\[
\text{(4.6)}
\]

where $x_i, y_j, z_k$, represent the coordinates for the eight grid points surrounding the point $(x,y,z)$. This is a continuous function, which is a product of linear functions in $x$, $y$, and $z$. The partial derivatives in equation 4.5 can be calculated, given the velocity model and the ray path from the earthquake to the observing station. The hypocentral partial derivatives satisfy (Lee and Stewart, 1981; Thurber, 1981),

\[
\begin{align*}
  \frac{\partial t}{\partial x_e} &= \frac{1}{v_e} \frac{dx}{ds} \\
  \frac{\partial t}{\partial y_e} &= \frac{1}{v_e} \frac{dy}{ds} \\
  \frac{\partial t}{\partial z_e} &= \frac{1}{v_e} \frac{dz}{ds}
\end{align*}
\]

\[
\text{(4.7)}
\]
where \( v_e \) is the velocity at the earthquake source (hypocenter), \( ds \) is the parameter of path length, and \( dx/ds, dy/ds \) and \( dz/ds \) are components of the unit vector tangent to the ray path at the hypocenter and pointing in the direction of ray propagation. This equation is valid for hypocentral perturbations that are small compared to the scale of velocity variations. The velocity parameter derivatives cannot be computed exactly because they involve integrals along the ray path

\[
\frac{\partial t}{\partial v_n} = \int_{source}^{station} \left( \frac{1}{V(x,y,z)} \right)^2 \frac{\partial V(x,y,z)}{\partial v_n} ds
\]  
\( (4.8) \)

where \( v_n \) is the \( n \)th velocity parameter. In practice, the partial derivative with respect to slowness is utilized for computational simplicity.

\[
\frac{\partial t}{\partial u_n} = \int_{source}^{station} \frac{\partial U(x,y,z)}{\partial u_n} ds
\]  
\( (4.9) \)

Usually all LET methods begin with equation (4.5) and then diversify to some extent, based on differing treatments of some or all of the following aspects of the problem.

(a) The scheme for the representation of the velocity structure,

(b) The technique for ray-path and travel time calculation.

(c) The treatment of the hypocenter velocity structure coupling

(d) The method of inversion and iteration

(e) The assessment of solution quality.

(f) The use of S waves.

The velocity \( V(x,y,z) \) and its partial derivative with respect to a model parameter can be calculated throughout an interpolation scheme and apparently the minimization of
the travel time residuals involves the solution of the forward and inverse problem during an iterative process, the above can be written in matrix notation as

\[ \Delta d = G \Delta m \]  \hspace{1cm} (4.10)

where \( G \) is the Jacobian matrix containing all the partial derivatives in equation (4.5), \( d \) are the residuals and \( \Delta m \) the perturbations of the model parameters. Since, the problem of passive tomography is usually underdetermined or mixed determined, the damped least-squares method is applied.

\[ \Delta m = \left( G^T G + \varepsilon^2 I \right)^{-1} G^T d \]  \hspace{1cm} (4.11)

where \( \varepsilon^2 \) is the damping factor and \( I \) is the identity matrix. Here the matrix size is fixed by the number of velocity model parameters and does not increase with the number of earthquakes included in the inversion (Spencer and Gubbins, 1980; Thurber, 1983; Thurber, 1993). The LET algorithm originally developed by Thurber (Thurber, 1981; Thurber, 1983; Thurber, 1993) uses this approach. The disadvantages of this method are: i) the loss of singular value information, ii) the sensitivity of the solution to the choice of damping parameter and iii) squaring of the condition number of the matrix to be inverted.

A reasonable damping value may be numerically obtained by running one step inversions with varying damping and constructing ‘trade-off’ curves comparing the data variance (residual size measure) and solution variance (model perturbation size measure). The optimum damping value is taken to be the one below which solution variance would increase rapidly with little or no decrease in data variance (Eberhart-Phillips, 1986; Eberhart-Phillips, 1993).
4.3 Steps in Local Earthquake Tomography (LET)

This method accomplishes a result through iterative refinement of liberalized solutions to a highly non-linear problem. The following steps are followed in the process of iteration (Thurber, 1983)

a) A model appropriate for the area being investigated for the P-wave velocity structure of the earth is assumed or if 1-D velocity model is available then that is used;

b) An initial estimate of the earthquake location and origin time (hypocentral parameters) is made;

c) Theoretical P-wave travel times are calculated from the estimated location to the observing stations, and the corresponding arrival time residuals are computed;

d) Partial derivatives of the travel times with respect to the variations in the earthquake location and origin time (hypocentral parameters) and with respect to variations in the velocity structure parameters are calculated;

e) For a set of earthquakes, a system of linearized equations is solved using inverse theory (inversion of seismic phases) to find adjustments to both the hypocentral and the velocity parameters, which minimize the arrival time residuals, obtaining a new estimate of the earthquake locations and subsequently the velocity structure.

4.4 Seismic Inversion and Tomography

Given a set of observed travel times, $t_1, \ldots, t_m$, from $m$ source-receiver pairs in a medium of slowness $S(X)$ and $P_i$ being the ray path connecting the $i$th source-receiver pair, neglecting observational errors, it can be written (Berryman, 1991)
\[ \int_{P_i} S(X) dl^{Pi} = t_i, \quad i=1, \ldots, m \] (4.12)

Assuming a block model of slowness, if \( l_{ij} \) is the length of the \( i \)th ray path through the \( j \)th cell then

\[ l_{ij} = \int_{P_i \cap \text{cell}_j} dl^{Pi} \] (4.13)

Assuming a model with \( n \) cells, Eq. 4.12 can then be written as

\[ \sum_{j=1}^{n} l_{ij} s_j = t_i, \quad i=1, \ldots, m \] (4.14)

Since a given ray path will in general intersect only a few of the cells in the model, for any given \( i \), the ray path lengths \( l_{ij} \) are zero for most of the cells \( j \). Figure 4.1 illustrates ray path segmentation for a 2-D cell model.

Equation 4.14 can be re-written in a matrix notation by defining the column vectors \( s \) and \( t \), and the matrix \( M \) is follows:

\[
S = \begin{pmatrix}
S_1 \\
S_2 \\
\vdots \\
S_n
\end{pmatrix}, \quad t = \begin{pmatrix}
t_1 \\
t_2 \\
\vdots \\
t_m
\end{pmatrix}, \quad M = \begin{pmatrix}
l_{11} & l_{12} & \cdots & l_{1m} \\
l_{21} & l_{22} & \cdots & l_{2m} \\
\vdots & \vdots & & \vdots \\
l_{n1} & l_{n2} & \cdots & l_{nm}
\end{pmatrix}
\] (4.15)

Equation 4.14 then becomes the basic equation of forward modeling for ray equation analysis.

\[ Ms = t. \] (4.16)

Travel – time for ray path \( i \):

\[ t_i = \sum_{j=1}^{16} l_{ij} s_j \] (4.17)
4.5 Ray Tracing Methods

Ray tracing is a two-point boundary value problem (BVP) since the end points of a ray are the source and receiver positions. Shooting, bending approximate or finite difference are the various available ray tracing methods.

The BVP is solved iteratively by solving an initial value problem in shooting method. This is done by keeping one end point fixed and varying the initial ray trajectory (Figure 4.2a). The ray path that reaches the other end point or comes closest to it is taken to be the required ray path (Koch, 1985; Lin and Roecker, 1990; Sambridge and Kennet, 1990). In the bending method (Figure 4.2b) the two end points are kept fixed and the path connecting these points is perturbed iteratively to obtain the minimum time ray path through the given velocity model (Wesson, 1971; Julian and Gubbins, 1977; Pereyra et al., 1980). It may be possible to obtain a global minimum path with negligible amplitude or a local minimum path instead of a global
one in both shooting method and bending method. The following studies have been
done in this direction:

- An approximate bending method was used by Thurber (1983) in which the ray
  paths are arcs of circles, i.e. paths of constant curvature
- An approximate but fairly accurate treatment of exact bending method was
done in the pseudo-bending method of Um and Thurber (1987) and Prothero et
  al. (1988)
- More advanced methods that can estimate the global minimum ray path are the
  finite–difference method of Vidale (1990) and the network theory method of
  Moser (1991). A basic estimate of arrival amplitude can also be made in
  Vidale’s (1990) method.

![Figure 4.2](image.png)

**Figure 4.2** Schematic representations of the (a) shooting method and (b) bending
method. Darker lines represent the minimum time ray path (Thurber, 1981).
4.6 Literature Review

Several workers have attempted to study the velocity structure of northeastern region of India previously. Mention may be made about the crustal P-wave velocity and velocity-ratio study in northeastern India by De and Kayal (1990). They studied the upper crustal P-wave velocity beneath Shillong plateau and Nagaon area and they came out with a generalised seismic velocity of the area using analog data which was later used for microearthquake location. Kayal and De (1987), has carried out the Pn velocity study using temporary seismographic network in Meghalaya Plateau. The analog data that was generated out of a temporary seismographic network was used to study the Pn velocity for uppermost mantle beneath the Shillong plateau. The average Pn velocity was found to be high compared to the average mantle velocity beneath other areas of the Indian subcontinent. High Pn velocity has been correlated to the high density materials beneath the plateau. A three dimensional seismic structure beneath Shillong Plateau and Assam Valley for northeastern India has been worked out by Kayal and Zhao (1998) using analog data. Tomography method was used to determine 3-D velocity structure of the crust and upper mantle in this area. The result revealed significant lateral heterogeneities. Also it was found that the tomographic images are compatible with major tectonic features such as active faults and seismicity trends. Mukhopadhyay et al. (1997), studied the crustal properties in the epicentral tract of the great 1897 Assam earthquake, northeastern India using analog data. The velocity structure for the upper and middle crust in the epicentral tract of the great 1897 Assam Earthquake (western half of the Shillong Plateau) was estimated using locally recorded micro earthquake data. Combining the estimated velocity structure with their previous work on seismicity it was proposed that the upper homogeneous crustal layer in the Shillong Plateau may be moving relatively
southward across an intracrustal thrust zone. Using both analog and digital data run on telemetric system, Bhattacharya et al. (2008), worked on the seismic structure of the northeastern India region to understand its local and regional tectonics. An attempt had been made to estimate the 3-D P-wave velocity structure of northeastern India region using the first arrival data of local earthquakes that were recorded by about 77 temporary/permanent local both digital and analog seismic stations during three months. A 1-D velocity model was developed based on travel-time RMS residual $\leq 0.49$ s and azimuthal gap $\leq 180^\circ$. Using this 1-D velocity model as a priori velocity model, a 3-D velocity model was developed which significantly improved the locations of previously located 980 earthquakes. The reconstructed P-wave velocity (Vp) structure, with the relocated events, reveal strong heterogeneity in lateral as well as in vertical directions corresponding to local and regional geology/tectonics of the region.

The present study comprehensively covers the entire northeastern India to study the 3-D velocity structure by using digital data that ensures accuracy of timing parameter as the seismic stations used are impinge with GPS timing system and that has manifest itself in more accuracy of results which helped in drawing better conclusion. Moreover, the digital data have been inverted to get the one-dimensional inverted optimum velocity model for the NE region of India by using local earthquake travel time tomography (LET) method given by Thurber (1983) and later modified by Eberhart-Phillips (1993). This “1-D inverted optimum velocity model has been fed into the tomoDD programme (Zang and Thurber, 2003) as a priori velocity model for 3-D tomographic inversion.
4.7 Methodology-Hypocentral Parameters and Seismic Tomographic Techniques Utilized in this Study

A vital exercise in any seismological research is a precise location of earthquakes which is fundamental. It is the determination of hypocentral coordinates and the source origin time. This is a complex process of nonlinear process of inverse modeling with velocity structure and hypocentral coordinates, wherein the observed phase data available in the form of phase arrival times must be solved for determination of hypocentral location that are consistent with the observed data. Location of an earthquake demands for the following:

1. Station Coordinates (seismometer locations),
2. A velocity model- to have an idea about the crustal structure beneath the network,
3. Control parameters- to constrain the programme, and
4. Phase data-from the seismograms.

Accurate determination of hypocentral parameters mainly depends on precise picking of phases supplemented by a preliminary true crustal velocity model as the input to the hypocentral location programme. Any seismological research initiative pertains to accurate epicentral location. Since location programme exclusively depends on velocity structure, so precise determination of velocity structure still remains a bigger challenge so far seismological studies is concerned. Although several studies on estimation of crustal velocity structure (De and Kayal, 1990; Kayal and Zhao, 1998; Bhattacharya et al., 2005) have been carried out in northeastern India. These studies are made based on analog database which lacks time accuracies; they have their own limitations to divulge much on the detailed 3D velocity structure in the region. With the advent of digital seismic network, there remains an excellent
scope to re-estimate the crustal velocity structure with better accuracy in northeastern India. Therefore, one of the objectives of this study is to estimate a better version of accurate velocity structure for the northeastern region of India by means of three-dimensional inversion of travel times, using an updated imaging technique.

The earthquake location programmes and the tomographic techniques employed and/or have some implications in this study have been described in the sections below as follows:

**4.7.1 Hypoinverse Earthquake Location Programme**

The Hypoinverse programme is an earthquake location programme and similar to HYPO 71 in terms of the basic input requirements such as station particulars and the phase data. The programme uses a generalised inverse method to solve for the adjustment vector (dt, dx, dy, dz) rather then stepwise multiple regression. To save space the programme uses a condensed station table and therefore, the coordinates of some reference origin near the centre of the array must be specified and this forms the first leg of information. Format of the phase data card is same as for HYPO71, however, if the type of phase arrival is left blank the phasewill be skipped. The phase will be also skipped if the arrival time is more than 4 minutes from the first card. Apart from this the version 1 of the programme accepts up to 3 crustal models, each of which may have up to 12 homogeneous layers including half space. A particular station uses the same crustal model. A feature of this programme is it can read and also use a travel time table generated independently of the location process.

**4.7.2 Double–Difference Earthquake Location Algorithm**

The double-difference earthquake location algorithm was developed by Felix Walhauser and William L. Ellsworth in the year 2000. This algorithm belongs to the same class of relative earthquake location method (Got *et al.*, 1994). The basic notion
on which this technique is based on the fact that if the hypocentral separation between two earthquakes is small compared to the event-station distance and the scale length of velocity heterogeneity, then the ray paths between the source region and a common station are similar along almost the entire ray path (Fréchet, 1985; Got et al., 1994). Taking into consideration this assumption, the difference in travel times for two events observed at one station can be attributed to the spatial offset between the events with high accuracy and so velocity model errors are minimized without the use of station correction. The residual between observed and calculated differential travel time for the two events is calculated as

$$ dr_{k}^{ij} = (t_{k}^{i} - t_{k}^{j})^{obs} - (t_{k}^{i} - t_{k}^{j})^{cal} $$

(4.18)

where $t$ is the travel time, superscript $i$ and $j$ corresponds to the two different events and the subscript $k$ corresponds to a particular observation (i.e one particular phase to one common station).

To determine high-resolution hypocentre locations over large distances an efficient method is by the application of the double-difference earthquake location algorithm. This method is based on the application of ordinary absolute travel-time measurement and/or cross-correlation $P$-and $S$-wave differential travel-time measurements. For pairs of earthquake at each station, residuals between observed and theoretical travel-time differences (or double-differences) are minimized while linking together all observed event-station pairs. The vector difference between hypocentral pairs are adjusted iteratively and a least-square solution is found. The errors which arise due to unmodeled velocity structure are minimized by the double-difference algorithm without the employment of station correction. It is because of the fact that catalog and cross-correlation data are combined into one system of equations,
Interevent distances within multiplets are determined to the accuracy of the cross-correlation data. On the other hand relative locations between multiplets and uncorrelated events are simultaneously determined to the accuracy of the absolute travel-time data. To make an estimate of data accuracy and location errors, statistical resampling methods have been used and the uncertainties are improved by more than an order of magnitude compared to catalog locations. The algorithm has been tested on the northern Hayward fault, California where it collapses the diffuse catalog locations into sharp images of seismicity and reveals horizontal lineations of hypocentres that define the narrow regions on the fault where stress is released by brittle failure.

Earthquake catalog and/or digital waveform data provided by almost any seismic network can apply this method. The method has the ability to maintain consistency to relocate the seismicity with high resolution along an entire stretch of a fault system which opens a plethora of research scopes from the level of an individual earthquake to the scale at which tectonic processes takes place. The double-difference algorithm finds its application in the hypoDD earthquake relocation computer programme.

The double difference equations are derived from differentiating Geiger’s equation for earthquake location (Geiger, 1912), and have the form:

\[
\frac{\partial t_k^i}{\partial m} \Delta m^i - \frac{\partial t_k^j}{\partial m} \Delta m^j = d r_{ij}^{k}
\]

(4.19)

Where \(\Delta m (\Delta x, \Delta y, \Delta z\text{ and } \Delta \tau)\) is the change in the four hypocentral parameters of the events. The above equation can be expanded as

\[
\frac{\partial t_k^i}{\partial x} \Delta x^i + \frac{\partial t_k^i}{\partial y} \Delta y^i + \frac{\partial t_k^i}{\partial z} \Delta z^i + \Delta \tau^i - \frac{\partial t_k^i}{\partial x} \Delta x^j - \frac{\partial t_k^i}{\partial y} \Delta y^j - \frac{\partial t_k^i}{\partial z} \Delta z^j - \Delta \tau^j = d r_{ij}^{k}
\]

(4.20)
This equation relates the residual, $dr$, between observed and theoretical travel time difference between the two events (Eq 4.18) to changes in each event’s location $(x,y,z)$ and origin time $(\tau)$ (Kayal, 2008).

In double-difference earthquake location method, the partial derivatives in equation (4.20) are estimated for each event-station pair combination, and the resulting equations build up a system of linear equations of the form:

$$W\mathbf{Gm} = W\mathbf{d} \quad (4.21)$$

where $\mathbf{G}$ is the matrix of the partial derivatives, $\mathbf{m} (\Delta x, \Delta y, \Delta z \text{ and } \Delta \tau)^T$ the matrix containing the changes in the hypocentral parameters required to improve the model fit to the data, $\mathbf{d}$ the data vector containing the double differences estimated through equation (4.18) and $W$ a diagonal matrix used to weight equations (Kayal, 2008).

The above equations are finally solved by the conjugate gradient algorithm LSQR (Paige and Saunders, 1982) or by Singular Value Decomposition (SVD) method depending on the large or small amount of data respectively (Kayal, 2008).

The modification of double-difference location algorithm over Joint Hypocentre Method (JHD) is the latter gives a diffused picture of seismicity, whereas double-difference earthquake location algorithm brings out the structural details such as the location of active fault planes very conspicuously. To sum up the double-difference technique uses any combination of ordinary phase picks from earthquake catalogs (referred to as catalog data) and/or high-precision differential travel times from phase correlation of P- and/or S-waves (cross-correlation data). The catalog data is expressed as differential travel times so that the same equation is used for both types of data. To link together all possible pairs of locations for which data is available, travel time differences are generated. The scheme of dynamic weighting accommodates different qualities of data and measurement accuracies, such that the
inter-event distances within clusters of correlated events (multiplets) may be determined to the accuracy of differential travel-time data, whereas relative locations between the multiplets and uncorrelated events are determined to the accuracy of the catalog data (Wadlhauser, 2001).

4.7.3 hypo DD Earthquake Relocation programme

The hypoDD is a computer programme designed by Felix Waldhauser in the year 2001, meant for earthquake relocation. It is based on the double-difference algorithm for earthquake location given by Waldhauser and Ellsworth (2000). This earthquake relocation programme calculates travel times in a layered velocity model (where velocity depends only on depth) for the hypocentral locations at the station where the phase was recorded. Using the method of Singular Value Decomposition (SVD) or the conjugate gradients method (LSQR, Paige and Saunders, 1982), the double-difference residuals for pairs of earthquakes at each station are minimized by weighted least squares. The vector difference between nearby hypocentral pairs are iteratively adjusted to obtain solutions, with the locations and partial derivatives being updated after each iteration.

The hypo DD earthquake relocation programme works in two steps. The first step involves determination of travel time differences for pairs of earthquakes by analyzing both catalog phase data and/or waveform data. The scrutinizing of the data is a vital exercise as it ensures the minimization of the redundancy in the data set. The differential travel time data obtained from step 1 is utilized for the derivation of double-difference hypocentre locations. In carrying out this step, hypoDD solves for hypocentral separation after ensuring that the network of vectors connecting each earthquake to its neighbours has no weak links that would lead to numerical instabilities (Wadhauser, 2001).
4.7.4 Double-Difference Seismic Tomography

The double-difference (DD) seismic tomography programme was developed by Hijiang Zhang (2003) as a part of his Ph.D programme in Geophysics. The programme takes into account both absolute arrival times and differential travel times.

It considers the path anomaly biases explicitly such that it enables the DD tomography to determine the absolute and relative event locations and velocity structure with more accuracy by directly using the more accurate differential travel times from catalog and/or waveform cross correlation (WCC) data. DD tomography has applicability for both local and regional scale problems. In the case of local scale (10s to 100s of kilometers) problem the earth is treated with a Cartesian (flat-Earth model) and a pseudo-bending ray tracing algorithm to find rays and travel times between events and stations. At a regional scale (100s to 1000s of kilometers), the sphericals shape of the earth is taken care of by parameterizing a spherical surface inside a Cartesian volume of grid nodes. The velocity discontinuities of the earth such as Conrad, Moho and subducting slab boundary have been dealt with by finite-difference ray tracing algorithms.

The local scale DD tomography algorithm was tested on a synthetic dataset on an idealized model of the velocity structure of the San Andreas fault in central California which produced more accurate results for earthquakes located compared to the ones located with the help of DD location and standard tomography. The true velocity structure was better represented by DD tomography than by standard tomography. The DD tomography algorithm when applied for regional scale problems such as to image the seismic velocity structure of the subducting plates beneath the Northern Honshu, Japan (Zhang, 2003) and Wellington region, New Zealand (Zhang, 2003) gave a surprisingly phenomenal resolution and also the model
substantiated the hypothesis that intermediate focus earthquakes are triggered by dehydration reaction of hydrous minerals.

At times it becomes difficult to interpret Vp/Vs ratio determined from Vp and Vs due to difference in data quality and distribution between P and S data. As such at the local level both absolute and differential S-P data have been employed in this programme to estimate the Vp/Vs ratio directly which is more in agreement to the local geological setup. Most seismic tomographic programmes uses a regular grid, but faces the problem of highly nonuniform sampling due to the distribution of earthquakes and the stations. Thus, to avoid this issue of mismatch between ray distribution and the regular inversion grid, an adaptive grid DD tomography based on tetrahedra and Voronoi diagrams was developed. Thus this procedure helped in producing a robust and profound velocity model near the source region.

Local earthquake tomography (LET) is now conventionally used in seismically active regions catered by dense seismic network but does not take into account many new developments focused towards improving relative and/or absolute location accuracy. The accuracy of events location of hypocentres of earthquakes is governed by several factors of which network geometry, available phases and arrival time accuracies are of paramount importance (Palvis, 1986). Due to presence of noise, the arrival times picked either manually or automatically is generally associated with errors (Douglas et al., 1997). But waveform cross-correlation (WCC) and event clustering techniques have drastically brought about a revolutionizing improvement in the way earthquakes and explosions are located by giving improved arrival times estimates or in the determination of high-precision relative arrival times (Van Decar and Crosson, 1990). Some notable examples of its applications are Mount St. Helen (Fremont and Malone, 1987), Hawaii (Got et al., 1994; Rubin et al., 1998), California
(Poupinet et al., 1984; Nadeau et al., 1994; Shearer, 1997; Rubin et al., 1999; Waldhauser et al., 1999; Waldhauser and Ellsworth, 2000, 2002) etc. These studies are based on the assumptions that two similar sources generates waves and propagates along similar paths, will generates similar waveforms and waveform cross-correlation technique may be engaged to determine precisely the relative arrival times. These studies combined with others have proved to be a boon in revising the definition of seismogenic features and in the improvement of accuracy with which relative locations of ground-truth events is done by using multiple-event methods with high-precision absolute or relative arrival-time data. Apart from this, tectonic processes, earthquake recurrence, and earthquake interaction are now being seen in a new perspective.

The prime difference between the two cardinal ways of using WCC data is as follows:

(1) the direct usage of relative arrival times, the relative event locations could be conveniently determined (Fremont and Malone, 1987; Got et al., 1994; Waldhauser and Ellsworth, 2000), or (2) secondly, by adjusting absolute time of arrival picks to minimize discrepancies among relative arrival times (Dodge et al., 1995, 1996; Shearer, 1997; Rowe et al., 2002). The approach stated in (1) uses all the available information which are incorporated within the plethora of relative arrival time differences with each datum clearly exhibiting the direct measure of quality i.e. the correlation value. Some assumptions needs to be made for the sake of making the derivation of the locations from arrival time differences, which is a disadvantage. To exemplify this mention may be made about the methods of Fremont and Malone (1987) and Got et al. (1994), where a cluster of events precisely have the same take-off angle and azimuth to each station. This consequences in relatively derived
locations rather than absolute one, which, therefore, requires some assumptions to be made to obtain useful coordinates (e.g., final locations are computed relative to a catalog-based cluster centroid). Waldhauser and Ellsworth (2000), propounded a location algorithm which takes into consideration the spatial derivatives for a set of events, evaluated at the current location of each event, but then it assumes that the path anomalies introduced by velocity heterogeneity are location independent. This assumption works well for closely spaced events and does not hold true for far off events. As such far off events locations may possess velocity heterogeneity biased (Got et al., 1994; Waldhauser and Ellsworth, 2000; Wolfe, 2002). On the other hand, the latter approach makes use of relative arrival times to determine a much smaller number of adjusted arrival time picks, but these picks are absolute arrival times and therefore, very useful in determination of absolute locations (for a given velocity model or using tomography).

The Double-difference seismic tomography method combines the advantages and avoids the disadvantages of the aforementioned approaches. The method is based on the hypoDD code of Waldhauser (2001), and employs both absolute and relative arrival time data. This method helps in to determine a three-dimensional (3D) velocity model simultaneously with the absolute and relative locations of events. The advantage of this method lies in the fact that without discarding off valuable information by only using adjusted picks, relative arrival times with their quality values are included along with absolute arrival times. This also ensures to do away with the concept of simplifying assumptions about ray path geometries or path anomalies and thus produces absolute locations and not just the relative ones. The velocity model obtained is far more superior than the one obtained through standard tomographic techniques. Event locations using standard tomographic techniques
results in scattering due to imperfection in picks and correlated errors, whereas this shortcoming has been overcome in DD tomography by making use of the differential arrival times (this includes both the WCC and catalog time difference data) that helps to get rid of most of these errors, thereby arresting “fuzziness” from the velocity model.

4.7.4.1 Double-Difference Tomography: The Method and its Mathematical Properties

Let $T$ be the arrival time of body wave from an earthquake $i$ to a seismic station $k$, which may be expressed in terms of ray theory as path integral

$$T_i^k = \tau^i + \int_{i}^{k} u ds$$

(4.22)

where $\tau^i$ is the origin time of event $i$, $u$ is the slowness field and $ds$ is an element of path length. The source coordinates ($x_1$, $x_2$, $x_3$), origin times, ray paths, and the slowness field are the unknowns. The relationship between the arrival time and the event location is highly nonlinear, so a truncated Taylor series expansion is generally used to linearize equation (4.22). This linearly relates the misfit between the observed and predicted arrival times $r_k^i$ to the desired perturbations to the hypocentre and velocity structure parameters

$$r_k^i = \sum_{l=1}^{3} \frac{\partial T_k^i}{\partial x_l^i} \Delta x_l^i + \Delta \tau^i + \int_{i}^{k} \delta u \ ds$$

(4.23)

Subtracting a similar equation for event $j$ observed at station $k$ from equation (4.23), we have

$$r_k^i - r_k^j = \sum_{l=1}^{3} \frac{\partial T_k^i}{\partial x_l^i} \Delta x_l^i + \Delta \tau^i + \int_{i}^{k} \delta u \ ds - \sum_{l=1}^{3} \frac{\partial T_k^j}{\partial x_l^j} \Delta x_l^j - \Delta \tau^j - \int_{j}^{k} \delta u \ ds$$

(4.24)
Assuming that these two events are near each other so that the paths from the events to a common station are almost identical and the velocity structure is known, then equation (4.24) can be simplified as

\[ d_{ij} = r^i_k - r^j_k = \sum_{l=1}^{3} \frac{\partial T^i_k}{\partial x^l_k} \Delta x^l_i + \Delta \tau^i - \sum_{l=1}^{3} \frac{\partial T^j_k}{\partial x^l_k} \Delta x^l_j - \Delta \tau^j \]

(4.25)

where \( d_{ij} \) is the so called double-difference (Waldhauser and Ellsworth, 2000). This term is the difference between observed and calculated differential arrival times for the two events, and can also be written as

\[ d_{ij} = r^i_k - r^j_k = (T^i_k - T^j_k)_{\text{obs}} - (T^i_k - T^j_k)_{\text{cat}} \]

(4.26)

The observed differential arrival times \((T^i_k - T^j_k)_{\text{obs}}\) may be computed from both waveform cross-correlation techniques for similar waveforms and absolute catalog arrival times. The relation (4.25) is recognised as the DD earthquake location algorithm (Waldhauser and Ellsworth, 2000). In this method, the path anomalies are assumed to be location independent such that the differencing process can remove such path anomalies among event pairs observed at common stations. In case of fast change of path anomalies from velocity heterogeneity with earthquake location, i.e. location dependent, the earthquake locations will be inevitably biased. This is expected when the event pairs are far apart and inter-event distances exceed the scale length of velocity variations. To reduce or exclude data from event pairs that are far apart Waldhauser and Ellsworth (2000), applied a distance-weighting factor. They still may be linked in the inversion via a series of intermediate pairs that are far apart, albeit the arrival difference data from such events may be excluded (Got et al., 1994). This may be exemplified by the pair \((T^i_k - T^j_k)\) may be linked if the two pairs
$(T^i_k - T^m_k)$ and $(T^m_k - T^j_k)$ are included which means that the travel times from the distant events will be still contributing consequentially to the solution. During the inversion, the influence of far apart earthquakes is only diminished when the distance weighting is combined with damping (Wolfe, 2002). But, this approach will result in less resolution of the relative event locations between earthquakes which are far apart compared to the locations between closely spaced events. Thus, the DD location algorithm shows its capability in only improving the relative locations between closely spaced events. One way to overcome this issue of discrete location-dependent path anomaly bias is to divide a region into small sub-regions and apply the DD algorithm to each sub-region separately (Wolfe, 2002), because the partial derivatives for events from a small cluster will essentially be the same. It is for this reason that the system is only able to resolve the relative location between earthquakes. If one ventures relative location of events considering different regions than it is for sure that the path anomalies from different regions would introduce biases. Therefore, to get rid of the limitations of the DD algorithm, differential arrival-time data and equation (4.24) takes care of the path anomaly bias between event pairs explicitly. The aim is not only to determine the relative event location, but also their absolute locations and the velocity structure. It is to be noted that the ray paths between two closely spaced events will considerably overlap, which means the model derivatives in equation (4.24) will grossly cancel outside the source region. Thus, we need to apply appropriate relative weighing between absolute and differential system, to combine them together. It is evident from relation 4.23 and 4.24, the relationship between arrival time residuals and slowness models are more linear than the event locations. Ultimately, this is indicative of the fact that the convergence of velocity structure is
faster than earthquake locations. Thus, in practice, the simultaneous inversion of event locations and velocity structure alternates with the inversion of only event location.

### 4.7.4.2 Data Compatibility

As mentioned in the previous sections the double-difference tomography algorithm makes use of both absolute and differential data to solve for both event location and velocity structure. The differential arrival times \((T_t - T_f)^{obs}\) are obtained from waveform cross-correlation techniques and/or absolute catalog data by directly subtracting arrival times from event pairs observed at common stations. A combination of correlated and random contributions defines the picking errors for arrival time picks. The errors related to the model or structure errors and picking bias are known as the correlated errors. The issue of correlated and random errors for differential arrival times has been taken into account by introducing the waveform cross-correlation techniques that makes sure that the errors mentioned remains smaller for the differential arrival times compared to the absolute arrival times. Considering the differential catalog data, the picking bias may be correlated like that of model errors for two events which are not far apart from each other because the velocity heterogeneities along the ray paths have the same effect on the waveforms. The correlated errors as well as picking bias are reduced substantially with the employment of differential process. But for far off events the picking bias may not be correlated and the differencing procedure could cause a further amplification of random errors. This is the cardinal reason for which the distance weighting application is critical to reduce random errors between far off events and not just to constrain path anomaly bias.

In the process of inversion three basic type of data are used viz. the absolute arrival times, the catalog differential arrival times and the waveform cross-correlation
data. Similar to hypoDD, the hierarchical weighting scheme during the inversion has been applied to combine the three types of data into one system. It is only within the same cluster waveforms usually correlate between event pairs. But when it comes to different clusters, the relative locations between two events needs to be controlled by differential catalog data with larger event separation. It is absolutely for this reason that the inversion is started by applying higher weightage to the catalog data (both differential and absolute catalog data) to establish the large-scale result (1 for absolute data, 0.1 for differential catalog data, and 0.01 for cross-correlation data), similar to hypoDD (Waldhauser, 2001). Examples of synthetic data shows that a higher model resolution and lower model uncertainties for location parameters are attained in the system which uses both the absolute and differential data with relative weighting 1:0.01.

To refine the event locations and the velocity structure near the source regions, the catalog differential data are weighted more (1 for differential catalog data, 0.1 for absolute catalog data, and 0.01 for cross-correlation data). In a final step towards further refinement of event locations and the velocity structure near the source region, waveform cross-correlation data if available, will then be given even more weightage than the catalog differential data (e.g. 1 for cross-correlation data, 0.01 for differential catalog data, and 0.001 for absolute catalog data). As the cross-correlation data are at least an order of magnitude more precise than the manual picks (Waldhauser and Ellsworth, 2000), the differential catalog data is downweighed by a factor of 100 in the final stage of the inversion. It may be mentioned that without sacrificing the lower model uncertainties that a system using differential data provides, additionally, by inclusion of absolute data into the system with relatively small weighing is helpful to make the system more stable.
**4.7.4.3 Realization of Double-Difference Tomography**

_tomoDD_ code was developed for double-difference tomography based on the DD location code _hypoDD_ (Waldhauser, 2001). The pseudo-bending ray-tracing algorithm (Um and Thurber, 1987) has been used in the current version of _tomoDD_ to find the rays and calculate the travel times between events and stations. A regular set of 3D nodes represents the model and the velocity values are interpolated by using the tri-linear interpolation method. The local velocity at the source and the hypocentral partial derivatives are calculated from the direction of ray (Lee and Stewart, 1981). By apportioning the derivatives to its 8 surrounding nodes according to their interpolation weights on the segment midpoint (Thurber, 1983), the model partial derivatives (calculated in terms of fractional slowness perturbation, so that the derivatives are related to path length) of ray paths divided into a set of segments are evaluated. If at a common station two rays are observed for which the model derivatives at the common inversion node for these two rays are close enough (i.e. within 5 m), then the corresponding elements are set to zero in the model derivative matrix to make the system more stable. Similar to Waldhauser and Ellsworth (2000), distance weighting is also used in the DD tomography to control the maximum separation between event pairs and apply higher weighting to the closer events.

By seeking first-order or second-order smooth model, velocity anomalies are constrained for simultaneous inversion of velocity structure and event locations. It is required that a smoothing regularization needs to provide a minimum-feature model that contains only as much as structure can be resolved above the estimated level of noise in the data (Constable _et al._, 1987; Sambridge, 1990; McCaughey and Singh, 1997). The slowness perturbations between neighbour grid nodes are constrained to be smooth by first-order smoothing constraint as follows:

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\[ \Delta m_i - \Delta m_j = 0 \]  \hspace{1cm} (4.27)

To penalize the roughness of the slowness perturbation between neighbour grid nodes, alternatively, the second-order smoothing constraints may be used as follows:

\[ \Delta m_i - \Delta 2m_j + \Delta m_k = 0 \]  \hspace{1cm} (4.28)

Smoothing weights could be either same or different in different directions. *tomoDD* code also offers the option to determine just the event locations by using different data types. The linearity of relationship between arrival time residuals and slowness model is well reflected by this fact. Consequently, the convergence of slowness structure to its optimum value is faster than the event locations. Normally, the joint inversion of velocity structure is alternated with event locations with the inversion for event locations only done in the applications of double-difference tomography.

The complete of linear equation to solve for velocity structure simultaneously with event locations by using both absolute and differential data is

\[
Y = (Q_{dd}E)_{WE}^{-1} (Q_{dd})_{WE} \Delta T
\]  \hspace{1cm} (4.29)

Equation 4.29 models the unknown parameters including hypocenter, origin time, and slowness models where \( w \) is the relative weighting and \( I \) is identity matrix and \( Q_{dd} \) is the difference operator which annihilate the station correction. This is the complete formula to solve for velocity structure simultaneously with event locations by using both absolute and differential data.

The above set of linear equation (4.29), in association with the smoothing constraints relations (4.27) and (4.28), is solved by means of the LSQR algorithm (Paige and Saunders, 1982) for the damped least squares problem (for the inversion for earthquake locations, the system is based on relation (4.30) as given below.
\[ \Delta X = \left( \mathbf{Q}_{\text{DDA}} \mathbf{w}^{-1} \right) \left( \mathbf{Q}_{\text{DD}} \right)^{-1} \Delta \mathbf{T} \] (4.30)

where the partial derivative operator used in double-difference location is \( \mathbf{Q}_{\text{DDA}} \). \( \left( \mathbf{Q}_{\text{DDA}} \mathbf{w}^{-1} \right)^{-1} \mathbf{Q}_{\text{DD}} \) may be used to indicate the dependence of the corresponding model parameter (location or origin time in this case) on all travel times.

The parameters on which each equation is weighted are \textit{a priori} data uncertainty, data type, distance between the event pair, and misfit during each iteration. The relative weighting for different data types and the distance weighting are determined \textit{a priori}, whereas the residual weighting is determined \textit{a posteriori}, with large residuals rejected or downweighted by a biweight function (Waldhauser and Ellsworth, 2000).

The number of iterations required to reach certain accuracy during LSQR depends strongly on the scaling of the problem (Paige and Saunders, 1982). Proper scaling of the rows or columns of matrix makes it easier to recover the solution. Especially, for this reason before the system is fed into the LSQR, each column is scaled so that the \( L^2 \) norm of each column is equal to 1.

4.7.4.4 Summary

The basic objective of designing the double-difference tomography code is to solve for earthquake location problem and simultaneously work out a velocity structure by using both absolute and differential data. By taking into consideration the path anomaly biases between event pairs explicitly, this system overcomes the limitation of the double-difference location method. Apart from characterizing the local velocity structure accurately, the double-difference tomography has the capability to relocate large number of earthquakes accurately both in relative and absolute locations.
The mathematical properties of D tomography based on 2D synthetic example has shown that the differential data has lower model uncertainties for both location and slowness model parameters, but has lower model resolution for some boundary slowness model parameters. This has been taken care of by including the absolute data with a relatively small weighting into the differential system is useful to bring stability to the system without compromising the low model uncertainties and high model resolution that the differential system can offer. To combine the different types of data into one system and also to solve for the event locations and velocity structure from a large to a small scale, a hierarchical weighting scheme has been introduced. To reduce random errors between far apart event pairs and to reduce path anomaly bias, the distance weighting technique has been used in the DD tomography.

The system has been regularized by using smoothing constraints to penalize the smoothness or roughness of the slowness perturbations between neighbouring grid points. In totality, the DD tomography and smoothing constraints is solved by means of the LSQR algorithm. The tomoDD package is an outcome to bring to fruition the DD tomography method.

4.7.5 Simultaneous Inversion Tomography Programme

One of the widely used Local Earthquake Tomography (LET) methods based on travel times is the simultaneous inversion programme given by Thurber (1983). This method iteratively simultaneously invert for three-dimensional velocity structure and hypocentral parameters for local earthquakes used in a study. The parameter separation method given by Pavlis and Booker (1980) to simultaneously estimate velocities along a three-dimensional flexible grid as well as hypocentral parameters of locally recorded earthquakes has been adopted. Thurber (1981) devised the forward
modelling for 3-D ray tracing to calculate P-wave travel time. To incorporate the S-wave arrivals for estimation of 3-D variation in S-wave velocity and Vp/Vs, Eberhart-Phillips (1993) modified the programme. The resolution of the data set and efficacy of the tomographic inversion are very much influenced by model parameterization of 3-D inversion which consists of grid parameterization, selection of initial 1-D velocity model, suitable damping parameter and the number of iterations for convergent and consistent solution. One has to perform several inversions by varying the model parameterization grid as well as the starting velocity model in order to obtain a well-constrained velocity structure.

The linearized equation for Simultaneous Inversion relating the travel time residual $r$ to model parameter, using a finite parametrization changes can be written as

$$r = \Delta t_e + \frac{\partial t}{\partial x_e} \Delta X_e + \frac{\partial t}{\partial y_e} \Delta Y_e + \frac{\partial t}{\partial z_e} \Delta Z_e + \sum_{n=1}^{N} \frac{\partial t}{\partial v_n} \Delta v_n$$  \hspace{1cm} (4.31)

where $\Delta t_e$, $\Delta X_e$, $\Delta Z_e$ and $\Delta v_n$ are perturbations to the hypocentral parameters (earthquake origin time and location) and the velocity parameters, and $\partial t/\partial x_e$, $\partial t/\partial y_e$, $\partial t/\partial z_e$ and $\partial t/\partial v_n$ are the partial derivatives of the arrival time respect to the earthquake coordinates and velocity parameters respectively. $N$ is the total number of velocity parameters. There is one such equation for each observed arrival.

The algorithm for the solution is analogous to hypocentre location, but then instead of solving for just one earthquake, simultaneously the programme inverts for travel times of many earthquakes, and instead of using an assumed velocity model, a new velocity model evolves where the output files of the previous inversion becomes the input file for the next inversion.
4.7.5.1 Numerical Solution of the Simultaneous Inversion Problem

(1) First a trial parameter vector $\zeta^*$ as given by equation is guessed

(2) Then the theoretical travel time $T_{jk}$ and its spatial partial derivatives are computed.

(3) The system of linear equations are solved for the adjustment vector $\delta \zeta$ using the normal equations approach or generalized inversion.

(4) The trial parameter vector $\zeta^*$ is replaced with $(\zeta^* + \delta \zeta)$.

(5) These steps are repeated until the solution converges to some pre-defined number of iterations.

4.8 Data Base Used and Data Analysis

4.8.1 Earthquake Location

In this study, local earthquake data of northeastern India was used, digitally recorded by various agencies having their own network in northeastern India. These agencies are viz. CSIR-North East Institute of Science & Technology (CSIR-NEIST), Jorhat, CSIR-National Geophysical Research Institute (CSIR-NGRI), Hyderabad, Indian Meteorological Department (IMD), Shillong, Manipur University and IIG-Shillong for the period 2001-2012. These events are recorded with higher signal-to-noise ratio and have clear P and S phases.

In order to accomplish inversion the travel time of P and S-P phases are used. S phases could not be inverted as there were not enough S-phase readings available to give a well resolved S-wave velocity structure. P-wave arrivals had the reading accuracy of 0.1 sec and the S-wave arrivals had the accuracy of 0.2 to 0.5 sec. The waveforms have been recorded by 47 digital seismic stations with three component
seismometers which is a mix of both permanent and temporary network operated by different agencies in different time and space. The locations of these seismic stations and the instrumentation used along with the network to which they belong have been detailed in Table 4.2. The locations of these seismic stations and the network to which they belong have been shown in Figure 4.6. GPS timing systems were impinged in the seismic stations to ensure time synchronisation with Co-ordinated Universal Time (UTC) to achieve high timing precision. The magnitude of completeness is M~2.5. During a span of 12 years, 1293 events were recorded with clear P and S phases of which 1087 events was initially selected which had at least six observations.

Initial location of events were obtained with a dataset containing 1087 events using the Hypoinverse earthquake location programme given by Fred W Klein (1978), by employing a 1-D priori velocity model of De and Kayal (1990). The location of the events and the stations are shown in Figure 4.3. It has been observed that the uncertainties in hypocentre locations contribute towards instabilities in the inversion process, therefore, scrutiny of preliminary locations was done (Kissling et al., 1994). This led to the selection of 892 events with low errors in hypocentral parameters (ERLN, ERLT, ERDP) & RMS i.e. ≤ 1 and properly constrained by azimuthal coverage. The year wise distribution of these 892 earthquakes have been shown in Figure. 4.7.

For the 3-D inversion the data sets were prepared by ensuring that each event hypocentral parameter is determined by minimum of 5 number of P and S arrival times, with at least two S phases. S phase readings are required for better precision of the earthquake locations particularly for constraining the depth (Bhattacharya et al., 2008). Travel times were restricted to residuals less than 1sec. When the initial weight
was given to be as 4, the travel time was kept at with P-residual less than 0.5 sec and S-residual less than 1 sec. The event pairs were constructed at the same station with inter-event distance less than 20 km and event-station distance less than 300 km. Likewise, two set of data were prepared, one with RMS < 0.3 and the other with RMS < 0.5.

The events chosen for the data set with RMS < 0.3 after differential time match is 399 along with catalog differential P arrival times of 4215 and catalog differential S arrival times of 2852. The catalog based absolute P arrival times is 2198 and absolute catalog S arrival times is 1680. Similarly for the data set with RMS < 0.5 the events chosen after differential time match is 493. The catalog based differential P arrival times is 5396 and the S differential arrival times is 3510. The absolute P arrival times and S arrival times are 2704 and 2002 respectively.

To obtain a grid configuration that matches well between model parametrization, spatial resolution and a reliable representation of the velocity structure, several trials of grid spacings were run by selecting spacing within the grid is so selected that it provides enough ray paths near each grid point. The partial derivatives of velocity and the velocity for a point along a ray path are computed by linearly interpolating between the surrounding eight grid points. This is how the velocity of the medium is parameterized by assigning velocity values at the intersections (grid points) of a non-uniform, three-dimensional grid. Thus, the velocity is solved to show gradational changes rather than sharp discontinuities (Bhattacharya, 2010). Likewise, after some trials, we selected a grid with an inter-node distance of 30 km from latitude 25° to 27° N and longitude 90° to 94° E and 10 km in depth is selected as the vertical grid separation in the entire study region.
During the inversion process at the beginning maximum weightage was given to the absolute data and then to the differential data. When all the parameters were adjusted, a comparison of the results of the inversions with RMS < 0.3 and RMS < 0.5 was done (Figures 4.19 & 4.20 respectively). It was observed that the velocity structure were very similar in both the cases, but rather somewhat clearer with RMS < 0.5 (Figure 4.20). Otherwise the control parameters such as variation reduction were quite comparable. Therefore, the results obtained with RMS < 0.5 with inter-node spacing of 30 km were accepted for tomoDD (Figure 4.20). In this tomographic programme the optimum inverted 1-D velocity model has been used to get 3-D crustal velocity structure of northeastern India. The derivation of inverted 1-D velocity model has been discussed in the next section under 4.8.2.

Similarly, another test was carried out with RMS < 0.5 and grid configuration with 26° N and 93° E at the origin covering an area of about 400 km x 400 km in the east-west and north-south directions, respectively. A 50 km inter-node distance was taken with 10 km in depth is selected as the vertical grid separation in the entire study region (Bhattacharya et al., 2008). This parameterization allowed ray paths which was superfluous and resulted in contamination of the results in the sense that there was much improvement in the resolution but velocity anomalies were smoothes out over a large volume, making it rather difficult to correlate with the tectonic elements. So the results of this test were discarded off (Figure 4.21).

### 4.8.2 Optimum 1-D Velocity Model

To devise the optimum 1-D model by inversion, the scrutinized 892 events were engaged in the local tomographic technique (LET) of simultaneous inversion given by Thurber (1983) and modified by Eberhart-Phillips (1993) to incorporate the
S-wave arrivals for estimation of 3-D variation in S-wave velocity and Vp/Vs. This method makes use of the parameter separation method of Pavlis and Booker (1980) to simultaneously estimate velocities along a three-dimensional flexible grid as well as hypocentral parameters of locally recorded earthquakes. Thurber (1981), made use of forward modelling for 3-D ray tracing to calculate P-wave travel time. Due to meagre number of S phase picks, S-wave velocity inversion was avoided.

Since LET is a nonlinear inversion problem, such that the results of inversion depends on the initial reference model (Kissling et al., 1994). An appropriate initial model for local earthquake tomography should be close to the true 1-D average model. Therefore, the velocity model of Bhattacharya et al. (2008) was used as a priori model for performing the inversion to obtain an optimum 1-D inverted velocity model for northeastern India (Table 4.1). This optimum 1-D P-wave velocity model could be generated after six inversion steps, wherein the output files of the previous inversion becomes the input files for the next one. The inverted optimum 1-D velocity structure suggests six-layered crust beneath the region. In the process, hypocentral parameters and simultaneously the velocity structure keeps on improving with every iteration and at one point of time we get the most refined and realistic hypocentral parameters of all earthquakes along with a near accurate 1-D velocity model. The 3-D tomographic inversion by tomoDD was then performed on 892 events using the optimum 1-D inverted velocity model as the priori input model for starting with the inversion. The efficacy of the tomographic inversion and the resolution of data set are strongly influenced by model parameterization of 3-D inversion which comprises of grid parameterization, selection of initial 1-D velocity model, suitable damping parameter and the number of iterations for convergent and consistent solution. In order to obtain a well-constrained velocity structure, several inversions had to be
performed by varying the model parameterization grid as well as the starting velocity model (Bhattacharya et al., 2008). This may be understood in the light of the fact that the number of rays passing near each grid intersection, which controls the resolution at that node, arises from the station coverage, earthquake distribution and node spacing. As such in this work maximum numbers of stations available in the database have been used. If we increase the node spacing, it is likely that the resolution of the velocity anomalies may be improved, but then velocity anomalies smoothens over a large volume which makes it difficult to correlate with the tectonic features. On the other hand, inverting for small anomalies by reducing the node spacing causes a substantial decrease in resolution. Decreasing the damping factor increases resolution at the expense of an increased standard error. Therefore, care was taken to choose an optimal value of the damping parameter that yields low data variance, low solution variance and low standard error with a relatively good average resolution (Bhattacharya et al., 2010).

To derive the optimum inverted 1-D velocity model to be used for 3-D inversion in this study, various models were tried for trial runs as apriori model, so as to obtain a mean initial velocity model. While running the tomographic programme using the velocity model of Bhattacharya et al. (2008) as the apriori model as discussed above, the optimum inverted initial velocity model was obtained which shows a decrease in velocity in all depths (Table 4.1) as compared to the models of De and Kayal (1990) and Bhattacharya et al. (2008). The epicentral distribution of relocated earthquakes using the optimum inverted 1-D velocity model has been shown in Fig. 4.4. The epicentral distribution is mostly clustered in the Shillong-Mikir Plateau, Kopili Fault zone and the IBR. The epicentres which were scattered in the vicinity of the northeastern region of India seem to have diminished drastically and
there seems to be sparse distribution of epicentres in the Arunachal Himalaya block. The inverted 1-D velocity model reduced the average error in longitude from 7.08 Km to 4.63 Km (Figure 4.13), average error in latitude from 9.26 Km to 4.86 Km (Figure 4.14), average error in depth from 8.39 Km to 5.73 Km (Figure 4.15) and average error in RMS from 0.75 sec to 0.39 sec (Figure 4.16) compared to that obtained by the 1-D priori model (De and Kayal, 1990). Thus the quality of relocated events with inverted 1-D velocity model improved significantly. The frequencies of distribution of the relocated earthquakes by using the 1-D inverted velocity model with respect to depth and RMS error have been shown in Figures 4.8 and 4.9 respectively. Large number of earthquakes is seen to occur below a depth of 100 Km and most of the earthquakes have a RMS or residuals below value 0.5 sec which means the database is of good quality. The depth wise distribution of the earthquakes relocated by 1-D inverted velocity model with respect to longitude and latitude is shown in Figures 4.10 (a) & (b) respectively. From both the diagrams it is clearly seen that the relocated earthquakes are distributed mostly within 100 km with scattered occurrences beyond 100 Km, extending upto a maximum depth of 200 Km. The deeper earthquakes are seen to occur mostly towards IBR region, which is expected due to subduction tectonics. The 3-D tomographic inversion was done using tomoDD by using this 1-D inverted velocity model as the starting velocity model or the priori velocity model.
Table 4.1: *Apriori* 1-D velocity models used to derive the optimum 1-D inverted velocity model for the 3-D inversion process in this study

<table>
<thead>
<tr>
<th>Depth (Km)</th>
<th>Vp (km/s)</th>
<th>Vp (Km/s)</th>
<th>Vp (Km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>De &amp; Kayal (1990) Using Hypo-71 (205 events)</td>
<td>Bhattacharya <em>et al.</em> (2008) Using SIMULPS Inversion (980 events)</td>
<td>This study (1-D Inverted P-velocity model derived) (892 events)</td>
</tr>
<tr>
<td>0</td>
<td>5.55</td>
<td>5.56</td>
<td>5.25</td>
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<tr>
<td>10</td>
<td></td>
<td>6.10</td>
<td>5.90</td>
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<tr>
<td>20</td>
<td>6.52</td>
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<td>7.60</td>
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<td>8.10</td>
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<td>46</td>
<td>8.57</td>
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<tr>
<td>50</td>
<td></td>
<td>8.40</td>
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</tbody>
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4.8.3 Inversion for a 3-D Velocity Structure

The three dimensional (3-D) velocity structure of NER of India was investigated by using the LET programme *tomoDD* designed by Zhang and Thurber (2003) which is based upon the double-difference location code *hypoDD* (Waldhauser, 2001). In this programme the travel times between the events and stations are calculated by tracing the rays by using pseudo-bending ray tracing algorithm given by Um and Thurber (1987). Both differential and absolute arrival time of the data is used to estimate locations and 3-D seismic velocity structure. There is a coupling effect between the event hypocentres and the velocity structure.
(Thurber, 1992). The advantage of using double-difference technique is to eliminate the overlapping path effect common to nearby events travelling to a single station. Thus, the model derivatives will essentially cancel outside the source region, helping to isolate the relative locations of the events. Apart from using the double differences of nearby earthquakes, the absolute arrival times are used in the inversion to resolve the velocity structure outside the source region. This helps to simultaneously determine the velocity structure and the relative event locations as well as the absolute event locations in an accurate manner. The velocity values are interpolated by using the trilinear interpolation method and the model is represented as regular set of 3D nodes. From the direction of the ray, the hypocentral partial derivatives are calculated and the local velocity at the source. The model partial derivatives are calculated in terms of fractional slowness perturbation and for that reason the ray path is divided into a set of segments.

4.8.4 Resolution Analysis

Accuracy and uniqueness of the imaged anomalies in inversion studies is of vital importance. A scrutiny of the reliability of the depicted images is crucial in seismic tomography, which makes it critical to go for a model resolution test (Menke, 1989). An analysis of resolution matrix R needs to be done so as to determine the minimum ray paths required to compute the velocity perturbation with higher precision in each block. The coupling of the solution and the model parameters are commonly described by the model resolution matrix R, i.e. $\Delta m^{est} = G^{-g} \Delta d = R \Delta m^{true}$, where $G^{-g}$ is the generalized inverse and $m^{est}$ is the estimation for the model $m$. The variation in model parameter $\Delta m$ expresses the disturbances of the data $\Delta d$, i.e. $\Delta d \sim G \Delta m$. The resolution matrix may be treated as a filter through which the estimated model is obtained from the true model. An idea about how well the velocity
is constrained at each grid point and then it is correlated with the number of ray paths passing near the grid point (Thurber, 1981). Station coverage, earthquake distribution and node spacing determines the total number of rays passing near the grid intersection which influences further the resolution at that node. Although, increasing the node spacing improves the resolution but smoothen out the velocity anomalies over a large volume making it increasingly difficult for a tectonic correlation. But then, there is a considerable decrease in resolution when inversion over small anomalies is done by reducing the node spacing. Maximum number of stations out of the ones available in the database was utilized and given the distribution of both seismic stations and earthquake sources, the model resolution varies drastically over the inverted volume. But the overall result is resolvable in the central part of the model at depth layers from the subsurface to the 50 km depth. Resolution seems to be poor in the fringe zone of the study area because the density of ray paths is poor. Only the resolvable parts of the velocity models could be analyzed. The resolution for Vp at different depth slices as obtained by Checkerboard Resolution Test (CRT) are illustrated by Figures of 4.22 which clearly indicates higher resolution in the central part of the study area. The purpose of carrying out this test was to assess the ability of the data and the method to recover existing velocity anomalies within the model. Checkerboard test was carried out with RMS < 0.5 with the inversion grid being same as has been taken for inter-node distance of 30 km. To make a checkerboard, positive and negative values of velocity perturbation with 5 % anomaly was assigned alternatively to all the 3-D grid nodes. The results of CRT were obtained for six representative depths for Vp which shows that 5 % anomalies were well recovered for most of the layers except for the fringe zone of the model. Ray density analysis gives a fair idea about the number of phase readings used for inversion of a model.
parameter at a given grid node. This depends on the quantity of ray paths passing at a given grid node. In Figure 4.17 the variation of the ray density parameter for various depth levels has been shown. In the study area values vary from 0 to 3500. The depth levels where the value is less 10, are not properly inverted, therefore, depths upto which the ray density parameter is more than 10 is only shown. Upto 50 km depth the ray coverage is observed to be good and proper inversion of data is observed. Beyond this depth, less ray coverage is seen due to less number of earthquakes at greater depths. Inversion was reasonably constrained with alternation of high and low velocity recovered in the central part of the model with Derivative Weight Sum (DWS) > 50, but not in the amplitude of perturbation, which means that the velocity changes is reliable but not their absolute value. Derivative weight sum (DWS) is the normalized ray path density. It is an index as to how important any data is and how it is contributing to the inversion. It is the ray path density weighted by ray-node separation and ray path length in the vicinity of the node. DWS describes the amount of data actually constraining the velocity at the nodes (Thurber et al., 1994). Figure 4.18 shows the variation of DWS for Vp at different depth slices. The DWS value for the study region ranges from 0 to 9991. From depth slices 0 to 50 Km the DWS values ranges from 60 to 8000 which is indicative of good ray path coverage. Beyond 50 Km the ray coverage gets reduced and shows poor inversion of travel times.

4.9 Results and Validation of the Tomographic Model

The understanding of the structure and petrology of the continental crust is of paramount importance in creating a knowledgebase about crustal generation and its evolution. Earlier the practice was to come out with rather simple layered models of the earth’s crust by employing the technique of seismic refraction. Of late more emphasis have been paid for a structurally validated gradient model with lateral
variations that takes care of the local structural complexities too and thus, the classical layered crustal model have become redundant. These new integrated seismic images are a reflection of the multiple episodes of accretion, deformation, metamorphism, plutonism and volcanism.

Keeping in view a similar endeavour the seismic tomographic study of the crust of the north-eastern region of India have been undertaken for this research programme. But, the goal of such an exercise remains unaccomplished without a proper geological correlation and corroboration of the model obtained. Therefore, in this LET study the seismological model obtained for northeastern India region by *tomoDD* tomographic programme is well represented in form of six depth slices (Figure 4.20) which are discussed as follows:

**4.9.1 For Depth Slice at 0 km (Figure 4.20 a)**

This depth slice represents the shallow subsurface almost near the surface of the earth. At this depth a conspicuous low velocity zone (LVZ) trending NW-SE is seen cutting through the central Assam Valley and is flanked on either sides by two high velocity zones (HVZ). Geologically, this LVZ (~4.5-5 km/s) is represented by the Kopili fault zone. It is well known that the NW-SE oriented Kopili fault zone dissects the Shillong Plateau and causes detachment of the Mikir Hills from the Shillong Plateau Faults zones are characterized by damaged zones which are caused by earthquakes over time and seismically characterized as LVZs composed of highly fractured materials. Cracks in the damaged zone are good conduits of fluid which further contributed in slowing down the seismic velocity (Yang and Zhu, 2010). The Kopili fault zone is seismically highly active and very heterogeneous and fractured as compared to Shillong Plateau due to the influence of multiple complex shear stress contributed by the collisional tectonics of the Himalayan arc in the north and the the
Moreover, the geologic formation of Kopili area itself is highly sedimentary in nature (Nandy, 2001). These factors do contribute for the LVZ nature of the fault zone and in this tomogram it has been imaged in the LVZ of ~ 4.5 to 5 km/s. The HVZs on either side are the Shillong Plateau and Mikir Hills which are hard rock terrains. Therefore, they are being imaged as HVZ (5.5-6 km/s) zones. The hard rocks are highly elastic and so offer a fast passage for the seismic waves which are highly elastic in nature. The Shillong-Mikir Plateau is characterised gneissic rock at the basement which are very old (Precambrian age), hard and rigid (Nandy, 2001). Hence, they act as HVZ material, which is why the two HVZs are to be seen parallel to the LVZ. There are some geological examples of exposure of hard rocks on the surface of the Shillong-Mikir Plateau and the Assam Valley. The prominent HVZ is because of the exposure of the basement gneissic complex mostly at the Mikir Plateau (Figure 3.1).

A comparison of global P-wave velocity (Vp) given by Christensen and Mooney (1995) for different rock types have been drawn. The values within the braces refer to these global Vp values at 5 km depth. There are exposures all over in the Shillong-Mikir Plateau and also in parts of Lower Assam Valley areas, of rocks like the granite gneiss (with velocity 5.994 km/s) and other hard igneous and metamorphic rocks like granite-granodiorite (6.192 km/s), quartzite (5.889 km/s), phyllites (6.073 km/s), slate (6.065 km/s), mafic granulite (6.746 km/s), Mica Quartz schist (6.161 km/s), metapelites, amphibolites (6.746 km/s) etc. Going by these global values it is expected that at this depth slice, on an average Vp is seen to range between 5.5-6 km/s on the higher side. The low velocity side (4.5-5 km/s) is a consequence of soft sediments like alluvium and silt and sedimentary rock like limestone (3.5-6 km/s), Dolomite (3.5-6.5 km/s), sandstone (2-3.5 km/s), shales and clay (1.1-2.5 km/s).
km/s) (coal (2.2-2.7 km/s) etc. being especially in abundance in the Assam valley and are also available in the Shillong-Mikir Plateau. The values of sedimentary rocks have been referred from the values given by Stanford Rock Physics Laboratory available online. The central and eastern part of Shillong Plateau and the Mikir Hills is conspicuously HVZ (~ 5.5-6 km/s). This is because these areas are dominantly occupied by granite gneiss, calc-silicate gneiss, basic granulites and Shillong group of rocks which are sedimentary rocks that have undergone metamorphism with intrusive plutons of Mylliem granite type including Khasi greenstone (Nandy, 2001 and Choudhury, 2008).

4.9.2 For Depth Slice at 10 km (Figure 4.20 b)

One of the chief features of this particular depth slice is the prominence of LVZ (~ 5.5 km/s) which appears in the western extremity, northwestern and northern areas of the Shillong Plateau and a NW-SE trend of densification of seismicity is seen. LVZ is because of the fact that the area is clustered with shallow heterogeneous structure as it is the seat of some shallow faults like Chedrang, Dudhnoi, Samin and Dapsi faults oriented NW-SE, N-S, NW-SE and E-W respectively which shows dense distribution of hypocentres. Seismicity is seen to be very sparse in Mikir Hills at 10 km depth and is imaged as HVZ (6 km/s) which is little reduced in its Vp from 6.5 km/s at depth 0 km/s. Alluvium covers the hard basement rocks (granite gneiss) because of which a gradation from high to low velocity (~5.5-6 km/s) is observed. Everywhere, around the Assam Valley there is a thick cover of alluvium upto ~ 4 km which is why a gradation from HVZ to LVZ (~ 5.5-6 km/s) is seen.

4.9.3 For Depth Slice at 20 km (Figure 4.20 c)

In this depth slice the dominance of low Vp is conspicuous. Marked low Vp (6 km/s) is witnessed in the parts of northwestern, southern, southwestern, southeastern
margins of the Shillong Plateau. Some of these areas show Cretaceous-Tertiary Shelf sediments (Nandy, 2001). These sedimentary terrains comprises of rocks such as limestones, sandstones, arkose, fossiliferous beds, alluvium etc. But it is unreasonable to expect a sedimentary terrain at a depth of 20 km. The sedimentary rocks may have undergone metamorphism. In that sense the presence of fluids and melts cannot be ruled out. In this region at this depth hard rocks are expected, since the crust is very thin (Borah et al., 2016) and surface exposures of basement gneissic rocks have been reported from this part of Shillong Plateau (Nandy, 2008) The clustering of faults viz. Samin faults, Chedrang fault, Dudhnoi fault, Dapsi and Dauki Thrust are obviously major contributor to the low Vp characteristics. As such the collective effect of the heterogeneity introduced to the crust looks like the primary cause behind the very low Vp (~5.5-6 km/s) zone in this part of the plateau at 20 km depth albeit the presence of melt which very much causes reduction in velocity of seismic waves may also be looked at in the future.

Coming to the Tyrsat-Barapani shear zone (BSZ) oriented in NE-SW, is a zone with drastic reduction of grain size of the country rocks, intense silicification, development of hydrous phase of minerals due to intense shearing (Nandy, 2001), which is indicative of an intense tectonic history. The intense shearing of Shillong Group rocks must have obviously caused lots of heterogeneity in the crust as well as produced heat and and expulsion of fluids which influences in the slowing down of Vp (~5.5-6 km/s) at this depth slab. Similarly the Brahmaputra valley comprises alluvium cover upto ~ 4 km especially in the upper Assam Basin which actively contributes in slowing down of Vp. Kopili fault, Naga thrust, Belt of Schuppen and Disang thrust do indicate evidence of low Vp (6 km/s) which is absolutely obvious because the structural complexity and the influence of intense
tectonism of multiple tectonic regime on these geological structure results in immense heterogeneity and might be the possible presence of fluids in these fault zones causes the slowing down of velocity of P-waves (Vp). Therefore, LVZs are seen to be associated with faults and shear zones all throughout the study area. The high Vp (7 km/s) observed along Indo-Burman Ranges at the junction where it takes a convex turn is the result of accumulation of high stress due to intense deformation which causes asperity zone like situation and hardening of the crust. Therefore, when seismic waves travel through this area, Vp shoots up (7 km/s). The conspicuous cluster of high Vp (~7-6.5 km/s) seen near Assam-Bhutan border where the MBT and MCT are juxtaposed, is a clear indication of asperity zone and may prove to be a potential source for future eventful earthquakes.

4.9.4 For Depth Slice at 30 km (Figure 4.20 d)

The tomogram at 30 km depth albeit the Vp ranges from 6-7.5 km/s but the dominance of LVZ (6-6.5 km/sec) is observed. As per Christensen and Mooney (1995), this velocity matches with granite gneiss (6.120 km/s) and felsic granulites (6.303 km/s). It is known in the study area that the basement rock has two types of rocks; the gneissic complex proper comprising of abundant granitic gneiss of amphibolites facies and migmatites too (Nandy, 2001). There are also exposures of rocks from upper amphibolites facies to granulite facies (Hornblende-pyroxene granulites and pyroxene granulites) in the Patharkhanga area of West Khasi Hills District of Meghalaya (Choudhury, 2008). It is likely that granitic gneiss underwent metamorphism and attained granulite facies equilibration.

4.9.5 For Depth Slice at 40 km (Figure 4.20 e)

At this depth slice, Vp ranges from 6.5-8 km/s, but it is unequivocally the predominance of low Vp (6.5-7 km/s), observed all over. With reference to the global
Vp values of Christensen and Mooney (1995), the Vp values at 40 km depth for amphibolites (6.751 km/s) and mafic granulites (6.685 km/s) matches with the dominantly low to intermediate Vp values (6.5-7 km/s) observed at this depth for the study area. As mentioned before, there are reports of occurrences of rocks such as granite gneiss, metapelites and basic granulites (Choudhury, 2008) apart from rocks like migmatites along with metasediments (Nandy, 2008) from the Shillong-Mikir Plateau area. Choudhury (2008) advocated that basic granulites, metapelites and granite gneiss attained a granulite facies equilibration which was followed by thorough annealing recrystallization and subsequent cooling and rehydration which were responsible for the retrogressed assemblage in different rock units. These examples of occurrence of metamorphic rock types in the plateau area obviously renders supports to the observed Vp (6.5-7 km/s) which is corroborated by global seismic P-wave velocities for the mentioned types of rocks at this depth. This correlation bears testimony to the fact that, it is most likely that at depths of around 30-40 km and beyond metamorphic processes resulted in the gneissic rocks and granulites and also such metamorphic processes might have resulted in the occurrences of melt too, as the basement gneissic complex show the presence of migmatite rocks which is a product of partial melting. The LVZs at this depth range may be related to the possible occurrence of melt. As evident from this study the dominance of low to intermediate Vp values (6.5-7 km/s) at this depth range and correlation with average global Vp values given by Christensen and Mooney (1995), it is indicating that possibly the composition of melt at this depth range could be between mafic granulite to mafic garnet granulite, the average global values of Vp for which are 6.685 km/sec and 7.011 km/sec respectively. But to establish this requires extensive petrological and geochemical studies corroborated and correlated with other geophysical parameters as
well. The availability of such high pressure and temperature metamorphic rocks at shallower depths and also as surface exposures may be due to the emplacement of these rocks to shallower depths and even to surface by episodes of tectonic uplift and exposure on the surface due to erosion of geological overburden.

At this particular depth two very LVZ (~ 6.5-7 km) is witnessed along northern extension of Kopili fault around Tezpur area in the northern bank of Brahmaputra which extends upto Main Boundary Thrust (MBT) in the Assam-Bhutan Border and the other one at Mikir Hills. From analysis of fractal dimension and b-value it is established that that the Kopili fault zone is highly fractured as compared to the Shillong Plateau due to complex multiple shear stress contributed by both Himalayan arc in the north and the Burmese arc in the east which is why this fault zone is very heterogeneous (Bhattacharya et al., 2010). The existence of brittle and fracture zone indicate low Vp in the region. The Tezpur area going upto MBT is criss crossed by fractures, faults and lineaments which renders to a very highly heterogeneous nature of the crust and a low Qc value was demonstrated around Tezpur and western part of Arunachal Himalaya near Bomdila based on their study of coda waves attenuation in northeastern India (Hazarika et al., 2009). This could be the reason for the remarkable low LVZ around this area. Another striking LVZ is described by the Mikir Hills area extending towards the Naga and Disang thrust. The clustering of earthquake hypocentres along the Kopili fault and the Mikir Hills following the NW-SE trend and running parallel to similar clustering of earthquakes in the western and southwestern extremity of Shillong Plateau overlapping these strikingly low velocity zone are features seen in all the lesser depth slices above. But then below this depth of 40 Km this trend goes missing and is replaced by prominently dominant intermediate to very high velocity zones. Therfore, this LVZs
at 40 Km depth is the reflection of the seismogenic base and the HVZ beneath is the
reserver of seismic stress in lowermost crust and uppermost mantle that triggers the
seismicity seen in the seismogenic base. It is obvious to expect LVZs around the Naga
and Disang Thrust, because these thrust zones are characterised by higher degree of
heterogeneity due to intense deformation and also might be due to the possible
presence of heat and fluids causes the slowing down of seismic waves (Kuwahara et
al., 2003) and, therefore, this fact accounts for the manifestation of low Vp (~ 6.5-7
km/s)

4.9.6 For Depth Slice at 50 km (Figure 4.20 f)

This depth slice describes the uppermost mantle and shows at velocity
variation of P waves from 7.5-9 km/s which is indeed a very high Vp zone. Since the
seismicity is also very sparse so undoubtedly this must be a very strong geologic layer
but to discuss in details about the exact lithology requires additional seismic and
geological investigations

Therefore, from the aforementioned results based on the correlation and
corroboration of the established geology of the study area and the variation of seismic
velocity (Vp) refers to the presence of strong lateral and depth wise heterogeneity in
the lithosphere of the study area. It also throws light on the fact that definitely strong
hard rock lithology defines ~10-40 km depth and even beyond. The gradual decrease
of earthquake hypocentres at higher depths (~ 40 km) and the shooting up of the Vp
values to 8 km/s and which goes up to 9 km/s in the uppermost mantle, is indicative of
an absolutely strong and rigid lowermost crust which is declining in its brittleness. A
reflection of rheological transition of lower crust-to-Moho-to upper most mantle is
understood.
4.10 Depth Cross-Section

The depth cross-sections for Vp were prepared along E-W profile at 25° N, 26° N and 27° N latitudes and the depth cross-sections have been shown as Figure 4.23, 4.24 and 4.25 respectively. In depth cross-section at 25° N latitude shown in Figure 4.23, dominance of LVZs underneath of Shillong Plateau and Dauki fault upto about 20 Km. This patch of LVZ shrinks and narrows down from Disang Thrust (DsT) to Indo-Burman Ranges (IBR). Disang Thrust show low velocity lithology approximately upto 8 Km. Thinning of low velocity material at IBR is seen. There is pinching in of HVZs beneath Shillong-Mikir Plateau at 40 Km. From tomograms in depth range 0 to 30 km (Figure 4.20 a-d) high seismic activity is observed to the west of Dauki Fault which diminishes beyond 40 km depth (Figure 4.20 e-f). Also prominent isolate patch of HVZ at 40 km depth tomogram (Figure 4.20 e) is seen overlapping the Dauki Fault. The HVZ may be representing the basement rocks which are hard rocks at a depth of 40 Km in the Dauki fault and may be influencing the seismicity being triggered at shallower depths of 0 to 30 km in the western part of the plateau due to localised built up of stress.

In the depth cross-section at 26° N latitude shown in Figure 4.24 maximum resolution is observed and hence the inversion is far more reliable. Correlating with the tomograms it is witnessed two conspicuous patches of earthquake hypocentres, one lying below Shillong Plateau (SP) and the other one beneath Mikir Hills region. A distinct patch of LVZ is seen upto about 12 Km which slims down from Naga Thrust (NT) to IBR. The low to intermediate velocity anomaly of Vp beneath the Indo-Burma Ranges (IBR) is seen conspicuously upto 15 km depth, beyond which it gradually gives way to higher values of Vp. The subducting lithosphere would be deforming when it bends down at shallower depths and that would cause an increase in the
heterogeneity and would lead to low velocity. Pinching in of HVZ of at 30 Km depth underneath IBR is seen. The metamorphosis of sediments at greater depth and the presence of colder subducting lithosphere relative to the temperature conditions in its vicinity would lead to such a velocity variation.

The depth cross-section taken at $27^\circ$ N latitude shown in Figure 4.25 is very heterogeneous laterally as evident from pinching in of high velocity materials into low velocity zone at depth range of 20 to 40 Km. There is scanty earthquakes beyond 40 Km depth slice which supports the HVZ seen at this depth section below 40 Km. The low velocity zone thickens from Kopili fault to the upper Assam basin and Disang Thrust upto about 5 Km which is indicative of thickening of the sediment column between Brahmaputra river and Disang Thrust. The heterogeneity of MCT and MBT is very prominent in this section upto about 5 Km. From the tomogram at 50 Km depth slice (Figure 4.20 f) it is seen that the seismicity is sparse and the the shooting up of the Vp values to above 8 km/s in this depth cross section, is indicative of an absolutely strong and rigid lowermost crust which is declining in its brittleness before giving way to upper mantle. A rheological transition zone from lower crust-to-Moho-to upper most mantle is seen. The Moho could be somewhere around 45 km depth.

4.11 Discussion

A visual inspection of the distribution of earthquake epicentres in relation to the major geological structures of the northeastern region of India (Figures 1.2 & 4.3) shows clustering of epicentres in four distinct tectonic domains of this region viz. (1) Shillong Plateau, (2) Mikir hills and Lower Assam valley, (3) Arunachal Himalaya and (4) the Indo-Burman Ranges (IBR). The seismic activity in the studied region shows major seismic activity mostly upto 40 km depth, beyond this depth the distribution of earthquakes is sparse and scattered (Figures 4.8). It is observed that in
the Indo-Burma Ranges (IBR) deep seated earthquakes mostly occur upto around 200 km of depth (Figures 4.10 a & b) located by inverted 1-D velocity model. The epicentres located by employing 3-D velocity model are mostly confined in the Shillong-Mikir Plateau and the Kopili Fault zone of Assam Valley (Figure 4.5). The density of epicentre distribution in the IBR has drastically declined. The epicentres are sparsely distributed in the Arunachal Himalyas and the Tripura Fold belt areas. The RMS values as seen during the location of the events using 3-D velocity model is a major improvement, which improved to 0.05 sec as compared to the ones located with Hypoinverse using a priori velocity model and by optimum inverted 1-D velocity model (Figure 4.16). A comparison of the average errors in longitude, latitude and depth have been shown in Figures 4.13, 4.14 and 4.15 respectively. The average errors in longitude, latitude and depth reduced to 2.16 Km, 2.98 Km and 2.08 Km respectively which is a marked improvement in location of events by employing 3-D velocity model. Figure 4.11 shows the Frequency distribution of the relocated earthquakes with respect to depth obtained with 3-D inverted velocity model which indicates that the 3-D velocity located events have an improved constraint over depth because the maximum depth to which events are distributed have come down to between 100.1 to 120 Km which was between 180.1 to 200 Km with the events located by inverted 1-D velocity model. Figure 4.12 shows frequency distribution of the relocated earthquakes with respect to RMS obtained with 3-D inverted velocity model which improved remarkably with 616 earthquakes having RMS in the range of 0-0.1 sec which was 278 earthquakes in the range of 0-0.1 sec located by 1-D inverted velocity model which means that the 3-D velocity relocated earthquakes are all very highly constrained and the results are very much reliable.
The Checkerboard Recovery Test (CRT) reveals a better recovery in the central portion of the model with DWS > 50 for most of the layers except the fringe areas of the model. The checkerboard obtained with alternative high and low velocity is in concurrence with the actual Vp (Figures 4.22).

The tomogram (Figure 4.20 a) at the shallow subsurface (0 km) shows a typical trend of LVZ flanked by HVZ on either side is noticed in Shillong-Mikir Plateau. The central and eastern part of the plateau and the Mikir Hills is conspicuously HVZ. The seismicity is very sparse mostly confined to western part of Shillong Plateau in NW-SE trend which is a LVZ and extends towards western Assam. At this depth slice seismicity associated with Mikir Hills and Kopili fault zone is observed to be very sparse. The HVZ seen in the shallow subsurface start gradually vanishing in the upper crust slice of 10 km depth (Figure 4.20 b). In the upper crust at depth slice of 10 km mostly LVZ (~5.5 km/s) is observed. The HVZ (6 km/s) is clearly seen to be diminishing everywhere in the resolved part of the depth slice. Prominently, LVZ (~ 5.5 km/s) appears in the western extremity of the Shillong Plateau which is in conformity with to the clustering of the shallow faults like Chedrang, Dudhnoi, Samin and Dapsi faults oriented NW-SE, N-S, NW-SE and E-W respectively. A NW-SE trend of densification of seismicity is clearly observed. Seismicity is seen to be very sparse in Mikir Hills at a depth of 10 km and is inferred as LVZ (~5.5 km/s). Seismicity is not very strong along Kopili fault zone at this depth which shows a concentration of HVZ (6-6.5 km/s). Now a HVZ is flanked by two LVZ which is the opposite situation of what it is witnessed at 0 km depth. The LVZs (~6 km/s) becomes more pronounced at depth of 20 km (Figure 4.20 c) with the NW-SE trend of seismicity becoming denser in the western part of Shillong Plateau. LVZ (~5.5-6 km/s) is also seen in the eastern plateau especially the Barapani Shear Zone.
(BSZ) (Kayal and Zhao, 1998; Kayal, 2008). Shear zones undergo intense shearing, deformation and fracturing and mylonitization especially in ductile shear zones. Mineral grains and crystals are reduced in size and lots of irregularities in the mineral fabric develop, micro faulting within the minerals and generation of heat are some characteristics of shear zones which is why the seismic wave velocity gets reduced. Appearance of marked LVZ (6 km/s) and seismicity with a NW-SE orientation in the Mikir Hills has been witnessed which is due to the heterogeneous and highly fractured nature of the fault zone. Apart from this area, transition from HVZ to LVZs or intermediate velocities (6-6.5 km/s) are to be seen all across the Brahmaputra valley, Kopili fault, Naga thrust, the Belt of Schuppen and the Disang thrust which are expected of thrust, fault and shear zones due to heterogeneity introduced into the crust and perhaps because of presence of fluids and possible generation of heat. Albeit, the Indo-Barman Ranges (IBR) is not resolved well, there is a noticeable cluster of HVZ (7 km/s) at which the IBR takes a hair pin arc like turn with meagre distribution of seismicity. The HVZ might act as a reserve of seismic stress and probably consequences in the frequent occurrence of earthquakes in and around Manipur (India)-Burma border, but then the entire stretch of IBR is imaged as LVZ which this becomes clear at 40 km depth. Two prominent clustering of HVZs (7-6.5 km/s) are seen along MBT and MCT which might be a reflection of asperity.

The most intriguing point is these fault zones and shear zones like Dudhnoi fault, Kopili fault zones, Barapani shear zone and all across western Shillong Plateau defined by faults are overlapping with LVZs and are defined by pronounced seismicity. The same trend NW-SE continues at 30 km depth (Figure 4.20 d) with LVS (~ 6.5 km/s) along Kopili fault trending NW-SE shows up with more of its seismic activities. The seismic activities along LVZs (~6.5 km/s) in the western
Shillong Plateau and Mikir Hills in NW-SE orientation continues but with relatively lesser density of hypocentres compared to 20 Km depth. Kopili fault seismicity is strongly visible. At 40 km depth (Figure 4.20 e) albeit prominence of LVZs (~ 6.5-7 km) continues, intriguingly two very conspicuously LVZs (~ 6.5 km/s) shows up in the northern Brahmaputra valley around Tezpur spreading up to MBT and another one at Mikir Hills passing over to the Naga and Disang Thrust (Figures 4.20 e). Hazarika et al. (2009) demonstrated a low Qc value around Tezpur and western part of Arunachal Himalaya near Bomdila based on their study of coda waves attenuation in northeastern India. This is in concurrence with dense fracturing and criss-crossing of faults and lineaments that characterize the particular site which complicates it further with the presence of MBT and MCT. Seismic wave attenuation is high (low Q) in LVZ areas and lower seismic wave attenuation (high Q) is associated with HVZ (Furumura and Singh, 2002; Ottemöller, 2002; Ojeda and Ottemöller, 2002; Ottemöller et al., 2002 and Kayal, 2008). Another striking LVZ is described by the area along the Naga and Disang thrust. This is obvious to expect because fault zones are characterised by higher degree of heterogeneity and likely presence of heat and fluids, causes the slowing down of seismic waves (Kuwahara et al., 2003) and, therefore, this fact accounts for the manifestation of low Vp (~ 6.5-7 km/s) along the Naga and Disang Thrust (Figure 4.20 e).

The seismicity in the western extremity of Shillong plateau starts dwindling and so is the concentration of hypocentres in the Mikir Hills diminished with Vp at ~ 6.5-7 km/s. The Kopili fault still shows seismic activity at this depth level too (Bhattacharya et al., 2008, and Bhattacharya et al., 2010). It is validated that Kopili fault is associated with intense deeper level of seismicity (35-40 km) and the roots of the fault may be traced to depth beyond 40 km. It should be somewhere between 40 to 50 km. As seen in Figure 4.20 f at 50 km slice that the zone along Kopili fault is
mostly replaced by HVZ (~8.5 km/sec) which is almost devoid of earthquake hypocentres. This depth could be the zone of concentration of seismic stress and fuels seismic activities at shallower depths above in this fault zone. It has been championed by many workers that 90% of the earthquakes occur at the base of the seismogenic zone (Sibson, 1986; Lay and Wallace, 1995; Kayal, 2000; Kayal et al., 2002; Mishra and Zhao, 2003). They asserted that large magnitude earthquakes tend to nucleate at the base of the seismogenic zone which is considered to be the fault end. Bhattacharya and Kayal (2003) and Kayal et al. (2006) identified it as a zone with a high risk for an impending large earthquake in future.

The image at 50 km depth (Figure 4.20 f) representing the uppermost mantle is almost devoid of hypocentres and strictly HVZs (~8.5-9 km/s) and almost a thorough replacement of LVZs is observed here. The occurrence of upper mantle seismicity is idiosyncratic to regions of recent continental convergence (Chen and Molnar, 1983).

The discussion in geological validation of the tomographic model unambiguously emphasized for the abundance of hard rock lithology especially from ~10-40 km crustal depth and beyond and to the existence of a strong and rigid lower crust with declining brittleness before the advent of the Moho. This is well substantiated by distribution of earthquake hypocentres and also by the Vp values at different depth slice obtained from the 3-D tomographic model which are supported by the global P-wave velocities for different hard rocks. The seismicity seen upto 40 km is unequivocally indicative of a hard brittle lithology to source the earthquakes at those depths (10-40 Km). Since there is evidence of seismicity from depth of 0 km upto 40 km, the study has reservation with the notion that an aseismic weak lowermost crust (35-45 km) remains sandwiched between the brittle lower crust and rigid uppermost mantle for Shillong-Mikir Plateau and Assam Valley as observed by Chen and Molnar (1983). It seems that the lowermost crust is hard and strong as
reflected by the shooting up of Vp (8 km/s) and presence of seismicity at 40 km depth, which is unlike the weak lowermost crust of the “Jelly-Sandwich” model of Molnar and Chen (1983) and Chen and Molnar (1983, 1990). Thus, this study also emphasizes that the entire crust beneath the study area is brittle and seismogenic with the seismogenic depth being at around 40 km and then it gives way to a strong and rigid lowermost crust before the crust assimilates with the Moho. The observation made by Maggi et al. (2000 a,b) and then reiterated by Mitra et al. (2005) based on recomputed depths of earthquakes employed by Chen and Molnar (1983, 1990), using receiver function techniques to arrive at a 1-D velocity model, concluded that the entire crust beneath northeastern India is seismogenic and the present study also reflecting the same. The lowermost crust (beyond 40 km) could thus be the reservoir of seismic stress that triggers seismicity at lesser depths above it, which is seen to be overlapping the LVZs and clustering of hypocentres.

Borah et al. (2016) based on their study of crustal structure beneath northeastern India inferred through receiver function techniques asserted that there is a significant variation in crustal thickness for northeastern India which increases from south to north. They argued based on inverted shear wave velocity models that crustal thickness varies from 32-36 km in Shillong Plateau (South), 36-40 km in Assam Valley and ~ 44 km in Lesser Himalaya (North). The average Vp/Vs for Shillong Plateau was found to be 1.73-1.77 compared which is lesser than the Assam valley (~1.73-1.84) and Lesser Himalaya (~ 1.80). In this study the base of the crustal seismogenic zone has been found to be ~ 40 km which seems like to be at the base of the upper part of lower crust and which finally gives way to a sparsely seismic HVZ in the uppermost mantle (~ 50 km). Therefore, it is expected that the Moho for the study area should be on an average ~ 5-7 km farther down from the seismogenic depth of 40 km, before the HVZ of the uppermost mantle sets in. Due to paucity of
earthquake hypocentres at depth greater than 50 km, the Moho boundary could not be properly ascertained but the seismogenic zones and their base is imaged clearly.

4.12 Conclusion

The optimum 1-D velocity model obtained in this study helped to generate a 3-D velocity structure for the study area which reveals various facets of this study. The 3-D velocity model also well constrained the location of the earthquakes. The tectonic regime of northeastern India is a very complex one which consequences in a complicated structural framework. The crustal stress distribution of this region is completely governed by the collisional tectonics of the Himalayan arc in the north and the subduction tectonics in the eastern margin of the region. This is well reflected by the presence of an entirely seismogenic crust in the study area. An intricate relationship between tectonics, structure, geology and the manifestation of seismicity is seen for the study area. This is validated by the fact that the structural trend of the study area which is NW-SE is conformity with the distribution of seismicity which is also the orientation of major seismogenic faults, fracture zones etc. and is overlapped by intermediate to low velocity zones. The Shillong-Mikir Plateau which is a hard rock terrain is expected to be HVZ, but then the dominant occurrences of intermediate to low velocity zones is somewhat conflicting. This may be understood in the light of the fact that the contributions of multiple tectonic regime in making the crust of study area highly heterogeneous is pivotal, which is why LVZs and fault, fracture and shear zones are seen to be overlapping. The occurrence of LVZ in the otherwise hard rock terrain of the Shillong-Mikir Plateau which dominantly covers the tomographically resolved part of the study area is also indicative of the presence of liquid phase (or melting phase), especially at a depth of 30-40 km. This study reflects a heterogeneous and seismogenic crust for the study area and the dominant presence of intermediate to low velocity zone could cause a substantial increase in seismic wave attenuation that
enhances the seismic hazard and seismic vulnerability of the study area. Identification of conspicuous clustering of extremely high velocity zone at Mikir Hills at depth of 0 km, MCT and MBT at depth of 20 km and IBR at 10-20 km could be a precursor to potentially powerful earthquakes in the future which is an index to the seismic threat posed by the study area. There is a valid reason to this interpretation because these identified high Vp zones could be asperity zones (Kato et al., 2007) as there are either sparse seismicity or no dense concentration of earthquakes are to be seen, which means stress has not been released. Such conditions cause hardening of the crust which results in a high Vp.

The findings of this study are properly substantiated by geology and tectonic of the study area which has been tomographically resolved mostly in the Shillong-Mikir Plateau and Assam Valley areas. Since whole of the study area has not been tomographically resolved, therefore, scope remains to further this study covering areas of Arunachal Himalaya, Indo-Burman Ranges (IBR) and the Tripura Fold with the availability of more robust data through enhanced network coverage.
Figure 4.3: Distribution of earthquake epicentres located by Hypoinverse earthquake location programme shown by red solid circles and seismic stations shown by blue solid triangles

Scale: $1^\circ = 111$ Km
Figure 4.4: Epicentral distribution of earthquakes relocated by using inverted 1-D velocity model

Scale: $1^\circ = 111$ Km
Figure 4.5: Epicentral distribution of earthquakes relocated by using 3-D velocity model

Scale: $1^\circ = 111$ Km
Figure 4.6: Map of northeastern region of India showing the distribution of seismic stations from different seismic networks, the data of which have been used in this study.

Scale: $1^\circ = 111$ Km
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Figure 4.7: Year wise distribution of 892 earthquakes used for inversion in the 1-D and 3-D velocity structure study for northeastern India
Figure 4.8: Frequency distribution of the relocated earthquakes with respect to depth obtained with 1-D inverted velocity model

Figure 4.9: Frequency distribution of the relocated earthquakes with respect to RMS error obtained with inverted 1-D velocity model
**Figure 4.10 (a)** Depth section of the located earthquakes with respect to longitude obtained with inverted 1-D velocity model

**Figure 4.10 (b)** Depth section of the located earthquakes with respect to latitude obtained with inverted 1-D velocity model
**Figure 4.11:** Frequency distribution of the relocated earthquakes with respect to depth obtained with 3-D inverted velocity model

**Figure 4.12:** Frequency distribution of the relocated earthquakes with respect to RMS obtained with 3-D inverted velocity model
Figure 4.13: Plot shows comparison of average error in longitude for earthquakes located with 1-D priori velocity model (De and Kayal, 1990), inverted 1-D velocity model and 3-D inverted velocity model.
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Figure 4.15: Plot shows comparison of average error in depth for earthquakes located with 1-D priori velocity model (De and Kayal, 1990), inverted 1-D velocity and 3-D inverted velocity.
Figure 4.16: Plot shows comparison of average error in RMS (in sec) for earthquakes located with 1-D priori velocity model (De and Kayal, 1990), inverted 1-D velocity and 3-D inverted velocity
Figure 4.17: Ray density for Vp. The depth of layers are shown at the top corner on the left. The longitudes are shown in the X-axis and latitudes in Y-axis in degrees.
Figure 4.18: Derivative Weight Sum (DWS) for Vp. The depth of layers are shown at the top corner on the left. The longitudes are shown in the X-axis and latitudes in Y-axis in degrees.
Figure 4.19 (a): Tomogram at a depth of 0 Km for the events having RMS < 0.3 inversion grid of 30 km. The following are the descriptions of the basic features of the tomogram which remain the same for all the tomograms till Figure 4.21. The red coloured triangle are the location of the seismic stations and the solid black squares are the clusters of earthquake hypocenters. The white border indicate the extent of the resolved part of the study area. The colour bar in the right indicates the velocity of P-wave (Km/s). Following are full form of the abbreviations used in the Figure.

MCT: Main Central Thrust; MBR: Main Boundary Thrust; SP: Shillong Plateau; DF: Dauki Fault; KF: Kopili Fault; MH: Mikir Hills; NT: Naga Thrust; DT: Disang Thrust; IBR: Indo-Burman Ranges.
Figure 4.19 (b): Tomogram at a depth of 10 Km for the events having RMS < 0.3 with inversion grid of 30 km

Figure 4.19 (c): Tomogram at a depth of 20 Km for the events having RMS < 0.3 with inversion grid of 30 km
Figure 4.19 (d): Tomogram at a depth of 30 Km for the events having RMS < 0.3 with inversion grid of 30 km

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Figure 4.20 (b): Tomogram at a depth of 10 Km for the events having RMS $< 0.5$ with inversion grid of 30 km.
Figure 4.20 (c): Tomogram at a depth of 20 Km for the events having RMS < 0.5 with inversion grid of 30 km

Figure 4.20 (d): Tomogram at a depth of 30 Km for the events having RMS < 0.5 with inversion grid of 30 km
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**Figure 4.21 (b):** Tomogram at a depth of 10 Km for the events having RMS < 0.5 with inversion grid of 50 km
Figure 4.21 (c): Tomogram at a depth of 20 Km for the events having RMS < 0.5 with inversion grid of 50 km

Figure 4.21 (d): Tomogram at a depth of 30 Km for the events having RMS < 0.5 with inversion grid of 50 km
Figure 4.21 (e): Tomogram at a depth of 40 Km for the events having RMS < 0.5 with inversion grid of 50 km

Figure 4.21 (f): Tomogram at a depth of 50 Km for the events having RMS < 0.5 with inversion grid of 50 km
Figure 4.22 (a): Results of the checkerboard resolution test for the image at depth of 0 Km with RMS < 0.5 with the inversion grid of 30 km

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Figure 4.22 (c): Results of the checkerboard resolution test for the image at depth of 20 Km with RMS < 0.5 with the inversion grid of 30 km

Figure 4.22 (d): Results of the checkerboard resolution test for the image at depth of 30 Km with RMS < 0.5 with the inversion grid of 30 km
Figure 4.22 (e): Results of the checkerboard resolution test for the image at depth of 40 Km with RMS < 0.5 with the inversion grid of 30 km

Figure 4.22 (f): Results of the checkerboard resolution test for the image at depth of 50 Km with RMS < 0.5 with the inversion grid of 30 km
Figure 4.23: Vp cross section in the E-W direction at 25° N latitude. IBR: Indo-Burman Ranges; DsT: Disang Thrust Longitude is represented in degree in the X-axis and depth in Km along Y-axis.
Figure 4.24: Vp cross section in the E-W direction at 26° N latitude. AV: Assam Valley; KF: Kopili Fault; MH: Mikir Hills; NT: Naga Thrust; Belt of Schp: Belt of Schuppen; DsT: Disang Thrust; IBR: Indo-Burman Ranges
Longitude is represented in degree in the X-axis and depth in Km along Y-axis
Figure 4.25:  Vp cross section in the E-W direction at 27° N latitude. 
MCT: Main Central Thrust; MBT: Main Boundary Thrust 
KF: Kopili Fault; DsT: Disang Thrust 
Longitude is represented in degree in the X-axis and depth in Km along Y-axis