Estimation of shear velocity contrast for dipping or anisotropic medium from transmitted Ps amplitude variation with ray-parameter

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SUMMARY

Amplitude versus offset analysis of P to P reflection is often used in exploration seismology for hydrocarbon exploration. In the present work, the feasibility to estimate crustal velocity structure from transmitted P to S wave amplitude variation with ray-parameter has been investigated separately for dipping layer and anisotropic medium. First, for horizontal and isotropic medium, the approximation of P-to-s conversion is used that is expressed as a linear form in terms of slowness. Next, the intercept of the linear regression has been used to estimate the shear wave velocity contrast ($\delta$) across an interface. The formulation holds good for isotropic and horizontal layer medium. Application of such formula to anisotropic medium or dipping layer data may lead to erroneous estimation of $\delta$. In order to overcome this problem, a method has been proposed to compensate the SV-amplitude using shifted version of SH-amplitude, and subsequently transforming SV amplitudes equivalent to that from isotropic or horizontal layer medium as the case may be. Once this transformation has been done, $\delta$ can be estimated using isotropic horizontal layer formula. The shifts required in SH for the compensation are $\pi/2$ and $\pi/4$ for dipping layer and anisotropic medium, respectively. The effectiveness of the approach has been reported using various synthetic data sets. The methodology is also tested on real data from HI-CLIMB network in Himalaya, where the presence of dipping Moho has already been reported. The result reveals that the average shear wave velocity contrast across the Moho is larger towards the Indian side compared to the higher Himalayan and Tibetan regions.

Key words: Body waves; Seismic anisotropy; Computational seismology.

1 INTRODUCTION

The P-to-s converted wave technique (usually termed as Receiver Function analysis) in passive seismology is routinely used to map the crust- and upper-mantle structures. In order to interpret the receiver function time series in a meaningful geological interpretation, we need to model it to retrieve the subsurface velocity information. The modelling is done either by forward or by an inversion (Ammon et al. 1990; Ligoria & Ammon 1999, etc.) approach to arrive at a structural model. In such modelling techniques, the information of the primary conversion and the free-surface multiple(s) are important to accurately constrain the model parameters. One such widely used technique is $h$–$k$ stacking technique (Zu & Kanamori 2000) to estimate the crustal parameters. A grid search approach has also been proposed by Julia (2007) to estimate the velocity and density contrast across an interface using the amplitudes of converted and reflected phases of receiver function data. Recently, Kumar et al. (2014) estimated the shear wave velocity contrast across an interface based on the transmitted amplitude versus ray-parameter in teleseismic records, similar to AVO analysis done in exploration seismology (e.g. Castagna & Backus 1993), to estimate elastic parameters. The method utilizes Zoeppritz equation to express the transmission coefficients as a function of slowness and independent elastic parameters. They presented approximations for the P-to-s transmission coefficient as a function of slowness. Subsequently, a linear relationship between the slowness-weighted SV-amplitude versus square of the slowness has been established. The linear regression yields the intercept $X$, which is next used to estimate the shear wave velocity contrast across an interface. The method works well for the horizontal layer and isotropic model and may serve as the first guess for velocity information before proceeding to a cumbersome forward or inversion approach. However, the limitation of Kumar et al.’s (2014) method is that it does not account for dip or anisotropy, therefore application of such formulation to the data, either from dipping layer or anisotropic medium may lead to erroneous estimation of shear wave velocity contrast ($\delta$). Therefore, it is important to isolate the effect of dip or anisotropy from the data before using the isotropic-horizontal layer formulation of Kumar et al. (2014). Hence, this work is motivated to devise a simple method to correct for either dip or anisotropy using shift-and-remove approach as discussed in the subsequent sections. The present method is tested on synthetic data and then applied to real field data from central Himalayan region.
2 Method

Exact reflection and transmission coefficients for an incident plane wave as a function of angle at an interface of two layers in welded contact are given by the Zoeppritz equations. We focus here on the analytic form for the converted wave transmission coefficient from P to S (P-to-s) at a planar boundary of two elastic media (Fig. 1). Fig. 1 demonstrates the possible combination of reflection and transmission ray geometries. The P-to-s transmission coefficient (e.g. Aki & Richards 2002), is approximated by Kumar et al. (2014) and then a relation has been presented under the assumption that \(i_1, i_2, j_1\) and \(j_2\) are small as,

\[
\frac{A_{ps}}{p} = X + Y p^2.
\]

Here, \(A_{ps}\) is the normalized P-to-s transmitted amplitude and \(p\) is the horizontal slowness. For P-to-s transmission energy which is a straight line for \((A_{ps}/p)\) versus \(p^2\) with intercept, \(X\) and slope \(Y\). They further derived intercept, \(X\) for P-to-s transmission energy as

\[
X = 2 (\beta_2 - \beta_1) \delta \beta,
\]

where \(\delta \beta\) is the shear wave velocity contrast across an interface. eq. (2) represents that the intercept is double to that of the shear wave velocity contrast across an interface.

Now, let us first apply Kumar et al.’s (2014) approach to a simple horizontal and isotropic model (shown in Fig. 2) before dealing with the dip or anisotropy case. We first generate synthetic P-receiver functions at various slowness values (Frederiksen & Bostock 2000; Fig. 2a) for a simple velocity model (shown in Fig. 2b). The model contains a crust at a depth of 40 km and the shear velocity contrast at crust to subcrust is \(\delta \beta = 0.5\) km s\(^{-1}\) (see Fig. 2b). The positive peaks at \(\sim 3-4\) s in the receiver function traces (Fig. 2a) are P-to-s conversions from the interface at a depth of 40 km. The amplitudes of these phases are picked automatically within the time window of 1–6 s. Once the amplitudes are picked, the slowness weighted amplitude \((A_{ps}/p)\) versus \(p^2\) (Fig. 2c) yields the intercept \(X\) by fitting a regression line through the data. Here, the estimated value of \(X\) is 1.04, which is almost double that of the \(\delta \beta\) (\(= X/2 = 0.52\) km s\(^{-1}\)) assigned to the model. Hence, the estimated shear wave contrast agrees well with that of the actual value. However, in case of either dip or anisotropy the SV amplitudes are altered and the eqs (1) and (2) remain no longer valid. In such case, the transverse component of receiver functions is also important to analyse in conjunction with the radial component. Therefore, in order to use Kumar et al.’s (2014) horizontal layer isotropic medium relations to retrieve the shear wave velocity contrast, a compensation to the SV amplitudes is necessary such that the resulting SV amplitudes are equivalent to that from a ‘horizontal layer’ or ‘isotropic medium’, as the case may be.

Recently, Schulte-Pulkum & Mahan (2014), demonstrated the isolation of azimuthally varying signal from radial component receiver functions by subtracting the average over backazimuth to retrieve dip and anisotropy. The radial signal matches the tangential components with an offset of \(\pi/2\) and \(\pi/4\) in backazimuth for dipping and azimuthal anisotropic cases, respectively. In case of hexagonal anisotropy with horizontal axis of symmetry a significant amount of energy of P-to-SH polarized shear waves are present in the transverse component of receiver functions. In such case, the medium exhibits periodic function of the azimuth of the ray path with period of \(\pi\) in azimuthal harmonic. Further, P-to-SH conversion results into two-lobed or four-lobed radiation pattern with backazimuth depending upon the symmetry axis and anisotropy (Savage 1998; Levin & Park 1998; Schulte-Pelkum & Blackman 2003). The P–SH scatters vanish when the incoming P-wave coincides with the strike of the horizontal symmetry axis.
Thus, the $\pi$ periodic signature induced by anisotropy can easily be discriminated from that of an isotropic stratified medium containing dipping layers ($2\pi$ periodic with respect to the backazimuth) (Girardin & Farra 1998; Vinnik et al. 2007; Piana Agostinetti et al. 2008). Based on this periodic nature arising due to dip, Girardin & Farra (1998), demonstrated an azimuthal filtering, which can readily be used to make the seismic signal containing purely anisotropy (Vinnik & Montagner 1996; Ohrels et al. 2010, etc.). Whereas, Bianchi et al. (2010) used harmonic decomposition of the receiver function data to retrieve isotropic structure. The radial and transverse components of receiver functions have backazimuth harmonics of $\pi/2k$, where $k$ is the harmonic degree and order (Shiomi & Park 2008). Thus, in presence of anisotropy, the amplitudes in SV and SH exhibit azimuthal dependence of $\pi/2$ periodicity and the phase lag between them is $\pi/4$. Therefore, in order to achieve in-phase SV and SH, the SH must be azimuthally shifted by $\pi/4$. After shifting the SH amplitudes, it is removed from SV to compensate for anisotropic effect. The arguments are mathematically explained below:

In case of homogenous anisotropy with hexagonal symmetry and a horizontal axis of symmetry, there are two principal directions of polarization, one parallel to and the other perpendicular to it. Let us suppose in a spherical isotropic earth, the radially polarized shear wave is $s(t)$, then the projections of the ground motion on the two principal directions are

$$s_1(t) = s(t) \cos \phi$$

and

$$s_2(t) = s(t - \delta t) \sin \phi,$$

where $\phi$ is the fast $S$-wave polarization with respect to the radial direction and $\delta t$ is the delay. The time domain relations for radial and transverse components can be derived projecting the above signals, one gets (e.g. Babuska & Cara 1991),

$$u_r(t) = s(t) \cos^2 \varphi + s(t - \delta t) \sin^2 \varphi$$

and

$$u_\phi(t) = 0.5 \left[ s(t) - s(t - \delta t) \right] \sin (2\varphi).$$

Now, shifting the transverse component by $\pi/4$, eq. (4) becomes

$$u_\phi^{\pi/4}(t) = 0.5 \left[ s(t) - s(t - \delta t) \right] \sin (2\varphi + \pi/2)$$

and

$$u_r(t) - u_\phi^{\pi/4}(t) = 0.5 \left[ s(t) + s(t - \delta t) \right],$$

where $\sin(2\varphi + \pi/2) = \cos(2\varphi) = \cos^2 \phi - \sin^2 \phi$. Assuming weak anisotropy, where $\delta t \to 0$, the above relation becomes

$$u_r(t) - u_\phi^{\pi/4}(t) \approx s(t)$$

which is the radial component in isotropic medium.

Hence eq. (6) transforms the anisotropic radial component into an equivalent isotropic form in the presence of anisotropy. Once this has been done, one can safely use eqs (1) and (2) of Kumar et al. (2014) to estimate the shear wave velocity contrast.

Similar argument can also be extended for dipping layer case. However, in case of dipping layer medium, SV shows polarity reversal with $\pi$ periodicity. P–SV will be a strong conversion in the up-dip and smallest conversion coefficients in the down dip side. Also, the radiation pattern of P-to-s conversions for dipping layer are two-sided lob (Levin & Park 1997) and this variation of amplitudes are associated with the variation of P-wave incidence angle at different dipping interface. At each angle of incidence of $P$ wave from teleseismic distance, from any azimuthal direction making an angle at the interface spiting into P–SV and P–SH waves with a phase lag of $\pi/2$. Therefore, in case of dip, the shift of SH-amplitude is $\pi/2$ in azimuth. The explanation has been numerically verified in the subsequent sections dealing with dipping and anisotropic case.

3 SYNTHETICS

In order to test the validity of the transformations as described above, we first apply to the synthetic data sets generated for various models as below.

3.1 Dipping layer case

Fig. 3 demonstrates the application of the algorithm to a dipping layer case. A dip ($\theta$) of 15° with strike ($\alpha$) of 0° are introduced to a model used in Fig. 2(b) and then synthetic $P$ receiver functions (Frederiksen & Bostock 2000; Fig. 3(a) for a source distribution as shown in Fig. 3(b). The sources are monotonically increasing with slowness and backazimuth. It is clearly seen that the SV- and SH-components of the seismograms (Figs 3(c) and (d) show polarity reversal by $\pi$. The first positive phase at SV components at $\sim 4$ s is the conversion (P-to-s) from 40 km deep Moho. Corresponding SH-phase is also seen at the similar timings (Fig. 3(d)). Now, a time window of 2–6 s has been set to pick the maximum amplitudes automatically. The picked amplitudes are displayed in Figs 3(e) and (f) with respect to backazimuth and slowness, respectively. In both the subplots the solid black and grey circles are SV- and SH-amplitudes, which show sinusoidal type of variation. As described in the theory, an azimuthal shift of $\pi/2$ is applied to the SH-component (shifted version: open diamond) and then removed from the SV-amplitudes (solid black circles). The operation is done within the azimuthal range of 90–360°. The results are displayed in Figs 3(e) and (f) (indicated as crosses). The transformed SV-amplitudes (crosses) should be dip compensated, that is equivalent to the horizontal layer SV data. In order to verify this, the above processes have been repeated using the same velocity model (as Fig. 3(b)) but now without any dip in the layer (as discussed by Kumar et al. 2014). As discussed above, the amplitudes are picked and displayed in Figs 3(e) and (f) (open squares). The plot shows that there is an excellent match between the ‘dip compensated’ and the ‘horizontal layer’ data. Now, the dip compensated data (i.e. crosses) are used to compute the slowness weighted amplitude (eq. 1) to estimate the $b\delta$ (Fig. 3(g)). In Fig. 3(g), we show the estimations for all three types of data, (i) the dip compensated data (crosses), (ii) data from horizontal layer (open squares) and (iii) data from the dipping layer (solid black circles). The intercepts ($\delta$s) are estimated for all the above data by fitting linear regression lines. The respective intercepts are converted to $b\delta$ for each case. It is clear that the dip compensated and horizontal layer data provide the consistent solutions of $b\delta$, that is 0.49 and 0.52 km s$^{-1}$, respectively corresponding to the actual value of 0.5 km s$^{-1}$. Whereas, the data from dipping layer provide overestimated value (0.86 km s$^{-1}$). Hence, numerically it is verified that the dip correction works well and yields a realistic estimate of the shear wave velocity contrast across an interface.

In the above example, the dispositions of sources have full back-azimuthal coverage, which is not always a likely scenario as far as the observed data are concerned. The effect of dip on receiver functions can only be validated with confidence, if we have good azimuthal coverage. In order to show that the transformation yields reasonable result even with azimuthal gap in data, we take another
Figure 3. Application of the present approach (as described in the text) to a synthetic data from dipping layer. (a) Shows a simple isotropic model of 40-km-thick crust with shear wave velocity contrast across crust to subcrust is 0.5 km s\(^{-1}\). The bottom part of the crust is dipping at an angle of 15° with strike 0° N. In order to generate synthetics, we select the events as shown in (b) from all backazimuths with monotonically increasing epicentral distances. The SV- and SH-components of synthetic receiver functions are shown in (c) and (d), respectively. Due to the dip an appreciable amount of energy is also seen at SH-component of seismograms at similar time with 180° periodicity of polarity. Subplots (e) and (f) depict the amplitudes of SV and SH with respect to backazimuth and slowness, respectively. The meanings of different symbols are explained in the index. In (d) \(A_{pS}/p\) versus \(p^2\) plot is shown for different data, that is for dipping layer (\(X_d\), solid circle), dip corrected (\(X_{dc}\), cross) and for the model with no dip (\(X_h\), open square). Three regression lines are fitted to respective data in (g) to estimate the intercepts (\(X_s\)). Once we know \(X_s\), the shear wave velocity contrast can be estimated as \(\delta \beta = X/2\) (according to eq. 2). It is clear that the dip corrected value is in good agreement with the actual value.
Figure 4. This is same as Fig. 3, but for different source distribution as shown in (b). Here, a large azimuthal gap has been introduced so as to mimic approximately real field scenario. The backazimuths lie almost in the first and fourth quadrants and epicentral distance are irregularly placed. The triangles in (b) are the linearly resampled data in azimuth with 5° spacing. The interpolation has been limited to the cluster and avoids in case of azimuthal gap greater that 20° (in the present study) to avoid unbiased prediction. The estimated $\delta \beta$ values are found to be consistent and it is seen that the dip correction works well and yields close to that of the actual one.

Example as depicted in Fig. 4. In this case, the events are randomly distributed in distance and the azimuths are restricted within the first and fourth quadrants (Fig. 4b). Since the data are placed unevenly, therefore we re-sample it to make evenly distributed. For regularization, we use a linear interpolation, where the interpolant is piecewise linear for each interval of the form

$$y = y_0 + (y_1 - y_0) \frac{x - x_0}{x_1 - x_0}$$

(7)
Figure 5. Application of the present approach to a synthetic data in case of hexagonal anisotropic crust with horizontal axis of symmetry. (a) Shows a simple anisotropic model of 40-km-thick crust and shear wave velocity contrast of 0.5 km s\(^{-1}\), similar to that used in earlier examples. The peak-to-peak S-wave anisotropy in crust is 8 per cent with fast direction along 30°N. The synthetic SV- and SH-components of receiver functions have been generated for the source distribution shown in (b). The primary conversions at around 5 s are observed and are quite different in both the components. Similar to the earlier examples the amplitude are picked and plotted in (c) and (d) with respect to backazimuth and slowness, respectively. \(A_{ps}/p\) versus \(p^2\) plot in (d) show three different types of data, that is for anisotropic layer (\(X_a\), solid circle), anisotropic corrected (\(X_{ac}\), cross) and for the model with isotropy (\(X_i\), open square). The fitting regression lines provide intercepts (\(Xs\)). Once we know \(Xs\), the shear wave velocity contrast can be estimated as \(\delta \beta = X/2\) (according to eq. 2). Here we observe that the anisotropic corrected value is in good agreement with the actual value 0.5 km s\(^{-1}\).
Where \( y \) is unknown at \( x \) and holds in the interval \((x_0, x_1)\). Linear interpolation is a robust estimator, however, in case of large gaps in azimuth, this may produce a biased prediction. To avoid such problem, we restrict the interpolation only within the data domain, that is in the present case the interpolation is avoided from \( \sim 100^\circ \) to \( \sim 250^\circ \) (Fig. 4b) azimuthal range. The resampled points at every 5° are denoted by triangles. The subplots in Fig. 4 explain all the steps sequentially, similar to Fig. 3. Fig. 4(g) depicts the final estimation plot of \( \delta \beta \), where it is clear that the dip corrected and flat layer estimations are quite close and they are 0.52 and 0.53 km s\(^{-1}\), respectively. On the other hand, if we use the dipping data directly to the eqs (1) and (2), they yield \( \delta \beta = 0.35 \) km s\(^{-1}\) in contrast to the actual value of 0.5 km s\(^{-1}\).
3.2 Anisotropic medium

After the successful application of the present methodology to dipping layer cases, we now, test the transformation to anisotropic synthetic data sets. In order to do so, we take a model same as in Fig. 2 and introduce crustal S-wave hexagonal anisotropy of 8% with horizontal symmetry axis having fast polarization direction along 30°N (Fig. 5a). The depth of the anisotropic layer is 40 km, followed by an isotropic half space. The SV- and SH-components of the receiver functions are generated (Figs 5c and d) for the specified source distribution (Fig. 5b). The P–SV conversions at ~4–5 s are from the discontinuity at 40 km depth. The P–SH phases are also seen in the similar timings (Fig. 5d). The maximum amplitudes for these phases are picked automatically and displayed in Figs 5(e) and (f) with respect to backazimuth and slowness, respectively. It is clearly seen that whatever be the value of the fast direction, the SV and SH- conversions will always be in a phase lag of π/4. Now, as described in the methodology section, SH-component is azimuthally shifted by π/4 (open diamonds) and then subtracted from SV-amplitudes. The results are shown in Figs 5(e) and (f) (crosses). Similarly, the process has been repeated for isotropic case and is superimposed on Figs 5(e) and (f) (as open squares). It is evident that the anisotropic compensated data and isotropic data agree well. Now, in order to estimate the contrast, all the three types of data are plotted as (Ap/p) versus p2 (Fig. 5g), for example (i) the anisotropy compensated data (crosses), (ii) data from isotropic medium (open squares) and (iii) the data from the anisotropy medium (solid black circles). The δβ estimations for anisotropic compensated and isotropic data are 0.50 and 0.52 km s⁻¹, respectively corresponding to the actual value of 0.5 km s⁻¹. On the other hand, the data without anisotropic correction yields δβ = 0.45 km s⁻¹, thus numerically verifying the efficacy of the anisotropy transformation.

We also test the present algorithm on anisotropic medium when the symmetry axis is tilted. Fig. 6 shows the demonstration, where we use a model same as Fig. 5 but anisotropy has a tilted axis of symmetry having plunge of 20°. The procedure has been repeated as discussed above with the shift of π/4 applied to the transverse component and finally we estimate δβ. The estimated value of δβ for anisotropic corrected one is 0.53 km s⁻¹, which is quite consistent against the actual value of 0.5 km s⁻¹.

4 OBSERVED DATA

It is clear from the synthetic sections that there is a substantial variation in amplitude with respect to the backazimuth in case of dipping layer or anisotropy medium. The discrimination of such variations due to dip and/or anisotropy present in the observed data in particular, is not a straightforward job. The factors like data quality and azimuthal coverage puts limitation to the analysis of dip or anisotropy modeling. Further, the two-lobed P–SH radiation pattern due to the conversion from P-to-s, in case of isotropic dipping layer (Levin & Park 1997; Savage 1998) is again a challenge to discriminate the anisotropy and deviation from the layered media. Thus, the simultaneous presence of both dip and anisotropy make the observed seismograms more complex. Nevertheless, in order to show the robustness of the present approach, it has been tested on real data from Himalaya array of HI-CLIMB network (Nábělek et al. 2009; Fig. 7). The experiment was in operation from 2002 to 2005 with average station spacing of ~5 km (Nábělek et al. 2009). The array covers the foot hill of Himalaya in Nepal, high Himalaya and Tibet. Using the present data set, a number of seismological works have been attempted in the recent past, however, the shear wave velocity contrast across the Moho has not been estimated yet. Nábělek et al. (2009) analysed the converted wave technique and studied the lithospheric structures, tracing the Indian Moho till 31° N latitude in Tibet. The crust and lithosphere–asthenosphere boundary along this array provides a 3-D subducting configuration of the Indian plate (Xu et al. 2011) using S-to-p conversions. Their results are quite consistent with that imaged by Kumar et al. (2006) and Zhao et al. (2010) in this region. The crustal and uppermost mantle S-velocity structure has also been attempted by Xu et al. (2012) using receiver function and surface wave studies. Further, the mantle shear wave birefringence analysis reveals very low or null spitting measurements in most of the stations in the southern part of Tibet especially along this profile (Fu et al. 2008; Chen et al. 2010). Here, we attempt to bring out the nature of the Moho discontinuities by estimating the shear wave velocity contrasts along two profiles, that is AB and CD (Fig. 7). Since the presence of anisotropy in the region is reported to be almost null, it simplifies our analysis, therefore, we apply only dip correction to the data. In order to estimate δβ across Moho, the first job is to compute P-receiver functions (Burridge & Langston 1977; Langston 1977; Vinnik 1977; Kumar & Kawakatsu 2011; Kumar et al. 2011) of all the available stations which have more than 100° of backazimuthal coverage. The source equalization is achieved using the time domain deconvolution technique (Berkhout 1977). Once the receiver functions are computed, the data are distributed uniformly in backazimuth by resampling the original data. In case of large azimuthal gap in data (more that 20° in the present case), we restrict the interpolation to avoid any biased estimation of shear velocity contrast...
Figure 8. Application of the present methodology to observed data from an individual station (H1040). (a) and (b) are the SV- and SH-components of receiver function images arranged with increasing order of slowness. The source distribution is shown in (c). In (a), \( \sum \) denotes the summation trace after moveout correction. The maximum amplitudes are picked by setting time window in (a) and (b) (thick black lines). Once the data are picked, they are linearly re-sampling and plotted in subplots (d) and (e). The meanings of the symbols are demonstrated in index. The slowness weighted amplitudes for dip corrected (black solid circle) and without dip corrected (grey solid circle) and shown in (f). As described in methodology, the intercepts \( X_n \) (=without dip correction) and \( X_d \) (=with dip correction) are estimated. In order to show the consistency of our estimate we invert the stack trace (black wiggle in g) for the model shown in (h). Grey wiggle in (g) is the synthetic receiver function trace.

predictions. The resampled data are further used to estimate \( \delta \beta \) as discussed in the previous sections.

4.1 Individual station examples

Figs 8 and 9 demonstrate the application of the present methodology to two individual stations, for example H1040 and H1270. The shear wave birefringence analysis reveals very low or null splitting measurements in this part of Tibet (Fu et al. 2008; Chen et al. 2010), hence the data are subjected to only dip compensation. Figs 8(a) and 9(a) show the P-receiver function images for the two stations, which clearly show P-to-s conversions from Moho at \( \sim 7-10 \text{ s} \) (Figs 8a and 9a). The same can also be seen in the stack traces shown for individual stations plotted in the same panel. Here, the source distributions are quite good with azimuthal coverage of 0° to \( \sim 180° \) (Figs 8c and 9c). In Fig. 8, we once again demonstrate the steps in detail similar to the description given in the synthetic section. Now, a time window (black box) is set to pick the SV- and SH-amplitudes. These are displayed in Figs 8(d) and (e) with increasing backazimuth and slowness, respectively. The backazimuth versus amplitude data are resampled using linear interpolation at every 5° azimuth for both SV and SH data. The interpolation has been avoided in case where the gap in backazimuth exceeds 20°. As described in the theory, the SH-components are shifted by \( \pi/2 \) and removed from SV. The resulting amplitudes are used to generate \( (A_{ps}/\rho) \) versus \( p^2 \) (black solid circles in Figs 8f and 9d). We show two sets of data in Figs 8(f) and (d), that is one ‘without dip compensated’ (grey solid circle) and other ‘with dip compensated’ (black solid circle). Fitting linear regression lines to both sets of data (i.e. dip compensated, black circle and no-dip compensated, grey circles) yield the intercepts...
Figure 9. This is same as Fig. 8, but for another station H1270. Here we only show the data in subplots (a) and (b). The source distributions are displayed in (c). The slowness weighted amplitudes for dip corrected (black solid circles) and without dip corrected (grey solid circles) and plotted in (d). Further, the inversion results are also shown in (e) and (f). It is clear that the $\delta\beta$ estimated using the present methodology after dip correction is quite consistent with the inversion value.

These intercepts are translated to $\delta\beta$ of 1.27 and 2.16 km s$^{-1}$ corresponding to ‘with’ and ‘without’ dip correction, respectively. It is obvious that the latter value, that is without dip compensated is unrealistically large.

Further, in order to compare our dip compensated result with that from the traditional modelling technique, we invert the receiver function stack. First, a stack receiver function trace for each station is generated from the Figs 8(a) and 9(a) after applying a moveout correction with a reference slowness of 6.4 s deg$^{-1}$ (Yuan et al. 1997) using IASP91 velocity model. The observed traces are shown in Figs 8(g) and 9(e) as black wiggles. Secondly, the stack traces are inverted for a optimum model as shown in Figs 8(h) and 9(f). The synthetic traces are also shown in Figs 8(g) and 9(e) as grey wiggles. In case of H1040, the inversion yields $\delta\beta = 1.27$ km s$^{-1}$, which matches well with that generated by the present approach ($\delta\beta = 1.27$ km s$^{-1}$). While for station H1270, the present method gives $\delta\beta = 0.8$ km s$^{-1}$, while inversion yields 0.802 km s$^{-1}$. Thus, we find that the estimation of shear wave velocity contrasts using the present approach for both the stations are well consistent. It should be noted that the derived contrast in fact, defines the shear wave velocity jump across the discontinuity, ignoring its transition (i.e. sharpness). For example, the contrasts below station H1040 is more than that below H1270, implies that the velocity gradient at the lower crust is more for H1270 and this large transition can also be seen in the traditional inversion modelling for the individual station (Figs 8h and 9f).

4.2 Shear wave velocity across Moho below Himalaya

In order to map the shear wave velocity contrast across Moho along a profile below Himalaya, we analyse seismic waveforms from all the available stations in a similar manner as described in the previous sections. Firstly, SV- and SH-components of receiver functions are generated and visually identify the base of the crust (similar to cited examples in Section 4.1) in the depths range presented by Nábělek et al. (2009). Once the Moho conversions are identified, the maximum amplitudes are picked automatically by setting a time window on either side of the identified Moho. The same time window is also set in the SH-components to pick the corresponding SH-amplitudes. We skipped some of the stations, where we do not have clear Moho phase. Once the amplitudes of SV and SH are picked, the backazimuth versus amplitude data are resampled using linear interpolation at every 5$^\circ$ azimuth. The interpolation has been avoided in case where the gap in backazimuth exceeds 20$^\circ$.

Next, the SH-amplitudes are shifted by $\pi/2$ along backazimuth and then removed from SV-amplitudes as described in the previous sections to estimate $\delta\beta$s for each station. The estimated $\delta\beta$s are plotted (solid black circle) corresponding to each station along the profiles AB and CD, shown as in Fig. 10(a) (profiles are also shown in Fig. 7). In order to compare the present estimated values with that derived from without dip correction, we estimate the $\delta\beta$ directly using eqs (1) and (2) (i.e. isotropic, horizontal layer relations). The results (Figs 10b and c) show that the $\delta\beta$ estimations without taking care of dip are unrealistically overestimated for most of the stations.
Figure 10. Shear wave velocity contrast ($\delta\beta$) across Moho along the HI-CLIMB profiles. The location of the seismic station and profiles are shown in (a). At the top panel of each subplot (b) and (c) the elevation and stations (inverted open triangle) are shown. In the bottom panel of each subplots (c) and (d), $\delta\beta$ values are plotted along the profiles. The grey and black solid circles represent the $\delta\beta$ without dip and with dip correction, respectively. The shaded region across the dark black lines is the standard error of $\pm 2SE$ using bootstrap technique. The dashed continuous lines in each subplots are mean variation of $\delta\beta$ across Moho. It is clearly seen that the average shear wave velocity contrast across Moho gradually decreases from Indian side to High Himalayan and then it flattens to almost constant values below Tibet.

However, the dip corrected $\delta\beta$s along both the profiles reveal a consistent picture of the contrast across the Moho within the errors ($\pm 2SE$) computed from bootstrap (Efron & Tibshirani 1993) analysis with 500 bootstrap families. It is interesting to note that shear wave velocity contrast across Moho along both the profiles are not smoothly varying, instead it has scatters. Such type of variations can also be observed in $V_p/V_s$ along a profile in a number of regions where we have closely spaced stations. This prompts us to suggest that the Mohorovicic discontinuity might be compositionally variable under the illuminating wavelength of our seismic waves. Further, it is clear from Fig. 10(b) that $\delta\beta$ across Moho along the profile AB has more scatters where the elevation is lesser compared to that along the profile CD, where we have higher topography (Fig. 10c). The mean $\delta\beta$ is larger by $\sim 0.2 \text{ km s}^{-1}$ towards the Indian side compared to the higher Himalayan and Tibetan regions. The plausible explanation for such scatters in $\delta\beta$ along profile AB which spans the Himalayan orogenic prism, could be due to inhomogeneities caused by the deformation of Indian crust due to the Indian–Asia collision. The regions near foothill and high Himalaya experienced intense deformation due to the onset of thrust and bending of lithosphere, possibly resulting into the heterogeneous crust. However, the possibility of other potential sources of bias or scatters cannot be ruled out, such as, the unaccounted part of anisotropy and the effect of topographic scattering on the converted phase. Here, we discard the influence of the former due to the lack of consistent effect on all the stations along the profiles. On the other hand, the conversions from $P$ to Rayleigh waves at the free surface may sometime be dissimulated as P-to-s phase. Mode conversions and focusing/defocusing effects due to hills and valleys can also be sources of observed scatters (Bannister et al. 1990; Gupta et al. 1990; Abers 1998; Jones & Phinney 1998; Monteiller et al. 2013).

5 CONCLUSIONS

The conclusions can be divided into two major parts, on methodology and on real field data example.
(i) An approach is demonstrated to compensate for dip or anisotropy to estimate the shear wave velocity contrast using teleseismic P-to-s transmitted wave amplitude variation with slowness. The present approach is essentially an extension of Kumar et al.'s (2014) method, which is based on isotropic and horizontal layer assumptions. The advantage of the method is that it directly estimates the shear wave velocity contrast across an interface that may serve as a first guess for velocity information before proceeding to a cumbersome forward or inversion modelling. The method takes care of dip or anisotropy compensation without their identification and numerical values. The limitations of the present method are: (1) it does not account for both dip and anisotropy simultaneously, (2) the backazimuthal coverage of the sources should be sufficiently large (at least more than 100°), (3) it also works well for low value of plunge of the symmetry axis and (4) importantly, the data should be of high signal to noise ratio with no interference by other phases. (ii) The real field example from HI-CLIMB show the shear wave velocity contrast across Moho along the two profiles encompassing Indian, Himalayan and Tibetan Moho. The results reveal that the δβs across Moho are overestimated if azimuthal variations are not considered, and that the correcting for azimuthal effects due to dip (or plunging axis anisotropy, which has same periodicity) reduces the estimated contrasts as well as the scatters along the profile significantly.

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REFERENCES


