Seismic structure of the Transverse Ranges, California

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ABSTRACT

Travel-time data obtained from both natural and artificial events occurring in southern California indicate a major, lateral crustal transition within the Transverse Range Province. The eastern crust is very similar to the adjacent Mojave region, where a crustal velocity of 6.2 km/sec is typically observed. The western ranges are dominated by an extensive 6.7 km/sec layer. P_n velocity beneath the western Mojave, Transverse Ranges, and northern Peninsular Ranges is 7.8 km/sec. The crustal thickness of these provinces is 30 to 35 km. The Transverse Ranges do not have a distinct crustal root. Unlike other provinces within southern California, the Transverse Ranges are underlain at a depth of 40 km by a refractor with a P-velocity of 8.3 km/sec. P-delays from a vertically incident, well-recorded teleseism suggest that this velocity anomaly extends to a depth of 100 km. These data indicate that this high-velocity, ridge-like structure is coincident with much of the areal extent of the geomorphic Transverse Ranges and is not offset by the San Andreas fault. Four hypotheses are advanced to explain the continuity of this feature across the plate boundary: (1) dynamic phase change; (2) a coincidental alignment of crust or mantle anomalies; (3) the lithosphere is restricted to the crust; (4) the plate boundary at depth is displaced from the San Andreas fault at the surface. Within the context of the last model, we suggest the plate boundary at depth is at the eastern end of the velocity anomaly, in the vicinity of the active Helendale-Lenwood-Camprock faults. The regionally observed 7.8 km/sec layer is suggested as a zone of decoupling necessary to accommodate the horizontal shear that must result from the divergence of the crust and upper mantle plate boundaries. The geomorphic Transverse Ranges are viewed as crustal buckling caused by the enhanced coupling between the crust and upper mantle which is suggested by the locally thin, 7.8 km/sec layer.

INTRODUCTION

The Transverse Ranges have long been an enigma in both the seismological and the geological understanding of the structure of southern California. These east-west–trending ranges clearly transverse the tectonic grain of the entire west Pacific coast from the Andes to the Aleutians, as well as that of southern California. Passing obliquely between the San Gabriel and San Bernardino Mountain Ranges, and yet not significantly offsetting the east-west topography, is the active San Andreas fault system with a horizontal dislocation of 250 km (Crowell, 1973).

The present style of faulting within this area is predominantly east-west, left-lateral, strike-slip, and thrust faulting, as typified by the San Fernando and Point Mugu earthquakes (Div. Geol. and Planetary Sciences, Caltech, 1971; Ellsworth and others, 1973). However, the structural geology of the region is extremely complex. The Transverse Ranges east of Cajon Pass are bound on the south by the Banning fault system (Allen, 1957), where crystalline rocks are thrust to the south over very young gravels. North of the Banning system, in the region of Lucerne Valley (Fig. 1), crystalline and metamorphic rocks are thrust to the north over young alluvium (Dibblee, 1964). Passing through and associated with these two systems is the San Andreas fault system. Thrust faulting in the western Transverse Ranges is similar to that in the eastern Ranges. Crystalline rocks again are thrust south over valley alluvium (Wentworth and others, 1971). However, the general style of tectonic deformation is somewhat different from that within the eastern portions of the range. Large, deep, actively subsiding basins such as the Ventura–Santa Barbara channel and the Los Angeles basin (Vedder and others, 1969; Yerkes and others, 1965) suggest a mode of vertical tectonics with great uplift and subsidence occurring within the same geologic province. In apparent conflict with this mode of vertical deformation within the Transverse Ranges are the available gravity data (McCulloh, 1960). A warp in the Moho corresponding to the 10-km subsidence within the Los Angeles basin and the great uplift within the San Gabriel Mountains is not indicated by the Bouguer gravity data. Instead, the data suggest a gentle thinning of the crust toward the offshore borderlands and an essentially uniform crustal thickness beneath the Transverse Ranges (S. Biehler, 1976, personal commun.).

Although many investigators have used extensive seismic data in the past to discern crustal structure within southern California (Gutenberg, 1944, 1951, 1952, 1955; Richter, 1950; Shor, 1955; Press, 1956, 1960; Roller and Healy, 1963; Kanamori and Hadley, 1975), the great number of seismic stations necessary to delineate structure within a single province are only recently available.

DATA

The southern California array operated by the California Institute of Technology since the early thirties has experienced of late a nearly exponential growth. The few critical stations initially installed were increased several fold in the early sixties. More recently, in cooperation with the U.S. Geological Survey, the number of stations has been increased to approximately 120. Most of these stations are within the Transverse Ranges, the eastern Mojave, and the Imperial Valley.

The present study was initially motivated by the need to interrelate many seismic-refraction profiles from different provinces of southern California. Recent testing at the Nevada Test Site (NTS) provided an ideal source for three distinct seismic-refraction profiles that extend to Santa Monica, Corona, and Eagle Mountain (Fig. 1). An upper crustal velocity of 6.2 km/sec was observed between the distances of 40 and 150 km. Consistent with the northeast-trending, reversed refraction profile of Roller and Healy (1963), an intermediate crustal

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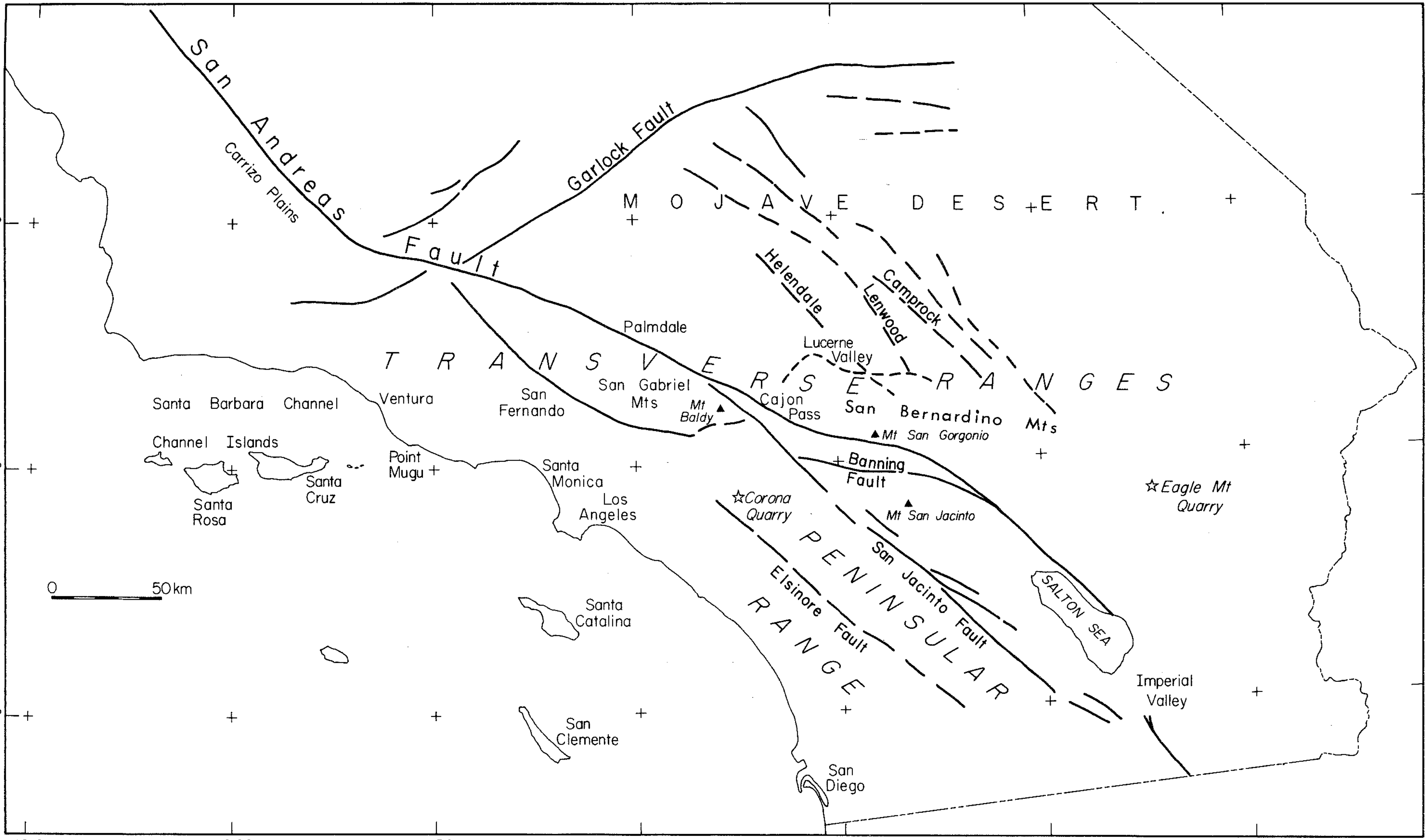

Figure 1. Index map for the southern California region.

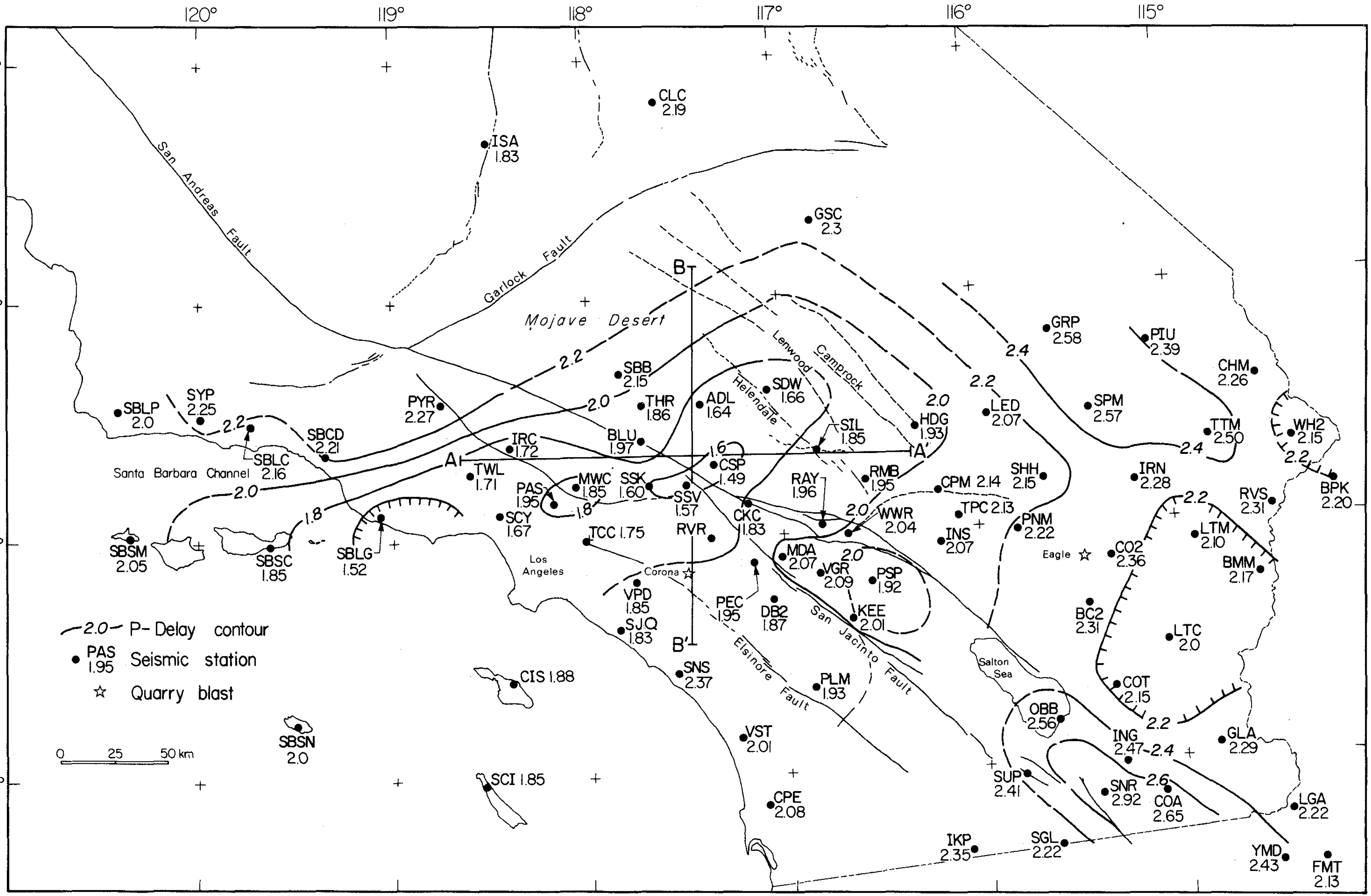

Figure 2. P-delay contours. The data for the reduction of the raw arrival time to the final station P-delay residual are listed in Table 1. As compared to the Jeffreys-Bullen travel-time tables, all stations are late by 1.5 to 2.9 sec. Station corrections for the sediment delay for the Santa Barbara and for the Imperial Valley stations have not been included.

branch with a velocity of 6.7 to 7.0 km/sec was not observed as a first arrival. P_n velocities appear to increase from 7.8 km/sec for the western profile to about 8.0 km/sec for the eastern. The crustal thickness of about 32 km derived from these profiles is similar to that derived in earlier work (Kanamori and Hadley, 1975). A Corona quarry blast in January 1975 was used to reverse the center profile. This reversed profile indicates that the Moho is essentially flat. Furthermore, even though the profile passed obliquely through the Transverse Ranges, the travel-time curves are identical. If the Ranges have a significant crustal root structure, then some shift in the reversed travel-time curve should have been observed. A minimum thickness of the 7.8 km/sec layer is estimated to be about 20 km because of the large distance over which this branch is observed.

The same Corona shot was used to establish travel-time curves northwest to the Carrizo Plain, east across the northern end of the Salton Trough and southeast down the Peninsular Ranges (Fig. 1). P_n data obtained from these profiles are consistent with the earlier interpretation of Kanamori and Hadley (1975) that southern California is underlain by a uniform 7.8 km/sec layer. Crustal velocities and crossover distances are 6.1 km/sec and 155 km for the Carrizo Plain and 6.1 km/sec and 130 km for the northern end of the Salton Trough. Unlike those obtained for other profiles to date, a crustal velocity of 6.6 km/sec was obtained for the profile down the Peninsular Ranges. The crossover to P_n is 165 km. Finally, a small blast located just south of the Salton Sea was detonated to calibrate the Imperial Valley array. P_n velocity along a composite profile striking approximately N30°W was 7.8 km/sec, crustal velocity was 6.3 km/sec, and the crossover distance was 100 km. The details of these results will be published elsewhere (in prep.).

The uniform 7.8 km/sec P_n velocity observed over much of southern California is similar to that typically observed for the tectonically active Basin and Range Province (Pakiser, 1963; Pakiser and Hill, 1963; Keller and others, 1975). The inversion of both Rayleigh-wave (Biswas and Knopoff, 1974) and body-wave (Archambeau and others, 1969) data for the Basin and Range indicate that the relatively low P_n velocity extends to depths of at least 150 km. This zone is interpreted as the low velocity zone (LVZ). The base of the crust is viewed as the top of the LVZ. The inferred minimum thickness of 20 km for the uppermost layer of the mantle and the low velocity of 7.8 km/sec suggest, by analogy, that the crust of southern California is riding directly upon the LVZ.

TABLE 1. OBSERVED PKP ARRIVAL TIMES, EPICENTRAL DISTANCES, JEFFRYS-BULLEN CALCULATED TRAVEL TIMES, STATION CORRECTIONS FOR ELEVATION AND SEDIMENTS, AND FINAL P-DELAY TIMES

Event Date: January 23, 1976; origin time: 5:45:30.7 GMT; location: 119.9°E, 7.5°S; depth: 620 km

Station	Δ(DEG)	T(6:03)	T_{JB}	T_{HTCR}	T_{SED}	T_{FINAL}
CKC	121.296	17.02	15.08	0.11		1.83
MDA	121.488	17.67	15.45	0.17		2.07
RAY	121.613	18.12	15.69	0.47		1.96
WWR	121.748	18.12	15.94	0.14		2.04
VGR	121.660	18.17	15.78	0.3		2.09
DB2	121.478	17.42	15.43	0.12		1.87
PSP	121.881	18.16	16.20	0.04		1.92
KEE	121.832	18.39	16.11	0.27		2.01
CHM	123.284	21.32	18.87	0.19		2.26
TTM	123.136	21.30	18.58	0.22		2.50
WH2	123.476	21.66	19.26	0.25		2.15
BPK	123.685	21.98	19.68	0.10		2.20
RVS	123.461	21.69	19.24	0.14		2.31
LTM	123.169	20.91	18.66	0.15		2.10
BMM	123.477	21.56	19.28	0.11		2.17
LGA	123.798	22.17	19.94	0.01		2.22
FTM	123.978	22.49	20.31	0.05		2.13
YMD	123.805	22.41	19.96	0.02		2.43
COT	122.966	20.41	18.26	0.0		2.15
OBB	122.767	20.45	17.89	0.0		2.56
SUP	122.664	20.11	17.70	0.0		2.41
SGL	122.814	20.21	17.99	0.00		2.22
ING	123.076	20.96	18.49	0.0		2.47
SNR	123.002	21.27	18.35	0.0		2.92
COA	123.258	21.51	18.86	0.0		2.65
THR	120.764	16.11	14.04	0.21		1.86
BLU	120.792	16.44	14.09	0.38		1.97
ADL	121.006	16.73	14.51	0.18	0.40	1.64
SDW	121.269	16.91	15.01	0.24		1.66
SIL	121.529	17.72	15.52	0.35		1.85
SSK	120.863	16.19	14.24	0.35		1.60
SSV	121.024	16.45	14.56	0.32		1.57
SYP	118.947	13.01	10.50	0.26		2.25
ISA	119.912	14.32	12.32	0.17		1.83
CLC	120.570	15.96	13.62	0.15		2.19
GSC	121.320	17.60	15.10	0.20		2.30
SBB	120.648	16.12	13.80	0.17		2.15
CSP	121.114	16.47	14.73	0.25		1.49

In attempting to relate differing crustal thicknesses and velocity profiles derived from these data, an impulsive, distant teleseism was studied. The initial hope was to map the changing crustal sections by observing subtle changes in the observed P-delay times throughout southern California. The teleseism chosen for this purpose was a deep earthquake occurring within the Java Trench. The U.S. Geological Survey hypocentral data are: 7.5°S, 119.9°E; depth = 620 km; origin time = 5:45:30.7 GMT; January 23, 1976; m_b = 6.4. The source station separation is approximately 120°. At this distance, the direct P-wave is not well observed, owing to the shadow zone of the Earth's core. However, a particularly strong PKP phase, emerging through the crust at an incidence angle of 4°, was well recorded. The curvature of the wavefront for this phase is extremely slight, $dt/d\Delta$ is constant, and the relative theoretical arrival times at all southern California stations are easily computed. All stations recorded on 16-mm film that is viewed at a scale of 1 sec/cm. All P-arrival times were read twice, and the average time was used. Fifty msec was typical of the difference between the two readings.

The theoretical travel times from the source to the stations were calculated by interpolating the Jeffreys-Bullen (J-B) tables over the appropriate depth and distance range. A simple correction for the Earth's ellipticity was calculated; although over the area of southern California, the relative variation of this correction is negligible. Finally, a height correction was introduced to reduce all stations to a common elevation, sea level. This correction was simply the height of the station divided by the average velocity of the uppermost layer of the crust,

TABLE 1. *(Continued)*

Station	Δ(DEG)	T(6:03)	T_{JB}	T_{HTCR}	T_{SED}	T_{FINAL}
RVR	121.116	16.61	14.84	0.05		1.72
PEC	121.363	17.29	15.22	0.12		1.95
TPC	122.212	19.10	16.82	0.15		2.13
PLM	121.726	18.19	15.92	0.34		1.93
VST	121.466	17.46	15.43	0.02		2.01
CPE	121.633	17.88	15.76	0.04		2.08
SCI	120.423	15.28	13.39	0.04		1.85
IKP	122.499	19.95	17.41	0.19		2.35
GLA	123.455	21.67	19.25	0.13		2.29
RMR	121.762	18.26	15.97	0.34		1.95
HDG	121.929	18.48	16.28	0.27		1.93
CPM	122.081	18.90	16.57	0.19		2.14
INS	122.135	19.09	16.68	0.34		2.07
PNM	122.442	19.71	17.26	0.23		2.22
LED	122.215	19.06	16.82	0.17		2.07
SHH	122.509	19.75	17.38	0.22		2.15
GRP	122.397	19.98	17.15	0.25		2.58
SPM	122.641	20.38	17.63	0.18		2.57
PIU	122.820	20.59	17.96	0.24		2.39
IRN	122.894	20.60	18.12	0.20		2.28
CO2	122.842	20.44	18.03	0.05		2.36
BC2	122.794	20.49	17.94	0.24		2.31
LTC	123.151	20.73	18.64	0.09		2.00
SBLP	118.598	11.84	9.81	0.03		2.00
SBSM	118.739	12.18	10.10	0.03		2.05
SBLC	119.167	13.32	10.92	0.24		2.16
SBSC	119.332	13.19	11.25	0.09		1.85
SBSN	119.581	13.79	11.74	0.05		2.0
SBCD	119.492	13.80	11.55	0.04		2.21
SBLG	119.770	13.68	12.09	0.08		1.52
SNS	121.148	17.22	14.81	0.04		2.37
SJQ	120.866	16.52	14.26	0.03	0.40	1.83
CIS	120.454	15.42	13.44	0.10		1.88
VPD	120.892	16.60	14.31	0.04	0.40	1.85
TCC	120.649	15.97	13.82	0.06	0.34	1.75
MWC	120.563	15.84	13.64	0.35		1.85
PAS	120.487	15.50	13.49	0.06		1.95
SCY	120.267	14.79	13.06	0.06		1.67
TWL	120.117	14.93	12.74	0.08	0.40	1.71
IRC	120.251	14.85	13.01	0.12		1.72
PYR	119.937	14.90	12.38	0.25		2.27

5.0 km/sec. In the worst case, for the highest station, if the velocity is locally as low as 3 km/sec, then this correction is too small by 0.3 sec. More typically, the height correction is probably in error by less than about 0.1 sec. Hence the average error of the relative arrival times across southern California, incoporating both uncertainties in the height correction and reading errors, is of the order of 0.1 sec for low elevation stations and 0.2 sec for the few stations with large height corrections. Table 1 lists the stations, distances, arrival times, calculated J-B times, height corrections and final P-delay residuals. Travel-time corrections for stations in the Los Angeles basin situated on deep accumulations of sediments were made on the basis of past refraction data. Similar corrections for the suite of stations north of the Santa Barbara channel have not been included, as our present seismic data do not provide a good estimate of these delays. However, from geologic considerations, these stations are probably delayed by 0.1 to 0.3 sec.

The consistency of the observed residuals (Fig. 2) attests to the quality of both the timing system and the impulsive, unambiguous arrival of this phase. The most striking feature seen in the pattern of P-delays is the very early arrivals recorded throughout most of the Transverse Ranges. The three earliest arrivals, stations SSK, SSV, and CSP, straddle the surface expression of the San Andreas fault and arrive 0.7 sec earlier than GSC, a northern Mojave station. This band of advanced arrivals includes the Los Angeles basin and part of the Santa Barbara channel. Corrections for sediment delay north of the Santa Barbara channel would probably broaden the band of early arrivals to include stations SBCD, SBLC, SYP, and SBLP. A second feature is the abrupt transition of delays in the eastern Mojave. Delays range from 2.07 to 2.15 sec at LED and SHH to 2.57 and 2.58 sec for SPM and GRP, respectively. Another region of large delays is within the Imperial Valley. The relative P-arrivals for stations within the valley, not corrected for the low-velocity sediments, are 2.5 to 2.9 sec; for stations on bedrock and on the periphery of the valley, they are about 2.3 sec. The variation in P-delay residuals within southern California is greater than 1 sec. However, the delay from varying crustal structures was expected to be about 0.2 sec. As a large contribution to this variation is associated with the Transverse Ranges, additional seismic-refraction data were collected for paths solely within this region. Two earthquakes with local magnitude of about 4.5 have occurred at opposite ends of the Transverse Ranges since the most recent augmentation of the array (Fig. 2, profile A-A′). The large number of stations that recorded both events provide excellent hypocenter locations and define several branches of the travel-time curves (Fig. 3). Epicentral distances to the stations used to locate both events were less than 120 km. More than 20 stations were used for both locations. The crustal-velocity model used for the hypocentral determinations was quite similar to that shown at point A′ in Figure 3. The epicentral uncertainty for both events is estimated to be less than 1 km. However, the determined depth for these earthquakes is dependent on the crustal-velocity model. The model we used to locate the eastern event (constructed in previous studies; Kanamori and Hadley, 1975), is probably quite realistic, and the hypocentral depth is not likely to be seriously in error. The depth of the western event is somewhat less certain. The calculated depth for this event is 12 km. However, uncertainty in the model suggests 3-km error bars on the location. This results in an uncertainty in the crustal-model determination, because the source depth, crustal thickness, and crossover distance from P_g to P_n all trade off. Given a crossover distance, the derived crustal thickness and the depth of the event are directly related. In the model presented, the event was located as deep as possible, and the resulting crustal thickness represents a maximum. The dissimilarity between the two curves is the result both of differing source depths and of dip on the top of the 6.7 km/sec layer. The solid lines are calculated from the velocity model shown. The thickness and velocity of the topmost layer, although not constrained by the present data, are very typical of many regions in southern California (Kanamori and Hadley, 1975). The eastern end of this

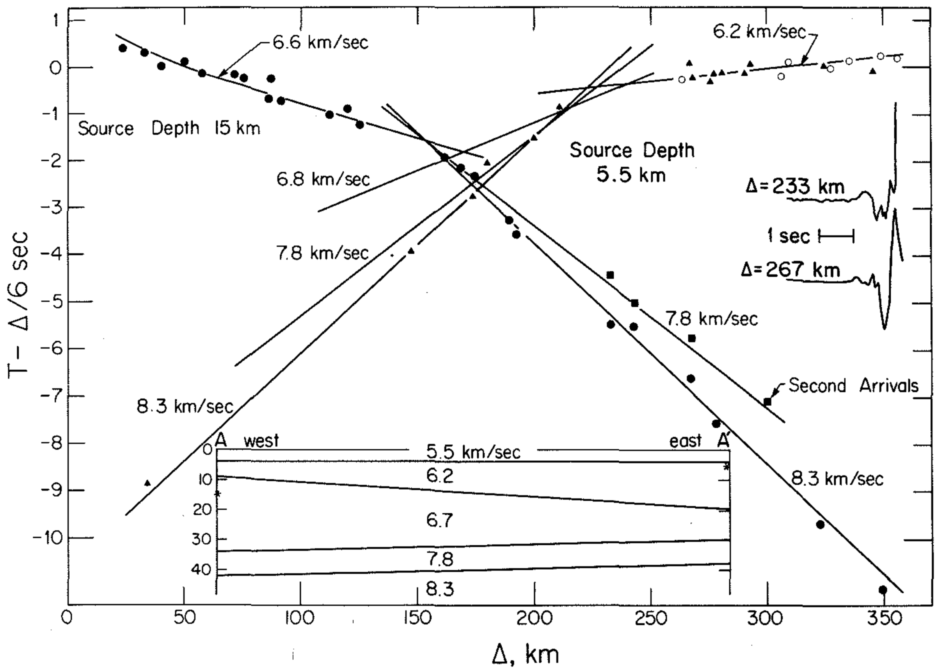

Figure 3. **Travel-time data from two magnitude 4.5 earthquakes which occurred and were recorded within the Transverse Range Province. Open-circle data points are from stations east of the profile. Solid lines are calculated from the inset model. Waveforms show the relative amplitudes of the initial and secondary arrivals. The location of this profile is shown on Figure 2.**

structure is similar to that derived by Kanamori and Hadley (1975). The western section has a very thick 6.7 km/sec layer. The shallow depth to this interface and the thickness of this unit are consistent with the crustal model derived by Stierman and Ellsworth (1976) for the Point Mugu region. Although a large secondary arrival with an apparent velocity of 7.8 km/sec was observed in the distance range 230 to 300 km for the western, deeper event, the apparent velocity of the first arrival from both earthquakes was 8.3 to 8.4 km/sec. The thickness of the 7.8 km/sec layer is about 8 km, and the top of the 8.3 km/sec layer is approximately 40 km.

INTERPRETATIONS

The observed P-delay times recorded both within and around the Transverse Ranges cannot be explained as differences arising from dissimilar crustal structure. The expected time difference for a ray passing vertically through the eastern or western ranges, or even through the Mojave crust, is about 0.1 sec. The remaining travel-time advance within this province of about 0.5 sec must then result from an upper mantle, high-velocity structure. The observation of a shallow, high-velocity, 8.3 km/sec structure beneath the Transverse Ranges is consistent with the P-wave advance. The observed velocity difference of 0.5 km/sec between the 8.3 km/sec and the normal P_n velocity in southern California (7.8 km/sec) suggests that the high-velocity structure beneath the region of greatest teleseismic P-wave advance must exist over a vertical distance of about 60 km. Figure 4 shows a north-south profile through the Cajon Pass area (profile B-B′ in Fig. 2) based on this interpretation. The lateral extent of this high-velocity body must coincide with the region of P-wave travel-time advance, the vertical thickness or velocity contrast decreasing to zero beneath regions reporting a more intermediate P-delay of about 2.2 sec. The cross-sectional shape of the body is not well constrained by the present data. An equally acceptable model is a lens-shaped body. The refraction data from both NTS and Corona suggest the 8.3 km/sec body is not at a shallow depth beneath either the southern Mojave or the northern Peninsular Range. This observation precludes a distribution of higher velocity material in a form with the upper and lower boundaries flat and concave upward respectively. Extending from 40 to 100 km in depth at the point of greatest thickness, this body forms a high-velocity ridge beneath much of the area of the Transverse Ranges.

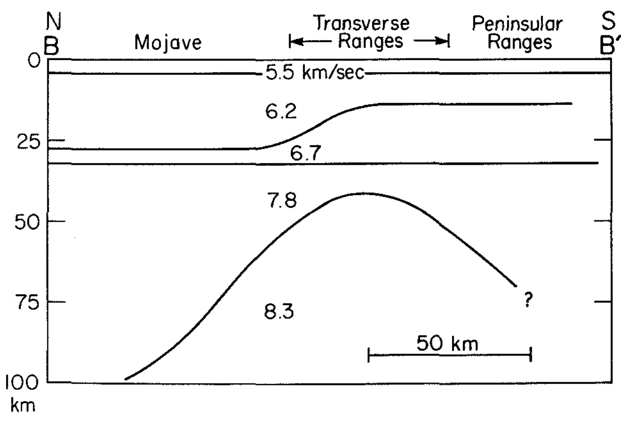

Figure 4. North-south cross section through the Transverse Ranges in the vicinity of Cajon Pass. Seismic sections for the crust from the Mojave, Transverse Ranges, and Peninsular Ranges have been smoothly connected in constructing this model. The thickness of the high-velocity upper mantle anomaly is constrained by the vertical travel-time of a well-recorded PKP phase.

An alternative contouring of the vertically incident P-delay data in the region bounded by stations SBB, PYR, ISA, and GSC would extend the high-velocity upper-mantle anomaly as far north as ISA. However, the teleseismic data of Raikes (1975) indicate that arrivals from the northeast at PYR, northwest at SBB, and southeast at ISA are all delayed relative to other azimuths. This suggests that the upper mantle in the PYR-SBB-ISA region is slow compared with the mantle southeast of SBB and PYR, as indicated by relatively early arrivals. In addition, stations TPC and PEC show early arrivals for northwest azimuths. The lateral extent and the travel-time advance of this anomaly are consistent with the azimuthal P-delay data presented by Raikes (1975).

The surface expression of the San Andreas fault traverses this high-velocity ridge at the point of greatest inferred thickness (Fig. 2). If the San Andreas fault represents the plate boundary between the Pacific and the North American plates, then one would normally expect to find this feature truncated by the fault. One of several hypotheses may explain the observed lateral continuity. (1) The anomaly is the result of a phase change that continuously adjusts to maintain this feature across the plate boundary. (2) The apparent continuity is simply the result of a coincidental alignment of crust or upper mantle anomalies. (3) The

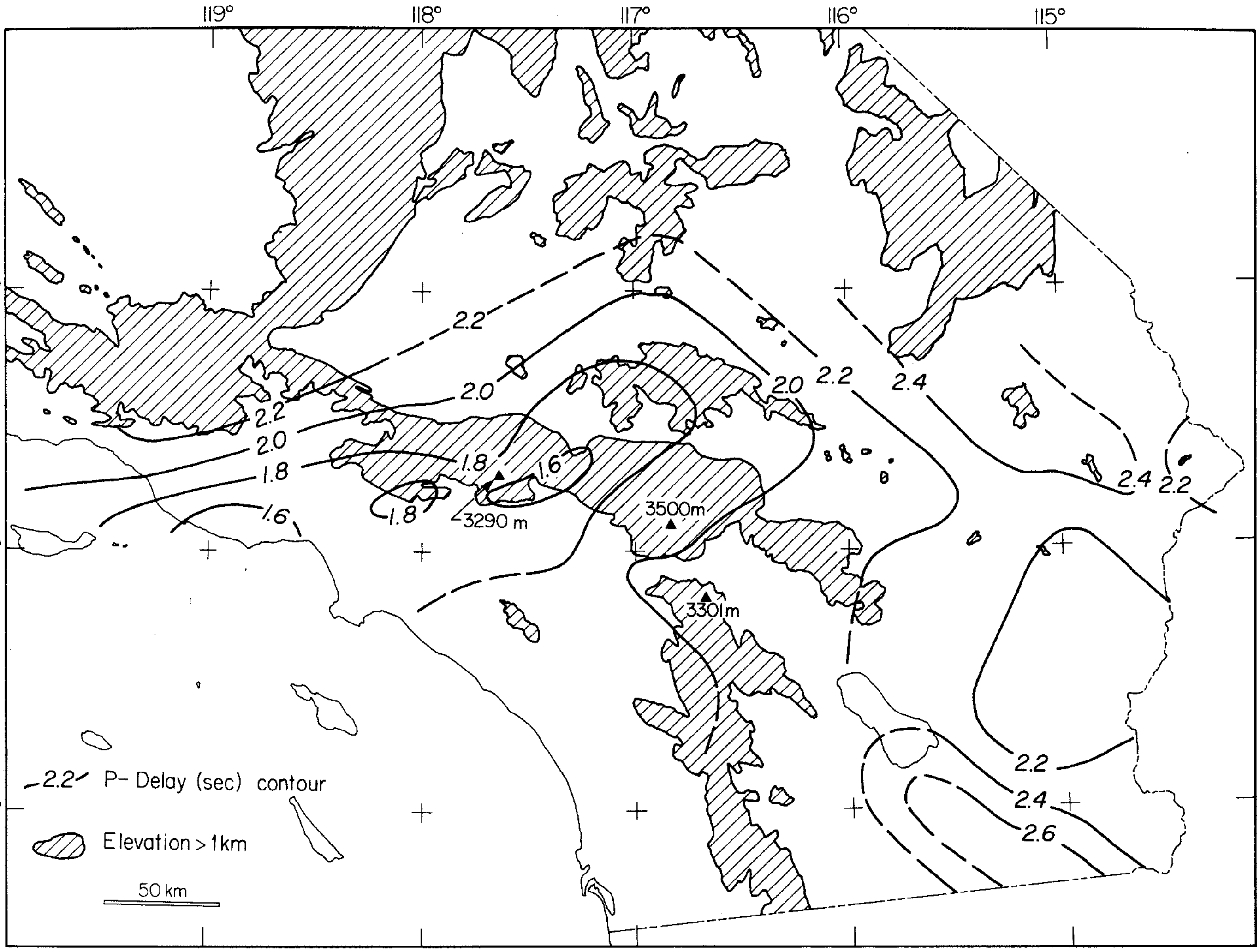

Figure 5. Relief map and P-delay contours. Note that much of the relief associated with the Transverse Ranges, including a broad uplift within the south-central Mojave, is underlain by the mantle anomaly. However, the anomaly is present beneath the relatively flat Los Angeles basin and absent beneath adjacent Mount San Jacinto. This suggests that the crust (via a crustal-loading, pressure-induced phase change) is not controlling the anomaly.

plate or lithosphere is restricted to the crust. (4) The plate boundary extends to the upper mantle but is displaced at depth from the San Andreas fault at the surface.

Either temperature or pressure could drive the phase change (Ringwood, 1975). However, changing phases with variable temperature appears difficult. An ad hoc mechanism that could alternately cool and reheat large sections of the upper mantle in the required geometry would be necessary. In addition, the time constant for changing temperatures by diffusion in a body with the observed physical dimensions is several million years; this time constant would be too long to maintain the continuity of the upper mantle feature which is continuously offset by the fault. Although temperature perturbations seem inadequate, stress deviations within the upper mantle could serve to drive the phase change. The lack of earthquakes at depths greater than 20 km implies that if such a stress deviation exists, its magnitude is not large. These arguments suggest that maintaining the continuity of the anomaly, a phenomenon that must continually readjust as the plates move, via a dynamic phase change driven either by the temperature or pressure fields, is unlikely.

Without additional data on the frequency of occurrence of anomalous upper mantle structures elsewhere along the present plate boundary, the probability of a coincidental alignment is difficult to evaluate. The possibility that the anomalous body results from a phase change resulting from a relatively long-lived lithostatic load is worthy of mention. In this model, the mantle anomaly is controlled simply by the crust. Indeed, much of the anomaly lies beneath a region of substantial relief (Fig. 5). Perhaps the continuity of the anomaly across the San Andreas fault is as coincidental as the continuity of the geomorphic Transverse Ranges. However, the absence of an anomaly beneath Mount San Jacinto (the second highest peak in southern California and adjacent to the anomalous area) and the presence of the anomaly beneath the low-lying Los Angeles basin suggest that the lithostatic load is not controlling the anomaly.

The third hypothesis shares an important implication with the fourth. The regionally observed 7.8 km/sec P_n layer, by analogy with the adjacent Basin and Range Province as discussed above, suggests that the low-velocity zone extends up to the base of the crust. If the low-velocity zone serves as a complete decoupling layer, and if plate motion is confined to the crust, then most of southern California is underlain by a major zone of horizontal shear at a depth of approximately 30 to 35 km. The San Andreas fault need not extend beyond the Moho; most of the southern California crust simply moves over a relatively stationary upper mantle. This consequence excludes the possibility that the local plate driving force arises from a viscous drag on the base of the crust from sympathetic motion in the upper mantle. The tractions necessary to move the plate must be applied to the lithosphere in regions exterior to southern California. If this is the present mechanism, then the concentration of strain release along the San Andreas fault, in contrast with the many other faults that could participate, is quite remarkable. One might more reasonably expect the horizontal shear between the North American and the Pacific plates to be manifested in a number of faults within the thin lithosphere.

If the upper mantle participates in the plate motion, then the question arises —where is the plate boundary at depth? The continuity of the high-velocity ridge suggests that the boundary, at depth, is removed either east or west from the San Andreas by at least the amount represented by the lateral extent of the early P-wave arrivals. As in the third hypothesis, this requires a zone of horizontal shear or decoupling within the upper mantle between the surface trace of the plate boundary and that at depth. Unlike the third hypothesis, the decoupling is largely limited to the region between the boundaries at the surface and at depth. Outside of this region, in the vicinity of the Transverse Ranges, minor decoupling must occur to accommodate the divergence of the plate boundary at depth from the San Andreas fault at the surface. The observed low P_n velocity of 7.8 km/sec, substantially lower than P_n velocities of 8.1 to 8.3 km/sec obtained in tectonically stable regions, suggests shear may be occurring within this layer. The structural rifting of the Gulf of California (Henyey and Bischoff, 1973) and the Salton Trough (Biehler and others, 1964) requires a plate boundary that extends at least as far north as the Salton Sea. Rather than inferring that the plate boundary at depth strikes west from the Imperial Valley to the Channel Islands (a trend contrary to all of the recent tectonics), we propose that the boundary extends northwest along the strike of the structural elements of the trough and passes through the eastern end of the anomaly in the vicinity of the active Helendale-Lenwood-Camprock faults (Fig. 2). Figure 6 is a schematic representation of the proposed model. The plate boundary beneath the Mojave desert is probably not a simple planar structure, but rather a broad zone of shear.

The present data do not answer the question of the northern continuation of this proposed plate boundary. The great crustal extension seen in the Basin and Range Province must affect the trend of any plate boundary within southern California. However, the nature of this interaction, particularly at subcrustal depths, is not yet understood.

The observed thinning of this proposed decoupling layer beneath most of the Transverse Ranges suggests that this region may result simply from greater local coupling between the mantle and the crust than occurs elsewhere. The differential motions between the crust and the upper mantle, which must occur under southern California if the crust and mantle plate boundaries diverge, would have a profound effect upon the crust in this region of greater drag. Within model 4, the morphologic Transverse Ranges (extending across the San Andreas fault) should appear as a constant phenomenon associated with the high-velocity upper-mantle structure and the thin decoupling layer. As the Pacific plate moves northwest, the Ranges should likewise migrate. Fault systems that penetrate only the crust should have only minor effects on the geomorphology of the range. The Transverse Ranges are viewed as a ripple in the crust that reflects a major upper-mantle structure. Great uplift (such as the San Gabriel Mountains), relatively minor, recent uplift (such as that at Palmdale; Castle and others, 1975), and major subsidence (Yeats, 1976) within the Transverse Ranges can all be reconciled with a mode of crustal buckling associated with a subcrustal viscous drag.

If the possibility that the high-velocity structure results from a dynamic, recent process is ignored, then the maximum emplacement age of the anomaly should coincide with the cessation of subduction, 15 to 20 m.y. ago (Atwater, 1970). If this structure were the remnant of subduction, one of two possible mechanisms might control its history: (1) a long-lived thermal perturbation or (2) an inherent change in petrology. As a rough estimate of the thermal time constants of this body, consider a slab 100 km thick with initial temperature T_0 welded between two half-spaces. After 10 m.y., the temperature 10 km from the boundaries has changed by only 25%. Although the anomaly is clearly not a slab, the time necessary to alter significantly by diffusion the thermal field of a body with dimensions approaching 100 km suggests that an emplacement temperature field is relatively stable over the time bracket of millions of years. This suggests that if an anomalous temperature field were present in the subducted slab, a major portion of that field would still remain today. If the temperature field controls a phase change, such as from garnet-granulite to eclogite, or from a partial melt to the solidus, then the heat of the phase change further stabilizes the field. Although a major, localized change in upper-mantle petrology beneath southern California seems unlikely, the phase transition boundary for garnet-

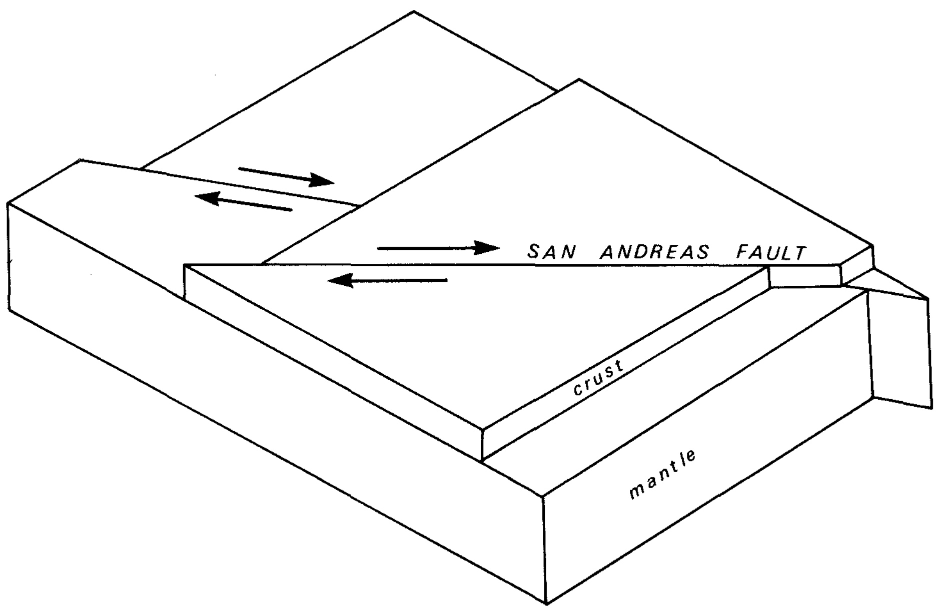

Figure 6. Block diagram of the proposed divergence of the crust and mantle plate boundaries: hypothesis 4. The crust and mantle boundaries are assumed to be coincident within the Salton Trough. The location of the mantle boundary in the Mojave region is inferred from the P-delay data to be in the vicinity of the active Helendale-Lenwood-Camprock faults.

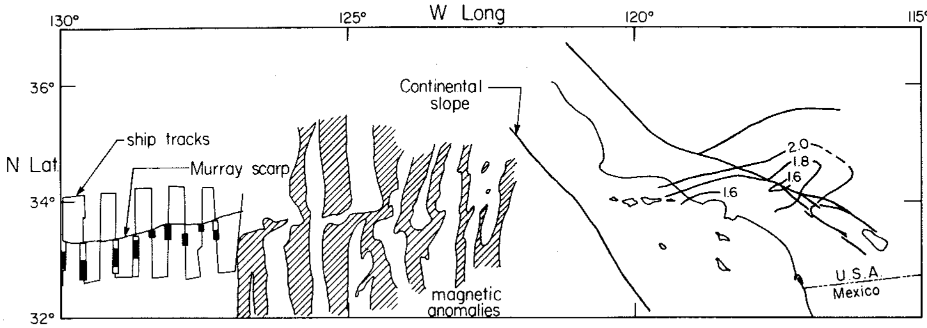

Figure 7. Murray Fracture Zone, magnetic anomalies, and the southern California P-delay contours. Solid and open rectangles along the ship tracks are normal and reversed magnetized bodies inferred to have been incorporated into the crust during a history of emplacement (Malahoff and Woollard, 1968). Cross-hatched anomalies are magnetized in the normal direction. There is a 160-km offset at 125°W long (Mason, 1958).

granulite to eclogite can be significantly modified by small variations in chemistry (Ringwood, 1975). These considerations suggest that the anomalous mantle structure could result from either an initial thermal field or minor petrologic differences. The regionally averaged Bouguer gravity data might allow a distinction between these two phase-change models. Less than 1% melt can reduce the P-velocity from 8.3 to 7.8 km/sec (Anderson and Spetzler, 1970). The associated density change is less than 0.03 g/cm^3. The minimum density change for an equivalent velocity reduction via an eclogite to garnet-granulite transition is 0.15 g/cm^3 (Ringwood, 1975). For the assumption of a buried horizontal cylinder with the appropriate physical dimensions, the expected maximum Bouguer anomaly is 30 and 150 mgal, respectively. The available data (S. Biehler, 1976, personal commun.) suggest that the partial-melt model is preferable.

Assuming that the anomaly is a relic of subduction, either of these two mechanisms implies an inherent, localized variation in the sea floor most recently subducted. The most obvious feature on strike with and approximately the same width as the anomaly is the Murray Fracture Zone (Fig. 7). South of the Murray scarp (the solid east-trending line in the left third of Fig. 7), there is a broad zone of ridges, troughs, and seamounts (Malahoff and Woollard, 1968). The juxtaposed normal and reverse magnetized bodies south of the scarp were interpreted by Malahoff and Woollard as resulting from the incorporation of new crustal material at different times. The magnetic offset around 125°W long indicates that about 25 m.y. ago the ridge system of the Pacific-Farallon plates was offset 160 km (Mason, 1958; Atwater and Menard, 1970; Atwater, 1970). The rapid decay of the peak-to-peak magnetic field intensity from 600 to 200 gamma as the continental slope is approached, compared with a background noise of 50 gamma for any given measurement, hinders attempts to estimate the magnitude of the more recent offset. McKenzie and Parker (1967) have suggested that the Pacific and American plates shared a common pole of rotation during subduction of the Farallon plate. Although the convergence rate of the Farallon-American plates has been estimated to have been 7 cm/yr (Atwater, 1970); the relative motion of the Pacific-American plates, for the same time period, is not constrained. The region of the present anomalous upper-mantle structure must have been the site of at least a trench-transform junction, or more probably a trench-transform-transform triple junction, depending upon the magnitude of the relative motion between the Pacific and American plates. This junction was active for about 2 m.y. (~ 160 km/7 cm/yr). However, because of the possibility of major, undetected right-lateral shear between the observable eastern extent of the Murray Fracture Zone and the anomalous, high-velocity structure, a clear connection between these two features cannot be made. The 60- to 100-km width of the subducted fracture zone (Menard, 1955), intruded by crustal material after the plate left the ridge, and the once-active triple junction suggest that the present anomaly is at least potentially the site of some unusual past tectonics.

We suggest that the upper-mantle high-velocity anomaly is a remnant of subduction, possibly associated with the Murray Fracture Zone. The regionally observed 7.8 km/sec layer is suggested as the zone of decoupling that is necessary to accommodate the horizontal shear that results from the divergence of the crust and upper-mantle plate boundaries. The morphology and location of the Transverse Ranges appear to be controlled by the enhanced coupling between the crust and upper mantle, as indicated by the locally thin 7.8 km/sec layer.

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