CRUSTAL STRUCTURE IN NORTHERN AND CENTRAL CALIFORNIA FROM SEISMIC EVIDENCE *

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The location of California along the continental margin, and the differentiation of the State into a number of strikingly dissimilar geologic provinces, have led to the expectation that within the State the gross composition and structure of the crust should vary considerably from place to place. Documentation, by seismic means, of major differences in crustal structure from region to region began more than 25 years ago, when Perry Byerly adduced evidence that the Sierra Nevada has a root that extends much deeper into the mantle than the base of the crust beneath the Coast Ranges. Recent seismic refraction studies by the U.S. Geological Survey in California and adjacent portions of the western United States have explored regional variations in crustal structure in greater detail.

This article will outline the nature and limitations of the principal seismic methods used in detailed investigations of the crust, and it will also review the seismic coverage available in northern and central California, summarize the observations and results so far obtained, and convey some impression of the reliability of those results. In the interest of brevity, no discussion of important supplementary techniques, such as analysis of P-wave delays and surface-wave dispersion, will be included. Not all of the earthquake and refraction studies carried out in the region will be considered, but only those which relate most directly to the present theme.

METHODS

Until quite recently, the opportunity to conduct detailed seismic studies of the deep crust and upper mantle of the earth has been limited to regions that have both frequent, shallow earthquakes and specialized networks of seismographs. The principal task of such networks, like that of the University of California in the central Coast Ranges, has been to study the earthquakes themselves; and stations have been concentrated in zones of high seismicity at sites that facilitate locating the epicenters of small to moderate earthquakes. Information on crustal structure in these regions has been obtained as a byproduct of earthquake-wave traveltime studies that have been carried out to improve the accuracy of epicenter and focal depth determinations.

Because the number of recording stations normally is small, data from many earthquakes with epicenters scattered at random through the network are usually combined to establish the empirical traveltime curves from which a mathematical crustal model is calculated. To determine the traveltime curve of waves refracted along the top of the mantle, observations at distant stations located outside the primary region of study, even in different geologic provinces, are frequently used. To interpret traveltime curves established by such heterogeneous data, it is customary to assume a uniform structure for the entire region studied. Traveltime perturbations caused by regional variations in structure are normally too poorly defined to bring such structural variations to light. The crustal model resulting from such an investigation is a poorly defined average of conditions over the entire region containing the earthquakes and the recording stations employed. Moreover, serious uncertainties in focal depth of most of the earthquakes studied further decrease the reliability of values obtained for the total thickness of the crust and for the depths of boundaries within it.

Many of the obstacles that impair determinations of crustal structure by near-earthquake traveltime studies are removed when accurately timed and located large explosive charges are substituted for natural earthquakes, and when well-laid-out profiles of portable high-performance seismic systems are used in place of regional near-earthquake seismic networks. Investigations can be carried out in aseismic as well as seismic areas, profiles can be laid out with regard to major geologic provinces, problems arising from uncertainties in the time and location of the seismic source are eliminated, and observations that are sufficiently dense and continuous along the profile to permit "tracking" of individual seismic phases through their entire range of occurrence can be obtained. On the other hand, the high cost of explosives requires the use of the smallest practicable charges for distant observations, and waves refracted from the mantle usually cannot be traced to the large distance that would be desirable. This difficulty is largely offset by the use of "later arrivals," including waves reflected from the top of the mantle and intermediate boundaries within the crust. Such waves can be identified and traced far more reliably on seismograms from profiles with close-spaced instruments along a single line than on the seismograms of scattered earthquakes recorded on instruments spaced irregularly throughout a large area.

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Since near-earthquake and explosion-seismic data are interpreted on the same theoretical basis (the analysis of traveltimes of refracted and, to a lesser extent, reflected waves), the two techniques are limited by some of the same factors. A region with a crust that is relatively homogeneous, or that varies only in a simple manner over a distance of 8 to 10 times the thickness of the crust, is essential if arbitrary assumptions about crustal parameters are to be avoided. Near-surface conditions, especially the thickness of very low velocity rocks, should be nearly uniform in the region studied. With unfavorable velocity-depth relationships, such as appear to exist in many regions, "masked" layers that are very difficult to detect may exist in the lower crust. Particularly complete (or better, moderately redundant) systems of observations are needed to detect such layers, which are not represented by "first arrivals" over some part of the profile.

Traveltime curves of compressional waves in most regions consist of two principal branches that are readily distinguishable because they represent the first waves recorded on the seismograms over considerable distances. Waves refracted through the upper, sialic or "granitic," part of the continental crust below the irregular superficial blanket of lower velocity near-surface rocks (P_g waves) fall close to a traveltime line with a reciprocal slope (velocity) of 6 km/sec and an intercept of about 1 sec. They are first arrivals from the vicinity of the shotpoint, or epicenter of a shallow earthquake, to a distance of 100 to 250 km, depending on the thickness of the crust. These waves through the upper part of the crust are succeeded as first arrivals at larger distances by waves that are refracted through the upper part of the mantle (P_n waves) below the Mohorovicic discontinuity. P_n arrivals fall close to a traveltime line with a velocity near 8 km/sec and an intercept between 5 and 10 sec, depending on the thickness of the crust. Other phases, with velocities in the range of 6.5 to 7.0 km/sec, that have been refracted horizontally through an "intermediate layer" in the lower crust can sometimes be detected as first arrivals at intermediate distances between the domains of P_g and P_n. More commonly, at least in continental regions, such intermediate phases (P^*) do not appear as first arrivals, and the existence of an intermediate layer is inferred from later arrivals that appear to be reflected from, or refracted along, its upper surface.

Although intermediate layers are difficult to detect and do not strongly influence first-arrival traveltimes, they tend to be very prominent in crustal models computed from seismic data. Healy (1963) has illustrated how a complex crust 37 km thick can produce the same first-arrival traveltimes as a single layer crust only 28 km thick, though such an extreme case seems unlikely. When rock compositions or densities are inferred from seismic-wave velocities, extremely serious errors in estimates of average values of these parameters in the crust will result if intermediate layers go undetected.

EVIDENCE ON CRUSTAL STRUCTURE FROM NEAR EARTHQUAKES

The only area in northern and central California that has had a sufficient concentration of earthquakes and seismograph stations to permit the calculation of a reasonably well-documented near-earthquake crustal model is the part of the Coast Ranges southeast of San Francisco.

The principal analysis of crustal structure in this region, which has been studied by Perry Byerly and his students at the University of California for several decades, was reported by Byerly (1939). He concluded that the crust is about 32 km thick and is divided into two principal layers. The upper crust appeared to vary in thickness from about 10 to about 20 km, and the velocity of P waves in it (established along paths that lay east of the San Andreas fault in a region with Franciscan basement rocks) was found to be 5.6 km/sec. P waves through the lower part of the crust, which appeared to have a velocity of about 6.7 km/sec, were nowhere observed as first arrivals; so their traveltime curve was not firmly established.

To determine the traveltime curve of waves through the upper mantle (P_n), Byerly was obliged to use observations at stations in southern California in addition to those in the central Coast Ranges. The P_n velocity so obtained (8.02 km/sec) and the total thickness of the crust reported were, therefore, influenced by conditions outside the region in which the velocity in the upper crust was obtained.

The most important seismic contribution to our understanding of central California crustal structure outside of the Coast Ranges was also made by Byerly (1938), who demonstrated that P_n waves from earthquakes along the San Andreas fault in the central Coast Ranges arrived late at stations in Owens Valley. He attributed this delay to a thick root of low-velocity crustal rock extending into the mantle beneath the high southern Sierra Nevada.

EVIDENCE FROM EXPLOSION SEISMIC REFRACTION PROFILES

In the fall of 1961 the U.S. Geological Survey began an extensive investigation of crustal structure in the western United States as part of the VELA Uniform project of ARPA. In northern and central California, major seismic refraction profiles were run longitudinally in the Coast Ranges between San Francisco and Santa Monica, with an intermediate shotpoint at Camp Roberts, and in the Sierra Nevada from Shasta Lake to China Lake, with an intermediate shotpoint at Mono Lake (fig. 1). Two other profiles were run transverse to these (and to major structures of the region): one between San Francisco and Fallon, and one between San Luis Obispo and the Nevada Test Site, through Owens Valley (fig. 1). Additional profiles were run south and east of these to study the bordering geologic

Figure 1. Map showing shot points and seismic refraction profiles made by the U.S. Geological Survey in the northern and central part of California.

provinces: the Transverse Ranges, Mojave Desert, and Great Basin.

Analyses of the Coast Range profile and the northern transverse profile have been reported previously (Healy, 1963; Eaton, 1963) and will be only summarized here. Results from these profiles will be compared with those from the Sierra Nevada profile, which will be more fully documented because it has not been reported previously. Both longitudinal profiles were sufficiently long and well recorded that we may accept their results with reasonable confidence. A few problems of wave identification and interpretation remain, but it is unlikely that the recording of additional explosions and reinterpretation of the augmented data would lead to significantly different results. The interpretation of the transverse profiles, however, is far less certain. They serve chiefly to set limits on where and how major units of the crust vary between the longitudinal profiles. Additional profiles along the strike of the major geologic structures will be required to resolve problems that have been identified from the transverse profiles.

Descriptions of instrumentation (Warrick and others, 1961) and field procedures (Jackson and others, 1963) used in the Survey's crustal refraction work, and discussions of techniques used in detailed analysis of the field data (Eaton, 1963), have been reported previously. The principal results of such work in the western United States, including California, were summarized by Pakiser (1963), who discussed the gross geologic implications of the pattern of crustal thickness and P_n velocities obtained.

Sierra Nevada Profile

Series of nitrocarbonitrate charges ranging in size from 2,000 to 6,000 lbs were detonated in Shasta Lake, Mono Lake, and in drill holes near China Lake, and were recorded along a pair of end-to-end profiles running the length of the Sierra Nevada. The northern profile crosses a portion of the Cascade Mountains west of Mount Lassen and then runs down the crest of the Sierra Nevada just west of Lake Tahoe and on to Mono Lake. The southern profile runs along the east face of the Sierra Nevada southeast of Mono Lake, along the western edge of Owens Valley, to China Lake. Because the eastern face of the Sierra Nevada is convex eastward in this region, refraction paths from the Mono Lake and China Lake shotpoints to recording points at distances greater than 150 km pass beneath the highest peaks of the range.

In the interpretation of the recorded seismograms, record sections were constructed to facilitate correlation of wave arrivals and identification of phases. Traveltimes of the principal phases were measured on the monitor records (or on playbacks, with filter corrections subtracted) and were plotted on a reduced scale (that is, T-$\Delta/6$ vs Δ, where T is traveltime and Δ is distance) to establish traveltime curves. Phases used in the calculation of the seismic cross section were either first arrivals over a reasonable distance or were second arrivals that were strongly supported by reflected phases. Parameters of the traveltime lines of these phases, and a statement of the nature of the wave arrivals and the approximate range over which they were recorded, are given in table 1.

The seismic cross section calculated from these curves (fig. 2) shows a three-layered crust with a maximum thickness of about 54 km beneath the high southern part of the range. Beneath a thin weathered(?) zone, the speed of longitudinal waves (in the granite) is very close to 6.0 km/sec. At a depth of about 14 km the velocity appears to increase to 6.4 km/sec; and at about 27 km it increases to 6.9 km/sec. The 6.4 km/sec layer produced first arrivals from 135 to 205 km on the Mono Lake to China Lake profile, and the 6.9 km/sec layer produced first arrivals from 160 to 250 km on the China Lake to Mono Lake profile.

Beneath the crest of the Sierra Nevada west of Lake Tahoe the crust thins to about 47 km and the top of the 6.4 km/sec layer rises to about 10 km. The best determination of the velocity of P_n was between Shasta Lake and Mono Lake, where it appears to be slightly more than 7.9 km/sec.

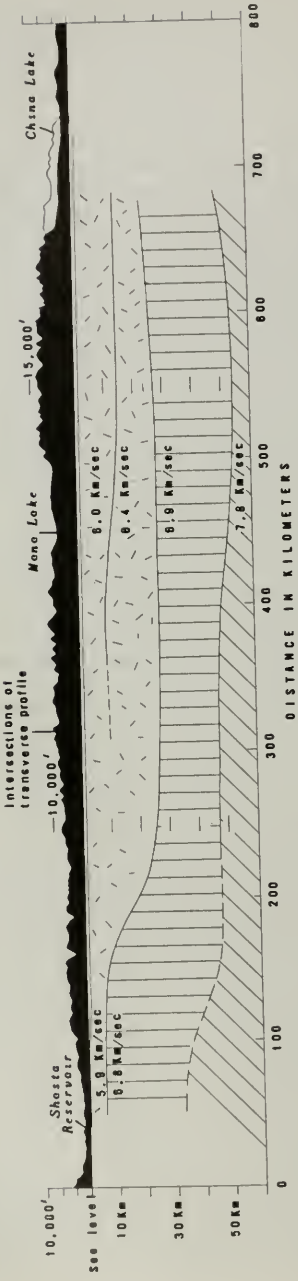

Figure 2. Longitudinal seismic cross section through the Sierra Nevada from near Shasta Reservoir to China Lake.

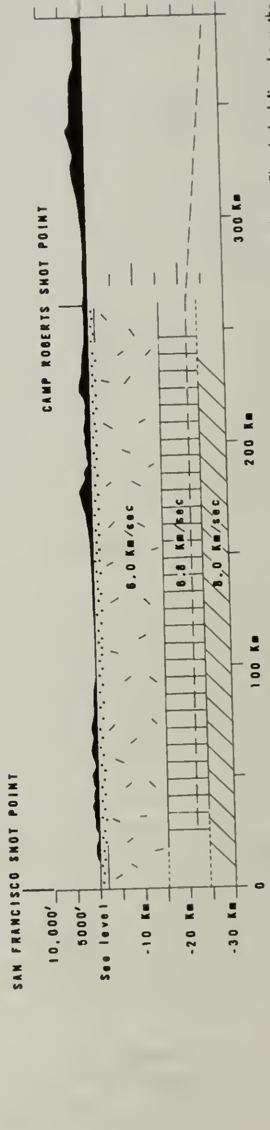

Figure 3. Longitudinal seismic cross section through the Central Coast Ranges from San Francisco to Camp Roberts. The dashed line shows the single-layer interpretation of the first-arrival traveltimes. Solid lines show the two-layer interpretation with the thickest intermediate layer of velocity 6.8 km/sec that would not produce first arrivals on some part of the profile.

Table 1. Traveltime curves of the Sierra Nevada refraction profile.

Phase	*Traveltime line*	*Nature and range of arrival*
(Shasta Lake to Mono Lake)		
P_g	$0.4 + \Delta/5.9$	First arrival, 10 to 40 km
P^*_a	$1.4 + \Delta/6.8$	First arrival, 40 to 80 km
P^*_b	$1.7 + \Delta/6.8$	First arrival, 90 to 140 km
P^*_c	$3.0 + \Delta/6.8$	First arrival, 215 to 250 km
P_n	$8.0 + \Delta/7.9$	First arrival, 250 to 405 km
(Mono Lake to Shasta Lake)		
P_g	$1.2 + \Delta/6.0$	First arrival, 10 to 85 km
P^*_1	$1.9 + \Delta/6.3$	First arrival, 85 to 175 km
P^*_2	$4.5 + \Delta/6.8$	Reflections, 100 to 250 km
P_n	$9.2 + \Delta/8.0$	First arrival, 210 to 400 km and Shasta Lake to Mono Lake reciprocal point
(Mono Lake to China Lake)		
P_g	$1.0 + \Delta/6.0$	First arrival, 10 to 135 km
P^*_1	$2.6 + \Delta/6.4$	First arrival, 135 to 205 km
P^*_2	$4.7 + \Delta/6.9$	Reflections primarily, 100 to 230 km
P_n	$10.6 + \Delta/8.1$	First arrival, 280 to 310 km, and reflections, 150 to 250 km
(China Lake to Mono Lake)		
P_g	$1.1 + \Delta/6.1$	First arrival, 5 to 160 km
P^*_1	$2.3 + \Delta/6.3$	Reflections, 55 to 170 km
P^*_2	$4.2 + \Delta/6.9$	First arrival beyond 160 km
P_n	$8.7 + \Delta/7.7$	Reflections, 100 to 250 km, and Mono Lake to China Lake reciprocal point

Just southeast of Shasta Lake a velocity of 6.8 km/sec was encountered at a depth of only 6 km. Details of deeper structure in this region are obscure. Rapid thickening of the upper crustal layers appears to begin near the canyon of the North Fork of the Feather River about 60 km south-southeast of Mount Lassen. Very similar conditions in the upper crust have been encountered in the Snake River plain south of Boise, Idaho (Hill and Pakiser, 1963).

Coast Range Profile

To insure compatibility with the Sierra Nevada profile, the data reported by Healy for the reversed profile between San Francisco and Camp Roberts were replotted to obtain reduced traveltime curves in the manner just indicated. P_n was well recorded between 195 and 305 km southeast of San Francisco and established the line $5.9 + \Delta/8.03$ sec. From Camp Roberts, P_n was less well recorded: a line drawn through the data points (175 to 245 km) and constrained to pass through the well-established San Francisco to Camp Roberts P_n reciprocal point is given by $5.8 + \Delta/8.01$ sec. The line $1.3 + \Delta/6.0$ sec fits 10 well-recorded P_g arrivals northwest of Camp Roberts; and a line $1.4 + \Delta/6.0$ sec is compatible with (but is not established by) four P_g(?) arrivals southeast of San Francisco. The 6.1 km/sec P_g velocity obtained by Healy fits the data equally well.

Inconclusive evidence for an intermediate layer is provided by later arrivals from both shotpoints and by two somewhat early P_g(?) arrivals about 100 km north of Camp Roberts. These heterogeneous arrivals are approximately represented by the line $3.5 + \Delta/6.8$ sec, which passes through the crossover point of P_g and P_n waves from the San Francisco shotpoint. High noise levels in the Salinas Valley seriously impaired the quality of recordings that were needed to test the existence of an intermediate layer.

If evidence for an intermediate layer is ignored, the crust appears to be about 22 km thick between San Francisco and Camp Roberts (dashed line, fig. 3). It thickens southeastward from Camp Roberts to Santa Monica (Healy, 1963), where it is 35 km thick. Inclusion of the possible intermediate layer discussed above yields the structure represented by solid lines in figure 3: low-velocity near-surface zone to 1.5 to 2 km, 6.0 km/sec to 15.4 km, 6.8 km/sec to 24.3 km, and 8.0 km/sec in the upper mantle.

When comparing the P_g velocity obtained along the San Francisco to Camp Roberts profile (6.0 to 6.1 km/sec) with that reported by Byerly for the central Coast Ranges (5.6 km/sec), it is important to note that the explosion-refraction profiles sampled only the granite corridor southwest of the San Andreas fault whereas Byerly's material on $\bar{P}$ (P_g in the present notation) sampled only the region of Franciscan basement rocks northeast of the fault. Additional evidence for an abnormally low P velocity northeast of the San Andreas fault in the San Francisco Bay region was provided by the San Francisco to Fallon profile (Eaton, 1963).

The velocity of P_n beneath the Coast Ranges deduced from the refraction profiles agrees closely with that found by Byerly.

Transverse Profiles

An attempt to construct a cross-section transverse to the Sierra Nevada on the basis of limited data from refraction profiles between San Francisco and Fallon has been reported by Eaton (1963). Inadequate recordings at large distances seriously restricted interpretation of those profiles, and the maximum thickness of the Sierra root was estimated by extrapolating results obtained on the flanks of the range toward its center. Recordings from the San Francisco shotpoint were especially difficult to interpret: the thick sediments of the Great Valley, with their high noise level and very low seismic velocity, nearly blotted out the profile in the critical range of 80 to 160 km. Only between 160 and 275 km, from the western foothills to the crest of the Sierra Nevada, were adequate seismograms recorded.

Better results were obtained from the Fallon shotpoint, which provided evidence on the location of the

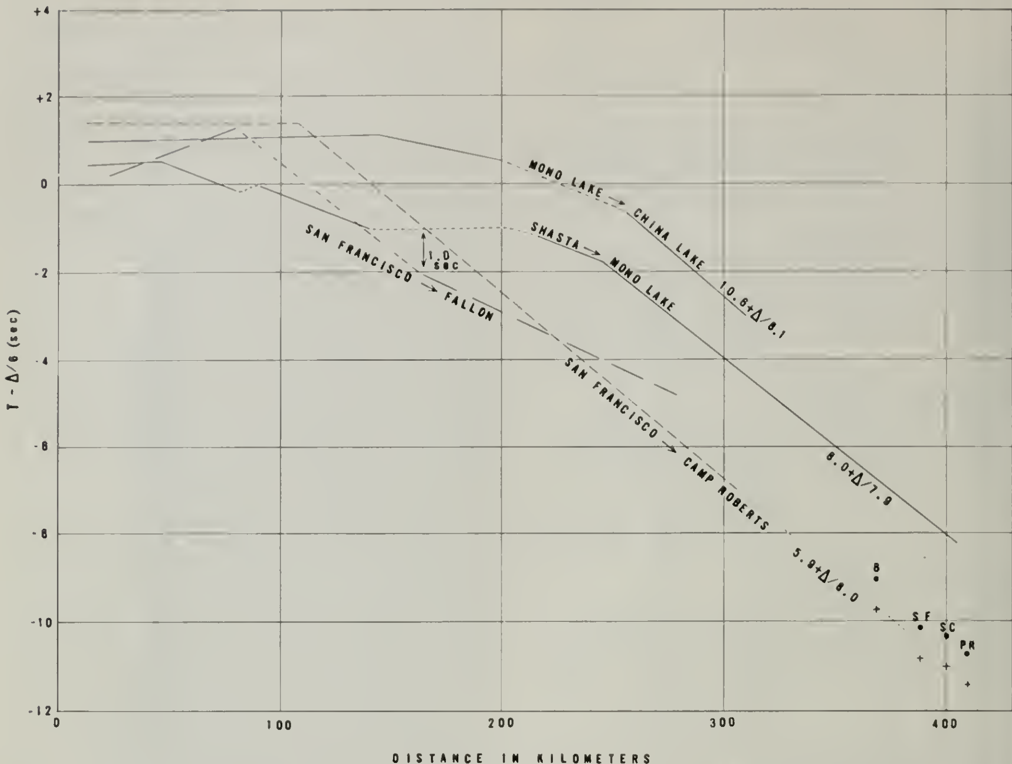

Figure 4. Reduced traveltime curves of first arrivals on selected profiles shown on figure 1. T-Δ/6, where T is traveltime in seconds and Δ is shot to receiver distance in kilometers, is plotted against Δ. Uncorrected reduced traveltimes versus distance from SHOAL to seismograph stations near the San Francisco Shotpoint are shown as solid circles (B, Berkeley; SF, San Francisco; SC, Santa Cruz; and PR, Point Reyes). After correction for diffraction around the Sierra Nevada root (fig. 5), they are shown as crosses.

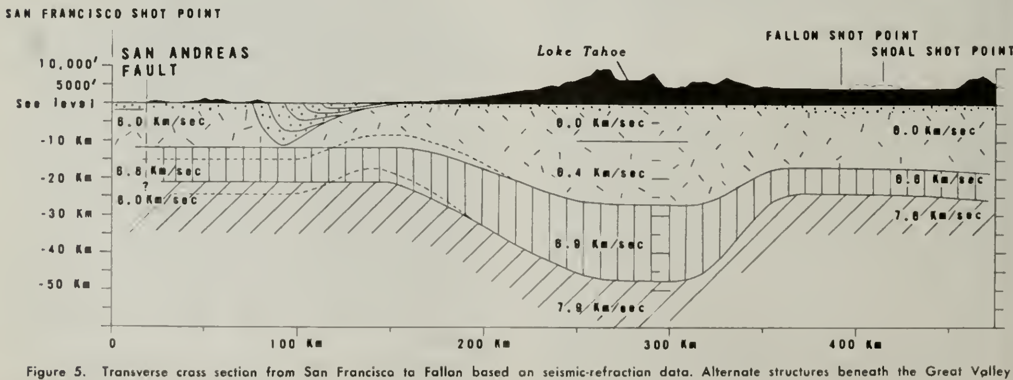

Figure 5. Transverse cross section from San Francisco to Fallon based on seismic-refraction data. Alternate structures beneath the Great Valley are shown in solid and dashed lines.

steep eastern boundary of the Sierra Nevada root. Profiles from Fallon to Owens Valley and between Fallon and Eureka further defined the eastern portion of the root and established the crustal structure of the Basin and Range province farther east (Eaton, 1963).

A more reliable picture of the structure of the crust between San Francisco and Fallon can now be drawn. Beneath the Coast Ranges and Sierra Nevada the structure is quite well determined by the longitudinal profiles; and results from the Fallon shotpoint serve to establish the structure of the eastern end of the section. Recordings between 160 and 275 km from San Francisco set additional constraints on the behavior of crustal boundaries beneath the Great Valley and Sierra Nevada foothills.

To evaluate the profile from San Francisco toward Fallon, it is useful to compare first arrivals recorded along it with those recorded along other critical profiles in the region. For this purpose, reduced first-arrival traveltime curves for the profiles San Francisco to Camp Roberts, Mono Lake to China Lake, and Shasta Lake to Mono Lake are plotted with that for San Francisco to Fallon in figure 4.

Plotting reduced time ($T-\Delta/6$) vs distance (Δ) permits the use of an expanded time scale and greatly increases the difference in slope of two lines representing slightly different velocities. At any particular distance, however, differences in arrival time on the various curves can be read directly from the graph. The near-horizontal portions of the curves at small distances represent the P_g phase, longitudinal waves propagating through the upper crust beneath superficial low-velocity materials. The nearly parallel portions of the curves at large distances represent the P_n phase, longitudinal waves with the deepest parts of their paths through the upper mantle just beneath the crust. Portions of the curves with intermediate slopes (for example, Shasta Lake to Mono Lake, Mono Lake to China Lake) usually represent waves refracted through rocks of intermediate-velocity in the lower part of the crust.

With certain reservations, the average thickness of the crust beneath the shotpoint and receiving stations is proportional to the intercept of the P_n line. Thus, for the profile San Francisco to Camp Roberts, which traverses a relatively regular crust less than 25 km thick, the P_n intercept is only 5.9 sec while for the profile from Mono Lake to China Lake, which runs longitudinally through the thick Sierra Nevada crust, the P_n intercept is 10.6 sec (fig. 4). With similar reservations, the average thickness of the upper part of the crust beneath the shotpoint and the receiving stations is proportional to the intercept of the traveltime line of any intermediate layer that might be detected. Steplike increases in the intercepts of the 6.8 km/sec intermediate layer traveltime line segments recorded southeast of Shasta Lake (fig. 4) suggest steplike increases in the thickness of the upper crust as that profile crosses from the southern Cascades into the Sierra Nevada.

The significance of the most crucial (and also the most reliable) recordings on the profile from San Francisco to Fallon is best brought out by comparison with arrivals at the same distance along the San Francisco to Camp Roberts profile. At 160 km, waves from the San Francisco shotpoint emerge at the western edge of the Sierra Nevada foothills on the transverse profile a full second earlier than they emerge at the same distance on the longitudinal profile. By 220 km first arrivals on the transverse profile have fallen back to the same time as those on the longitudinal profile, and by 275 km they have fallen nearly a full second behind.

Because these arrivals on the San Francisco to Fallon profile have an apparent velocity of about 7 km/sec, it is tempting to attribute them to a very shallow intermediate layer with such a velocity. This possibility appears to be ruled out by the persistence of the phase to the crest of the Sierra Nevada, which was shown by the longitudinal Sierra profile to have a thick section of rocks with velocities of 6.0 to 6.4 km/sec. Failure of the phase to appear as a first arrival west of the Great Valley also argues against this solution.

A more likely explanation is that this phase is P_n emerging from beneath a progressively thickening crust as the profile crosses into the Sierra Nevada. If thickening of the crust were due to an increase in thickness of a 6.0 km/sec layer in the upper crust, the Mohorovicic discontinuity would have to dip northeastward along the profile at about 11°; and if crustal thickening were due to the overall thickening of a composite crust with an average velocity of 6.4 km/sec, the dip of the Moho would be nearly 16°. Explanation of the early arrival of P_n at the western edge of the Sierra foothills requires a thinner crust beneath the Great Valley or higher crustal velocities beneath the valley and foothills, or both, than beneath the central Coast Ranges. As much as 0.3 sec of the 1 sec lead might be explained by a thinner zone of superficial low-velocity material in the Sierra Nevada foothills than in the Coast Ranges. To account for the remaining 0.7 sec by variations in the crust requires rather large changes in boundaries between layers.

Two modifications of the two-layer-crust solution for the Coast Range profile that would satisfy observations along the San Francisco to Fallon profile are indicated in figure 5, which summarizes current seismic refraction information on the structure of the crust between San Francisco and Fallon. The structure shown by solid lines (thin crust extending southwestward approximately to the San Andreas fault) also accounts for the early arrival of P_n in the Great Valley suggested by observations that were corrected for delays in the sediments (fig. 4 and Eaton, 1963). Other models with somewhat greater depths to the mantle and a very shallow intermediate layer beneath the eastern part of the valley and the Sierra Nevada foothills satisfy the observations equally well.

Preliminary analysis of the San Luis Obispo to Owens Valley transverse profile shows that it has striking similarities to the northern transverse profile. Waves from the San Luis Obispo shotpoint are late, as expected, where they emerge through the thick sediments of the valley; but beyond 180 km they fall very close to the line established by arrivals beyond 160 km on the northern profile. P_g arrivals and apparent reflections from the mantle and an intermediate layer in the southern Coast Ranges northeast of San Luis Obispo suggest a two-layer crust that is about 3 km thicker than that between San Francisco and Camp Roberts. As on the northern profile, thinning of the crust beneath the valley or increase in the velocity of crustal rocks beneath the valley and foothills is required to explain the early P_n arrivals in the Sierra Nevada foothills.

Failure to record P_n east of the Sierra Nevada from shots at San Francisco left that profile incomplete. This defect was remedied by recordings at the University of California seismograph stations Berkeley, Santa Cruz, San Francisco, and Point Reyes near the San Francisco shotpoint (Mikumo, 1965), of waves from the SHOAL nuclear explosion, about 50 km southeast of Fallon, fig. 1. SHOAL P_n arrivals at these stations are plotted as solid circles on figure 4. The increase in path length required for propagation around the Sierra Nevada root shown in figure 5 would delay these arrivals about 0.7 sec. The corrected arrivals, which are plotted as crosses, fall on the extension of the San Francisco to Camp Roberts P_n line. Thus, unless the average velocity of P in the upper mantle between San Francisco and Fallon is significantly higher than 8.0 km/sec, which seems unlikely, the crust at SHOAL cannot be significantly thicker than that near Camp Roberts. This result tends to corroborate the thin crust (24 km) near the Shoal site deduced from the Fallon to Eureka profile (Eaton, 1963).

SUMMARY

A longitudinal seismic refraction profile in the Sierra Nevada shows that the high southern part of the range is underlain by a crust about 54 km thick. The crust thins beneath the northern end of the range and changes drastically in character at the Sierra Nevada-Cascade Range boundary. A similar profile through the central Coast Ranges indicates that the crust is only 22 to 25 km thick in that region. Early wave arrivals at the western foothills of the Sierra Nevada from explosions along the Pacific shore suggest that the crust is thinner, or is composed of rocks with higher seismic velocities, beneath the Great Valley than beneath the central Coast Ranges.

The velocity of P waves in the upper mantle appears to be somewhat less beneath the Sierra Nevada (7.9 km/sec) than beneath the central Coast Ranges (8.0 km/sec). The velocity of P waves in the upper part of the crystalline crust is the same beneath the Sierra Nevada and the central Coast Ranges west of the San Andreas fault (6.0 km/sec). Earthquake studies and meager refraction data suggest that the velocity of P waves in the upper crust beneath the central Coast Ranges east of the San Andreas fault is only 5.6 km/sec. This lower velocity may characterize a very thick accumulation of Franciscan rocks, which form the basement in this region.

Rocks with intermediate P wave velocities (6.4 to 6.9 km/sec) form a major part of the Sierra Nevada root, but they thin to 10 km or less toward both the northeast and southwest. They are probably a major constituent of the crust beneath the eastern Great Valley and the Sierra Nevada foothills.

Evidence from seismology on the detailed structure of the crust and upper mantle in northern California outside the regions described above is almost entirely lacking. Except for the tantalizing but limited evidence from the two transverse refraction profiles, the crust beneath the Great Valley remains essentially unexplored; and the Coast Ranges north of San Francisco, as well as the Klamath Mountains, have received even less attention. Extrapolations of results from one region into a neighboring one is of dubious value in light of the drastic changes encountered at the boundary between the Sierra Nevada and the provinces that surround it.

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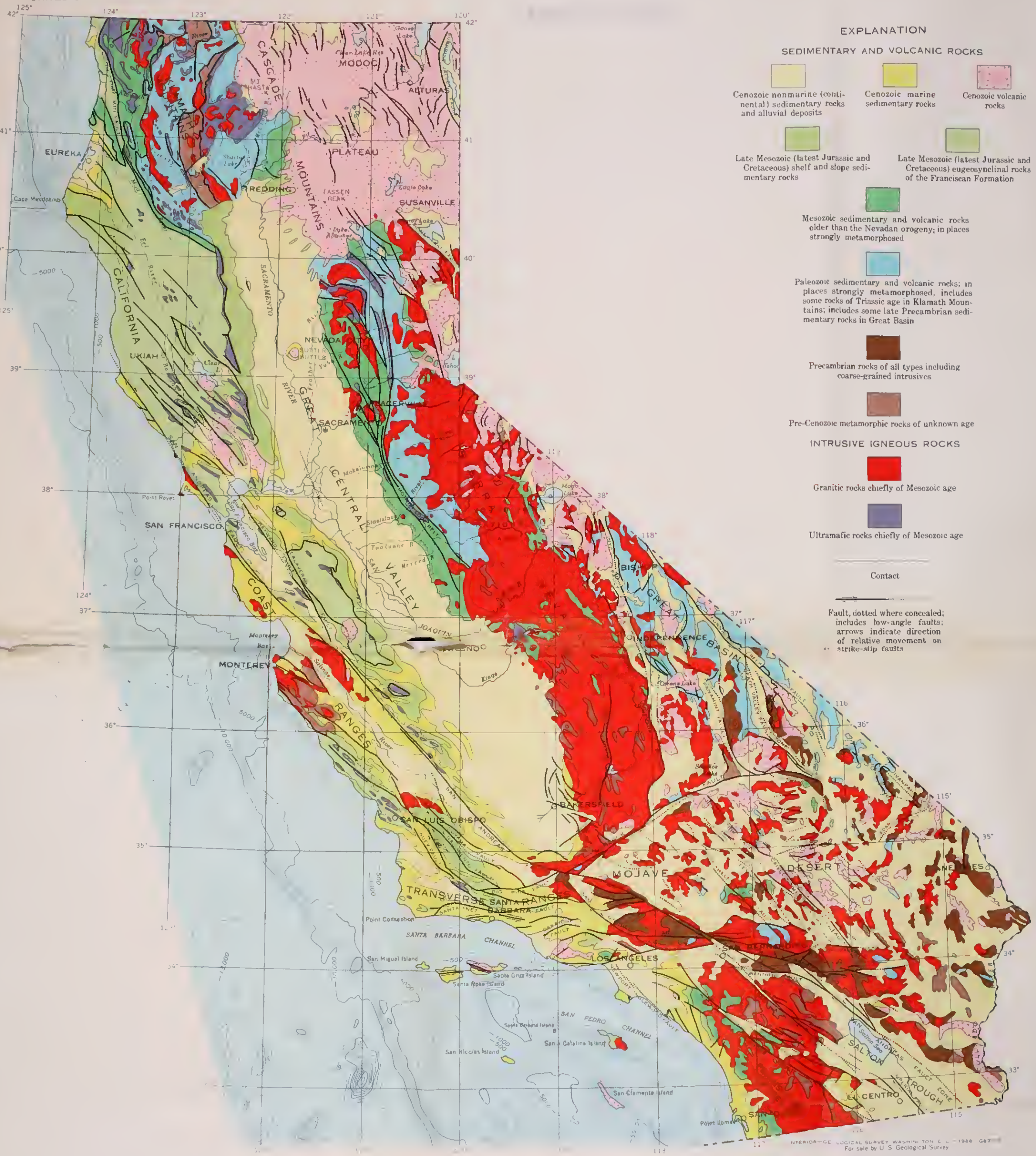
MISCELLANEOUS GEOLOGICAL INVESTIGATIONS
MAP I-512



GEOLOGIC MAP OF CALIFORNIA

COMPILED BY U.S. GEOLOGICAL SURVEY
AND CALIFORNIA DIVISION OF MINES AND GEOLOGY

SCALE 1:2 500 000

0 50 100 MILES

SUBMARINE CONTOUR INTERVALS 500 AND 1000 FEET DATUM IS SEA LEVEL

1966