Table 1. Normal mean square power in the deconvolved output

S. no.	Kalman predictor model	Levinson
Marine data		
1	0.6334	0.9507
2	0.4739	0.936
3	0.1899	0.7816
4	0.1033	0.7614
Land data		
1	0.1172	0.9397
2	0.0350	1.0403
3	0.0182	1.0756
4	0.0067	1.0337

predictor. The Robinson type predictor always results in a stable predictor due to the minimum phase criterion. However, deconvolution results are much better in the case of the Kalman predictor model. The above algorithm can fail to give satisfactory results when the data are nonstationary. Another limitation is that the innovation model is not suitable for direct estimation of the reflection sequence since all the information required is not available. One way of overcoming these difficulties to some extent is by making use of adaptive estimation techniques.

Acknowledgments

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Seismic 10—Migration

Iterative Depth Migration by Backward Time Propagation S10.1

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This paper discusses a method for doing depth migration of common-midpoint stacked sections in a variable velocity media. Migration is posed as a time variable boundary value problem where a form of the scalar wave equation is solved numerically (e.g., by finite difference techniques). The problem is solved in time domain with the time-reversed seismic section applied as upper surface boundary conditions. This is in contrast to many migration schemes which apply the surface section as initial values for a depth extrapolation technique based on the wave equation.

While the time-domain solution to the wave equation is generally more costly in terms of computer economics, it can handle arbitrary dip in a variable velocity field with no instabilities in the algorithm. Because of this flexibility, the migration procedure can be used for very complex velocity models. It velocity boundaries are consistent with geologic boundaries, then depth migration can be used to test the validity of a model. If some knowledge of the vertical sequence of the velocities is available at a few lateral locations, then in many cases depth migration can be employed to help develop a better model.

The migration procedure numerically solves a full scalar wave equation as a time regression problem, thus there is no structural dip limitation. In fact, waves which have encountered a turning point can be migrated, provided the reflection data are recorded and adequately sampled at the surface. An example of this is shown in Figure 1, which is a prototype for a Gulf coast salt dome. Because of the increasing interval velocity structure with depth, the normal incidence section from this structure contains seismic events which were generated from beneath the overturned portion of the salt dome. This time section was then used as boundary conditions for the migration procedure, which reconstructed all recorded dips, including the overturn.

The increased sensitivity of depth migration procedures to interval velocity brings with it a paradox: If the velocity model is known, then migration is not needed, and if the velocity model is not known, then an essential input for migration is absent. Because of this paradox, migration itself is not an inverse method. Depth migration is, however, a very useful tool in unraveling complex structure. It offers, at the very least, a procedure by which an assumed structural model may be tested by comparing the depth migration with a depth model. If the comparison is not favorable, then modifications in the model must be made. In the case where the vertical sequence of velocities is known at a few locations (e.g., from well control, velocity analysis, or migration before stack), then migration can be used iteratively to help develop an improved geologic model.

Shown in Figure 2 is a schematic of an iterative migration procedure. An initial guess of a model is made, incorporating all known external velocity information. The seismic section is migrated, and a comparison between the migrated section and the model is made, and a new model is constructed in an attempt to resolve these differences and the procedure is repeated. This is in contrast to iterative modeling, where synthetic and real time sections are compared.

An example of the iterative process is shown in Figures 3-6. A geologic model and synthetic seismic section are shown in Figure 3. In the field situation, the model would not be known and the synthetic section would be replaced by a common-midpoint stack of the field section. To demonstrate the iterative procedure, we assume that the only knowledge of the velocity field in this synthetic example is obtained from a well (as indicated on Figure 3). While this model is reasonably complex, the vertical trend of the velocity structure does not change over the span of the model. From the

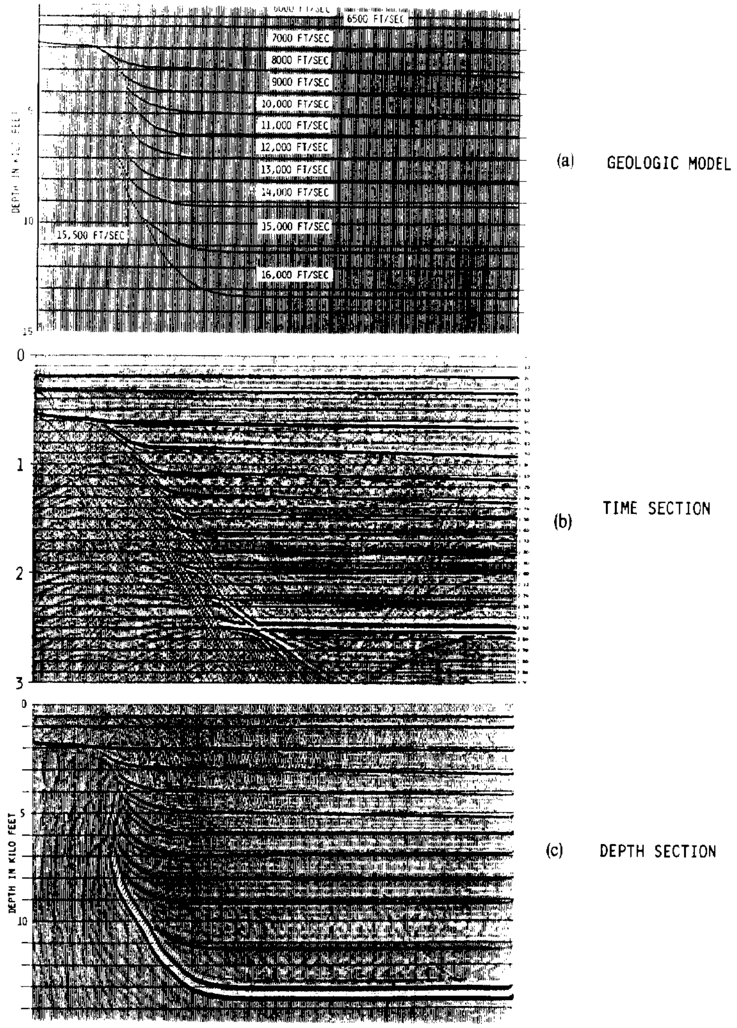


FIG. 1. (a) Geologic model. (b) Time section. (c) Depth section.

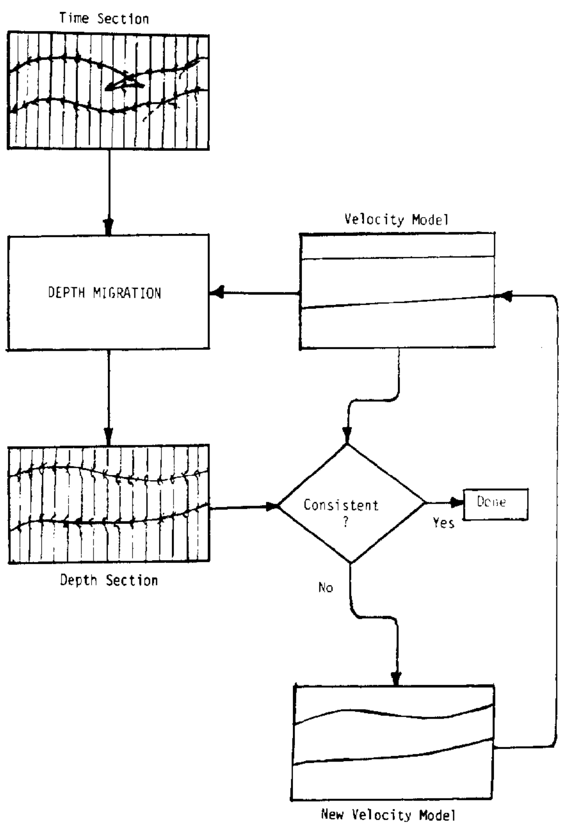


FIG. 2. Iterative depth migration.

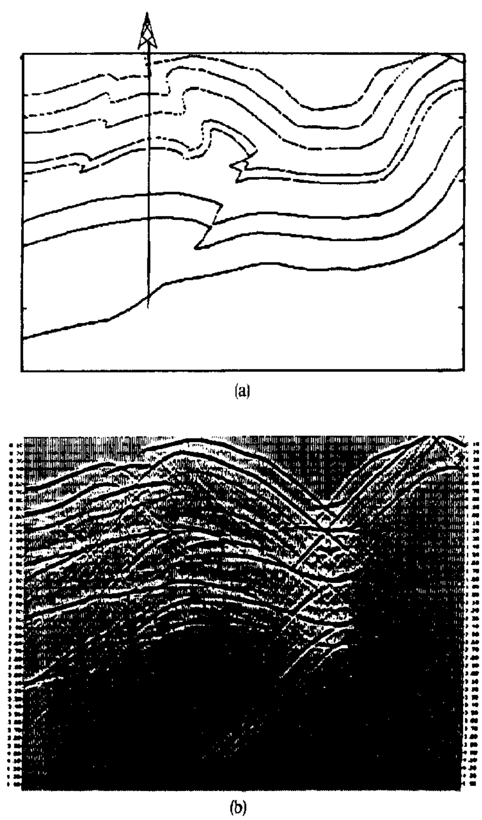
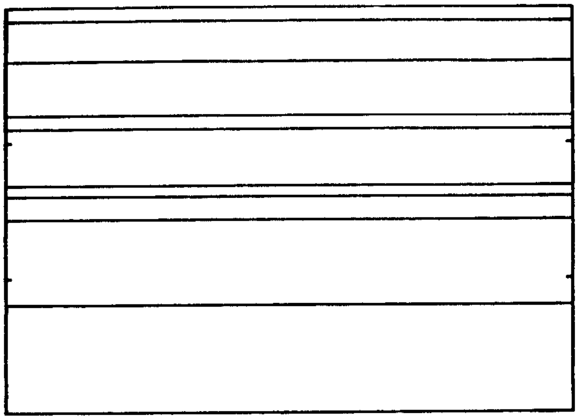
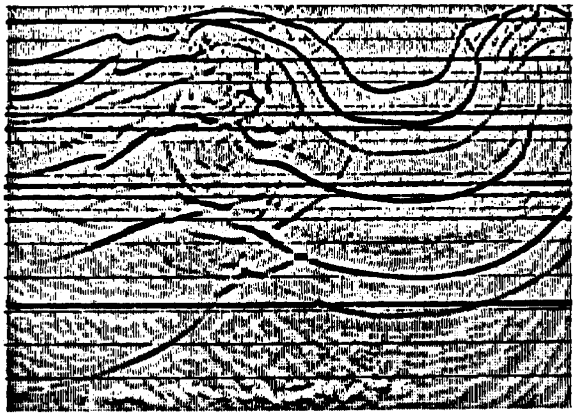


FIG. 3. (a) Geologic model. (b) Zero offset time section.

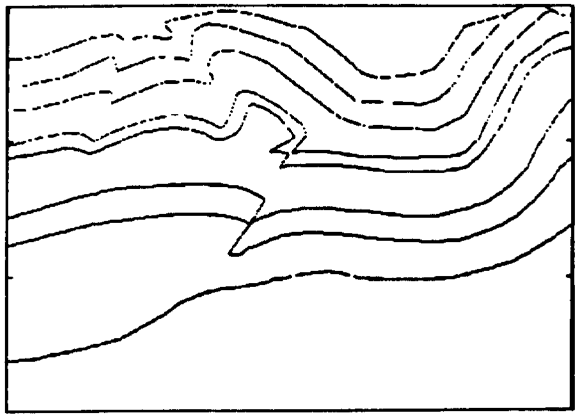


(a)

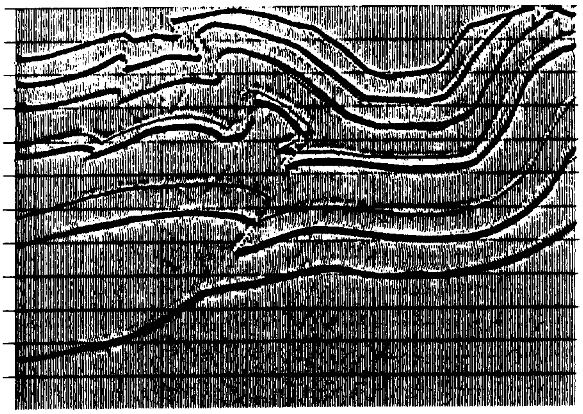


(b)

FIG. 4. (a) Initial velocity model. (b) Depth migration with model overlay.



(a)



(b)

FIG. 6. (a) Final velocity model. (b) Final depth migration.

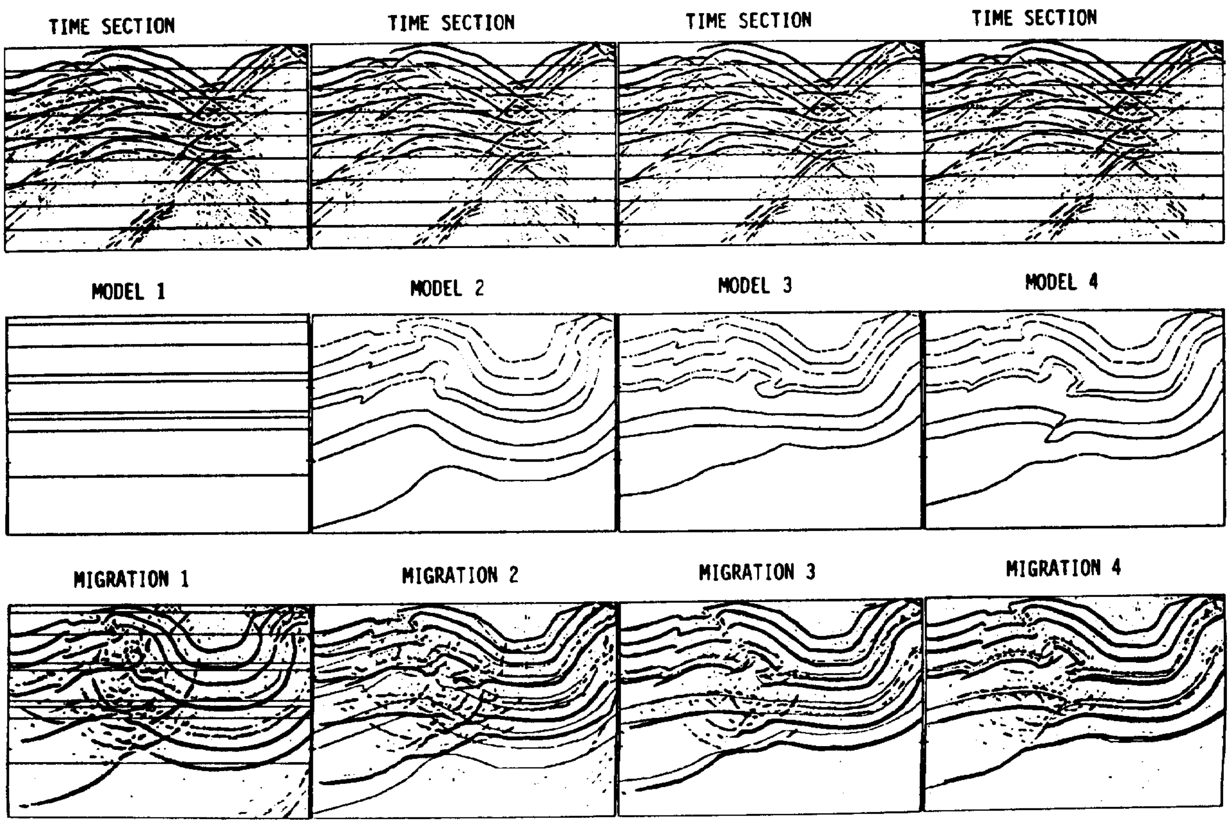


FIG. 5. Iterative depth migration of CMP sections.

"well" a horizontally layered model is constructed; the time section is depth migrated. A higher order guess might have been made if well dip information were available. The model and resultant migration (plus a model overlay) are shown in Figure 4. It is immediately obvious that the model, as indicated by the overlay, is not consistent with the depth section. Therefore, the model is not correct and must be modified. While there may be optimum ways of guessing a new model (e.g., there are obvious locations of over migration), a reasonable guess will be to assume that the reflectors themselves are closer to the correct geologic boundaries than the initial model was. The migrated section is therefore interpreted and digitized, producing a new model guess.

The migration procedure is repeated again, obtaining a new depth section and thus a new model; the process continues until a reasonable match is obtained between the migrated section and the model which produced it. A sequence of these iterations for this model is shown in Figure 5. The first row contains the input time section, the second row the model guess for each iteration (produced from the depth migration of the previous iteration), and the last row is the resultant depth migration with the model overlaid. It is worthwhile noting that the quality of the depth migration improves as each sequential model gets closer to the correct model. After a reasonable number of iterations, a final velocity model is obtained. In this case, the model and depth wave field are essentially coincident (Figure 6). This model can be compared to the actual model (for this synthetic case) and the comparison is quite good.

While this provides a format for doing iterative migration, the process is only as good as the data and reliable interval velocity information at each lateral location where the geology undergoes significant change in its vertical sequence. A CMP stack is generally only an approximation to the primaries only, zero offset section that all migrations assume. Furthermore, interpretation of the depth section is generally a nontrivial task. These pitfalls must always be observed in applying this procedure effectively.

Reliable structural definition requires a velocity sensitive migration procedure and adequate velocity control. If the interval velocity trend is known at representative lateral locations, interative depth migration can be used to help extrapolate this information away from the control locations and thus complex structure can be correctly imaged. In areas of good data quality, where seismic amplitudes are a relative measure of the change in interval velocities, a velocity estimation step could also be put in the migration procedure to help refine the velocity model.

Ray Asymptotic Migration (Basic Concepts) S10.2

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A new method of migration is presented. The method is based on local (ray) asymptotic formulas for the field of waves observed and calculated and for components of the Green's displacement and stress tensors. The method includes the following main steps.

(1) The fields

$$V_\alpha(0, A, t) = a_\alpha[0, A, t - \tau_g(A)] \cos \phi_\alpha[0, A, t - \tau_{ph}(A)] \quad (\alpha = 1, 2 \ldots),$$

of separate waves are detected from a given seismogram $U(0, A, t)$ (a source at 0, a receiver at A) by means of the automatic procedures of phase and group correlation. a_α and ϕ_α are the envelope and the phase of a wave, respectively. τ_g and τ_{ph} are the group and velocity traveltimes.

(2) The inverse kinematic problem for each αth wave is solved. This means that the position of the point of reflection M_α is found using the traveltime $\tau_{ph}^{(\alpha)}(\bar{A})$ and slowness $P_\alpha(\bar{A})$ for the middle point $\bar{A}$ for the array and knowing the structure of the overburden.

(3) If point $\bar{A}$ is not located in the special area on the front of αth wave, then the reflection coefficient K_α (M_α) at point M_α can be calculated using the mean value $V_\alpha(0, \bar{A}, t)$ of the observed field $V_\alpha(0, A, t)$, traveltime $\tau_{ph}(A)$, and formulas for the leading terms of the ray series for reflected and incident waves.

(4) The more universal method of an asymptotic migration is proposed for a general case. This method is based on formula of the Kirchhoff (Helmholtz) type for elastic waves, on the approximate formula for wave scattered by a body of arbitrary shape, and on asymptotic presentations for the Green's tensors.

A depth section for the reflectivities of the interfaces (inclusions) is constructed as a result of asymptotic migration.

Let the X_ith component (vertical ($X_3 = Z$) or one of the horizontal ($X_1 = X, X_2 = Y$)) $U_{xi}(X_n, X_m, t)$ of the displacement vector $\mathbf{U}(X_n, X_m, t)$ be given with the source and the receiver located at the points $0_n(X_m)$ and $A_m(X_m)$, correspondingly. The seismogram $U_{xi}(\xi_\ell, t)$ ($\ell = 1, 2 \ldots,$) is formed from a given set of traces $U_{xi}(0_n, A_m, t)$ ($n = 1, 2, \ldots, N, m = 1, 2, \ldots, N, m = 1, 2 \ldots, M$).

This seismogram can correspond, for example, to one of the following systems of observation:

$$U_{xi}(\xi_\ell, t) = U_{xi}(\mathbf{X}_{n\ell} = \text{const}, \mathbf{X}_{m\ell} = \xi_\ell, t), \text{ CSP method;}$$

$$U_{xi}(\xi_\ell, t) = U_{xi}(\mathbf{X}_{n\ell} = \xi_\ell, \mathbf{X}_{m\ell} = \text{const}, t), \text{ CRP method;} \quad (1)$$

and

$$U_{xi}(\xi_\ell, t) = U_{xi}(\mathbf{X}_{n\ell}, \mathbf{X}_{m\ell}, t),$$

method used when the position of the source and the receiver in the ℓth offset varies according to the given rule, as for example, in the common-midpoint (CMP).

The superposition (wavegram)

$$\bar{u}_{xi} = \sum_\alpha V_{xi}^{(\alpha)}(\xi_\ell, t) \quad (2)$$

of X_ith components $V_x^{(\alpha)}$, (ξ_ℓ, t) of separate waves is obtained as a result of application of the procedure of group and phase correlation (Gelchinsky et al, 1983) to each seismogram $U_{xi}(\xi_\ell, t)$.

Each component $V_{xi}^{(\alpha)}$ (ξ_ℓ, t) of the αth wave is presented in the form

$$V_{xi}^{(\alpha)}(\xi_\ell, t) = a_{xi}^{(\alpha)}[\xi_\ell, t - \tau_{gr}^{(\alpha)}(\xi_\ell)] \cos \phi_\alpha[\xi_\ell, t - \tau_{ph}^{(\alpha)}(\xi_\ell)] \quad (3)$$

where $a_{xi}^{(\alpha)}$ and $\phi_{xi}^{(\alpha)}$ are the envelope and the phase of the x_ith components, correspondingly, determined with the help of the Gilbert transform, τ_{gr} (ξ_ℓ) and $\tau_{ph}(\xi_\ell)$ are the group and the phase traveltimes, correspondingly. The envelope $a_{xi}^{(\alpha)}$ is a finite function