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Teleseismic tomography of the Loma Prieta region, central California

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TELESEISMIC TOMOGRAPHY OF THE
LOMA PRIETA REGION,
CENTRAL CALIFORNIA

A Thesis

Presented to

The Faculty of the Department of Geology
San Jose State University

In Partial Fulfillment
of the Requirements for the Degree
Master of Science

By

Yoko Takauchi

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ABSTRACT

TELESEISMIC TOMOGRAPHY OF THE LOMA PRIETA REGION, CENTRAL CALIFORNIA

by Yoko Takauchi

Three-dimensional velocity models reaching 116 km depth in the Loma Prieta region, central California, have been derived using teleseismic travel-time data and an advanced "ACH" inversion method. Upper crustal structure is in agreement with previous local-earthquake tomography results. Many velocity gradients fall beneath the San Andreas fault at all depths, suggesting that the San Andreas was a deep penetrating shear locus at least at one time. Near the Moho, a large low-velocity feature is observed beneath the Loma Prieta highlands. This low may be caused by a crustal root, approximately 2-5 km thick. Two upper mantle low-velocity features displaced by ~45 km along the San Andreas fault in a right-lateral sense also are found. This offset could have been caused either by narrow shear locus jumping landward with time or persistent broad (~85 km) shear zone.

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INTRODUCTION

The San Andreas fault is the principal manifestation of the northwest-striking boundary between the Pacific plate and the North American plate in California. In the San Francisco Bay Area, the fault splays and forms a complex fault system with many subsidiary faults. This fault system has active seismicity, with the Loma Prieta earthquake of 1989 causing severe damage ($M_S = 7.1$) and loss of human lives. The structure of the lithosphere in the Loma Prieta region is complicated, and many models have been proposed. However, little is known about the structure of the lower crust and upper mantle from direct observation. In this paper, I present teleseismic tomography models based on *P* wave travel time. In this study I use teleseismic tomography to shed light mainly on this deeper lithospheric structure. These results affect seismic hazard analysis by improving understanding of the Cenozoic tectonic evolution of central California.

GEOLOGIC SETTING

Geological Units

Figure 1 is a map of the study area showing the locations of the major faults. Figure 2 shows the distribution of geological units in the region of the 1989 Loma Prieta earthquake (Wentworth, 1993). Rocks on the southwest side of the San Andreas fault belong to the Salinian block, which include an assemblage of mainly Late Cretaceous granitic or metamorphic rocks ("Salinian basement") overlain by Tertiary marine sedimentary rocks (Irwin, 1990). Because of the similarity of the granitic rocks and their metamorphic roof pendants to rocks found in the Sierra Nevada, this block sometimes is interpreted as part of the Sierra Nevada batholith (e.g., Page, 1981). Paleomagnetic studies suggest that this block may have been transported from the southeast, and possibly from Mexico or Central America (Champion et al., 1984).

Another major assemblage, the Franciscan, occupies the northeast side of the San Andreas fault. This diverse assemblage is mainly of Late Cretaceous age and consists largely of graywacke and shale with radiolarian chert, serpentized ophiolites, and low-temperature, high-pressure metamorphic rocks (Irwin, 1990). This assemblage is believed to have been formed as a subduction complex (Ernst, 1970) along convergent boundary (e.g., Atwater, 1970) between the

Figure 1. Map of the study area showing the names of major faults and the cross section lines along which I show the inversion results. "LP" is Loma Prieta peak as a landmark.

Figure 2. Geological map of the study area (from Wentworth, 1993)

Farallon and North American plate, prior to the development of the transform boundary. From paleomagnetic data, some units of the Franciscan assemblage appear to have originated several thousand kilometers to the south (Alvarez et al., 1980). Seismic refraction studies indicate the possibility of a remnant slab of the Farallon plate beneath the Diablo Range (Fuis and Mooney, 1990).

The Franciscan assemblage is overlain locally by the Coast Range ophiolite and the Great Valley group (Bailey et al., 1970). The Coast Range ophiolite is exposed at only one place in the study area near the intersection of the San Andreas and Sargent faults. Paleontologic and paleomagnetic studies suggest that the Coast Range ophiolite formed in an equatorial setting and was transported northward (McLaughlin et al., 1988). The Coast Range ophiolite is overlain by the rocks belonging to the Great Valley group. There are no outcrops of the Great Valley group in the study area. The Great Valley Group is mainly Cretaceous and Early Tertiary. It includes marine sedimentary rocks with less deformation than the other two assemblages. This group is interpreted as forearc-basin sediments from the same subduction regime as the Franciscan (Dickinson and Seely, 1979). However, how these rocks and ophiolite were juxtaposed with the Franciscan rocks is the subject of debate. One hypothesis involves a midcrustal detachment fault (Coast Range fault) along which the upper plate was extended during the uplift of the

Franciscan assemblage as a tectonic wedge (Jayko et al., 1987, Platt, 1986).

Faults

San Andreas Fault

Near the Loma Prieta earthquake epicenter, the San Andreas fault (hereafter called "SAF") changes strike approximately from N40°W in the southeast to N50°W in the northwest. This left-stepping bend, hence a compressional or "restraining" bend, is also seen in the Zayante fault, which runs parallel to the SAF near the Loma Prieta, with almost the same strike change (Figure 1). The October 17, 1989, Loma Prieta earthquake ($M_S = 7.1$, $M_W = 6.9$) occurred along this left-stepping bend of the SAF. Its hypocenter depth is estimated to be 18 km (Oppenheimer, 1990), which is unusually deep for this area. The focal mechanism solutions of the Loma Prieta main-shock suggest a combination of reverse and strike-slip motion (Oppenheimer, 1990). Geodetic data also suggest right-lateral slip of 1.6 m plus 1.2 m of reverse slip (Lisowski et al., 1990). Dietz and Ellsworth (1990) investigated the aftershock sequence and identified a dipping zone that rises from the mainshock hypocenter (at 18 km depth) to 10 km depth. This zone is parallel to the mainshock nodal plane, strikes N51°W±2° and dips 65°SW±5°, and may be the strike and the dip of the SAF in this restraining bend. Dietz and Ellsworth suggest that the reverse slip

observed in the center of the aftershock sequence is consistent with the growth of fault-bend folding. Such folding was modeled later by Shaw et al. (in press). However, at the southern end of the rupture zone (about 18 km southeast of the epicenter), seismicity falls on a near-vertical plane (above 10 km depth) beneath the surface trace of the SAF.

Shaw et al. (in press) discuss the mechanism of the restraining bend in detail and conclude that right-lateral motion along the SAF and Zayante faults in these bends results in transpression and uplift of rocks adjacent to the fault bends. This uplift as well as folding of the hanging wall block are caused by thrusting over fault surfaces. Using a kink-band migration model to show the fold growth, they predict the non-planar subsurface fault geometry in this region. However, the depth of the SAF is not well constrained in their model, since the bottom edge of the rupture plane of the Loma Prieta earthquake is not necessarily the termination depth of the SAF.

Other Major Faults

The Sargent fault is a short, probably reverse fault that lies on the northeast side of the SAF. The southern portion of the Sargent fault is subparallel to the SAF. To the northwest it changes to N58°W, near the restraining bend of the SAF, and it is truncated by the SAF. The dip of this

reverse fault is averaging -60°SW near the surface and possibly flattening to the depth of 8 km where it is truncated by the SAF (McLaughlin et al., 1991).

The Zayante-Vergeles fault (hereafter called the Zayante fault) is also a reverse fault, but lying southwest of the SAF. It is subparallel to the SAF in the Loma Prieta region but changes strikes at both ends. The southern end (the Vergeles fault) converges to the SAF, while the northern end changes strike gradually to almost due west. As described earlier, Shaw et al. (in press) discussed the mechanism of the restraining bend in the Zayante fault. Its dip is about 60°W and its vertical extent is not known well (Clark et al., 1989).

The Calaveras fault runs from south of Hollister to Danville—more than 120 km. Its strike is about $\text{N}27^{\circ}\text{W}$ and it is a nearly vertical fault. This fault is a seismically active dextral strike-slip fault, but part of the strike-slip motion of the northern part of this fault is converted to oblique-slip or thrusting over a wide area (Page, 1982). One hypothesis is that the current lithospheric boundary between the Pacific and North American plates lies beneath this fault (Furlong et al., 1989). If this is true, the effective depth of this fault may be much greater than those of the other faults.

PREVIOUS STUDIES

Zandt (1981) carried out the first comprehensive teleseismic tomography study in central California. Figure 3 shows his results. By teleseismic *P* wave travel time inversion with 10-25 km horizontal block sizes, he found a low-velocity zone mainly on the northeast side of the SAF at 10-30 km depths. He attributed this low-velocity anomaly primarily to crustal thickening and additional effects of variations in P_n velocity. In the upper mantle, at 30-90 km depths, he found a linear zone of low-velocity running subparallel to the Calaveras fault. He interpreted this low as asthenospheric material rising to an unusually shallow depth.

The local-earthquake tomography of Lees (1990) yielded images of *P* wave velocity perturbations to 18 km depth. (Local earthquake tomography is effectively blind below the deepest hypocenters in a given area.) Figures 4(a)-(c) show his results. Lees found a low-velocity anomaly southwest of the SAF between $36^{\circ}58'$ - $37^{\circ}02'$ latitude north at $121^{\circ}46'$ longitude west, from near-surface to depths of 7-9 km. This low-velocity anomaly extends to the northwest, forming a low-velocity anomaly wedge with its long axis parallel to the SAF. Northeast of and below this low, a high-velocity anomaly is coincident with much of the aftershock activity. This high extends to 18 km depth. Lees (1990) inferred that

Figure 3. Compressional-wave velocity perturbation maps for central California from Zandt (1981). Layer depths and initial velocities are as shown. Squares represent the horizontal block size in each layer.

Figure 4(a) Map of the study area of Lees' (1990) local-earthquake tomography, showing the location of his cross section lines and receivers (triangles). Large hatched area is the approximate rupture zone of the Loma Prieta earthquake based on seismicity plots.

Figure 4(b) Cross sections of compressional-wave velocity perturbations along the lines shown in Figure 4(a) (after Lees, 1990). Dark area means low-velocity. Earthquakes used in the inversion model are shown as dots.

Figure 4(c) Compressional-wave velocity perturbation maps of Lees (1990). Otherwise are the same as 4(b).

this high-velocity material is relatively brittle and competent, capable of supporting high shear stress. Eberhart-Phillips et al. (1990) also present models derived by local-earthquake tomography, and interpreted in concert with resistivity models. Figures 5(a)-(c) show their results which are similar to Lees' (1990) but not identical. A low-velocity, low-resistivity wedge of material from near surface to depths of over 8 km between the SAF and Zayante faults was found. They interpreted it as Tertiary marine sedimentary rocks; it is consistent with Lees' (1990) result. They also found a high-velocity wedge within 2 km of the surface between the San Andreas and Sargent faults. They concluded that this high-velocity units may be Salinian granitic or mafic intrusive rocks. This result differs from Lees (1990).

Both Lees' (1990) and Eberhart-Phillips et al.'s (1990) results are also in good agreement with the results of local earthquake tomography by Lees and Shalev (1992), and with the *P* and *S* wave attenuation structure derived by Lees and Lindley (1994). Both papers support the location of low-velocity body southwest of the SAF. Lees and Shalev's (1994) model is consistent with Lees' model (1990) in terms of the location of the high-velocity body northeast of the SAF. From high Q_p and low Q_s , Lees and Lindley (1994) suggest that pores are fully saturated in the aftershock zone at depths of 8-15 km. The maximum depths of all these studies are about 18 km, so that little is known about the structure below.

Figure 5(a) Map of Eberhart-Phillips et al.'s (1990) local-earthquake tomography study area.

Figure 5(b) Velocity cross sections along the San Andreas fault after Eberhart-Phillips et al. (1990).

Figure 5(c) Initial model and velocity cross sections across the San Andreas fault after Eberhart-Phillips et al. (1990).

INVERSION TECHNIQUE

Seismic tomography is the technique of imaging zones of complex, heterogeneous velocity and/or attenuation structure of the Earth by using redundant measurements of some property of seismic waves (Iyer, 1989). Body waves, surface waves, normal modes, and refraction or reflection data can be used in tomography. Body wave travel times are the most commonly used data in tomography. Measurements of attenuation also are common and whole waveforms have been used in a few cases. Body wave tomography with natural sources is typically divided into local-earthquake tomography and teleseismic tomography. The former uses earthquakes generally occurring within the modeled volume, while teleseismic tomography sources usually are more than 25° away. Teleseismic tomography can be applied to a greater depth than local-earthquake tomography, although it does not deal with absolute velocities, which local-earthquake tomography can do.

Since the 1930s, seismologists have realized that distant earthquakes seldom reach the recording stations exactly at times predicted by available velocity models of the Earth. The differences, called travel-time residuals, are used to infer heterogeneities within the Earth.

Aki, Christofferson, and Husebye (1977) were the first to develop an inversion technique for teleseismic residuals

(dubbed the "ACH" method after their initials). The 3-D modeled volume is divided into blocks, each with an unknown velocity perturbation. From travel-time data and a reference velocity model, a system of equations is derived. By using the damped least squares mathematical inversion technique these unknowns are solved for. The ACH method is still in wide use. It is a robust method because of the absence of ray turning points inside the modeled volume. Hence raypaths depend only weakly on the velocity structure being determined. Local-earthquake tomography, on the other hand, is strongly non-linear due to this interaction of velocity and raypath.

In the following sections, steps in the technique, as used by Evans and Achauer (1993), are described.

Calculation of Relative Residuals

The first step in tomography is to calculate the travel-time residual R_{ij} at the i th station for the j th event with respect to a known one-dimensional Earth model. However, this residual has errors due to origin-time uncertainty, hypocenter mislocation, and velocity heterogeneity outside the modeled volume. To eliminate these errors, relative residuals (RR_{ij}) are calculated by subtracting a reference residual (R_j), thought to be the representative of the receiver array for the j th event, from R_{ij} . Thus,

$$RR_{ij} = R_{ij} - R_j .$$

For R_j , the average residual computed over the whole network is sometimes used. In this research, a weighted mean residual is used instead. Hence,

$$R_j = \frac{\sum_{i=1}^{I_j} w_{ij} r_{ij}}{\sum_{i=1}^{I_j} w_{ij}}$$

where w_{ij} is a weight derived from pick quality metrics determined by the analyst (from signal-to-noise ratio and signal-coherency considerations).

Parameterization of the Modeled Volume

The next step is to parameterize the target volume, in this case by dividing it into layers and subdividing each layer into blocks. In this research, the modeled volume is divided into layers of about equal vertical travel-time, so layer thickness increases slowly with depth. The total thickness of the modeled volume should be roughly equal to the aperture of the receiver array, because below this depth, P wave ray bundles diverge, so that deeper objects cannot be resolved well. The appropriate block width is about the same as the average station spacing. If the blocks are narrower, the inversion result will become oscillatory, with highs and lows appearing repeatedly. If the blocks are wider, one is not making maximum use of the data set. For simplicity and consistency with other studies, and to avoid ill-tested software conditions when rays exit along the side of the model, I sized and positioned each layer so that all rays are

confined within it. Since rays diverge at depth, this process yields a pyramidal model volume. Each block has unknown fractional slowness deviation from the initial-model slowness of the layer to which the block belongs ("slowness" is the reciprocal of wave-velocity). As rays approach the surface, it accumulates travel-time deviation depending on which blocks the ray travels through. The sums of the travel-time deviations (i.e., travel-time residuals) are expressed in a matrix equation

$$\mathbf{b} = \mathbf{G}\mathbf{m}$$

where $\mathbf{b}$ is a vector essentially containing RR_{ij} , $\mathbf{m}$ is also a vector containing the unknown fractional perturbation of slowness that each block has, and $\mathbf{G}$ is a matrix derived from partial derivatives, which turn out to be the unperturbed travel times of ray ij in block k . (I write "essentially" and "derived from" because I have skipped the details at data weighting and parameter separation in this presentation.)

Choice of Damping

To solve this equation, the damped least squares method is used in this study. Namely, $\mathbf{G}$ is singular—there is no unique solution. (There are many choices of additional constraint to make $\mathbf{G}$ non-singular. Other common ones are smoothness constraints and the "generalized inverse.") Therefore,

$$\hat{\mathbf{m}} = (\mathbf{G} + \theta^2 \mathbf{I})^{-1} \mathbf{b}$$

where $\hat{\mathbf{m}}$ is a solution of the previous equation, and θ^2 is a damping constant that weights the additional constraint that the length (complexity) of $\hat{\mathbf{m}}$ also be minimized. The damping constant is chosen empirically by plotting a trade-off curve between residual (i.e., unmodeled) data variance and squared model length (i.e., model complexity), while varying the damping constant over a wide range. If damping is too small, then overfit of noise in the data set appears as a large increase in model complexity. On the other hand, if a damping value is too large, then one discards real anomalies in the data set and leaves excessive data variance. The optimal damping value is near the origin of the curve, where neither extreme is evident, and near a residual data variance comparable to the square of expected reading errors.

Resolution and Covariance

During the calculation of a solution, we also derive two, important, related matrices called the resolution matrix, $\mathbf{R}$, and the covariance matrix, $\mathbf{C}$. Both are used to help evaluate the reliability of the computed model. $\mathbf{R}$ describes the relationship between the inversion result $\hat{\mathbf{m}}$ and the parameterized "true" (unknown) Earth model $\mathbf{m}$ (where "true" is limited by the parameterization). $\mathbf{R}$ is defined so that

$$\hat{\mathbf{m}} = \mathbf{R}\mathbf{m}.$$

$\mathbf{R}$ is a symmetric, singular matrix (i.e., one can not find $\mathbf{m}$ by inverting $\mathbf{R}$). Each row or column of $\mathbf{R}$ is what an isolated 1% anomaly in that block of $\mathbf{m}$ would look like in $\hat{\mathbf{m}}$. In general, resolution is sharpest in the center of the modeled volume and changes gradually to much poorer resolution at the periphery. Furthermore, applying damping to a subvertical ray set weakens the resolution between layers, while the horizontal resolution remains sharp. Although an evaluation of $\mathbf{R}$ is required when one interprets $\hat{\mathbf{m}}$, $\mathbf{R}$ fails to predict some artifacts. Evans and Achauer (1993) examined several cases where caution is required in interpretation, for example horizontal slabs.

The posteriori covariance matrix, $\mathbf{C}$, describes the effect of uncertainty in the data on the values in $\hat{\mathbf{m}}$. The square root of the diagonal elements of $\mathbf{C}$ are the standard errors of the model perturbations. These uncertainties, together with the size and shape of the columns of $\mathbf{R}$, help one determine the smallest magnitude and spatial size of an interpretable velocity anomaly.

Smoothing

The inversion result using the layer thickness and block width described above will have a 'blocky' appearance that can be difficult to interpret and may contain certain artifacts of the parameterization itself, particularly for spatially small features. To make interpretation easier and

reduce artifacts, it is desirable to smooth the images.

Vertically, we can thin each layer by dividing it into thinner layers, in this case layers of equal travel time. This is possible only because damping in the absence of subhorizontal rays weakens resolution between layers. The total model depth will, however, stay the same. The damping constant must be reduced by the same ratio that the number of modeled blocks increases. If the damping value stays the same and is applied to every layer (the number of layers in the thinned-layer model is three times greater than in the base model), the result is overdamping. That is, the balance between G and $\theta^2 I$ would be thrown off, since the same number of data constrain the model (G) while damping is applied more extensively ($\theta^2 I$ where I is three times as many elements as before). Therefore, damping value must be reduced to compensate for the greater number of blocks in the modeled volume.

Because diagonal elements of R decrease rapidly with layer thinning, a new way to evaluate resolution is required. Instead of requiring large values in diagonal elements of R which is no longer possible, it is necessary as well as more informative to understand the size and shape of larger, imaginary 'blocks' that would have diagonal elements d of some desired magnitude. This presentation was named "volume metric" by Evans and Achauer (1993). "The 'volume metric of R for equivalent diagonal element of d ' is the volume of the

model within which the elements of that column of $\mathbf{R}$ sum to d , when summing the largest positive elements of $\mathbf{R}^*$ (Evans and Achauer, 1993). Introducing the volume metric also gives information about the shape and orientation of the columns of $\mathbf{R}$, yielding a more complete picture of the information in $\mathbf{R}$.

For horizontal smoothing, block narrowing can not be used, since it induces an oscillatory velocity pattern. If the block width is too small, then there will be some blocks through which few rays travel. Therefore, an offset-and-average method is employed for horizontal smoothing. By offsetting each layer of the thinned-layer model by increments of $1/n$ th of the block width along each horizontal axis and inverting models with all no-redundant permutations of offsets, n^2 inversions result. Horizontal smoothing is done by taking the average of these n^2 perturbation results, within each $1/n$ th- $1/n$ th-width block overlap region. This process should be done after layer thinning, since the thickness of each layer is kept constant during the offset-and-average process. The result is a thinned, smoothed (but not yet stripped) model. By this scheme, one expects to reduce noise, mitigate the artifacts caused by parameterization, and better locate anomaly edges.

Stripping of the Shallow Structure

The shallowest layer can be treated differently from the others. The shallowest layer often has large travel time

perturbations because of local geology, such as unconsolidated sediments, high porosity, and weathering. The perturbations caused by these shallow upper-crustal structures can distort deeper features (i.e., they smear downward when filtered through R). Also, there is often independent information about shallow structure. Therefore, to interpret deeper structure better, methods to eliminate the effects of shallow features are desired. Consequently, an iterative stripping of the first layer is used. The shallowest layer of the model is inverted alone and repeatedly. At each inversion, the calculated velocity perturbations of the prior inversion are added to the perturbation input to that layer as correction terms. This procedure continues until these correction velocities converge to stable values. Then the last of these single-layer inversion results is used in the inversion of the whole modeled volume. By this stripping, all travel time perturbations which can be explained by shallow crustal velocity structure are removed, so that the deeper structure can be evaluated more accurately. The output of this technique is a thinned, smoothed, and stripped model. If deeper features are stable between the stripped and unstripped models, one is confident that these features are not artifacts of shallow structure.

Dawson et al. (1990) pointed out that iteratively stripped models may be overcorrected and some travel-time

perturbations in fact caused by deeper structure are removed. Both unstripped and stripped models are probably extreema in this sense, so they are end members. A best estimate is thought to fall between these two extreema. However, this in-between model, obtained from local-earthquake tomography, is beyond the scope of this thesis and is planned for a subsequent *JGR* paper.

TELESEISMIC DATA

Design of the Receiver Array

Using portable instruments to augment permanent "CalNet" stations of the U.S. Geological Survey (USGS), the array used here was designed by Dr. John Evans at the USGS to yield three nested studies (Figures 6(a) and 6(b)).

(1) A regional study with resolution about the same as the mean CalNet station spacing (about 12 km) to a depth of 110-120 km.

(2) A 6-km horizontal-resolution study to a depth of 60-70 km in and near the Loma Prieta aftershock zone. This study is the principal result of this thesis.

(3) A 3-km horizontal resolution study to a depth of 30-40 km in the vicinity of the San Andreas and Sargent faults' intersection, where structure was thought to be complicated and informative.

Table 1 shows the locations of receivers. An "X" in Table 1 means that data from that station were used in the models described in the modeling section. (Two stations which have not been analyzed are shown for completeness in Table 1.)

Teleseisms used in this study were recorded during the period of August, 1990, to December, 1990. CalNet stations have vertical 1-Hz sensors with signals telemetered to a central recording facility at the USGS in Menlo Park,

Figure 6(a) Distribution of temporary seismic stations used in this study.

Figure 6(b) Distribution of permanent CalNet stations used in this study.

Table 1. Station Coordinates

Station	Latitude	Longitude	Elevation	Used in Model*		
	North	West	(m)	D	P	R
CAD	37 ° 09.83 '	121 ° 37.55 '	207			X
CCO	37 ° 15.46 '	121 ° 40.35 '	366			X
CMH	37 ° 21.57 '	121 ° 45.38 '	518			X
CSC	37 ° 17.11 '	121 ° 46.35 '	128			X
HAZ	36 ° 53.08 '	121 ° 35.45 '	122			X
HBT	36 ° 51.01 '	121 ° 33.04 '	98			X
HCA	37 ° 01.52 '	121 ° 29.02 '	332			X
HCB	36 ° 55.88 '	121 ° 39.63 '	219			X
HCO	36 ° 53.25 '	121 ° 42.49 '	134		X	X
HCR	36 ° 57.46 '	121 ° 35.01 '	241		X	X
HDL	36 ° 50.12 '	121 ° 38.64 '	204			X
HER	36 ° 47.90 '	121 ° 42.64 '	69			X
HFE	36 ° 59.00 '	121 ° 24.09 '	323			X
HFH	36 ° 53.29 '	121 ° 28.13 '	101			X
HGS	37 ° 05.75 '	121 ° 26.83 '	778			X
HGW	37 ° 01.02 '	121 ° 39.20 '	133		X	X
HJG	36 ° 47.88 '	121 ° 34.43 '	171			X
HKR	36 ° 54.10 '	121 ° 25.56 '	66			X
HOR	36 ° 55.03 '	121 ° 30.46 '	98			X
HPH	36 ° 51.38 '	121 ° 24.37 '	122			X
HPR	36 ° 57.19 '	121 ° 41.70 '	94		X	X
HSF	36 ° 48.72 '	121 ° 29.97 '	340			X
HSP	37 ° 06.91 '	121 ° 30.94 '	850			X
JAL	37 ° 09.50 '	121 ° 50.82 '	244		X	X
JBC	37 ° 09.62 '	122 ° 01.57 '	660	X	X	X
JBG	37 ° 20.52 '	122 ° 20.34 '	158			X
JBL	37 ° 07.69 '	122 ° 10.08 '	792			X
JBM	37 ° 19.09 '	122 ° 09.16 '	820			X
JBZ	37 ° 01.07 '	121 ° 49.15 '	213	X		X
JCB	37 ° 06.71 '	121 ° 41.33 '	192		X	X
JEC	37 ° 03.04 '	121 ° 48.56 '	438		X	X
JEL	36 ° 55.64 '	121 ° 49.61 '	85			X
JHL	37 ° 06.54 '	121 ° 49.99 '	908	X		X
JHP	37 ° 26.65 '	122 ° 18.09 '	347			X
JLX	37 ° 12.11 '	121 ° 59.17 '	244	X	X	X
JNA	37 ° 10.62 '	121 ° 50.68 '	512	(Not used)		
JPL	36 ° 58.62 '	121 ° 49.93 '	158		X	X
JPP	37 ° 15.81 '	122 ° 12.78 '	186			X
JPS	37 ° 11.94 '	122 ° 20.90 '	84			X
JRG	37 ° 02.22 '	121 ° 57.87 '	213		X	X
JRR	37 ° 03.27 '	121 ° 43.61 '	408	X	X	X
JSC	37 ° 17.07 '	122 ° 07.42 '	357			X
JSF	37 ° 24.31 '	122 ° 10.55 '	143			X

Table 1. Station Coordinates (Continued)

Station	Latitude		Longitude		Elevation (m)	Used in Model*		
	North		West			D	P	R
JSG	37 °	16.96 '	122 °	03.00 '	198			X
JSJ	37 °	20.03 '	122 °	05.48 '	122			X
JSM	37 °	12.74 '	122 °	10.06 '	262			X
JSS	37 °	10.17 '	121 °	55.84 '	946	X	X	
JST	37 °	12.41 '	121 °	47.84 '	149		X	X
JTG	37 °	01.71 '	121 °	52.58 '	253	X	X	X
JUC	37 °	00.07 '	122 °	02.91 '	177			X
JUM	37 °	09.65 '	121 °	53.86 '	1048	X		X
LD1	37 °	09.62 '	121 °	58.31 '	216	X	X	X
LD2	37 °	06.22 '	121 °	55.87 '	549	X		
LD3	37 °	04.60 '	121 °	56.65 '	302	X	X	
LD4	37 °	02.60 '	121 °	55.04 '	360	X		
LD5	37 °	08.53 '	121 °	51.48 '	372	X	X	
LD6	37 °	04.66 '	121 °	52.98 '	652	X		
LD7	37 °	05.02 '	121 °	51.37 '	506	X		
LD8	37 °	02.92 '	121 °	50.39 '	634	X		
LD9	37 °	05.24 '	121 °	47.28 '	378	X		
LDA	37 °	04.17 '	121 °	46.04 '	351	(Not used)		
LDD	37 °	08.55 '	121 °	48.97 '	605	X	X	
LT1	37 °	13.77 '	122 °	03.49 '	549		X	X
LT2	37 °	07.14 '	122 °	04.49 '	378		X	X
LT3	37 °	04.00 '	122 °	02.21 '	195		X	X
LT4	37 °	13.88 '	121 °	55.58 '	183		X	X
LT5	37 °	08.01 '	121 °	56.59 '	387	X	X	
LT6	37 °	06.00 '	121 °	59.48 '	317	X	X	X
LT7	37 °	06.07 '	121 °	53.60 '	597	X	X	
LT8	36 °	59.80 '	121 °	58.12 '	72		X	
LT9	36 °	59.32 '	121 °	53.99 '	98		X	X
LTA	37 °	13.64 '	121 °	50.77 '	155		X	
LTB	37 °	06.72 '	121 °	45.08 '	207		X	X
LTC	36 °	57.81 '	121 °	46.40 '	55	X	X	
LTD	37 °	10.42 '	122 °	05.68 '	384		X	X
LTE	36 °	54.40 '	121 °	47.61 '	20		X	
LTF	37 °	09.40 '	121 °	43.35 '	219		X	X
LTG	37 °	03.85 '	121 °	37.96 '	111		X	X
LTH	36 °	59.58 '	121 °	43.89 '	163	X	X	
LTI	37 °	07.72 '	121 °	54.21 '	655	X	X	

* An "X" means that the data from that station are used in the model series D, P, and R. See Table 5 for explanation of definitions of these model series.

California. The temporary instruments in Figure 6(a) have both vertical and two horizontal (north-south and east-west) components, and events were recorded locally on magnetic tapes.

Digitizing and Data Selection

All events and associated time codes were digitized at 100 samples per second at the USGS except stations LDA, LD8, and LD9, which were digital recorders. The data used in this study were selected to make a reasonably homogeneous distribution of rays within the target volume beneath the receiver array. Figure 7 shows the epicenters of events used in this study. *P* phases with $25^{\circ} \leq \Delta \leq 100^{\circ}$ and *PKIKP* phases with $\Delta \geq 110^{\circ}$ are used in this study. Rays from these ranges travel with steep inclination angles in the modeled volume. There are three reasons to use the rays from these ranges. One is that they usually record well. Another reason is that they have no turning points in the modeled volume. Finally, they span the available set of incidence angles about as well as can be. Table 2 shows event origin-times and hypocenters. Most *P* phases recorded in the western U.S. come from three directions: (1) the northwest, such as Japan, Kamchatka, and Alaska, (2) the southwest, such as Tonga, Fiji, and the New Hebrides, and (3) the southeast, such as Central and South America. Only a few events come from the northeast, such as the Mid-Atlantic Ridge. Five *PKIKP* phases are also included

Figure 7. Azimuthal equidistant maps of epicenters (dots) of the earthquakes used in this study. Diamond in (a) is centered on the study area. Its antipode is shown as a diamond in (b). Dark circles in (a) are at 25° and 100° distance from the center of the study area. The circle in (b) is at 110° from the study area (hence, 70° from the antipode).

TABLE 2. Dates and Locations of Events Used in This Study

Location	Date	Origin Time UTC	Latitude North	Longitude West	Depth km
Chile	08/15/90	13:23:36.3	-18°50.0'	69° 5.8'	120
Hokkaido	08/15/90	23:08:56.0	43°45.2'	-143°17.8'	162
Xinjiang	08/16/90	04:59:57.6	41°33.8'	-88°46.2'	0
Fox island	08/17/90	01:28:43.8	52°28.4'	167°29.9'	33
Solomon	08/17/90	13:07:17.4	-11° 9.8'	-161°59.8'	29
Nicobar	08/18/90	18:52:36.8	7°30.4'	-93°58.5'	21
Fiji	08/19/90	08:03:44.0	-20°37.1'	178° 2.3'	556
Sakhalin	08/20/90	00:03:52.7	46°11.3'	-142°17.3'	309
Solomon	08/20/90	14:11:43.0	-11°14.8'	-161°55.1'	44
Kermadec	08/21/90	00:57:28.5	-33°22.4'	178° 2.8'	33
Easter Island	08/21/90	14:13:04.8	-27°29.2'	104°13.6'	11
Colombia	08/25/90	11:47:31.2	5°47.5'	77°30.2'	10
Cuba	08/26/90	07:53:41.6	19°33.5'	77°52.4'	10
Honshu	08/27/90	10:52:53.0	29°21.7'	-142°58.3'	34
Chile	08/28/90	08:58:05.8	-19°38.7'	69°52.6'	64
North Atlantic	08/30/90	15:54:29.8	15°58.1'	47°55.2'	10
Panama	08/30/90	18:33:00.8	6° 8.8'	82°38.0'	10
Ecuador	09/02/90	04:26:48.0	-00° 8.6'	80°17.0'	14
Vanuatu	09/03/90	01:02:51.0	-14°29.3'	-167°47.9'	84
Cuba	09/04/90	08:03:04.2	19°48.2'	75°41.4'	33
Colombia	09/04/90	17:51:41.6	6°51.1'	72°59.3'	157
Argentina	09/04/90	23:15:11.3	-31°15.7'	69° 3.8'	112
Sandwich	09/08/90	20:21:48.3	-57°53.3'	25°26.0'	33
Tonga	09/08/90	20:31:57.7	-20°32.9'	174°10.4'	33
New Britain	09/09/90	05:35:45.9	-5°10.7'	-151°43.7'	37
Andreanof	09/10/90	07:30:07.6	52°18.1'	178°41.5'	143
Kamchatka	09/12/90	22:45:14.7	52°38.1'	-157°41.6'	134
Honshu	09/13/90	02:58:46.5	33°29.7'	-140°59.5'	51
Alaska	09/14/90	07:00:02.6	51°27.4'	164° 7.1'	33
Sandwich	09/14/90	09:20:18.9	-60°41.3'	22°47.6'	33
Easter Island	09/15/90	12:21:43.6	-24°27.3'	117°34.7'	10
Iceland	09/15/90	23:07:42.8	64°39.3'	17°37.0'	10
Sumatra	09/17/90	11:57:24.1	-5°55.0'	-103°47.8'	59
Aleutian Island	09/22/90	08:52:36.9	50° 0.5'	178° 3.1'	26
Honshu	09/23/90	21:13:07.4	33°16.0'	-138°38.6'	10
Honshu	09/23/90	22:09:04.2	33°11.3'	-138°41.1'	10
Peru	09/24/90	15:33:15.6	-10°37.1'	73°26.6'	135
South Pacific	09/26/90	15:40:42.4	-36°11.4'	100°47.7'	10
Rat Island	09/27/90	21:07:38.4	51°41.3'	-176°18.4'	34
Vanuatu	09/28/90	19:44:47.1	-13°33.5'	-167° 4.7'	176
Mariana	09/29/90	21:39:43.7	13° 6.0'	-145°13.9'	53
Tonga	10/02/90	15:06:44.6	-24° 2.2'	174°38.8'	9
Carlsberg	10/06/90	09:08:09.1	-2°42.7'	-67°54.5'	10
Fiji	10/10/90	05:54:53.5	-23°29.8'	-179° 1.7'	549
Bonin Island	10/10/90	08:22:53.6	27°11.9'	-140° 7.1'	384
Shikoku	10/10/90	15:32:16.2	29° 9.9'	-132° 0.0'	33
Dominican	10/11/90	15:14:45.0	18°43.6'	69°22.6'	101
Mariana	10/13/90	00:20:23.3	15°43.4'	-147°55.9'	37
Ecuador	10/13/90	11:14:33.1	-2°13.5'	80° 2.6'	76
North Atlantic	10/15/90	06:30:20.3	15°18.7'	46°21.0'	10
Kuril	10/16/90	06:13:13.7	49° 2.6'	-155° 4.6'	83
Peru	10/17/90	14:30:13.1	-10°58.2'	70°46.6'	599
Leeward Island	10/18/90	19:12:15.9	17°38.6'	62°10.3'	54
Peru/Brazil	10/20/90	10:01:31.2	-7°43.7'	74°24.8'	162
Peru	10/21/90	15:10:43.7	-3°59.3'	77°16.4'	116
Novaya Zemlya	10/24/90	14:57:58.1	-73°21.7'	-54°42.4'	0

TABLE 2. Continued

Location	Date	Origin Time	Latitude	Longitude	Depth km
		UTC	North	West	
Solomon	10/25/90	16:33:33.4	-6°11.3'	-154°57.6'	142
Kuril	10/28/90	06:53:01.2	49°40.9'	-155°53.1'	48
Mexico	10/29/90	02:21:28.5	18°56.5'	102° 0.2'	99
Chile	10/31/90	14:39:27.2	-26°38.0'	70°28.9'	60
Hokkaido	11/02/90	12:50:17.8	41°11.3'	-142° 9.5'	71
Peru	11/04/90	18:13:43.0	-15°43.2'	72°37.1'	121
Rat Island	11/04/90	22:04:45.4	51°11.2'	-178°56.9'	33
Komandorsky	11/06/90	20:14:29.7	53°27.1'	-169°52.3'	25
Nicaragua	11/08/90	13:12:54.9	11°45.4'	85°52.3'	163
Mexico	11/10/90	01:14:12.7	15°26.5'	94°42.2'	39
Taiwan	11/11/90	23:59:03.5	24° 5.5'	-121°45.5'	34
Peru	11/22/90	14:25:12.7	-10° 7.8'	78°37.0'	48

to complement *P*, which arrive in the upper crust at inclinations of about 15° - 30° from the vertical. *PKIKP* travels through the outer and inner core (Figure 8) and provides a fifth group of rays that arrives nearly vertically. Some teleseisms from abundant source regions have been excluded to avoid the distortion of resolution kernels due to oversampling along these ray directions. The source magnitudes for *P* phases are mainly greater than 5.5 while those for *PKIKP* are about 0.3 magnitude units higher than for *P*.

Time correction

Automatic and manual time correction is applied to each trace by reading WWVB radio time codes. This precision time code is transmitted continuously by the National Institute of Standards and Technology on 60 kHz A.M.

Figure 8. A diagram showing the path for a typical *PKIKP* ray.
(Modified from Wallace and Lay, 1992)

DATA REDUCTION AND DEBUGGING

P Wave Arrival Time Picking

The first arrival of compressional waves on each trace is picked by hand and subjectively assigned a pick quality. First, a narrow-band filter is chosen and applied to each trace to optimize the signal-to-noise ratio. To pick at best available precision, one splits the area of a clear, early-arriving peak or trough of the waveform. Since teleseismic tomography does not constrain absolute velocity, absolute first arrival picks are not required (relative picks are sufficient). Because relative picks are more precise and less subjective, they are desirable. Later, one revisits the original (unfiltered) traces to verify coherence visually and particularly to guard against cycle skipping. This method is reliable, since many branches of teleseismic phases tend to be coherent over distances of 1000 km and more (Evans and Achauer, 1993). However, coherence decreases into the coda of each phase. Therefore, picking should be done on the earliest readable peak or trough in each phase.

A pick quality is assigned subjectively to each pick, depending on any distortion of the waveform and on the signal-to-noise ratio. The best quality is 'a', which is nominally ± 0.05 s at the 3σ (standard deviation) level, then 'b' with ± 0.1 s followed by 'c' with ± 0.25 s. These

qualities and estimated errors are used in calculation of weighted mean residuals.

Data Debugging

After all picking is done, the whole data set must be reviewed in depth to detect and eliminate reading errors. Evans and Achauer (1993) demonstrated the effects of a small number of extreme data errors. Such errors produce strong radial streaks through the whole model thickness, even in the well-resolved parts. Therefore, debugging is essential. For this debugging procedure, residuals relative to the event-median residuals are plotted at each station as a function of ray back-azimuth and slowness (i.e., of ray orientation). Since relative residuals generally vary smoothly with ray orientation, rays yielding residuals significantly different from surrounding rays may have errors. Once identified, these rays are checked for time- code reading errors and for picking errors. The original traces are consulted and command the process. That is, one must never allow the *suspicion* of an error to force a change to the data set. An error must be verified in the original seismograms to avoid biasing ("cooking") the data.

The final results at each station, versus weighted mean residuals are plotted in Appendix A. The final data set has 2705 usable picks for 68 events. There are 544 'a' picks, 1193 'b' picks, and 968 'c' picks.

INVERSIONS

Parameterization

Table 3 shows the reference velocity model of the study area from Dietz and Ellsworth (in press). They obtained the result in Table 3 by an inversion process using the Loma Prieta aftershocks recorded at stations on the southwest side of the SAF for model 1 and stations on the northeast side for model 2. The starting model is not critical to the teleseismic tomography result, since the ray turning points are excluded from the modeled volume. That is, the starting model has only secondary effects on ray locations, so also on velocity perturbations. However, it is better to use a starting model consistent with other studies of the region. I use the average of models 1 and 2 as the starting crustal velocity structure for my inversions. This crustal velocity structure was attached to Herrin's (1968) deeper model via a linear gradient from 25 km to 2894 km depth to create an Earth velocity structure. Once this velocity structure of the Earth was available, an Earth-flattening transform (Aki and Richards, 1980) and averaging of the Earth model within layers designated for my inversions were applied to obtain my initial models. The Earth-flattening transform enables more accurate ray tracing, since the inversion uses plane layers. Tables 4(a) to (f) show my initial-velocity models. Hereafter, the "detailed model" stands for the model with

TABLE 3. Reference compressional-velocity models
from Dietz and Ellsworth (in press)

Depth to Top of Layer (km)	Southwest of San Andreas (km/s)	Northeast of San Andreas (km/s)
0.0	3.35	2.99
1.0	4.58	3.99
2.0	5.26	5.57
5.0	5.96	5.98
8.0	6.16	6.26
13.0	6.22	6.44
18.0	6.52	6.58
25.0	8.00	7.98

Table 4a. Initial Base Detailed Model

Layer	Thickness km	Horizontal Block Size km x km		Initial Velocity km/s
1	2.90038	*		4.05638
2	4.00329	3.00	x 3.00	5.66981
3	4.40599	3.30	x 3.30	6.15671
4	4.45967	3.30	x 3.30	6.29700
5	4.64337	3.50	x 3.50	6.45997
6	4.64647	3.50	x 3.50	6.57318
7	5.72535	4.30	x 4.30	8.03653
8	5.72973	4.30	x 4.30	8.06595

Table 4b. Initial Thinned-layer Detailed Model

Layer	Thickness km	Horizontal Block Size km x km		Initial Velocity km/s
1	2.90038	*		4.05638
2	1.25081	3.00	x 3.00	5.41849
3	1.33124	3.00	x 3.00	5.60788
4	1.42124	3.00	x 3.00	5.97554
5	1.44171	3.30	x 3.30	6.03235
6	1.47208	3.30	x 3.30	6.21788
7	1.49220	3.30	x 3.30	6.22024
8	1.44347	3.30	x 3.30	6.22310
9	1.51268	3.30	x 3.30	6.32121
10	1.50352	3.30	x 3.30	6.34487
11	1.51911	3.50	x 3.50	6.34660
12	1.54956	3.50	x 3.50	6.46283
13	1.57470	3.50	x 3.50	6.57035
14	1.55527	3.50	x 3.50	6.57104
15	1.51525	3.50	x 3.50	6.57381
16	1.56594	3.50	x 3.50	6.57469
17	1.90762	4.30	x 4.30	8.02623
18	1.90819	4.30	x 4.30	8.03614
19	1.90953	4.30	x 4.30	8.04723
20	1.88911	4.30	x 4.30	8.05551
21	1.93014	4.30	x 4.30	8.06634
22	1.91048	4.30	x 4.30	8.07590

Table 4c. Initial Base Regional Model

Layer	Thickness km	Horizontal Block Size km × km		Initial Velocity km/s
1	6.90367	*		4.85802
2	8.86567	6.00	× 6.00	6.22649
3	9.27984	6.20	× 6.20	6.51604
4	11.45508	7.60	× 7.60	8.05124
5	11.52660	7.70	× 7.70	8.10527
6	11.56705	7.70	× 7.70	8.13664
7	11.59806	7.70	× 7.70	8.16378

Table 4d. Initial Thinned-layer Principal Model

Layer	Thickness km	Horizontal Block Size km × km		Initial Velocity km/s
1	6.90367	*		4.85802
2	2.91379	6.00	× 6.00	6.12468
3	2.94518	6.00	× 6.00	6.22107
4	3.00669	6.00	× 6.00	6.33392
5	3.03820	6.20	× 6.20	6.40293
6	3.12043	6.20	× 6.20	6.57123
7	3.12120	6.20	× 6.20	6.57388
8	3.81581	7.60	× 7.60	8.03118
9	3.81887	7.60	× 7.60	8.05163
10	3.82039	7.60	× 7.60	8.07092
11	3.84332	7.70	× 7.70	8.09087
12	3.83494	7.70	× 7.70	8.10659
13	3.84834	7.70	× 7.70	8.11838
14	3.84991	7.70	× 7.70	8.12739
15	3.86257	7.70	× 7.70	8.13666
16	3.85457	7.70	× 7.70	8.14588
17	3.86687	7.70	× 7.70	8.15419
18	3.86001	7.70	× 7.70	8.16464
19	3.87118	7.70	× 7.70	8.17251

Table 4e. Initial Base regional Model

Layer	Thickness km	Horizontal Block Size km x km		Initial Velocity km/s
1	13.73951	*		6.44329
2	18.09678	12.00	x 12.00	8.06806
3	18.24916	12.20	x 12.20	8.13272
4	18.35196	12.30	x 12.30	8.17605
5	18.42542	12.30	x 12.30	8.21866
6	18.50980	12.40	x 12.40	8.26259

Table 4f. Initial Thinned-layer Regional Model

Layer	Thickness km	Horizontal Block Size km x km		Initial Velocity km/s
1	11.30966	*		5.29298
2	4.47985	12.00	x 12.00	6.29736
3	4.59310	12.00	x 12.00	6.45932
4	4.66656	12.00	x 12.00	6.57347
5	6.02642	12.00	x 12.00	8.03706
6	6.03213	12.00	x 12.00	8.06843
7	6.03823	12.00	x 12.00	8.09879
8	6.09370	12.20	x 12.20	8.11771
9	6.04936	12.20	x 12.20	8.13285
10	6.10610	12.20	x 12.20	8.14764
11	6.11119	12.30	x 12.30	8.16166
12	6.11783	12.30	x 12.30	8.17633
13	6.12294	12.30	x 12.30	8.19017
14	6.12884	12.30	x 12.30	8.20432
15	6.13550	12.30	x 12.30	8.21883
16	6.16109	12.30	x 12.30	8.23281
17	6.17747	12.40	x 12.40	8.24707
18	6.15255	12.40	x 12.40	8.26193
19	6.17978	12.40	x 12.40	8.27882

* Rays are associated with stations rather than geographic blocks in layer 1.

about 3-km horizontal resolution, the "principal" model for 6-km resolution, and the "regional" model for 12-km resolution.

Two types of models are shown in Tables 4(a)-(f). "Base models" refer to the models shown in Tables 4(a), 4(c), and 4(e), and "thinned-layer models" mean the rest. Base models were used first for preliminary inversions to obtain (1) appropriate damping values and (2) meaningful diagonal elements of **R** and **C**. The thickness of layers in base models is about 1.5 times the mean station spacing. However, the inversion results using base models will have a 'blocky' appearance and may have parameterization artifacts as discussed in the inversion technique section. Therefore, thinned-layer models were made later by dividing each layer of the base models, except the first, into three layers. The vertical travel times were kept about constant in each layer within a given model. The nominal depth of the Moho is 25 km, which is common in each model, and I put a layer boundary there to improve ray tracing accuracy. The block width is also kept the same as the mean station spacing in each base model and each thinned-layer model. The size and position of each layer was chosen so that all rays were contained within the modeled volume.

Since there are so many variants of the models, I chose a systematic model naming convention to assist the reader. This convention is described in Table 5, using single

Table 5. Model Naming Convention

Experiment	Model Series	Layer Thickness	Stripping
Loma Prieta	D = Detailed	B = Base	U = Unstripped
1990	P = Principal	T = Thinned and averaged	S = Stripped
Teleseismic: LP90-	R = Regional	T* = Thinned, but not averaged	

D = "Detailed" model series derived from 3-km spaced subarray

P = "Principal" model series derived from 6-km spaced subarray

R = "Regional" model series derived from 12-km spaced subarray

B = "Base" models without layer thinning

T = "Thinned-layer" models with offset and averaged

T* = "Thinned-layer" models without offset and averaging

U = Directly inverted models ("unstripped")

S = Iteratively "stripped" models

For example, "LP90-P/T/S" means the principal series of models, with layer thinning, offset and averaged, and iteratively stripped.

characters to indicate base (B) versus thinned (T), stripped (S) versus unstripped (U), and (P) principal- versus (D) detailed- versus (R) regional-series models.

Damping Values

Figures 9(a)-(c) show the damping trade-off curves for base models of each series (D, P, R). Damping values selected on the basis of these curves are 5, 20, and 20 units for detailed (D), principal (P), and regional (R) model series respectively. For thinned-layer models (T), they are 2.00, 7.70, and 7.82, in the same order, reduced to compensate for the greater number of modeled blocks in these thinned-layer models.

ACH Inversions

Once appropriate damping values are chosen, images of velocity perturbation, along with resolution and covariance matrices, were derived through ACH inversion. Because of constraints on disk space and expected small gradual variations in the characteristics of **R** across each model, full resolution and covariance matrices were not calculated in thinned-layer models. Instead, resolution columns for representative blocks were calculated to obtain representative "volume metrics." Full **R** and **C** matrices were evaluated for base (B) models to obtain diagonal values used

Figure 9(a) Damping trade-off curve for the base model of the detailed series. "Residual error" refers to the standard deviation of the data portion left unmodeled by the inversion. One hopes to place this unmodeled residual near the estimated picking errors, nominally 0.05 s. Damping value selected for use in this model series is circled.

Figure 9(b) Damping trade-off curve for the base model of the principal series. Other explanations are the same as in Figure 9(a).

Figure 9(c) Damping trade-off curve for the base model of the regional series. Other explanations are the same as in Figure 9(a).

as summaries of the variations in model quality from one place to another.

Smoothing and Stripping

For horizontal smoothing, 3×3 offset-and-averaging was applied to the thinned-layer models. Finally, iterative stripping was applied to the first layers of each model, with these first-layer stripping velocities as corrections. Thinned, stripped models were also offset-and-averaged. Table 5 shows the names of all models presented in this thesis and its appendices.

INTERPRETATION OF RESULTS

In this section, I describe some prominent velocity anomalies, discuss resolution and standard errors and interpret the significant velocity features. The principal model series (Figures 10, 11, 12, and 13) is discussed and illustrated in the main text, and others are put in the Appendices C and D. The regional and detailed series mainly show modest differences due to different horizontal resolutions.

In Figures 10(a)-(g), the left color frame is the velocity perturbation map of the thinned, offset-and-average smoothed, stripped principal series model—my main result. In layer 1 the stripping velocities have been added back into the stripped model. Only every third layer is shown because the others are largely redundant subsequent to layer-thinning. Small dots are earthquakes occurring within the depth range of this thin layer from January 1, 1990, through May 31, 1994. The right color frame is the diagonal elements of the resolution matrix of the base model in the corresponding layer. (Base model layers are thicker, so layer numbering differs.) The gray scale frame on the subsequent page is the standard errors for the base model for the same layer as in the resolution plot. Resolution and standard errors for unstripped and stripped models are effectively identical; the former are shown here. To show

Figure 10(a) Results for the principal model. See explanation in the text (p. 50).

Figure 10(a) (Continued)

Figure 10(b) (Continued)

Figure 10(c)

Figure 10(c) (Continued)

Figure 10(d)

Figure 10(d) (Continued)

Figure 10(e) (Continued)

Figure 10(f) (Continued)

Figure 10(g) (Continued)

Figure 11(a) Results for the principal model. See explanation in the text (p. 86).

Figure 11(b)

Figure 11(c)

LP90-P/T/U Velocity SAF

LP90-P/T/S Velocity

Knowns added in first layer.

SAF

Figure 12(a) Results for the principal model.
See explanation in the text (p. 86).

S A F

LP90-P/T/U Velocity SAF

LP90-P/T/S Velocity

Knowns added in first layer.

SAF

Figure 12(b)

LP90-P/B/U Resolution

SAF

Contour interval= 0.1

LP90-P/B/U Standard Errors
SAF

Contour interval= 0.025%

Figure 12(b) (Continued)

LP90-P/T/U Velocity SAF

LP90-P/T/S Velocity

SAF

Figure 12(c)

SAF

Contour interval= 0.1

LP90-P/B/U Standard Errors
SAF

Contour interval= 0.025%

Figure 12(c) (Continued)

Figure 12(d)

Figure 12(d) (Continued)

LP90-P/T/U Velocity SAF

LP90-P/T/S Velocity

Knowns added in first layer.

SAF

Figure 12(e)

SAF

Contour interval= 0.1

LP90-P/B/U Standard Errors SAF

Contour interval= 0.025%

Figure 12(e) (Continued)

LP90-P/T/U Velocity

LP90-P/T/S Velocity

Knowns added in first layer.

Figure 12(f)

Figure 12(f) (Continued)

LP90-P/T/U Velocity

LP90-P/T/S Velocity

Knowns added in first layer.

Figure 12(g)

Contour interval= 0.1

LP90-P/B/U Standard Errors

Contour interval= 0.025%

Figure 12(g) (Continued)

LP90-P/T/U Velocity

LP90-P/T/S Velocity

Knowns added in first layer.

Figure 12(h)

Figure 12(h) (Continued)

LP90-P/T*/U Volume-Metric Equivalent Diagonal Elements

Figure 13. Cross sections of volume metrics for selected representative blocks (the outlined blocks) along planes defined in Figure 1. The darkest-shaded blocks are required for equivalent resolution-matrix diagonal element 0.8, darkest plus medium for 0.9, and all shadings together for 0.95. Loosely, the irregular shapes of these volume metrics represent the smallest resolvable objects. Equivalently, they represent the size and shape of a "block" that can be resolved well (i.e., modeled distinctly from a hypothetical neighboring "block" at similar size and shape).

LP90-P/T*/U Volume-Metric Equivalent Diagonal Elements

Figure 13 (Continued)

differences in shallower modeled structure between unstripped (U) and stripped (S) principal models, map view results of both for layers 1, 4, and 7 are shown together on the same pages in Figures 11(a)-(c). The left color frame is the velocity perturbations in the unstripped principal model, and the right is of the stripped principal model. Other explanations are the same as for Figure 10. Cross sectional velocity perturbations, resolution, and standard errors along the planes defined in Figure 1 are also presented in Figures 12(a)-(h). The top color frame of each pair is the velocity perturbations of the unstripped principal model, and the bottom frame is of the stripped principal model. Small dots are earthquakes occurring during the same period as in Figure 10, within 5 km of the cross section plane. On the subsequent page the top gray scale frame is the diagonal elements of the resolution matrix of the base model, and the bottom frame is the associated standard errors. Block sizes are larger in the base model, since it is neither layer-thinned nor offset-and-average smoothed.

Velocity

In general, many velocity gradients fall beneath the SAF, although vertical smearing may amplify this effect (see discussion section below). The difference between unstripped and stripped models (Figures 11(a)-(c)) is minor below layer 7 (i.e., in the lower crust and upper mantle), demonstrating

that the effects of shallow crustal structure do not play an important role at these depths.

In the uppermost crust (layer 1), a strong velocity contrast is observed across the SAF (Figure 10(a)). This velocity gradient is prominent on cross sections A-A', B-B', and C-C' (Figures 12(a)-(c)). The high-velocity feature lies on the northeast side of the SAF, near the intersection with the Sargent fault, while the low-velocity feature is seen on the southwest side of the SAF.

In the upper mantle (layer 10), another large low-velocity object (marked as "L" in velocity map of Figure 10(d)) is observed with maximum amplitude -3.5% . This low is very evident on cross section D-D' (Figure 12(d)). In layer 13 through layer 19, two low-velocity features (marked as "M1" and "M2" in Figures 10(e)-(g)) are displaced beneath the surface trace of the SAF in right-lateral sense. Near the southern edge of the study area (Gabilan Range), a high-velocity anomaly is seen, although it can not be resolved in any detail with these data.

Resolution and Standard Errors

The resolution and standard errors are typical of teleseismic tomography (Evans and Achauer, 1993). A well resolved area lies at the center of the study area, where the coverage of rays is nearest optimal. Resolution decreases

gradually toward the periphery, where fewer ray orientations constrain the models.

Figure 13 is cross sections of volume metrics along the lines C-C' and G-G' shown in Figure 1. Near the center of the study area, where resolution is good, the total volume to get particular equivalent diagonal element of R is smaller than for the blocks at periphery. This means that smearing of anomalies is less at the center and rather bad at the periphery. For blocks at the periphery the direction of smearing is systematically radial to the sensor array (i.e., it follows the locally-dominant ray path direction).

Block standard errors show similar patterns. However, the smaller values near periphery are not intuitively reasonable. Since resolution is poor there, a peculiarity of ACH tomography decreases these periphery-block values and they are meaningless. Near the center, where resolution is fair to good (≥ 0.5), these values are meaningful estimates of error. Generally speaking, standard errors are much less than velocity perturbations in this area, which implies that the velocity perturbation results are interpretable. For example, in Figure 10(a) standard errors are less than 0.6 % while velocity perturbations are up to 7.0%.

Interpretation

The high- and low-velocity anomalies in the upper crust show general agreement with previous local-earthquake

tomography studies, and correlation with surficial geological units can be inferred. The low observed on the southwest side of the SAF is thought to be the Tertiary marine sedimentary rocks overlying the Salinian basement (Eberhart-Phillips et al., 1990). I concur with this inference. In cross sections like A-A' (Figure 12(a)), the thickness of this low-velocity body appears to be more than 10 km. However, the thickness is not well constrained due to resolution effects (damping weakens the resolution between layers, so that the true thickness of this body can be smaller). Hence, local-earthquake tomography provides the better estimate of true thickness. Eberhart-Phillips et al. (1990) estimated the thickness of this body as approximately 8 km, while in Lees' (1990) model, it extends to 7-9 km depth.

The high-velocity anomaly in layer 1 on the northeast side of the SAF (near the Sargent fault) also confirms the results of the local earthquake tomography, although there is some disagreement about its location. Eberhart-Phillips et al. (1990) map this high between the SAF and Sargent fault. They interpret it as granitic or mafic intrusive rocks of the Salinian block. In my model and that of Lees (1990), the high extends eastward from near the Sargent fault. Except for the differences, shallow crustal structures found by local earthquake tomography are repeated here by teleseismic tomography.

The low-velocity body near the Moho (feature L in layer 10) is elongated parallel to the Sargent fault. In cross sections D-D' and C-C' (Figures 12(d) and 12(c)), this low-velocity feature appears diamond-shaped, which is the expected pattern for a horizontal slab or horizontal lenticular body (Evans and Achauer, 1993). Although the thickness of this body appears to be as much as 20 km, the true thickness could be much less. Hence, the diamond-like shape of this low-velocity body and its location between the highest part of the mountain range may imply that it is caused by a crustal root. Alternatively, it may be caused by high temperature due to laterally discontinuous shear heating of the uppermost mantle by shearing beneath the SAF. These possibilities are elaborated below, in the discussion section.

There are several possible causes of the low-velocity features (M1 and M2) observed between 45- and 70-km depth. These lows could be caused by temperature differences, compositional differences, or partial melting. One scenario to create such conditions is a subducted ridge segment. Atwater (1989) reconstructed the plate configuration of the eastern Pacific and western North America since 160 Ma, using paleomagnetic data. Her study shows that pieces of the Farallon plate which were separated from the Pacific plate by spreading centers have been rotated and then subducted beneath the North American plate. Another plate

configuration model including this area (Skaer, 1989) also shows that there is enough complexity in plate geometry during the final stages of subduction to create velocity perturbation in many ways. If such subducted ridge segments are still hot (since they are covered by the North American plate), the temperature contrast would contribute a low-velocity effect. Asthenosphere welling up through a break developed in a weak portion of a subducted slab could also decrease the velocity because of upwelling material's high temperature or even partial melting. A similar velocity low associated with an inferred slab window north of Clear Lake, central California, was also identified by Benz et al. (1992).

The high-velocity feature observed in the southern edge of the study area is very poorly resolved, particularly as to its depth, so the cause of this high remains enigmatic. It could, for example, be caused by compositional differences such as a slab remnant that has been subducted beneath this area or a gabbro accumulation. Further discussion of this high is beyond the scope of this paper.

Discussion

Velocity Gradients Beneath the San Andreas Fault

Tables 6(a)-(g) show the ratio of the fraction of the length of several arbitrary lines (defined in Figure 14) that happens to fall along any velocity gradient. To make a simple test of how significant it is that velocity gradients underlie the SAF, I ask how often velocity gradients underlie these lines, using the same subjective valuation applied to the SAF. Tables 6(a)-(f) suggest that velocity gradients falling beneath the SAF are somewhat more common than those falling beneath arbitrary lines, although the difference is not definitive. Indeed, some of the arbitrary lines match gradients over larger fractions of their length in some layers. However, Table 6(g) shows the totals across all measured layers, with the SAF ranking first. This result supports the inference of a correlation between velocity gradients and the SAF.

Low-velocity Body in Layer 10

The travel-time difference caused by the layer-10 low-velocity body is about 0.09 s, which is equivalent to a 2-5 km thick crustal root if caused by a root alone. The range of this estimate depends on the *P* wave velocity contrast across the Moho. Therefore better-constrained velocities would lead to a better estimate of the thickness of such a

Figure 14. A map of the lines used for the velocity-gradient tests shown in Table 6.

Table 6. Fraction of lines falling along velocity gradients.

(a) Layer 4

	Total Length (km)	Observed with Gradient (km)	Fraction with Gradient (%)
Line B	52.3	30.6	59.8
Line F	63.6	31.6	49.6
SAF	67.6	32.1	47.4
Line D	42.2	19.2	45.5
Line C	41.4	10.8	26.2
Line E	47.8	7.9	16.5

(b) Layer 7

	Total Length (km)	Observed with Gradient (km)	Fraction with Gradient (%)
Line E	53.0	22.7	42.8
SAF	75.4	32.1	42.5
Line B	60.2	20.7	34.4
Line F	69.0	23.7	34.3
Line D	46.0	13.8	30.0
Line C	42.9	0.0	0.0

(c) Layer 10

	Total Length (km)	Observed with Gradient (km)	Fraction with Gradient (%)
Line E	67.6	42.4	62.7
SAF	93.7	58.2	62.1
Line D	55.7	31.6	56.7
Line B	74.0	29.6	40.0
Line C	52.3	15.8	30.2
Line F	80.9	11.8	14.6

(d) Layer 13

	Total Length (km)	Observed with Gradient (km)	Fraction with Gradient (%)
SAF	107.5	69	64.2
Line E	75.9	38.5	50.7
Line B	80.7	38.5	47.7
Line D	67.1	21.7	32.3
Line C	64.6	19.7	30.5
Line F	97.6	19.7	20.2

(e) Layer 16

	Total Length (km)	Observed with Gradient (km)	Fraction with Gradient (%)
Line E	86.8	39.4	45.4
SAF	128.7	57.2	44.4
Line F	107.5	42.4	39.4
Line C	72.5	27.6	38.1
Line D	69.5	22.7	32.6
Line B	79.9	25.6	32.1

(f) Layer 19

	Total Length (km)	Observed with Gradient (km)	Fraction with Gradient (%)
Line F	118.3	48.3	40.8
Line C	84.3	33.5	39.8
Line E	93.2	35.5	38.1
SAF	138.0	52.3	37.9
Line B	79.9	26.6	33.3
Line D	81.9	21.7	26.5

Table 6. Continued

(g) All Layers

	Total Length (km)	Observed with Gradient (km)	Fraction with Gradient (%)
SAF	610.9	300.9	49.3
Line E	424.3	186.4	43.9
Line B	427.0	171.6	40.2
Line D	362.4	130.7	36.1
Line F	536.9	177.5	33.1
Line C	358.0	107.4	30.0

Lines are defined in Figure 14. The lines are presented in the order at descending fractions.

root. The location of the low coincides with the highest elevations in that part of the Coast Range, the Loma Prieta highlands (compare Figures 1 and 15). Burgmann et al. (1994) evaluated fission track ages and other data of the rocks in the Santa Cruz Mountains. They infer that the region northeast of the SAF has been uplifted about 4 km during the last 4.6 m.y. (0.8 mm/yr), while southwest of the SAF, the region is characterized by slower uplift rates (0.1-0.4 mm/yr) and less total uplift. Therefore, a crustal root only on the northeast side of the SAF is not surprising. The inference of a crustal root in this area was also made by Zandt (1981).

Heat flow patterns weakly support this crustal root interpretation. In the vicinity of the Loma Prieta earthquake epicenter, heat flow is lower than in the most of the Coast Range which shows generally high heat flow (Lachenbruch and Sass, 1980, Figure 16). Sibson (1983) pointed out that the maximum focal depth of earthquakes along the San Andreas fault system is inversely correlated with surface heat flow. He suggests that the maximum depth of earthquakes coincides with the temperature-dependent transition from brittle failure in the seismogenic upper crust to aseismic, quasi-plastic flow in the lower crust. Although temperature alone is not the only factor controlling earthquake occurrence, the unusual depth of the Loma Prieta earthquake hypocenter may imply that the crust is thicker,

Figure 15. Topography map of the study area. Color bar is annotated in meters.
(Supplied by Robert Jachens, 1994, from USGS database.)

Figure 16. Available heat flow measurements in the central Coast Ranges. (Modified from Lachenbruch and Sass, 1980.)

hence having a lower temperature gradient, beneath the Loma Prieta highland.

To test this interpretation, one might examine the Bouguer gravity map (Figure 17) of the study area (graphical communication, R. Jachens, 1994, from USGS database). If the low is caused by the crustal thickening, then a low-gravity anomaly might be anticipated, but none is evident in Figure 17. To examine this possibility, I calculated the gravity anomaly caused by a simple model of a crustal root comparable to the one described above. The root is approximated by two, thin, concentric disks at Moho depth, the smaller disk below the larger. The combined volume of these disks equals that of a "root" of 3 km maximum thickness and with cosine cross-section wavelength 28 km. The gravity anomaly caused by each disk, measured above the center of it was obtained by Singh (1977):

$$g = 2\pi G\sigma\Delta z \left[1 - \frac{z}{\sqrt{z^2 + a^2}} \right]$$

where G is the gravitational constant, σ is the density (contrast), Δz is a thickness of the disk, z is the distance from the disk to the surface, and a is the radius of the disk. Assuming σ is typical of much of the U.S., 0.4 g/cm^3 (personal communication, R. Plouff, 1994), calculated sum of the gravity anomaly caused by the two disks mentioned above is 3.0 mgal at its maximum (above the center of the disks). Figure 18 shows the attenuation of this anomaly with distance

Figure 17. Bouguer gravity anomaly map of the study area. Color bar is annotated in milligals. (Supplied by Robert Jachens, 1994, from USGS database.)

Figure 18. Estimated gravity anomaly from a crustal root in layer 10. Root is approximated by two thin disks centered at 0 km with 0.4 g/cm^3 density contrast (see text).

from this center. Such a gravity anomaly is too weak and too broad to be detected within the complex gravity field of Figure 17, presumably caused by shallower structures. Therefore the crustal root interpretation is not excluded by the gravity anomaly shown in Figure 17, especially the crust is not in isostatic equilibrium.

The second possibility, shear heating, also cannot be excluded at this stage. Although Zoback et al. (1987) pointed out that the heat flow expected from mechanical shear in the upper crust along the SAF is not detected, which suggests that the shear strength of the SAF is extremely low, their observation does not exclude the possibility of shear heating at the depth that I observe the low-velocity (near the Moho). There may be other reasons why such deep-source heat flow is not detected.

Lachenbruch et al. (1976) modeled heat conduction from a simple "magmatic" heat source. Figure 19 shows the fractional increase in surface heat flow Q above the center of a constant disk-shaped source of strength Q_0 and radius R at depth a . The variable l is a characteristic distance from the heat source to a plane where the temperature perturbation will be appreciable after the passage of time t :

$$l(t) = \sqrt{4\alpha t}$$

Assuming (Lachenbruch et al., 1976) that α , the thermal diffusivity, is $0.008 \text{ cm}^2/\text{sec}$ ($25 \text{ km}^2/\text{m.y.}$), l will be 35 km when $t = 12.3 \text{ m.y.}$ That is, it takes about 12 m.y. for the

Figure 19. Heat Flow conduction model for a magmatic source (Lachenbruch et al., 1976). Variable a is the depth to the source (~35 km), R is the radius of the disk-shaped source (~19 km), Q_0 is the heat flow from the source disk, and Q is the resulting heat flow measured at the surface, at the origin. The abscissa represents time in terms of a "conduction length", $(l(t) = \sqrt{4\alpha t})$, the distance from source Q_0 at which a heat anomaly is "significant" after t m.y. for thermal diffusivity α . In this case, $t \approx 8$ m.y. so $l \approx 28$ km (i.e. $l/a \approx 0.8$). Also R/a is about 0.5, yielding 0.03 Q_0 at the surface.

heat flow anomaly from layer 10 (at ~35 km depth) to reach the surface whereas faulting has been active there for only about 8 m.y. (Atwater, 1989). Therefore, any surface heat flow anomaly would still be far from the equilibrium state.

The long axis of the low-velocity body is ~38 km and the distance from the surface to layer 10 is 35 km. Therefore in Figure 19 $R/a = 0.5$ corresponds to the case of the low-velocity body in layer 10. Reading the graph in Figure 19, only 3% of Q_0 would have reached the surface if heating began at 8 Ma. Figure 20 shows the heat flow measured at sites shown in Figure 16. The deviations from the mean are typically 0.25 HFU or 10.5 mW/m². This "noise" and paucity of heat flow measurements in the vicinity of Loma Prieta highlands would make it difficult to detect a heat flow anomaly from layer 10 shear heating. The long time needed for this heat to reach the surface (>12 m.y.) sharply reduces the chance of observing an anomaly. Therefore, shear heating is a plausible candidate for this low-velocity feature, while yielding no observable heat flow perturbation at the surface.

It is not clear why low-velocity caused by shear heating would not be observed except beneath the Loma Prieta highlands. Since heat generation is strongly dependent on rock type and other poorly known facts (personal communication, W. Thatcher, 1994), shear heating is laterally discontinuous due to either variations in rock strength or strain rate along the fault. On the other hand, the crustal

Figure 20. Heat flow measured in the southern San Francisco Bay (data from Figure 16). (From Lachenbruch and Sass, 1980.)

root interpretation is not surprising. Therefore, I prefer the crustal root interpretation.

Low-Velocity Offset in the Upper Mantle

The two low-velocity features observed in layers 13 through 19 may be caused at least partly by smearing from the shallower low-velocity body of layer 10—a resolution artifact. To examine this possibility, columns of resolution matrix for the blocks belonging to M1 and M2 were calculated. The results indicate that such contributions are the second order.

For M1, the sum of the elements of resolution columns corresponding to the low-velocity body in layer 10 is -0.027 , and for M2, it is 0.198 . This implies that M1 may actually be muted slightly by smearing effect from layer 10. M2 probably is reinforced by smearing from feature L but probably a minority fraction, perhaps 20%. The bulk of the anomalies in layers 13 through 19 appear to be generated by features mainly from 45 to 70 km depth. Two cross sectional plots of resolution for a block that belongs to L are also shown in Appendix E. The diagonal element of this block is 0.340 , while its off-diagonal elements in layer 13 are only about 10% (0.035) as large.

Whatever their origin is, the M1-M2 separation is in a right lateral sense beneath the surface trace of the SAF, suggesting an important result for tectonics. Suppose that

this offset was caused by a shearing locus beneath the SAF. Then the limited amount of offset (~45 km) appears too small, that is, less than surface displacements along the SAF. Figure 21 shows that offset of various geological units of known age along the SAF (Dickinson et al., 1972). A more recent study by Sedlock and Hamilton (1991) shows a similar displacement history for the SAF, though with a more abrupt rate change at about 6 Ma. Both studies use geologic markers distributed mainly in the central and southern portion of the SAF. They lack data in the San Francisco Bay area per se, where the SAF formed later than the southern portion of the SAF (and shear may be distributed between the Hayward fault and SAF). However, an accelerating trend of displacement along the SAF is well documented in both studies, though with large uncertainties in constraining data. From the work of Atwater (1989), Furlong et al. (1989) inferred that the Mendocino Triple Junction passed the southern San Francisco Bay at about 8 Ma. The offset since 8 Ma is 225 km (Figure 21), much greater than the observed M1-M2 offset.

Current geodetic motions also imply faster shear across the SAF than the M1-M2 offset supports. Using geodetic data, Lisowski and Savage (1992) measured a benchmark velocity field for the San Francisco Bay area. Figure 22 shows their result while Figure 23 shows the N33°W velocity components of this field near the line marked "Figure 3" in Figure 22. They found a very broad shear zone that is approximately 85

Figure 21. Plot of offset versus time for various geologic units along the San Andreas fault. (Modified from Dickinson et al., 1972.) Boxes enclose age span inferred and uncertainty in the amount of apparent offset: 1, lower Pliocene rocks (their estimate); 2, uppermost Miocene (Delmotian?) rocks (their inference from Pinnacles-Parkfield-Neenach felsite relations noted by Huffman 1970; 3, lower upper Miocene (Mohnian?) rocks (after Fletcher, 1967; and Huffman, 1970); 4, middle Miocene rocks (their interpretation after Addicott, 1968); 5, lower Miocene volcanic rocks dated at 21.5 and 23.5 m.y. old (circles indicate most likely offset inferred by Huffman, 1970). Dashed line is Dickinson et al.'s inference. Vertical lines indicate initiation of faulting along the San Andreas fault (8-6 Ma) and along the Hayward-Calaveras fault system (3.6-2.67 Ma). Arrows spanned by the irregular stippled band show the amount of offset between the M1 and M2 low-velocity features (Figure 10 (e)).

Figure 22. Benchmark velocity field in the San Francisco Bay Area reported by Lisowski and Savage (1992). Measurements were by trilateration, Very Long Baseline Interferometry (VLBI), and Global Positioning System (GPS) surveys spanning 15 years. VLBI stations are marked with solid squares labeled PR (Point Reyes), P (Presidio) and FOS (Fort Ord South). Open squares show GPS stations.

Figure 23. N33°W component of the benchmark velocity field along the "Fig. 3" line of Figure 22. (From Lisowski and Savage, 1992.) The error bars represent one standard deviation on either side of the plotted point. Solid line is the profile predicted by a dislocation model including the San Andreas, Hayward, and Calaveras faults, plus fault creep. The dashed line is the same model but without fault creep.

km wide across the Loma Prieta highlands (and over 100 km wide further north). The distributed shear is $\sim 35\text{mm/yr}$ in this region of which 19mm/yr is assigned to the SAF by those authors. At this rate, the displacement between M1 and M2 would accumulate in about 2.4 m.y., if the mantle is shearing at the same rate as the crust. These considerations suggest that if the M1-M2 offset was created by the SAF, then it was not created throughout the history of the SAF, but only part of its history, stopping at some time in the past.

Coupling thermal and rheologic properties of the rocks in the Loma Prieta region with the tectonic setting, Furlong and Langston (1990) proposed a geodynamic model of the Loma Prieta earthquake. They suggest that the deeper lithospheric boundary between the Pacific and the North American plates in the San Francisco Bay area is now offset to the east of the SAF and underlies the Hayward-Calaveras fault system. Figure 24 shows the differential movement beneath the San Francisco Bay area that Furlong et al. (1989) modeled. This three dimensional geometry requires a mid-crustal detachment surface beneath the Bay area, extending from the SAF on the west to the region below the Hayward-Calaveras fault on the east. In the Loma Prieta region, the upper crust of the Pacific plate moves in a more westerly direction (i.e., parallel to the SAF) than gross plate motions while the mantle maintains the gross plate motion parallel to the Hayward-Calaveras fault. The discrepancy in motion

Figure 24. Cross section and relative plate velocities across the Pacific-North American plate boundary in the southern San Francisco Bay Area (From Furlong et al., 1989.). Plate boundary configurations are shown in top frame with creeping faults shown as solid lines and locked fault as dashed lines. Hatched area in top frame shows the location of Furlong et al.'s proposed detachment fault. Results of a displacement model are shown by a solid line in the bottom frame; asterisks are observed displacements reported by Prescott et al. (1981).

directions between the crust and mantle of the Pacific plate causes strain accumulation in the lower crust in the vicinity of the Loma Prieta, accumulating that strain by ductile deformation. The lithologic and thermal structure of this region suggests that only the lowermost Pacific crust can deform in a ductile fashion at such strain rates. The base of the brittle crust is the locus of maximum strain-energy accumulation. Therefore, elastic strain accumulation driven through the decoupling of the crust and mantle of the Pacific plate was the primary source of energy for the Loma Prieta earthquake. This detachment fault, connecting the SAF and Calaveras fault at depth, has not been confirmed by the local earthquake tomography or other geophysical methods. Only Brocher et al. (1994) imaged a prominent lower crustal reflector at ~15 km depth beneath the San Francisco Bay Area, which they interpret as an evidence of the detachment fault. Zandt (1981) also argued for the possibility of differential crustal/mantle displacement in central California in interpreting his teleseismic tomography model. He infers that volcanic fields at the surface have been displaced from associated low-velocity bodies (inferred melt sources) in the upper mantle. If this detachment fault exists, then the SAF needs not be active in the upper mantle, and the offset of M1 and M2 could have been created before the shear locus jumped to beneath the Calaveras fault. Since the Calaveras fault is estimated to have formed 2.67-3.6 Ma (Page, 1982), the M1-M2

offset probably was created before then. Another study (Graham and Dickinson, 1978) suggests that the differential movement along the SAF has begun by about 6 Ma, and possibly earlier—6 Ma is the latest plausible inception time for shear along the SAF proper. The triple junction passed this way at 8 Ma, which I take to be the earliest plausible inception time for the SAF. Surface slip along the SAF near Loma Prieta beginning at 6–8 Ma and ending at 2.67–3.6 Ma falls between about 60 and 120 km (Figure 21), compared to my observed 45 km offset between M1 and M2. Hence, the M1–M2 offset is near the minimum of this range (corresponding to 6 Ma through 3.6 Ma). Although these estimates have large uncertainties, it is possible that the M1–M2 offset preserves mantle shear displacement beneath the former SAF during the shortest plausible residence of the shear locus beneath the SAF—6 Ma to 3.6 Ma.

An alternate interpretation of the M1–M2 offset is that mantle shear has always been broadly distributed, not constrained to any single narrow locus. Such a pattern would be similar to modern surface displacements reported by Lisowski and Savage (1992). If broad shear predominates, the portion of the shear falling between M1 and M2 (separated by about 17 km center-to-center in the northeast-southwest direction) is about 20% of the total (17 km of an 85 km-wide shear zone in the San Francisco Bay area). The shear for the whole of the Coast Ranges since 8 Ma must fall between the

total plate motion (370 km; Atwater and Molnar, 1973) and the amount of slip along the SAF proper during that interval (225 km; Figure 21). Thus, northwest-southeast displacement between M1 and M2 would fall between 74 km and 46 km. This range is as good a match to the M1-M2 offset as is the narrow, jumping shear zone model described first.

Hence, both narrow and broad shear-zone models are plausible, as are intermediate models with simultaneous shear on several, parallel, narrow shear loci. The single- and multiple-locus models are better at predicting the shape of the M1-M2 anomaly, which is discontinuous. A fully distributed, broad shear model would produce an elongated but contiguous feature, so this end-member model is not as plausible as the single- and multiple-locus models. What is definite, if M1 and M2 once belonged together and predate 8 Ma, is that the mantle shearing was not confined for the whole period to a single, narrow locus beneath the SAF.

CONCLUSIONS

Three-dimensional inversion of teleseismic *P* wave travel-time from the Loma Prieta region provides images of lithospheric structure beneath the area. Upper crustal structure shows good correlation with geology and local-earthquake tomography results. Near the Moho, a large low-velocity body is observed near the intersection of the SAF and Sargent faults, which may be due to a crustal root of 2-5 km thick, or to laterally discontinuous shear heating. In the upper mantle, two low-velocity features show a right lateral offset of ~45 km along the surface trace of the SAF. If this offset was created by a narrow mantle shear locus beneath the SAF, its limited amount of offset suggests that the SAF was active at this depth during only a portion of the surface fault's history, from 6-8 Ma until ~4 Ma, when the shear locus may have jumped to beneath the Calaveras fault. This interpretation favors the existence of the lower-crustal detachment fault bridging the gap between the SAF and Calaveras fault, as proposed by Furlong et al. (1989). Alternatively, a persistent broad shear zone or multiple, narrow shear loci could have limited the mantle offset.

FUTURE STUDIES

Both the unstripped and stripped models presented here are useful for determining the deep structure in this region. A model which is derived by merging my data with local-earthquake tomography results and other studies would be the most definitive. "Receiver functions", if they can be computed from these short-period instruments, could help find accreted slab fragments and possibly the detachment fault. These should be included in forward modeling and in the starting model for inverse modeling. Reflection/refraction seismology results can similarly be added to tomography data set. Teleseismic anisotropy studies may find the extant shear strain pattern to help differentiate between narrow-locus shearing and alternate models. For complicated structures such as at the San Andreas and Sargent faults intersection, NeHT tomography will be useful in sorting out structure within the seismogenic upper crust.

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Appendices

Appendix A. Residuals relative to the weighted mean residuals at each station. Each ray is plotted as a function of ray back-azimuth and slowness. The bounding circle shows slowness of 10 s/degree.

Appendix B. How to Read Appendices C and D
 (Also see explanation in the text on page 50 and 86)

Appendix B. (Continued)

Appendix C

Figure C. 1(a)

Figure C. 1(b)

Figure C.1(c)

Figure C. 1 (d)

Figure C.2(a)

Figure C. 2(b)

Figure C. 2(c)

Figure C. 2(d)

Figure C. 3(a)

Figure C. 3(b)

Figure C. 3(c)

Figure C. 3(d)

Figure C. 4(a)

Figure C. 4(b)

Figure C. 4(c)

Figure C. 4(d)

Figure C. 5(a)

Figure C. 5(b)

Figure C. 5(c)

Figure C. 5(d)

Figure C. 6(a)

Figure C. 6(b)

Figure C. 6(c)

Figure C. 6(d)

Figure C. 7(a)

Figure C. 7(b)

Figure C. 7(c)

Figure C. 7(d)

Figure C. 8(a)

Figure C. 8(b)

Figure C. 8(c)

Figure C. 8(d)

Figure C. 9(a)

Figure C. 9(b)

Figure C. 9(c)

Figure C. 9(d)

Figure C. 10(a)

Figure C. 10(b)

Figure C. 10(c)

Figure C. 10(d)

Figure C. 11(a)

Figure C. 11(b)

Figure C. 11(c)

Figure C. 11(d)

Figure C. 12(a)

Figure C. 12(b)

Figure C. 12(c)

Figure C. 13(a)

Figure C. 13(b)

Figure C. 13(c)

Figure C. 13(d)

LP90-D/T*/U Volume-Metric Equivalent Diagonal Elements

Figure C. 14

LP90-D/T*/U Volume-Metric Equivalent Diagonal Elements

Figure C. 14 (Continued)

Appendix D

Figure D. 1(a)

Figure D. 1(b)

Figure D. 1(c)

Figure D. 1(d)

Figure D. 2(a)

Figure D. 2(b)

Figure D. 2(c)

Figure D. 2(d)

Figure D. 3(b)

Figure D. 3(c)

Figure D. 3(d)

Figure D. 4(a)

Figure D. 4(b)

Figure D. 4(c)

Figure D. 4(d)

Figure D. 5(a)

Figure D. 5(b)

Figure D. 5(c)

Figure D. 5(d)

Figure D. 6(a)

Figure D. 6(b)

Figure D. 6(c)

Figure D. 6(d)

Figure D. 7(a)

Figure D. 7(b)

Figure D. 7(c)

Figure D. 7(d)

Figure D. 8(a)

Figure D. 8(b)

Figure D. 8(c)

Figure D. 8(d)

Figure D. 9(a)

Figure D. 9(b)

Figure D. 9(c)

Figure D. 9(d)

Figure D. 10(a)

Figure D. 10(b)

Figure D. 10(c)

Figure D. 10(d)

Figure D. 11(a)

Figure D. 11(b)

Figure D. 11(c)

Figure D. 11(d)

Figure D. 12(a)

Figure D. 12(b)

Figure D. 12(c)

Figure D. 12(d)

Figure D. 13(a)

Figure D. 13(b)

Figure D. 13(c)

Figure D. 13(d)

Figure D. 14(a)

Figure D. 14(b)

Figure D. 14(c)

Figure D. 14(d)

Figure D. 15(a)

Figure D. 15(b)

Figure D. 15(c)

Figure D. 15(d)

LP90-R/T*/U Volume-Metric Equivalent Diagonal Elements

Figure D. 16

LP90-R/T*/U Volume-Metric Equivalent Diagonal Elements

Figure D. 16 (Continued)

Appendix E. Resolution elements of the blocks belonging to the low-velocity body in layer 10.

Contour interval = 0.025

Appendix E (Continued)