Missing roots and mantle "drips": Regional P_n and teleseismic arrival times in the southern Sierra Nevada and vicinity, California

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Previous seismological studies have placed the source of the uplift of the Sierra either within the crust, suggesting a Mesozoic age for the source of the uplift, or in the upper mantle, consistent with late Cenozoic creation of the buoyant material producing the uplift of the range. We deployed 16 temporary seismometers in the high part of the southern Sierra Nevada to augment the permanent Southern California Seismic Network and record arrivals from regional and teleseismic earthquakes. Arrival times of P waves from 54 teleseisms recorded at these stations are advanced by over a second by a high-velocity body in the upper mantle west and northwest of Lake Isabella. Inversion of the arrival times indicates that this "Isabella anomaly" is of limited north-south extent (about 40-60 km), has compressional velocities about 4-5% higher than its surroundings, and probably extends from about 100 to 200 km depth. The limited north-south extent of the "Isabella anomaly" indicates that it is unrelated to the Sierra; we speculate that it is the downgoing part of a small scale convection system similar to that inferred beneath Southern California. This inversion does not clearly reveal either a large crustal root or a substantial low-velocity body in the upper mantle beneath the Sierra. Although the presence of either degrades the fit to the arrival times and requires high-velocity material beneath the low-velocity material of either root, Bouguer gravity anomalies require low-density material under the Sierra. Assuming that arrival times from earthquakes 150-350 km north and south from the southern Sierra come from a common refractor (the one-layer structure), the upper mantle P wave velocity (P_n) beneath the High Sierra is about 7.6-7.65 km/s; if the arrivals from north and south are from different refractors (the two-layer structure), material with a P wave velocity greater than ~7.2 km/s (the "7.x" layer) would lie under a nearly flat interface from more normal crustal velocities and be separated by a north-dipping interface from underlying mantle with velocities about 7.9-8.1 km/s. The P_n velocity beneath the region immediately to the east is significantly greater (7.9-8.0 km/s) than that of material at equal depths under the Sierra. For the one-layer structure, further assuming that mean crustal velocities are uniform along north-south lines, we find little dip on the Moho in the area; using the arrival times from earthquakes to the south, we infer a depth of 33 ± 5 km for the Moho beneath the southern High Sierra. This structure of a thin to normal crust over a low-velocity mantle can be reconciled with earlier observations that were used to infer a thick crust under the Sierra. By considering the Bouguer gravity anomaly, the surface geology, refraction profiles in this region, and our own observations, we suggest that 1/3 to 1/2 of the modern elevation in the range is supported by lateral (east-west) density contrasts in the crust; the remainder is supported by density contrasts in the uppermost mantle or lateral variations in the thickness of the "7.x" layer. Our interpretation is that the southern Sierra overlies mantle lithosphere that has been thinned and warmed in response to regional lithospheric extension in Neogene time. This part of the upper mantle might have provided the melt that migrated to the east and produced volcanics in the southwestern Great Basin; depletion of the upper mantle might have increased the seismic velocity and decreased the density of material about 60-100 km beneath the southern Sierra.

INTRODUCTION

Most of the relief on the Earth's surface can be understood in the framework of plate tectonics. However, the causes of deformation in areas such as the Cordillera of the western United States remain enigmatic because significant uplift has occurred without

compressive tectonism and its attendant thickening of the crust. Some of the most spectacular deformation in the western Cordillera is in the southwestern Great Basin, where Death Valley, the lowest point in North America, lies about 100 km from the southern High Sierra Nevada, which contains the highest peak in the western United States. The recent and continuing deformation in this region provides an opportunity for detecting geophysical signatures directly related to this deformation and not masked by subsequent deformation. In this paper we report the results of a passive seismic investigation of the structure of the crust and upper mantle of the southern Sierra Nevada and adjoining areas and discuss the implications for deformation processes in this region.

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Geology

Physiographically dominating its surroundings, the Sierra Nevada is unusual in many respects (Figure 1). Although voluminous intrusions formed the range's crust in the Mesozoic, the modern day heat flow observed over most of the range is well below the global average [e.g., *Lachenbruch and Sass, 1977; Saltus and Lachenbruch, 1991*], though some portions of the range have heat flows more characteristic of the Basin and Range to the east [*Saltus and Lachenbruch, 1991*]. In the tectonic chaos of the Cenozoic western United States, the Sierra has been relatively undisturbed by internal faulting and has risen largely as a block. The >2 km uplift of the range relative to sea level was accomplished not by contractional strain or voluminous volcanism [*Christensen, 1966; Huber, 1981*] but within a tensile stress regime [e.g., *Jones and Dollar, 1986; Lockwood and Moore, 1979*]. Although lacking throughgoing active faults, the southern Sierra is among the most seismically active regions of California

[*Jones and Dollar, 1986*]. In short, the range is full of contradictions.

The bedrock geology of the range has been examined extensively over the past few decades. Two facts are of primary importance to this study: (1) intrusive rocks grade from very silicic to ultramafic from east to west across the range [*Oliver et al., 1987, 1993; Saleeby, 1981*], and (2) middle to lower crustal plutonic rocks crop out in the southernmost Sierra and Tehachapi Mountains [*Pickett and Saleeby, 1993; Ross, 1989; Sharry, 1981*]. The former relation probably reflects the presence of a late Paleozoic or early Mesozoic suture between continental and oceanic basement in the western foothills of the Sierra [*Saleeby, 1981*]. The cause of the latter observation remains unknown, but the unroofing of the southernmost Sierra must have been essentially completed before onlapping sediments were deposited in the Eocene [*Goodman, 1989; Goodman and Malin, 1992; Nilsen, 1987; Reid, 1988*].

Regional Earthquakes Producing P_n Arrivals and older refraction lines indicating a thick crust

Fig. 1. California and Nevada with the locations of seismic profiles used in inferring a thick crust in the southern Sierra Nevada. Small dots are temporary stations deployed in the summer of 1988; outlined area is shown in Figure 2.

Tectonic subsidence seems to have occurred throughout the San Joaquin Valley from this time forward, though apparently most of the subsidence in the southernmost San Joaquin Basin occurred during extension in the early Miocene [Goodman, 1989] while subsidence farther north apparently continued or accelerated throughout the Neogene [Moxon and Graham, 1987; Rentschler and Bloch, 1988]. This Neogene subsidence in the main San Joaquin Basin (north of Bakersfield) has been attributed to the passage of the Mendocino Triple Junction [Graham and Williams, 1985; Zandt and Furlong, 1982] and thrust loading of the western San Joaquin Valley accompanied by weakening of the lithosphere in the San Joaquin Valley and Sierra Nevada [Rentschler and Bloch, 1988].

Because relief across this region exists in near isostatic equilibrium [Oliver and Robbins, 1982], a mass deficit must exist under the Sierra that is large enough to isostatically support the range. The two principal hypotheses presented to date are that the range is compensated by an Airy-type crustal root [e.g., Lawson, 1936] or that it is underlain by low-density mantle [e.g., Crough and Thompson, 1977].

Despite decades of study, the location of the mass deficit compensating the southern Sierra Nevada remains controversial (see Jones [1987] for a more detailed summary). Proponents of a crustal root point to two longitudinal seismic refraction profiles along the range [Eaton, 1966; Pakiser and Brune, 1980], the likely existence of thick crust beneath the Mesozoic arc, and gravity profiles constructed using constraints from the seismic profiles [Bateman and Eaton, 1967; Oliver, 1977]. Those proposing a low-velocity upper mantle note observations of low seismic velocities from a teleseismic study extending north from Mono Lake [Mavko and Thompson, 1983], surface wave phase velocities [Crough and Thompson, 1977], and low P_n velocities observed on refraction profiles across the range [Carder, 1973]. Although the two hypotheses need not be mutually exclusive, the interpretations of the several refraction profiles (the most direct observations of the lower crust and upper mantle in the region) have differed radically and are discussed more below.

Previous Seismic Investigations

The strongest evidence for a thick crust under the Sierra Nevada comes from an active refraction profile [Eaton, 1966] and an unreversed passive profile [Pakiser and Brune, 1980] along the axis of the range (Figure 1). Eaton originally identified a P_n phase at both ends of the profile with apparent velocities of 7.7 km/s near Mono Lake and 8.1 km/s near China Lake. These velocities and the travel time intercepts led him to infer that the Sierran crust was more than 50 km thick and the P velocity of the underlying mantle was 7.9 km/s. Prodehl [1979] reanalyzed the same seismograms and reached a somewhat different conclusion in part because he could identify neither the P_n phase near China Lake nor phases from a high-velocity lower crust; the difference produced a thinner crust (33 km near China Lake) and an undetermined P_n velocity. Using arrivals from the 1966 Truckee earthquake ($M = 6.5$) and its aftershocks recorded at horizontal-component sensors placed within the range, Pakiser and Brune [1980] made the only measurements of seismic velocities along the axis of the range (Figure 1). This sparsely recorded, unreversed profile produced apparent P_n velocities of 7.7 km/s as far south as their station NEL (Figure 2), but they interpreted the observed apparent P_n velocity of 8.6 km/s between NEL and ISA (Figure 2) as reflecting a southward thinning of the crust. Thus Pakiser and Brune interpreted their results as manifestations of a thick (50–55 km) crustal

root, thinning south of about 36°30'N, above mantle with a P_n velocity of 7.9 to 8.0 km/s.

One alternative to a thick crust is a normal crust overlying low-velocity upper mantle; this structure is largely based on seismic refraction profiles transverse to the range. These profiles used nuclear explosions at the Nevada Test Site (NTS) as sources; one profile near 37°N was partially reversed by an earthquake in Monterey Bay [Carder, 1973; Carder et al., 1970]. Arrivals in the western Sierran foothills were too early to be consistent with a crustal root; this led Carder [1973] to infer that the Sierran crustal thickness was normal to slightly thin and that low-velocity (7.6–7.7 km/s) upper mantle was present and was, in a sense, a seismic root to the Sierra. Although this interpretation was disputed by Pakiser and Brune [1980], later reanalysis supported Carder's interpretation [Bolt and Gutdeutsch, 1982], though that analysis failed to resolve the discrepancy between the transverse and longitudinal profiles.

Other seismic studies generally support the presence of low-velocity mantle but are not conclusive. Low phase velocities of surface waves traveling down the Sierra led Crough and Thompson [1977] to suggest a low-velocity upper mantle, but these surface waves might reflect velocities beneath both part of the volcanic Cascade Range along the path and the Basin and Range just east of the path. Mavko and Thompson [1983] found that teleseismic arrival times observed north from Mono Lake were consistent with lower P wave velocities in the mantle beneath the higher parts of the Sierra, but their network was too sparse to resolve the depth of the anomalies they inferred. Previous inversions of teleseismic arrivals in southern California produce structures beneath the Sierra with very limited resolution, but these inversions have tended to show the presence of higher velocities in the mantle beneath the Sierra than elsewhere in southern California [Humphreys et al., 1984; Raikes, 1980].

Faced with the ambiguous and conflicting results of earlier studies, we attempted to locate the mass deficit supporting the Sierra Nevada by using a temporary network of seismic stations in the southern Sierra that augmented the permanent Southern California Seismic Network. Observations of teleseismic P wave arrival times and upper mantle P wave refractions from regional earthquakes are evaluated to constrain the thickness of the crust and the velocity structure of the upper mantle. These efforts also encompass attempts to constrain the location of high-velocity material previously noticed. Reconciling our observations with previously gathered data indicates that roughly half of the elevation of the Sierra is supported by density variations within the crust and that the remaining half (representing the difference in elevation between the High and southernmost Sierra) comes from density variations in the upper mantle or, perhaps, the existence of an exceptionally high-velocity lower crustal root.

DATA COLLECTION

We deployed one Kinematics DR-200, six "LBS" ("Land Based Seismometer," a prototype instrument designed as a terrestrial equivalent to ocean bottom seismometers (OBSs)), and 10 Kinematics DR-100 seismic recorders in the southern Sierra Nevada in the summer of 1988; all but one (station pmt) recorded all three components of ground motion (Table 1 and Figure 2). Sample rates varied from 25 to 240 samples per second (sps). The temporary network spanned the entire width of the southern topographic margin of the High Sierra with a typical spacing of 15 km (Figure 2). This spacing and the network's width were designed to distinguish changes in the thickness of the crust from lateral ve-

TABLE 1. Locations of Temporary Stations

Station	Latitude (N)	Longitude (W)	Elevation, m	Station Delay, s	Seismometer/Recorder
arc	36° 00.74'	118° 19.91'	2614	-0.02	three L4s/DR100
bar	36° 31.17'	118° 31.68'	2972	-0.04	L4-3D/LBS
bkr	35° 50.78'	118° 31.32'	2265	-0.02	S6000, then L4-3D/DR100
bla*	36° 05.79'	118° 16.62'	2556		three L4s/DR100
blb*	36° 05.79'	118° 16.62'	2556		L4/DR100
bmd	35° 52.38'	118° 19.33'	2454	-0.01	three L4s/DR100
ccr	36° 36.40'	118° 45.36'	2195	0.01	Ranger, then three Rangers/DR100
cdg	36° 47.32'	118° 40.48'	1494		S6000/DR100
cha	36° 27.51'	118° 24.51'	2082	-0.03	L4-3D/LBS
cou	36° 21.92'	118° 17.66'	2731	0.07	L4-3D/LBS
jhn	35° 57.85'	118° 32.62'	1460	-0.01	three L4s/DR100
jun	36° 34.88'	118° 24.67'	2510	0.00	L4-3D/LBS
mnk	36° 27.26'	118° 36.40'	2362	-0.01	Ranger, then L4-3D/DR100
ndl	36° 05.81'	118° 28.88'	1829	-0.05	three L4s/DR100
pmt	36° 01.69'	118° 11.61'	2256	0.02	L4/DR100
twr	36° 20.87'	118° 24.75'	2109	0.02	L4-3D/LBS
wil	36° 14.27'	118° 25.00'	2009	0.03	L4-3D/LBS
wmd	36° 11.94'	118° 34.52'	2615	0.00	three Rangers/DR100

* These trial stations were deployed during only the first week of the experiment.

locity variations in the mantle. Because the depth of a velocity anomaly is constrained by rays traveling in different directions through the same space (Figure 3), we anticipated that our combination of station spacing, array size, and teleseism locations would permit us to resolve the extent of velocity anomalies between about 20 and 150 km depth.

All of the temporary stations were deployed on crystalline bedrock of the Mesozoic Sierra Nevada batholith, which produced excellent signal-to-noise ratios (Figure 4). This allowed the event-triggered temporary instruments (except the DR-200) to record a total of 300 teleseisms, 136 of which were recorded by five or more stations. More than 1000 regional and local earthquakes

were also recorded. The internal clocks on all instruments were calibrated to WWVB radio with an estimated accuracy of about 10 ms. Electronic records from the permanent Southern California Network were preserved for over 400 teleseisms detected either by key stations in the Southern California Network or by the temporary network. All seismograms of the permanent network were timed relative to WWVB and were sampled at 62.5 or 100 sps.

Vertical Component Seismograms

Sea of Okhotsk, $z=645$ km
14 August 1988, origin 10:56:57.6 Z

Fig. 4. Seismograms from the temporary (top four traces) and permanent arrays. Stations cou and wil used land based seismometer (LBS) recorders with the low-end boost of Roberts [1989]. Tic marks on seismograms denote predicted Jeffreys-Bullen arrival times. Horizontal axis is time after the picked first arrival.

Fig. 3. Cartoon illustrating a cross section through anomalous rays from buried velocity anomalies that would be produced by (a) a simple, roughly spherical anomaly and (b) an anomaly elongated with depth. Darker rays are more anomalous (have greater travel time residuals). Note that the smoothest horizontal gradients in travel time anomaly occur at the mean depth of the anomalous body. The patterns of circles at given depths illustrate the idea behind the "pauper's tomography" in Plate 2.

Fig. 5. Teleseisms analyzed in this paper. Azimuthal equidistant projection centered on the southern Sierra Nevada; outer limit of the plot is 100° from the center; inner circles are at 30° increments. Size of the circles is in proportion to the number of stations recording the earthquake.

We used a low-pass filter with a corner frequency of either 5 or 10 Hz in the amplifier of the DR-100 recorders to prevent aliasing and improve the detection of teleseisms. The DR-100s' seismometers all had a 1-Hz natural frequency except for the 2-Hz S-6000. The LBS recorders utilized L4-3D seismometers but contained a circuit to amplify low frequencies, providing an effective frequency response down to 0.1 Hz [Roberts, 1989]. Stations of the permanent network used L4 seismometers with a 30-Hz low-pass antialiasing filter.

A subset of 54 of the teleseisms were selected for this analysis. These events provide good azimuthal coverage and avoid the strong azimuthal bias of the complete set of teleseisms (Figure 5). Seismograms from each event were timed by eye relative to the trace with the best signal-to-noise ratio through the use of an interactive code that superimposed each trace in turn on the reference trace. Each arrival was assigned an integral quality code from 0 to 4 that corresponds to an estimated uncertainty of 0.05, 0.1, 0.2, 0.3, and 0.4 s, respectively; there were 676 quality 0 picks, 316 quality 1, 157 quality 2, 106 quality 3, and 72 quality 4 picks. The waveforms recorded by the permanent stations and the DR-100s are all nearly identical for at least the first second after the *P* arrival, greatly improving the quality of our picks of arrival times compared with picks made without reference to a common waveform. To permit direct comparison with the DR100 waveforms, the recordings from the LBS instruments were filtered using a causal three pole Butterworth filter with a passband of 0.5 to 20 Hz. This passband yielded seismograms very similar to those from the other instruments. To determine if the use of this filter biased the arrivals, we compared arrival times derived from the filtered seismograms for some of the more impulsive earthquakes

to times picked directly from the unfiltered seismograms. No significant difference between the two sets of arrival times was found.

QUALITATIVE DISCUSSION OF THE TELESEISMIC ARRIVALS

Residuals calculated from the arrival times of teleseisms clearly demonstrate the presence of several large velocity contrasts within and near the southern Sierra; locations of the most significant anomalies can be inferred from a simple examination of these residuals. These residuals are calculated by first determining variations in travel times below a depth of 270 km using Herrin travel time tables [Herrin, 1968], then by calculating the travel times through a local one-dimensional (1-D) structure that combines the upper 240 km of the mantle structure of the Jeffreys-Bullen (J-B) velocity structure [Jeffreys and Bullen, 1940] with a 30 km thick crustal structure modified from structure A of Jones and Dollar [1986] for the southern Sierra (Table 2). The J-B

TABLE 2. One-Dimensional Seismic Velocity Structure of the Southern Sierra

<i>P</i> Wave Velocity, km/s	Top of Layer (km below mean sea level)
5.80	- 3
6.25	0
7.80	30
7.92	70
8.03	110
8.16	150
8.29	190
8.42	230

structure approximates the upper mantle structure under southern California without resorting to poorly resolved low-velocity zones inferred in regional studies [Burdick and Helmberger, 1978; Iyer and Hitchcock, 1989; Walck, 1984, 1985]. Rays were traced to the stations at their proper elevation.

To remove systematic variations due to local, upper crustal velocity perturbations, we derive station corrections from the arrival times. Usually we would average the weighted travel time residuals calculated at each station to obtain the station correction. The large anomalies (~ 1 s) clearly related to long-wavelength sources in the upper mantle (discussed below) would tend to bias these corrections from those solely reflecting local upper crustal velocity perturbations; therefore we have instead averaged the residuals remaining after a single pass of a least squares inversion designed to remove the effect of smooth, long-wavelength structures. This

appears to have succeeded as the resulting corrections lack long-wavelength patterns (Figure 6). Corrections are near zero for the temporary stations, which were on crystalline bedrock, but become significant in the vicinity of Indian Wells Valley (Figure 2), where residuals tend to reflect both sedimentary fill under some stations and upper crustal velocity anomalies previously imaged by Walck and Clayton [1987] from local earthquake travel times and associated by them with magmatic systems in the area.

Arrivals from earthquakes to the west contain the largest anomaly in the region (Plates 1e and 1f), a group of residuals up to 1 s early relative to surrounding residuals. This anomaly, here termed the "Isabella anomaly" for the many observations made at station ISA, has long been recognized but has never been well located [Aki, 1982; Hadley and Kanamori, 1977; Humphreys *et al.*, 1984; Humphreys, 1985, 1987; Raikes, 1980]. The anomaly is ob-

Fig. 6. Station corrections employed in this study. Symbol sizes and fills are proportional to the magnitude of the correction. Circles denote stations with local delays, and squares are stations with advanced arrivals of teleseisms.

Plate 1. Teleseismic residuals for groups of earthquakes in different backazimuths; direction of travel of seismic waves is indicated by heavy arrows. Residuals from several events have been averaged together. To remove the source origin time, station Goldstone (GSC), southeast of our map area (Figure 1), was set to zero residual to permit comparison of these maps with *Raikes's* [1980] maps for all of southern California (but we do not use the crustal corrections that she used). For those events where Goldstone did not yield a clear arrival, we chose to make our residuals at stations ISA, CLC, and BMT (Figure 2) match those for similarly located events studied by *Raikes* [1978, 1980]. Stations FRI (on the northwest corner) and BNP (on the top edge) are off of the map (as indicated by the small arrows) but have been plotted at their proper latitude and longitude, respectively.

served farther south for earthquakes with more northerly backazimuths (Plates 1g and 1h); simply projecting the edges of the anomaly back toward the source gives an indication of the north-south extent and position of the causative velocity anomaly.

Arrivals from other azimuths do not present such a dramatic pattern of residuals, especially in the High Sierra. If the High Sierra is supported by either a crustal root or density variations in the crust, we would expect to see delayed arrivals at the High Sierra stations from most azimuths because low-density materials commonly have low velocities. If the High Sierra is supported by a low-density, low-velocity feature in the mantle, we would expect delayed arrivals to be observed at stations on the far side of the Sierra from the source. Examination of the arrivals does not consistently reveal either pattern; most noticeably, arrivals from earthquakes located due north or south (Plates 1d and 1h), which are least affected by the "Isabella anomaly," fail to indicate the clear presence of any low-velocity material under the Sierra. However, the pattern of residuals observed for earthquakes to the north is strikingly different from that for events to the north-northwest (Plate 1h vs. 1g). Stations in the Sierra in the northern part of our network have near zero residuals for events to the north-northwest, but these same stations have strong negative residuals (about 0.5 s earlier) for the events from the north; thus the early arrivals in the Sierra from the earthquakes to the north might indicate the presence of a very deep (>200 km) velocity anomaly to the north that might mask shallower, lower velocities. This difference between the north-northwest backazimuths and the northerly backazimuths suggests that two different velocity anomalies might be responsible for the notably early arrivals at ISA from these azimuths, instead of a single broad high-velocity anomaly under the Sierra as suggested from earlier studies [e.g., Raikes, 1980]. A similar explanation might account for the differences seen between two easterly backazimuths (Plates 1a and 1b).

The anomaly patterns can be presented in a second manner that both demonstrates how well these arrivals sample the Earth beneath the study area and provides a simple method of discerning the major velocity anomalies under the network. We determine the point at which each ray between every earthquake-receiver pair crosses a fixed depth; we then map these points, coloring each point to reflect the anomaly observed for that ray as cartooned in Figure 3. As for the residuals in Plate 1, we have fixed the residual at GSC to zero. Any anomalies near GSC will thus bias the residuals from arrivals in certain backazimuths and can thus create artifacts in this image; previous work using the entire southern California network indicates such effects are minimal [Humphreys *et al.*, 1984; Raikes, 1978, 1980]. These plots can be interpreted as maps of relative velocity perturbations if all observed residuals were caused by anomalies at a single depth. The difference in coherence of an anomaly at different depths provides some insight into the depth of the anomaly; in the absence of other anomalies, the depth at which an anomaly is most coherent is probably a good estimate of the depth of that anomaly (Figure 3).

Once again, the "Isabella anomaly" is dramatically illustrated by these "pauper's tomographic plots" (Plate 2) and cross sections through these plots (Figure 7). The early arrivals defining the anomaly appear to be best focused between depths of 110 and 190 km, where the scatter in residual among adjacent rays is minimized to about 0.2 s; at 70 km depth the edges of the anomaly are very incoherent with a mixture of residuals differing by as much as 0.8 s on the south edge. At depths of 190 km and greater the anomaly becomes less tightly focused but remains fairly coherent owing to the absence of large variation of the azimuths of rays at any given point. By analogy with Figure 3, it would seem that the

center of the Isabella anomaly is probably somewhere near 150 km depth.

Patterns under the Sierra Nevada in this simple backprojection reinforce the impression that a significant localized low-velocity body is absent beneath the Sierra, particularly at shallow depths (Figure 7e). Simple interpretation of Plate 2 and Figure 7 is hindered by the presence of the large high-velocity body so clearly evident to the west, but the locus of generally later arrivals along and east of the crest at 70 km depth is fairly incoherent (scatter among neighboring rays is about 0.6 s) and appears to terminate northward at Owens Lake. The plot of residual versus distance along the crest (Figure 7e) indicates that the dominant pattern is for more delayed arrivals to be observed to the south, which is opposite that expected from the increased elevation to the north. At greater depths the delayed arrival anomaly does become more coherent and shifts west under the range; at 150 km depth there are delayed arrivals clustered under the High Sierra and under the southernmost Sierra. At these and greater depths these clusters are formed largely of arrivals observed at stations in the Coso area from earthquakes to the northwest, west, and southwest.

Although we lack sufficient coverage east of the Coso region to make very definitive conclusions, an interesting pattern exists to the east of our network. The advanced arrivals observed in the Coso region from events to the east (Plate 1b and 1c) tend to focus at depths of 70 to 110 km beneath or a little west of the southern end of Panamint Valley near the Garlock fault (Plate 2). There is also a small collection of early arrivals clustering near the northern end of Panamint Valley. These clusters are rather puzzling if they do represent high-velocity material at approximately these locations because there seems to be no correlation with the surface geology. The generally early arrivals observed from events to the east might not reflect mantle with a higher velocity than that near GSC but could reflect generally thinner or higher velocity crust east of the Coso area.

The relatively simple presentation thus far strongly suggests that a large high-velocity body about 100 to 200 km beneath the foothills of the Sierra Nevada is responsible for the early arrivals observed in the southern Sierra from events located to the west. To isolate other anomalies from this large body we have conducted a series of inversions using the arrival time residuals presented above.

INVERSION OF TELESEISMIC ARRIVALS

Technique

We use the arrival times of P waves from teleseismic events recorded at the regional array to determine the three-dimensional seismic velocity structure beneath the array. The technique used, based on the method of Aki *et al.* [1977], is similar to that used by Shedlock and Roecker [1987] and Abers and Roecker [1991] with some additional modifications discussed below. The crust and upper mantle above a depth of 270 km is parameterized as a set of contiguous volumes of possibly irregular shape; each volume has a constant slowness. To solve for perturbations to the initial set of slownesses, we use the nonlinear method described by Tarantola and Vallette [1982] as summarized in the formula

$$\begin{aligned} \vec{p}_{k+1} = & \vec{p}_k + \left(\mathbf{G}_k^T \mathbf{C}_{dd}^{-1} \mathbf{G}_k + \mathbf{C}_{pp}^{-1} \right)^{-1} \cdot \\ & \left\{ \mathbf{G}_k^T \mathbf{C}_{dd}^{-1} [\vec{d} - g(\vec{p}_k)] - \mathbf{C}_{pp}^{-1} [\vec{p}_k - \vec{p}_0] \right\} \end{aligned} \quad (1)$$

where $\vec{p}_k$ is the vector of slowness at the k th iteration, $\vec{p}_0$ is the initial slowness model, $\mathbf{G}_k$ is the matrix of (demeaned) partial

Cross Sections of "Pauper's Tomography"

Fig. 7. North-south sections through the "pauper's tomography" of Plate 2. Vertical axis is travel time residual and horizontal axis is distance along section lines either (a-d) through the "Isabella anomaly" or (e-h) along the axis of the Sierra. As with Plate 2, distances are calculated from the positions of rays at depths of 70 km (Figures 7a and 7c), 110 km (Figures 7b and 7f), 150 km (Figures 7c and 7g), and 190 km (Figures 7d and 7h). Note that for the "Isabella anomaly," the scatter of residuals at a given horizontal distance tends to be minimized at depths of about 110-150 km. The section along the Sierra yields a pattern of decreasing delays to the north in the shallowest section (70 km), but the extreme scatter of residuals suggests that these arrivals are strongly affected by anomalies at other depths.

Plate 2. "Pauper's tomography" for the southern Sierra Nevada. Each point represents the intersection of a single raypath with a depth of (a) 70 km, (b) 110 km, (c) 150 km, and (d) 190 km beneath the southern Sierra Nevada. The color of each point reflects the travel time residual associated with that ray, as shown in the histogram at the bottom. Note that the zero incidence of the histogram is not at the histogram base (there are very few points with residuals > 0.7s). Residuals have been normalized so that station GSC is zero. Note the greater area mapped than in Figures 2 and 6 or Plate 1. In the northeast quadrant, the rightmost valley (between parallel faults) is Death Valley, the next valley to the west is Panamint Valley, and the Coso region is west-southwest of Panamint Valley (also see Figure 1). The north end of Panamint Valley lies at the north end of the fault defining its western side. Brackets indicate the position of the sections in Figure 7.

derivatives at the k th iteration, C_{dd} is the data covariance matrix, C_{pp} is the model covariance matrix, g is the relation used to calculate the travel time (the forward problem), and $\vec{d}$ is the vector of demeaned travel time residuals.

In applying this formula, we presume that a priori data uncertainties (as summarized in C_{dd}) are uncorrelated, but that a priori model uncertainties may be correlated. In other words, we presume that heterogeneities exist in the Earth that will cause arrival times to be systematically delayed or advanced at stations that are sufficiently close together. We model this correlation using the covariance function

$$C_{ppij} = \sigma^2 e^{-\left[\frac{d_{ij}^2}{2\Delta_h^2} + \frac{z_{ij}^2}{2\Delta_z^2} \right]} \quad (2)$$

where d_{ij} and z_{ij} are the horizontal and vertical distances between the centers of blocks i and j , respectively, Δ_h and Δ_z represent the characteristic horizontal and vertical dimensions of the heterogeneity, and σ is the a priori uncertainty in the strength (i.e., size of the slowness anomaly) of the heterogeneity. In practice, the effect of incorporating this term into the inversion is to impose a degree of smoothness on the solution, particularly where large contrasts are not strongly demanded by the data. When using this term, we normally choose Δ_h and Δ_z to be 40 km, which is based on the width of robust anomalies clearly visible in the raw residuals (e.g., Plate 2). The value of σ is more arbitrary as it represents an a priori estimate of the variance of the block velocities; we presume that a proper value would correspond to a velocity standard deviation of about $\pm 4\%$, which is approximately the scatter in the P_n velocities and upper mantle velocities inferred from body waves [Burdick and Helmberger, 1978; Iyer and Htchcock, 1989; Walck, 1984, 1985].

Our use of the smoothing effect of the C_{pp} matrix is limited by numerical instabilities probably introduced by the multiple inversions of C_{pp} in our current formulation of the inversion (1). These instabilities are expressed as divergence of the inversion and generally become more significant as the characteristic dimensions become larger, the a priori model uncertainty (σ) becomes larger, and as the number of parameters grows. Comparison of results with and without the off-diagonal terms of C_{pp} indicates that for our data and choice of parameters, the smoothing effect of these off-diagonal terms is pronounced in the crustal layer(s) (0–30 km) and decreases with depth to the point of not affecting the results below about 110 km.

Our inversion minimizes the variance

$$\sigma^2 = \frac{n_o}{(n_o - n_f)} \left(\frac{\sum_i w_i \text{res}_i^2}{\sum_i w_i} \right) \quad (3)$$

where n_o is the number of observations, n_f is the number of free parameters, res_i is the demeaned residual calculated from observation i , and w_i is the weight assigned observation i . Note that because the residuals have been demeaned, the mean residual term usually present in the calculation of the variance is zero. On occasion, we will mention the rms as a measure of the fit to the travel times without reference to the number of degrees of freedom in the structure. This rms is the square root of the variance in (3) with n_f set to zero.

We conducted several different inversions, varying the starting structure, parameterization (both block size and grid orientation), characteristic dimensions of the model covariance matrix, and a

priori estimates of the variance of model parameters. In general, we found that significant results were insensitive to the starting structure and only weakly affected by the characteristic dimensions of the C_{pp} matrix. Some significant variations were caused by the choice of parameterization, where “significant” means an anomaly present in one parameterization but absent in another, and where an anomaly is a group of three or more contiguous blocks differing from the one-dimensional (1-D) velocity structure by more than one standard deviation. Although such variations are associated with differences of several percent in the a posteriori data variance and thus might reflect a true improvement of one structure relative to another, it is difficult to evaluate the significance of these differences in the absence of an exhaustive investigation of different parameterizations. This is particularly true when elements of the parameterization include not only the position and orientation of block boundaries but also the choice of constant velocity blocks versus constant velocity gradients between nodes and the applicability of the simple scheme for tracing rays in the presence of large velocity contrasts.

One method of dealing with errors introduced by either the choice of parameterization or inaccuracies in tracing rays is to use an a priori theoretical variance to account for scatter in the observations produced by errors in the parameterization used [e.g., Roecker, 1993; Tarantola, 1987, p. 158]. This term is added to the a priori data variance (C_{dd} in (1)). Adding 0.1 s to the estimated a priori data variances does, in fact, significantly reduce the differences between structures with differing parameterizations, and experience indicates that the statistically significant anomalies remaining are relatively independent of the parameterization chosen. This improved agreement among different parameterizations does, of course, increase the overall variance somewhat for any given parameterization (e.g., 0.00980 to 0.0146 s² for the parameterization with the smallest variance (Plate 3; Table 3)), but this increase is small compared with the overall decrease in variance from that of the 1-D structure, 0.0570 s². Adding this term to compensate parameterization errors also increases the a posteriori standard errors from about 1% of the velocity to about 1.5% and reduces the magnitude of the diagonal elements of the resolution matrix by about 30%.

The structure presented in Plate 3 has the lowest variance of the inversions described above; however, many of its 589 free parameters are unnecessary to account for most of the observed residuals. To separate the important anomalies and present them without the clutter of the full inversion, we have constructed a much simpler parameterization using “metablocks” (39 degrees of freedom; variance = 0.0172 s²) that encapsulates features common to most of the inversions with several hundred degrees of freedom (Figure 8; Table 3). Although this parameterization is not unique, it has the property that removing any of the degrees of freedom remaining will increase the variance; thus all of the variations in this inversion most likely represent some real variation in the earth, though our choice of boundaries might not be entirely correct. In examining both structures, note that this inversion constrains only lateral variations in velocity referenced to some one-dimensional structure. Thus comparisons between layers should be made with care.

Results

The purpose of the inversions is to identify important velocity anomalies and constrain their location and intensity. We especially seek to locate velocity perturbations that might be associated with the isostatic compensation of the Sierra Nevada. Constraints are established by comparing the different inversions

Plate 3. Results of the inversion of teleseismic travel-time residuals for the best-fitting parameterization attempted for (a) velocity perturbations only and (b) velocity perturbations overlain by resolution (diagonals of the resolution matrix; black pattern). Except for grouping of certain blocks into "superblocks" near the edge of the inverted volume, each block's velocity could vary independently. Note that the blocks are 20 km on a side in the top layer and 27 km on a side in all other layers.

b

Plate 3. (continued)

TABLE 3. Principal Statistics of Several Structures and Inversions

Structure/Inversion	rms (s)	Percent Decrease From 1-D	Degrees of Freedom	Variance, s ²	Percent Decrease From 1-D
<i>A Priori Structures</i>					
1-D	0.2386	0	-	0.0570	0
A priori crustal root	0.2737	-14.7	-	0.0749	-31.4
A priori mantle root	0.2582	-8.2	-	0.0667	-17.0
<i>Inversions Starting From 1-D*</i>					
Completely independent blocks	0.0738	69.1	589	0.0098	82.8
Completely independent blocks, but added 0.1 s to C _{dd} (Plate 3)	0.0902	62.2	589	0.0146	74.3
Simplified parameterization (Figure 8)	0.1293	45.8	39	0.0172	69.8
<i>Inversions Testing a Priori Structures[†]</i>					
1-D start, highly damped $z < 50$ km	0.1088	54.4	494	0.0189	66.9
A priori crustal root, highly damped $z < 50$ km	0.1134	52.5	491	0.0204	64.2
A priori crustal root, highly damped $30 < z < 50$ km	0.1043	56.3	592	0.0197	65.5
A priori mantle root, highly damped $z < 70$ km	0.1133	52.3	446	0.0193	66.1
A priori mantle root, highly damped $30 < z < 70$ km	0.1024	57.0	530	0.0175	69.3

Statistics are based on 1325 arrivals and 54 earthquakes. 1-D is one-dimensional.

*Parameters for these inversions are 20 x 20 km blocks in layer 0-30 km depth and 27 x 27 km blocks in layers 30-70, 70-110, 110-150, 150-190, 190-230, and 230-270 (e.g., Plate 3). Horizontal and vertical characteristic dimensions are $\Delta_h = \Delta_v = 40$ km.

[†] Parameters for these inversions are 20 x 20 km blocks in layers 0-15, 15-30, and 30-50 and 27 x 27 km blocks in layers 50-70, 70-110, 110-150, 150-190, 190-230, and 230-270.

to identify anomalies common to all inversions with acceptably low variances, then considering the standard error and diagonal element of the resolution matrix of each anomaly. Generally, we prefer to interpret anomalies with resolution diagonals greater than 0.5 and with amplitudes exceeding one standard deviation; blocks with resolution diagonals > 0.5 are termed "better resolved" below. Resolutions greater than 0.5 indicate that the anomalous values are derived more from the arrival time observations than the a priori structure [Jackson and Matsu'ura, 1985], while anomalies with amplitudes less than one standard deviation are not considered to be statistically significant. In addition, the off-diagonal terms of the resolution and covariance matrices indicate how distinctly a given block is resolved from any other block. Throughout discussions of velocity structures derived from teleseismic arrival times we will refer to "low" and "high" velocity material; this refers to the velocity of the material relative to other material at the same depth. Low-velocity material beneath high-velocity material, in this context, need not mean that there is a velocity inversion; such an occurrence would depend on the increase of the velocity of the mean Earth structure with depth. For this reason, we quantify velocity perturbations as a percentage difference from the mean (or another body) at that depth range.

The better resolved region in the 0 to 30-km layer extends about 80 km north and 40 km east from Lake Isabella, with another equally resolved area just east of the Sierra in the Coso region (Plate 3); velocity patterns in these regions indicate that velocities consistently are lowest in the High Sierra and greatest in the area around Lake Isabella with intermediate velocities in the Coso area. Although this pattern within the Sierra suggests that we have imaged some part of the low-density body supporting the High Sierra, both the presence of lower velocities under Coso than under the topographically higher Isabella region and the absence of this feature from simpler structures (e.g., Figure 8) indicates that these variations might not be significant.

The anomalies inferred to lie in the uppermost mantle layer (30 to 70 km depth) differ sharply from those within the crustal layer. The better resolved region lies beneath and between the areas resolved at shallower depth and extends south to the Garlock fault. Within this better resolved region, the dominant contrast is

between high-velocity material roughly west of Lake Isabella and lower-velocity material to the east with the boundary running roughly north-northeast from or slightly east of Lake Isabella. This contrast reaches or exceeds a magnitude of about 5%, or about 0.40 km/s. This pattern places the higher velocities in the upper mantle beneath the lower velocities inferred in the crust.

The relationship between the layers above and below the Moho deserves special attention as the discussion of compensation of the Sierra focuses on these depths. Velocities in these two layers are coupled because of the steep incidence angles of the teleseismic rays at these depths; this results in poor resolution vertically but good resolution horizontally. Because blocks within the upper layer (0 to 30 km depth) are most strongly coupled to the blocks beneath them in the second layer (30 to 70 km), the patterns of anomalies in these layers might be combined with minimal effect on the variance. This was tested by running an inversion identical to that shown in Plate 3 except the velocities in the top layer (0-30 km) were forced to remain laterally constant. The variance of the resulting structure is identical to that for the structure that allows 3-D variations in the crust. Velocities in the upper mantle (30-70 km) appear to combine the velocity variations in the top two layers in Plate 3, with the net result more closely resembling the pattern in the 30 to 70-km layer in Plate 3. This can be seen to some degree in the simplified inversion in Figure 8. Velocities below 70 km were nearly identical to those seen in Plate 3.

The 110- to 150-km layer is the most symmetrically resolved: off diagonal terms of the resolution matrix are both small and of similar magnitude for blocks above or at the side of any given block. Most of the southern Sierra and Coso region is well resolved, though the resolution starts to decrease from this layer downward as the number of hits per block decreases. This depth range is the shallowest in which the source of the "Isabella anomaly" is clearly evident. The anomaly can be produced in several ways; the most parsimonious (shown in Plate 3 and Figure 8) is a single body, roughly 90 km in north-south extent, extending west from roughly the longitude of ISA for an unknown distance, and with a velocity about 5% higher than its surroundings. Material with a velocity slightly (2%) lower than average seems to

Fig. 8. Inversion results for a very simple parameterization constructed using the results of several different inversions and examination of the "pauper's tomography" from Plate 2.

lie under the eastern High Sierra at these depths. Most inversions tend to place high-velocity material beneath the intersection of the Garlock and Panamint Valley faults between 110 and 150 km (e.g., the isolated high-velocity block to the east in the 110-150 km layer in Figure 8), but that region is poorly resolved.

The Isabella anomaly appears to extend down to depths between 150 to 190 km, though its eastern edge appears to be 30 to 50 km west of ISA. This apparent west dip of the eastern edge of the anomaly is probably more an artifact of the raypaths used here (which, at these depths, are mostly coming from earthquakes to the west; Plate 2) than a real feature, but the position of the eastern edge of the anomaly should accurately reflect the easternmost possible extent of the anomaly at these depths. The amplitude of the anomaly would once again be about 4%.

The velocity structure below 190 km generally lacks consistent patterns from inversion to inversion and is poorly resolved; the velocity variations in these layers probably reflect a combination of real anomalies at these depths, anomalies outside our structure, and artifacts produced by the parameterization. Because azimuths of raypaths within any single block in these layers tend to be very limited (i.e., there are few crossing rays), depth resolution becomes very poor. There seems to be some pattern of anomalies similar to that of the Isabella anomaly above; this might either be smearing of the Isabella anomaly downward, or the Isabella anomaly could extend to these depths but be poorly resolved. As indicated by the simple parameterization of Figure 8, the "Isabella anomaly" can be explained largely by velocity contrasts above 190 km, so the velocity anomaly need not extend below 190 km.

Tests of a Priori Structures

Unconstrained inversions of the teleseismic data do not yield a simple image of an isolated low-velocity body under the High Sierra where a low-density mass must compensate the elevation of the range; to satisfy the teleseismic observations, any low-velocity body probably overlies somewhat higher velocities. The possible tradeoffs between layers, discussed above for the 0- to 30-km and 30- to 70-km layers, prevents us from resolving the low velocity in the crust from deeper high-velocity material as imaged by the unconstrained inversion (Plate 3). This suggests that introducing an a priori structure with low velocities under the Sierra might lead to a structure containing such low velocities but with a variance comparable to the best fit structure illustrated in Figure 10. We test two possible structures: a thick crustal root, using the contour map of *Mooney and Weaver* [1989], and a "mantle root," using a velocity of 7.65 km/s from 30 to 70 km depth under the range north from 30 km north of ISA and 7.9 km/s elsewhere.

As we would expect from the absence of a large cluster of delayed residuals under the High Sierra in the "pauper's tomography," the variance associated with these two structures before inversion is substantially greater than the variance associated with the initial 1-D structure of Table 2 (Table 3). This clearly indicates that these a priori structures are unacceptable with respect to the teleseismic arrival times. As noted earlier, this could be a bias produced by the high velocities under the Sierran foothills; thus we wish to invert for an optimal velocity structure that preserves either the crustal or mantle root. Because other experiments have shown that a final velocity structure is independent of the starting velocity structure for a given parameterization if all blocks are equally damped, we invert one test structure by holding the upper 50 km fixed for the crustal root and the other by holding the upper 70 km fixed for the mantle root. After these converge to a solution, we permit the upper 30 km to vary. As a control, we also inverted a starting 1-D structure holding the upper 50 km fixed. Our

parameterization is identical to that in Plate 3 except that we have subdivided the upper layers to form layers from 0-15 km, 15-30, 30-50, and 50-70 km. The vertical interfaces in the layers above 50 km are the same as those in the 0- to 30-km layer in Plate 3.

Although these inversions do reduce the variance of the travel time residuals by about 65% from the variance of the 1-D structure (Table 3), requiring the inclusion of either type of root produces a variance about 30% greater than that of our best fit structure of Plate 3. Indeed, both types of root fit the data no better than a 1-D structure in the upper 50 km. The substantially larger variance than that of the best fit structure (Plate 3) would lead us to reject the presence of either type of root, were there not other evidence of some such body, such as the P_n travel times and Bouguer gravity anomalies discussed below. If we assumed that one type of root or the other must exist, the results of this inversion would not permit us to choose one root over the other, though the mantle root does have a somewhat smaller variance.

The freely inverted parts of the velocity structures produced by the inversion of both crustal and mantle roots are fairly similar. Both produce a strong high-velocity body beneath the base of the root, between 50 and 110 km for the crustal root and 70 to 110 km for the mantle root. The velocities under the roots are generally 3-8% higher than the region southeast of a northeast trending line through ISA. Differences between the crustal and mantle roots are few outside of the 30- to 70-km depth range: The velocities beneath the mantle root are somewhat higher than beneath the crustal root (4.5-8% versus 3-6%), and the mantle root produces a larger low-velocity body under the High Sierra west of Owens Lake between depths of 110 and 150 km. Other than this intensification of the low velocity between 110 and 150 km depth, both structures are very similar to that in Plate 3 below 110 km.

Teleseismic observations are not compatible with the crustal and mantle roots previously proposed for the Sierra Nevada. Two hypotheses regarding compensation of the High Sierra are consistent with the teleseismic observations: either densities beneath the Sierra do not correlate with P wave velocity, or a low-velocity body near the base of the crust overlies a high-velocity body roughly coincident with the low-velocity body. To distinguish between these possibilities, we must turn to other observations.

P_n ARRIVALS AND THE MISSING ROOT

Because steep teleseismic rays can average structure vertically, analysis of teleseismic travel times cannot entirely rule out the possibility that a low-velocity body compensating the range somehow overlies higher velocity material, though it does rule out a low-velocity body lacking a companion high-velocity body. The analysis of regional upper mantle refractions (P_n arrivals) can provide good resolution between the upper mantle and the crust and can help separate low-velocity upper mantle from any high-velocity material that might underlie it.

We were fortunate in recording several earthquakes about 200-400 km distant from the center of our network. The three most interesting events lie north and south of the network: in the Garfield Hills (western Nevada), near Santa Monica, and southeast of San Clemente Island (Figure 1). These events provide clear P_n arrivals, thus constraining the Moho beneath the southern Sierra.

The Garfield Hills earthquake provides the most startling results, particularly in view of the earlier observation of *Pakiser and Brune* [1980] of an apparent P_n velocity of about 8.6 km/s in this same region between their two southernmost stations NEL and ISA from an event to the north-northwest near Truckee, California (Figures 1 and 2). We observe two very different apparent P_n ve-

Fig. 9. (a) Record section of P_n arrivals observed in the southern Sierra from an earthquake in the Garfield Hills (September 19, 1988, 02:56:31.74 UT, 38.459°N, 118.347°W, $z = 10.4$, $M_D = 5.0$, $M_b = 4.5$ [Lat, 1989]); reduction velocity of 7.8 km/s is used. Elevation corrections assuming an upper crustal velocity of 5.5 km/s and station corrections from Figure 6 have been applied. Seismograms from both the LBS stations and the permanent network have been filtered with a low-pass, three-pole Butterworth filter with a 10-Hz corner frequency. Heavy traces are for those stations within the backazimuth ranges used in Figure 13. (b) P_n arrivals observed in the Coso region. (Station JAW, picked in Figure 10, not shown for clarity).

Fig. 10. (a) Reduced travel time plot of arrivals from the earthquake in the Garfield Hills ($M = 5.0$) observed in the southern Sierra (backazimuths from 359° to 5°) and Coso/Indian Wells Valley (backazimuths from 349.8° to 355.7°). Error bars denote station corrections (see Figure 6). "Good" picks should be accurate to ± 0.1 s; "poor" picks are usually accurate within 0.4 s. All picks were made on unfiltered seismograms. Solid and dashed lines are linear regressions through the Sierra and Coso picks, respectively. (b) Same as Figure 10a for an earthquake near Santa Monica (September 12, 1988 13:24:34.18 UT, 33.858° N, 118.441° W, $z = 9.4$, $M_L = 3.9$). Short dashed line is an alternative fit suggesting that a small shadow zone exists around Lake Isabella with low P_n (7.63 km/s) velocities farther north. (Sierran stations have backazimuths from 176.4° to 184° ; Coso stations' backazimuths from 195° to 205°). Gray lines indicate arrival times predicted for Moho depths and local P_n velocities labeled at right. (c) Same as Figure 10b for an earthquake southeast of San Clemente Island (August 20, 1988 18:15:27.89 UT, 32.503° N, 117.907° W, $z = 6.0$, $M_L = 3.8$). Short dashed line indicates fit through Coso stations north of the Garlock fault (Sierran stations have backazimuths from 167° to 176° ; Coso stations' backazimuths from 179.6° to 185°).

localities: 7.48 ± 0.1 km/s within the Sierra Nevada, and 8.13 ± 0.1 km/s to the east, in the Coso/Indian Wells Valley area (Figures 9 and 10). We consider these arrivals to be P_n because of the magnitude of the apparent velocity, the great distance from the source, and the linearity of the arrival times over a considerable distance, although we will discuss the possibility that these arrivals are re-

fractions from an interface above the Moho. We here make the simplest interpretation of these data by assuming that the arrivals are normal refractions and not diffractions or "tunneling waves," so called because seismic energy traveling along an interface (such as a P_n phase) enters the medium above the interface at a sharp corner and then tunnels through that lower-velocity

medium. Our assumption can be judged by the overall consistency of our final interpretation. Because of the linearity of the arrival times and because of the 2-D control provided by the network geometry, we do not think these estimates are contaminated by energy traveling out of the plane between source and receiver for the arrivals shown in Figures 9 and 10. Other stations in other azimuth ranges do show erratic arrival times suggesting that they are contaminated by complex raypaths, particularly near the eastern edge of the Sierra Nevada.

Unlike the earlier study of *Pakiser and Brune* [1980], we can obtain an estimate of the true P_n velocity by using the two events to the south (Figure 10). The earthquake near Santa Monica was the better recorded in the Sierra; the rays that sample the same mantle as those from the Garfield Hills event arrive at stations north of ISA. Two interpretations are possible: first, the mean apparent P_n velocity is 7.80 km/s (solid line, Figure 10b) and the early arrivals about 10–20 km north of ISA reflect locally thin or fast crust somewhat south of ISA; second, the apparent P_n velocity varies, with a clear value of 7.63 beyond 20 km north of ISA and more uncertain values to the south, ranging from 7.8 to 8.3 km/s. The arrivals near ISA are within about 40–50 km of the crossover from P_g (direct crustal phase) first arrivals to P_n arrivals; thus these arrivals, which have greater reduced travel times than arrivals to the north or south (Figure 10b), might lie in a shadow zone produced by complex structure in the lower crust or upper mantle near the Garlock fault, a suggestion supported by the absence of similarly late arrivals at the same stations from the San Clemente earthquake (Figure 10c). The San Clemente earthquake arrivals have a mean apparent P_n velocity of 7.70 km/s.

The San Clemente earthquake was better recorded by the permanent network than the Santa Monica event and so provides a better collection of arrivals in the Coso region (Figure 10c). The overall mean apparent P_n in the Coso region was observed to be 7.81 km/s, but this includes arrivals south of the Garlock fault. Considering only the arrivals north of the vicinity of the Garlock, the best fit apparent P_n velocity is 7.97 km/s. Only the rays reaching the three or four northernmost Coso stations sample the same mantle path as the rays reaching the three or four southernmost stations picked from the Garfield Hills earthquake (Figure 10a), so interpretations of the apparent velocities in the Coso region in terms of a true upper mantle velocity are poorly constrained.

Using these velocities and fairly conservative uncertainties

(0.1 km/s if no range is stated above), we estimate the Sierran P_n velocity to be between 7.5 and 7.7 km/s and the P_n velocity beneath Coso to be between about 7.8 and 8.1 km/s. Although our inferred P_n velocity under the Sierra is unusually low, it seems too high for crustal materials, and the absence of first-arriving phases with greater apparent velocities to distances of about 400 km indicates that these velocities do not reflect a pillow of high-velocity material at the base of the crust (though such a possibility is considered below). Thus we will term material which has velocities in excess of 7.5 km/s as mantle through the remainder of this paper. If the crust is laterally homogeneous, then the Moho would dip 0° to about 2° south under the Sierra. Because of both the limited reversal of mantle paths in the Coso area and the known presence of velocity perturbations in this area [Walck and Clayton, 1987], it is difficult to infer the N-S component of Moho dip. If the Moho dips away from Coso both to the north and south, true P_n velocities might be as low as 7.8 km/s, but if it is nearly flat, the true P_n would be greater than 8.0 km/s.

We can formalize this estimate and use the total travel time from earthquake to receiver to estimate the crustal thickness near the receiver; this is more difficult than inferring P_n velocities or Moho dip because we must know the location, origin time, source structure, and mean mantle velocity along the raypath to estimate the crustal thickness. The travel time t is the sum of the time spent in the crust at the source (t_s), the time in the mantle (t_m), and the time in the crust at the receiver (t_r):

$$t = t_s + \int_{L_s}^{L_e} \frac{dx}{v(x)} + t_r \quad (4)$$

where $v(x)$ is the P wave velocity at the top of the mantle at a distance x from the earthquake and L_s and L_e are the positions where the ray enters and exits the mantle. If we separate the mantle path into discrete segments with a constant gradient in v and d (the depth of the Moho) between nodes (Figure 11), we can rewrite (4) as

$$t = t'_s + s_1(l_0 + \frac{1}{2}l_1) + \sum_{j=2}^{n-1} \frac{1}{2}s_j(l_j + l_{j-1}) + s_n(\frac{1}{2}l_{n-1} + (1 - \frac{f}{2})l_n) + s_{n+1}\frac{f}{2}l_n f + \frac{(1-f)d_n + fd_{n+1}}{v_{cr} \cos \alpha_{cr}} + \frac{h}{v_e \cos \alpha_e} \quad (5)$$

Fig. 11. Illustration of the parameterization used in the simple inversion to fit arrival times of seismic waves inferred to have refracted along the Moho. Seismic energy travels down from the earthquake at left to the mantle, where it is assumed to travel horizontally to an exit point from the mantle, where it ascends to the station h km above sea level. Moho depth and upper mantle P wave velocity are considered to vary linearly between values at nodes (large dots).

where $l_j (= x_j - x_{j-1})$ is the distance between nodes j and $j+1$, $s_j = 1/v(x_j)$, f is the fraction of the distance from node n to node $n+1$ where the ray exits the mantle ($f = (L_e - x_n)/(x_{n+1} - x_n)$), h is the station elevation, v_{cr} and v_e are the mean velocities of the crust below and above sea level, respectively, and α_{cr} and α_e are the incidence angles within the crust above and below sea level, respectively, and

$$t'_s = t_s - \int_0^{L_s} \frac{dx}{v(x)} = \frac{(d_0 - z) \cos \alpha_{cr}}{v_{cr}} \quad (6)$$

Note that the expression of total travel time in (5) is a linear function of t'_s , s_j , d_j , and $1/v_e$. Thus we can construct a series of linear equations relating observed travel times to model parameters t'_s , s_j , d_j , and $1/v_e$ and invert this relation using (1). As for the teleseismic inversion, we can recover resolution and a posteriori model variances. We can combine travel times from several earthquakes provided they lie along a single profile and the arrivals all are refractions from the same layer.

We invert the arrival times from the Santa Monica, San Clemente, and Garfield Hills earthquakes for two different assumptions: (1) that all the arrivals from these three earthquakes are refractions from a common interface and (2) that the arrivals from the Santa Monica and San Clemente earthquakes are from a deeper interface with a higher velocity than that for the arrivals from the Garfield Hills earthquake. In the first case, we can solve both for mantle velocities and Moho depths, but in the second case we must fix either the geometry of the interface or the velocities.

Assuming a common refractor from both north and south, we find that P_n velocities are best resolved between about $x = 200$ km and $x = 275$ km (roughly from station ISA to Mount Whitney); this is easily understood from the ray geometries shown in the top of Figure 12. The Moho dips slightly south at $\sim 1.5^\circ$, the crust is about 33 ± 5 km thick (incorporating uncertainties from the earthquakes' origin times and depths), and the mean P_n velocity is about 7.6 ± 0.1 km/s. The depths of the Moho at individual nodes are quite well resolved (resolution diagonals > 0.95), but the P_n resolutions do not exceed 0.6 for the parameterization shown in Figure 12. This is because P_n values of adjacent nodes are closely linked because of the geometry of rays; the mean P velocity of 7.6 from $x = 200$ to $x = 275$ km is much better resolved than is apparent from individual nodes; a different parameterization yields an estimate of 7.62 ± 0.08 km/s from $x = 200$ to $x = 275$ km.

For the case where arrivals from earthquakes to the north and south do not share a common refractor, we assume that the refractor for earthquakes to the south has a velocity of 8.0 km/s and that the refractor for the earthquake to the north is at a depth of 31.4 km and has a velocity of 7.5 km/s; the crust above 31.4 km has an average P wave velocity of 6.1 km/s. From these assumptions, we find that the 8.0 km/s refractor must dip north $\sim 8^\circ$, deepening from 41 to 47 km going from station ISA ($x = 200$ km) north to $x = 240$ km; there is no control farther north (Figure 13). Changing the velocity of this refractor has only a minor change on its depth at ISA, but a change of ± 0.1 km/s changes the depth by ± 2.7 km at $x = 240$ km. The nominal uncertainty on these depths is ~ 0.5 km, which reflects their relative accuracy, but the absolute depth might vary by about 5 km owing to uncertainties in the earthquakes' origin times and depths. Decreasing the velocity of the upper refractor from 7.5 to 7.2 km/s reduces the thickness of that 7.x km/s layer by 25–30% and requires the upper refractor to dip to the north.

We cannot conclusively reject either of the two scenarios discussed above. At present, the assumption of a common refractor

for arrivals from earthquakes to the north and south produces the simplest explanation of the data. If there are separate refractors from north and south we must explain our failure to observe any arrivals from the Garfield Hills earthquakes that are consistent with the presence of a layer with a velocity between 7.9–8.1 km/s. Perhaps there is some shadowing of seismic energy that prevents significant energy from entering this higher-velocity layer, or perhaps the layer is laterally discontinuous and so fails to transmit seismic energy. These explanations seem increasingly less likely as the velocity of the shallower refractor is reduced from ~ 7.5 km/s, because the lower the velocity of this layer, the thinner it becomes and the more steeply it dips to the north. This combination greatly increases the chances of the deeper refractor producing first arrivals for the Garfield Hills earthquake. Because we cannot easily bound the mechanism that shadows the deeper refractor, we cannot fully bound the range of velocities of the upper refractor, but this argument suggests that velocities below 7.2 km/s are quite unlikely for this "wedge layer" between the upper and lower refractors. Examination of earthquake arrival times throughout the Sierra (M. K. Savage et al., Earthquake refraction profiles of the root of the Sierra Nevada, submitted to *Tectonics*, 1993, hereafter M. K. Savage et al., submitted manuscript, 1993) indicates that material with velocities between about 7.2 and 7.65 km/s overlies a deeper 8.0 refractor in much of the range. Depending on the exact velocity of this material under the southern Sierra, this double refractor structure might be preferred despite its greater complexity.

The first hypothesis of a single refractor with lateral gradients in the P wave velocity beneath the refractor fits the observations quite well; the weighted rms travel time residual for the structure in Figure 12 is 0.07 s. In contrast, if we assume the structure of a crustal root suggested by *Pakiser and Brune* [1980], the weighted rms travel time residual is 0.65 s (0.54 s if the mean is removed from the residuals for each earthquake). Even after allowing the inversion to adjust the P_n velocities, the rms residual is 0.19 s, and the P_n velocities would be below 7.3 km/s under much of the Sierra. If the arrival times are from a common refractor, we can confidently rule out the possibility that there is a significant crustal root under the High Sierra. This assumption produces Moho depths comparable to those inferred by *Prodehl* [1979] and an estimated P_n velocity very nearly that inferred by *Carder* [1973].

Although the double refractor interpretation does differ substantially from the single-refractor model, both share one central feature. Both models require velocities in excess of about 7.2 km/s at depths of only about 32 km. In either model this "7.x" layer must be at least about 20 km thick under the High Sierra, a substantial departure from older interpretations. This "7.x" layer could be crust or mantle, but it seems likely that it is supports, in part, the elevation of the Sierra Nevada.

INTERPRETATION AND DISCUSSION

Sierra Nevada

Analysis of the teleseismic and regional P_n arrivals produces a picture of the Sierra Nevada at odds with the traditional view dating back to *Lawson* [1936] of the mountains as a manifestation of a thick crust. The principal observations that constrained the inference of a thick crustal root were seismic refraction profiles. The large Bouguer gravity anomaly, while consistent with a thick root farther to the north [*Oliver*, 1977; *Oliver and Robbins*, 1982], does not by itself require a thick crust and will be discussed later.

Two profiles paralleling the axis of the southern Sierra have

Fig. 12. Structure derived assuming that arrivals from regional earthquakes to the north and south are from a common refractor. Top panel shows the raypaths used and the depth of the refractor (interpreted as the Moho) with uncertainties drawn at the nodal points of the inversion. Gray line at the top is the mean elevation along the profile from *Diment and Urban* [1981]. Second panel shows the raypaths without vertical exaggeration. Third panel shows the inferred P_n velocities with uncertainties (leftmost velocity was highly damped). Bottom panel shows the diagonal of the resolution matrix from the inversion for the P_n velocity. Strong correlations between the P_n velocities of adjacent nodes suggests that the mean of the P_n velocity over a larger distance is the most robust parameter; the horizontal bar in the third panel illustrates the mean over the portion of the Moho with the best resolution.

Fig. 13. Two-refractor interpretation of arrival times observed from regional earthquakes. Arrivals from the Garfield Hills earthquake to the north (not shown) would be from the top of the 7.5 km/s wedge, while arrivals from the two earthquakes to the south would be from the refractor at the base of the 7.5 km/s wedge. Rays shown are for a structure with a lower refractor velocity of 8.0 km/s. The lower refractor's velocity is 7.8 km/s for $x < 115$ km, and rises linearly to the value labeled at right between 115 and 150 km.

produced the only direct measurements of a thick crust (Figure 1): an active, reversed profile between China Lake and Mono Lake [Eaton, 1966; Prodehl, 1979], and an unreversed profile extending south from Truckee to station ISA [Pakiser and Brune, 1980]. While interpretations of these data differ somewhat, estimates of crustal thickness have ranged from 42 to more than 50 km. Interpretations of both profiles assume that the structure is laterally continuous and that no energy travels out of the vertical plane between source and receiver.

The sensors used in Eaton's [1966] profile were deployed along the east flank of the Sierra in Owens Valley south into the China Lake area (Figure 1). The P_n arrivals from the Mono Lake blast were recorded south of Owens Lake into Indian Wells Valley. The arrivals we observed from the $M = 5.0$ event in the Garfield Hills clearly indicate that the east flank of the Sierra separates high (8.13 km/s) apparent P_n velocities to the east from low (7.5 km/s) apparent P_n velocities to the west (Figures 9 and 10). Eaton's profile trends across this boundary at an oblique angle, suggesting the assumption used in interpreting this profile, that is, that the seismic velocities in the upper mantle are laterally uniform, is probably erroneous. The presence of profound lateral velocity contrasts might also explain the different phase picks made by Eaton [1966] and Prodehl [1979]. An alternative interpretation of this profile might be that the 7.7 km/s apparent P_n velocity near Mono Lake and the 8.1 km/s apparent P_n velocity near China Lake are accurate estimates of the average P velocity under each region; this variation would largely explain the differences in the intercepts of lines fitting the P_n arrival times between the two

endpoints of the profile, thus removing most of the justification for an extremely thick crust (>50 km) in the southern Sierra.

Using travel times from aftershocks of an earthquake near Truckee, California to stations within the Sierra (Figure 1), Pakiser and Brune [1980] observed a mean apparent P_n velocity of 7.7 km/s over a distance 60 to 260 km north of ISA, but they used the 8.6 km/s apparent P_n velocity observed between NEL (Figure 2) and ISA to argue that the mean P_n velocity is really 7.9 to 8.0 km/s. Our stations cover the same 60 km interval north of ISA, but we observed an apparent velocity of 7.5 km/s from our Garfield Hills event. The difference in backazimuth to the two sources (near 0° for the Garfield Hills event and about 330° for the Truckee earthquake) cannot account for the difference in apparent P_n velocity if the 8.6 km/s observation truly reflects a dipping Moho. At least some of the arrivals from the two events must have traveled along a different refractor; an explanation consistent with each of the two refraction structures presented above is possible. If the arrivals from the Garfield Hills earthquake traveled along the Moho, then arrivals from the Truckee aftershocks at one or more stations must have traveled in higher-velocity material. Numerous workers have documented high upper mantle velocities (>8.0 km/s) under the Great Valley [e.g., Holbrook and Mooney, 1987; Oppenheimer and Eaton, 1984]; this led Jones [1987] to suggest that the ISA observation from the Truckee event is an out-of-plane refraction that traveled under the Great Valley. Alternatively, the arrival at ISA traveled through a deeper, high-velocity body of uncertain geometry that was undetected by the arrivals from the closer Garfield Hills earthquake. Recent exami-

nation of earthquake arrival times for events and stations in the entire Sierra Nevada suggest that this latter possibility could be the case (M. K. Savage et al., submitted manuscript, 1993), or that most of the arrivals studied by Pakiser and Brune travelled under some thickness of material with velocities between about 7.2 and 7.65 km/s.

If there is a 7.x km/s "wedge" layer under the Sierra, then the arrivals from the Garfield Hills earthquake might be from a shallower layer than arrivals from the Truckee aftershocks. The original interpretation of the Truckee aftershocks included identification of arrivals with an apparent velocity of 7.2 km/s [Brune and Archambeau, 1967]. Possibly these arrivals came from the same "wedge" layer as the Garfield Hills arrivals. An ongoing investigation into observations of arrival times of earthquakes within the Sierra Nevada indicates that material with velocities from 7.2 to 7.65 km/s exists under much of the Sierra (M. K. Savage et al., submitted manuscript, 1993).

Pakiser and Brune [1980] also noted that ISA had a later arrival than CLC to the east or WDY to the west and thus inferred a greater crustal thickness under ISA than to the west or east. The earlier arrival at CLC mirrors our observations and can be explained by the higher P_n velocity east of the Sierra; the earlier arrival at WDY to the west probably has a similar explanation. Thus the earlier observations interpreted as products of a thick root under the Sierra are consistent with the thinner crust and lower P_n velocities documented here.

The remaining refraction profiles are transverse across the southern Sierra and led Carder [1973] and Carder et al. [1970] to propose a thin crust under the Sierra underlain by a low-velocity upper mantle. The southernmost of their profiles lies just north of our array; Carder's [1973] original interpretation and the later variation used by Bolt and Gutdeutsch [1982] are in excellent agreement with the crustal thicknesses and upper mantle velocities that we infer in our "single refractor" structure. To date, there is no investigation of the possibility that a 7.x km/s layer is compatible with arrivals on these cross profiles.

Thus all the seismological evidence from regional earthquake travel times and refraction profiles can be explained by a near normal crust about 33 km thick under the Sierra underlain by rocks with P wave velocities between about 7.2 and 7.7 km/s to depths of at least 50 km. Our arrivals are incompatible with the structures proposed by Pakiser and Brune [1980] and Eaton [1966] for the sub-Sierran structure. In common with those older models, though, our inferred velocities are less than velocities at the same depths to the east and west of the Sierra. Our "single-refractor" structure fits all the observations available to date with a minimal complexity. This interpretation requires rocks with P wave velocities of 7.6 ± 0.1 km/s at about 33 km depth; this velocity is probably best considered a very low upper mantle velocity, perhaps representing very warm mantle containing some melt. Note that this interpretation does not rule out the presence of layers with velocities near 7.2 km/s at the base of the crust. An alternative interpretation (the "two-refractor" structure) permits the presence of rocks with velocities of ~7.2-7.5 km/s at about the same depth, and these rocks would then overlie a Moho(?) at about 50 km depth with underlying mantle having a near normal velocity about 7.9-8.1 km/s. In addition to being more complex than the single-refractor model, this structure does not explain our failure to observe arrivals from the deeper refractor from the Garfield Hills earthquake. Because of the additional modifications to this structure that are required to prevent observation of arrivals from this structure's ~50-km-deep-Moho at our Sierran stations, we prefer the single-refractor structure.

"Isabella Anomaly"

Unlike the High Sierra, the western foothills lack spectacular relief but do lie over a large velocity anomaly in the mantle. The high-velocity body that we have dubbed the Isabella anomaly lies slightly west of our network; for this reason we cannot constrain the western edge of the anomaly and cannot put tight bounds on the top and bottom of the anomaly. The pattern of travel time anomalies does indicate a definite southern and northern edge to the Isabella anomaly (at least at its eastern end; Plate 2 and Figure 7); this pattern is evident in the shape of the velocity anomaly produced by the inversion (Plate 3 and Figure 8). The anomaly appears to extend from about 110 km to 200 km depth, though the anomaly could extend to greater depth farther to the west. The overall pattern evident through several different inversions is that the anomaly's amplitude is greatest from about 150 to 190 km depth and is somewhat less above that. The anomaly does not extend above about 100 km depth unless low-velocity material is forced to exist near the Moho under the Sierra. As discussed above, forcing a low-velocity root to exist causes the inversion to bring high-velocity material to the base of the root, though over a broader area than that defined below 100 km.

The limited range of this high-velocity body was inferred by Aki [1982] from his compilation of earlier tomographic studies over the whole state of California and reinforces our inference that the anomaly is local to the southern part of the San Joaquin Basin. More recent inversions of teleseismic observations at permanent stations in California have tended to confirm the limited north-south extent of the anomaly below 100 km depth, though these studies have poor resolution of the eastern side of the anomaly owing to the distribution of stations [Benz and Zandt, 1993; Biasi and Humphreys, 1992]. Unlike the anomaly under the Transverse Ranges [Hadley and Kanamori, 1977; Humphreys et al., 1984; Humphreys and Clayton, 1990; Raikes, 1980], this anomaly is not easily connected with any topographic feature or geologic event. The great north-south extent of the San Joaquin Valley, the Sierra Nevada, and the Coast Ranges all argue against an association that was suggested by Humphreys [1987] between this limited anomaly and any of these features. The anomaly apparently has an east-west dimension similar to the east-west extent of the south dipping thrusts at the south end of the San Joaquin Valley [Biasi and Humphreys, 1992], but these thrusts lie about 100 km to the south and root away from the anomaly [e.g., Namson and Davis, 1988; Goodman et al., 1989]. The only geological feature that could possibly be spatially associated with the Isabella anomaly is the Bakersfield arch, which overlies the northern edge of the Isabella anomaly. The arch separates the 12 km deep Maricopa subbasin [Goodman et al., 1989] from the main San Joaquin Valley and its 6-km-thick fill, and it contains the largest outcroppings of Tertiary rocks in the eastern San Joaquin Valley.

Two clues suggest one possible interpretation of the Isabella anomaly. First, the velocity contrast between the anomaly and the surrounding mantle appears to decrease from about 190 km up to 100 km depth; the anomaly might not exist at shallower depth. This indicates that the anomaly might represent cooler, denser material from the upper mantle that is now descending from near the base of the crust to depths of near 200 km. Second, we observed a low-velocity body under the Tehachapi Mountains that is probably in the upper 100 km of the mantle. We suggest that the Tehachapi low represents low-velocity material replacing the high-velocity material now descending to form the Isabella anomaly to the north.

We can further speculate that the process creating the Isabella anomaly started in the Miocene when the upper crust in the south-

ern San Joaquin Valley and the Tehachapi Mountains was extended. During this time the sediments exposed near the edge of the modern basin were deposited at great depths (perhaps 2000 m below sea level; [Olson *et al.*, 1986]). After considerable vertical motions in the Miocene, the basin rebounded significantly [Goodman, 1989; Olson *et al.*, 1986], perhaps caused by the transport of dense upper mantle from this region down and to the north. The significantly greater exposure of Tertiary rocks on the east side of the San Joaquin Valley south of and including the Bakersfield arch might reflect some continuing regional uplift of this region accompanying the thinning of high-velocity upper mantle below. The lower velocities in the upper mantle beneath the Tehachapi Mountains (Plate 3 and Figure 8, 30–110 km depth) might also reflect removal of older mantle lithosphere and subsequent replacement with warmer mantle. This system of mantle downwelling and upwelling resembles that proposed by Humphreys *et al.* [1984] and Humphreys and Hager [1990] for the Transverse Ranges and Salton Sea to the south, but this system is smaller and has not produced crustal features comparable to the Salton Sea or Transverse Ranges. A somewhat similar interpretation has been advanced by Zandt and Carrigan [1993], who prefer to derive the material within the Isabella anomaly from the north.

The only noteworthy geophysical phenomenon at the surface that seems spatially associated with the Isabella anomaly is a paucity of seismicity. This absence of seismicity, combined with the coincident transition from extensional stress to the east to compressional stress to the west, led Mount [1989] to infer a point of isotropic stress in the vicinity of the Isabella anomaly. It is beyond the scope of this work to explore mechanisms that might link a mantle velocity anomaly with such major transitions in the stress field in the crust, but we do note that Sonder [1990] has shown that a high-density body in the mantle can perturb the regional stress field in the crust. Thus it is possible that this anomaly does produce some effect on the crust, and should such a link be established it might be possible to constrain other information about the Isabella anomaly from this effect.

The Isabella anomaly apparently represents some dynamic of the upper mantle not simply reconciled with the tectonics at the surface. Unlike the anomaly under the Transverse Ranges, this anomaly is not easily explained as the consequence of thickening of the lithosphere itself induced by the kinematics of the plate margin. The presence of this anomaly has not led to the construction of mountains or the foundering of basins. This is somewhat surprising because the Isabella anomaly produces travel time residuals equal to or greater than those associated with the Transverse Ranges anomaly [e.g., Humphreys and Clayton, 1990]. If the Isabella anomaly does represent some large downward flux of cold upper mantle, then large motions of mantle in or below the lithosphere need not produce sizable tectonic features at the surface.

Isostasy and Gravity

The large Bouguer gravity low associated with the relief of the Sierra Nevada (Plate 4) has generally been attributed to an Airy-type crustal root beneath the range; disagreements have been limited to the depth and size of the root and whether the range is in isostatic equilibrium [e.g., Oliver, 1977; Saltus and Lachenbruch, 1991] or overcompensated [e.g., Kennelly and Chase, 1989]. Although interpretations of gravity are notoriously nonunique, gravity gradients provide minimum estimates of the depth of density anomalies and the gravity anomalies themselves are the only direct observation of the density contrasts at depth responsible for support of topography. Because we have inferred that no sub-

stantial (>5 km thick) Airy root lies under the High Sierra, we here reexamine the gravity data to learn what constraints it places on the compensation of the Sierra. For completeness, we consider three simple end-member models of compensation that could both support the range and produce the observed gravity anomaly (Figure 14): a classic Airy root, lateral contrasts within the crust ("crustal Pratt root"), and lateral contrasts within the mantle ("mantle Pratt root"). By comparing the gravity anomalies calculated from these idealizations with observed anomalies and by using constraints from measured densities and surface geology, we show that a near-equal combination of the crustal "Pratt" mechanism with either of the other two can support the Sierra and satisfy the gravity measurements.

East-west profiles across the Sierra have often been used to demonstrate the presence of an Airy root beneath the range [e.g., Oliver, 1977; Kennelly and Chase, 1989]. This is curious because gradients on the west side of the Sierra are so steep that they require substantial density contrasts in the upper crust [Kennelly and Chase, 1989; Oliver and Robbins, 1982]; thus even overcompensation at the base of the crust will not fit the observed anomaly without upper crustal anomalies. Our east-west profile across the Sierra (Figure 15) strongly indicates that lateral density contrasts in the crust (i.e., the Pratt crustal root) fits the observed Bouguer anomaly much better than either an Airy root or a Pratt mantle root. These latter models produce the residual high in the foothills and low in the High Sierra depicted on isostatic residual maps [e.g., Jachens and Griscom, 1985]; these residual anomalies have in turn led to the suggestion that the Sierra is overcompensated [Kennelly and Chase, 1989].

The necessity of shallow density variations in the Sierran foothills has long been noted, but most workers tend to use the thinnest bodies possible so that most of the gravity low along the Sierra might be explained by a large crustal root [e.g., Kennelly and Chase, 1989]. The choice of such thin density variations is largely arbitrary and seems contrary to geologic and geophysical inference. The density variations measured at the surface mirror the topography to a large degree [Oliver *et al.*, 1987; Saltus and Lachenbruch, 1991] and are comparable to the 0.225 Mg m^{-3} variation calculated for our Pratt root (Figure 15). These density variations mimic the structure of the Sierran crust, which is dominated by the juxtaposition of silicic plutonic rocks along the Sierran crest with mafic to ultramafic rocks exposed in the western foothills [Saleeby, 1981]. Because of the large-scale vertical transport of material in the Mesozoic magmatic arc [Saleeby, 1990], it seems reasonable to assume that the gross variations in density observed at the surface extend down many kilometers into the crust. Gravity gradients associated with individual plutons in the upper crust indicate that these plutons extend down to about 10 to 13 km depth [Oliver, 1977; Oliver *et al.*, 1987, 1993]. A more diffuse density contrast below 10 km between the San Joaquin crust and the Sierran crust could exist, for the deep levels of the Sierran batholith exposed in the southernmost part of the range remain fairly silicic to paleodepths approaching 30 km, and densities rarely exceed 2.8 Mg m^{-3} even in these deep levels of exposure [Ross, 1989; Saleeby, 1990]. The bulk of geologic evidence along an east-west profile across the Sierra thus supports the possibility that much, or even all, of the gravity low of the Sierra comes from lateral density contrasts in the crust; the gravity data require that some of that low must be in the upper crust.

Like the east-west profile, a north-south profile of the Bouguer gravity anomaly along the Sierran axis is best matched by a "crustal Pratt" compensation (Figure 16). However, the north-south gravity gradient is gentle enough that the causative density

Plate 4. Gravity map with profile locations (heavy, outlined, straight white lines). Faults (solid lines) and 304 m (1000 feet) contour (white line) are shown for reference.

contrasts could be placed at or below the Moho. This latter possibility is reinforced by the absence of significant differences in the bedrock along the profile. Unlike the east-west profile, no dramatic change in the surface geology accompanies the ~120 mGal variation in the Bouguer anomaly south from the High Sierra (Figure 16). The paleodepths of exposed rocks increase only slightly, if at all, from the High Sierra to the Lake Isabella region [Ague and Brimhall, 1988], suggesting that the gravity anomaly does not originate in a shallow body increasingly denuded toward the south. Thus any of the three mechanisms could be operating along this profile; other information must be used to constrain the depth of compensation on this profile.

Constraints from the seismic observations. It is along this north-south profile that the seismic experiment presented here provides the most information about the compensation of the Sierra Nevada. As all the stations in the Sierra north of 35°15' are located on crystalline rock, station corrections are small and artifacts produced by sedimentary basins and other strong, upper

crustal velocity contrasts are absent. Inversion of the teleseismic residuals produced generally lower seismic velocities in the crust toward the northern end of the profile (Plate 3). Although this pattern reflects the variation we expect for compensation of the Sierra, it is not significant, for as we noted earlier, an inversion where the 0- to 30-km layer is forced to remain uniform results in a structure fitting the arrival times as well as the best fit structure of Plate 3.

Lacking good control from the teleseismic inversion, we must use constraints from the P_n observations. The pattern of low P_n velocities under the High Sierra and high P_n velocities under Coso more closely resembles the 0- to 30-km pattern from the teleseismic inversion than the 30- to 70-km pattern (Plate 3 and Figure 8). This, combined with the nearly uniformly thick crust between about 35°15' and 36°30'N inferred from the P_n observations, leads us to infer that variations in the compensating mass under the Sierra along a north-south profile must lie in the upper mantle (or the "7.x" wedge layer). Further experimentation with

Fig. 14. Three idealized mechanisms for compensating a mountain range: the Airy root, "crustal Pratt" root, and "mantle Pratt" root.

a East-West Profile
through the Sierra Nevada
Comparing Different Compensation Styles

Fig. 15. (a) East-west profile across the Sierra Nevada at 36°30'N illustrating the gravity signal from a crustal (or "Airy") root, from lateral density variations in the entire crust ("Pratt crustal root"), and from lateral density variations in the upper 30 km of the mantle ("Pratt mantle root"). The density contrast at the Moho is assumed to be 0.5 Mg m⁻³ for the Airy root and the depth of the Moho is assumed to be 30 km for the Pratt roots. The density variations are calculated assuming that the smoothed Sierran topography (in Figure 15b) [Diment and Urban, 1981] is in perfect local isostatic equilibrium. Profile line is located in Plate 4; the Sierran crest is near 0 km. (b) Smoothed topography and compensation variations for the east-west gravity profile.

b East-west topography
and isostatic density anomalies
(assuming local isostatic equilibrium)

Fig. 16. (a) North-south section averaging Bouguer gravity from $118^{\circ}15'$ and $118^{\circ}30'W$. Compensating density variations as in Figure 15. Density variations for all the calculated gravity anomalies extend 55 km away from the profile line. Profile lines are located on Plate 4. (b) Topography and compensation variations along the north-south profile.

alternate parameterizations for the teleseismic inversion indicates that, if the Sierran crust lacks any variations, the high velocities in the 30- to 70-km layer (Plate 3) could largely reside at depths closer to 50–80 km without increasing the variance over the structure in Plate 3. The 30–50 km depth range in this alternate structure lacks any significant velocity variations.

Considering this information, it seems likely that some part of the variation along this profile comes from variations in the lower crust (the 7.x wedge) or the upper mantle. Neither teleseismic nor P_n observations can rule out lateral variation in the crust, but the absence of a clear change in the surface geology likely to accompany a large crustal density variation north of about $35^{\circ}15'$ suggests that such variations are likely to be much smaller than on the east-west profile. Additionally, Jones [1987] inferred from a dissected pediment surface that this part of the Sierra had been tilted up to the north in late Cenozoic time, which seems incompatible with a driving force emplaced in Mesozoic time.

The difference in P_n velocities (or thickness of the 7.x wedge) between the Sierra and both Coso to the east and the Great Valley to the west suggests that a large low-velocity body probably lies in the mantle under the Sierra. To the west of the range, P_n velocities are about 8.0 km/s [Holbrook and Mooney, 1987; Mooney and Weaver, 1989; Oppenheimer and Eaton, 1984], and our observations indicate P_n velocities of 7.6 km/s under the Sierra and about 7.9–8.0 km/s to the east. These observations mirror those made by Hearn and coworkers [Hearn et al., 1991; Hearn and Clayton,

1986] on the basis of tomographic inversions of regional P_n arrivals in southern California and the western United States. In the absence of partial melt, such velocity contrasts can be associated with density contrasts of the order of 0.03 to 0.10 $Mg\ m^{-3}$ if thermal variations or phase changes are responsible for the velocity variations [Aki, 1982; Birch, 1961]; thus we shall assume that the 7.6 km/s material is about 0.10 $Mg\ m^{-3}$ less dense than its surroundings.

As noted earlier, the elevation and Bouguer gravity anomaly associated with the High Sierra are roughly halved going south to the Lake Isabella region; we infer that this portion of both the relief and gravity anomaly of the High Sierra is produced by density variations in the mantle. The other half of the relief and gravity anomaly probably comes from the density variations observed in the upper crust across the western foothills. The east-west gravity anomaly can be reproduced by a simple structure combining a mantle anomaly shaped as a rectangular prism between 30 and 70 km depth, 110 km wide and 0.1 $Mg\ m^{-3}$ less dense than the surrounding mantle with a lateral variation in the upper 15 km of the crust of about 0.2 $Mg\ m^{-3}$, a value similar to that actually measured by Oliver et al. [1987] (Figure 14). We will address the implications of this model for our teleseismic results after noting other evidence suggesting that part of the support of the Sierra lies in the mantle.

Mantle anomalies and regional refraction results. Seismic structures derived from refraction profiles can be used to support

our inferred partition of the density anomaly that supports the Sierra. Recent U.S. Geological Survey seismic refraction profiles in the San Joaquin Valley provide good control on the crustal structure to the west [Holbrook and Mooney, 1987], but the structure in the Sierra is poorly known at present. We have constructed two possible seismic velocity structures for the southern Sierra Nevada, one generally based on a structure suggested by J. Eaton and used by Jones and Dollar [1986] to locate earthquakes in the southern Sierra Nevada, the other based on Prodehl's [1979] 1-D interpretations of seismic refraction lines running north and northwest from China Lake (Table 4). We require the two structures to have a total vertical travel time through the crust below sea level of 5.12 s (equal to 32 km thickness / 6.25 km/s mean velocity), about the same as that inferred from our analysis of the P_n travel times. These are also compared with two interpretations of the profile just east of Death Valley between NTS and Ludlow, California [Gibbs and Roller, 1966; Prodehl, 1979] (Table 4); we use the structure near the Ludlow end of the profile because it agrees more closely with the two-way travel time to the Moho of about 10 s in the Death Valley area [Serpa et al., 1988].

Using a modified Nafe-Drake curve [Ludwig et al., 1970], we derive density structures from the velocity structures and find that the mean density in the Sierran crust appears to be about 0.1 Mg m⁻³ lower than in the San Joaquin Valley. This is only half of the difference necessary to support the range by density variations in the crust, as is illustrated by calculating the elevation at the top of each structure assuming a homogeneous mantle and local isostatic equilibrium. Holding the elevation of the San Joaquin Valley section at sea level, we find that this procedure produces approximately the correct elevation (about 700 m) for the Ludlow-NTS line. The estimated Sierran elevations of 1 to 1.5 km fall 1 to 1.5 km short of the smoothed elevation of the high part of the range (Table 4). This shortfall cannot be eliminated as long as a single velocity-density relation similar to either this modified Nafe-Drake curve or a linear relation [e.g., Birch, 1961] is applied to all of the velocity structures equally. Although systematic variations from these relations are quite possible (and even likely in the presence of ultramafic rocks), the modified Nafe-Drake velocity-density relation produces densities that in turn produce gravity anomalies in good agreement with those observed over a seismic profile in western Nevada [Thompson et al., 1989]; the velocity variations observed along that profile are comparable to the differences between the Sierra and the adjoining areas. Thus these refraction profiles over the region tend to support our inference that about half of the isostatic support for the High Sierra comes from below the upper 30 km of the crust.

The existence of a "7.x" km/s layer at the base of the crust would reduce the discrepancy but not eliminate it. If the estimates from the two-refractor structure are used, adding a 15-km-thick 7.5 km/s (3.09 Mg/m³) layer to the base of the Sierran structures of Table 4 adds about 500 m to the elevation supported by crustal buoyancy. An additional 250 m is added if the depth to the top of the asthenosphere is flat over the area. Only a few hundred additional meters of elevation support would be required, but we note that the 7.5 km/s layer might be less buoyant if this material is mafic to ultramafic, for such rocks have a greater density than would be inferred from the Nafe-Drake curve [Birch, 1961].

Reconciling low-density upper mantle to the teleseismic observations. We tested the possibility that a large low-velocity body is in the upper mantle beneath the High Sierra similar to that described above (see section on tests of a priori structures) and found that such a feature increases the variance of the teleseismic travel time residuals by about 20 to 30% compared with the best

TABLE 4. Seismic Velocity Structures in the Vicinity of the Southern Sierra Nevada

Bottom of Layer, km (Depth Below Mean Sea Level)	P Velocity, km/s	Density, Mg/m ³
<i>San Joaquin Valley [Holbrook and Mooney, 1987] (Southeastern end of profile)*</i>		
3	2.2	2.01
5.5	6.1	2.72
9	6.5	2.81
12	6.6	2.84
16	6.65	2.85
29	7.2	3.01
<i>Southern Sierra Nevada (Modified From Work of J. Eaton, Cited by Jones and Dollar [1986])†</i>		
1	3.5	2.3
4	5.8	2.66
22	6.2	2.74
31.8	6.9	2.91
<i>Southern Sierra Nevada (Modified from Prodehl [1979])‡</i>		
3.7	5.05	2.54
5.3	6.00	2.69
8.2	6.09	2.72
10.1	6.17	2.73
11.8	6.24	2.74
22.4	6.0	2.69
27.0	6.48	2.81
31.1	6.72	2.87
<i>Ludlow toward NTS (Death Valley) [Prodehl, 1979]§</i>		
4.5	5.00	2.21
7.3	6.05	2.71
9.2	6.12	2.72
14.1	6.22	2.74
20.2	6.34	2.77
22.9	6.45	2.80
25.1	6.67	2.85
26.5	6.99	2.95
27.1	7.30	3.04
28.3	7.65	3.17
<i>Ludlow toward NTS (Death Valley) [Gibbs and Roller, 1966] </i>		
0.9	2.50	2.10
13.9	6.1	2.72
26.9	6.8	2.89

Densities inferred using relation of Ludwig et al. [1970]. Mantle (lithosphere) density defined as 3.25 Mg m⁻³. In the footnotes, $\bar{v}$ is mean crustal velocity, $\bar{\rho}$ is mean crustal density, CIE is crustal isostatic elevation, and L_m is the thickness of mantle lithosphere using relations of Lachenbruch and Morgan [1990] and assuming an asthenospheric density of 3.20 Mg/m³.

* $\bar{v} = 6.37$ km/s, $\bar{\rho} = 2.82$ Mg m⁻³, CIE = 0 m (= local mean elevation), $L_m = 71$ km

† $\bar{v} = 6.21$ km/s, $\bar{\rho} = 2.77$ Mg m⁻³, Mean Elevation = 2500 m, CIE = 1440 m, $L_m = 4$ km

‡ $\bar{v} = 5.96$ km/s, $\bar{\rho} = 2.72$ Mg m⁻³, Mean Elevation = 2500 m, CIE = 1640 m, $L_m = 16$ km

§ $\bar{v} = 6.16$ km/s, $\bar{\rho} = 2.71$ Mg m⁻³, Mean Elevation = 700 m, CIE = 560 m, $L_m = 62$ km

|| $\bar{v} = 6.25$ km/s, $\bar{\rho} = 2.77$ Mg m⁻³, Mean Elevation = 700 m, CIE = 260 m, $L_m = 43$ km

fit structure (Table 3). Introducing such a large low-velocity body into the upper mantle also increases the velocities of a high-velocity body beneath the low-velocity body. We can reconcile the low P_n velocities, low gravity anomalies, and relatively early teleseismic arrivals in two ways: (1) place a low-density body in the

upper mantle (about 30–80 km depth) with strong seismic anisotropy, faster (by about 6%) for vertically traveling (teleseismic) rays than for horizontally traveling (P_n) rays or (2) place a low-density, low-velocity body in the upper mantle over a high-velocity body with unknown density. Certainly, comparing the P_n results with the 30–70 km layer in the teleseismic inversion (Plate 3) lends credence to the possibility that the upper mantle is seismically anisotropic, and *Savage and Silver* [1993] have documented large-magnitude horizontal anisotropy in the northern Sierra Nevada, but there is no good reason why a highly anisotropic medium would be buoyant and support the southern High Sierra. Thus we opt for an isotropic medium and option 2.

We reconcile the seismic and gravity observations by choosing a structure with a low-velocity, low-density body in the uppermost mantle (30–60 km) and a high-velocity body beneath it (60–100 km) that has either the same or lower density than its surroundings. Minimizing the thickness of the low-velocity upper mantle will best fit the teleseismic observations. The depth of 60 km was considered the shallowest both to provide an adequate thickness of buoyant material and to fit the lack of an observed first arrival in the P_n observations from the high-velocity material below the low-velocity uppermost mantle. A simple density structure based on this interpretation can fit the observed east-west gravity variations across the Sierra (Figure 17). This choice demands that the relation between seismic velocity and density in the uppermost mantle is not unique. Variations in the mean atomic weight of materials produces different velocity-density relationships [e.g.,

Birch 1961]; a mechanism for such variation is the depletion of upper mantle material through the derivation of melt [e.g., *Jordan*, 1979; *Dawson et al.*, 1990]. Experimental and theoretical studies of the velocity and density properties of residuum from mantle melt support the inverse correlation of velocity and density suggested here [Akimoto, 1972; *Jordan*, 1979]. The low P_n velocity under the Sierra inferred from the single refractor structure and the moderate to high heat flow recently observed under the eastern Sierra near Lake Isabella indicates that the underlying mantle either is warmer than that to the west or became warm earlier than that to the west [Black and Braile, 1982; *Saltus and Lachenbruch*, 1991]. Thus we suggest that the uppermost mantle under the Sierra is unusually warm and composed in part of melt derived from a volume beneath it; the depleted mantle would then have higher velocities but lower densities than the surrounding mantle (Figure 18).

Implications of a possible second refractor. The preceding discussion has presumed that the single-refractor structure is correct, but most of its conclusions would still apply if the two-refractor structure better represents the southern Sierra. As before, higher-velocity material would have to lie under the low-velocity “7.x wedge” between refractors in order to reconcile this structure to the teleseismic observations. This interpretation differs most from that above if this “wedge” is considered to be within the crust and was formed in the Mesozoic. If so, and if the underlying mantle is relatively cool and thus presumably antibuoyant, this would be a modified version of the classic crustal root for the Sierra Nevada, but this inference is difficult to reconcile with recent heat flow measurements indicating an elevated heat flow in the southeastern Sierra [Saltus and Lachenbruch, 1991] and seismological and geological observations indicating a close association between the tectonics in the Sierra and in the Basin and Range [Jones, 1987; Jones and Dollar, 1986]. Additionally, if the mantle is antibuoyant beneath the Sierra, the isostatic calculations presented earlier would require the presence of a significant buoyant body that is not present within the seismic structure. An intriguing observation is the inference of a similar “double Moho” of 7.5 km/s material over 7.9 km/s mantle under the Wasatch front in Utah [Loeb, 1986; Loeb and Pechmann, 1986; Pechmann et al., 1984; Smith et al., 1989]; if this has the same cause as the similar sub-Sierran structure, it suggests a Cenozoic origin for such a feature, because the Wasatch front area, unlike the Sierra, is not a batholith, but both are in similar tectonic environments on the margins of the Basin and Range. Thus our preferred alternative is that this material, whether called crust or mantle, has been derived from the mantle in Cenozoic time and that, in some sense, this is the igneous “rift pillow” apparently absent from regions to the east [Beck and Zandt, 1991; Patton et al., 1991] (G. Zandt, et al., unpublished data, 1992). The tectonic implications of this interpretation differ little from that of the “single refractor” structure, for it still suggests that uplift of the Sierra is due to the creation of warm, buoyant material beneath the range in the past 10–20 m.y.

Implications for the Tectonic History of the Sierra

Partitioning the isostatic support for the Sierra between crust and upper mantle also tends to partition the uplift of the range between Mesozoic and late Cenozoic time. Density contrasts in the upper crust were clearly in place by the close of the Mesozoic; in the absence of density anomalies in the mantle and assuming local isostasy, we would expect the range to have had elevations between about 1.0 and 1.5 km; it is quite possible the elevations

Fig. 17. East-west profile across the Sierra Nevada (same as Figure 15) illustrating fit possible with a simple combination of lateral density contrasts in the crust and mantle. The mantle anomaly extends from 30 to 70 km depth and is 0.10 Mg m^{-3} less dense than its surroundings. The crustal lateral density contrasts extend from sea level to 15 km depth. The intentional misfit near 20 km reflects the gravity anomaly associated with basin fill under Owens Lake.

Fig. 18. Perspective block diagram illustrating principal interpretations of this study. No vertical exaggeration (except for the topography); bottom of the blocks is at 250 km depth. Block faces are aligned along north-south (left) and east-west orientations. Stippling in the crust reflects density variations shown in Figure 17.

were somewhat lower because of the elastic strength of the crust [Chase and Wallace, 1986, 1988]. These paleo-elevations are similar to those inferred by Huber [1981, 1990] farther north and Christensen [1966] in this region. Because low P_n velocities tend to be associated with high heat flow [Black and Braile, 1982], the low heat flow over much of the Sierra through most of the Cenozoic [Dumitru, 1990; Saltus and Lachenbruch, 1991] indicates that the velocity and density variations inferred to lie in the upper mantle were acquired in only the past 10 to 20 m.y.

The creation of a new low-density body under the High Sierra in late Cenozoic time obviates the need, suggested by Chase and Wallace [1986, 1988], for an extremely strong lithosphere through the Cenozoic to suppress a large, Mesozoic crustal root. Removing the large crustal root also simplifies the lithospheric history to the west, as inferred by Rentschler and Bloch [1988] from flexural models of the San Joaquin Valley through the Tertiary. They assumed an even thicker Sierran crust than that used by Chase and Wallace and then found that to match the depositional history in the San Joaquin Valley, they needed to weaken the lithosphere on the west side of the Sierra starting at about 8 Ma. This weakening could reflect instead the addition of a load beneath the Sierra about this time; a generally weaker lithosphere could be assumed for times before 8 Ma because there would not be a need for a strong lithosphere to suppress a large crustal root. This would tend to make the strength of the Sierran lithosphere about the same or somewhat weaker than that under the San Joaquin Valley to the west, a reasonable situation considering the more quartz-rich (and thus weaker) rheology of the Sierra compared to that of the basement of the San Joaquin Valley.

We infer that low-velocity, low-density upper mantle beneath the High Sierra lying over high-velocity upper mantle reflects derivation of melt from the upper mantle from beneath the Sierra; this implies that the mantle lithosphere has been greatly thinned beneath the Sierra (Figure 18). There are only some small volcanic centers in the southern Sierra, but significant eruptions of volcanic rocks have occurred over the past 15 m.y. in the Basin and Range to the east. We propose that the melt derived from the upper mantle under the Sierra migrates up and east through the lithosphere, erupting in the adjoining Basin and Range. The preferential eastward migration of melt might be an effect of west dipping structural fabrics through the crust in this region [e.g., Wernicke *et al.*, 1988] or could even be entrained in wholesale flow of the lower crust to the east, as suggested by Wernicke [1990, 1992] and Walker and Coleman [1991]; migration of melt along a west dipping fault has been proposed in the Death Valley area from the analysis of reflection profiles across the area [Serpa *et al.*, 1988]. This kind of structure agrees with the inference that extension at this latitude is localized under the Sierra Nevada in the upper mantle and under Death Valley in the crust (Figure 18) [Jones, 1987; Wernicke, 1985].

CONCLUSIONS

Teleseismic and P_n travel times observed at temporary and permanent stations in the southern Sierra are inconsistent with the presence of thick (>40 km) crust under the southern High Sierra as previously envisioned. The travel time residuals of teleseismic rays penetrating the crust and upper 60 km of the mantle beneath the High Sierra are not noticeably late (Plate 2 and Figure 7), thus an isolated low-velocity, low-density body sufficient to support the southern Sierra does not exist under this part of the range. The most dramatic anomaly seen in the teleseismic residuals is from a

large high-velocity body under the Sierran foothills somewhat west of our network; this body appears to extend from about 100 to 200 km depth (Figure 18). The limited north-south extent of this anomaly indicates that it is not responsible for the large north-south physiographic features (the Sierra Nevada, the San Joaquin Valley, and the Coast Ranges) in the area.

Although a small low-velocity body under the High Sierra was found in our best fit inversion of the teleseismic arrival times (Plate 3), a large body sufficient to support the range produces a significant misfit to the observations (Table 3); a velocity anomaly in the mantle ("mantle root") fits the observations only slightly better than a crustal root. We use arrival times of P_n phases from earthquakes to the north and south to help distinguish among these structures. Assuming that these arrivals come from a single refractor, the crust is 33 ± 5 km thick under the High Sierra and the upper mantle P wave velocity is about 7.6 km/s. P_n velocities to the east are greater, perhaps 7.9–8.1 km/s. If refractions from north and south are from different refractors, then a ~20-km-thick wedge of material with a P-wave velocity between about 7.2 and 7.5 km/s could underlie the southern High Sierra, but because this structure requires complications not present in the single refractor interpretation, we prefer the single-refractor interpretation.

We chose to reconcile the P_n , teleseismic, and gravity data by suggesting that the mass deficit supporting the southern High Sierra is present in both crust and mantle. In the crust, we infer from surface geology and gravity and density measurements that the upper 15 km of the crust beneath the Sierran crest is about 0.2 Mg m⁻³ less dense than that beneath the foothills to the west (Figure 18). Because of the P_n observations and the Bouguer gravity anomalies, we infer that low-velocity, low-density upper mantle exists between about 30 and 60 km beneath the range north of about 36°N. Because of our teleseismic arrival times and the Bouguer gravity anomalies, we believe that high-velocity, low-density material lies under the low-velocity material to depths of perhaps 100 km. Low velocities might then lie under that material under the region west of Owens Valley.

Our interpretation of these anomalies (Figure 18) is that they represent greater thinning of the mantle lithosphere beneath the Sierra than of that under the Basin and Range; this thinning corresponds to the deeper levels of crustal extension centered in the Death Valley region. This thinning of the lithosphere has caused the asthenosphere to approach the crust beneath the High Sierra. Melt derived from the upper 40–60 km of the asthenosphere has risen upward, leaving higher-velocity but lower-density residuum between 60 and 100 km depth. Mantle above this depth might not be substantially depleted but is probably quite warm, producing the low-velocity, low-density layer immediately beneath the crust. This forms the reservoir of mantle melt that migrates up and east toward volcanic centers in the western Basin and Range. Creation of this deformational system provided the buoyancy necessary to raise the High Sierra 1–2 km to its present elevation.

To the west of the Sierra, the lateral contrasts in the crust were emplaced in the Mesozoic during the construction of the crust in this region. These contrasts provided the buoyancy to support the ancestral Sierra through much of the Tertiary. The substantial anomaly under the western foothills (the "Isabella anomaly") lacks any obvious connection to the tectonics at the surface despite the large velocity contrast (~+4%) with its surroundings. We have suggested that it is downwelling lithosphere, probably derived from the south but possibly also from the north. This process might have started in the Miocene, when extensional tectonism deformed the southern San Joaquin Valley and Tehachapi Mountains; this deformation might have introduced an instability

in the cold mantle lithosphere beneath the southern Sierra, permitting it to start to move downward. We have indicated a counterflow to this process under the Tehachapi Mountains; this is based on poorly constrained low velocities between about 30 and 100 km depth.

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