Seismic structure of the upper mantle in a central Pacific corridor

James B. Gaherty and Thomas H. Jordan
Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge

Lind S. Gee
Seismographic Station, University of California, Berkeley

Abstract. The seismic structure of the Tonga–Hawaii corridor has been investigated by combining two data sets: Revenaugh and Jordan's reflectivity profile from ScS reverberations, which provides travel times to and impedance contrasts across the major mantle discontinuities, and 1500 new observations of frequency-dependent phase delays for the three-component S, SS, and SSS body waves and the $R_1$ and $G_1$ surface waves, which constrain the velocity structure within this layered framework. The shear waves turning in the upper mantle showed significant splitting of the $SI$ and $SV$ components, indicative of shallow polarization anisotropy. The data set was inverted in conjunction with attenuation and mineralogical constraints to obtain a complete spherically symmetric, radially anisotropic structure. The final model, PA5, is characterized by a high-velocity, anisotropic lid, bounded at 68 km depth by a large (negative) G discontinuity; a low-velocity, anisotropic layer below G, extending to a small L discontinuity at 166 km; an isotropic, steep-gradient region between 166 km and 415 km; and transition-zone discontinuities at 415, 507, and 651 km. The depth of the radial anisotropy in PA5 is shallower than in most previous models based on surface waves and higher modes. The average value of radial shear anisotropy in the lid, +3.7%, is consistent with the magnitude expected from the spreading-controlled models of olivine orientation, while anisotropy in the low-velocity zone, which is required by our data set, could be induced either by paleostrains that took place near the ridge crest or by shearing in the asthenosphere as a result of present-day plate motions. On the basis of recent work by Hirth and Kohlstedt, we suggest that the G discontinuity is caused by a rapid increase in the water content of mantle minerals with depth, marking the fossilized lower boundary of the melt separation zone active during crust formation. The high-gradient zone between 200 and 400 km is a characteristic feature of convecting oceanic upper mantle and is probably controlled by a steady decrease in the homologous temperature over this depth interval. The average shear-velocity gradient in the transition zone is lower than in most previous seismic models, in better agreement with the predictions for a pyrolitic composition.

Introduction

From a seismological perspective, the Pacific Ocean basin is the best natural laboratory for studying the structure of the oceanic mantle because it is nearly surrounded by active earthquake source regions, including deep-focus zones, that frequently illuminate its interior with seismic waves. It has been the testbed for many innovative studies in structural seismology, including seminal work on the three-dimensional (3-D) variability [e.g., Leeds et al., 1974; Suetsugu and Nakanishi, 1987], anisotropy [e.g., Forsyth, 1975a; Regan and Anderson, 1984; Farra and Vinnik, 1994], and discontinuity structure [e.g., Revenaugh and Jordan, 1987, 1989; Vidale and Benz, 1992] of the upper mantle. Most recently, global and regional 3-D tomography has begun to provide more detailed maps of the lateral heterogeneity in the Pacific mantle [e.g., Woodhouse and Dziewonski, 1984; Montagner and Tanimoto, 1991; Zhang and Tanimoto, 1993; Su et al., 1994]. While such tomographic studies have yielded useful images of the major geographic features, their data sets and model parameterizations have thus far suffered from inadequate resolution of vertical structure, especially with regard to the location and amplitudes of high-gradient regions (usually approximated as discontinuities) and the distribution of seismic anisotropy. These deficiencies have been particularly acute in the upper mantle, where typical global 3-D models display far less variability than observed among regional 1-D models [Nolet et al., 1994]. Discontinuities and anisotropy are structural features that have considerable diagnostic value in the continuing assessment of the upper mantle's properties and processes, including its composition, phase, temperature, and deformation history [e.g., Iess, 1964; Ringwood, 1975; Bass and Anderson, 1984; Weidmer, 1985; Revenaugh and Jordan, 1991a,b,c; Karato, 1992].

In this study, we use a novel and powerful combination of data to investigate the radial anisotropic structure in a narrow corridor from the Tonga-Fiji seismic zone to seismographs on the island of Oahu, Hawaii (Figure 1). We start with the results of Revenaugh and Jordan [1991c], who stacked ScS reverberations to obtain whole-mantle reflectivity profiles for this and other corridors in the southwest Pacific and Australasia. For most individual corridors, including Tonga-Hawaii, they were able to detect, locate, and measure the amplitudes of all mantle discontinuities with reflectivity peaks greater than about 1%. We use this precise, layered framework of vertical travel times and impedance contrasts in our construction of a regional upper...
mantle structure. This framework is tied together by a new data set derived from the Tonga-Hawaii seismograms: 1500 frequency-dependent travel times measured from direct surface waves ($R_f$, $G_s$) and body phases ($S$), as well as from a variety of wave groups whose ray-theoretical decompositions contain the surface-reflected shear waves (e.g., $SS$, $sSS$, $SSS$).

At the epicentral distances and frequencies used in this study ($\Delta = 39^\circ - 58^\circ$; $f = \omega/2\pi = 10 - 45$ millihertz (mHz)), the latter arrivals are essentially guided waves, arising from the complex interference of multiple refractions, reflections, and conversions from upper-mantle discontinuities. We account for these complexities in our waveform modeling, which is done by complete mode summation, and in our phase-delay inversions, which are based on Gee and Jordan's [1992] theory of generalized seismological data functionals (GSDF). In the past, seismologists have typically applied different analysis techniques to the recordings of low-frequency guided waves on different seismographic components: ray-theoretical formulations in the case of transverse components of ground motion, which involve only horizontally polarized ($SH$) shear waves [e.g., Grand and Helmberger, 1984a, b], and mode-theoretical formulations for shear waves on the vertical and radial components, where the $P-SV$ interactions are more complex [e.g., Nolet, 1975; Lerner-Lam and Jordan, 1983, 1987; Zietzsch and Nolet, 1994]. (An exception is the higher-mode study of Cara and Lévéque [1988], who derived dispersion curves for Pacific-crossing paths from both $P-SV$ and $S-H$ wave trains. However, this frequency-wavenumber technique requires a large-aperture array for spatial filtering, severely limiting its applicability.) In the GSDF formulation of the structural inverse problem, the phase-delay spectra of all selected wave groups are measured by applying the same types of time- and frequency-localizing operations to the cross-correlations between synthetic wave groups (isolation filters) and the data seismograms. The partial derivatives of each phase delay are calculated with respect to all model parameters, including discontinuity depths, density, and anisotropic elastic parameters by formulæ that account for the finite-bandwidth wave propagation and interference effects. GSDF thus provides a unified methodology for the analysis of complex wave groups on all three seismographic components.

Applying GSDF to the Tonga-Hawaii seismograms, we demonstrate that the $SH$ and $P-SV$ components of the surface and guided waves are significantly split by anisotropy situated in the uppermost mantle. This anisotropy is most likely related to the lattice-preferred orientation (LPO) of olivine in upper mantle peridotites caused by the horizontal shearing during plate formation and translation, which induces azimuthal asymmetry in the wave speeds [Hess, 1964; Nicolas and Christensen, 1987]. The observations presented here cannot resolve azimuthal variability, however, and we show that they can be satisfied by a transversely isotropic (radially anisotropic) model that incorporates no azimuthal variations. While such a model restricts our ability to address azimuthal anisotropy and its geodynamical significance, it does allow us to assess the distribution of anisotropy with depth and its relationship to discontinuities and other radial features of the mantle. The accuracy and uniqueness of the mapping between the actual anisotropy and our radial average using data from a single seismic corridor depends on the details of the olivine alignment, which are unknown. For example, if the LPO is largely horizontal but varies with respect to the propagation direction along the path, then the inferred radial model approximates an
azimuthal average of the local anisotropy [Estey and Douglas, 1986; Jordan and Gaherty, 1995]. If the LPO is coherently aligned at an oblique angle to the propagation direction, the resulting path-averaged anisotropy will be similar to the azimuthal average, but if coherent alignment coincides within ±20° of the propagation direction, the magnitude of the anisotropy obtained assuming transverse isotropy will underestimate the azimuthal average [Lévêque and Cara, 1983; Maupin, 1985]. As discussed below, our data, as well as other geological and seismological observations, indicate that the condition of parallel alignment is not likely to apply to the geometry of the central Pacific corridor investigated here.

We therefore invert the phase delays of the surface, guided, and body waves jointly with the ScS reverberation data for a path-averaged, layered, radially anisotropic model via an iterative, linearized scheme that perturbs the wave speeds, density, and discontinuity depths for each layer. As an aid to finding a geophysically plausible structure, we condition the inversion with a prior probability distribution on the model space that biases the estimate in favor of constraints derived from laboratory measurements and other seismic experiments relevant to the oceanic upper mantle. We also test a set of hypotheses that are alternatives to those imbedded in the model prior, with a specific focus on assessing the depth extent of seismic anisotropy in this part of the Pacific Ocean.

### Tonga-Hawaii Corridor

The paths between earthquakes in the Tonga-Fiji seismic zone and the seismographic stations KIP and HON, both on Oahu, sample a relatively homogeneous corridor of old Pacific lithosphere (Figure 1). The crustal ages along this profile fall in a narrow range, 100-125 Ma [Mueller et al., 1993], and its bathymetry is relatively uniform, with an average ocean depth of 5.1 km. The sediment thickness averages about 200 m [Ludwig and Houtz, 1979]. The corridor crosses the Line Islands and the eastern margin of the Darwin Rise, an area of prolific Cretaceous volcanism [Menard, 1984], but lies to the west of the disturbed eastern margin of the Darwin Rise, an area of prolific Cretaceous volcanism [Menard, 1984], and Houtz, 1979]. The corridor crosses the Line Islands and the eastern margin of the Darwin Rise, an area of prolific Cretaceous volcanism [Menard, 1984], but lies to the west of the disturbed upper mantle associated with the Pacific Superswell [McNutt and Fischer, 1987].

This relative homogeneity is reflected in seismic tomographic models [Zhang and Tanimoto, 1993; Su et al., 1994; Ekström and Dziewonski, 1995]. Lateral velocity variations imaged in such models are small along the corridor, with higher shear velocities in the lid and lower shear velocities below the lid relative to the preliminary reference Earth model (PREM) of Dziewonski and Anderson [1981]. The corridor-averaged lower mantle velocity is slow relative to PREM, and the root-mean-square (rms) variability increases substantially in region D" [Su et al., 1994; Ekström and Dziewonski, 1995; Garnero and Helmberger, 1996].

The velocity of the Pacific plate in the hot-spot reference frame is roughly perpendicular to the corridor at a rate of 10.9 mm/yr [Gripp and Gordon, 1990]. However, the exact history of plate formation, of interest to us because the high strains associated with seafloor spreading largely determine the anisotropy of shallow mantle peridotites [Ribe, 1989, 1992], is difficult to discern, owing to the absence of magnetic reversals during the Cretaceous quiet period (~80-110 Ma) when much of the crust in this region was formed. The continuity of the major fracture zones across the Line Islands implies that the seafloor within the northern half of the corridor was formed at the Pacific-Farallon ridge in a geometry consistent with that preserved along the 75-Ma anomaly [Menard, 1967; Joseph et al., 1987]. South of the equator, the bend in the 125-Ma contour and the presence of the rift-associated Manihiki Plateau imply a complex history of triple-junction spreading involving an ancient Phoenix plate [Joseph et al., 1987]. The seafloor along the southern half of the corridor may have been formed on a NNW-SSE trending Pacific-Farallon or Phoenix-Farallon ridge, or along a WSW-ENE trending Pacific-Phoenix ridge. Either possibility implies that spreading-related fossil anisotropy is oblique to the corridor along much of its length.

Recent surface-wave and body-wave studies in this region have attempted to relate both the tectonic history of the plate, and the underlying mantle flow, to observed seismic anisotropy. Surface-wave observations include differential phase-velocity anomalies of Love and Rayleigh waves (polarization anisotropy) [Cara and Lévêque, 1988; Nishimura and Forsyth, 1989], azimuthal velocity variations of (primarily) Rayleigh waves [Nishimura and Forsyth, 1989; Montagner and Tanimoto, 1991], and surface-wave coupling [Yu and Park, 1994]. The first two indicate that the depth distribution and long-wavelength azimuthal variations of anisotropy are generally consistent with strain induced by both past and present plate motion, but the details of such structure are poorly resolved. The last is most sensitive to sharp gradients in anisotropy, and its presence implies that the anisotropy is spatially variable. Shear-wave splitting observed in body waves [Farra and Vinnik, 1994; Su and Park, 1994] provides localized estimates of azimuthal anisotropy, and they are also spatially variable. These estimates are sometimes correlated with either past or present plate motion, but they often disagree, and the sampling density of these waves is too low to precisely map the orientation and depth distribution of the azimuthal component in this region. Overall, the observations imply that multiple mechanisms, with various orientations and length scales, give rise to seismic anisotropy in Pacific upper mantle. Along our corridor, the likely alignment directions vary from oblique to perpendicular to the propagation direction. The radial anisotropy derived in this study can therefore be interpreted as a path average of this azimuthal variability, which should be close to a complete azimuthal average of the local anisotropy along the path [Maupin, 1985].

### ScS Reverberations

Much of what is known about the details of mantle structure in this part of the Pacific has come from the study of SH-polarized ScS reverberations, which are very well excited at Oahu by dip-slip earthquakes in Tonga-Fiji. Reverberation phases are classified by their core-mantle boundary (CMB) reflection number, \( n = 1, 2, \ldots \), and their reverberation order, \( m = 0, 1, \ldots \), which is the number of times they reflect off one or more internal mantle discontinuities. We follow the conventions of Revenaugh and Jordan [1991b] in designating internal discontinuities above 400 km by a capital letter (M, Mohorovičić; H, Hales; G, Gutenberg; L, Lehmann) and transition-zone discontinuities by their nominal depths in kilometers, that is, 410, 520, 660, 710, 900. The upper mantle refers to the region between M and 660, whereas the lower mantle refers to the region between 660 and the CMB.

The zeroth-order reverberations are the primary ScS\(_n\) and sScS\(_n\) phases, which have long been used to study mantle attenuation [Press, 1956; Kovach and Anderson, 1964]. Sipkin and Jordan [1980b] applied a phase-equalization and stacking algorithm to 12 ScS\(_n\) phase pairs to estimate a path-averaged quality factor of \( Q_{ScS} \approx 141 \pm 16 \) for the Tonga-Hawaii corridor.
Revenaugh and Jordan [1987] expanded this data set to 26 ScS and sScS phase pairs, obtaining the discrete discontinuity model shown in interpretation panel, from Revenaugh and Jordan [1991c]. The amplitudes and travel times contrast across d, reverberation observations [Revenaugh and Jordan, 1987]. (c) times from the surface to d and the reflectivity or impedance discontinuity d yield precise estimates of the vertical SV travel discontinuities as individual seismic phases. When referenced to observations for the Tonga-Hawaii corridor.

Revenaugh and Jordan [1987] found the travel times of reverberations from the 660 discontinuity placed its apparent depth along the Tonga-Hawaii profile at 650 km, and an amplitude analysis combining the zeroth- and first-order reverberations yielded a reflectivity of \( R_{660} = 0.080 \pm 0.004 \), and quality factors of \( Q_{UM} = 82 \pm 18 \) and \( Q_{LM} = 231 \pm 60 \) for the upper mantle and lower mantle, respectively. We adopted these quality factors (which pertain to the 10-35 mHz band) as constraints on a simple forward modeling procedure to determine the attenuation model used in this study (Figure 2). They are lower than the global values of \( Q_{UM}^{PREM} = 133 \) and \( Q_{LM}^{PREM} = 312 \) calculated for the preliminary reference Earth model, which has an attenuation structure determined from low-frequency (3-10 mHz) free oscillations [Dziewonski and Anderson, 1981]. This discrepancy has been interpreted as a manifestation of the frequency dependence of mantle attenuation [Sipkin and Jordan, 1979].

Revenaugh and Jordan [1989, 1991a-d] developed a technique for stacking first-order ScS reverberations to obtain whole-mantle reflectivity profiles, which they applied to 18 seismic corridors. Their results for Tonga-Hawaii, reproduced in Figure 2 and Table 1, are used directly as data in our inversions. The reflectivity profile shows well-developed positive peaks from the 410 (4.6 \( \pm \) 1.0%) and 660 (7.8 \( \pm \) 1.0%) and a strong negative peak (5.5 \( \pm \) 2.0%) from a shallow G discontinuity; their apparent depths calculated from PREM and corrected for 3-D heterogeneity of Woodhouse and Dziewonski [1984] are 415, 653, and 59 km, respectively. The one-way differential travel time, \( t_{660} - t_{410} = 45.5 \pm 1.0 \) s, is slightly less than the global average (47 \( \pm \) 1.5 s), suggesting that the transition-zone temperatures in this region are slightly higher [Revenaugh and Jordan, 1991b]. The shallow depth of G, which defines the top of the shear-wave low-velocity zone (LVZ), is consistent with inferences from surface-wave studies that properly account for polarization anisotropy in the uppermost mantle [Regan and Anderson, 1984].

A secondary but robust feature in the reflectivity profile is the distinctive loop between the 410 and 660 peaks introduced by the 520 discontinuity [Revenaugh, 1989; Shearer, 1990], which implies a small impedance contrast (2.1%) at an apparent depth of about 500 km. Notably absent is evidence for an H discontinuity near 60 km, which may be obscured by G [Revenaugh and Jordan, 1991c], and an L discontinuity near 220 km, which shows up in Australia as the sharp base of an anisotropic mechanical boundary layer [Revenaugh and Jordan, 1991c; Gaherty and Jordan, 1995].

The ScS reverberation data also allow us to address further the question of along-path heterogeneity. Although the path was chosen to sample what we assume is a relatively homogeneous part of the oceanic upper mantle, Sipkin and Jordan [1980a] observed that the differential travel times between multiple-ScS

![Figure 2. (left) Left side of panel compares the observed reflectivity profile for the Tonga-Hawaii corridor (dots) with a synthetic profile (solid line) obtained from the discrete discontinuity model shown in interpretation panel, from Revenaugh and Jordan [1991c]. The amplitudes and travel times associated with these discontinuities are the observed values listed in Table 1. (right) Shear Q structure for the Tonga-Hawaii corridor. (a) Average mantle Q from multiple-ScS observations [Revenaugh and Jordan, 1987]. (b) \( Q_{UM} \) and \( Q_{LM} \) from ScS reverberation observations [Revenaugh and Jordan, 1987]. (c) \( Q \) structure assumed for the modeling in this paper, based on the values shown in column b and our own surface and body wave reflectivity profile for the Tonga-Hawaii corridor.]

### Table 1. ScS Reverberation Data and Model Values for the Tonga-Hawaii Corridor

<table>
<thead>
<tr>
<th>Discontinuity</th>
<th>Observed Data</th>
<th>Model PA2</th>
<th>Model PA5</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>( t_d ), s</td>
<td>( R_d ), %</td>
<td>( t_d ), s</td>
</tr>
<tr>
<td>M</td>
<td>1.7 ( \pm ) 0.5</td>
<td>12.0 ( \pm ) 5.0</td>
<td>2.3</td>
</tr>
<tr>
<td>G</td>
<td>12.0 ( \pm ) 1.0</td>
<td>55.5 ( \pm ) 2.0</td>
<td>21.8</td>
</tr>
<tr>
<td>L</td>
<td>----</td>
<td>0.0 ( \pm ) 1.0</td>
<td>41.1</td>
</tr>
<tr>
<td>410</td>
<td>91.1 ( \pm ) 1.0</td>
<td>46.0 ( \pm ) 1.0</td>
<td>89.7</td>
</tr>
<tr>
<td>520</td>
<td>1077.5 ( \pm ) 2.0</td>
<td>21.6 ( \pm ) 1.5</td>
<td>----</td>
</tr>
<tr>
<td>660</td>
<td>1366.5 ( \pm ) 1.0</td>
<td>7.8 ( \pm ) 1.0</td>
<td>1412</td>
</tr>
</tbody>
</table>

*Variable \( t_d \) is one-way, vertical SV travel time from the solid surface in seconds; \( R_d \) is the shear-wave impedance contrast at vertical incidence. Observations are from Revenaugh and Jordan [1991a, c].*
The initial Airy phase of the observed Rayleigh wave for the SS, sSS, and SSS. To minimize unmodeled source effects, we choose from 1983-1991, with hypocentral depths ranging from 10 to 663 km and epicentral distances from 39 ø to 58 ø (Figure 1). The horizontal-component seismograms were rotated into the radial and transverse directions, and all three components were limited our selection to Tonga-Fiji earthquakes of moderate size (Mw < 6.6) having well-determined Harvard centroid moment tensor (CMT) solutions [Dziewonski et al., 1981]; 55 were chosen from 1983-1991, with hypocentral depths ranging from 10 to 663 km and epicentral distances from 39 ø to 58 ø (Figure 1). The horizontal-component seismograms were rotated into the radial and transverse directions, and all three components were low-passed with a zero-phase filter having a corner at 45 mHz.

Evidence of polarization anisotropy can be seen directly on the seismograms. Figure 3 compares the vertical and transverse seismograms observed from two events with synthetic seismograms computed for the isotropic model PA2 [Lerner-Lam and Jordan, 1987]. (All synthetic seismograms used in this paper were calculated by complete mode summation to 50 mHz, assuming the Harvard CMT and low-pass filtered like the data.) The initial Airy phase of the observed Rayleigh wave for the shallow-focus event (Figure 3a) is reasonably well matched by PA2 (which was derived from fundamental and higher-mode Rayleigh waves propagating across the northwest Pacific), while the observed Love wave clearly travels faster than the synthetic. The magnitude of the misfit cannot be rectified by perturbing the isotropic structure within reasonable bounds. Similarly, the SSS wave group from a deep-focus event (Figure 3b) shows clear evidence of shear-wave splitting, with SSSv advanced by approximately 6 s relative to SSSp. Splitting of this magnitude is typical of upper-mantle guided waves observed on other paths [Gee and Jordan, 1988; Gaherty and Jordan, 1995] and is diagnostic of anisotropy in the uppermost mantle.

Further evidence of this anisotropy may be seen on the vertical-component seismogram for the shallow-focus event, as a small arrival at about 18 min that is not well modeled by PA2 (Figure 3a). It is possible that this waveform represents Love-to-Rayleigh coupling from laterally variable azimuthal anisotropy in the central Pacific, as discussed by Yu and Park [1994]. In this distance range, however, this arrival closely coincides with the theoretical arrival time of SSv, whose excitation in this case is very sensitive to source depth. By increasing the source depth from 10 to 20 km, we can generate SSv from the vertical-component seismogram for the shallow-focus event, as a diagnostic of anisotropy in the uppermost mantle.

Frequency-Dependent Travel Times From Surface and Turning Waves

Frequency-dependent travel times have been measured from three-component, long-period seismograms at HON and KIP for an extensive set of direct Rayleigh (R) and Love (G) surface waves, direct S body waves, and a series of wave groups containing the surface-reflected shear waves, which include the phases SS, sSS, and SSS. To minimize unmodeled source effects, we limited our selection to Tonga-Fiji earthquakes of moderate size (Mw < 6.6) having well-determined Harvard centroid moment tensor (CMT) solutions [Dziewonski et al., 1981]; 55 were chosen from 1983-1991, with hypocentral depths ranging from 10 to 663 km and epicentral distances from 39 ø to 58 ø (Figure 1). The horizontal-component seismograms were rotated into the radial and transverse directions, and all three components were low-passed with a zero-phase filter having a corner at 45 mHz.

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GSDF Analysis

This procedure for measuring and inverting frequency-dependent travel times has its roots in the residual-dispersion methods pioneered by Dziewonski et al. [1972] and Herrin and Geforth [1977] and has theoretical affinities to the wave-equation travel-time inversion of Luo and Schuster [1991]. GSDF processing comprises six steps: (1) Synthesizing a target wave group to create a time-limited isolation filter. Although any convenient method for seismogram synthesis will do (e.g., asymptotic ray methods), we employ Gee and Jordan's [1992] mode-theoretic formulation, which obtains the isolation filter by summing traveling modes weighted according to the group and phase velocities of the target arrival. This synthesis procedure is exceptionally well suited for upper-mantle surface and guided waves, because it requires no high-frequency approximation and includes all wave interactions. (2) Cross-correlating the isolation filter with both the observed seismogram and the complete
synthetic seismogram. (3) Windowing the two broadband cross-correlograms in the time domain. This time-localization operation reduces (but does not eliminate) the interference from other wave groups. (4) Filtering the windowed cross-correlograms in a discrete set of narrow frequency bands \( \{ \omega_i \pm \sigma_i \} \). We typically center these filters at 5-mHz intervals across the frequency range from 10 to 45 mHz, which yields up to eight spectral estimates per isolation filter. This frequency-localization operation produces waveforms that under appropriate (and always enforceable) conditions can be approximated as Gaussian wavelets. (5) Fitting the windowed and filtered correlograms in the time domain with two five-parameter Gaussian wavelets. One of the five parameters is the (known) value of the bandwidth, \( \sigma_i \); the other four can be written as a set of time-like quantities that include a phase delay, a group delay, and two equivalent amplitude parameters. On the synthetic cross-correlogram, nonzero values of these four parameters measure the interference between the isolation filter and other wave groups; subtracting these from the values estimated for the observed seismogram yields the interference-corrected data [Gee and Jordan, 1992, equations (9)-(12)]. (6) Correcting these data for windowing and filtering effects using equations (56)-(59) of Gee and Jordan [1992].

We limit our analysis to the differential phase delay, denoted \( \delta \tau_p(\omega_i) \), which is our definition of a frequency-dependent travel time. Among the four generalized seismological data functionals recovered by the GSDF procedure, this observable is the most robust and useful in structural inverse problems. In particular, it is relatively straightforward to compute the Fréchet kernel \( g_p(\omega_i) \) expressing its first-order sensitivity to a model perturbation \( \delta m \). For a spherically symmetric model,

\[
\delta \tau_p(\omega_i) = g_p(\omega_i) \cdot \delta m. \tag{3}
\]

The inner product in (3) includes a sum over the discontinuities in \( m \), as well as an integral over depth. Under a set of approximations valid for the isolation filters used here, the Fréchet kernels can be written as a simple sum of the kernels for individual traveling modes:

\[
g_p(\omega_i) = \sum_{n=1}^N c_n g_p^*(\omega_i). \tag{4}
\]

The coefficients \( \{ c_n \} \) depend on the isolation filter and the details of its interference with other wave groups on the seismograms, and they account for the excitation amplitude and phase of the source [Gee and Jordan, 1992, equations (73), (74), and (95)]. We have extended this formulation of the Fréchet kernels to include radially anisotropic, transversely isotropic media [Gaherty, 1995]. In this paper we utilize Dziewonski and Anderson's [1981] notation for the six medium parameters: mass density, \( \rho(z) \); the speeds of horizontally and vertically propagating \( P \) waves, \( v_{PH}(z) \) and \( v_{PV}(z) \); the speed of horizontally propagating, transversely polarized shear waves, \( v_{SH}(z) \); the speed of a shear wave propagating either horizontally with a vertical polarization or vertically with horizontal polarization (e.g., \( S_{cS} \) reverberations), \( v_{SV}(z) \); and a parameter that governs the variation of the wave speeds at oblique propagation angles, \( \theta(z) \).

Therefore \( \delta m \) comprises perturbations to these six functions of depth, as well as the perturbations to the depths of the internal mantle discontinuities.

Examples of GSDF Processing

Since this paper discusses the first full-fledged application of the GSDF methodology to a structural inverse problem, we illustrate some of its advantages with a few examples of waveform processing. Figure 4a displays the data, a full synthetic seismogram, and an isolation filter for a vertical-component \( SSS \) phase. The phase delays recovered by the GSDF procedure (Figure 4b) show that the \( SSS \) waveform is dispersed relative to the synthetic model; the low frequencies arrive approximately as predicted, whereas the higher frequencies arrive up to 8 s early. (A standard cross-correlation analysis yields only a single travel-time residual; in this case, approximately 6 s, which corresponds to the phase delay at \(-35\) mHz.) The Fréchet kernels relating these phase delays to model perturbations are also frequency dependent (Figure 4c). At low frequency, this \( SSS \) phase is sensitive primarily to the average \( SV \) velocity in the upper mantle, with two broad peaks near 100 and 300 km depth. At higher frequencies, the kernel becomes more oscillatory and more sharply peaked in the transition zone below 400 km depth, with the peak location eventually converging to its ray-theoretical bottoming depth of 480 km. Note that these phase delays do not depend on \( v_{SH} \). They are sensitive, however, to the other model parameters, whose kernels are not plotted here.

The full power of GSDF analysis is illustrated by two versions of its application to the same multimode oceanic Love wave from a shallow-focus earthquake (Figures 5 and 6). Love waves provide critical information about anisotropic structure, but the lid/LVZ structure in oceanic regions results in very strong interference between the fundamental and higher modes, making traditional (single-mode) dispersion measurements difficult, especially along short propagation paths [Thatcher and Brune, 1969; Forsyth, 1975b; Schue and Knopoff, 1977]. The isolation filter in Figure 5 (Filt1), obtained by integrating just the fundamental-mode branch, is a poor representation of the Love
We generally rejected summation, and used GSDF processing to extract 1497 earthquakes, generated their isolation filters by weighted mode.

Figure 5. Example of the ability of GSDF to account for interference of complex waveforms. (a) (top) Observed, (middle) full synthetic, and (bottom) isolation filter synthetics for transverse-component data from a shallow-focus (h = 10 km) event. The S wave arrives at -14.5 min, the multimode Love wave arrives at -18 min. The isolation filter (Filt1) is constructed by summing only the fundamental-mode branch, and thus it does not include higher mode contributions and is a poor match to the full synthetic Love wave. (b) Initial phase-delay estimates $\delta t = t$, and final phase delays $\delta t$, as a function of frequency. The $\delta t$ have been corrected by up to 6 s by removing the interference term $I_p$. (c) Fréchet derivative kernels for $v_{SH}$ (solid line) that are used to invert the $\delta t_p$ observations. The dashed line represents the kernels that would be used if the Love wave were assumed to be fundamental mode only (i.e., the interference corrections were not applied).

Data Summary

We selected 233 waveforms from the 55 Tonga-Fiji earthquakes, generated their isolation filters by weighted mode summation, and used GSDF processing to extract 1497 frequency-dependent travel times. We generally rejected waveforms which had low signal-noise ratio or were poorly matched by the synthetics (often indicative of errors in the source parameters or strong multipathing). All travel times were corrected for ellipticity, receiver, and source anomalies, which were calculated by averaging the tangential and vertical broadband S-wave delays for each event.

Figure 6. (a) Same data and full synthetic as Figure 5a, but with a new isolation filter (Filt2) constructed by summing all mantle modes with group velocities $4.34 \pm 0.2 \text{ km/s}$. This filter represents an excellent match to the full synthetic Love wave, and interference corrections are negligible. (b) Final phase delays $\delta t$ for both Filt2 and Filt1 from Figure 5b. (c) The $v_{SH}$ Fréchet derivative kernels for Filt2 and Filt1. The nearly identical behavior of the phase delays and kernels for these two filters demonstrates the robustness of the GSDF procedure.

The key features of the data set can nevertheless be recognized in Figure 7, and a feel for what they imply about mantle structure can be gleaned from examples of the Fréchet kernels for each of the eight wave types displayed in Figures 8 and 9. Differences between the $R_1$ and $G_1$ residuals are everywhere positive and reach 25 s near the low end of the frequency band. This is an example of the ubiquitous Love-Rayleigh (LR) discrepancy [Anderson, 1961; McEvilly, 1964; Anderson and Dziewonski, 1982]. It is indicative of polarization anisotropy of the normal type, that is, where $v_{SH} > v_S$ and $v_{PH} > v_P$. The Love waves are not strongly dispersed relative to PAZ, but their mean is about 17 s faster, requiring $v_{SH}$ in the lid and LVZ to be significantly
mantle than their high-frequency kernels, this frequency dependence limits the depth extent of the anisotropy. The S waves are more concentrated in the shallow part, owing to their greater depth of penetration, and it also decreases with frequency. Since the low-frequency kernels also require by the ScS reverberation data and GSDF theory for the 1497 frequency-dependent travel times. In the latter case we integrated the Fréchet kernels in (4) over layers to obtain the partial derivatives computed for \( m_k \). The partials were calculated using ray theory and plane-wave reflection coefficients for the 11 waves, which have high-frequency kernels that peak near their ray-theoretical turning depths (~1000 km), show little evidence of splitting or relative dispersion, requiring the anisotropy to be small in the transition zone and lower mantle.

Assuming an LPO origin of upper-mantle anisotropy, the observations imply an olivine a-axis alignment that is predominantly horizontal and not coherently parallel to the propagation path [Maupin, 1985; Cara and Le Berre, 1988], consistent with the inferences we have drawn from the geological and prior seismological data. This consistency further justifies using a transversely isotropic model for determining the depth distribution of anisotropy.

As emphasized by Anderson and Dziewonski [1982] and Regan and Anderson [1984], Rayleigh-wave travel times are sensitive to density, \( P \) velocities, and \( \eta \), and the kernels for the other \( SV \) phases in Figures 8 and 9 show a similar dependence. In particular, their times all show a negative sensitivity to \( v_{PV} \) and a positive sensitivity to \( \eta \). The sensitivity to \( v_{PV} \) is also positive, except near the surface. On the other hand, the SV times do not depend on \( v_{SH} \), whereas the \( SH \) times are (negatively) sensitive to \( v_{SV} \). Such behavior reinforces the importance of inverting for a fully radially anisotropic structure.

Figure 7. Summary of the frequency-dependent travel time data. Phase delays are categorized by phase type (surface wave, SSS, SS, or S), separated into tangential (SH, circles) or radial/vertical (SV, squares) observations, and averaged in each frequency band. Symbols represent the residuals relative to PA5 to these data. Dispersion relative to PA2 is indicated by a frequency-dependent trend, while evidence of anisotropy can be observed as a separation of the \( SH \) and \( SV \) observations for a given phase type: (a) Love and Rayleigh waves; (b) SSS and sSS waves; (c) SS waves; and (d) S waves. Error bars are determined by a weighted average of estimated a priori errors.

The inversion was initiated with a starting model \( m_0 \), chosen to be the isotropic PA2 structure of Lerner-Lam and Jordan [1987]. PA2 is parameterized in terms of a discrete set of layers with linear gradients bounded by sharp discontinuities corresponding to M, G, L, and the two major transition-zone discontinuities. We retained this basic parameterization throughout the inversion, except that we allowed a new discontinuity to develop in the transition zone in order to satisfy the ScS reverberation data for the 520-km discontinuity. Because an L discontinuity is not observed in the reflectivity profile for Tonga-Hawaii, we included \( R_L = 0 \pm 1 \% \) as a data point. A four-layer attenuation structure was constructed to match the ScS reverberation data and surface-wave amplitudes for this path, with layers separated by the G, L, and 660 discontinuities (Figure 2). The depths of these and all other discontinuities were allowed to vary during the inversions, but the four attenuation quality factors were held fixed. We accounted for the attenuative dispersion by assuming the layer \( Q \) values were frequency-independent [Liu et al., 1976], which is a good approximation across the frequency band of the observations [Sipkin and Jordan, 1979]. We calculated the elastic parameters in the models at a reference frequency of 35 mHz.

The inversion sequence was carried through three iterations. In the 4th iteration the difference between the true earth model \( m \) and the estimate \( m_{k-1} \) was assumed to satisfy the linearized system \( A_k \delta m + n = \delta d_{k-1} \), where \( \delta d_{k-1} \) is the N-dimensional data-residual vector and \( A_k \) is the \( N \times M \) matrix of partial derivatives computed for \( m_{k-1} \). The partials were calculated using ray theory and plane-wave reflection coefficients for the 11 ScS reverberation data and GSDF theory for the 1497 frequency-dependent travel times. In the latter case we integrated the Fréchet kernels in (4) over layers to obtain the partial derivatives of the data with respect to the layer slopes and intercepts. The data-error vector \( n \) was assumed to have zero mean, \( \langle n \rangle = 0 \), and a diagonal covariance matrix \( C_m = \langle nn^T \rangle = \text{diag}(\sigma_{n1}^2, \sigma_{n2}^2, \ldots, \sigma_{nN}^2) \). The standard deviations assigned to the ScS reverberation data depended primarily on the strength of the reflector, with \( \sigma_{n1} \) and \( \sigma_{n2} \) ranging from 0.005 and 1 s for the
Our prior knowledge of earth structure was specified by the $M$ vector $\mathbf{m}$ and the symmetric $M \times M$ matrix $C_{mn}$. Several different types of information and assumptions were combined in the construction of these quantities. For example, at six discrete depths between 250 and 720 km we constrained the density $\rho$ and bulk sound velocity $v = (v_p^2 - 4v_s^2/3)^{1/2}$ to satisfy the estimates obtained by Ita and Stixrude [1992] for a pyrolite mineralogy, assigning a standard error of $\pm 1\%$ to each estimate (Figure 10). These parameters, which are complementary to our shear-wave-dominated data set, can be inferred with reasonable precision from laboratory observations; moreover, the choice of a pyrolite composition is not particularly restrictive, since the $v(z)$ and $\rho(z)$ profiles for most competing mineralogical models (e.g., high-aluminum piclogite) differ from pyrolite by less than the assigned errors [Ita and Stixrude, 1992]. We also assumed that the jumps in both $v_p$ and $\rho$ were small ($\pm 0.5\%$) at the G and L discontinuities.

After each step, the model was evaluated by a complete remeasurement of the data using synthetic seismograms and isolation filters computed for the new model $\mathbf{m}_k = \mathbf{m}_{k-1} + \delta \mathbf{m}_k$. Our prior knowledge of earth structure was specified by the $M$ vector $\mathbf{m}$ and the symmetric $M \times M$ matrix $C_{mn}$. Several different types of information and assumptions were combined in the construction of these quantities. For example, at six discrete depths between 250 and 720 km we constrained the density $\rho$ and bulk sound velocity $v = (v_p^2 - 4v_s^2/3)^{1/2}$ to satisfy the estimates obtained by Ita and Stixrude [1992] for a pyrolite mineralogy, assigning a standard error of $\pm 1\%$ to each estimate (Figure 10). These parameters, which are complementary to our shear-wave-dominated data set, can be inferred with reasonable precision from laboratory observations; moreover, the choice of a pyrolite composition is not particularly restrictive, since the $v(z)$ and $\rho(z)$ profiles for most competing mineralogical models (e.g., high-aluminum piclogite) differ from pyrolite by less than the assigned errors [Ita and Stixrude, 1992]. We also assumed that the jumps in both $v_p$ and $\rho$ were small ($\pm 0.5\%$) at the G and L discontinuities. This requirement was loosened to $\pm 1\%$ for $\rho$ at the 520-km discontinuity, where a density increase of 0 to 2% is expected for the $\beta - \gamma$ transition in $(Mg, Fe)_2SiO_4$.
We built into the model prior three types of constraints on the amount of anisotropy. We assumed that below some depth $z_{\text{max}}$, which we varied in different inversions (described below), the mantle is isotropic, that is, $v_{PH} = v_{PV}$, $v_{SH} = v_{SV}$, and $\eta = 1$. In the anisotropic region above $z_{\text{max}}$, we assumed the anisotropy ratio for $P$ waves, $\Delta v_P = 2\left(v_{PH} - v_{PV}\right)/\left(v_{PH} + v_{PV}\right)$, is equal to that of $S$ waves to within $\pm 2\%$, consistent with theoretical calculations for a pyrolite mineralogy [Estey and Douglas, 1986] as well as observations of mantle rocks [Christensen, 1984]. Finally, we constrained $\eta(z)$ in anisotropic layers to be close to those in PREM (± 5%), which also conforms to mineralogical expectations [Nataf et al., 1986; Montagner and Anderson, 1989].

To complete the prior, we assumed that the shear velocities and all otherwise unconstrained model parameters had expected values equal to those in the PA2 starting model. These estimates were assigned large standard deviations (±0.5 km/s for $v_s$, ±20 km for discontinuity depths), except in the crust and shallowest portion of the mantle, where we had independent estimates from rock samples and shallow seismic refraction data [e.g., Shearer and Orcutt, 1986]. In the crust, for example, we took the prior uncertainties to be 0.1 km/s for wave speeds and 0.1 g/cm$^3$ for density.

Our preferred model for the Tonga-Hawaii corridor, PA5, is presented in Figure 11 and Table 2. The progression from PA2 to PA5 involved three complete iterations of the linearized inversion. In the first iteration, we reduced the sediment thickness from 1 to 0.2 km to match better the geological estimates for the corridor [Ludwig and Houtz, 1979], and we restricted the anisotropy to the seismic lid above the G discontinuity by setting $z_{\text{max}} = Z_G$ in the model prior. The resulting perturbation accounted for some of the larger discrepancies between PA2 and the observations. To satisfy the $ScS$ reverberation data in Table 1, the lid was thinned substantially, the 410 was deepened, the 660 was raised, and a 520 was introduced. The fit to the average surface-wave and body-wave phase delays was improved by thinning the lid and raising its shear velocities, and the splitting observed for these phases was accommodated by introducing radial anisotropy.

Significant residuals remained, however, including large negative residuals in the Love-wave phase delays that indicated the need for more anisotropy. In the next iteration, we conducted a series of experiments to investigate the depth extent of the anisotropy. We first reinverted the data set (remeasured using isolation filters calculated for the new reference model, PA3) retaining the constraint $z_{\text{max}} = Z_G$, that is, under the hypothesis that the anisotropy is dominated by lattice preferred orientation frozen into a thin mechanical boundary layer (MBL), identified with the lid. We found that models of this type could not fit both the Love and Rayleigh waves. The problem is most evident in the Love waveforms, as illustrated in Figure 12; models with...
anisotropy restricted to the lid produce simple, impulsive groups, like the trace labeled $A_{70}$, whereas the data show a more complex, emergent waveform caused by a dephasing of the first three overtones with respect to the fundamental. (Note that the first two higher modes are larger than the fundamental.) To accomplish this dephasing, it was necessary to increase the $SH$ velocity in the LVZ, which advances the 0th- and 1st-mode branches relative to the 2nd and 3rd. Anisotropy in the LVZ was thus required in order to maintain the $SV$ contrast across the G discontinuity, as dictated by the $ScS$ reverberation data.

We therefore relaxed the constraint on the depth extent of the anisotropy by setting $z_{max} = z_{410}$, which resulted in a model (PA4) with anisotropy extending throughout the LVZ. The anisotropy in the high-gradient zone (HGZ) between 200 and 400 km was small, however (~1% or less). Further experiments demonstrated that the velocity gradients in this region are well determined by both the $SH$ and $SV$ dispersion data, from which we concluded that the anisotropy of the HGZ must be minor. In the third and final iteration, we set $z_{max} = z_L$, the depth to the interface (internal to the LVZ) that defines the top of the HGZ. This depth is unconstrained by the $ScS$ reverberation data, although the $SV$ contrast across it was required to be small (0 ± 1%) to satisfy the lack of an $L$ discontinuity peak in the reflectivity profile. The resulting model, PA5, provides a good overall fit to the data set (Table 1, Figure 7), yielding a variance reduction of over 80% relative to PA2. Anisotropy in the model extends to $z_L = 166$ km, which gives a good match to the Love waveforms not obtained by models with shallower anisotropy (Figure 12).

Observed seismograms are compared with the full mode-synthetic seismograms calculated for PA5 in Figure 13. These events were chosen solely because they represent a variety of source depths and distances with mechanisms that yield strong signals on both transverse and vertical components; hence they are truly representative of our complete data set. The overall match is quite good, especially considering that the misfits are

![Figure 10](image1.png)

**Figure 10.** Values of density $\rho$ and bulk sound velocity $v_s$ calculated by Ita and Stixrude [1992] for the pyrolite model (circles with error bars). These were used to construct prior distributions for $\rho$ and $v_s$ in the inversion. The values calculated from resulting model PA5 (solid lines) are shown for comparison.

![Figure 11](image2.png)

**Figure 11.** Model PA5. From left to right, $\eta$, density, shear velocities, and compressional velocities are plotted as a function of depth. The model is radially anisotropic through the lithosphere and the low-velocity zone, with $v_{SH}$ and $v_{PH}$ being higher than $v_{SV}$ and $v_{PV}$, respectively. Below 800-km depth, the model is the same as the preliminary reference Earth model (PREM) at a reference frequency of 35 mHz.
Table 2. Model PA5

<table>
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<tr>
<th>z, km</th>
<th>( \rho, \text{Mg/m}^3 )</th>
<th>( v_{SV}, \text{km/s} )</th>
<th>( v_{SH}, \text{km/s} )</th>
<th>( v_{PV}, \text{km/s} )</th>
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Model is calculated at a reference frequency of 35 mHz; it is identical to PREM below 801 km.

Discussion

The formal standard errors in the model parameters are easily calculated from the a posteriori covariance matrix associated with (5), but these values tend to underestimate the actual uncertainties in structural inverse problems, which are often dominated by hidden biases in the data and assumptions underlying the model parameterization [Woodhouse and Dziewonski, 1984]. In the discussion that follows, our assessment of the structural features of PA5 therefore combines these posterior uncertainties with other measures, such as the structural variations associated with plausible modifications to the model prior, the sensitivity of the model parameters to possible bias in subsets of data, and the compatibility of PA5 with previous studies of oceanic upper-mantle structure not modeled by PA5 [Katzman et al., 1996].

almost always associated with phases such as sS, ScS, and sScS, which show considerable variability caused by near-source and lower-mantle structure not modeled by PA5 [Katzman et al., 1996].

Radial Anisotropy

The depth of anisotropy in PA5 is 166 km, shallower than in previous studies based on surface waves and higher modes, which range from 220 km [Regan and Anderson, 1984] to over 300 km [Cara and Lévêque, 1988; Nishimura and Forsyth, 1989]. This difference probably reflects the model parameterizations and minimization criteria more than any significant observational discrepancies. Regan and Anderson [1984] adopted a PREM-like structure with anisotropy extending to an L discontinuity fixed at 220 km, while the other two studies employed smooth parameterizations that permitted small amounts of anisotropy at greater depths. Between G and 410, PA5 has only two layers separated by a (small) L discontinuity, which delineates the base of the anisotropy. The formal error in the depth to L (±10 km) is considerably smaller than that estimated for the depth of anisotropy (±30 km), although in PA5 these depths are constrained to be identical. Additional inversion experiments using more layers indicated that the data marginally prefer a distribution of anisotropy that extends slightly deeper and an L that is smaller than in PA5. However, in all of these experiments, the large splitting observed for the surface waves combined with the smaller splitting of the reflected shear waves (Figure 7) constrained the integral of the anisotropy ratio \( \Delta v_S/v_S = 2(v_{SH} - v_{SV})/(v_{SH} + v_{SV}) \) to be small (< 1%) between 200 km depth and the 410 discontinuity. Because these models contained more free parameters but did not significantly improve the fit to the data, we maintained the simpler parameterization of PA5.

The frequency-dependent travel times used in the PA5 inversion are integrals of a projection of the total (3-D) anisotropy onto the direction of wave propagation. Seismic refraction experiments and laboratory measurements on rock samples from ophiolites have demonstrated that the anisotropy in the lid is primarily due to lattice preferred orientation in olivine, with the fast a axis nearly horizontal and aligned with the initial seafloor spreading direction [Nicolas and Christensen, 1987]. On a hand-sample scale the peridotite anisotropy ranges from 3-8% in P waves and 3-6% in S waves [Christensen, 1984]. The projection of this fossilized anisotropy onto the propagation path is expected to yield radial anisotropy ratios \( \Delta v_S/v_S \) that are positive and large (up to 5%) for paths perpendicular or oblique
to the spreading direction, but small (and even negative) for paths parallel (within ±20°) to this direction [Kawasaki and Kon'no, 1984; Maupin, 1985]. Although the plate reconstructions for the southern part of the corridor are uncertain [Joseph et al., 1987], the corridor axis probably has a mean orientation of about 50°-60° relative to the seafloor-spreading directions. The average value of $\Delta v_S / \bar{v}_S$ in the PA5 lid, +3.7%, is thus consistent with the magnitude expected from the spreading-controlled LPO models [e.g., Kawasaki and Kon'no, 1984]. This shear anisotropy is also comparable to the PREM global value and to azimuthally averaged regional models of the central and western Pacific [Nishimura and Forsyth, 1989], as well as that observed for the uppermost mantle beneath Australia [Gaherty and Jordan, 1995]. Models with nearly isotropic lid structures [e.g., Schlue and Knopoff, 1977; Regan and Anderson, 1984] do not appear to be capable of satisfying the Tonga-Hawaii data.

Models with anisotropy confined to the lid [e.g., Mitchell and Yu, 1980] also fail to fit the data (Figure 12), so there must be some anisotropy in the sub-G layer of the LVZ. Our inversion experiments suggest that the average shear anisotropy in this layer must be greater than ~2%. In PA5 the shear anisotropy ratio increases slightly, though insignificantly, across G (from 3.4% to 4.3%), and then decreases linearly to 1.9% at the L discontinuity. While the lack of a resolvable peak in $S_{CS}$ reflectivity profile indicates that the $SV$ impedance contrast across L is probably less than ~1%, a more substantial (negative) contrast in $v_{SH}$ is allowable. For example, a model with $v_{SH}$ set to a constant value of 4.45 km/s in the sub-G layer fits the data as well as PA5. Such a structure implies a sharp decrease in the Voigt average of the shear velocities at L, however, which is difficult to explain in the context of a peridotitic mantle of constant composition [Leven et al., 1981; Gaherty and Jordan, 1995]. For this reason, the L discontinuity in PA5 was constrained to be small.

The explanation of sub-G anisotropy is ambiguous. It could be caused by LPO in olivine [Ribe, 1989] or by the preferred orientation of melt-filled pockets [Schlue and Knopoff, 1977]; moreover, the alignment of the LPO and/or melt pockets could be induced by paleo-strains that took place near the ridge crest [Ribe, 1989, 1992; Zhang and Karato, 1995] or by shearing in the asthenosphere due to present-day plate motions [Montagner and Tanimoto, 1991]. The plate-motion changes during the long cooling history of the central and western Pacific have been large enough to potentially complicate the depth distribution of azimuthal anisotropy. For example, the seafloor-spreading direction differs by about 40° from the azimuth of the current absolute plate motion where the Tonga-Hawaii corridor crosses the Clipperton fracture zone. There is some seismic evidence for an azimuthal variation with depth; the high-frequency Rayleigh-wave propagation in the central and western Pacific as a whole is fastest roughly parallel to the fracture zones [Nishimura and Forsyth, 1988], while longer-period Rayleigh waves [Nishimura and Forsyth, 1988; Montagner and Tanimoto, 1991] and vertical shear-wave splitting [Farra and Vinnik, 1994; Su and Park, 1994] imply a fast direction roughly parallel to absolute plate motion. The termination of the anisotropy at the L discontinuity in PA5 could correspond to the base of the zone of coherent horizontal deformation (either fossilized or active) or to a change from dislocation-dominated to diffusion-dominated creep within an active shear zone, as suggested by Karato [1992].

**G Discontinuity**

The PA5 shear velocities drop by about 6% across a G discontinuity at 68 ± 4 km. This depth may be an overestimate, because PA5 predicts a value of $t_G$ that is 2.2 s larger than the $S_{CS}$ reverberation datum. If the $SV$ lid velocity is fixed at the PA5 value of 4.66 km/s, the observed time is fit exactly by $t_G = 59$ km. However, thinning the lid while still maintaining the fit to the surface waves requires an increase of ~0.15 km/s to both $v_{SV}$ and $v_{SH}$, driving them to values on the order of 4.8 and 5.0

![Figure 12. Seismograms showing a multimode Love wave that is diagnostic of high $v_{SH}$ in the LVZ. The observed seismogram (center trace) is the transverse component recorded at KIP from a shallow-focus event. The direct S wave arrives at ~13.5 min and the emergent Love-wave group at ~16 min. The trace labeled $A_{70}$ is a complete synthetic seismogram for an Earth model where anisotropy is restricted to the upper 70 km. Synthetics of the first four mode branches (0th-3rd) for $A_{70}$ are displayed above the full synthetic. The trace labeled $A_{170}$ is a complete synthetic for PA5, where the anisotropy extends into the LVZ. Synthetics of the first four mode branches (0th-3rd) for PA5 are displayed below the full synthetic. Comparison of the mode-branch synthetics for the two models demonstrates that the increased $SH$ velocity in the LVZ in PA5 enhances and advances the 0th and 1st mode branches, thereby generating a complete synthetic with emergent, dispersed behavior similar to that observed in the data.](image)
Figure 13a. Comparisons between observed and synthetic seismograms from four representative shallow- and intermediate-focus events. Each event is represented by both tangential-(LPT) and vertical- (LPZ) component seismograms; for each event pair, the top trace is the data, the bottom trace is the synthetic for final model PA5. The traces are aligned (at 0 min) on the S wave, with the synthetic seismograms further corrected for an event static, and major phases are labeled. All traces are low-passed with a 45-mHz corner.

km/s, respectively. Lid velocities of this magnitude are difficult to reconcile with $S_n$ apparent velocities of 4.70-4.80 km/s typically observed for earthquakes in older oceanic regions [Walker, 1977] and the values of 4.55-4.75 km/s found in a refraction experiment using a borehole seismometer just east of the Tonga trench [Shearer and Orcutt, 1986]. (Although $S_n$ times are usually read from short-period, vertical-component seismograms, the 3-D scattering of these high-frequency waves tends to couple the two quasi-shear waves, so that their travel times are expected to give an average intermediate to the $SH$ and $SV$ velocities; because they are picked as first arrivals, they may be weighted more heavily towards $SH$ [Gee and Jordan, 1988].) Since it is possible that the G peak in the reflectivity profile of

Figure 13b. Same as Figure 13a, except for four representative deep-focus events. The large amplitude phases arriving just prior to SS are $sS$ and $ScS$. An example of misfit due to poorly modeled $ScS$ can be seen near 3 minutes on LPT for 870210 and 841117.

Figure 2 is biased by an unresolved H discontinuity [Revenaugh and Jordan, 1991c], we accepted the $t_G$ misfit.

In any case, the path-average lid velocities for the Tonga-Hawaii corridor are higher and the lid/LVZ transition is shallower than in most previous isotropic models. For example, $z_G$ is about 100 km in PA2 and Grand and Helmberger's [1984b] western-Atlantic model ATL (Figure 14), and a comparable value has been inferred for the central Pacific by Zhang and Tanimoto [1993] from their global 3-D model [see also McNutt, 1995]. On the other hand, the PA5 estimate of $z_G$ is similar to those found in the radially anisotropic models by Regan and Anderson [1984] and Nishimura and Forsyth [1989]. It is also consistent with the lid thickness directly beneath Hawaii observed in $S$-to-$P$ converted phases [Bock, 1991].

Heat-flow and bathymetry data indicate that the thermal boundary layer (TBL) for 100-Ma oceanic plates extends to depths of 100-120 km [Parsons and Sclater, 1977; Stein and
Stein, 1992; Carlson and Johnson, 1994; McNutt, 1995). Therefore, in the central Pacific, G is a feature internal to the oceanic TBL rather than defining its base. The existence of a sharp reflectivity peak implies that G must be a relatively narrow transition, with most of the impedance drop occurring over an interval of less than 30 km [Revenaugh and Jordan, 1991b]. The data on surface-wave attenuation are consistent with (though do not require) a large drop in Q across this boundary (Figure 2), suggesting that this feature corresponds to an abrupt drop in the solidus temperature with depth. In an isochemical mantle, this decrease could be caused by the release of water owing to the destabilization of hydrated phases like amphibole [cf. Ringwood, 1975, Figure 4.7]. However, the amount of water in the normal oceanic upper mantle appears to be small enough, of the order of a few hundred parts per million [Dixon et al., 1988; Michael, 1988], to be accommodated in the nominally anhydrous phases such as pyroxene [Bell and Rossman, 1992] and olivine [Bai and Kohlstedt, 1992], in which case hydrous minerals would not play a significant role in determining the depth dependence of the solidus.

We prefer an alternative scenario in which the decrease in the solidus temperature corresponds to an increase in the total amount of water. A chemical transition of this sort could be the result of the decompression melting at the ridge crest, where the separation of basaltic magma from peridotites is an efficient mechanism for drying out the uppermost mantle [Hirth and Kohlstedt, 1996]. In Hirth and Kohlstedt's [1996] model, the water content in olivine increases rapidly below the zone of residual harzburgite, and since water has a substantial effect on the long-term mechanical strength of olivine [e.g., Karato and Wu, 1993], this causes a large downward decrease in viscosity. This increase in water will also decrease the seismic velocities [Karato, 1995]. In this scenario therefore, G would represent the fossilized lower boundary of the melt separation zone. This hypothesis has the advantage of potentially explaining the peculiar observation of a thicker but slower seismic lid in the Philippine Sea [Gaherty et al., 1995].

High-Gradient Zone

The gradient in shear velocity between the L and 410-km discontinuities is high, $(2.2 \pm 0.2) \times 10^{-3} \text{ s}^{-1}$, and well determined by the data. This HGZ appears to be a characteristic feature of oceanic upper mantle. The PA5 velocities in this range nearly coincide with those obtained from 1-D isotropic models of the North Atlantic [Grand and Helmberger, 1984a], the East Pacific...
Rise [Grand and Helmberger, 1984b], and the northwest Pacific [Lerner-Lam and Jordan, 1987] (Figure 14). The global 3-D tomographic models that use PREM as a reference, such as S12_WM13 of Su et al. [1994], typically have smaller gradients at these depths (Figure 15), and they are consequently less successful in matching the waveforms at higher frequencies (Figure 16).

The high gradient below L excludes the possibility that there can be a significant increase in either the SV or SH impedance at L, which is consistent with the absence of an L peak in the SV reflectivity profile. This implies that there is no distinct base to the LVZ in this depth range, and that the association of L with such an increase [e.g., Lehmann, 1959; Anderson, 1979; Hales, 1991] can be discounted, at least in the upper mantle beneath old ocean basins. From the available data it appears that L is more accentuated beneath continents, where it represents a fairly abrupt transition from an anisotropic zone to a more isotropic region of the upper mantle [Revenaugh and Jordan, 1991c; Gaherty and Jordan, 1995]. As in the other oceanic models in Figure 14, the low-velocity "channel" in PA5 extends from G all the way to the 410.

In this region of the mantle, the continental cratons typically display higher values and lower gradients in v_s than old ocean basins [Lerner-Lam and Jordan, 1983, 1987; Grand and Helmberger, 1984b], an inference which supports the hypothesis that the continental cratons are characterized by a very thick (300-400 km), chemically stabilized TBL [Jordan, 1975, 1988]. Some authors have questioned this inference on the grounds that the previous seismic studies did not explicitly account for upper-mantle anisotropy [e.g., Anderson, 1979; Karato, 1992]. However, the shear velocities in Gaherty and Jordan's [1995] 1-D, anisotropic model for Australia (AU3), obtained by an analysis nearly identical to that used here, are fast relative to PA5 down to depths exceeding 300 km, consistent with the thick-TBL hypothesis.

The HGZ may be due in part to the gradual solution of pyroxene into garnet component over a broad depth interval [Akaogi and Akimoto, 1977; Irfune and Ringwood, 1987], but the much lower gradient in the AU3 continental model suggests that the HGZ is primarily governed by the thermal structure of the convecting oceanic upper mantle. The geotherm T(z) in this convecting region should approximate an adiabat with a gradient of about 0.3 K/km [Ito and Stixrude, 1992]. Since the solidus temperature T_m(z) for mantle peridotites has a significantly larger gradient, of the order of 1.0 K/km or greater [Thompson, 1992], the homologous temperature T /T_m will decrease by 10% or more over the HGZ. If the mantle adiabat approaches the mantle solidus at the top of the HGZ, then d v_s / d(T /T_m) may be sufficiently large [e.g., Sato et al., 1989] to explain the high shear-velocity gradient.

Transition Zone

The 410 discontinuity in PA5 occurs at 415 ± 3 km, somewhat deeper than that in either ATL or PA2, but nearly identical to the average value of 414 ± 2 km obtained by Revenaugh and Jordan [1991b] for the southwest Pacific and Australasia and Shearer's [1993] global average of 413 km. The Δv_s of this discontinuity is 4.1 ± 0.9%, similar to other regional and global seismic models, and its velocity-density ratio, Δv_s / Δρ = 1.1 ± 0.3, is consistent with a pyrolite mineralogy [Weidner, 1985; Ito and Stixrude, 1992]. The 660 discontinuity is shallow (651 ± 4 km) relative to Revenaugh and Jordan's average (660 ± 2 km), but similar to Shearer's (653 km). Compared with most other studies,
its shear-wave amplitude is large (\(\Delta V_s = 9.9 \pm 1.5\%\)). Although PA5 fits the reflectivity datum for 660 (Table 1), \(\Delta V_s / \Delta \rho\) for this discontinuity is high (1.5) relative to the values of 0.6-1.0 estimated for isothermal pyrolite models [Bass and Anderson, 1984; D. Weidner, personal communication, 1995]. The significance of this discrepancy should be questioned, however, because the GSDF data constraints in this region are weaker than at shallower depths, and this ratio could be biased upward by our parameterization of structural variations below 660 (we tied PA5 to PREM at 801 km, for example). PA5 includes a 520 discontinuity at 507 \(\pm 10\) km, with \(\Delta V_s = 1.5 \pm 0.5\%\) and \(\Delta \rho = 0.7 \pm 0.5\%\). These combine to give a shear impedance contrast of 1.3%, which is within the error of the ScS-reverberation data (2.1 \(\pm 1.5\%\)). The 520 appears to be a global feature [Shearer, 1990; Revenaugh and Jordan, 1991b] most likely caused by the \(\beta-\gamma\) transition in (Mg, Fe)\(_2\)SiO\(_4\), and the velocity and density contrasts that we obtain for this discontinuity are consistent with the available mineralogical data [Rigden et al., 1991; Bina, 1991]. However, the width of the \(\beta-\gamma\) transition is probably at least 30 km [Akagi et al., 1989], and the exsolution of calcium perovskite may also contribute to this feature [Gasparik, 1990; Bina, 1991]. Since these effects tend to broaden the transition, modeling it as a sharp discontinuity, as done with the ScS-reverberation data in recovering the impedance contrast, could underestimate the total impedance contrast of the transition [Revenaugh and Jordan, 1991b].

The shear-velocity gradients in the transition zone are lower in PA5 than in PREM and most other global seismic models (except below 600 km, where PREM has a slope break to a low gradient that is similar to PA5). The gradient between the 410 and the 520 is very close to ATL, but between the 520 and the 660 it is lower (Figure 14), owing to the jump at 520 and the larger contrast across the 660. The lower gradients bring the seismically determined shear velocities more in line with the predictions for a pyrolite composition [Weidner, 1985; Rigden et al., 1991].

Conclusions

PA5 results from a first attempt to satisfy both ScS-reverberation data on mantle discontinuities and frequency-dependent travel times of PSV- and SH-polarized phases that are sensitive to velocity gradients and anisotropy. The data have been collected using a consistent methodology from a common source-receiver array that samples old oceanic lithosphere in a corridor traversing the central Pacific. We have shown that these data can be successfully modeled in conjunction with attenuation data and mineralogical constraints on bulk sound velocity and density to obtain a complete spherically symmetric, radially anisotropic structure for this path.

The vertical resolution of mantle layering that results from this procedure is superior to previous studies. In Figure 15, the PA5 upper mantle is compared with the corresponding path average of the aspherical global tomographic model S12_WM13, and in Figure 16, synthetic seismograms calculated from these models are compared with data for vertical and transverse components of ground motion. At low frequencies (~15 mHz), both models yield good fits to these observations, but at higher frequencies (~35 mHz), the fit of the tomographic model is poor. While the incorporation of higher-frequency data into the aspherical models can be expected to improve the fits in the near future [e.g., Ekström and Dziewonski, 1995], it will be some time before tomography can achieve this level of vertical resolution.

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J.B. Gaherty and T.H. Jordan, Massachusetts Institute of Technology, Department of Earth, Atmospheric, and Planetary Sciences, Rm. 54-512, Cambridge, MA 02139. (e-mail: gaberty@quake.mit.edu; thj@mit.edu)

L.S. Gee, Seismographic Station, ESB 475, University of California, Berkeley, CA 94720. (e-mail: lind@seismo.berkeley.edu)

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