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Evaluation of Crustal Structure Beneath The Central Part of the "Bay Of Bengal" from Surface Wave Dispersion Studies.

Dibakar Ghosal*
National Geophysical Research Institute, Hyderabad, India, Email- dibakarghosal@yahoo.co.in

Summary

Surface waves from aftershocks (magnitude>5) of the great Sumatra Earthquake recorded at the STS-2 broad-band seismometer, located at Hyderabad, are used to estimate the average shear wave velocity structure of the Bay of Bengal. The fundamental mode Rayleigh waves are used to construct Rayleigh wave group velocity distribution maps of the region using multiple filtering techniques. These group velocities are plotted against the corresponding periods, and thus obtained the dispersion curve for each of these events. Later these events are classified into two clusters based on source location and similarity of structure sampled. An error weighted average, group velocity dispersion curve is obtained for each of these clusters and are inverted individually to look at the one-dimensional shear wave velocity structure beneath the Bay of Bengal. Two velocity models were obtained for each cluster with a good match between the calculated and observed dispersion curves. The first is the model that gives the number of interfaces while the second yields a model in terms of the number of layers. In order to have a good resolution between the layers including the thin strata, the velocity gradient need not be extremely smooth which otherwise will produce an average effect. In determining the model with the minimum number of layers, the early stage inverted models were examined and adjoining layers that had similar wave speeds were merged into a single homogeneous layer. This model was used as the starting model for subsequent inversion. This process was repeated until satisfactory model whose forward solution matched the significant features of the measured dispersion curves. The final models suggest a thick continent-like crust beneath the Bay of Bengal. The average depth to the crust-mantle boundary is 30±3km. Later the sensitivity of the obtained model by inversion is checked by perturbing the layer thickness. A forward modeling is done after each perturbation and the synthetic dispersion curve is compared with the observed dispersion within a predefined error bound. The test is done on two boundaries of each of the two clusters; on the metamorphosed sediment-upper crust boundary and on the crust-mantle boundary.

Introduction

Classical plate tectonic theory tells us that a stronger lithosphere overlies a relatively weaker asthenosphere and the lithospheric plates remain unreformed except at the plate margins. Indian plate is the most significant exception of this theory. It can easily be observed that deformations are so much prominent in the central and northern Indian ocean, Indian subcontinent (Peninsular India), and the adjoining Bay of Bengal which is unique among the world's oceanic areas, having the thickest (~22 km beneath the Bangladesh shelf) young sedimentary deposits, caused by accumulation of rapidly eroded sediments of Himalayas at the hard waters of the Ganges and the Brahmaputra. Marine reflection profiling and gravity studies reveal widespread deformation and originally flat lying sediments, particularly south of the Bay of Bengal , which includes folding, caused by N-S compressive stress of 4-5 kbars, reverse faulting and unusual undulations

To study velocity variations, vertical component of Raleigh waves of Java – Sumatra – Andaman earthquake of December 26, 2006, were analyzed to estimate the surface-wave group dispersion for known propagation paths. The earthquake initiated where the incoming Indian Plate lithosphere is the warmest and the dip of the Wadati-Benio. Zone is least steep along the subduction zone extending from the Andaman Trench to the Java Trench. Group velocities were estimated using a multiple filter method upon 29 events, recorded at broadband seismological observatory, Hyderabad. To reduce the data the earthquakes are grouped into 2 clusters and the group velocities of each clusters are averaged. Finally, we have found out the velocities at different depth of layered structures of Bay of Bengal, applying the inversion theory upon the clustered average group velocities.

Theory and Method

Constructive interference of incident P and S -waves arriving at the free surface causes the generation of the Surface waves (Rayleigh and Love waves) which propagates through the upper surface of the Earth Rayleigh wave particle motion, which is the combined effect of P and SV waves, is retrograde elliptical in the direction of propagation to a depth $1/5^{th}$ of the wavelength of the Rayleigh wave, where it goes to zero. Below this depth, elliptical particle motion becomes prograde. A Rayleigh wave is found in the vertical and radial plane with no tangential motion. . The particle motion for Love waves, which caused due to the conversion of SH waves, is parallel to the surface but perpendicular to the direction of propagation and found on the transverse record of the seismogram. Because the amplitude of surface waves decreases with increasing depth hence, they are affected by lateral variations in structure.

During propagation through an anelastic medium, both the amplitudes of Rayleigh waves and Love waves decay exponentially with depth in the half space and the solution exists as a fundamental mode and a finite number of modes for which amplitudes become zero, before the entire wave field vanishes, at a number of points (nodes) equal to the overtone number minus one. Fundamental modes, which are slower than that of the higher modes and easily observed in seismograms, have chosen as the velocity measuring parameter of surface-waves.

Dispersion, where the velocity of a wave on the surface is dependent on its frequency (or period), is the governing property of Surface wave. Surface-wave-trains behave like a energy packet and propagate with their own velocity called the group velocity, with different frequencies arriving at different times. Longer period waves are faster because they sample deeper into the Earth where the velocities are higher on the other hand, short period waves behave just the opposite; they are slower because they travel through lower velocity material. Because surface waves only sample the upper few hundred kilometers of the crust and upper mantle, phase and group velocities can be used to determine upper crustal structure by looking at velocity changes with depth. Group velocities can be defined as the material parameter and are slightly more sensitive to the medium parameters than phase velocities, which favor their use in a S wave velocity inversion study.
In this study, we use fundamental mode and vertical component of Rayleigh waves to calculate group velocity dispersion for earthquake-receiver paths across the central part of the Bay of Bengal from the Sumatra aftershocks recorded at the Hyderabad Broadband seismological observatory and invert these dispersion data to obtain one-dimensional shear wave velocity structure beneath the region.

1. Dispersion Measurements

Deconvolution removes the instrumental response from the received data which are then filleted by applying a band pass Gaussian filter (10-120 sec) to prevent emphasizing of low frequency and high frequency noise. Signal-to-noise ratio gives the idea about the quality of recorded signals .If this ratio becomes greater than two, then it will be feasible for dispersion analysis. Background "noise" may be other energy (e.g. body waves) that arrives before the arrival of the surface waves.

2. Group Velocity Measurement

Multiple Filter Technique (MFT), developed by Dziewonski et al. (1969), is used to determine the group velocities of surface waves, depending on its dispersive nature. It is composed by mainly an array of narrow filters which may resolve transient signals comprised of several dominant periods that arrive at the recording station almost simultaneously. Using the filtered amplitude, we calculate the group velocity for each of the above station–earthquake pair.

Group velocity is a dispersive characteristic where constructive patterns travel along the surface as wave packets and the individual waves that make up the wave packet travel with phase velocity.

The packet of energy that propagates as a surface wave contains a spectrum of periods and each of these periods can be measured from the time between successive peaks or troughs. The wave with longest periods travels fastest, and appears first on the seismogram. The group velocity is found by dividing the distance between the source and the receiver by the travel time of the wave group.

This method can be applied in a more sophisticated way by using Multiple Filter Analysis to isolate wave groups of different periods. This filter is a Gaussian filter with a parameter α controlling the band width of the filter. The result is plotted as a function of period which is called the dispersion curve (Fig -1).

A Gaussian filter with peak amplitude centered at the desired period is applied to the seismogram in the frequency domain. The peak of the envelope of the corresponding time domain signal is used to estimate the group travel time. An instantaneous period is measured at the time of the envelop peak. The group velocity for a given period is estimated by dividing the distance between the station and source by the group arrival time. The process is repeated for each period in a specific range and is plotted verses group velocity. Isoclines are drawn connecting the same energy levels for the traces with different frequencies of the same seismogram. Now we pick the group velocity points continuously in the high energy contour region and is plotted against the period, gives the dispersion curve (Fig-1).

After the estimation of group velocities, a mode isolation filter, having a smooth, stable spectral amplitudes and eases estimation of signal phase, is applied to the signal to isolate the fundamental mode from the generated signal and ambient Earth noise.

After the completion of the dispersion calculation for all the events, we apply a spline fit and interpolation of the dispersion data to obtain a continuous group velocity values and to smoothen out the dispersion curves. This spline fitted dispersion curves of all events are clustered into two groups based on their back azimuth and epicentral distance.

3. Clustering

3.1 Introduction

Based on lattitude and longitude of events locations, I have clustered the whole of 29 broadband seismograms into two groups to reduce the data points and to examine whether all paths are having the same structure.

For each clusters we obtain an error weighted average dispersion curve of all individual dispersion curves. The error in group velocity is not the true error, since the dispersion is based on only one observation. Since the Gaussian filter has a longer impulse response at longer periods, a reading error could be associated with a misplaced maximum – the number computed here assumes that the travel time can be mis-measured by one filter period. We are applying an inverse error weight for each group velocities so as to assign a low weight for group velocity having large error value and to give more weight for group velocities with low error value.

Error weighted group velocity,

$$V_{ew} = \{(V_1*1/E_1 + V_2*1/E_2 + \ldots\ldots\ldots + V_N*1/E_N) / (1/E_1 + 1/E_2 + \ldots\ldots\ldots\ldots + 1/E_N)\}$$

Then we find the average of all the error weighted velocities corresponding to each period for all the events in a particular cluster. This forms the new group velocity and the average of the standard deviations of the error weighted average velocity and individual velocities are also being calculated.

3. 2 Cluster 1

Location of Cluster 1-Events are in Figure 2 (right) along with the individual spline fitted dispersion curves for all the events in this cluster plotted together in one plot. An error weighted average dispersion curve is calculated for this cluster and is shown in Figure 4 (right). The highlighted curve is the error weighted dispersion curve within error bounds, while the other curves are the individual observed spline fitted dispersion curves for cluster1. The average of the standard deviations of the error weighted average velocity and individual velocities for a particular period accounts for this error bound for the average one dimensional shear wave velocity structure beneath the region, for this error weighted average dispersion curve.

3.3 Cluster2

Positions of events in Cluster 2 are shown in Figure 3 (right) along with the individual spline fitted dispersion curves for all the events in this cluster plotted together in one plot.

4 Inversion: An Overview

Using all the estimated group a depth dependent shear wave velocity structure is estimated with the help of inversion program which is based on inverse theory. A common approach to the problem of estimating elastic properties of the Earth from seismological observations consists of assuming that the Earth is composed of horizontal homogeneous layers which are infinitesimally small, and then inverting the observations for the seismic properties of each layer. An inversion in seismology usually involves having more observation than unknowns, making it an over determined problem. The general form for this type of inversion is:

$$d = G\delta m$$

where G is a matrix containing partial derivatives of the group velocities with respect to the layer shear velocity, compressional velocity, and group velocity, d is a vector containing the observations (i.e. dispersion estimates), and δm is the correction vector of model parameters (i.e. shear-velocity model). Model parameters are perturbed from the initial guess of the model and fit to the observed values. If the size of the perturbations is kept small (by use of a damping factor) then the relation between the observable and parameter perturbation is linear.

In this study we use the model suggested by Brune and Singh (1986) as the initial model. The inversion process is repeated through *iterations* where the model is perturbed and the resulting fit is observed. The number of iterations is monitored to minimize extraneous artifacts being introduced into the model (e.g. low-velocity zones). During each iteration, the standard deviation between the observed and predicted models should reach a point where the change in the model is negligible (i.e. the model converges).

As mentioned earlier, models from previous studies are used as a starting point for the inversion. The ultimate goal for inversion is to resolve the upper crustal structure in term of the shear velocity in order to better understand how energy passes through a given region. We thinly parameterized the initial model for which the crust and the upper mantle lie on top of a half-space. The crust consists of three layers: a sedimentary cover, an upper crust, and lower crust; the half-space velocities were consistent with values observed on the upper mantle. The initial models contained thin 2 km thick layers for 100 km in order to better resolve the crustal and the upper mantle structure. Below 100 km lies an infinite half space. The upper crust is especially sensitive to short-period a surface wave, which is while the short period part of the dispersion curve reveals the shallow earth structure; the higher periods give upper mantle structure. We use the same initial model in order to obtain final velocity structures reflecting the differences in the dispersion curves of different path groups but free from bias from initial models. In doing the above analysis, we have assumed that the medium is isotropic and we only do simple one-dimensional modeling.

Once both the predicted and the observed model converge, I coarsely parameterized the layers having nearly same velocities. This gives the thickness of different layers. Also a model with minimum number of interfaces is made by dividing the thick layers into small individual layers and applying a gradational increase or decrease in velocities to those layers. The final model is chosen on how well is fits the observations and if the model is consistent with the regional geology

We obtained two velocity models were obtained for each cluster with a good match between the calculated and observed dispersion curves. The first is the model that gives the number of interfaces while the second yields a model in terms of the number of layers. In order to have a good resolution between the layers including the thin strata, the velocity gradient need not be extremely smooth which otherwise will produce an average effect. In determining the model with the minimum number of layers, we examined the early stage inverted models and for adjoining layers that had similar wave speeds, and merged these into a single homogeneous layer, and this model was used as the starting model in the subsequent inversion. We repeated this process until satisfactory model whose forward solution matched the significant features of the measured dispersion curves. The two final models for each of the 2 clusters cluster 1, and cluster 2 are displayed in Figure 5 (left & right) respectively.

Figures

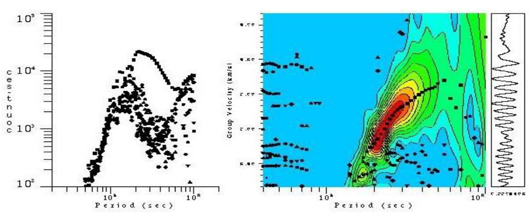

Fig-1: Screen plot of the MFT program. At the right side, the plot of the group velocity dispersion curve with period is displayed; the color represents the filtered envelope values as a function of velocity and period. The red color represents the highest amplitude. At the left side the plot shows the maximum amplitude of the envelop per period.

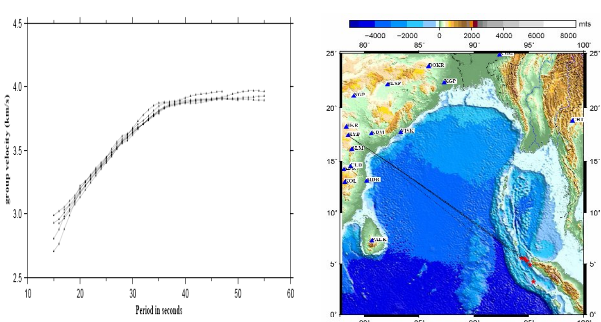

Fig-2: Dispersion curves of all events in cluster 1 and the right hand figure shows the location and the travel path of all the events in cluster 1

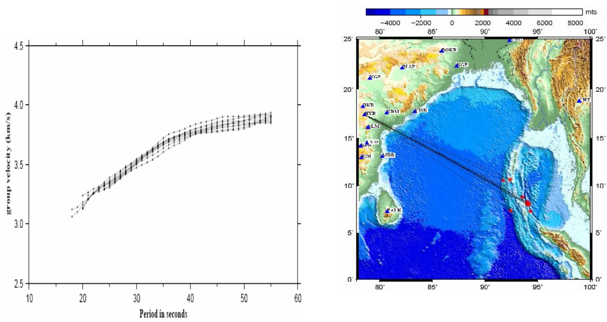

Fig-3: Dispersion curves of all events in cluster 2 and the right hand figure shows the location and the travel path of all the events in cluster 2.

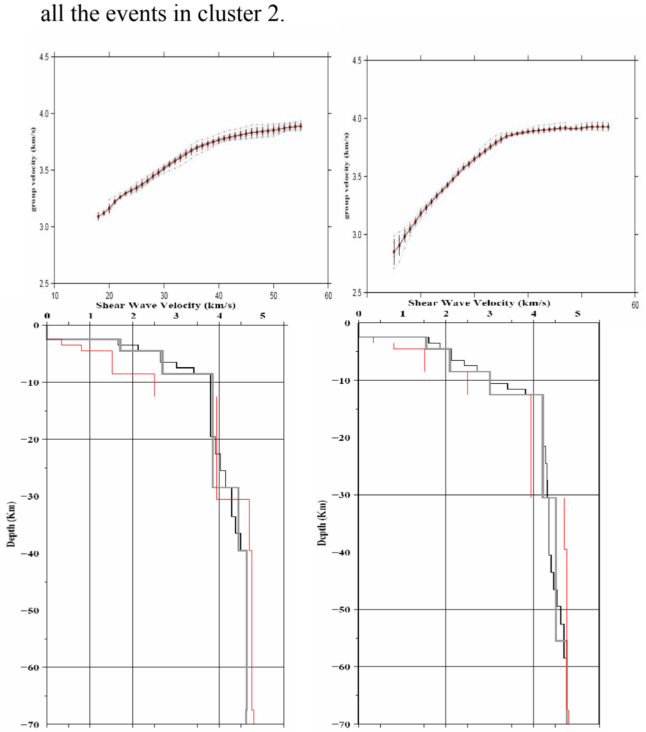

Fig-5: Final velocity models for cluster 1(left) & 2 (right) line is the minimum layer and minimum interface velocity models respectively. The thin black line is the model suggested by Brune and Singh (1986) for the area which we considered as the initial model for the inversion.

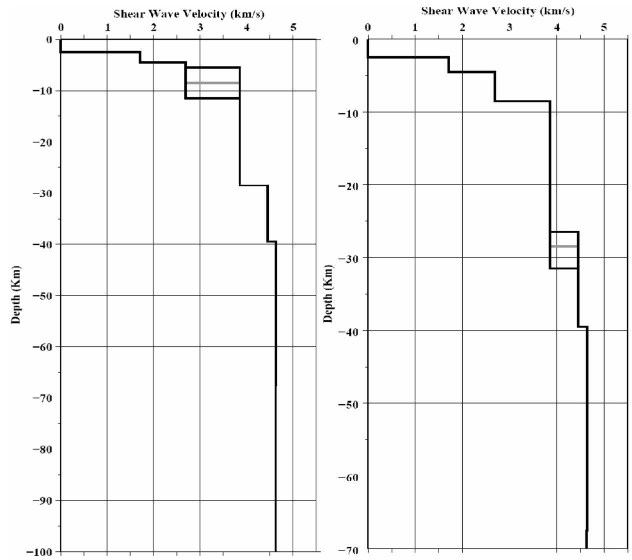

Fig-6: Boundary test(meta-sedi layer & Moho)- A + 3 km thickness perturbation to the base of the assumed metamorphosed sediments layer in cluster we showed a misfit in the period range 22-28 sec and a -3 perturbation lead to the misfit in the period range 21-29 sec. Cluster 1 data doesn't show any clear crust-mantle boundary. Even a ± 10 km thickness perturbation the boundary located at a depth of 37 km doesn't cause any change in the fit of the data. So the boundary check in this particular depth show that dispersion curve obtained for cluster 1 is not sensitive to a depth beyond 30 km

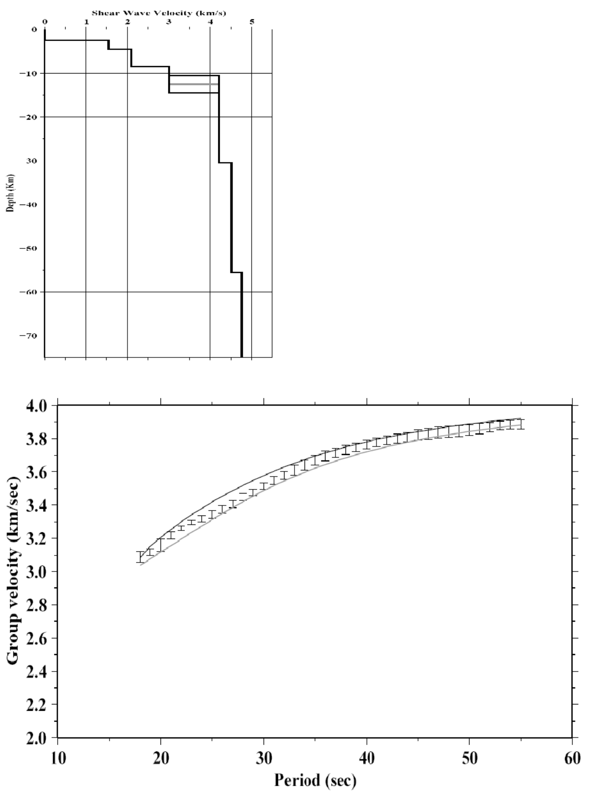

Fig-7- Boundary test of cluster 2 for meta-sedimentary layer only.

Conclusions

In this work we obtained group velocity dispersion curves for fundamental mode Rayleigh waves. These dispersion curves are important because they contain information about deep structure of the Earth.

Cluster 1, which is the closest cluster to the source, showed extremely low velocities which may be due to the thick sediment blanket of the Bengal fan. The inversion result from all the two clusters show an average unconsolidated sediment thickness of 2 km and consolidated sediment thickness of 4 km. A layer of Meta sediments with an average thickness of about 4 km with velocities close to the continental upper crust is found beneath the consolidated sediments in Fig-10(left). This layer which seems to be the crust is actually a layer of sediments. Brune et al, (1992) suggested that these deeply buried sediments beyond 6-7 km in the Bay of Bengal are metamorphosed to attain velocities of over 6 km/s, leading to their mis-identification as continental crustal rocks rather than sediments. This is followed by a more continent-like thick crust of an average thickness of 22±2 km (Fig-10right). Several causal factors for the anomalous crustal thickness have been proposed, among which the sediment loading has been a plausible explanation, leading to a dynamic adjustment at or near the Moho boundary. Also the sediment blanket temperature perturbations may have led to rejuvenation, partial melting and chemical differentiation leading to increased crustal thickness (Brune and Singh, 1986). As the dispersion calculation is giving an average crustal structure of the source receiver travel path, the large crustal thickness shown by cluster 2 can be due to the fact that it sampled the Andaman and Nicobar island regions too in addition to this large sedimentary blanket.

Cluster 2 shows much faster dispersion curve than that obtained for cluster 1. It gives a crustal thickness of 16±3km and a metamorphosed sedimentary layer thickness of 6±1 km (Fig-11upper). the crust-mantle boundary is at a depth of 32±3 km. the inversion results point to a low velocity zone in the upper mantle at a depth of about 50-65 km of V_s 4.3 km s^{-1}. This is a feature usually seen under the continental crusts. Both the minimum layer and minimum interface model fitted the dispersion data equally well.

The results from clusters 1 & 2 are consistent in the average Moho depth of 30±3 km and revealed a continent like thick crust beneath the Bay of Bengal. The results are consistent with the previous works done in the area. The reason for this anomalous crustal thickness is assumed to be due to the high sediment accumulation in the Bengal fan which is leading to a dynamic adjustment at or near the Moho boundary. Also the sediment blanket temperature perturbations may have led to rejuvenation, partial melting and chemical differentiation leading to increased crustal thickness (Brune and Singh, 1986).

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