Physical properties of deep crustal reflectors in southern California from multioffset amplitude analysis

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ABSTRACT

Analysis of reflection waveforms before stack can constrain the physical properties of reflectors in the deep crust. To simplify this analysis, recorded amplitudes are assumed to be reflections from weak elastic heterogeneities. With these assumptions, trends in reflection amplitudes with offset may indicate whether the signs of a reflector’s density and rigidity contrast agree with or oppose the sign of its Lamé’s parameter contrast. The slope of the trend indicates the degree of Poisson’s ratio contrast. No attempt is made to invert for the individual modulus or density contrasts. By examining only gross amplitude-versus-offset (AVO) trends, deep reflections constrain some crustal properties.

INTRODUCTION

The purpose of this work is to explain the physical nature of reflectors in the deep crust. While the geometries of such structures have been imaged by several projects, their interpretation has been guided by indirect evidence. My approach is to analyze the reflected energy at varying angles of incidence and thereby estimate the sense of Poisson’s ratio contrasts, to constrain possible lithologic models.

Reflections from structures deep below the Mojave Desert in southern California were first recognized by Dix (1965). The extensive COCORP survey across the western Mojave imaged several deep crustal reflectors over a wide area (Cheadle et al., 1986). With the exception of one reflection which they believe can be traced to surface exposures, the geologic interpretation of middle and deep crustal reflections by Cheadle et al. (1985, 1986) was guided by reflector geometry. They consider the major reflectors to be low-angle fault structures.

To complement information on reflector geometry, amplitudes on prestack shot records can be used to constrain physical properties, such as changes in compressional and shear velocity or density. Reflection amplitude-versus-offset (AVO) relations can be found from the solutions of the Zoeppritz equations (Aki and Richards, 1980). Reflection power is, however, a complicated function of several variables, even at precritical angles. Approximations of the Zoeppritz equations, such as those given by Shuey (1985) and Wu and Aki (1985), help simplify the interpretation of observed AVO relationships.

Shuey’s (1985) simplifications point out, as Koefoed did in 1955, that the variation in Poisson’s ratio across an interface is a key parameter in determining how reflection amplitude varies with offset. Ostrander (1984) recognized that the saturation of a sandstone reservoir by natural gas would produce a remarkable decrease in Poisson’s ratio in comparison to surrounding unproductive rocks, increasing the amplitude of its reflection with offset. The validity of this type...
of analysis has been verified in several cases, such as by Chiburis (1984), in which information on subsurface properties was available from a large number of drillholes.

The amplitudes of deep crustal reflections at nonnormal angles of incidence have been considered by Richards (1961), Davydovalo (1972), Davydov et al. (1972), and Tulina et al. (1972), among others. These workers were principally concerned with establishing whether step discontinuities or layered transition zones would better fit the surprisingly large amplitudes often observed for crustal and Moho reflections. They did not consider the possibility that variations in Poisson's ratio might also produce strong reflections. For many reasonable deep-crustal structures, which would follow Koeoed's simplifying assumptions, changes in Poisson's ratio may produce strong AVO trends as well as large amplitudes at certain offsets.

To analyze physical properties at depth, one must find the AVO variations at the reflector. Unfortunately, there are many factors not related to the deep reflectors that often interfere with the seismic reflection amplitudes recorded, and all of these factors must be accounted for in some way before reflector properties can be estimated. Some of these effects were addressed by Richards (1961) and by O'Doherty and Anstey (1971). The factors affecting AVO relations can be divided into four broad categories: (1) factors due to the methods used to record, process, and interpret the seismic data; (2) surface-consistent factors related to near-surface phenomena in the sense defined by Taner and Koehler (1981); (3) factors due to the propagation of the seismic waves through the crust between the surface and the deep reflector; and (4) amplitude effects at the reflector not directly related to its contrast in physical properties. The exact evaluation of all these factors would itself require more subsurface information than is generally available. I will suggest how, under certain assumptions, gross AVO trends at a reflector can be separated from other effects.

I do not attempt to invert a measured AVO relationship for individual modulus and density contrasts. The data from the Mojave Desert are contaminated by a variety of effects that preclude measurement of precise amplitude responses of reflectors. However, strong AVO trends that are observed on regionally prominent deep reflections may be interpreted in terms of Poisson's ratio variations, helping to constrain geologic interpretations.

**METHODS**

**Data sets used**

The Calcrust consortium of California universities collected a total of 108 km of seismic reflection profiling in May and June of 1985. The survey was located along five lines in the Ward, Rice, and Vidal Valleys of the eastern Mojave Desert, southeastern California. The survey's main objective was to collect high-resolution seismic reflection data from the shallow part of the crust. The consortium was able to augment the main survey with secondary recording on stationary long-offset receiver spreads. Sorting these together with the main roll-along survey resulted in reversed, high-fold common-midpoint (CMP) records with offsets to 15 km over a substantial portion of three lines. This paper analyzes some of the highest quality data, recorded on line WM-1, where the long-offset CMP gathers span 8.3 km.

The COCORP Mojave survey (Cheadle et al., 1985, 1986) recorded several lines in the western Mojave. Line 3 is examined here; it was the longest survey line, totaling 87 km, running from northeast of the Rand Mountains southwest toward California City. It includes offsets to 10 km.

**Reflection amplitude analysis**

This analysis determines what aspects of the physical nature of reflectors can be constrained by multioffset amplitude information within the Mojave Desert data sets. I use the approach of Wu and Aki (1985) to simplify the relations between reflector physical properties and reflection coefficients. Assuming that the variations in density $\rho$, Lamé's parameter $\lambda$, and rigidity $\mu$ at a point scatterer are small, one can derive a system of equivalent forces for each reflected phase from each of the three types of scatterer. The combination of equivalent forces determines the AVO variation. These effects are linear; the forces will simply add if physical properties are combined. If the signs of the $\rho$ and $\mu$ variations are the same as the sign of the $\lambda$ variation, amplitude will decrease with offset. Conversely, if the signs of the $\rho$ and $\mu$ variations are the opposite of the sign of the $\lambda$ variation, amplitude will increase with offset.

The assumptions made by Koefoed (1955) and Shuey (1985) are a special case of the variations possible. Viewing amplitude effects as resulting from differences between the sign of $\delta\rho$, $\delta\mu$, and the sign of $\delta\lambda$ gives a more complete picture. Any combination of small physical property variations in the crust can be considered for its effect on AVO trends.

If additional information is available on the reflector, then an AVO trend can strongly constrain its nature. If a reflector is known to be a $+\delta\lambda$ variation, then the above relations show that increasing amplitude with offset results from an increase in Poisson's ratio $\sigma$, and decreasing amplitude would result from a decrease in $\sigma$. Conversely, a $-\delta\lambda$ variation would show increasing amplitude from a decrease in $\sigma$ and decreasing amplitude from an increase in $\sigma$.

These relations have been verified by simply calculating the Zoeppritz coefficients as given in Aki and Richards (1980, p. 150) at subcritical angles. For an example reflector at 12 km depth within a standard Mojave crustal velocity model (Kanamori and Hauk, 1975) with $+\delta\lambda$, a 13 percent Poisson's ratio increase from 0.347 to 0.392 will produce a 22 percent increase in reflection coefficient from 0 to 10 km offset. A 16 percent Poisson's ratio decrease from 0.347 to 0.292 will cause a 39 percent amplitude decrease at 10 km. Both effects would be greater if larger offsets were available.

With this strategy a relatively simple examination of the data will provide constraints on the physical nature of deep reflectors. By employing the AVO relations given above, the sign of Poisson's ratio variation can be estimated. Particularly strong $\sigma$ discontinuities can be identified by simply discerning whether reflection amplitude increases or decreases with offset, in the manner of Long and Richgels (1985). It may not be necessary to look for variations between amplitudes at three ranges of offset, as was done by Onstott et al. (1984), or to try to quantify the steepness of the
trends. Such a simple look at trend directions may be practical for deep crustal data sets.

**Effects on seismic amplitudes**

A number of factors that could corrupt amplitude observations with offset need to be taken into account. Some are listed by Ostrander (1984). These, and additional factors, can be divided into four categories based on where they arise in a seismic experiment.

**Recording and processing effects.**—The two reflection surveys used here, like most data sets, lack absolute calibration of geophone sensitivities or source strengths. In the manner of Wiggins et al. (1985), I use a quantile-based trace equalization to adjust for these variations. Amplitude histograms of each trace between 9 and 11 s two-way traveltime allow a quantile to be selected which separates background noise from identifiable reflections. Amplitudes above this quantile are assumed to be from reflections. With the assumption that background noise is constant, each trace will be multiplied by a constant amplitude factor to set all trace amplitudes at this quantile to the same value. Wiggins et al. suggest that the results of setting the quantile anywhere between the 50th and 90th percentile are indistinguishable. Tests on a few shot records agreed with that conclusion. I select the 70th percentile of the amplitude samples as a compromise that will allow a range of signal-to-noise ratios in the 2 s window at large time to be properly equalized.

A direct way to examine AVO trends is to incorporate the analysis into the stacking process. Only reflections which stack well will be examined for AVO trends, which should minimize the effects of phases other than primary P-wave reflections, although out-of-plane reflectors may also stack well. This analysis concentrates on the more horizontal, laterally continuous reflections, which best obey the assumptions of the stacking process.

To determine which reflections stack best, I use the methods of Harlan et al. (1984). Their technique calculates how well any particular result of a linear transform, such as stacking, fits the objective assumptions of the transform. For each sample of the stacked trace, the percentage of hyperbolic “signal” within the corresponding multifield gather will be calculated. Weighting the stacked section by the distribution of signal content emphasizes those events that best fit the hyperbolae defined by the stacking velocities.

To find the AVO trends, a linear regression of the root-mean-square (rms) amplitude of a window along the stacking hyperbola is performed against offset. These trends are plotted as a section similar to the stacked section, in the manner of Long and Richgels (1985), which I call an “amplitude trend stack.”

The process of calculating the linear regressions of amplitude against offset can be corrupted by source-generated noise in the gathers. Yu (1985) pointed out how multiple reflections could falsely bias the linear trends. Air and surface waves could also be sources of interference. However, I look for the more rapid variations of trend with respect to intercept time on the amplitude trend stack. Because source-generated noise has a slower apparent velocity across the gather than the reflections, it produces a broad bias in the amplitude trend section that will not vary rapidly with respect to intercept time. The trend bias could be reduced, for example, from the Calcrust data set by band-pass filtering that attenuated source-generated noise. The trends of primary reflections stand out against any remaining bias (e.g., on Figure 5 above 7 s) and can be evaluated relative to it.

**Surface-consistent effects.**—One surface-consistent effect concerns the directivity inherent in the use of arrays of vertical vibrators and geophones. In Ward Valley, in the eastern Mojave, first-arrival times on shot records show the existence of a surficial layer between 50 and 120 m thick with a velocity between 0.9 and 1.1 km/s. Such velocities are so low compared to the 5.5 to 6.8 km/s velocities at the depths of interest that any precritically reflected rays were bent to near-vertical incidence at the surface. Incidence angles at the surface were always less than 12 degrees from vertical. At 12 degrees from vertical, with a surface velocity of 1 km/s, the 14 geophones in two parallel arrays 30 m long used by Calcrust should have attenuated 35 Hz reflection amplitudes less than 5 percent relative to normally incident waves. The source arrays were shorter, with fewer elements, so they had less effect on far-offset amplitudes. In analyzing the Calcrust data set, source and receiver directivities are ignored.

The limit on incidence angles may not hold for the COCORP lines in the western Mojave. About half of line 3 traverses low-velocity 1 km/s surficial layers at least 30 m thick, but other areas have surficial velocities above 4 km/s. Reflections from a structure at 15 km depth recorded at the 10 km maximum offset would have arrived at 18 degrees from vertical or less. COCORP used long 200 m arrays of 24 geophones. These would have attenuated a 35 Hz arrival, at a 4 km/s surface velocity, by 35% relative to normally incident waves. This degree of receiver directivity would make an apparent trend of amplitude decreasing with offset of doubtful validity. However, measuring an apparent trend that increases with offset would simply underestimate the true trend at the reflector.

A more serious problem within this category is the effect of near-surface lateral heterogeneities. Goupillaud (1961) cautioned that the velocity contrasts near the surface are likely to be sharper and more inhomogeneously distributed than anywhere else in the crust. The multiplicity and reciprocity available from the two high-resolution seismic reflection experiments are relied on to address this problem. Such multiplicity allows the amplitude effects of lateral heterogeneities to be averaged out during the linear regression process. Lateral continuity of a reflection’s AVO trend in the trend stack suggests that surface heterogeneities do not interfere with the analysis.

**Propagation effects.**—The simplest effect on amplitudes is that of geometric spreading. I correct the amplitudes for the length of the travel path by assuming that the medium above the reflector has a constant velocity equal to the stacking velocity. In this case the correction is obviously $$G_t = tV_{\text{stack}}$$, where $$G_t$$ is the length of the travel path from the source, $$t$$ is the two-way travel time of the reflection, and $$V_{\text{stack}}$$ is the stacking velocity at that time. Newman (1973) derives a similar formula, $$G_{t0} = tV_{\text{stack}}^2/2$$, where $$v_0$$ is the surface velocity, for a multilayered medium at offset distances near zero, as a specific case of a multifield formula.
Figure 1 shows four different geometric spreading corrections for the range of experimental offsets. It is clear that Newman's zero-offset formulation $G_m$, the traveltimes and stacking velocity product $G_l$, and the actual path length $G_R$ through a homogenous crust topped by a low-velocity sedimentary section will all correct the far offsets relative to the near offsets with the same proportion. However, the simple $G_l$ correction appears to match the true path length best. In the trend stack, any under- or overcorrection of geometric spreading will result in a broad bias of the traces, rather than the sharp changes in AVO with respect to time that would come from reflection arrivals.

The focusing and defocusing of waves through lateral inhomogeneities may unpredictably affect AVO trends (Hubral, 1983, 1984). While heterogeneities as large as the maximum offsets will not allow events to stack well, smaller heterogeneities may disturb the linear regression analysis. Looking only at trends with lateral continuity in the trend stack may mitigate this problem.

A third effect on reflection amplitudes due to propagation is attenuation. Assuming that the medium above the reflector has a constant velocity equal to the stacking velocity, the AVO trend $T_Q$ due to apparent attenuation can be evaluated at different intercept times by

$$T_Q(\tau) = \frac{A_s}{h} \left[ \exp \left( -\frac{\pi f}{Q_{ap}} \frac{\tau^2 + h^2}{V_{stack}^2} \right) - \exp \left( -\frac{\pi f}{Q_{ap}} \tau \right) \right].$$

(1)

$\tau$ is the intercept time, $A_s$ is the amplitude of the wave at the source, $h$ is the offset range of the experiment, $f$ is the frequency, $V_{stack}$ is the stacking velocity at time $\tau$, and $Q_{ap}$ is the apparent attenuation. Calculations made varying $h, f$, and $V_{stack}$ within reasonable ranges for the experiments in the Mojave Desert show that the shape of $T_Q$ as a function of $\tau$ is sensitive almost exclusively to $Q_{ap}$, which allows $Q_{ap}$ to be determined from the amplitude trend stacks. As for spherical divergence, $Q_{ap}$ heterogeneities can produce differences only in the overall biases of the trend stacks, which will vary far more slowly in $\tau$ than the trends due to reflections.

The effects of transmission through intermediate interfaces at different angles of incidence were shown to have an order of magnitude larger effect on AVO trends than the properties of objective reflectors by Gassaway (1984). An intermediate layer with strong offset-dependent transmission effects produces a bias on the amplitude trend stack of all reflections below it. This bias allows such intervals to be identified on AVO trend stacks from breaks in the lateral continuity of the trends of the underlying structures.

**Reflector effects.**—A reflector including a contrast in anisotropy can interfere with the interpretation of amplitude trends. Wright (1984) presented physical and numerical models indicating that the inclusion of anisotropy can reverse an AVO trend. However, Daley and Hron (1977) gave calculations indicating that, for a physical-property contrast including anisotropy, increasing the anisotropy up to a maximum of 20 percent mainly affects the position of the critical angle. While the critical angle may change by up to 15 degrees, reflected amplitudes will change by less than 25 percent within the precritical range. Such variations should not affect the interpretation of Poisson's ratio changes from the direction of precritical amplitude trends.

Fisher and Gardner (1984) have suggested that reflector layering, as well, can reverse the amplitude trends expected from simply considering the properties of a step discontinuity. To test this, I calculated the bounds on thin-layer spectral interference for a variety of canonical physical contrasts. The bounds are independent of the thickness of the layer and of the frequency of the waves. These bounds, for an isolated thin layer, were compared to the reflection coefficient of a step discontinuity having the same properties. Comparisons were made for contrasts of up to 10 percent for the ten cases of $\pm \varrho, \pm \alpha, \pm \beta$, and $\pm \alpha, \pm \beta$ ($\alpha$ and $\beta$ are the compressional and shear velocities). In all cases the envelope of the thin-layer spectral interference exhibits the same trend as the coefficient for a step discontinuity. Severely band-limited data may show a different trend over small ranges of offset, but combining a reasonable range with some frequency bandwidth should remove that effect.

Reflector curvature can have serious effects on amplitude trends (Shuey et al., 1984). For a deep reflector, a curvature with a large enough radius to cause a far-offset amplitude increase should be observable as such on a stacked section. The reflector would have several kilometers of relief. On the other hand, a deep reflector with a radius of curvature so small that its depth variations cannot be observed on a stack can attenuate only far-offset amplitudes. Increasing AVO trends would still be valid.

**RESULTS**

This section demonstrates the application of the above AVO analysis principles to deep-crustal data sets from the eastern and western Mojave Desert.
To stack the 1985 Calcrust gathers from line WM-1, each trace was first despiked and labeled with information on its source and receiver positions and offset. Then trace equalization, on the 70th percentile of the amplitude levels, and the spherical divergence correction \( G \), were applied as discussed above. Band-pass filtering between 15 and 25 Hz was used to attenuate air waves and ground roll. Mutes were applied to remove source-generated noise where band-pass filtering failed to do so.

Stacking velocities were evaluated by calculating constant-velocity stacks, at 0.1 km/s intervals, from all of the gathers. The constant-velocity stacks were weighted by the percentage of hyperbolic signal, calculated using the method of Harlan et al. (1984). These stacks gave clear indications of the velocities that resulted in the best images of many different reflections throughout the stack, allowing velocity functions to be picked for 19 different midpoint intervals. Velocities were picked only to yield a well-focused stack. While the resulting interval velocities were often unreasonable, this simply reflects the degree of lateral heterogeneity present below Ward Valley.

Examples of prestack deep reflections before and after filtering are shown in Figure 2. They are corrected for geometric spreading and normal movement (NMO) using the picked stacking velocities. Lateral heterogeneity makes the
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Fig. 4. Stack of deep reflections of the merged roll-along survey and far-offset stationary arrays from Calcrust line WM-1 in Ward Valley, after weighting by the proportion of hyperbolic signal in the CMP gathers, in the manner of Harlan et al. (1984). It incorporates 8.33 km of line composed of 219 CMP gathers spaced at 38 m between VP 686 and VP 1019. The north end of the stack, on the right, is 14.7 km from the northern terminus (34°26'58" North, 115°02'13" West) of line WM-1. The strongest reflections are from the top of the basal-crustal zone at 8.4 s and from the Moho at 9.5 s.

Fig. 5. Linear AVO trend stack of Calcrust line WM-1, for each point of the stack of Figure 4. Clip at 0.0001 counts/m. Dark areas indicate amplitude increases with offset. The arrows indicate the trends of the example basal-crustal reflections of Figure 3 at midpoints 1470 (right) and 1959 (left).
reflections in Figure 2b appear to be overcorrected, even though the NMO correction gives the best-focused stack at that midpoint. The indicated events at both midpoints are strong at some offsets and weak but visible at others. Their amplitude trends are shown in Figure 3. Gross trends can be discerned in the AVO of these basal-crustal reflections.

After recalculating the spherical divergence corrections with the picked stacking velocities, the gathers were stacked into a deep crustal time section. The stack was then weighted with the calculated hyperbolic signal percentages, producing the image in Figure 4. This stack, representing a section of the deep crust from about 12 to 28 km depth, and 8.3 km in width, shows north-dipping midcrustal reflections and relatively flat basal-crustal reflections near the Moho. The strongest deep reflections, that best fit the assumptions of the stacking process, stand out as “bright spots.” Portions of the midcrustal events stand out clearly, as do most of the basal-crustal events. The strongest, most continuous reflection is, in fact, the top of the basal-crustal zone, at about 23 to 24.5 km depth.

The next step in this AVO analysis is to plot the amplitude trend stack, in the manner of Long and Richgels (1985). The stacking algorithm collects statistics on the rms amplitude of a 0.08 s window of each offset trace centered about the NMO time, at each intercept time point of each stacked trace. These statistics are reduced by simple linear regression to yield the linear trend of amplitude with offset at each intercept time point. The correlations of each regression are also found.

The amplitude trend stack and an amplitude trend correlation stack from Calcrust line WM-1 are plotted separately. Figure 5 gives the trend stack, while Figure 6 gives the correlation stack. Comparing these two derived sections with the weighted stack of Figure 4 shows how to avoid some of the pitfalls of the interfering factors discussed in the section on amplitude effects. The most reliable information on AVO variations at the reflector will come from events that (1) are strong on the weighted stack, (2) show a high degree of amplitude linearity with offset on the correlation stack, (3) deviate significantly and sharply from the background slopes on the trend stack, and (4) show some degree of lateral continuity of all these properties on all three stacks.

The most prominent events of Figures 4 and 5, which

<table>
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<th>Layer</th>
<th>$\sigma$</th>
<th>$V_P$ (km/s)</th>
<th>$V_S/V_P$ (%)</th>
<th>$\rho$ (g/cm$^3$)</th>
<th>$\delta_\lambda$ (%)</th>
<th>$\delta_\mu$ (%)</th>
<th>AVO trend</th>
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<td>5.8</td>
<td>68 —versus— 2.72</td>
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<td>-7.0</td>
<td>+103</td>
<td>Increase</td>
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<td>6.3</td>
<td>86</td>
<td>2.92</td>
<td>+9.8</td>
<td>+64</td>
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<td>6.3</td>
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<td>2.92</td>
<td>+42</td>
<td>-8.5</td>
<td>Increase</td>
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<td>-8.5</td>
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<tr>
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<td>77 —versus— 2.92</td>
<td></td>
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<td>-36</td>
<td>Increase</td>
</tr>
<tr>
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<td>7.9</td>
<td>46</td>
<td>3.32</td>
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<td>-36</td>
<td>Increase</td>
</tr>
</tbody>
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Fig. 6. Correlations $r$ of the linear regressions on AVO for each point of the stack in Figure 5. Clip at $r = 0.6$. Dark areas indicate positive $r$.  

Table 1. AVO trends of hypothetical deep interfaces.
appear to represent reflector variations in the deep crust, are the basal-crustal reflections between 8.5 and 9.5 s. The reflections and their derived offset trends are reasonably flat and continuous over the 8.3 km length of the stack. Three examples of the AVO trends, at the arrows in Figure 5, were shown in Figures 2 and 3. At the northern example midpoint, CMP 1470, the 8.15 s reflection just above the top of the basal zone has a decreasing trend on Figure 3a. This clearly comes from higher near-offset reflection amplitudes that can be seen in Figure 2a. In Figure 2a the traces were not band-pass filtered. In Figure 3b the 8.65 s reflection shows no trend, in contrast to the 8.15 s reflection. The band-pass filtering attenuated air-wave energy at the near offsets, giving the 8.65 s event a slight increasing AVO trend in the trend stack, Figure 5 (lower arrow on right). Similar relations are shown on Figures 2b and 3c for the 8.7 s reflection from the top of the basal-crustal zone at the southern example midpoint, CMP 1959. The high amplitudes at 9 km offsets in Figure 3c come from reflection wavelets clearly visible in Figure 2b.

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**Fig. 7.** Shot gathers from COCORP Mojave line 3 in the Rand Mountains, southern California. As for the gathers in Figure 2, both have had quantile trace equalization and a spherical divergence correction applied. A strong midcrustal reflection arrives at 5-6 s on both shot gathers. A is from the vibrator point at VP 527. B is from VP 550. The arrow indicates a strong near-surface triplication and reflection. On both gathers the receivers are southwest of the vibrator point.

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**Fig. 8.** AVO trends of the 5-6 s reflection of Figure 7, with amplitude plotted on a log scale. A is from VP 527; B is from VP 550. The shallow triplication has depleted the far-offset amplitude of the reflection.
By themselves, the trends in Figures 2 and 3 may not be convincing. However, in the context of the lateral continuity of the basal reflector on the trend stack (Figure 5), the trends may be valid. The trend stack shows reflections having amplitudes which increase or decrease as much as 20 percent or more over a 12 km offset range. These strong trends are apparent despite interference from the effects of recording, surface variations, propagation, or reflector geometry. According to Shuey's (1985) relations and the Zoeppritz calculations discussed in the section above on amplitude analysis, such AVO trends may well arise at reflectors with at least a 10 percent change in Poisson's ratio.

Analysis of moveout at long offsets suggests that the reflection between 8.5 and 9 s arises at the top of a 6.6 km/s basal-crustal zone below a 5.8 km/s middle crust. The events near 9.5 s likely arise at the Moho. Both of these deep interfaces therefore represent increases in compressional velocity, meaning that the sum of $\lambda$ and $\mu$ must increase. The increasing AVO trends imply that the signs of the $\lambda$ and $\mu$ changes oppose one another at these interfaces. Deep shear-velocity structure is not well known in the Mojave. Table 1 summarizes hypothetical properties and AVO trends of the basal-crustal and Moho reflectors. For the basal crust to provide an increasing AVO trend against the midcrust, it must have a higher Poisson's ratio. A $\sigma$ much lower than that of the midcrust could produce the increasing AVO trend, but only if the shear velocity becomes unreasonably high. Table 1 also suggests that the Moho reflector includes an increase in Poisson's ratio. Such a high $\sigma$ may be confined to a thin layer at the Moho. Detailed shear-velocity information could test whether these reflectors actually incorporate the decreases in shear velocity implied in Table 1.

Another prominent reflection is seen at 6 s near the center of the stack (Figures 4 and 5). One of the strongest reflections after weighting by the hyperbolic signal content, it shows a large positive amplitude trend. Below it, the trends of the underlying reflections have been muted, and in some cases reversed. This is an example of how transmission through a strong reflector can affect the apparent trends of the underlying reflections, as suggested by Gassaway (1984). Since the stack was constructed from CMP gathers, the strong 6 s event adds a negative bias only to the events immediately below it. This is the effect that disrupts the continuity of the basal-crustal reflections. The negative bias may be useful, since it shows that the reflector produces increased reflection and decreased transmission amplitudes at larger precritical incidence angles.

**Western Mojave**

From the extensive COCORP data set in the western Mojave Desert, Cheadle et al. (1985, 1986) identified several shallowly dipping, regional reflections. They interpreted these reflections as arising from horizontal shear zones, with several models available for their ages and sense of motion. The data from line 3 are considered here. This line contains the best-imaged reflections and was the basis for the geometric interpretation of Cheadle et al. (1985).

The data were recorded using off-end spreads with 96 offsets from 0.4 to 10 km. While the offsets are unreversed, the coverage of the offset range is continuous, which allows the effects of at least the midcrustal reflectors to be analyzed. Figure 7 shows two shot gathers from within the Rand Mountains on line 3. Source-generated noise does not interfere with the visibility of a prominent reflection at 5–6 s, so no band-pass filtering was performed.

To speed up processing, I did not sort the shot gathers into CMP gathers before stacking. Instead, common-shotpoint gathers were stacked directly. While the data set retains the advantages of multiplicity, this analysis of shot gathers will force additional assumptions about the lateral homogeneity of the reflectors. A comparison of the shot-gather stack derived here (not shown) with the full midpoint stack of Cheadle et al. (1986) showed that the differences are few enough to suggest that lateral reflector homogeneity on the scale of a few kilometers is not unreasonable.

For an initial analysis, a small area of line 3 directly beneath the Rand Mountains was selected for an evaluation of stacking velocities. This was done with constant-velocity stacks at 0.1 km/s intervals. I selected velocities to emphasize the strongest flat-lying reflections. The trace-amplitude equalization was set to the 70th percentile, and the spherical divergence correction and air and surface wave mutes were applied as before. The shot-gather stack (not shown), weighted by the hyperbolic signal content, shows an impressively strong reflection between 5 and 6 s, at about 16 km depth. It also has indications of a reflection from the Moho between 10 and 11 s, at about 32 km depth. The lack of offsets of less than 400 m forced the direct-wave mute to eliminate all data in the stack from less than 0.9 s.

During stacking, AVO trends were tracked within 0.2 s windows surrounding the stacking hyperbolas. The larger window compensates for the multicycle character of prominent reflections in this data set. Two examples of AVO trends from the 5–6 s reflection in Figure 7 are shown in Figure 8. For this strong event, definite differences in AVO trend can be seen between the two vibrator points. The amplitude trend stack is shown in Figure 9. Several features are apparent. First, the trends exhibit a broad negative bias across the section that increases strongly at shallower times. This feature is due to the apparent attenuation of waves during propagation. Second, the trends begin to lose coherency below 6 or 7 s. It seems that, at greater times, no reflections were recorded that showed enough amplitude change, within the 10 km offset range of this survey, to stand out from the background noise level. The range of incidence angles within the deeper parts of the crust is too small to permit analysis of many reflectors, such as the Moho, in the western Mojave.

The reflection between 5 and 6 s shows clear amplitude trends. A trend correlation stack (not shown) was also calculated. The 5–6 s event shows good trend correlations. The trend, however, inverts between the center and the southwest side of the section (Figure 8; Figure 9, circles). In the trend stack, the large positive trend on the southwest side between 1 and 2 s is a hint that the problem lies along the propagation path. Specifically, the problem involves lateral heterogeneities at the interface between the alluvium and the basement, which are the strongest in the entire crust.

In Figure 7a, the 5–6 s reflection has high amplitudes at a large range of offsets, indicating that energy is penetrating down to it at the full range of incidence angles. The refracted
arrivals at the farthest offsets also have high amplitudes that show the same effect. On the other hand, the gather to the southwest (Figure 7b) shows a much weaker reflection, visible only at inner offsets. Refracted arrivals also change character, with their amplitude decreasing more at farther offsets.

The reason for these changes lies with a strong triplication at 1 s (Figure 7b, arrow) apparent on the gather to the southwest. The interface between alluvium and granitic basement has changed character such that it reflects almost all the energy incident upon it back toward the surface. The deep reflection is depleted at long offset because very little energy penetrates into the basement at larger incidence angles. This effect reverses the derived amplitude trend (Figure 8). Fortunately, as can be seen at the left side of Figure 9, the strong 1 s reflector biases the trends of reflections to 7 s beneath it, making its effect easy to identify on the trend stack.

With this analysis in mind, the entire line 3 data set was stacked in the same manner as for the section under the Rand Mountains. This shot-gather stack is not shown, but all of the reflections identified by Cheadle et al. (1985, 1986) on COCORP’s midpoint stacks could be identified.

To calculate an AVO trend stack for all of line 3, the same procedures were used as for Figure 9. In addition, I made a correction for apparent attenuation to remove the broad negative bias at shallow times. The traces of an original trend stack (not shown) were averaged over all of line 3, producing the trend profile of Figure 10a. Aside from high-amplitude glitches caused by summing in the shallow AVO trends of a few gathers contaminated by ground roll or first arrivals, the profile shows the gradual decline of the influence of effective attenuation with depth. This profile can be modeled, using equation (1), producing the profile in Figure 10b. The shape of the curve between 1 and 3 s is so strongly affected by the value of $Q_{ap}$ that the average $Q_{ap}$ over line 3 is constrained to be between 10 and 30. Such a low value is not surprising. The average trend, as modeled, is affected mostly by the shallowest section of the crust, where highly heterogeneous alluvium causes extensive scattering and mode conversion of high-frequency reflections.

This model trend was fit to the average AVO trend of the entire data set. Then I subtracted it from each trace of the trend stack to yield the trend stack in Figure 11, which is thus corrected for the average effective attenuation. The section shows dark areas, where $Q_{ap}$ is greater than 20 near the surface, and light areas where it is less than 20.

Overall, some major midcrustal reflections show increasing AVO trends, including a southwest-dipping event (Figure 11, arrow), and the flat event below it at 5–6 s, at ~16 km depth. These trends are reversed in a few places by strong near-surface reflectors. There are also hints near the center of the section that the Moho reflection at 10 s may have a positive trend. The reflection is not strong enough over the whole section, however, to yield a definite trend. Without corroborating information indicating the contrast in compressional velocity at these reflectors, the signs of their Poisson's ratio contrasts cannot be determined. However, Poisson’s ratio variations of at least 10 percent must occur at these prominent midcrustal reflectors.

CONCLUSIONS

Seismic reflection surveys in many areas have brought to light the existence of previously unsuspected structures that have led to new models of crustal phenomena. To date these surveys have employed techniques oriented toward defining the geometric structure of these reflectors. While knowledge
of reflector geometry allows hypotheses on the geologic nature of the reflectors to be proposed, in many cases it has not provided tests for competing models.

The need for testing models can be addressed through expanding the analysis of seismic data to include prestack information. Approximations to the Zoeppritz equations (Shuey, 1985; Wu and Aki, 1985) link the presence of AVO trends to variations in Poisson's ratio. If the signs of the density and rigidity variations agree with the sign of the variation in Lamé's parameter, then amplitude will decrease with offset. If the signs are opposite, amplitude increases with offset. If the sign of the compressional velocity variation is also known, the sense of variation in Poisson's ratio can be found. These relationships were verified by calculating solutions of the Zoeppritz equations over a set of canonical reflector models. This simple method may yield fundamental constraints on the Poisson's ratio variations of deep reflectors.

Looking only for the direction of linear AVO trends, this analysis was applied to two deep-crustal data sets from the Mojave Desert, southern California. Calcrust line WM-1, through Ward Valley in the eastern Mojave, and COCORP line 3, in the Rand Mountains of the western Mojave, imaged deep structures about which very little is known. Reflector density or modulus variations cannot be individually identified from AVO trends. However, both data sets contain reflections with strong AVO trends that can be explained by reflectivity variations as a function of incidence angle.

Gross linear AVO trends were found by least-squares from prominent, stackable reflections. The multioffset records were trace-equalized using an assumption of constant noise at large time and corrected for geometric spreading and NMO using velocities picked from constant-velocity stacks. AVO trends were calculated for each point of the stacked sections and presented as trend stacks in the manner of Long and Richgels (1985). Both increasing and decreasing AVO

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**Fig. 10.** (a) The sum of the raw linear AVO trends as a function of intercept time from all 804 gathers of COCORP Mojave line 3. The trend is truncated by the direct and refracted arrival mutes at less than 0.9 s. The high-amplitude spikes result from including the AVO trends of shot gathers contaminated by ground roll or first arrivals at shallow times. Both curves are clipped at 3.14 counts/m, with the dark areas indicating increases of amplitude with offset. (b) Model trend calculated from equation (1) for an apparent Q of 20, with $A_0 = 4.8 \times 10^3$ counts, $h = 10 \text{ km}$, $f = 20 \text{ Hz}$, and the picked stacking velocities from vibrator point 500, to match the above average trend. The fit of these two curves at times of less than 3 s indicates that the average $Q_{ap}$ over line 3 is 20 ± 10.

**Fig. 11.** Linear AVO trend stack of all of COCORP Mojave line 3, from vibrator point 1 (right, northeast) to 868 (left, southwest). The trend due to an apparent Q of 20, with the other parameters as shown in Figure 10b, has been subtracted. Dark areas indicate amplitude increases with offset. The dashed lines denote the area shown in Figure 9. The arrow indicates a prominent southwest-dipping reflection between about 12 and 16 km depth that exhibits an increasing AVO trend, suggesting Poisson's ratio variation of at least 10 percent.
trends were identified on multioffset gathers and trend stacks of the Calcrust data set. Interpretation of the COCORP data set had to be limited to the increasing AVO trends. Long geophone arrays used there had enough directivity to mimic a decreasing trend.

The trend stack from COCORP Mojave line 3 was used to estimate total apparent attenuation, including both scattering and intrinsic attenuation. In the western Mojave the average attenuation $Q_{av}$ is 20 ± 10, over a distance of almost 90 km and within 3 km of the surface. This added a negative bias to the first 3 s of the trend stack, which was corrected.

In the Calcrust line WM-1 stacks, a transition from the 5.8 km/s midcrust to a 6.3 km/s basal-crustal zone at ~23 km depth produces a strong reflection having a laterally continuous increasing AVO trend. Increasing trends also appear at the Moho reflector, at ~26 km. Both of these interfaces are increases in P-wave velocity. Therefore Poisson's ratio may increase by at least 10 percent at the basal-crustal zone and at the Moho. Less continuous reflections in the midcrust show strong AVO trends, also suggesting large variations in Poisson's ratio.

Strong reflections in the COCORP line 3 stacks with increasing AVO trends suggest the presence of Poisson's ratio variations of greater than 10 percent below the southwestern flank of the Rand Mountains. These reflections are coincident with regional low-angle fault structures interpreted by Cheadle et al. (1985, 1986) from the same data set. In both the eastern and western Mojave Desert, reflection AVO trends suggest that prominent crustal structures include large variations in Poisson's ratio.

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