The Moho in the northern Basin and Range province, Nevada, along the COCORP 40°N seismic-reflection transect

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ABSTRACT

COCORP seismic-reflection profiles across Nevada at about 40°N image a prominent, essentially continuous band of reflectors at a two-way traveltime of 9 to 11 s. The approximate correspondence of this reflection time with estimates of the two-way traveltime to the Moho in this area provided by seismic-refraction data suggests that the prominent reflections are from the Moho. The relief on these reflectors (the "reflection Moho") beneath Nevada, across a latitudinal transect of >450 km, is only 2.2 s (~7 km) after correction for basin effects, or only ~50% of the Moho topography previously inferred from refraction data. Observed variations in traveltime to the reflection Moho are gradual, with no evidence for major offsets. The reflection Moho in this part of the Basin and Range province is interpreted as the base of a complex group of reflections. A striking aspect of the reflection data is the resolution of this group of reflections, in many places, into 2 distinct reflections as much as 1.2 s (~4 km) apart, which elsewhere merge into a single reflection. The reflection Moho appears to be continuous beneath terranes that experienced very different tectonic histories in the Paleozoic and Mesozoic and apparently truncates crustal reflectors in some areas. The reflection Moho in its present configuration thus appears to be a young feature in the Basin and Range province, and it may be related to Cenozoic magmatism and extension. In some areas of rifting and volcanism outside the Basin and Range province, the observed reflection Moho closely resembles that seen in the Basin and Range province. In other areas with other tectonic settings, including the Sierra Nevada and the Colorado Plateau, the reflection Moho has variable seismic character and may represent different kinds of geologic boundaries from that of the Basin and Range province.

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INTRODUCTION

COCORP (Consortium for Continental Reflection Profiling) has collected near-vertical seismic-reflection data along a 1,000-km transect in the western United States at about 40°N, starting in the Sierra Nevada in California (Nelson and others, in press), continuing across the width of the Basin and Range province (Hauge and others, 1984; Hauser and others, 1984; Potter and others, 1984; Oliver and others, 1985; Lemiszki and others, 1985; Allmendinger and others, 1983) onto the Colorado Plateau in Utah (H. G. Farmer and others, unpub. data) (Fig. 1). This data set provides a detailed look at the whole crust and Moho across an entire, active extensional province and its margins. This paper reports on observations of reflections from ~9 to 11 s two-way traveltime (TWTT) (times given with respect to a datum 1 km above sea level) termed here the "reflection Moho." Other major findings of the 40°N transect are discussed by Allmendinger and others (in press). These results include the distinction of the different seismic-reflection characteristics of the Sierra Nevada, the Basin and Range province, and the Colorado Plateau; the remarkably uniform apparent crustal thickness of the Basin and Range province and the pronounced horizontal fabric of much of its lower crust; and the variable fault geometries and basin morphologies in the upper crust. The basin-range segment of the COCORP 40°N transect includes Nevada lines 1 to 8 and Utah line 1. This paper reports mainly on Nevada lines 1 to 7 and refers also to Utah line 1, previously discussed by Allmendinger and others (1983). Nevada line 8 was recorded a year after lines 1 to 7 and will be discussed elsewhere (P. Knuepfer and others, unpub. data).

The high-amplitude reflections at 9 to 11 s across northern Nevada are believed to be images of the Moho for three reasons. First, these reflections mark the most impressive change in seismic character visible on the COCORP basin-range data, between an often highly reflective zone above and a normally nonreflective zone below (Figs. 2 and 3). Second, traveltimes of 9 to 11 s to the Moho are broadly consistent with previous refraction results on the depth to the Moho. Third, the interpretation of the deepest bright reflections as the Moho is consistent with interpretations proposed for other reflection data sets that are better constrained by associated refraction profiles. These points are elaborated in later sections.

Throughout this paper, the bright reflections at 9 to 11 s TWTT are referred to as the "reflection Moho." Although one might wish to avoid proliferation of terminology, the new term seems required because of the way in which the Moho is currently defined. The crust-mantle boundary, subsequently named for Mohorovičić, was originally detected as a very rapid increase in P-wave velocity (V_p) from 6.5 or 7.0 km/s to ~8 km/s (Mohorovičić, 1910). Steinhart (1967) defined the Moho as the depth at which V_p first increases rapidly or discontinuously to 7.6–8.6 km/s, or, in the absence of steep gradients, the level where V_p first exceeds 7.6 km/s. All such definitions, however, are based on average rock properties detected by relatively low-frequency refracted waves and are not applicable in practice to near-vertical reflection experiments, because such experiments do

Geological Society of America Bulletin, v. 97, p. 603–618, 11 figs., May 1986.

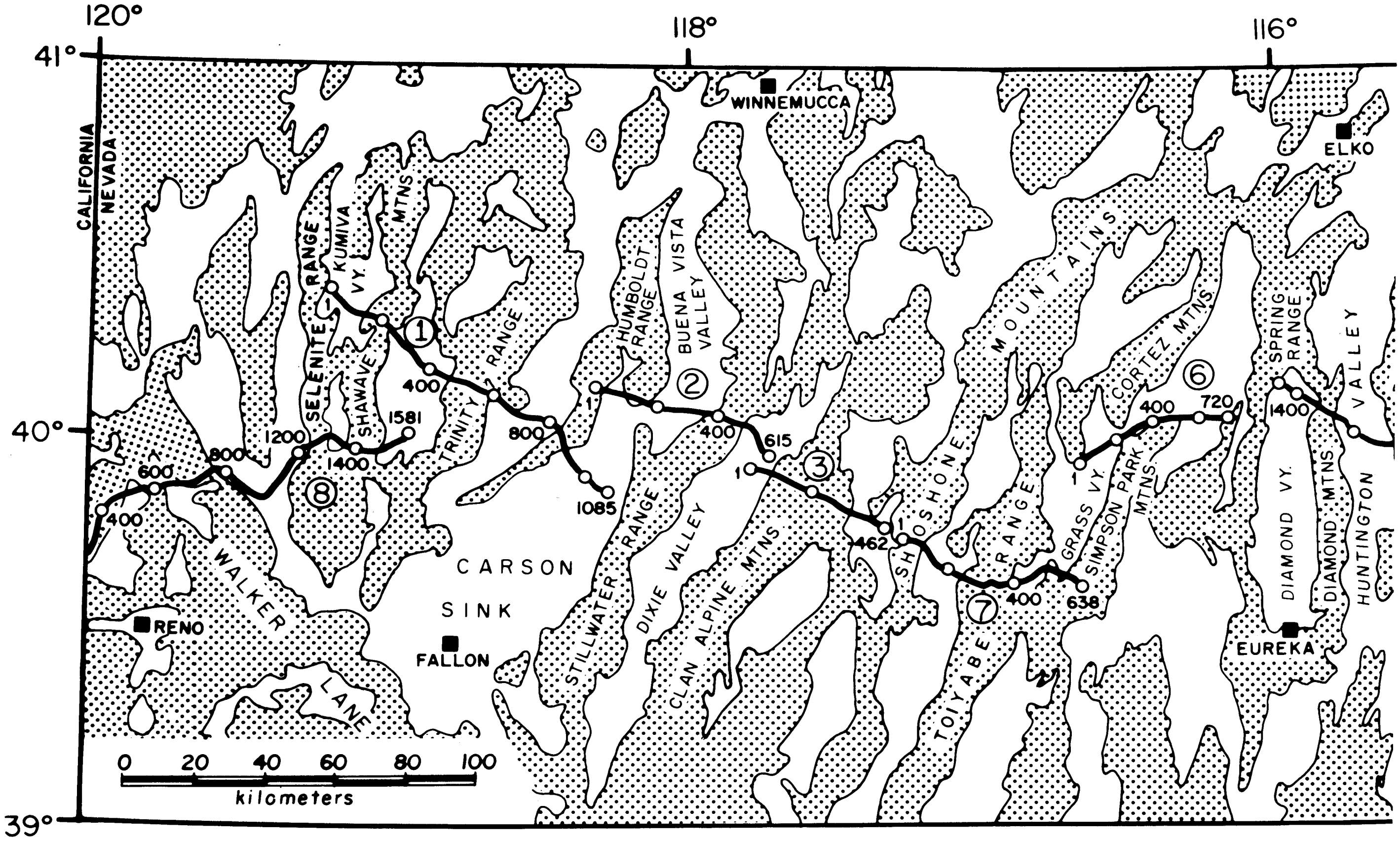


Figure 1. Line location map showing basin-range segment of the COCORP 40°N transect. COCORP lines from west to east are Nevada lines 8, 1, 2, 3, 7, 6, 4, and 5 and Utah line 1. Inset shows additional COCORP lines in California and the Colorado Plateau. Numbers along COCORP lines are source-point (SP) numbers. SP spacing is 100 m (except line 6; SP spacing is 66 m).

not normally provide accurate velocities at Moho depths. The evidence of reflection profiles is that at the scale of vertical resolution of these studies (100–200 m) crustal velocity is not a smooth function with depth, but a highly irregular (reflective) function. Small volumes of crustal rock may have velocities normally characteristic of the mantle (>7.6 km/s). The potential detection of such rocks by reflection seismology must vitiate any attempt to define the Moho in terms of a unique velocity contrast. The concept of a velocity function that fluctuates rapidly while increasing from "crustal" to "mantle" velocities has led to the notion of the Moho as a transition zone (for example, Meissner, 1967; Davydova and others, 1972; Hale and Thompson, 1982) consisting of alternating lamellae of higher and lower velocity material, despite the original definition of the Moho as a boundary or discontinuity.

In order to capitalize on the unique resolving power of near-vertical reflection seismology, the "reflection Moho" is here defined on the basis of attributes detected by reflection profiling. In this paper, the reflection Moho is defined as the deepest, high-amplitude, laterally extensive reflection or group of reflections (in which case, the reflection Moho is taken as the base of the group of reflections) present at traveltimes (depths) approximately commensurate with other estimates of crustal thickness. The reflection Moho may commonly separate highly reflective lower crust from largely reflection-free upper mantle. This definition of the reflection Moho as a sharp boundary, rather than as a transition zone, is intended for operational convenience and is not intended to imply that the change from crust to mantle necessarily occurs at a single lithologic boundary rather than as a transition zone. The definition, in common with the definition of the refraction Moho, is easily applied in some areas but only with difficulty, if at all, in others. This definition, however, unlike the traditional definition of Moho, satisfies the requirement that it can be applied to the data of reflection seismology. The reflection Moho may also be the refraction Moho, but this equivalence cannot be proved without coincident refraction and reflection profiles and may not be proved with certainty even then. In such tests, the refraction Moho and the reflection Moho have been found to be coincident to within 0.2 s in the North Sea (Barton and others, 1984), and to within 1 s of the refraction Moho in the Eromanga basin, Australia (calculated from data in Finlayson and others, 1984).

PROCESSING THE COCORP NEVADA DATA FOR MOHO REFLECTIONS

Figure 2A is a summary line drawing of the major features of COCORP Nevada lines 1 through 7 for the lower part of the crust, and figure 2B is a diagram showing the interpreted positions of the reflection Moho (M) and a prominent overlying reflection (X). Figures 3A and 4 show portions of the actual data in the form of 100% (single-fold) common-midpoint (CMP) sections. The COCORP data are collected so as to provide nominal 4,800% (48-fold) CMP data; that is, 48 traces sharing a common source-receiver midpoint are normally summed together, with appropriate compensation for velocity and geometric effects, to produce a single

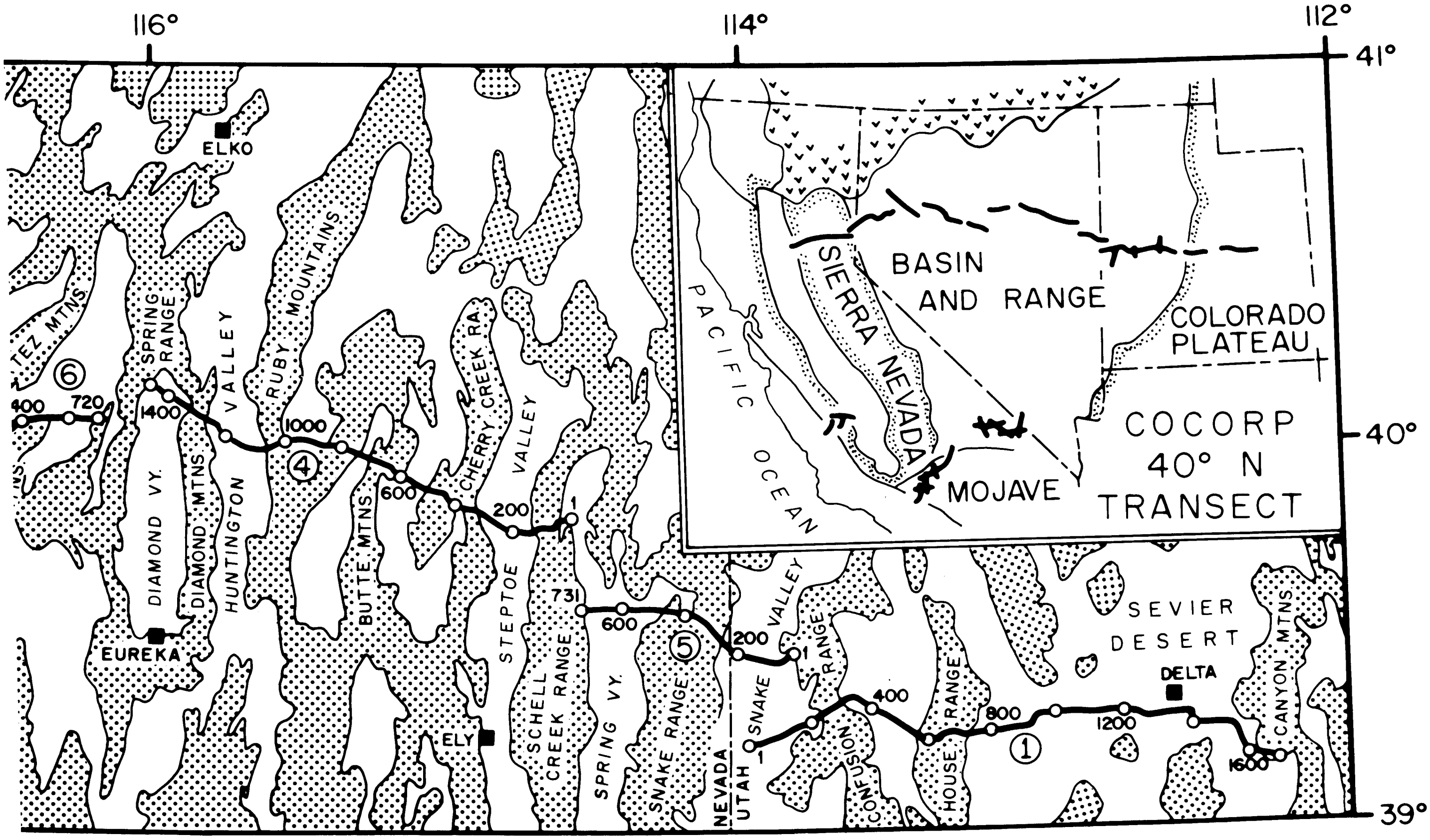


Figure 1. (*Continued*).

stacked trace. The stacking technique is generally an extremely effective method of improving signal-to-noise ratios, and it allows calculation of stacking velocities (for example, Sheriff and Geldart, 1983). For several reasons, however, single-fold rather than stacked data are emphasized in this paper. First, the Moho reflections in some places are already of high signal-to-noise ratio on individual traces. Second, the COCORP data quality is often quite variable between even closely spaced source points (SPs), due to factors such as different ambient noise or different source or receiver coupling (Fig. 5). It may be better in such cases to discard noisy data than to rely on stacking procedures to extract the signal from the noise. For the COCORP Nevada data, an improvement in signal-to-noise ratio, compared with the 48-fold CMP stack, was obtained by selecting only a single good trace from each CMP gather. Third, the basin-range structures in Nevada produce considerable shallow velocity heterogeneity along the COCORP lines which generally run transverse to strike. In these circumstances, single-fold sections show a simple pattern of pull-down beneath basins (Fig. 6B), whereas CMP stacking tends to smear out individual reflections into a complex pattern (Fig. 6C) (Peddy and others, 1985). These effects are considerable in the COCORP Nevada data, as shown in Figure 3, which compares the single-fold section for Nevada line 5 with the equivalent 48-fold section. For this study, therefore, all data are presented as 100%, that is, single-fold, sections; 48-fold CMP stacks for the entire COCORP 40°N transect are available from the Institute for the Study of the Continents.

The procedure followed was to visually identify SP gathers with good signal penetration and low ambient noise and to select from these gathers the best traces to produce a 100% section. The 100% sections shown consist of successive groups of 12 to 96 traces, each group selected from a single SP gather. On the sections presented in Figures 4 and 9 (see below), arrowheads at the base of the sections mark the joins between different SP gathers. Within an SP gather (between arrowheads), source-receiver offsets change by 100 m from trace to trace, but adjacent traces either side of an arrowhead are from different SP gathers and may have source-receiver offsets differing by as little as 100 m (SP spacing) or by as much as 9.5 km (spread length). The quality of the 100% data presented (Fig. 3A) might lead one to question the utility of CMP recording and stacking (Fig. 3B), but although only 2% of the 48-fold COCORP data has been used by this single-fold procedure, there is no way of predicting in advance where good data will be found. The field procedure thus cannot be readily simplified to produce only such data. Vibrating at every receiver point (thus building up 48-fold data with the COCORP 96-channel recording system) is also necessary to image the shallow structure along the seismic profile.

All of the 100% data have had velocity corrections (normal moveout) and elevation corrections (statics) applied and have been placed in correct geometric position (CMP sorting). Data were deconvolved to compress the source pulse and f-k filtered to remove surface waves. High-amplitude air waves were surgically muted. (See, for example, Sheriff and Geldart, 1983, for a discussion of seismic-reflection processing.) In addition, basin-stripping statics to correct for low-velocity basin fill were estimated from 48-fold CMP stacks and from the refracted arrivals routinely recorded as part of the COCORP data set and have been applied to the data displayed in this paper. The data displayed have a uniform datum level of 1 km above sea level, and basins deeper than this level have been replaced by material with a velocity of 4 km/s. Statics corrections of as much as 1.0 s were calculated and applied over the deepest basins, on the basis of the simplifying assumption of vertical raypaths.

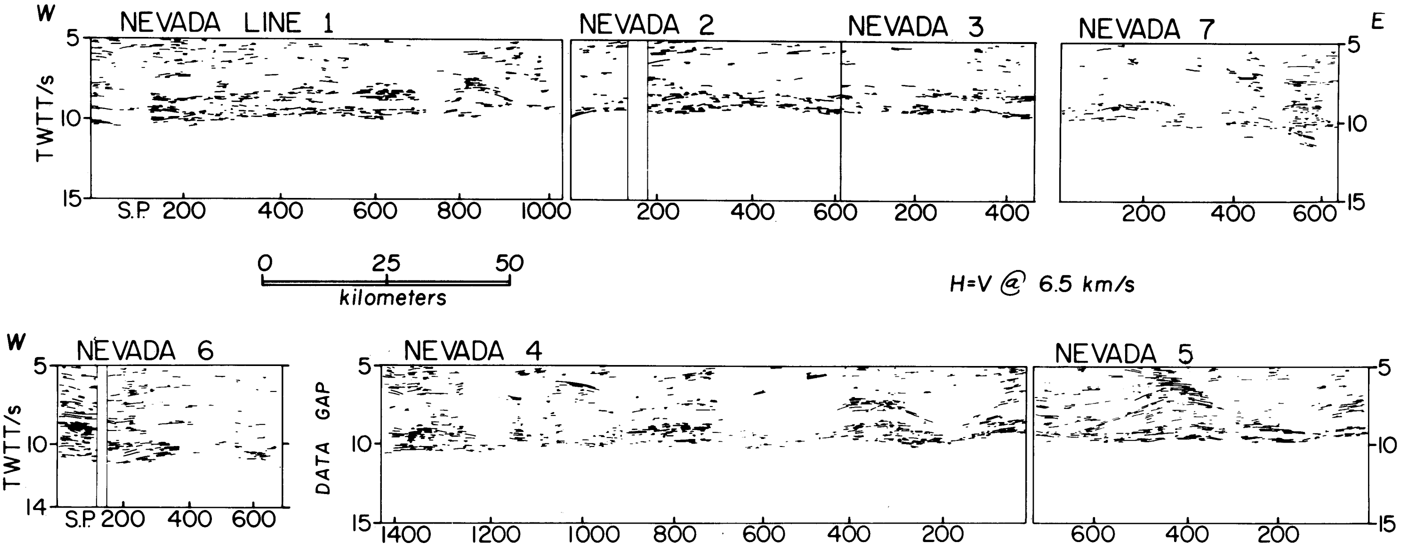


Figure 2A. Line drawing of 100% CMP section of COCORP Nevada lines 1 through 7 between 5 and 15 s two-way traveltime (TWTT). The lines are shown with correct spacing when projected north-south to latitude 40°N. Numbers beneath line drawings are SP numbers. Note considerable offsets in north-south direction between Nevada lines 1 and 2, lines 7 and 6, and lines 4 and 5 (Fig. 1).

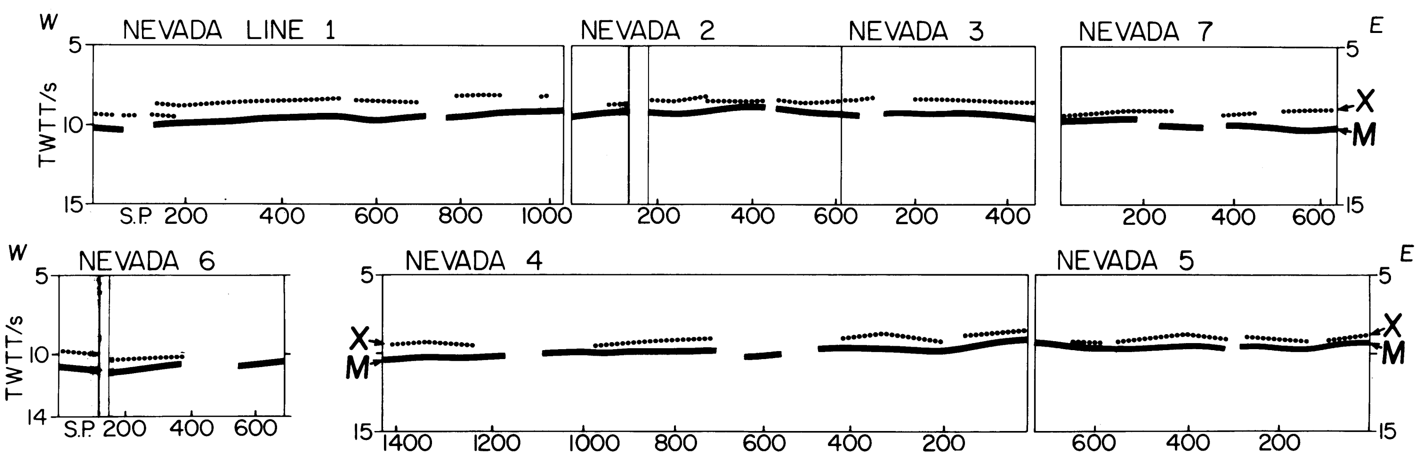


Figure 2B. Diagram illustrating interpreted position of reflections X and M on COCORP Nevada lines.

Some uncertainty remains in the velocity structure within the basins and in the detailed shape of some basins, but these errors are probably less than ~0.25 s. The possible existence and effects of deep lateral velocity variations are not addressed by this procedure, but percentage-wise, such variations are likely to be much smaller than shallow lateral velocity variations.

The reflection Moho was picked on the 100% sections on the basis of amplitude and continuity. The reflection Moho, marked M in Figure 4, typically shows a wavelet consisting of several cycles (0.05 to 0.2 s long), even after deconvolution. It is not always certain whether this complex waveform represents actual complexity of the Moho reflector or, alternatively, an incompletely compressed source pulse. Vibroseis (TM Conoco Inc.) data, such as these COCORP data, are normally assumed to be zero-phase (for example, Waters, 1981), and so the reflection Moho was picked as the maximum amplitude of the reflection wavelet M in Figure 4. The precise traveltime at which the reflection Moho is picked will depend on the source-wavelet characteristics (bandwidth and center frequency). Picking the Moho one cycle too high or too low would, however, give an error of only 0.05 s (for a 20 Hz wavelet), which is probably less than the errors involved in calculating the statics corrections required to reduce the data to a uniform datum level.

Across much of Nevada, the deepest continuous reflection is of higher amplitude than are superjacent reflections, and it is easily picked as the reflection Moho. In some portions of the COCORP data, however, identification of the reflection Moho requires more care. On the west end of line 4 (Fig. 4C), the deepest continuous reflections at ~10.3 s, picked as the reflection Moho, are of lower amplitude than is the overlying reflection at ~9.3 s. On the east end of line 7 (see Fig. 9A below), the reflection Moho is

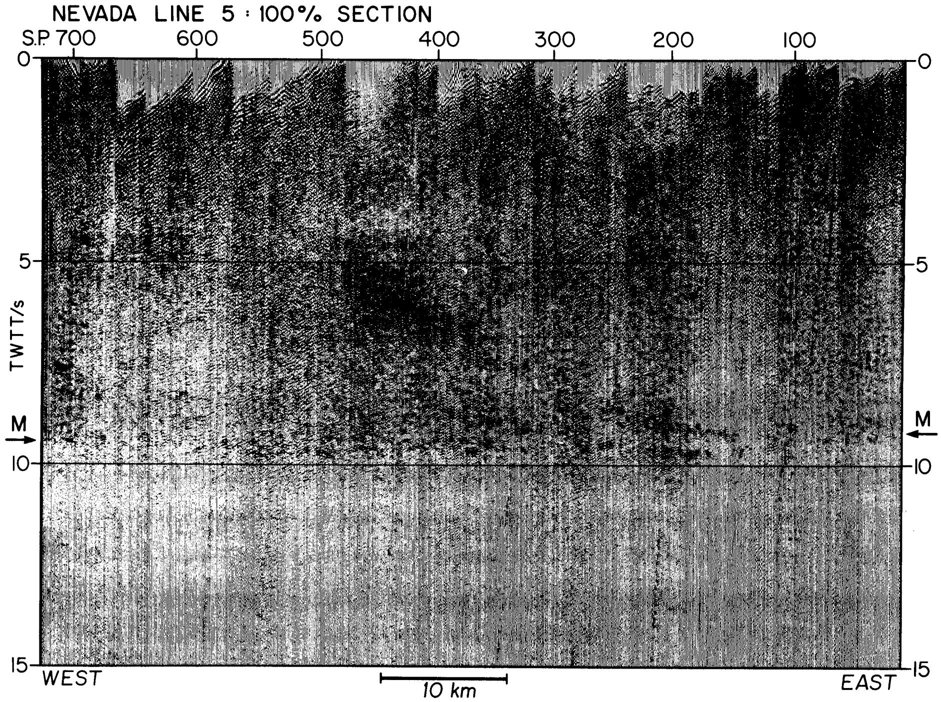


Figure 3A. 100% CMP section for Nevada line 5, 0 to 15 s. Note clarity of deep reflections at 9 to 10 s, and absence of images of shallow basins. Basin stripping statics have been applied. Datum is +1 km.

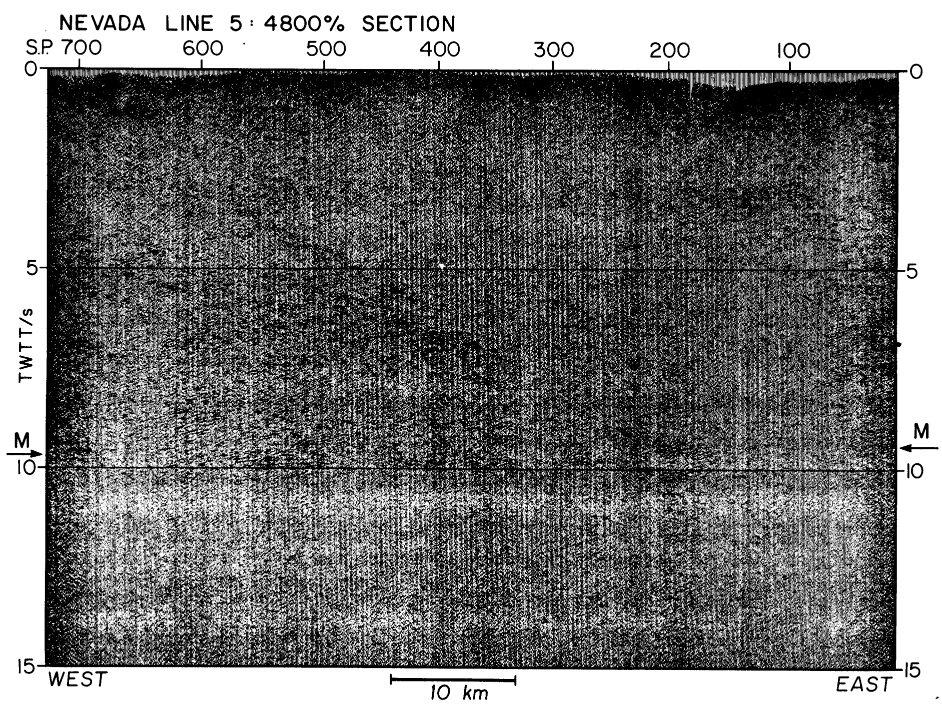


Figure 3B. 4,800% CMP section for Nevada line 5, 0 to 15 s. Note apparent complexity of deep reflections, but greater resolution in upper crust (0 to 5 s). No stripping statics have been applied. Datum is +1.46 km. Figure from E. Hauser and others, unpub. data.

picked at 10 s, even though reflections are observed to 11.5 s, because these later reflections are dipping and do not form a laterally continuous feature. Throughout this paper, traveltimes to Moho are used rather than depths to Moho, in order to avoid errors that may arise in depth calculations if the velocity structure of the basement is imprecisely known. All sections displayed in this paper have no vertical exaggeration for an assumed velocity of 6.5 km/s.

DESCRIPTION OF DATA: CONTINUITY, RELIEF, AND CHARACTER OF MOHO REFLECTIONS

The COCORP data show several features that are constant across the Nevada transect. First, the Moho reflectors are apparently continuous across all of the COCORP Nevada lines, with no evidence of abrupt offset (Fig. 2). Second, the reflection Moho undulates gently on a time section (Figs. 2 and 4). Third, where signal quality is good, there are frequently 2 distinct reflections from the deepest part of the crust (the lower being termed the "reflection Moho"), separated by as much as 1.2 s (~4 km) (Figs. 2, 3, 4, 7, and 8A). Fourth, in several places, dipping reflections are imaged that appear to truncate near the Moho (Fig. 9). Fifth, with few exceptions, the seismic sections are almost free of reflections beneath the reflection Moho (Figs. 2, 3, and 4). These observations are further detailed below.

In Figure 2, the reflection Moho is very clear across most of Nevada (~75% of the COCORP Nevada traverse), but it is not evident on some other parts of the COCORP Nevada data. The occasional absence of Moho reflections from the Nevada data is regarded as an artifact of locally higher noise levels, reduced signal penetration, or wavefront distortion, rather than as geologic reality. The degradation in data quality occurs commonly, but not ubiquitously, beneath basins where signal attenuation, surface-wave amplitudes, wavefront distortion, and cultural noise are generally higher than in the adjacent ranges. Figure 7 shows amplitude decay curves for portions of lines 1 and 5, illustrating failure to observe Moho reflections due to signal attenuation (Fig. 7A) and due to cultural noise (Fig. 7B). It is concluded that the apparent lack of Moho reflections on some parts of the Nevada data is not evidence against the presence of a continuous Moho reflector.

COCORP data from the Utah portion of the Basin and Range province show clear Moho reflections only locally (Allmendinger and others,

Figure 4. 100% CMP sections for portions of COCORP Nevada lines. Only 5 s of data displayed, from 7 to 12 s TWTT. SPs numbers given below sections refer to line location map, Figure 1. 100 SPs represent ~10 km. M: interpreted reflection Moho. X: reflection (or band of reflections) ~1 s above M. Arrowheads along base of sections divide groups of traces from separate SP gathers. Stars indicate positions of amplitude analyses shown in Figure 7.

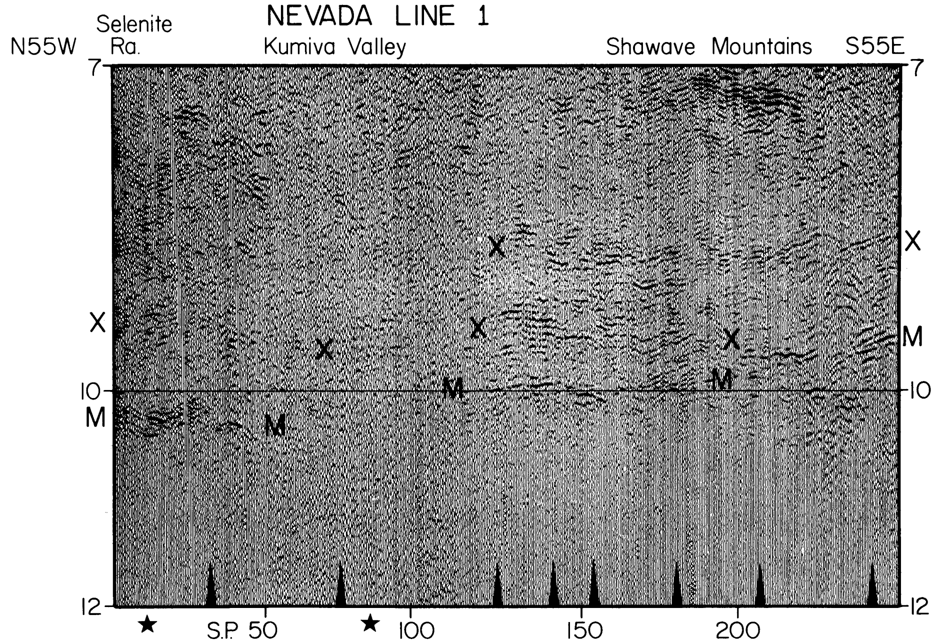


Figure 4A. NV 1, west end.

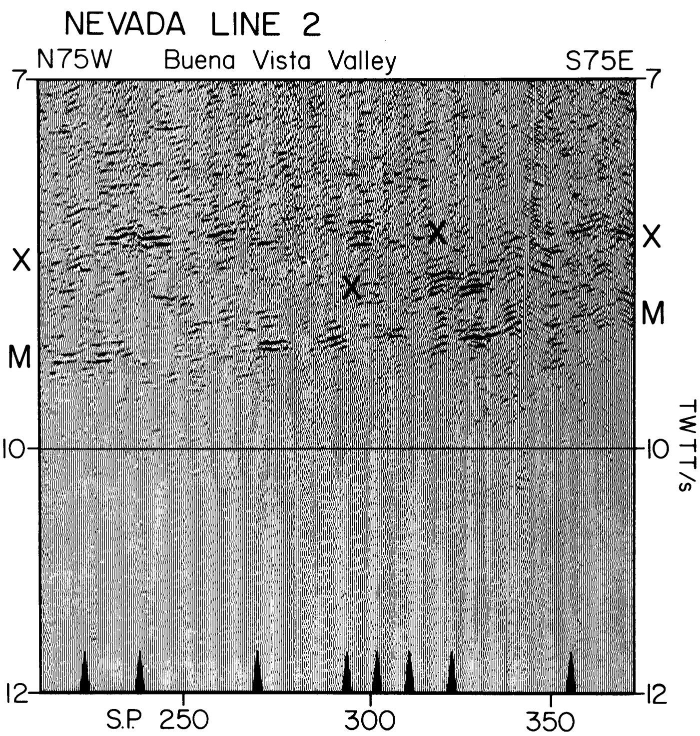


Figure 4B. NV 2, central part.

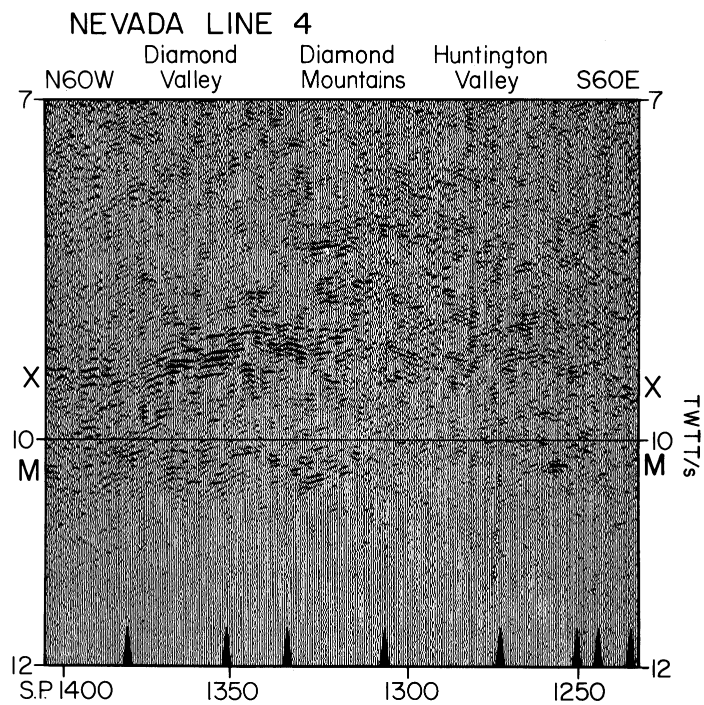


Figure 4C. NV 4, west end. Note amplitude of X greater than amplitude of M.

1983). It is uncertain at present whether the Moho reflector imaged in Nevada continues unchanged beneath Utah line 1 and was simply not well imaged by the COCORP Utah survey, or whether the Moho reflector is truly discontinuous. Even the best data from Utah line 1 (Fig. 8B), from the Sevier Desert, show a lower signal-to-noise ratio than do data from Nevada line 5 in Snake Valley (Fig. 8A). The presence of locally clear Moho reflections (~35% of Utah line 1), the evidence for generally higher noise levels (Fig. 8B), and the absence of Moho reflections from Utah line 1 data in Snake Valley only 20 km south of excellent Moho reflections on Nevada line 5 (Fig. 8A) all suggest that a Moho reflector may be present in the Utah Basin and Range. A second distinct reflection above the Moho was not observed in the Utah data.

The data in Figure 2 show apparent Moho offsets between Nevada lines 1 and 2 and also between lines 7 and 6, associated with north-

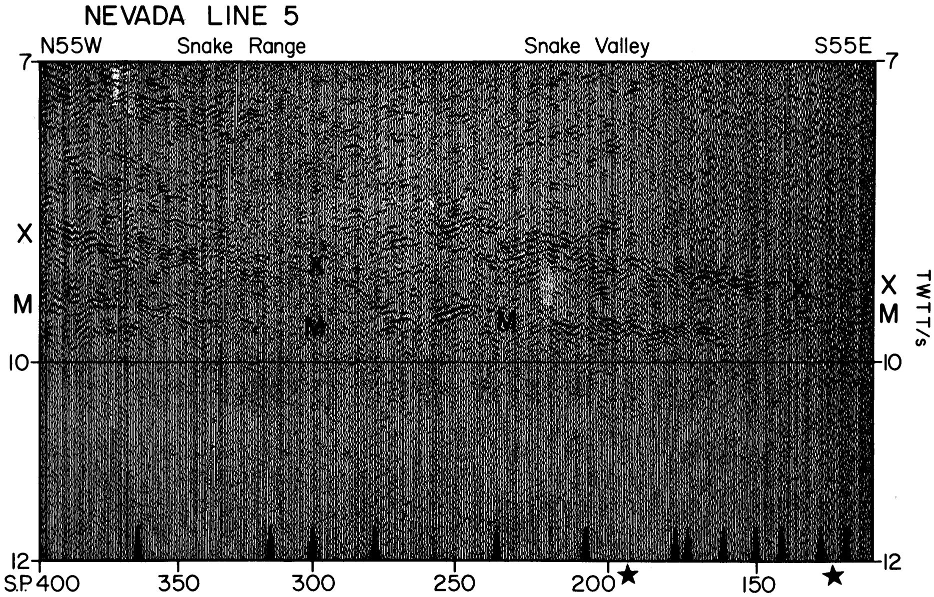


Figure 4D. NV 5, east end. Note amplitude of M greater than amplitude of X.

south offsets of the seismic lines (Fig. 1), thus providing evidence for Moho relief in the north-south direction. The data in Figure 4 also show minor apparent offsets of the reflection Moho. These minor offsets are almost certainly due to uncorrected near-surface statics, because they occur between traces from separate SP gathers (shown by arrowheads in Figs. 4 and 9) rather than between traces recorded from a single SP. No unequivocal, abrupt offsets on the Moho reflector were noted in this study. The theoretical horizontal resolution of the reflection method is ~2 km (Fresnel zone radius; for example, Sheriff and Geldart, 1983) at the Moho, much better than that of the refraction method as normally applied. The vertical resolution (ability to distinguish between closely spaced reflectors, rather than the minimum size of object that can be detected or the accuracy with which arrival times can be measured) is ~50 to 100 m (quarter-wavelength) under favorable conditions. Offsets of a reflector should produce overlapping diffractions rather than an apparently sharp offset of the reflections, and offsets of the Moho reflector smaller than the vertical resolution cannot be detected with this data set. Although the possibility of small (<0.5 km) offsets cannot be ruled out, certainly the data do not support the existence of vertical Moho offsets of the magnitude observed in the surface faulting, up to 5 km (Stewart, 1980). If, however, these faults dip less steeply at depth, or if the fault zones broaden with depth (Wernicke, 1985), the throw would decrease with depth, assuming constant offset along the faults. Of course, if the Moho reforms on timescales less than the time interval between major fault offsets, the lack of observable offsets need not imply that faults do not cut the Moho.

The data as presented show gently varying traveltimes, with a total relief of 2.2 s TWTT on the Moho across Nevada. If this traveltime difference represents thickness changes in a layer with a constant V_p of 6.5 km/s, the Moho relief would be ~7 km, with slopes of as much as ~8° locally. (Note that across distances of 100 km or more, Moho slopes are apparently only ~0.5° to 2.5°.) Alternatively, the variation in Moho TWTT could be entirely due to lateral velocity variations, with no physical relief on the Moho. A 32-km crust would need to have average velocity ranging from 5.7 to 7.1 km/s (a variation of 20%) and locally by as much as 0.5%/km laterally to satisfy the traveltime data. The answer probably is intermediate between these two possibilities, but these possibilities cannot be further tested with this data set due to the current lack of precise velocity estimates in the deep crust. Additional expanding spread surveys of the type carried out by COCORP in Utah (Liu and others, 1985) or other surveys recording both near- and far-offset data are needed to obtain accurate velocities for the deep crust. Gravity modeling may also provide some additional constraints in the future (Holbrook, 1984).

The reflection character of the Moho is important as a constraint on hypotheses of the nature of the Moho. The regional perspective provided by the COCORP Nevada transect allows the recognition that certain features of the Moho reflectors occur over an area 450 km wide, particularly that the Moho reflection (M) is very commonly overlain by a second reflection (X), with a vertical separation of as much as 1.2 s (~4 km) (Figs. 4, 7, and 8A). The two distinct reflections may have other, generally weaker, reflections between them (for example, Fig. 4A, SP 125–200), forming in this case a band of reflections at the base of the crust. The pair or band of reflections is visible on ~75% of the data where signal quality is sufficiently good to recognize clearly the reflection Moho, or ~50% of the entire Nevada data set. The separation of the two reflections (X and M) on the COCORP Nevada data varies laterally by as much as 0.1 s/km. The relative amplitude of the two reflections is variable; each is locally stronger than the other (compare Figs. 4C and 4D). The lateral persistence of the pair or band of events, their variable separation, and the frequently higher amplitude of the deeper reflection (Figs. 7A and 8A) show that the lower reflection is not a multiple of the upper reflection.

Where the COCORP data show a band of reflections at the base of the crust, this band might be modeled either as interfering reflections from a series of thin layers at the base of the crust (for example, Hale and Thompson, 1982) or as sideswipe from a single undulatory surface (for example, Behrens and others, 1972). The clear separation, however, of the two distinct reflection peaks, where present (Figs. 7 and 8A), and the partial 3-D control on X and M provided by the north-south offset of Nevada lines 6 and 7 (Fig. 2) both argue against sideswipe as a possible model and appear to require two separate reflectors. These two reflectors might represent the top and bottom of a transition zone at the base of the crust.

Another feature of interest on the COCORP basin-range data is the apparent truncation of dipping crustal, and rarely mantle, reflections by the reflection Moho. If these dipping reflections are from beneath the seismic lines and do not represent sideswipe, a hypothesis that can be properly tested only by the acquisition of cross-lines, then truncation relationships are present along the 40°N transect across Nevada on lines 8, 7, 4, and 5 (for example, Fig. 9) and on Utah line 1 (Allmendinger and others, 1983, their Fig. 2). The data shown in Figure 9 are unmigrated. Migration increases the angular discordance between the dipping reflections and the reflection Moho, thus accentuating the truncation relationship, and moves the dipping reflections in Figure 9 above both X and M. Similar truncation relationships have previously been noted on COCORP data from the Mojave Desert (Cheadle and others, in press, their Figs. 3 and 6). A single

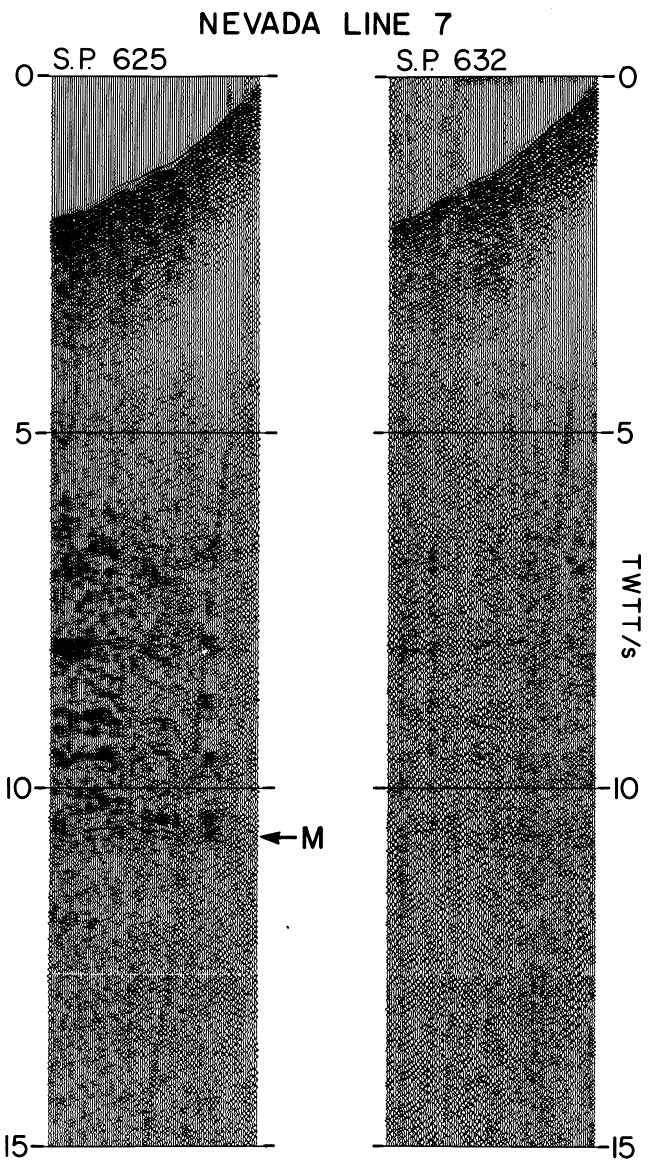


Figure 5. SP gathers for Nevada line 7 collected at an interval of 0.8 km. Spread length is 9.5 km, 15 s of data are shown. Note clear reflections on left-hand SP gather from 6.5 to 10.5 s and near-absence of reflections on right-hand SP gather.

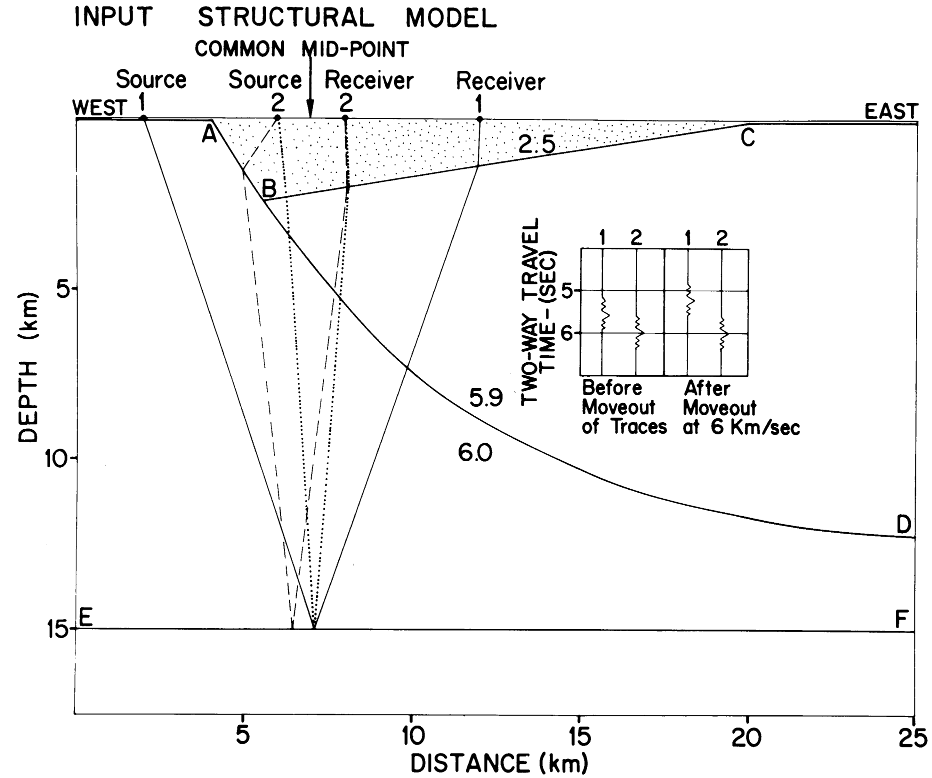


Figure 6A. Simple velocity model for a basin, showing source-receiver combinations for CMP shooting. Figure from Peddy and others (1985).

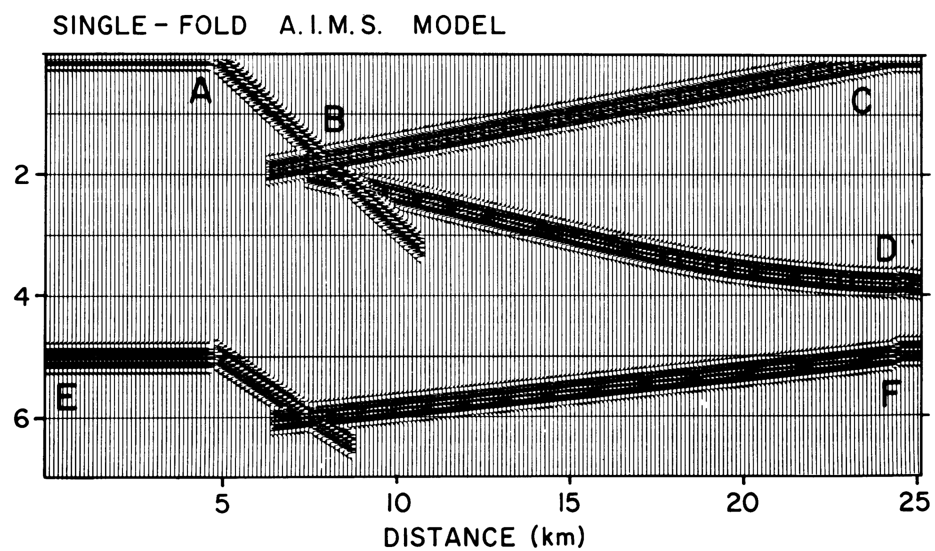


Figure 6B. 100% CMP section for model in Figure 6A. Figure from Peddy and others (1985).

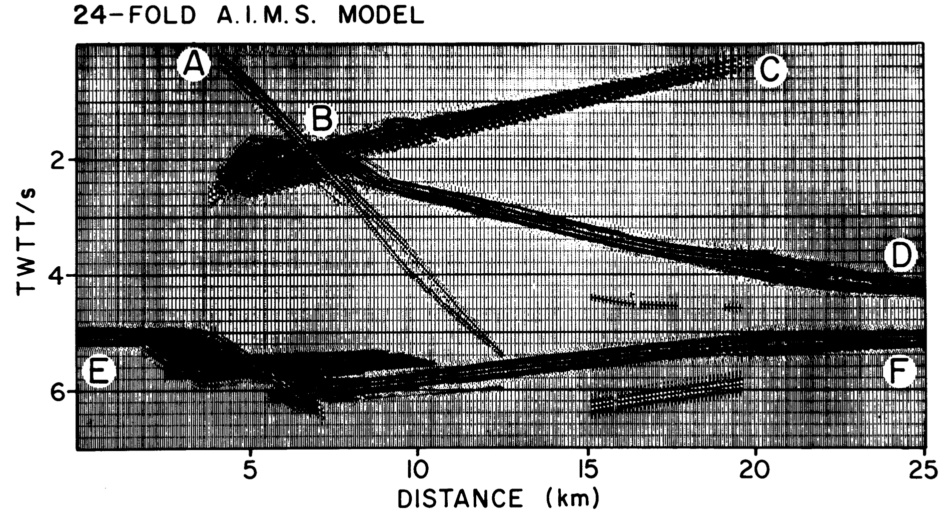


Figure 6C. 2,400% CMP section for model in Figure 6A. Note complexity and smearing of sub-basinal reflections. Synthetic seismic sections prepared using Advanced Interactive Modeling System (AIMS), a trademark of Geoquest International. Figure from Peddy and others (1985).

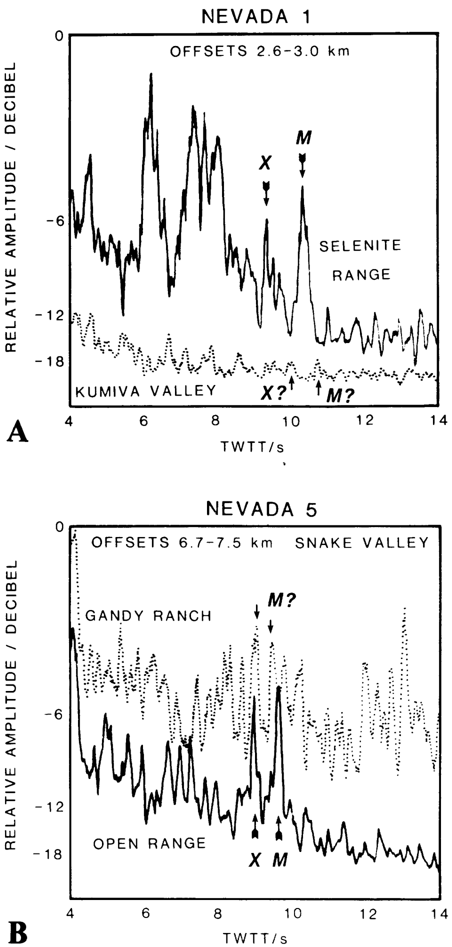


Figure 7. Amplitude plots for groups of traces from pairs of nearby SP gathers. Curves shown from 4 to 14 s only. All 4 curves are scaled to the same maximum amplitude, and curves are smoothed over a 100 ms window. Positions of amplitude analyses shown by stars in Figure 3A and 3D.

A. Solid curve: source in Selenite range; good signal penetration with clear reflections X and M. Dotted curve: source in Kumiva valley; poor signal penetration probably due to attenuation in Cenozoic basin fill. SPs for the 2 curves are separated by 11 km.

B. Solid curve: recorded in open range away from cultural noise sources; reflections X and M are clearly visible; amplitude curve continues to decay beyond 14 s. Dotted curve: recorded with SP at Gandy Ranch; background noise is greater than the amplitude of the Moho reflections; M cannot be confidently distinguished. SPs for the 2 curves separated by 7 km.

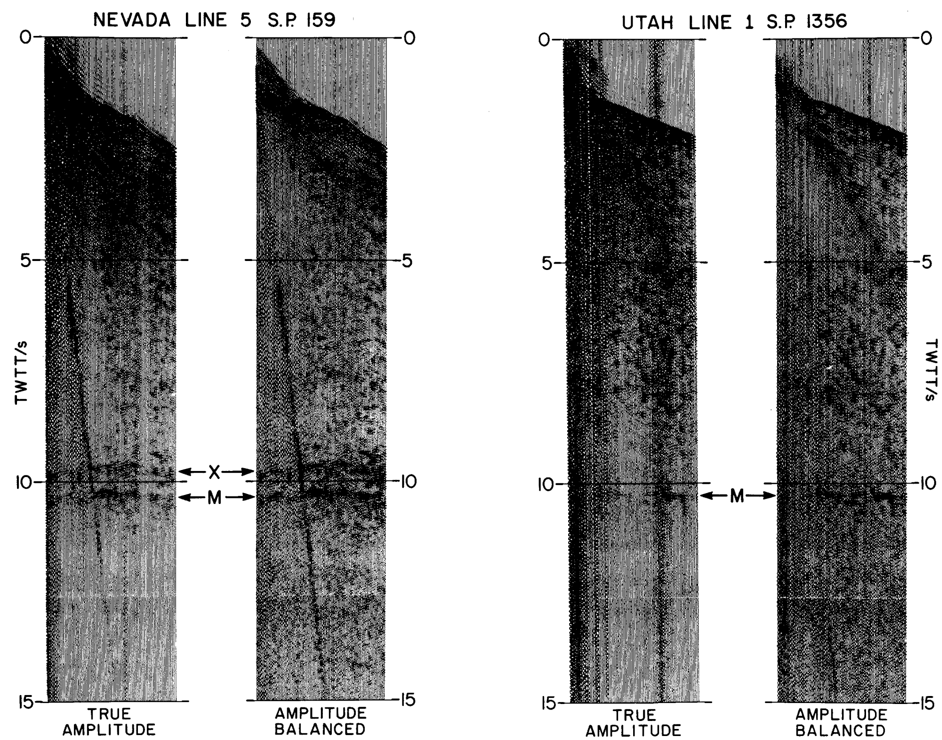
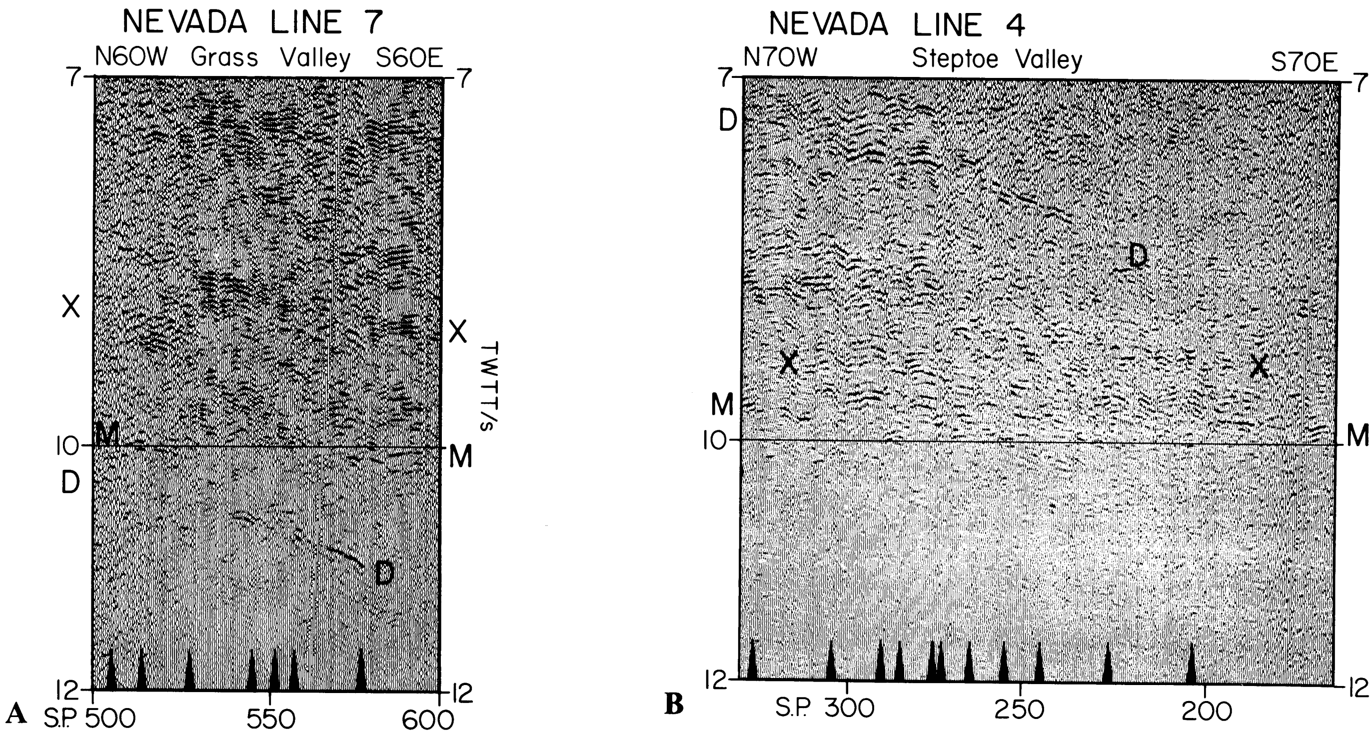


Figure 8. Comparison of data from Nevada line 5 and Utah line 1. A. Best data recorded in Snake Valley, Nevada line 5. B. Best data recorded in Sevier Desert, Utah line 1. Data are uncorrected for elevation and near-surface statics.



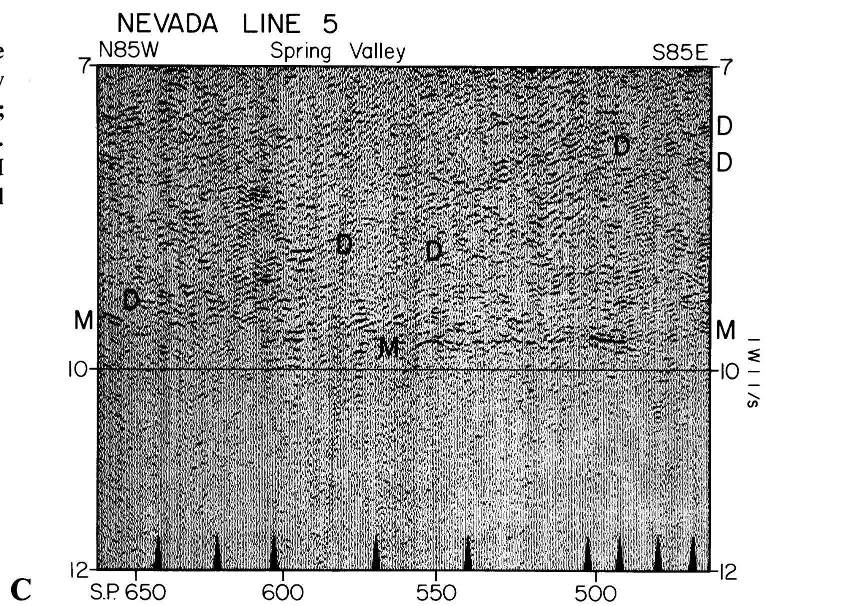


Figure 9. 100% CMP sections to illustrate apparent truncation of dipping reflections by Moho reflections. D: dipping reflections; other symbols as Figure 4. A. NV 7, east end. Note: reflections marked D move above M after 2-D migration. B. NV 4, central part. C. NV 5, west end.

such occurrence might permit the interpretation that the down-dip terminations of the dipping crustal reflectors are unrelated to the Moho. The frequent occurrence of truncation relationships suggests, however, that the Moho has overprinted older crustal fabrics and thus has been active as recently as or more recently than the age of the dipping fabrics.

Beneath the reflection Moho, reflections are rarely seen (Figs. 2 and 3). The lack of mantle reflections is probably due to the absence of highly reflective structures in this area, rather than to lack of signal penetration, because in at least some areas (Fig. 7B) the amplitude curve continues to decay beneath the Moho, showing that source-generated energy still exceeds ambient noise.

RELATION OF COCORP DATA TO PREVIOUS SEISMIC STUDIES OF THE MOHO IN THE NORTHERN BASIN AND RANGE PROVINCE

Previous seismic studies in the Basin and Range province have estimated Moho depth based on P_n and wide-angle P_mP arrivals. Because the primary observation in reflection profiling—as in all seismic experiments—is traveltime, we have chosen to convert other seismic estimates of crustal thickness and velocity to two-way traveltimes to the Moho. Figure 10 shows the range of Moho TWTT calculated from the layer thicknesses and velocities of published refraction models. The principal scatter arises from the conflicting interpretations of the reversed Fallon-to-Eureka refraction line of Eaton (1963) and Prodehl (1979). Whereas Eaton (1963) fit the data with a simple crustal model containing only one or two layers of constant velocity, Prodehl (1979) interpreted the same data using an approximation method that does not require sharp discontinuities, that takes into account velocity gradients, and that includes the effects of low-velocity zones. Eaton and Prodehl gave similar estimates of crustal thickness near Eureka (11.1–11.2 s at half the critical distance west of the Eureka shot-point) but very different estimates of crustal thickness near Fallon (Eaton, 9.0 s; Prodehl, 9.8 s at half the critical distance east of the Fallon shotpoint). The one-layer model of Eaton (1963) gave even lower estimates of crustal thickness (8.6 s near Fallon and 10.6 s near Eureka). In Figure 10A, data points are plotted at the locations given by the authors whose results are shown. This location is at half the critical dis-

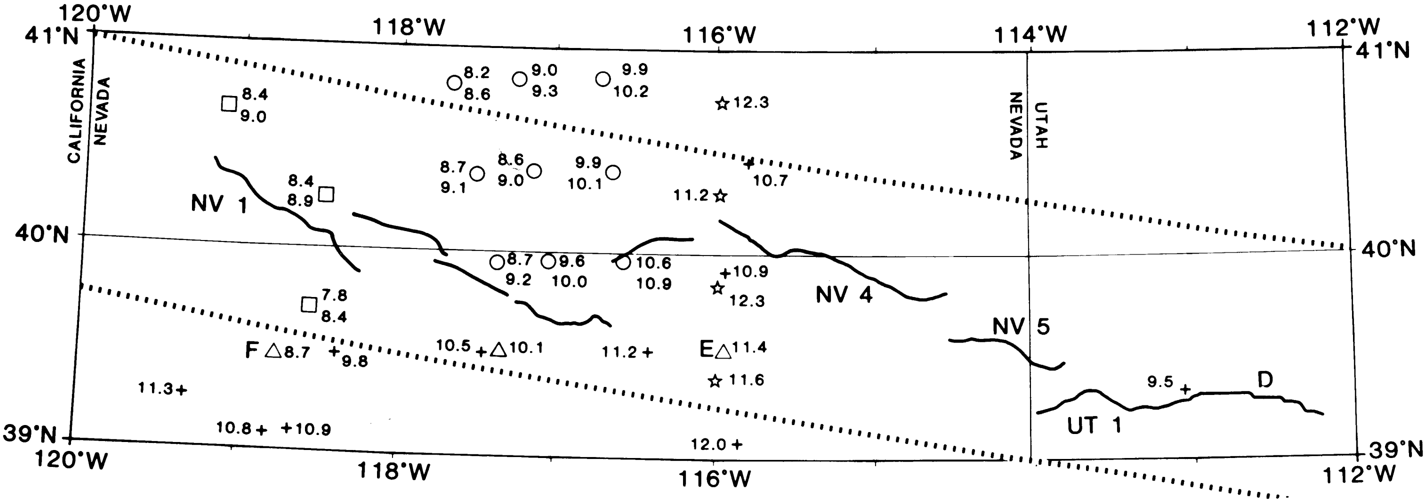
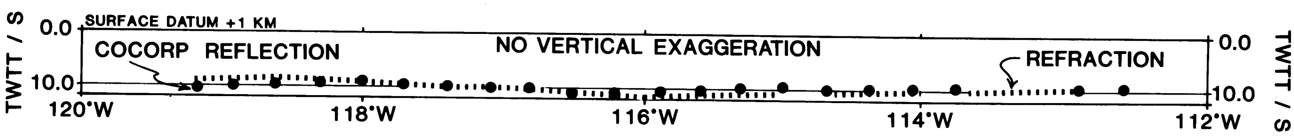


Figure 10A. Map covering same area as Figure 1, showing COCORP lines and depths to Moho interpreted from refraction data. Moho depths are given as TWTT/s from a datum 1 km above sea level. Triangles: Eaton (1963) (two-layer model) (one-layer model of Eaton is not shown, but would plot ~0.5 s above two-layer model); stars: Hill and Pakiser (1966); crosses: Prodehl (1979); circles: Stauber and Boore (1978); squares: Priestley and others (1982). Pairs of numbers by circles and squares represent unreversed data related to two-layer model of Eaton (1963) (top number) and to model of Prodehl (1979) (bottom number). Only selected, representative points are plotted for each survey. D, E, F: Delta, Eureka, and Fallon, locations of major refraction source points. Data falling between dotted lines are plotted in Figure 10B.



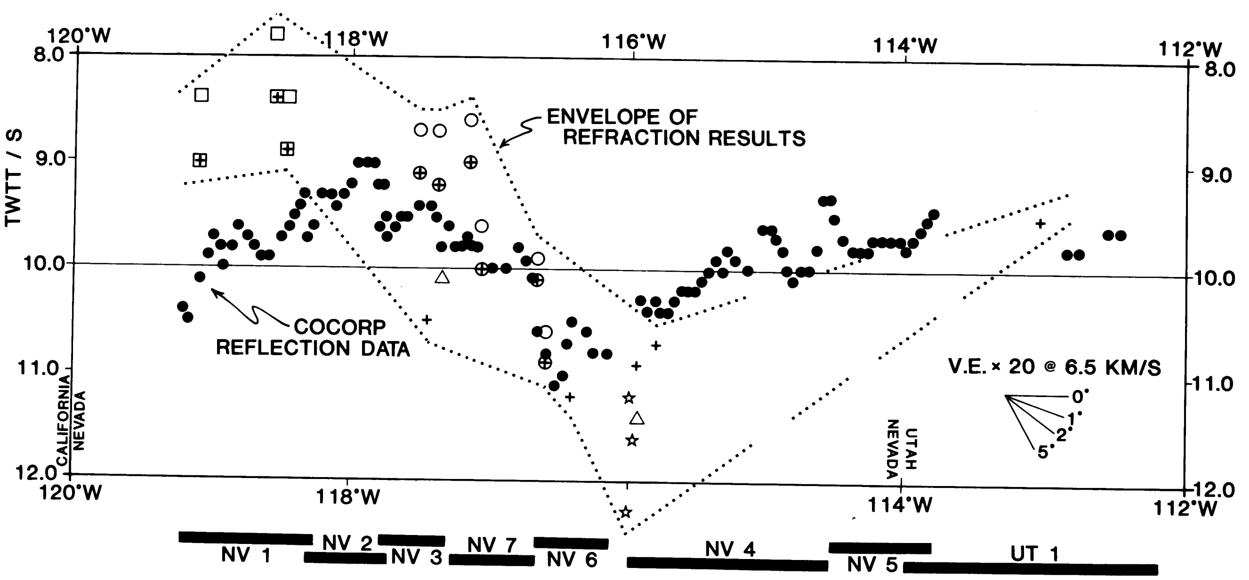


Figure 10B. Cross sections showing data of Figure 10A projected north-south onto latitude 40°N. Upper section is true scale at 6.5 km/s and shows section from surface to 12 s TWTT; lower section has vertical exaggeration of 20:1 and is from 8 to 12 s TWTT only. Symbols as Figure 10A; open squares and circles are data tied to model of Eaton (1963); squares and circles with crosses are data tied to model of Prodehl (1979). Filled circles are TWTT to reflection Moho from COCORP data. Heavy black lines show projections of COCORP lines onto 40°N.

tance from the shot-points for Prodehl (1979), but directly beneath the shot-points for Eaton (1963), even though a refraction line is only strictly reversed between the critical-reflection points, and Moho traveltimes beneath shot-points are necessarily extrapolations of the reversed data. Stauber and Boore (1978) and Priestley and others (1982) recorded nuclear explosions, quarry blasts, and local earthquakes in northern Nevada in unreversed profiles, and they estimated crustal thickness in northwestern Nevada using Eaton's (1963) one-layer model as an absolute datum to which they tied their models. In Figure 10, representative data points of Stauber and Boore (1978) and of Priestley and others (1982) are recalculated in the form of Moho TWTTs referenced both to Eaton's two-layer model and to Prodehl's more complex model that uses continuous velocity gradients and low-velocity zones.

The large spread of interpreted traveltimes to the refraction Moho exceeds the analytic errors involved in fitting models to the chosen phases (estimated by Prodehl, 1979, for example, to be ~3%, or 0.3 s at Moho traveltimes). Part of the spread in Figure 10B is undoubtedly due to projecting results as much as 80 km to the plane of the section at 40°N, but the large differences between the interpretations of Eaton and Prodehl probably correspond to different correlations of the seismic phases by the two writers (for example, Mooney and Gettings, 1984; Prodehl, 1979, p. 3). The spread of Moho TWTTs picked from the COCORP sections is also due in part to projecting data, collected as much as 40 km apart, to the same line, and partly to residual near-surface statics errors. The reflection data are not, however, generally subject to the large errors that may arise in refraction experiments due to problems of picking and correlating phases between widely spaced receiver stations.

Refraction data from the Basin and Range province have generally been interpreted as showing areas of thin crust (~20 km on the basis of the simplest model of Eaton, 1963) in northwestern Nevada and in western Utah and considerably thicker crust in eastern Nevada (~33 km) (for example, Smith, 1978). The COCORP data show the same general trend of Moho depths, but with considerably less Moho relief. The range of Moho TWTTs observed on the COCORP data is 9.0 to 11.2 s (times given after application of basin-stripping statics), a vertical relief of 6.6 km (at 6 km/s) to 7.7 km (at 7 km/s). To first order, on the scale of the thickness of the crust, the TWTT to the refraction Moho corresponds to the TWTT to the reflection Moho observed by COCORP (Fig. 10B). Differences are typically much less than 1 s (10%), except locally at the west end of Nevada line 1, and the sign of the differences is variable. In Nevada, the closest available refraction data are as far as 50 km from the reflection data. Given the evidence from the reflection lines for both east-west and north-south variations in crustal thickness, detailed correlations are of questionable value. Given also the uncertainties in inverting the refraction data, the different scales of resolution of the reflection and refraction methods, and the likelihood of some degree of crustal anisotropy, it seems probable that the reflection Moho and the refraction Moho are in fact the same boundary beneath northern Nevada.

In addition to comparing total traveltimes to the Moho, the velocity gradient at the base of the crust inferred from the refraction data has also been examined. Prodehl (1979) modeled the crust with a continuously varying velocity-function, including a gradient zone of rapidly increasing velocity at the base of the crust. A zone of steep velocity gradient might correspond to a zone of reflections from successive layers of increasing velocity. The thickness (in TWTT) was calculated for the zone at the base of the crust in Prodehl's model, in which the vertical velocity gradient exceeds 0.1 km/s/km, typically increasing to 1 km/s/km at the refraction Moho. The range of values calculated (0.4 to 1.2 s TWTT) is similar to the observed separation of reflections X and M on the COCORP sections (0.0 [only one reflection observed] to 1.2 s TWTT). Spatial correlation of the thickness of Prodehl's gradient zone with the separation of the two COCORP reflections is not possible, because Prodehl's results reflect lateral averaging of crustal properties, and because the vertical separation of the COCORP reflections X and M is laterally variable on a scale less than the separation between the COCORP lines and Prodehl's refraction lines. Even though the determination of a gradient zone at the base of the crust in Prodehl's models is apparently predicated by his interpretational methods, the approximate correspondence of the two zones interpreted independently from reflection and refraction data may suggest that the COCORP pair or band of reflections is a transition zone of increasing velocity at the base of the crust. A wide-angle or expanding-spread seismic-reflection experiment (for example, Liu and others, 1985) will be required to test this hypothesis.

Controlled-source and earthquake-source experiments also provide data on the velocity structure of the upper mantle. Thompson and Burke (1974) and Smith (1978) reviewed evidence for anomalously low P_n velocities in the Basin and Range province of ~7.8 km/s, significantly less than the normal velocity of ~8.2 km/s observed in stable regions. The low P_n velocities apparently persist to ~130 km depth, forming a thick low-velocity zone, or anomalous upper mantle (Thompson and Burke, 1974). In subsequent discussions of the nature of the Moho in the Basin and Range province, it is important to bear in mind that the Moho, the upper boundary of the anomalous upper mantle, may itself be anomalous in this region.

THE AGE OF THE MOHO IN THE BASIN AND RANGE PROVINCE

Evidence for the youth of the Moho is found in the apparent truncation by the Moho of overlying, dipping reflections (Fig. 9). Such truncations were taken by Roy Chowdhury and Hargraves (1981) to imply that the Moho formed subsequently to the truncated crustal fabric. Until the dipping crustal reflections are identified and dated by sampling or tracing to outcrop, however, these truncations will not provide tight constraints on the age of the Moho.

Better constraints on the age of the Moho arise from the continuity (Fig. 2) and similarity (Figs. 2 and 4) of the Moho across the entire width of Nevada, because the Basin and Range province is composed of regions thought to have had separate and contrasting tectonic histories prior to the Cenozoic. The western margin of the late Proterozoic North American craton (marked by the western extent of Paleozoic miogeoclinal strata and by the initial $^{87}Sr/^{86}Sr = 0.7060$ isopleth) runs north-south at ~117°W in northern Nevada (Stewart and Poole, 1974; Kistler and Peterman, 1978), crossing eastern line 3 or western line 7. During the mid-Paleozoic, the Antler allochthon of oceanic affinity was emplaced across this continental margin (Stewart, 1980; Speed, 1983). In Permo-Triassic time, a terrane of arc and oceanic-arc affinity (Sonomia) was accreted from the west and the Golconda allochthon emplaced above the older Antler allochthon (Speed, 1983). The apparent continuity and similarity of the Moho across suture zones implies that the Moho in Nevada has been formed since the accretion of Sonomia in the Permo-Triassic.

In early to middle Mesozoic time, an active margin was established farther to the west, producing extensive arc magmatism in California and western Nevada. The apparent continuity of the Moho beneath the eastern boundary of the Sierra Nevada batholithic terrane, which runs approximately north-south in the vicinity of lines 2 and 3 (Crowder and others, 1973), suggests that the Moho in western Nevada is late Cretaceous or younger.

Tighter constraints on the age of the Moho in the Basin and Range province depend on comparisons of the reflection Moho in Nevada and western Utah. The characteristic pair of reflections (X and M) observed in the Nevada Basin and Range has not yet been observed in the Utah Basin and Range (Fig. 8). One possible interpretation is that the dichotomous character of the reflection Moho may have resulted from different tectonic styles during the Sevier orogeny across the present-day Basin and Range province. In the Jurassic and Cretaceous, eastern Nevada was the site of greenschist- to amphibolite-facies metamorphism and calc-alkalic plutonism, whereas the hinterland of the Sevier orogenic belt in western Utah and the Snake Range of easternmost Nevada was characterized by nonductile deformation and insignificant metamorphism (Allmendinger and Jordan, 1981). The existence, however, of high-amplitude Moho reflections on Utah line 1 (Fig. 8) as well as across Nevada suggests that the Moho may well be as young as the Cenozoic extension and volcanism that characterize the Basin and Range province in both Utah and Nevada. Differences between the seismic character of the reflection Moho in Nevada and Utah, if real, could be due to possibly differing styles of basin-range extension in these areas. For example, the Utah Basin and Range at 40°N is characterized by major, low-angle normal faults (Sevier Desert detachment; Allmendinger and others, 1983). In contrast, in Nevada such faults have not been imaged, and extension may be accommodated by distributed ductile flow in the lower crust (R. W. Allmendinger and others, unpub. data; T. Hauge and others, unpub. data).

Even if the formation of the Moho reflectors predates formation of the Basin and Range province, the present relatively horizontal, large-scale geometry of the reflectors (as opposed to the physical nature of the impedance contrasts producing them) must be due to Cenozoic extension. This is necessitated by the vertical movements of the Moho required during regional isostatic adjustment of the lithosphere to crustal basin-range extension (Thompson and Burke, 1974). Adjustment of Moho depth during Cenozoic extension is also implied by the interpretive pre–basin-range crustal isopach maps of Coney and Harms (1984), which show a much thicker crust than at present exists, including a crustal welt formed by Sevier and Laramide shortening along the Utah-Nevada border.

In summary, the formation of the observed Moho reflections in the northern Basin and Range province appears to postdate Triassic microplate accretion and Cretaceous arc-magmatism, and the province-scale geometry of the reflection Moho appears to be a product of basin-range extension. The formation of the observed reflectors is arguably a result of hinterland magmatism and deformation during Mesozoic and early Cenozoic regional compression or, more probably, is a result of middle to late Cenozoic extension, high heat flow, and magmatism.

SOME OBSERVATIONS ON THE REFLECTION MOHO IN OTHER AREAS

This paper does not attempt a comprehensive review of observations of the reflection Moho, although there have been many such observations since the compilation and analysis of Hale and Thompson (1982). Without, however, minimizing possible effects of variable data collection techniques and upper-crustal geologic features on observations of the reflection Moho, two observations are presented here that may be of relevance in subsequent discussion of the Moho in the Basin and Range province. First, seismic-reflection data indicate that the character of the Moho is different in different tectonic provinces. Second, a prominent pair of reflections at the base of the crust has been found in areas other than the Basin and Range province, although this pair of reflections has not previously been recognized as a widespread feature.

In contrast to the clear reflection Moho in the Basin and Range province, no prominent reflection Moho was detected beneath the Sierra Nevada (Nelson and others, in press), whereas beneath the Colorado Plateau there is a rapid drop-off of reflected energy at Moho depths, and, locally, horizontal bright reflections are detected (H. G. Farmer and others, unpub. data). The Basin and Range province is therefore apparently distinguished from adjacent tectonic provinces by the prominent reflection Moho as well as by its history of prominent Cenozoic rifting and volcanism. Similarly, in the southern Appalachians, Cook and others (1983) observed Moho reflections from beneath the Coastal Plain (the Mesozoic rift margin of the Atlantic) and the adjacent Carolina Slate Belt, but no comparable reflections farther to the northwest beneath the Inner Piedmont and the Blue Ridge. Even this small sample of the many COCORP surveys shows that the expression of the Moho on reflection profiles is variable, and hence, perhaps, that the Moho has a variable origin and/or a variable evolution (for example, Oliver, 1982). These associations of a sharp, reflective Moho with recently rifted areas, including the Basin and Range province, and of a gradational crust-mantle boundary with areas of thicker crust, including the Sierra Nevada and the Colorado Plateau, were previously made by Prodehl (1977) on the basis of refraction observations.

Paired reflections at the base of the crust have been observed elsewhere, for example, on recent COCORP data from northeastern Washington and northern Idaho (Fig. 11), near the edge of the Neogene Columbia River flood basalts in a region of Eocene volcanism and coeval extension (for example, Ewing, 1980). Two distinct reflections are also visible in the Gabilan mountains, California (Hale and Thompson, 1982,

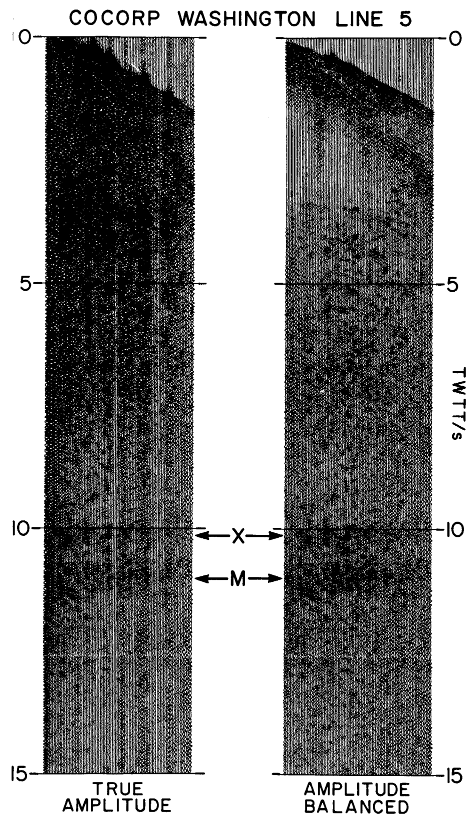


Figure 11. SP gather from COCORP Washington line 5, illustrating pair of reflections at base of crust.

their Fig. 2; Feng and McEvilly, 1983); in the Bowen basin, northeastern Australia (Mathur, 1983, his Fig. 5); in the Ukraine, on the margin of the Dnieper-Donets graben (Pavlenkova and Smelyanskaya, 1970, their Fig. 2); and in the Urach geothermal and Neogene volcanic area, West Germany (Bartelsen and others, 1982, their Fig. 9). For the seismic lines in California, Australia, and Germany, wide-angle reflection and refraction data were used to identify the zone between the two deep reflections as having a velocity transitional between those of the crust and mantle. The Urach area is ~100 km east of the Rhinegraben, beneath which a "rift cushion," with velocities intermediate between those of the crust and the mantle, has been detected by refraction seismology (Mueller and others, 1973). This rift cushion is interpreted to extend beneath the Urach area as a transition zone with a thickness of ~5 km (Zucca, 1984) and is thus commensurate with the thickness of the zone between the deep reflections observed by Bartelsen and others (1982). Transition zones, or anomalous lower-crustal layers, with velocities between 7 and 7.5 km/s have been recognized in many other continental rifts (Mooney and others, 1983), and it is possible that the zone between X and M in the Basin and Range province, and between paired reflections at the base of the crust elsewhere, may have an origin related to these anomalous lower-crustal layers. A pair of reflections separated by ~4 km at the base of the crust was also detected beneath the Hawaiian seamount chain (Watts and others, 1985), in contrast to a single sharp Moho reflection observed away from the island chain. Watts and others (1985) used expanding-spread profiles to show that the deep-crustal body bounded by these reflections has a velocity transitional between those of the crust and mantle, and they interpreted the deep-crustal body as a complex of tholeiitic sills intruded into upper-mantle rocks. Any genetic relation of this pair of oceanic reflections, or of a similar pair of reflections recognized in the western North Atlantic (Mutter and others, 1985), to the pair of reflections in the Basin and Range province is, of course, speculative at the present time.

THE ORIGIN OF THE MOHO IN THE BASIN AND RANGE PROVINCE

Any model of the Moho in the Basin and Range province must explain the sharp, continuous Moho reflections, varying in depth only by ~7 km, that may truncate dipping crustal reflections and also the two distinct reflections or zone of reflections straddling as much as 4 km at the base of the crust, possibly encompassing a zone of velocity intermediate between that of the crust and the mantle. Attention has historically focused on the alternatives of either a chemical discontinuity, perhaps along an intrusive contact or a shear zone, or an isochemical phase change in rocks of basaltic composition. Although the phase-change model makes specific predictions about the seismic nature of the Moho, more detailed expositions of models of the Moho as a chemical discontinuity will be needed before such models can be adequately tested against the present seismic data.

Phase-Change Model

Although the lack of global correlation between surface heat flow and Moho depth has been taken to imply that the continental Moho cannot everywhere represent the gabbro to garnet granulite to eclogite phase change (for example, Pakiser, 1965), Ringwood (1975) recognized that the gabbro-eclogite transition may play a significant role at the Moho in tectonically active regions with thick layers possessing P-wave velocities between 7 and 8 km/s. Ito and Kennedy (1970) specifically identified the upper mantle beneath the Basin and Range province, with its anomalously low P_n velocity (7.5 to 7.8 km/s; for example, Thompson and Burke, 1974; Smith, 1978), as garnet granulite, and therefore we pay particular attention here to the phase-change hypothesis for the Basin and Range province. Using the new COCORP data, two specific tests of this hypothesis are possible. First, one can compare pressure-temperature pairs, calculated for the Moho from COCORP traveltime data and from surface heat flow data, with the experimental gabbro to garnet granulite phase transition (for example, Mareschal and others, 1982). Although the COCORP data yield reasonable pressure estimates at the Moho, uncertainties in the Moho temperature due to the likelihood that measured heat flow contains laterally variable effects of magmatic heat transfer and ground-water convection and due to possible changes in the thermal state of the crust within the thermal time constant of the crust (10 to 15 Ma) preclude the resolution by this test of whether a phase change can satisfy the COCORP data on the Moho (Klemperer, 1985).

Second, although the mechanisms by which a phase change might produce reflections at the wavelengths in use in this experiment are uncertain, one can compare COCORP observations of thickness and relief of the reflection Moho with experimental evidence for density changes during the phase change. Ito and Kennedy (1970) suggested that the appearance of garnet in the gabbro to garnet granulite transition produces a 5% increase in density within a pressure interval of ~200 MPa. On account of the linear relationship between density and velocity (Birch, 1961), a 5% jump in density should produce a seismic impedance contrast (reflectivity) of ~5%. In contrast, Green and Ringwood (1972) preferred a smooth increase in density over a much wider range of pressures. The COCORP data show two reflections or a zone of reflections straddling traveltimes of no more than 1.2 s (often much less), corresponding to a pressure range of no more than ~120 MPa. The COCORP data thus show a crust-mantle transition that is probably too thin to satisfy the experimental data of Ito and Kennedy (1970) and is much too thin to satisfy the data of Green and Ringwood (1972). Green and Ringwood also pointed out the existence of variations between the equilibrium curves for phase transitions in different basaltic compositions of several hundred megapascals (MPa) (several seconds TWTT), or approximately twice the maximum Moho relief seen on the COCORP data. Unless, then, the lowermost crust along the COCORP transect is chemically homogeneous, the subdued Moho relief across Nevada, coupled with the sharpness of the reflection Moho observed in the COCORP data, seems to rule out the gabbro to garnet granulite transition as the cause of the reflection Moho in this region.

Additional evidence mitigates against the phase-change model of the Moho in the Basin and Range province. First, at the only xenolith locality in central or northern Nevada, the Lunar Crater area (38.5°N, 116°W), many ultramafic tectonites (of presumed mantle origin) and gabbroic xenoliths with igneous textures (of presumed crustal origin) are found, whereas no garnet-bearing rocks have been reported (Trask, 1969; H. G. Wilshire and J.E.N. Pike, 1985, personal commun.). Second, Bamford and others (1979) stated a case for as much as 3% anisotropy in P_n velocities beneath the Basin and Range province. If substantiated, this anistropy would argue against garnet granulite or eclogite, both relatively isotropic rocks (Kumazawa and others, 1971; Babuska, 1984), as the major rock types in the uppermost mantle.

Intrusive Boundary

Almost one-quarter of the surface area of Nevada is made up of igneous rocks younger than 43 Ma (Stewart, 1980), and hence it seems likely that magmatic activity has had an important effect on the tectonics and structures of the Basin and Range province, including the Moho. Magmatic underplating or ponding at the Moho has in the past been advocated as a mode of crustal formation (for example, Fyfe, 1974). In the model of Cox (1980), picritic magmas underplate the crust and fractionate until the residual liquid is basaltic; the Moho then becomes the petrologic interface between ultramafics added to the mantle and gabbros added to the lower crust. Herzberg and others (1983) speculated that the Moho represents a density boundary through which low-density magmas, evolved or hydrous, may pass, but beneath which higher-density, more primitive magmas are underplated. They further suggested that the underplating model would produce a sharply defined Moho, as opposed to a disrupted Moho produced by Moho-penetrating intrusions. The Moho also might vary locally between a single petrologic boundary and a transition zone of interlayered intrusions and crustal rocks (for example, Ivrea Zone; Mehnert, 1975), offering an explanation for the variations in thickness of the basal crustal layer seen on the COCORP data. If the Moho were intrusive, it might also tend to overprint older crustal structures, thus truncating dipping features above the Moho, as seen on the COCORP data.

An intrusive complex at the base of the crust might be reflective due to juxtaposition of sills and country rock with differing lithologies. Additionally, magma may be actually be present at the Moho, at least locally. Nevada lines 1, 2, 3, and 6 run along the southern edge of the Battle Mountain heat-flow high, for which Moho temperatures have been estimated to be 1050 to 1300 °C (Lachenbruch and Sass, 1977), close to or exceeding the dry basalt solidus.

The COCORP observations of the Moho seem to be consistent with the intrusive model of the Moho, and it is intriguing that Neogene volcanism has been important not only in the Basin and Range province but also in some other areas where a pair of basal crustal reflections are observed (northern Idaho–Washington, Urach geothermal anomaly, Hawaiian seamount chain). As noted above, however, models of intrusion do not yet offer clear predictions to be tested by the present reflection data.

Shear Zone

Stewart (1978) reviewed estimates of the amount of basin-range crustal extension that vary from 10% to 100%. Upper-crustal extension has been fundamental in the Neogene tectonic history of the Basin and Range province, but at some depth in the crust or mantle the amount of extension must decrease, and return flow inward beneath the Basin and Range province is required (Thompson, 1972). It is possible, among other alternatives, that the rates of extension in the lowermost crust and in the uppermost mantle are different, and that the Moho is acting as a shear zone. Rheological models of a quartzo-feldspathic crust overlying an olivine-rich upper mantle have a strength minimum at the Moho (for example, Meissner and Strehlau, 1982; Smith and Bruhn, 1984), so that strain might be preferentially concentrated at this depth. A zone of high strain could be seismically reflective, either by juxtaposing rocks of different petrology and seismic impedance, or by producing mylonites that are reflective due to their anisotropic fabric, internal layering, chemical alteration, or elevated pore pressure (Jones and Nur, 1984). Chemical alteration, elevated fluid pore pressures, and anisotropic fabrics are not, of course, limited to shear zones, and Berry and Mair (1980), for example, discussed the possibility that the Moho results from olivine recrystallization in an anisotropic stress field.

Shear zones offset or transpose pre-existing crosscutting features and thus could explain the truncation relationships observed on the seismic data. Crustal shear zones can certainly have the dimensions required to explain the COCORP basal crustal reflections. For example, the west Greenland Norde Stromfjord shear zone, developed at granulite facies, is exposed for a length of 170 km and has a vertical width of 2 to 3 km (Sorensen, 1983). The Whipple detachment fault, developed at 9 to 12 km depth, or possibly much deeper (Anderson, 1985), underlies 3,000 km^2 and exhibits >3.9 km of mylonitic rocks bounded by a sharp (3 to 30 m) transition to relatively undeformed rocks (Davis and others, 1979). This mylonite zone is also intruded by syntectonic plutonic sheets, as much as a few hundred metres thick, which might be expected to give planar reflections.

CONCLUSIONS

COCORP data from the 40°N transect provide various constraints on the Moho beneath the Basin and Range province, most notably with regard to its age. The age constraints and the detailed geometry of the reflection Moho both suggest that extension and igneous intrusion have been more important than isochemical phase changes in forming the reflection Moho in the northern Basin and Range province. The pair of reflections X and M may be genetically related to similar pairs of reflections observed elsewhere in the world. The lower crust is probably as variable in composition and structure as the upper crust, as evidenced by recent studies of exposed crustal cross sections (for example, Fountain and Salisbury, 1981; Percival and Card, 1983), and different models for the Moho may therefore be appropriate in different areas. Reflective transitions of different types may even arise in the same area. In areas of mixed rock type, as typified by the interleaved metabasalts and metasedimentary granulites of the Ivrea Zone (Mehnert, 1975), both phase changes and chemical discontinuities may produce reflections within the same crustal column. In future work, it will be important to develop more specific models for Moho formation that are testable with seismic-reflection data. One potential test may arise from the identification of areas where steeply dipping lithospheric reflectors transect Moho reflectors. The discovery of offsets of the Moho, or of offsets of dipping reflectors at the Moho, or of the disappearance of dipping reflectors above or below the Moho, will help to constrain the nature of the Moho in different localities (for example, Peddy, 1984; Phinney, 1985). The best test, however, is to trace the reflection Moho to the surface or to drillable depths.

ACKNOWLEDGMENTS

As with all COCORP projects, this study represents the collective efforts of the entire COCORP staff and members of the COCORP committees. We particularly acknowledge helpful discussions with Rick Allmendinger, Muawia Barazangi, Larry Brown, Robert Kay, Suzanne Kay, and Walter Mooney. We thank L. W. Braile, R. B. Smith, J. H. Stewart, G. A. Thompson, and M. L. Zoback for proposing the basin-range seismic lines to the COCORP Site Selection Committee, and R. A. Phinney, G. A. Thompson, and an anonymous reviewer for their comments. This research was supported by National Science Foundation Grants EAR82-12445 and EAR 83-13569. Klemperer received partial funding from the Milton Dobrin Scholarship of the Society for Exploration Geophysicists. Data acquisition was carried out by Crew no. 6834 of Petty-Ray Geophysical, a division of Geosource Inc., and data were processed on the COCORP Megaseis (TM Seiscom Delta) computer system.

REFERENCES CITED

Allmendinger, R. W., and Jordan, T. E., 1981, Mesozoic evolution, hinterland of the Sevier orogenic belt: Geology, v. 9, p. 308–313.

Allmendinger, R. W., Sharp, J. W., von Tish, D., Serpa, L., Kaufman, S., Oliver, J., and Smith, R. B., 1983, Cenozoic and Mesozoic structure of the eastern Basin and Range from COCORP seismic-reflection data: Geology, v. 11, p. 532–536.

Allmendinger, R. W., Hauge, T., Hauser, E. C., Potter, C., and Oliver, J., in press, Tectonic heredity and the layered lower crust in the Basin and Range province, western United States: Geological Society of London Journal.

Anderson, J. L., 1985, Contrasting depths of "core complex" mylonitization: Barometric evidence: Geological Society of America Abstracts with Programs, v. 17, p. 337.

Babuska, V., 1984, P-wave velocity anisotropy in crystalline rocks: Royal Astronomical Society Geophysical Journal, v. 76, p. 113–119.

Bamford, D., Jentsch, M., and Prodehl, C., 1979, P_n anisotropy studies in northern Britain and the eastern and western United States: Royal Astronomical Society Geophysical Journal, v. 57, p. 397–429.

Bartelsen, H., Lueschen, E., Krey, Th., Meissner, R., Schmoll, H., and Walter, Ch., 1982, The combined seismic reflection-refraction investigation of the Urach geothermal anomaly, *in* Haenel, R., ed., The Urach geothermal project (Swabian Alb, Germany): Stuttgart, Schweizerbart'sche Verlagsbuchhandlung, p. 247–262.

Barton, P., Matthews, D., Hall, J., and Warner, M., 1984, Moho beneath the North Sea compared on normal-incidence and wide-angle seismic records: Nature, v. 308, p. 55–56.

Behrens, J., Bortfeld, R., Gommlich, G., and Kohler, K., 1972, Interpretation of discontinuities by seismic imaging: Zeitschrift für Geophysik, v. 38, p. 481–498.

Berry, M. J., and Mair, J. A., 1980, Structure of the continental crust: A reconciliation of seismic reflection and refraction studies, *in* Strangway, D. W., ed., The continental crust and its mineral deposits: Geological Association of Canada Special Paper 20, p. 195–213.

Birch, F., 1961, The velocity of compressional waves in rocks to 10 kilobars, part 2: Journal of Geophysical Research, v. 66, p. 2199–2224.

Cheadle, M., Czuchra, B., Byrne, T., Ando, C., Oliver, J., Brown, L., Kaufman, S., Malin, P., and Phinney, R., in press, The deep crustal structure of the Mojave Desert, California, from COCORP seismic reflection data: Tectonics.

Coney, P. J., and Harms, T. A., 1984, Cordilleran metamorphic core complexes: Cenozoic extensional relics of Mesozoic compression: Geology, v. 12, p. 550–554.

Cook, F. A., Brown, L. D., Kaufman, S., and Oliver, J. E., 1983, The COCORP seismic reflection traverse across the southern Appalachians: American Association of Petroleum Geologists Studies in Geology 14, 61 p.

Cox, K. G., 1980, A model for flood basalt vulcanism: Journal of Petrology, v. 21, p. 629–650.

Crowder, D. F., McKee, E. H., Ross, D. C., and Krauskopf, K. B., 1973, Granitic rocks of the White Mountains area, California-Nevada: Age and regional significance: Geological Society of America Bulletin, v. 84, p. 285–296.

Davis, G. A., Anderson, J. L., Frost, E. G., and Shackelford, T. J., 1979, Regional Miocene detachment faulting and early Tertiary(?) mylonitization, Whipple-Buckskin-Rawhide mountains, southeastern California and western Arizona, *in* Abbott, P. L., ed., Geological excursions in the Southern California area: San Diego, California, San Diego State University, p. 75–108.

Davydova, N. I., Kosminskaya, I. P., Kapustian, N. I., and Michota, G. G., 1972, Models of the Earth's crust and M-boundary: Zeitschrift für Geophysik, v. 38, p. 369–393.

Eaton, J. P., 1963, Crustal structure from San Francisco, California to Eureka, Nevada, from seismic refraction experiments: Journal of Geophysical Research, v. 68, p. 5789–5806.

Ewing, T. E., 1980, Paleogene tectonic evolution of the Pacific Northwest: Journal of Geology, v. 88, p. 619–638.

Feng, R., and McEvilly, T. V., 1983, Interpretation of seismic reflection profiling data for the structure of the San Andreas fault zone: Seismological Society of America Bulletin, v. 73, p. 1701–1720.

Finlayson, D. M., Collins, C.D.N., and Lock, J., 1984, P-wave velocity features of the lithosphere under the Eromanga basin, eastern Australia, including a prominent mid-crustal (Conrad?) discontinuity: Tectonophysics, v. 101, p. 267–291.

Fountain, D. M., and Salisbury, M. H., 1981, Exposed cross-sections through the continental crust: Implications for crustal structure, petrology, and evolution: Earth and Planetary Science Letters, v. 56, p. 263–277.

Fyfe, W. S., 1974, Archaean tectonics: Nature, v. 249, p. 338.

Green, D. H., and Ringwood, A. E., 1972, A comparison of recent experimental data on the gabbro-garnet granulite-eclogite transition: Journal of Geology, v. 80, p. 277–288.

Hale, L. D., and Thompson, G. A., 1982, The seismic reflection character of the continental Mohorovicic discontinuity: Journal of Geophysical Research, v. 87, p. 4625–4635.

Hauge, T., Allmendinger, R., Burgess, S., Caruso, C. Gibbs, A., Hauser, E., Huang, J., Klemperer, S., Liu, C-S., Nelson, K., Opdyke, S., Potter, C., Sanford, W., Serpa, L., Zheng, L., Zhu, T., Brown, L., Kaufman, S., and Oliver, J., 1984, The COCORP 40°N transect of the North American cordillera: Part 1: Geological Society of America Abstracts with Programs, v. 16, p. 532.

Hauser, E., Allmendinger, R., Burgess, S., Caruso, C., Gibbs, A., Hauge, T., Huang, J., Klemperer, S., Liu, C-S., Nelson, K., Opdyke, S., Potter, C., Sanford, W., Serpa, L., Zheng, L., Zhu, T., Brown, L., Kaufman, S., and Oliver, J., 1984, The COCORP 40°N transect of the North American cordillera: Part 2: Geological Society of America Abstracts with Programs, v. 16, p. 532.

Herzberg, C. T., Fyfe, W. S., and Carr, M. J., 1983, Density constraints on the formation of the continental Moho and crust: Contributions to Mineralogy and Petrology, v. 84, p. 1–5.

Hill, D. P., and Pakiser, L. C., 1966, Crustal structure beneath the Nevada Test Site and Boise, Idaho, from seismic-refraction experiments, *in* Steinhart, J. S., and Smith, T. J., eds., Earth beneath the continents: American Geophysical Union Geophysical Monograph 10, p. 391–419.

Holbrook, S., 1984, Constraints on the crustal structure of Nevada imposed by a regional gravity model: Geological Society of America Abstracts with Programs, v. 16, p. 531.

Ito, K., and Kennedy, G. C., 1970, The fine structure of the basalt-eclogite transition: Mineralogical Society of America Special Paper 3, p. 77–88.

Jones, T. D., and Nur, A., 1984, The nature of seismic reflections from deep crustal fault zones: Journal of Geophysical Research, v. 89, p. 3153–3171.

Kistler, R. W., and Peterman, Z. E., 1978, Reconstruction of crustal blocks of California on the basis of initial strontium isotopic compositions of Mesozoic granitic rocks: U.S. Geological Survey Professional Paper 1071, 17 p.

Klemperer, S. L., 1985, The continental lower crust and Moho: Studies using COCORP deep seismic reflection profiling [Ph.D. thesis]: Ithaca, New York, Cornell University, 146 p.

Kumazawa, M., Helmstaedt, H., and Masaki, K., 1971, Elastic properties of eclogite xenoliths from diatremes of the east Colorado Plateau and their implication to the upper mantle structure: Journal of Geophysical Research, v. 76, p. 1231–1247.

Lachenbruch, A. H., and Sass, J. H., 1977, Heat flow in the United States and the thermal regime of the crust, *in* Heacock, J. G., ed., The Earth's crust: American Geophysical Union Geophysical Monograph 20, p. 626–675.

Lemiszki, P. J., Knuepfer, P.L.K., Hauge, T. A., Brown, L. D., Oliver, J. E., and Kaufman, S., 1985, Crustal structure of the Basin and Range–Sierra Nevada transition from COCORP seismic reflection profiling: EOS (American Geophysical Union Transactions), v. 66, p. 979.

Liu, C-S., Zhu, T., Farmer, H., and Brown, L., 1985, An expanding spread experiment during COCORP's field operation in Utah, *in* Barazangi, M., and Brown, L., eds., Reflection seismology: A global perspective: Washington, D.C., American Geophysical Union Geodynamics Series 13, p. 237–246.

Mareschal, J.-C., Gangi, A. F., and Lamping, N. L., 1982, The Moho as a phase change: A test of the hypothesis: Journal of Geophysical Research, v. 87, p. 4723–4730.

Mathur, S. P., 1983, Deep reflection experiments in northeastern Australia, 1976–1978: Geophysics, v. 48, p. 1588–1597.

Mehnert, K. R., 1975, The Ivrea Zone: A model of the deep crust: Neues Jahrbuch für Mineralogie Abhandlungen, v. 125, p. 156–199.

Meissner, R., 1967, Exploring deep interfaces by seismic wide-angle measurements: Geophysical Prospecting, v. 15, p. 598–617.

Meissner, R., and Strehlau, J., 1982, Limits of stresses in continental crusts and their relation to the depth-frequency distribution of shallow earthquakes: Tectonics, v. 1, p. 73–89.

Mohorovičić, A., 1910, Das Beben vom 8.X.1909, Jahrbuch des Meteorologischen Observatoriums in Zagreb (Agram) für das Jahr 1909: Jahrgang 9, Teil 4, Abschnitt 1, p. 3–63.

Mooney, W. D., and Gettings, M. E., 1984, One-dimensional velocity-depth functions determined for the Arabian shield and the southwestern Red Sea: A comparison of models, *in* Mooney, W. D., and Prodehl, C., eds., Proceedings of the 1980 Workshop of the International Association of Seismology and Physics of the Earth's Interior on the Seismic Modeling of Laterally Varying Structures: Contributions based on Data from the 1978 Saudi Arabian Refraction Profile: U.S. Geological Survey Circular 937, p. 125–139.

Mooney, W. D., Andrews, M. C., Ginzburg, A., Peters, D. A., and Hamilton, R. M., 1983, Crustal structure of the Mississippi embayment and a comparison with other continental rift zones: Tectonophysics, v. 94, p. 327–348.

Mueller, S., Peterschmitt, E., Fuchs, K., Emter, D., and Ansorge, J., 1973, Crustal structure of the Rhinegraben area: Tectonophysics, v. 20, p. 381–391.

Mutter, J. C., and North Atlantic Transect Study Group, 1985, Multichannel seismic images of the oceanic crust's internal structure: Evidence for a magma chamber beneath the Mesozoic Mid-Atlantic Ridge: Geology, v. 13, p. 629–632.

Nelson, K. D., Zhu, T., Gibbs, A., Harris, R., Oliver, J. E., Kaufman, S., Brown, L., and Schweikert, R. A., in press, COCORP deep seismic reflection profiling in the northern Sierra, California: Tectonics.

Oliver, J. E., 1982, Changes at the crust mantle boundary: Nature, v. 299, p. 398–399.

Oliver, J., Allmendinger, R. W., Hauge, T. A., Hauser, E. C., Potter, C. J., Klemperer, S. L., Nelson, K. D., and Knuepfer, P., 1985, COCORP 40°N transect in the North American cordillera: EOS, (American Geophysical Union Transactions), v. 66, p. 1073.

Pakiser, L. C., 1965, The basalt-eclogite transformation and crustal structure in the western United States: U.S. Geological Survey Professional Paper 525-B, p. B1–B8.

Pavlenkova, N. I., and Smelyanskaya, T. V., 1970, The nature of the group of reflected seismic waves from the base of the Earth's crust: Akademiya Nauk SSSR, Izvestiya, Physics of the Solid Earth, no. 1, p. 10–18.

Peddy, C. P., 1984, Displacement of the Moho by the Outer Isles Thrust shown by seismic modelling: Nature, v. 312, p. 628–630.

Peddy, C. P., Brown, L. D., and Klemperer, S. L., 1985, Interpreting the deep structure of rifts with synthetic seismic sections, *in* Barazangi, M., and Brown, L., eds., Reflection seismology: A global perspective: Washington, D.C., American Geophysical Union Geodynamics Series 13, p. 301–311.

Percival, J. A., and Card, K. D., 1983, Archaean crust as revealed in the Kapuskasing uplift, Superior Province, Canada: Geology, v. 11, p. 323–326.

Phinney, R. A., 1985, A seismic cross section of the New England Appalachians: The orogen exposed, *in* Barazangi, M., and Brown, L. D., eds., Reflection seismology: The continental crust: Washington, D.C., American Geophysical Union Geodynamics Series 14, p. 157–172.

Potter, C., Allmendinger, R., Burgess, S., Caruso, C., Gibbs, A., Hauge, T., Hauser, E., Huang, J., Klemperer, S., Liu, C-S., Nelson, K., Opdyke, S., Sanford, W., Serpa, L., Zheng, L., Zhu, T., Brown, L., Kaufman, S., and Oliver, J., 1984, The COCORP 40°N transect of the North American cordillera: Part 3: Geological Society of America Abstracts with Programs, v. 16, p. 626.

Priestley, K. F., Ryall, A. S., and Fezie, G. S., 1982, Crust and upper mantle structure in the northwest Basin and Range province: Seismological Society of America Bulletin, v. 72, p. 911–923.

Prodehl, C., 1977, The nature of the crust-mantle boundary beneath North America and Europe as derived from explosion seismology, *in* Heacock, J. G., ed., The Earth's crust: American Geophysical Union Geophysical Monograph 20, p. 349–369.

——— 1979, Crustal structure of the western United States: U.S. Geological Survey Professional Paper 1034, 74 p.

Ringwood, A. E., 1975, Composition and petrology of the Earth's mantle: New York, McGraw-Hill, 618 p.

Roy Chowdhury, K., and Hargraves, R. B., 1981, Deep seismic soundings in India and the origin of continental crust: Nature, v. 291, p. 648–650.

Sheriff, R. E., and Geldart, L. P., 1983, Exploration seismology, Volume 2: Data-processing and interpretation: Cambridge, England, Cambridge University Press, 221 p.

Smith, R. B., 1978, Seismicity, crustal structure and intraplate tectonics of the interior of the western Cordillera, *in* Smith, R. B., and Eaton, G. P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p. 111–144.

Smith, R. B., and Bruhn, R. L., 1984, Intraplate extensional tectonics of the eastern basin-range: Inferences on structural style from seismic reflection data, regional tectonics, and thermal-mechanical models of brittle-ductile deformation: Journal of Geophysical Research, v. 89, p. 5733–5762.

Sorensen, K., 1983, Growth and dynamics of the Nordre Stromfjord shear zone: Journal of Geophysical Research, v. 88, p. 3419–3437.

Speed, R. C., 1983, Pre-Cenozoic tectonic evolution of northeastern Nevada, *in* The role of heat in the development of energy and mineral resources in the northern Basin and Range province: Davis, California, Geothermal Resources Council, Special Report 13, p. 11–24.

Stauber, D. A., and Boore, D. M., 1978, Crustal thickness in northern Nevada from seismic refraction profiles: Seismological Society of America Bulletin, v. 68, p. 1049–1058.

Steinhart, J. S., 1967, Mohorovicic discontinuity, *in* Runcorn, K., ed., International dictionary of geophysics, Volume 2: Oxford, England, Pergamon Press, p. 991–994.

Stewart, J. H., 1978, Basin-range structure in western North America: A review, *in* Smith, R. B., and Eaton, G. P., Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p. 1–31.

——— 1980, Geology of Nevada: Nevada Bureau of Mines and Geology Special Publication 4, 136 p.

Stewart, J. H., and Poole, F. G., 1974, Lower Paleozoic and uppermost Precambrian Cordilleran miogeocline, Great Basin, western United States, *in* Dickinson, W. R., ed., Tectonics and sedimentation: Society of Economic Paleontologists and Mineralogists Special Publication 22, p. 28–57.

Thompson, G. A., 1972, Cenozoic basin-range tectonism in relation to deep structure: International Geological Congress, 24th, Montreal, Section 3 (Tectonics), Proceedings, p. 84–90.

Thompson, G. A., and Burke, D. B., 1974, Regional geophysics of the Basin and Range province: Annual Review of Earth and Planetary Sciences, v. 2, p. 213–238.

Trask, N. J., 1969, Ultramafic xenoliths in basalt, Nye County, Nevada: U.S. Geological Survey Professional Paper 650-D, p. D43–D48.

Waters, K. H., 1981, Reflection seismology: New York, Wiley Interscience, p. 453

Watts, A. B., ten Brink, U. S., Buhl, P., and Brocher, T. M., 1985, A multichannel seismic study of lithospheric flexure across the Hawaiian-Emperor seamount chain: Nature, v. 315, p. 105–111.

Wernicke, B., 1985, Whole-lithosphere normal simple shear: An interpretation of deep-reflection profiles in Great Britain, *in* Barazangi, M., and Brown, L., eds., Reflection seismology: The continental crust: Washington, D.C., American Geophysical Union Geodynamics Series 14, p. 331–339.

Zucca, J. J., 1984, The crustal structure of the southern Rhinegraben from reinterpretation of seismic refraction data: Journal of Geophysics, v. 55, p. 13–22.

Manuscript Received by the Society February 25, 1985
Revised Manuscript Received September 30, 1985
Manuscript Accepted November 20, 1985
Institute for the study of the Continents Contribution to Geology No. 5