Direct calculation of interval velocities and layer thicknesses from seismic wide-angle reflection times

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SUMMARY
Wide-angle seismic reflection traveltimes generated by large explosions are used extensively to obtain both velocity and regional structure of the crust and upper mantle. The method most commonly used to calculate interval velocities and thicknesses (layer parameters) of a multilayered earth, supposed to consist of homogeneous, isotropic and horizontal layers, is based on Dix's (1955) interval velocity formula, which needs a prior estimate of the rms velocity and traveltimes at zero offset. Use of Dix's hyperbolic approximation to the non-hyperbolic wide-angle reflection times causes large errors in the determination of layer parameters. Currently, no direct method exists to extract the velocity information from such wide-angle seismic data. Here we propose a layer-stripping method, using the ray parameter, to calculate the layer parameters directly in a vertically heterogeneous earth from a set of wide-angle seismic reflection data. Synthetic reflection times contaminated by some realistic errors have been used to demonstrate the efficiency and reliability of the algorithms. Field examples using well-identified wide-angle reflection times illustrate the practical feasibility of the proposed method.

Key words: crustal structure, deep seismic reflection, seismic velocities.

1 INTRODUCTION
The reflection coefficients of seismic waves often increase with offset, and reach a maximum value at the critical angle (Richards 1961; Winterstein & Hanten 1985). The amplitudes of the post-critical (wide angle) reflections are large due to total internal reflection. Such reflections are usually prominent on seismograms, even in a noisy area with a single fold coverage. The recording of wide-angle seismic waves in controlled-source seismology has become an important tool for deciphering the variations in both seismic velocity and crustal thickness in order to understand the tectonic framework of an area. A review of such studies in central Europe, edited by Giese, Stein & Prodehl (1976), contains many excellent examples. For the study of deep crustal structure, extensive use has been made of wide-angle reflections in the USSR, parts of Europe, Canada, USA and India.

In a multilayered earth consisting of horizontal, homogeneous and isotropic layers, the reflection traveltime, \(T(X)\), at an offset, \(X\), is expressed by the Taner & Koehler (1969) series as

\[
T^2(X) = C_0 + C_1 X^2 + C_2 X^4 + \ldots
\]  

(1)

The first two coefficients are respectively equal to the square of the zero-offset two-way time and the inverse square of the rms velocity. The other coefficients (\(C_2, C_4, \ldots\)) are related to layer parameters in a complicated way. The conventional hyperbolic equation becomes inaccurate when the spread-length-to-depth ratio and heterogeneity in the subsurface increase. To improve the accuracy, some authors (Shah & Levin 1973; Al-Chalabi 1974; May & Straley 1979; Hake, Helbig & Madsen 1984; Castle 1994; de Bazelaire & Vialli 1994; Kaila & Sain 1994; Thore, de Bazel & Ray 1994; Tsvankin & Thomsen 1994) have suggested the use of higher-order terms, but the gains do not compensate for the dramatic increase in computing time. In the past, the use of higher-order terms has not received a wide application in the exploration industry. The hyperbolic approximation (neglecting the third- and higher-order terms of the reflection series), which is valid only at small-offset ranges, is still used for seismic data processing work.

Interval velocities and layer thicknesses from large-offset reflection times have been calculated in various ways by different authors. Dix's (1955) interval velocity formula can be used only when the rms velocity and two-way traveltime at zero offset are known for a set of reflectors. As the offset distance increases, the estimated velocity, here called the slant-path rms velocity (Robinson 1983), obtained by hyperbolic fitting, is different from Dix's rms velocity due to the cumulative
effect of refractions of a single ray at each overlying interface in a multilayered earth model. The percentage error in rms velocity and zero-offset two-way traveltime, as determined using the hyperbolic approximation to the increasing offset reflection times without the control of offsets associated with a low-angle region, becomes very large at the wide-angle ranges (Kaila & Sain 1994). Al-Chalabi (1974) pointed out that the use of the hyperbolic approximation to the large-offset reflection times causes large errors in the determination of interval velocities and layer thicknesses unless an appropriate adjustment is made. Velocity analysis from wide-angle reverse reflection times using the effective velocity, defined by Kaila & Krishna (1979), has remained rather difficult.

For the calculation of interval velocities and thicknesses from a series of large-offset reflection traveltimes, Sattlegger (1965) proposed an iterative method that considers exact ray paths in a multilayered earth in order to compute the forward traveltimes, which are compared with the observed traveltimes by a least-squares (Gauss–Newton) method. This is applicable only to the small-offset traveltimes, which can be reasonably approximated by a hyperbolic equation. For wide-angle reflection times, the solution obtained by the Gauss–Newton method does not always converge, particularly when the initial model does not lie very close to the true solution. Limond (1975) made use of Sattlegger’s (1965) method in the computation of a forward response in order to match the wide-angle reflection times in marine sonobuoy work by a least-squares method in which only the velocity is varied while the zero-offset traveltimes are constrained from the normal-incidence reflection profile. Sain & Kaila (1994) show that, even without constraints, layer parameters can be calculated accurately from wide-angle reflection times by an iterative method. As compared with the time-consuming iterative method, the present method is direct and faster.

In order to calculate the interval velocity from large-offset reflection times, many authors (Garmany, Orcutt & Parker 1979; Diebold & Stoffa 1981; Stoffa et al. 1981, 1982; Schultz 1982) used the $\tau$-$p$ transformation, where $p = (dT/dX)$ is the ray parameter and $\tau$ equals $T - pX$, called the intercept time. $T$ is the two-way reflection time at an offset $X$. Our proposed method does not need any transformation of $(X, T)$ data. Direct methods (Gonzalez-Serrano & Claerbout 1984; Nowroozi 1989, 1990) of calculating layer parameters do not, of course, need any transformation, but they require the recognition of an $(X, T)$ pair with the same ray parameter at the top and bottom of the layer concerned. For field data containing many errors at a particular wide-angle recording spread, recognition of the same ray parameter at the top and bottom of a layer can be difficult. For a given ray parameter associated with a particular wide-angle data point on reflections from the bottom of a layer, the corresponding data point for reflection from the top of the layer may lie outside the recording range. This is because wide-angle reflections are very sensitive to ray parameters.

Wide-angle reflections and refractions are now used extensively in crustal seismics to obtain the 2-D velocity structure of the subsurface earth. This can be done either by ray-trace forward modelling (McMechan & Mooney 1980; Cerveny & Psencik 1984; Spence, Whittall & Clowes 1984; Zelt & Ellis 1988) or by inversion (Huang, Spence & Green 1986; Lutter, Nowach & Braile 1990; Zelt & Smith 1992; Zelt & Forsyth 1994; Zelt et al. 1994). For fast computation, a spatially non-varying velocity structure is assumed in 1-D forward traveltime and amplitude modelling (Fuchs & Mueller 1971), Wiechert-Herglotz inversion (Bullen & Bolt 1985) and $\tau$-$p$ inversion (Diebold & Stoffa 1981). Based on this assumption, we have formulated a new method to calculate the interval velocities and layer thicknesses directly from a series of wide-angle reflection times.

2 THEORY OF THE PROPOSED METHOD: VELOCITY ANALYSIS

We follow the conventional procedure of processing reflection data in a layer-by-layer fashion working downwards. This means that when we deal with the reflections from the $n$th interface, all parameters for the overlying $(n-1)$ layers are known.

In a multilayered medium consisting of horizontal, homogeneous and isotropic layers, the offset $X(p)$ for the $n$th interface and corresponding reflection time $T(p)$ are parametrically expressed (Taner & Koehler 1969) as

$$T(p) = 2 \sum_{k=1}^{n} \frac{z_k}{v_k \left[1 - \left(1 - p^2 \frac{v_k^2}{v_X^2}\right)\right]^{1/2}},$$

(2)

$$X(p) = 2p \sum_{k=1}^{n} \frac{z_k v_k}{\left[1 - p^2 \frac{v_k^2}{v_X^2}\right]^{1/2}},$$

(3)

where the ray parameter $p$ is defined by Snell’s law as

$$p = \frac{\sin \theta_k}{v_k},$$

(4)

$\theta_k$ is the angle of incidence at the $k$th layer. $\theta_k$, $z_k$, $v_k$ and $X_k$ are defined in Fig. 1. $X$ is the distance between the source and the receiver corresponding to a ray parameter. The slope of a straight line fitted to a segment of synthetic reflection data $(X, T)$ produces the ray parameter quite accurately for the centre of the segment, a point which is here referred to as the mid-distance of the segment $(X_m)$. The determination of the ray parameter remains accurate even if the synthetic data are contaminated by any percentage of realistic errors. The exact ray parameters at various offsets for the $n$th interface of a given velocity model can be calculated using eq. (4) corresponding to different prescribed take-off angles at the source. For the velocity model tabulated in Fig. 2(a), the true ray parameters from the bottom interface of the model are shown in Figs 2(a) and (b) against the offset-to-depth ratio. Corresponding to various ray parameters, the reflection times at different offsets are calculated using eqs (2) and (3) respectively. Here the true model has six layers. In this case, the various take-off angles lie below a maximum value, which is defined as

$$\theta_{\text{max}} = \sin^{-1} \frac{\frac{v_X}{v_h}}{1},$$

(5)

where $v_1$ and $v_h$ are the velocities of the first and the greatest velocity layers in the model. To simulate field situations, we add to each travelt ime some percentage of itself in order to create realistic noise levels. We now show how accurately we can calculate ray parameters from any increasing offset reflection times. For this, we segment the calculated traveltimes into groups containing, say, 50 offsets, an arbitrary number of data points (as large a data set as possible should be used for noisy data), such that the first segment includes the first to
Figure 1. Reflection ray path in a homogeneous and isotropic medium consisting of horizontal layers, with various parameters defined.

Figure 2. A comparison between true ray parameters and those determined by the straight-line for various segments with increasing offset reflection times is shown for (a) noise-free data and (b) noisy synthetic data calculated from the bottom interface of the model tabulated in (a). A comparison between true interval velocity and thickness of the sixth layer and those calculated by the present method is shown for (c) noise-free data and (d) noisy data. The comparison is shown with increasing offset reflection times.
Calculation of interval velocities

fifty-first data points, the second segment contains the second to fifty-first, and so on until all data points are included. We calculate the ray parameter through a straight-line fit to the traveltimes of each segment and assign it to the mid-distance of the segment. These ray parameters have been plotted in Figs 2(a) and (b), respectively, against the offset/depth ratio for noise-free and 1 per cent noisy synthetic data. We have seen that any type of error can be added, and the straight-line fit gives similar values for the ray parameter. It should be noted that a 1 per cent error for the reflection time for the last interface of the model includes a maximum value lying between ±100 and ±300 ms, which is rather high in comparison to the noise encountered in the field. It is evident from Fig. 2 that this method allows the ray parameters to be calculated quite accurately. Since wide-angle reflections are observed beyond the offset-to-depth ratio of 2, we show the calculation up to an offset-to-depth ratio of 6, within which all wide-angle reflections may be recorded on the seismograms. Of course, the ray parameter can also be calculated from further offset traveltimes.

For the calculation of interval velocity and thickness of the nth layer, we calculate the traveltime, \(T_{\text{md}}\), at the mid-distance, \(X_{\text{md}}\), of each segment by smoothing the available traveltimes of the segment through a local fit using a hyperbolic equation such as:

\[
T_{\text{md}}^2 = C_{\text{or}} + C_{\text{fr}} X_{\text{md}}^2,
\]

where \(C_{\text{or}}\) and \(C_{\text{fr}}\) are the fitted coefficients. For downward processing of reflection times from the nth interface, we know all layer parameters for the top \((n-1)\) layers. Referring to Fig. 1 and using eqs (2) and (3) we can write

\[
2 \frac{z_n}{v_n (1 - p^2 v_n^2)^{1/2}} T_{\text{md}} = 2 \sum_{k=1}^{n-1} \frac{z_k}{v_k (1 - p^2 v_k^2)^{1/2}} = K_1,
\]

\[
2 p \frac{z_n v_n}{(1 - p^2 v_n^2)^{1/2}} X_{\text{md}} = 2 \sum_{k=1}^{n-1} \frac{z_k v_k}{(1 - p^2 v_k^2)^{1/2}} = K_2,
\]

where \(K_1\) and \(K_2\) can be computed from the layer parameters of all overlying layers. Eqs (7) and (8), representing time and distance for the nth layer, differ from eqs (12) and (13) given by Nowroozi (1989). The time and distance in Nowroozi’s (1989) paper are the respective differences in traveltimes and offset distances for the reflected rays from the top and bottom of the layer in which rays start at the source with the same take-off angle or ray parameter. Such recognition of the ray parameter is not needed in this work. The present method works for any ray parameter, using a layer-stripping method.

Now, from eqs (7) and (8) we have

\[
(1 - p^2 v_n^2)^{1/2} = \frac{2 p z_n v_n}{v_n K_1} \frac{K_2}{K_2}.
\]

Using eq. (9), the velocity, \(v_n\), and the thickness, \(z_n\), of the nth layer can be directly calculated as

\[
v_n = \left( \frac{K_2}{p K_1} \right)^{1/2},
\]

\[
z_n = \frac{1}{2} \left[ K_1 K_2 \left( \frac{1 - p^2 v_n^2}{K_1} \right)^{1/2} \right].
\]

Figs 2(c) and (d) show the calculated layer parameters for the sixth layer from various segments of noise-free and noisy synthetic data, respectively. For clarity, layer parameters have been shown for every fifth segment. This demonstrates the accuracy of the present method for calculating interval velocity and layer thickness from large-offset or wide-angle reflection times. Eqs (10) and (11) indicate the velocity-depth (or thickness) ambiguity problem (Bickel 1990; Lines 1993; Ross 1994), which is inherent in velocity estimations for the case of vertically propagating waves \((p = 0)\).

3 MEAN ERRORS

In reality, the seismic velocity in a given layer varies both laterally and vertically, and the interfaces undulate. Therefore, the assumption of the flat-layered earth model produces averaged interval velocities and thicknesses of various layers. Hence, each layer parameter is associated with some mean error which can be calculated for the nth layer as follows.

Because the interval velocities and thicknesses of all \(n\) layers are known, the ray parameters at various offsets of the reflection data from the nth interface can be calculated using eq. (3), which is based on Sattlegger’s (1965) algorithm in a modified form. In this scheme, we first calculate the maximum take-off angle, \(\theta_{\text{max}}\), for the above \(n\)-layered velocity model using eq. (5). A take-off angle equal to half of \(\theta_{\text{max}}\) (or the average between the maximum take-off angle and the zero take-off angle) is provided at the source, and the ray parameter is calculated using eq. (4). Then we calculate the offset for the ray reflected from the nth layer as per eq. (3). If the calculated offset is more (or less) than the required offset, a take-off angle equal to the average of the previous take-off angle and the zero (or maximum) take-off angle is determined and the offset is again calculated using eq. (3). Next, we decide on a value for the take-off angle by taking the average of the previous take-off angles for which offsets lie immediately on either side of the required offset. Thus the take-off angle or ray parameter is searched for iteratively until the reflected ray reaches the required offset within a given tolerance. The process is repeated for all offsets, and the calculated ray parameters are then used in eqs (2) and (3) for the computation of traveltimes and offset contributions from all \((n-1)\) overlying layers. Subtraction of these traveltimes and offsets from the respective traveltimes and offset distances coming out of the nth interface produces the traveltimes and offsets associated only with the nth layer. The subtracted traveltimes, \(\Delta T_n\), and offsets, \(\Delta X_n\), are related to the interval velocity, \(v_n\), and the thickness, \(z_n\), of the nth layer through a hyperbolic equation:

\[
\Delta T_n^2 = \frac{\Delta X_n^2}{v_n^2} + \frac{4 z_n^2}{v_n^2} = a \Delta X_n^2 + \frac{b}{v_n^2},
\]

where

\[
a = \frac{1}{v_n^2},
\]

\[
b = \frac{4 z_n^2}{v_n^2}.
\]

The errors in observed \((X, T)\) data and the assumption of a flat-layered model will produce mean errors both in the interval velocity and the thickness of the nth layer. The mean errors \(\delta(a)\) of \(a\) and \(\delta(b)\) of \(b\) as illustrated by Topping (1963) are
given by

$$\delta(a) = m(\Delta T_i^2) \left( \frac{K}{K} \right)^{1/2} \left[ \sum_{i=1}^{K} \Delta X_{n-i}^2 - \left( \sum_{i=1}^{K} \Delta X_{n-i}^2 \right)^2 \right]^{1/2},$$

$$\delta(b) = m(\Delta T_i^2) \left( \frac{K}{K} \right)^{1/2} \left[ \sum_{i=1}^{K} \Delta X_{n-i}^2 - \left( \sum_{i=1}^{K} \Delta X_{n-i}^2 \right)^2 \right]^{1/2},$$

where $m(\Delta T_i^2)$ is the mean error of squared offset times between the subtracted travel times, $\Delta T_m$, and the travel times of the nth layer only, calculated using the hyperbolic relation between the velocity, $v_n$, and thickness, $z_n$, at various offsets, $\Delta X_{n-i}$. $K$ is the total number of $(\Delta X_{n-i}, \Delta T_m)$ data points from the nth interface. Now, from eqs (13) and (15), and from (14) and (16), the mean error in the velocity, $\delta(v_n)$, and the mean error in the thickness of the nth layer, $\delta(z_n)$, can be calculated as

$$\delta(v_n) = \left( \frac{v_n^2}{2} \right) m(\Delta T_i^2) \left( \frac{K}{K} \right)^{1/2} \left[ \sum_{i=1}^{K} \Delta X_{n-i}^2 - \left( \sum_{i=1}^{K} \Delta X_{n-i}^2 \right)^2 \right]^{1/2}.$$

4 SYNTHETIC AND FIELD EXAMPLES

Before any given method is applied to the field data, it should first be tested on synthetic data where the model parameters are known. A simplified six-layer velocity model of the continental crust is shown in Fig. 3(d) by the solid curve. The true velocities are 2.10, 3.00, 4.00, 5.00, 6.00 and 7.00 km s$^{-1}$ respectively, from the first to the sixth layers. The corresponding true thicknesses are 1.50, 2.00, 2.50, 3.00, 10.00 and 15.00 km respectively. Some wide-angle reflection times are analytically computed using the parametric equation (2) for offsets defined by eq. (3) by providing different take-off angles at the source corresponding to the various interfaces. 1 per cent random errors with Gaussian distributions are then added to these reflection times to simulate a field situation. The synthetic data limited to the wide-angle range are shown in Figs 3(a), (b) and (c) for the first and second, third and fourth and fifth and sixth layers, respectively. The six-layer true velocity model is compared in (d) with the velocity model obtained by the present method.

Figure 3. Synthetic reflection times with added noise, which are computed for various interfaces of the true model shown in (d), are compared with the travel time curves generated for the estimated model in (a), (b) and (c) for the first and second, third and fourth and, fifth and sixth layers, respectively. The six-layer true velocity model is compared in (d) with the velocity model obtained by the present method.
The random errors added to the first to sixth reflected phases are in the ranges $\pm 40$, $\pm 75$, $\pm 100$, $\pm 120$, $\pm 150$ and $\pm 200$ ms, respectively. A reduced time scale with a reduction velocity of 6.0 km s$^{-1}$ has been used throughout the paper to present all traveltimes in a compact form. By applying the proposed method to these synthetic data in a layer-by-layer fashion, we obtain the velocity model shown by the dashed line in Fig. 3(d). The interval velocities associated with mean errors, calculated as per eqs (10) and (17) for the first to sixth layers, are $2.07 \pm 0.02$, $3.03 \pm 0.04$, $4.03 \pm 0.07$, $5.06 \pm 0.11$, $6.05 \pm 0.17$ and $7.06 \pm 0.19$ km s$^{-1}$, respectively. The corresponding thicknesses and associated mean errors, calculated as per eqs (11) and (18), are $1.51 \pm 0.05$, $2.03 \pm 0.18$, $2.53 \pm 0.35$, $3.08 \pm 0.48$, $10.15 \pm 0.55$, and $14.90 \pm 0.62$ km.

The rms residual between the synthetic data and the computational layer of velocity 5.09 km s$^{-1}$ is $0.016$ and $0.032$ s. The traveltimes shown in Figs 3(a), (b) and (c) indicate a very good match between the synthetic data and the computed traveltimes of the calculated model. It is also evident that the calculated velocity model agrees quite well with the true model.

We now use two field data sets to demonstrate the practical feasibility of the method. Since we have assumed a 1-D model, the method produces averaged interval velocity and thickness of various layers. Fig. 4 shows a trace-normalized record section on a reduced time scale corresponding to shot point 235 along the Hirapur–Mandla deep seismic sounding profile (Kaila et al. 1987, 1989). The traveltimes of first arrivals (marked by $P_S$ and $P_p$ phases) observed in the record section are plotted in Fig. 5(a) on a reduced time scale. The analysis of the first-arrival traveltimes using the method of Kaila & Narain (1970) produces a very thin (0.83 $\pm 0.02$ km) sedimentary layer of velocity 5.09 $\pm 0.01$ km s$^{-1}$ overlying the crystalline basement of velocity 5.97 $\pm 0.03$ km s$^{-1}$. The rms residuals between the data being fitted and the fitted lines for the above two refraction line segments are 0.016 and 0.032 s. Two intracrustal reflections, marked by $P^1$ and $P^2$, have been identified by comparing seismograms from the reciprocal shot points along the profile. Some 'floating reflections' appear on the seismogram (Fig. 4), but they are not considered as we did not observe any similar reflected phases on seismograms for reciprocal shot points. Phases have been identified based on the reciprocity test (Zelt & Forsyth 1994) of the picked traveltimes from the reciprocal shot points. The $P^1$ reflected phase is regarded as wide-angle reflections from the bottom of the crystalline basement, as it extends asymptotically to the basement first arrivals. The strong reflected phase marked by $P^M$ represents the wide-angle reflections from the Moho boundary, and this phase has been identified from the reciprocal shot point (SP 80 along the profile). The downward extension of the basement is obtained from the traveltimes of the $P^1$ phase using the present method. This gives an interval velocity of 5.93 $\pm 0.02$ km s$^{-1}$ and thickness of 7.05 $\pm 0.04$ km for the second layer with an rms residual of 0.029 s. The velocity of the second layer is taken as the average (5.95 km s$^{-1}$) of the velocities obtained for the layer from both the first-arrival and the reflection data. The second reflected phase ($P^2$) produces an interval velocity of 6.65 $\pm 0.02$ km s$^{-1}$, corresponding to a thickness of 15.18 $\pm 0.03$ km and an rms residual of 0.026 s. The velocity of 6.98 $\pm 0.03$ km s$^{-1}$ and the thickness of 21.69 $\pm 0.21$ km calculated from the traveltimes of the wide-angle Moho reflections ($P^M$) are the averaged interval velocity and thickness of the lower crust in the Vindhyan basin, India.

The present study reveals the Moho depth of 44.75 km, which is comparable to the reported Moho depth (42–46 km) in this region (Kaila et al. 1987, 1989). The velocity structure and the Moho depth are also comparable to the worldwide average velocity structure and the thickness (41.4 $\pm 6.2$ km) of the crust in the shield regions (Christensen & Mooney 1995). The traveltimes for the $P^M$ phase are 0.05 s. Fig. 5(b) shows the estimated velocity model.

Kaila et al. (1990) have identified two wide-angle reflected phases in a field seismogram (marked by $P^5$ and $P^6$ in Fig. 7 of their work) from layers below the basement, corresponding to shot point 25 along the Sadra–Mehmadabad–Degam part of a long DSS profile in the North Cambay and Sanchor basins, India. The traveltimes of these two phases are shown in Fig. 6(a). Since we process the data in a layer-by-layer fashion, we need velocity information from the surface down to the basement. As the 2-D velocity model at shallow levels...
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\[ V_1 = 5.05 \text{ KM/S} \]

\[ V_2 = 5.95 \text{ KM/S} \]

\[ V_3 = 6.65 \text{ KM/S} \]

\[ V_4 = 6.98 \text{ KM/S} \]

**Figure 5.** (a) Traveltimes of all refraction and wide-angle reflection phases marked by \( P_s \), \( P_p \), \( P^1 \), \( P^2 \) and \( P^M \) on the field seismogram (Fig. 4). (b) The velocity model determined from the above traveltimes of various phases.

( obtained by forward modelling of seismic refraction data) in this region (Kaila et al. 1990) is more or less flat, we have averaged out the velocity structure in the first three layers to a simplified three-layer 1-D velocity model, indicated by velocity values (in km s\(^{-1}\)) at the right-hand side and thicknesses (in km) by dashed lines in Fig. 6(b). The first layer, with a velocity of 2.00 km s\(^{-1}\) and thickness of 1.30 km, the second layer, with a velocity of 3.30 km s\(^{-1}\) and thickness of 3.90 km, and the third layer, with a velocity of 4.80 km s\(^{-1}\) and thickness of 1.50 km, are fixed while calculating the layer parameters of deeper layers. The first wide-angle reflected phase produces an interval velocity of 6.38 ± 0.04 km s\(^{-1}\) and a thickness of 3.57 ± 0.06 km with an rms residual of 0.031 s. The second reflected phase produces an interval velocity of 6.04 ± 0.03 km s\(^{-1}\) and a thickness of 2.21 ± 0.08 km with an rms residual of 0.028 s. The estimated layer parameters for the fourth and fifth layers are shown by velocity values in km s\(^{-1}\) at the right-hand side of Fig. 6(b), and the thickness or depth, by dashed lines. The theoretical reflection times corresponding to the estimated velocity model are shown by the curves in Fig. 6(a). As the present method is based on ray theory and the calculation of ray parameters at the mid-distance of a data set is quite accurate, there is no problem even with the estimation of a low-velocity zone (LVZ), as has been calculated for the second reflected phase in the upper crust. The 2-D velocity model (Kaila et al. 1990) has been superimposed on the calculated 1-D velocity model in Fig. 6(b) to show the comparison and reliability of the proposed method.

5 CONCLUSIONS

The proposed method of calculating interval velocities and layer thicknesses is applicable to a set of reflection times at any non-zero offset distance. The method is very simple and direct in nature, and can be easily implemented on a computer for a quick estimation of both velocity and structure. The 1-D velocity models, obtained by the present method from different sets of wide-angle reflection times corresponding to various shot points along a deep seismic reflection profile, will produce a pseudo 2-D velocity picture of the sub-surface when integrated together, and can provide a good starting model for further refinement of the velocity structure by 2-D modelling (Bishop et al. 1985; Bording et al. 1987; Lutter et al. 1990; Zelt & Smith 1992).

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Figure 6. Traveltimes of two wide-angle reflected phases (marked by $p^{3,8}$ and $p^{6,6}$ in Fig. 7 of Kaila et al. 1990) from layers below the basement and their comparison with the traveltimes generated for the estimated velocity model. (b) The estimated 1-D model is shown superimposed on the available 2-D velocity structure (Kaila et al. 1990) for which thickness or depth is indicated by a solid line and velocity values (km s$^{-1}$) are shown in the middle of the model. The velocities of the 1-D model are shown in italics and the thicknesses by the dashed lines. The first three layers of the Kaila et al. (1990) model have been averaged into a 1-D model, and the fourth and fifth layer parameters have been calculated by the present method.

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