Crustal structure of north-central Nevada: Results from COCORP deep seismic profiling

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ABSTRACT

Deep seismic-reflection data which were collected by COCORP in central Nevada at ~40°N latitude reveal a horizontally laminated, reflective lower crust and an upper crust containing prominent moderately dipping reflections from Cenozoic lava flows, basin fill, normal faults, older thrust faults, and Paleozoic sediments. Prominent Moho reflections are seen at depths of 31–35 km, in agreement with refraction-based Moho depths for this area. The strongly layered character of the lower crust is best developed near the mid-Miocene Northern Nevada rift, suggesting that Cenozoic magmatism and ductile extension have played an important role in the development of this fabric. East-dipping reflections beneath southern Grass Valley in Lander County may locally disrupt an otherwise continuous "reflection Moho."

INTRODUCTION

The COCORP deep seismic-reflection profiles in central Nevada, part of the COCORP 40°N Transect of the North American Cordillera (Allmendinger and others, 1987; Hauge and others, 1987; E. C. Hauser and others, unpub. data), lie near the Proterozoic–early Paleozoic rifted edge of North America and cross major structures related to Paleozoic convergence and Cenozoic extension. The survey thus provides a good opportunity to evaluate the relative importance of Proterozoic through Recent events in forming the crustal structure of central Nevada. The seismic data presented in this paper indicate that a strong overprint associated with Cenozoic extension and magmatism has nearly completely obscured older structures; this overprint consists of a layered middle and lower crust and very prominent Moho reflections. Half-grabens in which basin-fill reflections are truncated by range-bounding faults are the main features in a relatively unreflective upper crust.

TECTONIC SETTING

COCORP Nevada lines 6 and 7 (Fig. 1) traverse the westernmost reaches of early Paleozoic North America, where the Antler allochthon, an imbricate assemblage containing both oceanic rocks and North American strata, was emplaced above the North American passive margin along the Roberts Mountains thrust during the Devonian-Mississippian Antler orogeny (Roberts and others, 1958; Johnson and Pendergast, 1981; Murphy and others, 1984). There is no clear imprint of early to mid-Mesozoic compression related to Sonoman and Nevadan accretionary tectonics in this area. Rocks in the Cortez Range, just north of line 6 (Fig. 1), have been affected by Jurassic-Cretaceous folding (Stewart, 1980a; Ketner and Smith, 1974), which is probably related to plate convergence to the west (Speed, 1983).

Oligocene through Recent extension occurred in the area of the central Nevada COCORP lines. Smith (1984) documented Oligocene extension in the Toiyabe Range in the vicinity of line 7 (Fig. 1). Line 6 (Fig. 1) crosses the pre-basin-range mid-Miocene Northern Nevada rift (Zoback and Thompson, 1978; Oregon-Nevada lineament of Stewart and others, 1975). This rift is defined by a north-northwest–trending zone of 14- to 17-m.y.-old diabase dike swarms, basaltic andesite flows, and an associated short-wavelength linear magnetic anomaly (Zoback and Thompson, 1978). Its north-northwest trend has been overprinted by the northeasterly trending basins and ranges associated with post–10-m.y. basin-range extension (Zoback and others, 1981).

Seismic-refraction surveys in the vicinity of the central Nevada COCORP lines suggest crustal thicknesses of ~30 km in this area (Eaton, 1963; Stauber, 1983). These surveys also suggest that upper-mantle velocities are low (~7.8 km/s), similar to those measured by refraction surveys throughout the Basin and Range Province.

DATA ACQUISITION AND PROCESSING

Two east-west seismic-reflection lines, Nevada line 6 (48 km long) and Nevada line 7 (64 km long), constitute the central Nevada part of COCORP's 40°N Cordilleran transect (Fig. 1). The data were acquired in mid-September through early November, 1983, using the VIBROSEIS technique. Line 6 was shot using a 96-channel recording system and 68-m

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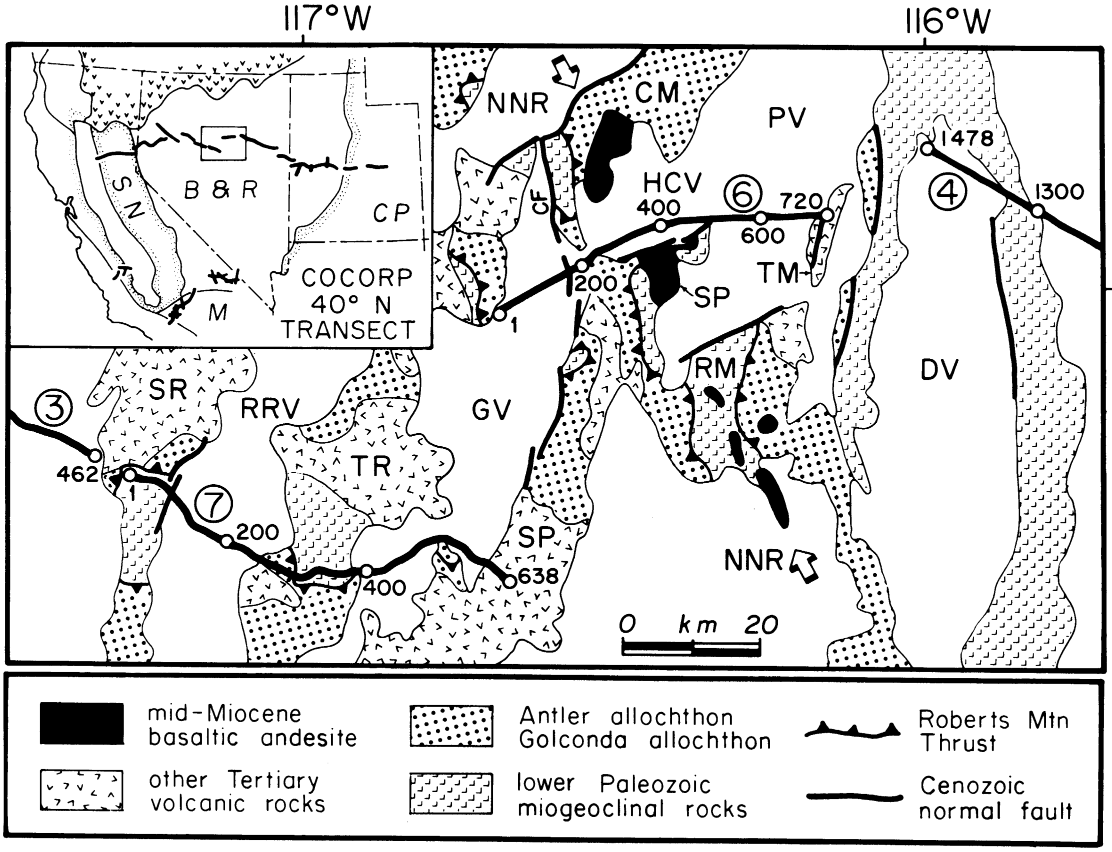

Figure 1. Location map showing the central Nevada part of COCORP's transect of the North American Cordillera at 40°N latitude. Circled numbers designate COCORP line numbers. NNR and large arrows indicate location of mid-Miocene Northern Nevada rift (after Zoback and Thompson, 1978). SR = Shoshone Range; TR = Toiyabe Range; CM = Cortez Mountains; SP = Simpson Park Range; RM = Roberts Mountains; TM = Table Mountain; GV = Grass Valley; RRV = Reese River Valley; PV = Pine Valley; HCV = Horse Creek Valley; CF = Cortez fault; DV = Diamond Valley. Inset: SN = Sierra Nevada; B&R = Basin and Range; CP = Colorado Plateau; M = Mojave.

station spacing; nominal near and far offsets were 272 and 6832 m, respectively. A 4-ms sampling rate was used, and record length was 14 s (~42 km depth). Four sweeps per station were vibrated for most of the line (8 sweeps per station on stations 148–272). Line 7 was shot using similar equipment and a 100-m station spacing; near and far offsets were 400 m and 9.9 km, respectively. The sampling rate was 8 ms, and the record length was 16 s (~48 km depth). On line 7, as well as on line 6, every station was vibrated along most of the line (terrain permitting), producing data of nominal 48-fold. The minor variations in field technique in different localities along this transect do not significantly affect the geological interpretation of the seismic data.

Data processing for these two seismic lines is summarized in Table 1 of Allmendinger and others (1987). Although automated migration was not performed, "hand migrations" were calculated for many of the dipping reflections and are included in the following presentation of the data.

CENTRAL NEVADA COCORP RESULTS

Lower Crust

The most striking aspect of COCORP data in central Nevada is the nearly horizontal layering in the lower crust evident in the western part of line 6 and the eastern part of line 7 (Figs. 2, 3, and 4). (The symmetric basinward dips of deep reflections shown on Figs. 2 and 3 beneath V.P.s 50–200 near Grass Valley on line 6 are an effect of velocity pull down related to low-velocity basin fill.) Farther east, a similar lower-crustal character is evident on the west end of COCORP Nevada line 4 (E. C. Hauser and others, unpub. data) and to the west on COCORP Nevada lines 1 and 2 (Hauge and others, 1987). Depths to the upper and lower bounds of this reflective zone can be calculated from their two-way traveltimes on the COCORP sections using an approximate upper-crustal velocity of 6.0 km/s and an average crustal velocity of 6.3 km/s, in accordance with results from refraction surveys in this part of the Great Basin (Eaton, 1963; Stauber, 1983). In central Nevada, the top of the reflective zone is at 5.5–6.0 s or 16.5–18 km. A well-defined band of 2- to 3-cycle reflections forms the base of the reflective zone at ~10.5 s, or 33 km, on the east half of line 7 (Fig. 4). On line 6, the basal reflections are less well defined but occur between 11 s (near the east end of the line beneath Pine Valley) and 12 s (near the west end of the line beneath Grass Valley, shown on Fig. 3). An approximate depth calculation which corrects for velocity pull down yields a depth of 35 km for these reflections beneath northern Grass Valley, which represents the deepest lower boundary of reflective lower crust observed by COCORP in Nevada.

The general agreement of the calculated depth to the lower bounding reflections on line 7 with the depth to the Moho determined from an intersecting refraction line (Stauber, 1983) suggests that these are Moho reflections. The widespread persistence of similar high-amplitude basal reflections on the COCORP Nevada transect (Klemperer and others,

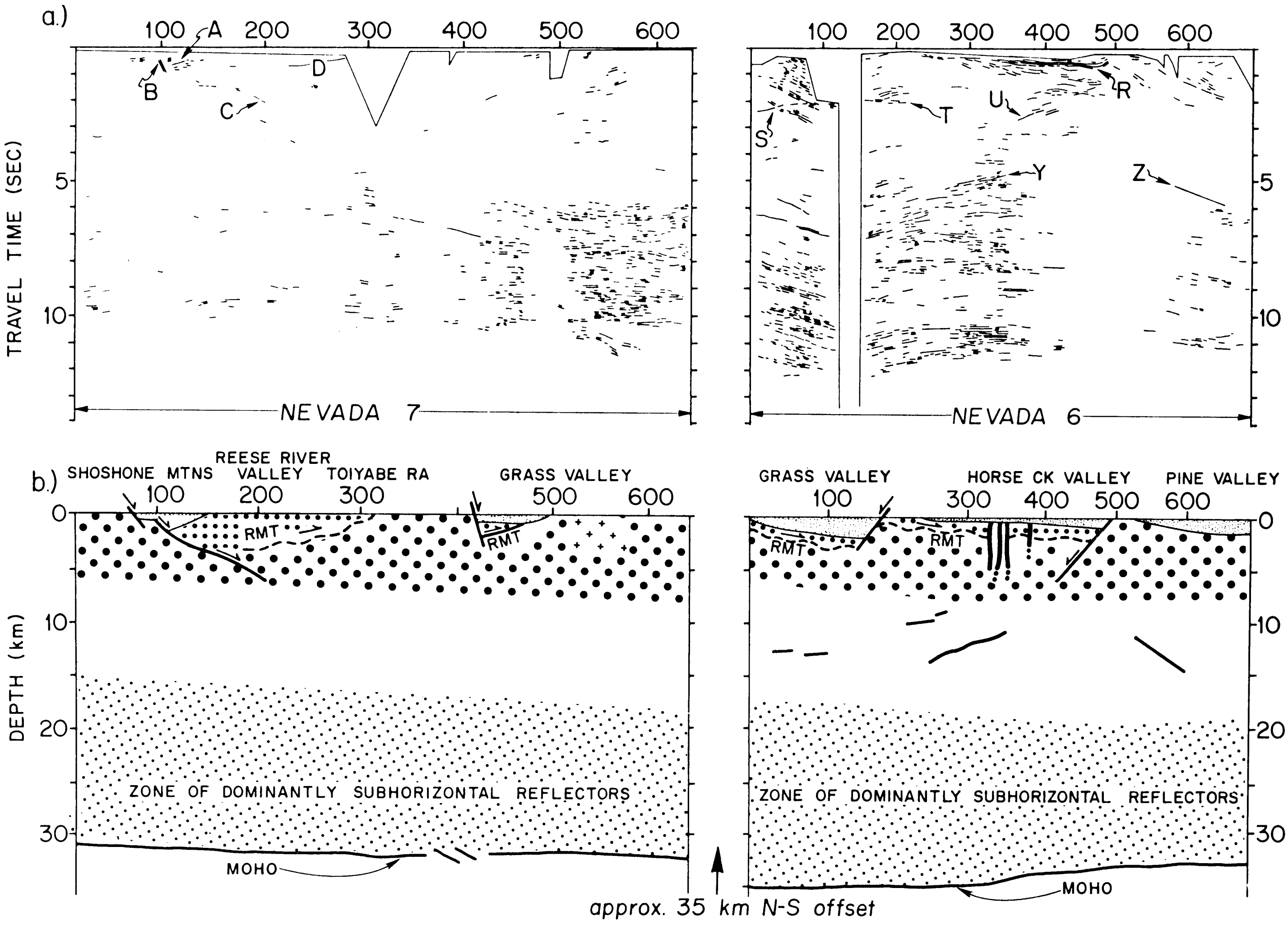

Figure 2. a. Line drawing of unmigrated stacked seismic sections of COCORP Nevada lines 6 and 7. Letters refer to features discussed in text. b. Interpretive depth section. In the upper crust, fine stipple = Cenozoic basins, + = Jurassic intrusion, small dots = Antler allochthon, and large dots = miogeoclinal strata. Shading in the lower crust denotes zone of largely subhorizontal reflections. RMT = Roberts Mountains thrust.

1986; Hauge and others, 1987; E. C. Hauser and others, unpub. data) supports their interpretation as the Moho.

On the east end of line 7 (V.P.'s 525–575), there is a reflection sequence that has an easterly apparent dip to a maximum time of ~12 s, beneath the prominent horizontal 10.4-s reflection interpreted as the Moho (Figs. 2a and 4). Hand migration, if an average crustal velocity of 6.3 km/s and an upper-mantle velocity of 7.8 km/s are assumed, places these reflections beneath V.P.'s 380–400 and dipping ~28° east. The migrated position of the deepest dipping reflections is 31 to 33 km deep, intersecting the Moho (Fig. 2b). There are no horizontal Moho reflections at the migrated position of this dipping sequence, and the dipping sequence may represent a local interruption in a reflection Moho which varies smoothly on a province-wide scale (Klemperer and others, 1986). Alternatively, it is possible that these dipping reflections originated outside the vertical plane of the seismic section, in which case, they represent features totally within the crust. Another persistent, low-amplitude east-dipping reflection occurs within horizontally laminated lower crust on line 7 between V.P.'s 300 and 425 at 6 to 7.5 s.

The best developed strongly layered fabric in the mid- to lower crust in the COCORP Basin and Range data appears beneath northern Grass Valley and Horse Creek Valley, on the west part of line 6 (V.P.'s 1–375; Fig. 3). The eastern part of this zone coincides with the Northern Nevada rift (Zoback and Thompson, 1978; Stewart and others, 1975), a zone of extensive mid-Miocene dike intrusion and volcanism. The especially prominent horizontal lower crustal layering thus may, at least in part, be related to mid-Miocene deep crustal intrusions.

On the east end of line 7 in southern Grass Valley (V.P.'s 500–636), high-amplitude reflections are evident on individual shot records at 7.6–8.0 s (two-way traveltime). Amplitude studies of these data demonstrate the existence of an amplitude anomaly which is prominent but of lesser magnitude than those associated with deep magma bodies on other COCORP surveys (Rio Grande rift: Brown and others, 1980; Death Valley: de Voogd and others, 1986). These events are the highest amplitude events in the strongly layered lower crustal sequence on the east end of line 7 (Fig. 4); the entire sequence between 6.0 and 10.5 is characterized by a plateau on amplitude-decay plots which steps down markedly at ~10.5 s (~31 km), as illustrated in Figure 6.

Upper Crust

The upper crust (0–5 s two-way traveltime; ~0–15 km depth) is characterized by tilted basinal sequences overlying gently to moderately dipping reflections. Basin-bounding faults are identified on the basis of termination of basin-fill and sub-basin reflections beneath the mapped positions of exposed normal faults; in one case (Reese River Valley), deeper, moderately dipping fault-plane reflections also occur. Upper crustal features are described below, from west to east.

Beneath Reese River Valley on line 7, west-dipping basinal sediments (Fig. 2a, event A) terminate along a reflection dipping steeply to the east, inferred to be a normal fault (B). The downdip continuation of this fault may be represented beneath the basin by a moderately east-dipping reflection (C). As shown on the time sections (Fig. 2a), the normal fault defined by B and C appears to have an abrupt change in dip. This is partly the result of velocity pull down beneath the Reese River Valley. The east dip of this fault contrasts with the pattern of west-dipping faults which control upper crustal extension beneath Antelope Valley and the Clan Alpine/Augusta Range to the west (COCORP Nevada line 3; Hauge and others, 1987). This change in pattern observed in the seismic data corresponds to a tilt-domain boundary along the Shoshone Mountains, identified by Stewart (1980b).

Gently dipping faint reflections beneath the Toiyabe Range on line 7 (D, Fig. 2a) project toward the mapped location of the Roberts Mountains thrust (Stewart and Carlson, 1976). Smith (1984), however, has reinterpreted the contacts between the Antler allochthon and autochthon in this area to be Oligocene normal faults which cut the Roberts Mountains thrust. The COCORP data show no clear indication of the Roberts Mountains thrust beneath the Toiyabe Range or Reese River Valley.

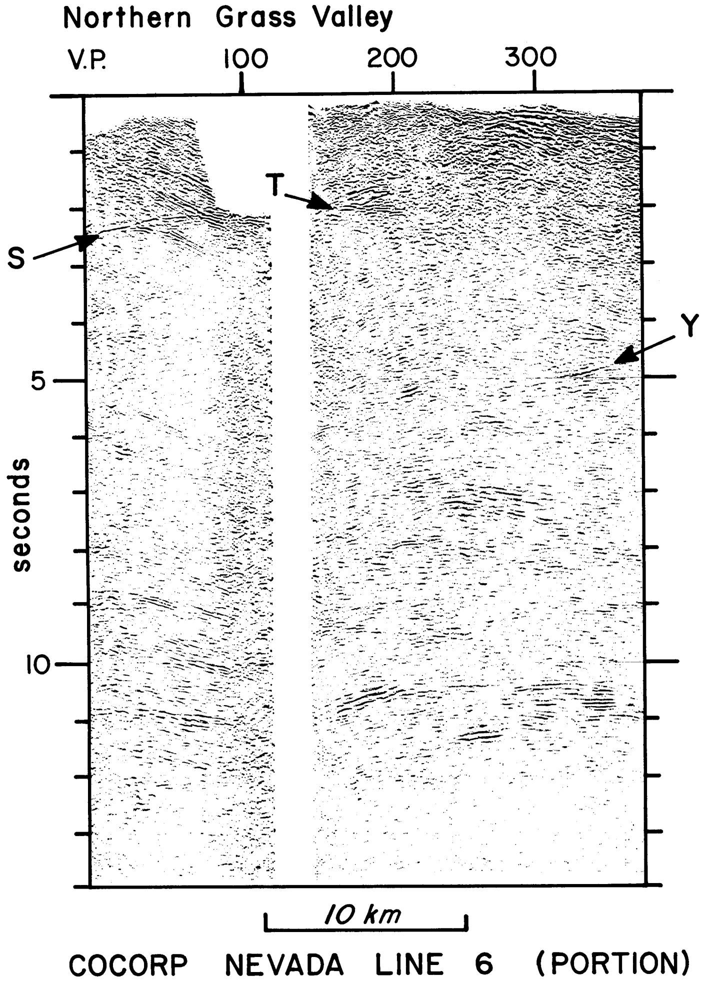

Figure 3. Unmigrated stacked seismic section of the western part of COCORP Nevada line 6. Vertical axis: two-way traveltime. See text for details.

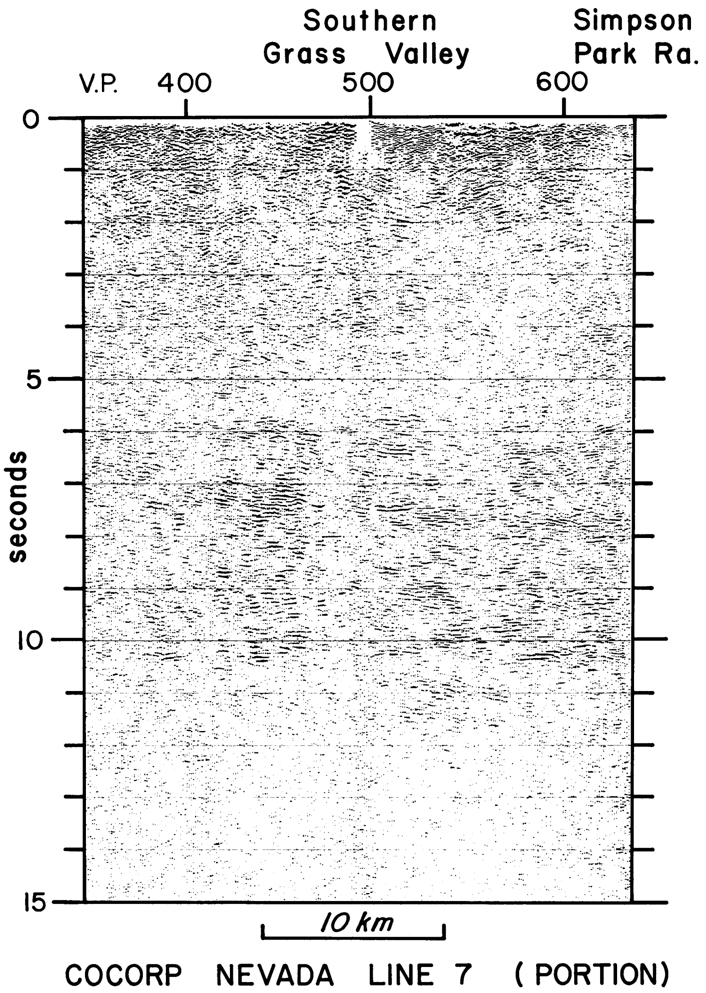

Figure 4. Unmigrated stacked seismic section of the eastern part of COCORP Nevada line 7. Vertical axis: two-way traveltime. See text for details.

The eastern two-thirds of line 7 is part of a 150-km-long northwest-trending transverse zone in central Nevada which is an across-strike tilt-domain boundary (Stewart, 1980b) and is characterized by the absence of major range-front faults. The COCORP data do not provide a clear image of a Tertiary basin in southern Grass Valley (Fig. 4), but gravity data which were collected along line 7 by the U.S. Geological Survey suggest that there is a shallow basin ~0.8 km thick beneath V.P.'s 400–500 (Chuchel and others, 1984).

West-dipping reflections at 1–2 s beneath V.P.'s 500–575 on line 7 (Figs. 2 and 4) may be from the base of a Jurassic granitic pluton (Stewart and Carlson, 1976) exposed in the basement prong just south of the seismic line.

Line 6 crosses a domain in which east-northeast–striking faults bound the northwest sides of east-northeast–trending ranges, such as the Cortez Mountains, Simpson Park Mountains, and Roberts Mountains (Fig. 1). North-northwest–striking faults, including the Cortez fault (Fig. 1), probably developed during mid-Miocene (14–17 m.y.) rifting to accommodate east-northeast–west-southwest–directed extension. The east-northeast–

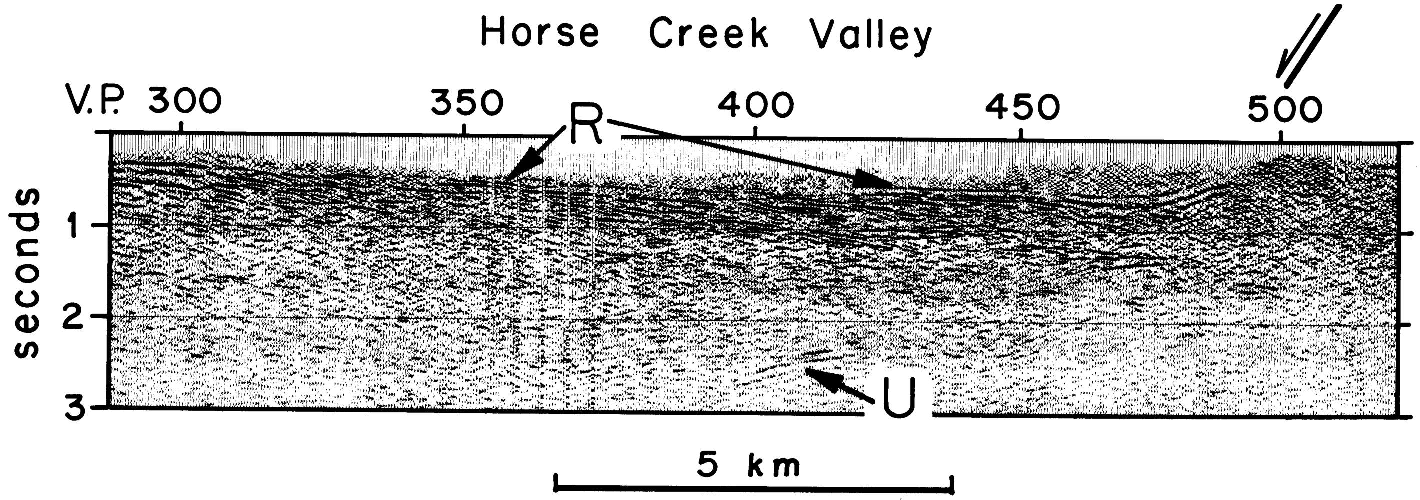

Figure 5. Unmigrated stacked seismic section of the shallow structure beneath Horse Creek Valley, COCORP Nevada line 6. Vertical axis: two-way traveltime. H:V = 1:1 for velocity of 2.5 km/s (approximate velocity of basin fill).

trending zones of weakness may also have developed at this time, possibly as transform faults (Zoback and Zoback, 1980). At some later date (post-13 m.y., pre-6 m.y.), major east-northeast range-bounding faults developed, and the two fault sets subsequently acted in concert during basin-range west-northwest–east-southeast–directed extension (Zoback and Thompson, 1978).

The west end of line 6 crosses the north end of Grass Valley; an east-dipping basinal sequence is apparent beneath V.P.'s 1–100 (Figs. 2 and 3). The reflections have an apparent dip of 11° to the northeast, toward the north-northwest–trending Cortez fault, which was not imaged because of a gap where no data were collected due to wet conditions. Mabey (1966) and Zoback (1978) presented gravity data suggesting that Grass Valley is an eastward-tilted asymmetric basin; this is consistent with the shallow reflection pattern.

A prominent gently west-dipping reflection (S on Figs. 2a and 3) is imaged at ~2 s beneath northern Grass Valley (line 6; V.P.'s 1–70). Because it passively crosses the east-dipping "grain" of pulled-down sub-basinal reflections, it is probably an out-of-plane reflection. Possible sources of this reflection include the folded Roberts Mountains thrust, exposed in a south-plunging antiform north of line 6 in the Cortez Range (Gilluly and Mazursky, 1965) or a dipping Paleozoic stratigraphic contact.

On line 6, beneath V.P.'s 175 to 225 at 1.6 to 2.0 s (Figs. 2 and 3), reflection T dips gently eastward and crosses the west-dipping fabric caused by velocity pull down beneath northern Grass Valley. Like reflection S, T is probably from slightly out of the vertical plane of the seismic section and could have originated from the Roberts Mountains thrust, Paleozoic sedimentary rocks, or a Cenozoic normal fault.

The basinal sequence beneath Horse Creek Valley has an easterly apparent dip (Figs. 2 and 5; line 6, V.P.'s 280–490). Gravity data (Mabey, 1966; Zoback, 1978) and locations of major range-bounding faults suggest that Horse Creek Valley is a southeast-tilted half-graben crossed by line 6 at a highly oblique angle. Miocene basaltic andesite flows capping the Cortez Range dip 5°–10° southeast (Gilluly and Mazursky, 1965). Down-dip projection of these flows from the Cortez Range suggests that the continuous east-dipping reflections beneath Horse Creek Valley on line 6 (Figs. 2a and 5; event R, at ~0.4 s at V.P. 300 and 0.6 s at V.P. 450) represent these flows. The depth-corrected apparent dips on event R and on faint shallower reflections from the basin sequence are 1°–2° east, consistent with true dips of 5°–10° southeast. Identification of the flow units as a mid-Miocene "basement horizon" provides a convenient marker for deformation that postdated the mid-Miocene Northern Nevada rift (Zoback and Thompson, 1978) and indicates that there is a maximum of ~1 km of basin fill which is younger than 14 m.y. in Horse Creek Valley. Reflections from the flows and sub-basinal east-dipping reflections, at least as deep as 2 s, are truncated on the seismic section (V.P. ~490; Fig. 5) at the northwest flank of a basement high which is the northeast subsurface continuation of the Simpson Park Range. This truncation is probably the seismic expression of the east-northeast–trending fault which bounds the northwest flank of the Simpson Park Range and projects into line 6 in this area. The apparent dip of this fault is ~50° west (60° northwest true dip); there is no evidence that its dip becomes shallower with depth. Reflections U, which have a gentle westerly dip as shown on the seismic section (Figs. 2 and 5; V.P.'s 375–420; 2.2 to 3.0 s), project toward the inferred position of the range-bounding fault below the bottom of the basin, but U occurs on a segment of line 6 which is nearly parallel to the fault and can most readily be interpreted as an out-of-plane reflection (sideswipe) from this fault.

To the east, the adjacent structural block (line 6; V.P.'s 490–700) is the northeastern continuation of the Simpson Park Range, another southeasterly tilted block. A narrow basement high occurs beneath southern Pine Valley where the basinal section is very thin, producing no reflections (Fig. 5; V.P.'s 490–520). Line 6 crosses this area at a high angle to geologic strike, and reflections dipping ~28° east (migrated dip, 32°) in the upper 2 s are apparent within the basement high and beneath Pine Valley (Fig. 2). These reflections may be interpreted as Tertiary volcanic units, Paleozoic sedimentary layering, or pre-Tertiary thrust faults. Basin fill in southern Pine Valley appears to dip east, as would be expected from the southeast dip of Tertiary volcanic rocks in the Simpson Park Range, and is ~1.2 s thick (~1.5 km). At the surface, this basin is bounded on its east side by a west-dipping normal fault which is exposed 2 km east of the termination of CDP coverage on line 6.

Two gently dipping reflections (Y and Z, Fig. 2) occur just above the horizontally laminated middle to deep crustal zone on line 6. Reflection Y (V.P.'s 290–360) dips westward at ~13° (Figs. 2 and 3) at traveltimes 4.8–5.2 s, corresponding to a depth of ~13 km. Y may originate within the lower Paleozoic or Proterozoic stratigraphic section or the underlying

Precambrian basement. Adequate stratigraphic constraints on depth to basement do not exist in this area (Stewart and Poole, 1974). Z is less prominent and dips 30° east (35° east migrated), at 4.8–6.2 s (~13–15 km), beneath V.P.'s 550–650 on line 6. This reflection appears to "sole into" flat reflections at ~6 s beneath V.P. 640–650 (Fig. 2a).

DISCUSSION

Detailed syntheses of regional geologic and geophysical data (Zoback and Thompson, 1978; Zoback, 1978) indicate that north-northwest–trending faults in the vicinity of line 6 (Fig. 1) were initiated during mid-Miocene rifting, accommodating east-northeast–west-southwest extension. The east-northeast faults may lie along zones of weakness which were initiated, possibly as transform faults, during mid-Miocene rifting (Zoback and Zoback, 1980). Both sets were active during later (post–10 m.y.; Zoback and others, 1981) basin-range extension, directed west-northwest–east-southeast. Simultaneous activity of these fault systems requires that the faults interact in a complex way at depth. Steep to moderately dipping basin-bounding faults crossed by line 6 have no obvious relationship to gently dipping deeper reflections beneath the basins; these faults may retain their steep dip at depth. Detailed conclusions regarding fault geometries and kinematics await three-dimensional seismic control, which could be achieved by additional profiling.

Sub-basinal reflections in the upper crust are difficult to interpret. As indicated above, there are several reflections beneath line 6 (S, T, Y, and Z) which cannot be traced to the surface, and so their identity is not well constrained. It is possible that both S and T are out-of-plane reflections from the folded Roberts Mountains thrust, but there is no way to prove this. West-dipping reflection Y (Figs. 2 and 3) from ~13 km depth probably originates within the Proterozoic miogeoclinal section or its basement. Y is probably too deep to be an Antler-age structure, as Antler-age deformation in this region is thin skinned (Roberts and others, 1958; Johnson and Pendergast, 1981; Murphy and others, 1984). It could conceivably be part of a Sevier-age thrust, resolved in the middle crust as distributed deformation or as a major detachment (Speed, 1982, 1983). The updip end of Y is directly beneath the projected position of the mid-Miocene dike swarm on line 6, and so Y may be an inclined magma feeder for the Northern Nevada rift. Y has no clear relationship to basin-range structure but may have accommodated midcrustal extension in conjunction with Z. Y and Z dip in opposite directions on either side of the subsurface basement high between Pine and Horse Creek Valleys and may form parts of the upper boundary on an extension-related midcrustal lens, similar to the model suggested by Hamilton (1982). The east-dipping reflection at 7–8 s directly below Y (Fig. 3) could form part of the lower boundary for such a lens.

Two independent reversed refraction studies suggest that the depth to Moho in the vicinity of line 7 is ~30 km (Stauber, 1983; Eaton, 1963). This agrees well with depth calculations for the base of the deep reflective sequence (at 10.0–10.5 s; ~31–33 km) on line 7 (Figs. 2 and 4). These basal events, and those at 11–12 s on line 6 (Figs. 2 and 3), are herein considered to be the "reflection Moho" also defined and identified by Klemperer and others (1986) beneath a large part of the COCORP 40°N Cordilleran transect. Although the Moho reflections in central Nevada fit most of the general observations on the reflection Moho described by Klemperer and others (1986), there are at least two ways in which they do not fit the general observations. First, Klemperer and others (1986, p. 604) defined the reflection Moho as "the deepest, high-amplitude, laterally extensive reflection or group of reflections . . . present at traveltimes (depths) approximately commensurate with other estimates of crustal thickness," and they added the observation that "the reflection Moho may commonly separate highly reflective lower crust from largely reflection-free upper mantle." Moho reflections beneath the west part of Nevada line 6 clearly separate highly reflective lower crust from a nearly reflection-free upper mantle, but these Moho reflections are of lesser amplitude than are many of the slightly shallower reflections in the lower crust and thus do not stand out as high-amplitude reflections. The second important exception to Great Basin reflection-Moho style described by Klemperer and others (1986) occurs beneath Nevada line 7, where the migrated position of deep east-dipping reflections (beneath V.P.'s 380–400, Fig. 2b) appears to interrupt an otherwise continuous and smoothly varying Moho. As mentioned above, there is a possibility that these dipping reflections originated outside the vertical plane of the seismic section, in which case they are crustal reflections and do not suggest Moho complexity.

Figure 2b illustrates significant variation in depths to both the upper and lower (= Moho) bounds to reflective lower crust on lines 6 and 7. This is most easily seen by comparing the depth section for the east end of line 7 with the adjacent section for the west end of line 6. There is a 35-km north-south offset between these two lines, and in that interval, there is a 1–4 km difference in the depth to Moho (34–36 km on west end of line 6; 32–33 km on east end of line 7) and a 1–3 km difference in the upper bound to reflective lower crust (15–16 km, west end of line 6; 17–18 km, east end of line 7). Additional seismic profiling would be necessary for the

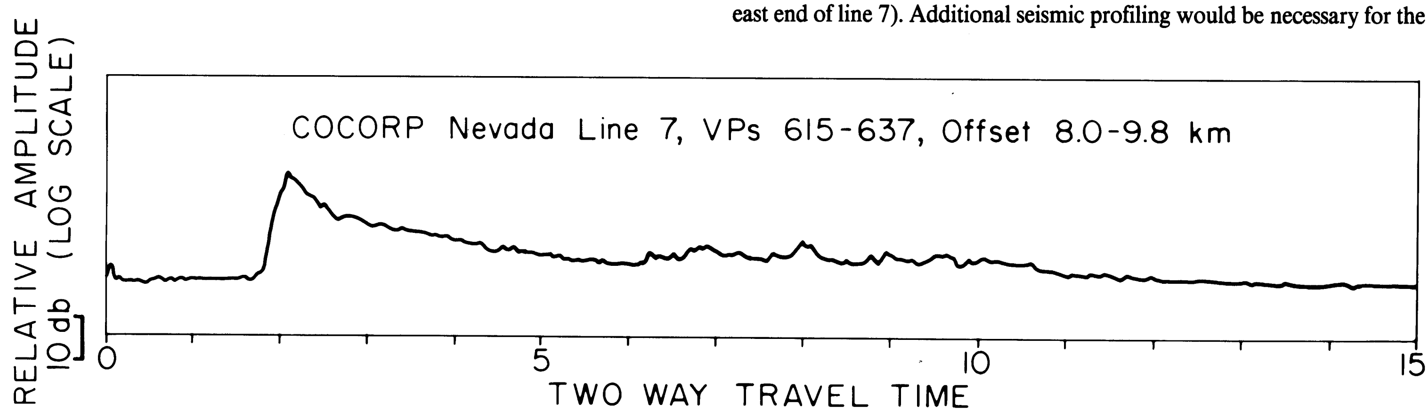

Figure 6. Plot illustrating relative amplitude for the far traces (shot receiver distances 8.0–9.8 km) on 23 consecutive shot records beneath southern Grass Valley on COCORP Nevada line 7. Note the "plateau" on the amplitude decay curve in the zone of reflective lower crust between 6 and 11 s two-way traveltime and the amplitude maximum at ~8 s two-way traveltime.

definitive three-dimensional control, in order to distinguish between smooth transitions and abrupt offsets in depth. No abrupt offsets in these boundaries were observed elsewhere in the Great Basin along the COCORP 40°N Transect. The eastern part of line 7, however, lies along a major transverse tilt-domain boundary identified by Stewart (1980b) whereas line 6 lies entirely north of this boundary. It is possible that the difference in Moho depth between lines 6 and 7 is related to the transverse tilt-domain boundary. This cannot be tested using the present data set, as no seismic lines on the COCORP 40°N Transect cross well-defined transverse tilt-domain boundaries (Stewart, 1980b).

Both Eaton (1963) and Stauber (1983) found evidence for a midcrustal, wide-angle reflection at depths of 15–20 km in this part of central Nevada. This reflection may correspond to the top of the horizontally layered, reflective mid- to lower-crustal sequences on lines 6 and 7. An upper-crustal velocity of 6.0 km/s (above this midcrustal wide-angle reflector) and lower-crustal velocity of 6.6–6.7 km/s (below the reflector) were determined by both Eaton (1963) and Stauber (1983). This is compatible with a model developed below for the existence of numerous mafic intrusions in the reflective middle and lower crust.

Zoback and Thompson (1978) presented data suggesting that the pre–basin-range Northern Nevada rift, which intersects line 6 at about V.P.'s 350–400, was part of a 700-km-long north-northwest–trending mid-Miocene (14–17 m.y.) rift zone. To the north, this rift is defined by feeder dikes to Columbia River basalts in southwest Washington, northwest Oregon, and west Idaho (Swanson and others, 1975); the graben of the western Snake River Plain; and volcanic rocks marking the mid-Miocene position of the Yellowstone hot spot (Armstrong and others, 1975). The bimodal, but predominantly basaltic, composition of mid-Miocene volcanic rocks associated with this rift and its long linear trend indicate that it was a discrete, mantle-tapping feature. Its principal seismic expression appears to be the prominent mid- to lower-crustal subhorizontal layering on the west part of line 6. Stauber and Zoback (1979) and Stauber (1983) presented arguments for excess mass in the lower crust along the Northern Nevada rift. Their conclusions were based on gravity modeling of independently determined crustal thicknesses, which if compared to a smoothed observed Bouguer field, result in a broad residual gravity high along the rift. This suggests that the reflective character of the middle and deep crust on the west half of line 6 may be due, at least in part, to the presence of high-density mid-Miocene lower-crustal intrusions. The detailed sequence of formation of the horizontal layering may involve vertical movement of mafic magma in lower-crustal dikes, resulting in heating and anatexis of silicic lower crust, followed by lateral spreading of mafic magma beneath a viscous plug of silicic melt (Lachenbruch and Sass, 1978). Introduction of basaltic material into the lower crust would also have the effect of creating softened zones, which would become subhorizontal ductile high-strain zones accommodating lower-crustal extension and in turn aiding in further horizontal transport of basaltic melt. The horizontal middle- and lower-crustal fabric may thus be the result of a complex interplay between intrusion and ductile deformation. According to this model, some of the mafic magma would manage to continue to ascend, forming shallow dikes and feeding volcanoes, along a zone of weakness defined by the Northern Nevada rift.

CONCLUSIONS

The central Nevada COCORP lines demonstrate a strong dichotomy between deep crustal and shallow crustal reflection character. The horizontally layered, reflective fabric in the lower crust probably records pervasive ductile deformation and intrusions. There is no strong evidence in the COCORP data for the existence of molten zones in the lower crust in this region; if such zones are present, their dimensions are too small to be detected given the profiling parameters used. The presence of volcanic rocks as young as 14 m.y. in the vicinity of the COCORP survey suggests that intrusions did play a role during earlier Cenozoic crustal evolution.

The reflection Moho seen in central Nevada and elsewhere on the COCORP 40°N Transect must be a young feature, as it is not offset beneath Cenozoic normal faults and varies smoothly beneath a province which has a complex pre-Tertiary compressional history (Hauge and others, 1987; E. C. Hauser and others, unpub. data; Klemperer and others, 1986). Similarly, the horizontally layered lower-crustal fabric is apparently not offset by Cenozoic normal faults. The reflection Moho and layered lower crust thus formed during Cenozoic extension, or they represent older features (related to ductile deformation during Proterozoic/early Paleozoic rifting, or Paleozoic/Mesozoic convergence) which were significantly reactivated and strongly overprinted during Cenozoic extension.

If the lower crustal fabric is a Cenozoic extensional feature, it can be viewed as having developed along symbiotically related networks of intersecting intrusions and high-strain zones. In this view, intrusions were localized along high-strain zones which aided magma migration; heat from the intrusions produced further softening and strain localization in discrete zones. This model is supported by the enhanced development of the horizontal lower crustal fabric in the area of the mid-Miocene Northern Nevada rift. Fountain and Salisbury (1981) discussed the existence of numerous ductilely deformed and transposed mafic and ultramafic intrusions in several exposed lower crustal sections, notably the Ivrea zone. These may have formed in the same general way as the model which we are suggesting.

The upper crust in central Nevada is relatively nonreflective but appears to contain discrete dipping reflections from Cenozoic sediments, lava flows, and normal faults; Paleozoic sediments; and, possibly, pre-Tertiary thrust faults. Range-bounding normal faults cannot be traced to midcrustal depths in this region. The interface between this upper crustal character and the horizontally layered lower crust lies at depths of 16–18 km and probably corresponds to the midcrustal, wide-angle reflection documented in the Eureka area by Eaton (1963).

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