Full-wavefield inversion of wide-aperture SH and Love wave data

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Accepted 1991 January 15. Received 1990 November 15; in original form 1990 July 20

SUMMARY
Reverse-time linearized inversion is implemented for synthetic wide-aperture SH and Love wave data, including multiple reflections, to estimate the S-wave velocity in a two-dimensionally inhomogeneous medium. Complete wide-aperture wavefields, in which triplications, pre- and post-critical reflections, and surface waves (Love waves) are present, can be imaged by reverse-time inversion to estimate the SH velocity distribution. The algorithm operates on common-source gathers and involves cross-correlation of the source and recorded wavefields to define the steepest descent direction to update the velocity at each point on a finite-difference grid. Effects of noise can be reduced by stacking information from different sources.

Key words: inversion, wide aperture.

INTRODUCTION
Wide-aperture data have recently been receiving considerable attention in the United States because of their increased availability to the academic community via national programmes such as EDGE and PASSCAL. Activity is also high in many other countries; representative examples are the BIRPS project in the United Kingdom and LITHOPROBE in Canada. Processing of data from such experiments still poses difficulties. The approaches that are currently common are based on either traditional refraction analysis, such as ray tracing or synthetic seismogram modelling (cf. Hamilton 1989; Jardine 1988; Sun & McMechan 1989) or on traditional reflection analysis, such as normal moveout corrections or slant stacking (cf. Plappert 1987; Chang & McMechan 1989).

Recent advances in processing of wide-aperture data include tomographic imaging (cf. Novotný 1981; Lutter & Nowack 1990; Lutter, Nowack & Braile 1990; Zhu & McMechan 1989; Hamilton 1989) and pre-stack migration of common-source gathers (cf. Chang & McMechan 1989; Chang, McMechan & Keller 1989). Both of these are steps in the direction of full wavefield processing, but are still limited in significant ways. Tomography is based on time picks and so uses only a small subset of the total data; pre-stack migration requires prior estimation of velocities, and most importantly, does not correct for the dependence of amplitude upon incident angle (McMechan & Fuis 1987).

These restrictions can be largely overcome by full-wavefield inversion.

This paper is the first of two in which we demonstrate full-wavefield inversion of wide-aperture seismic data. In this paper we consider inversion of SH and Love surface wave data (i.e., the transverse component of a three-component wavefield); in the companion paper we consider inversion of P and SV body waves and Rayleigh surface wave data (i.e., the vertical and radial components of a three-component wavefield). These two are functionally equivalent to the full acoustic and full elastic cases, respectively. The discussion and examples are restricted to 2-D media.

SEISMIC INVERSION
Overview
Linearized inversion of full-wavefield seismic data has existed conceptually for some time [see Jurkevics, Wiggins & Canales (1980) for an early example], but it has been only during the past five years that the required computer capabilities have been widely available.

Inversion of 2-D seismic data has been discussed by Tarantola (1984a, b, 1986, 1987). The object of inversion is to directly estimate the physical parameters of a medium; the procedure is to iteratively adjust the model parameters to minimize the residual between the observed seismograms and seismograms computed for a trial model.

Sun & McMechan (1986) for elastic VSP data. The relation between inversion and migration has been discussed by Tarantola (1984a) and Bourgeois, Jang & Lailly (1989) among others; basically, migration may be thought of as one iteration in iterative inversion.

The present paper involves estimation of the S-wave velocity distribution in a 2-D model using synthetic SH and Love waves for sources and recorders on the Earth's surface. This is an inherently non-linear problem; its solution is approximated by a sequence of small linear steps from a starting model that is close to the correct solution. The main difference of this work from previous 2-D full-wave inversions is the explicit inclusion of post- as well as pre-critical reflections, and of multiples. The inversion algorithm is, like that of the previous work on elastic VSP data by Sun & McMechan (1988), obtained by combining the theory of linearized inversion with concepts borrowed from reverse-time migration. The imaging condition in reverse-time migration is the time-coincidence between a reflected wave and the forward propagating wave from the primary source. The steepest descent direction of velocity adjustment in inversion is determined by cross-correlation, which is simply a time-coincidence condition, between the forward propagating wavefield from the source and the reverse-time propagating reflection residual wavefield. All reflected waves, both pre-critical and post-critical, are automatically, simultaneously included in the calculation of the cross-correlation which defines the velocity adjustment at each iteration.

The remainder of this paper consists of a summary of the theory followed by a presentation and discussion of examples of processing both noise-free and noisy data. The examples include both laterally homogeneous and inhomogeneous media imaged using pre- and post-critically reflected SH-waves, and Love waves [which are, physically, the superposition of closely spaced, post-critically reflected, free-surface SH multiples (Kelly 1983)].

### Generation of synthetic test data

The propagation of an SH-wave in a locally homogeneous, isotropic, elastic medium follows the wave equation (e.g. Aki & Richards 1980)

\[
\frac{\partial^2 u(x, t)}{\partial t^2} = \beta^2 \nabla^2 u(x, t) + f(x, t),
\]

where \(u\) is the SH particle displacement, \(\beta\) is the S velocity, \(t\) is time, \(x\) is the spatial position vector, and \(f\) is the source function. For the 2-D problems considered below, \(x\) has two components \((x, z)\) in the horizontal direction and \(z\) in the vertical direction. Synthetic test data are computed for a specified model, using a second-order, explicit finite-difference solution of equation (1). The upper (free surface) boundary of the finite-difference grid is assumed to be stress-free. The absorbing boundaries of Cerjan et al. (1985) are applied at the other three sides of the grid.

An SH-wave is emitted from a source at location \(x_s\). The (one-component) particle displacements recorded on the surface of the medium, as a function of a series of receiver locations \(x_r\) and time \(t\), are the observed data \(u_{ob}(x_r, t; x_s)\). For real data, this modelling step would be replaced by a physical experiment.

### Inversion procedure

To estimate the S velocity distribution in the medium, the first step is to estimate the smoothed (low-wavenumber) content of that distribution; for real data, this should be constrained by all available geologic information, and produce synthetic data whose arrival times are within one-half period of those in the observed data at the lowest reliable frequency (cf. Mora 1989). Using the same seismic source location and time function as that used in creating the observed data, the seismic response to the trial model can be computed using finite differences; for real data processing, this assumes that an independent estimate of the source wavelet has been made (or alternatively, that the source wavelet is estimated as part of the velocity inversion).

The resulting responses to the trial model at the receiver locations \(x_r\) (at the same locations at which the observed data \(u_{ob}\) were recorded) are the theoretical data \(u_{th}(x_r, t; x_s)\). The residual common-source wavefield is the difference between the theoretical and observed wavefields

\[
u_{res}(x_r, t; x_s) = u_{th}(x_r, t; x_s) - u_{ob}(x_r, t; x_s).
\]

This residual waveform is input to the inversion. The \(L_2\) norm of the residual is taken as the misfit function in defining convergence.

The goal of inversion is to adjust the velocity in the trial model such that \(u_{res}\) is minimized. A minimum is found along the steepest descent direction of the misfit function.

The procedure used here is similar to that described by Sun & McMechan (1988) for elastic waves, so only a brief summary is presented here.

The common-source residual wavefield is propagated, backward in time, into the trial velocity distribution (cf. Sun & McMechan 1986). The data act as time-dependent boundary conditions that drive the finite-difference mesh by insertion of successive time slices from \(u_{res}\), in reverse-time order (one slice at each time step), into the finite-difference grid, at the recorder locations. Let \(p(x, z, t)\) be the resulting SH displacement field at all grid points at all time steps. Similarly, let \(u(x, z, t)\) be the SH displacement field at all grid points, at all time steps, produced by the forward propagating wave from the source location.

An equation to find the steepest descent direction, in inversion of an acoustic wavefield, for a distribution of acoustic impedance, was given by Tarantola (1984a), in terms of cross-correlation of the time derivatives of \(u(x, z, t)\) and \(p(x, z, t)\). The main differences in our algorithm are that we are inverting directly for the shear wave velocity \(\beta\) in equation (1) and we cross-correlate \(u(x, z, t)\) with \(p(x, z, t)\). Implementation is approached from the point of view of reverse-time migration; i.e., to try to find the steepest descent direction for the reflections, by the time-coincidence imaging condition. The parts of the structure that are imaged (in migration), or where velocities are to be updated (in inversion), are those where the incident wave from the source \([u(x, z, t)]\) and the residual wavefield from the medium \([p(x, z, t)]\) are time and space coincident (or are in-phase). In inversion, we implement this condition by cross-correlation of \(u(x, z, t)\) and \(p(x, z, t)\) at each time step in the finite difference extrapolation of these wavefields.
The steepest descent direction is determined from
\[
m(x, z) = \left( \int_0^T u(x, z, t) p(x, z, t) \, dt \right) \nabla \phi(x, z, t) \bigg|_{t=T},
\]
where \( T \) is the final recording time.

Equation (3) does not give the exact amount of velocity adjustment, but gives the relative adjustment at each point (i.e., the steepest descent direction). \( m(x, z) \) in equation (3) is multiplied by a factor \( r \) before it is applied to update the velocity in the trial model. That is, if \( \beta_i(x, z) \) is the current velocity distribution, then
\[
\beta_{i+1}(x, z) = \beta_i(x, z) + rm(x, z)
\]
is the updated velocity distribution. The factor \( r \) is determined such that the \( L_2 \) norm of the residual becomes a minimum after the velocity in the trial model has been adjusted along the steepest descent direction (see appendix C of Gauthier et al. 1986). One iteration involves looping over the data from all sources. The velocity \( \beta_{i+1}(x, z) \) in the adjusted (updated) model is then used to compute the theoretical data for the next iteration. This process is repeated until the \( L_2 \) norm of the residual reaches a minimum, or falls to an acceptable level.

Other approaches to determination of model updates exist (cf. Tarantola 1984a, b, 1987; Al-Yahya 1989) that differ in some details; each may be more or less computationally demanding, or produce faster or slower convergence, or be slightly more or less stable. In the present context, in which our emphasis is on new applications, we opted for the steepest descent approach because of its reliability and stability; we are more interested in getting a reliable solution than a fast solution. In the final analysis, the details are not crucial as long as the model perturbations produce convergence; the path of convergence has negligible effect on the final solution. The only exception to this is if the solution corresponds to a local minimum, which usually indicates insufficient accuracy of the starting model or inadequate model constraints, or a high level of noise, rather than the algorithm chosen for model perturbations.

SYNTHETIC EXAMPLES

Since all synthetic SH data were produced with a finite-difference solution of equation (1), all scalar wavetypes are present. Wide-aperture SH-wavefields contain direct waves, pre- and post-critical reflections, wide-aperture refractions, triplications, internal multiples, and Love waves. For illustration and discussion, we present four wide-aperture examples; these are for SH-waves and Love waves in vertically inhomogeneous media, and SH-waves and Love waves in 2-D heterogeneous media.

Laterally homogeneous model: SH-waves

Figure 1(a) shows a laterally homogeneous model consisting of three layers overlying a half-space, and six source locations. Synthetic wide-aperture data were computed for this model for these six sources; a representative common-source gather, for the source at \( x = 0.0 \) km, is shown in Fig. 2(a). The data clearly show a triplication in the primary branch (D-P-R) and free surface multiples (M1 and M2), and thus contain most of the features usually associated with wide-aperture data.

Figure 1. SH velocity distribution for a 1-D test model. (a) is the actual distribution (the correct solution); (b) is the starting model for inversion; (c) is the image at six iterations; (d) at 13 iterations; and (e) at 24 iterations. In all panels, the SH velocity vector, at each horizontal location, is plotted with zero amplitude if equal to 2.0 km s\(^{-1}\), blackened positive amplitude if larger than 2.0 km s\(^{-1}\) and unblackened negative amplitude if less than 2.0 km s\(^{-1}\). Maximum and minimum SH velocities are 3.0 and 1.5 km s\(^{-1}\) respectively. The six black dots in (a) are the locations of the six sources used. Representative data for the source at \( x = 0.0 \) km are shown in Fig. 2(a).
Figure 2. SH-wave responses to the laterally homogeneous model in Fig. 1, for the source at $x = 0.0$ km. (a) is the complete observed response; (b) is the residual wavefield before iteration 1; and (c) is the residual wavefield at iteration 24. In (a), D is the direct SH-wave, P is the primary reflection, R is a refraction in the lower half-space, and M1 and M2 are the first and second free-surface multiple reflections. For clarity, only every tenth trace is plotted in each panel.

Figure 1(b) shows the starting model for inversion; it contains the low wavenumbers in the model, but not the high-frequency details. For real data, such a smooth initial model may be obtained by tomographic imaging of travel times (see Zhu & McMechan 1989). The corresponding residual wavefield is shown in Fig. 2(b). The output SH velocity distribution, after 24 iterations (Fig. 1e) closely approximates the correct solution, and the corresponding residual wavefields (that portion of the input wavefield that is not accounted for by the model at iteration 24) is very small (Fig. 2c). After six iterations (Fig. 1e), the main features of the solution are already visible; as iterations increase (toward convergence), the high-wavenumber details emerge. This behaviour is characteristic of all the examples in this paper.

In Figs 1(c), (d), and (e), there is a faint ghost image (at approximately 3 km depth) corresponding to the first free-surface multiple reflection. Such ghost images gradually disappear as the number of iterations increases; the multiple reflections in the data are automatically accounted for by the shallow structures.

**Heterogeneous model: SH-waves**

A second example is provided by the 2-D heterogeneous model in Fig. 3(a). The structure includes curved layers and a vertical fault. Six sources are used. The synthetic observed SH data for the representative source at $x = 0.0$ km are shown in Fig. 4(a); pre- and post-critical reflections and triplications are clearly present. The starting velocity model for inversion and the residual SH-wavefield at the first iteration are shown in Figs 3(b) and 4(b), respectively.

Figures 3(c), (d), and (e) contain the inversion outputs after six, 13, and 24 iterations. The output after 24 iterations is quite close to the correct solution (Fig. 3a). The residual SH-wavefield at iteration 24 (Fig. 4c), is negligible compared to that at the beginning (Fig. 4b). Note that convergence generally proceeds downward from the surface; deep structure cannot fully converge until the uppermost structure is nearly correct so that distortions of the deeper penetrating waves, by the near-surface structure, are accounted for.

The foregoing examples demonstrate that reverse-time inversion works well even with wide-aperture reflection data in which post-critical reflections or triplications are dominant.

**Laterally homogeneous model: Love waves**

Love waves are the result of superposition of multiple free-surface post-critical SH reflections (cf. Kelly 1983). Since post-critical reflections can be inverted by the procedure shown above, this suggests that it can also be used directly for reverse-time inversion of Love waves.

Figure 5(a) contains a laterally homogeneous test model for Love wave inversion. 10 sources are located on the surface of the medium. The Love wave is clearly seen in the synthetic observed data for the representative source at $x = 0.0$ km (Fig. 6a).

The starting velocity model for inversion is shown in Fig. 5(b); the corresponding residual wavefield is in Fig. 6(b). Note that, although the starting model is similar to the correct solution, the residual wavefield is large; Love waves are sensitive to structural details. The estimated velocity distribution after 24 iterations and the corresponding residual wavefield are shown in Figs 5(c) and 6(c), respectively. Reverse-time inversion is clearly an effective technique for Love wave processing, as well as SH-wave processing.

**Heterogeneous model: Love waves**

Figure 7(a) contains a 2-D heterogeneous SH velocity distribution. This model includes curved interfaces and a near vertical fault. 10 sources are used. The synthetic observed Love wave data for the representative source at $x = 0.0$ km are shown in Fig. 8(a); Love waves (L1) propagating in the left part of the model are reflected from the fault (L2) and transmitted through the fault (L3).

The starting velocity model for the inversion is shown in Fig. 7(b); the corresponding residual Love wavefield at the
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Figure 3. SH velocity distribution for a 2-D test model. (a) is the actual distribution (the correct solution); (b) is the starting model for inversion; (c) is the image after six iterations; (d), after 13 iterations; and, (e), after 24 iterations. In all panels, the SH velocity vector, at each horizontal location, is plotted with zero amplitude if equal to 2.0 km s\(^{-1}\), blackened positive amplitude if larger than 2.0 km s\(^{-1}\), and unblackened negative amplitude if less than 2.0 km s\(^{-1}\). Maximum and minimum SH velocities are 4.5 and 2.0 km s\(^{-1}\) respectively. The black dots in (a) are the locations of the six sources used. Representative data for the source at \(x = 0.0\) km are shown in Fig. 4(a).

First iteration is shown in Fig. 8(b). The estimated SH velocity distribution after 30 iterations and the corresponding residual wavefield are shown in Figs 7(e) and 8(c), respectively. The correct model (Fig. 7a) is effectively estimated, except for the velocity of the lowermost layer; even so, note that the velocity contrast at the interface at the top of this layer is accurately estimated. This is expected as the corresponding low-wavenumber information was not present in the starting model, nor recoverable from the bandlimited data.

Inversion of Noisy Data

Noise present in the seismic data from individual sources will produce artifacts in cross-correlation that are not related to the velocity deviation to be estimated. If more than one source is used, the image related to the actual velocity deviation is coherent, in contrast to the random image due to the noise. By stacking the cross-correlations obtained from all sources, parts of the images related to velocity deviation will be enhanced because they are coherent, whereas the parts of the images related to noise tend to cancel because they are incoherent.

The stacked image corresponds to a different steepest descent \(m(x, z)\) distribution (equation 3) and a different model adjustment (equation 4) to provide more reliable model updates than the image from any single source. Since this effect is similar on artifacts in images produced from noise-free data, we used this approach for all the examples in this paper. The result is a more stable solution; a direct parallel may be drawn between the relative instability of algebraic reconstruction, and the relative stability of simultaneous iterative reconstruction in tomography (cf. McMechan, Harris & Anderson 1987). In the end, the noise and artifacts tend to remain in the unfitted part of the residual traces for each common-source gather. In this way the disturbance to the solution, due to noise, can be reduced to a negligible amount. Even coherent noise is not expected to have a significant effect unless it is coherent across shots.

All the foregoing examples used noise-free data to illustrate the behaviour of the inversion algorithm under ideal conditions. All these examples were re-done after the addition of random noise (with signal-to-noise ratio approximately 2 dB), to each common-source data set; the model, the procedure, the numbers and locations of sources and receivers, and the starting model, were identical.

Figures 9 and 10 contain an illustrative example that is the noisy equivalent of that in Figs 3 and 4. A representative noisy common-source gather, for the source at \(x = 0.0\) km is shown in Fig. 9(a). The residual wavefield at the first iteration is shown in Fig. 9(b). The output velocity estimate after 24 iterations and the corresponding residual SH-wavefield are shown in Figs 10 and 9(c), respectively. Comparing the noise free data and residuals (Figs 4a, b and c) with the noisy data and residuals (Figs 9a, b and c) at various iterations, shows that inversion tends to fit the coherent parts of the data, leaving the incoherent noise in the residual. This is a very desirable characteristic. Similar results were obtained in tests performed on other noisy data.

Discussion and Synopsis

Inversion of synthetic SH data, from surface sources and receivers may be used to estimate the SH velocity in the medium. The complete SH-wavefield, including multiples, pre- and post-critical reflections, and Love waves, can be simultaneously inverted. To minimize the \(L_2\) norm of the
Figure 4. The SH-wave response of the heterogeneous model in Fig. 3, for the source at \(x = 0.0\) km. (a) is the complete observed response; (b) is the residual wavefield before iteration 1; and, (c) is the residual wavefield at iteration 24. For clarity, only every tenth trace is plotted in each panel.

residual, the steepest descent direction is followed. The steepest descent direction is found by considering the imaging condition calculated by the cross-correlation in equation (3).

As in all linearized inversions, convergence to the global minimum can be assured only if the solution is sufficiently constrained. These constraints are largely implicit in this algorithm and take a number of forms. First, it is assumed that the low wavenumbers in the \(SH\) velocity distribution are known \(a\ priori\) and are contained in the starting model. In this context, low wavenumbers are below those corresponding to the source wavelet. If there is no overlap between the wavenumbers in the input model and those represented in the data, convergence may be to a local minimum as a wavenumber window will be unrestrained. If there is some overlap, updates of the corresponding model wavenumbers will occur as iterations proceed.

Mora (1989) provides important insights into the interaction between high and low wavenumbers and the conditions necessary for resolution of low wavenumbers. In the present context, the low wavenumbers are expected to be provided, for example, by pre-inversion traveltime tomography using time picks from the data to be inverted; thus, tomography and iterative migration are complementary components of a complete inversion scheme (Mora 1989).

If the observed and synthetic data for the primary reflections are within one-half period of each other, the solution will move toward the global minimum (as these reflections contain the largest amplitudes and hence correspond to the largest reduction in residual). If they are not within one-half period of each other, cycle skipping may occur and the solution may move toward a local minimum. To facilitate systematic convergence at the higher frequencies we applied a low-pass filter to the cross-correlation; the high cut corner was moved to progressively higher frequencies as iterations proceeded. The high-wavenumber limit of resolution is also limited by the highest frequencies in the data. Within the reliable bandwidth in the data, all features can be accurately estimated.

The large amplitude primary reflections will control the model updates in the first few iterations. Then, as the velocity distribution becomes more accurate, the (low amplitude) multiples will also satisfy the half-period criterion and will begin to contribute to a higher level of detail in the model.

For application to real data, it is necessary to estimate the source function. This can be done empirically by deconvolution, or by solving for the source wavelet as additional unknowns in the inversion. Because the angle dependence of reflectivity is the key information that is inverted, the source directivity is as important as its time function.

For the synthetic inversion examples above, input data traces were available at each finite-difference grid line. For real data, the spatial density is expected to be less, and the spacing, more random; in practice, the requirement is only that the data not be spatially aliased so that reverse-time
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Figure 6. The Love wave responses of the laterally homogeneous model in Fig. 5(a), for the source at \( x = 0.0 \) km. (a) is the complete observed response; (b) is the residual wavefield before iteration 1; and (c) is the residual wavefield at iteration 24. For clarity, only every tenth trace is plotted in each panel.

Figure 7. 2-D \( SH \) velocity distribution for Love wave synthesis and inversion. (a) is the actual distribution (the correct solution); (b) is the starting model for inversion; (c) is the image after 13 iterations; (d) is after 20; and (e) is after 30. In (a) the velocity of the upper layer is 2.5 \( \text{km s}^{-1} \); of the intermediate layer, 3.5 \( \text{km s}^{-1} \); and, of the lower half-space, 4.0 \( \text{km s}^{-1} \). The black dots in (a) are the locations of the 10 sources used. Representative data for the source at \( x = 0.0 \) km are shown in Fig. 8(a).

Figure 8. The Love wave responses of the laterally heterogeneous model in Fig. 7(a), for the source at \( x = 0.0 \) km. (a) is the complete observed response; (b) is the residual wavefield at iteration 1; and (c) is the residual wavefield at iteration 30. For clarity, only every tenth trace is plotted in each panel.

Ghost images corresponding to multiple reflections may be present at early iterations. However, ghost images will be suppressed as iterations proceed; the velocity distribution is constrained by multiple, as well as primary, reflections.

Inversion of noisy data is facilitated by the stacking of velocity perturbation images from different sources at each iteration. This concept is borrowed directly from exploration seismology; incoherent noise tends to remain in the unfit residual, while coherent signals add constructively to the solution. A high level of noise will make the solution more non-linear and more prone to converge to a local minimum. The same is true if an approximate method were used to compute the theoretical data, as energy not accounted for would be treated as noise.

Full-wavefield inversion, as demonstrated above, can be extended to the elastic (P-SV-Rayleigh wave) case (which is currently in progress) and to three spatial dimensions. Ideally, real data processing will be done in 3-D, if 2-D processing is attempted, amplitude corrections for geometrical spreading will have to be applied (Crase et al. 1991). Simultaneous solution for other parameters, such as density and anelastic attenuation, which have previously only been done in 1-D (cf. Singh et al. 1989; Martinez & McMechan 1990) can also be extended to 2- and 3-D.

Since the scalar wave equation (1) is equally applicable to SH and acoustic waves (Kelly 1983) the algorithm presented above can also be directly applied to compressional (acoustic) seismic waves. The fully coupled P-SV case requires replacement of the scalar wave equation with the elastic wave equation. The elastic case also requires two-component (vertical and in-line horizontal) recording.

The wide-aperture (post- as well as pre-critical) example and the conclusions of this paper are consistent with those of Mora (1989), who provides lucid summary of the principle of inversion in the pre-critical regime. The key points are: (a) the use of iterative linearized inversion removes the necessity of knowing the reflector depths a priori; (b) the use of very low wavenumbers (those not present in the data) are implied a priori; (c) to avoid local minima, the initial velocity model should describe the kinematics of wave propagation to within about a half period of the source wavelet; (d) the velocity model tends to converge from top down, because the half wavelength criterion (c) is satisfied at increasing depths as iterations proceed; and (e) the presence of multiple reflections corresponds to non-linearity that slows convergence, but does not preclude obtaining satisfactory solutions.

Full-wavefield inversion involves higher than average computational costs, but much lower than average human effort in processing and interpretation. No a priori identification of arrivals is necessary, and the output is already in 'geologically meaningful' form (i.e., velocity rather than reflectively). The examples above typically took \( \approx 3.5 \) min of CPU time per iteration per source on CRAY-XMP, so a data set can easily be completely processed overnight.

The cost may be reduced by substitution of more efficient wave propagation algorithms. It is important, however, that the full wave equation be used to ensure accurate simulation of all angle-dependent effects. Convergence may be accelerated by substituting other methods (e.g., conjugate gradients) for the steepest descent solution or by posing the solution as a non-linear problem that uses predictions based on the model update trajectory; these involve few iterations, but more computations at each iteration, and may be less stable. Also, as each common-source gather extrapolated independently, a parallel computing environment can be used to advantage by extrapolating a number of common-source gathers simultaneously.
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This paper, is to our knowledge, the first test of the feasibility of full-wavefield inversion of wide-aperture (including post-critical) data containing multiples and surface waves as well as body waves.

ACKNOWLEDGMENTS

The research leading to this paper was funded by the sponsors of the UT-Dallas Geophysical Consortium and by the National Science Foundation under grant EAR-9015852. Computations were performed at the Center for High Performance Computing of the University of Texas System. Constructive reviews by R. Nowack and two other reviewers were appreciated. The manuscript was expertly typed by Anna Radasinovich. Contribution No. 671 from the Programs in Geosciences at The University of Texas at Dallas.

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