Lateral variations in mantle $P$ velocity from tectonically regionalized tau estimates

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Summary. Estimates of tau functions for a tectonically regionalized earth are obtained from over 1.25 million seismic ray paths of ISC Bulletin data to study the correlation of potential lateral variations in mantle $P$-wave velocities with recognized surface heterogeneity. The use of regionalized travel-time data relies on statistical regularity criteria to check the consistency of the global regionalization. Estimates of tau perturbations attributed to velocity structure at the seismic source and receiver regions are derived by a simple algebraic formulation. Contributions from near-surface heterogeneity are thereby removed and permit the assessment of lateral velocity variations at depth. Estimates of 'single region' equivalent tau functions are constructed and inverted to obtain velocity-depth functions and extremal bounds at the 99.9 per cent confidence level for seven different tectonic regions.

Deviations from a regionally weighted reference mean velocity function which agrees well with PREM (Dziewonski & Anderson) indicate significant differences, particularly between oceanic and continental tectonic regions, extending to a depth of 700 km. $P$-wave velocity anomalies are typically less than ±1 per cent of the reference velocity. The age-dependence of the shallow oceanic mantle is observed by increased velocities from young ($< 25$ Myr) to old ($> 100$ Myr) regions, and similarly from orogenic zones and magmatic belts to stable continental regions. Unlike for young and old oceanic regions, stable continental regions do not show a pronounced gradient in the mean velocity residuals above 250 km. Continental platforms and particularly shields, however, show a compensation in the sign of the mean velocity residual at depths between about 350 and 700 km. Evidence for a velocity anomaly between 700 and 950 km is indicated. Significant negative residuals are observed centred at a depth of about 780 km below oceanic ridges and about 880 km below active continental regions. This anomaly may be related to a boundary layer between the upper and lower mantle. The level of velocity variations decreases below 950 km. Lateral variations are also suggested within 250 km of the core boundary.

Key words: mantle, $P$ velocity, lateral variations, tau, regionalization
1 Introduction

Geophysical research into the lateral structure of the mantle marks a relatively new direction towards understanding the Earth's interior. The lateral heterogeneity of the crust and tectonic processes suggest that the mantle is also heterogeneous. However, problems concerning the correlation of surface heterogeneity with lateral variations at depth are by no means resolved.

Several groups, concentrating on the global mapping of the heterogeneity in the velocity of seismic body waves (Dziewonski, Hager & O'Connell 1977; Clayton & Hearn 1982; Clayton & Comer 1983; Dziewonski & Anderson 1983; Dziewonski 1984) and using surface waves and free oscillation modes (Silver & Jordan 1981; Masters et al. 1982; Nakanishi & Anderson 1982, 1983, 1984a,b; Woodhouse 1984; Woodhouse & Dziewonski 1984; Tanimoto & Anderson 1984), have provided results suggesting the existence of significant lateral variations in mantle structure. Body wave studies of the lower mantle point to variations in the seismic P-wave velocity on the order of 1–2 per cent of the lateral average, with the larger values occurring near the core boundary. Surface wave studies indicate a correlation of anomalous velocity features and azimuthal variations with tectonic regions and the motions of plates. However, although the above groups employ a variety of inversion techniques—tomographic back-projection, spherical harmonic expansion, measurements of surface wave phase and group velocities, and waveform inversion—the problem of resolution needs further work (Tanimoto & Anderson 1985) and it is difficult at this time to assess the significance of the suggested heterogeneities.

The method described below is based on a global regionalization of the Earth to characterize those types of tectonic regions where velocity anomalies are statistically significant. A simple algebraic formulation is derived which applies the tau method (Bessonova et al. 1974, 1976) to regionalized travel-time data. The contributions from crustal and shallow mantle structure to observed travel times are determined. This allows the assessment of lateral variations at depth and the estimation of 'single-region' equivalent tau functions. These tau functions are then inverted to obtain velocity functions for each tectonic region.

2 Method

2.1 Tau Function

The analysis of travel-time anomalies is undertaken in the form of perturbations on the tau function, \( \tau(p) \), a continuous, decreasing, and single-valued function of slowness \( p \), related to the travel time \( T(p) \) and epicentral distance \( \Delta(p) \) by a Legendre transformation,

\[
\tau(p) = T(p) - p\Delta(p).
\]  

(1)

Since

\[
\Delta(p) = -\frac{d\tau(p)}{dp},
\]  

(2)

tau contains all the information of \( T(p) \) and \( \Delta(p) \). Bessonova et al. (1976) show that while the statistical interpretation of seismic data contains measurement errors in both travel times and epicentral distances, the latter errors are second order in estimates of tau.

One advantage of using the tau function is that it satisfies a Clairaut first-order differential equation and there exists a very simple method of estimating tau and its uncertainty directly from raw travel-time data (Bessonova et al. 1974, 1976; Lee 1981; Lee & Johnson 1984).
These authors have shown the tau method to be a reliable and efficient means of analysing global data. Furthermore, the inverse problem of mapping a tau function into a velocity-depth function has an analytical solution for either the mean tau or its extremal bounds (Johnson & Gilbert 1972; Bessonova et al. 1974, 1976). Although the tau method was developed assuming a spherically symmetric model, Frazer & Phinney (1980) and Buland & Chapman (1983) show that the basic concepts and properties of the tau function can be extended to studies of laterally heterogeneous models.

2.2 REGIONALIZATION

The method of regionalized tau estimates relies on the ability to divide the surface heterogeneity of the Earth into a few types of tectonic regions and on the premise that with each such region is associated a characteristic velocity structure which extends through the crust and into the upper mantle. The question of how deeply into the mantle a tectonic regionalization based on surface features can be traced is important and somewhat controversial (Toksoz & Anderson 1966; Jordan 1975; Okal 1977; Anderson 1979; Jordan 1978, 1981, and further references in the latter). Numerous studies, beginning with that of Toksoz & Anderson, have attempted to use the known lateral variations at shallow depths to remove from the data the effects attributable to the individual tectonic regions. Inverting these regionalized data sets then leads to velocity models which can be compared to determine the depth extent of the differences. An advantage of this method of \textit{a priori} regionalization is that the known heterogeneity of the shallow mantle is explicitly taken into account and has less chance of masking the unknown heterogeneities at greater depths. However, heterogeneities that do not correlate with the tectonic regionalization are difficult to uncover.

In this study, a regionalization based on global tectonics is adopted to subdivide the data and estimate regionalized tau functions. The regionalization is based on features that are readily apparent in the crust and shallow mantle. The hypothesis that this regionalization extends to greater depths is checked progressively throughout the entire mantle by using preliminary tau estimates to test the consistency of the chosen regionalization. Failure of the tau functions to satisfy the expected behaviour of a realizable velocity structure for each tectonic region suggests modifications to the regionalization. An important aspect of this study is that statistical estimates of uncertainty are propagated throughout the entire process so that the significance and resolution of all results can be quantitatively assessed.

If there is no significant correlation of lateral velocity variations with surface regionalization, significant differences between the tau functions of the independent data sets are not expected. In this case, there are two conclusions which must be considered. (1) The mantle is inhomogeneous in its (compressional) velocity structure but the inhomogeneities cannot be characterized entirely according to surface tectonic features; there is sufficient ‘mixing’ of the regions in depth that the seismic waves essentially sample a mean earth for any given path. (2) There are no lateral variations at the depths considered or they are not resolvable. On the other hand, a positive correlation of anomalous tau estimates with surface regionalization would indicate that differences in structure may be related to variations manifested at the surface.

2.3 REGIONALIZED TAU FUNCTION

The lateral heterogeneity of the Earth’s crust and shallow mantle is characterized by \textit{N} tectonic regions, each of which is taken to be relatively homogeneous in its lateral velocity
structure. Assume that the travel-time properties of a typical seismic body wave can be characterized by the material near the turning point, in which the major portion of the travel time is spent, and perturbations from the shallow structure near the source and receiver where lateral variations in velocity are known to exist. The corresponding tau function is written

$$\tau_{ijk}(p) = \tau^T_i(p) + \delta\tau^S_i(p) + \delta\tau^R_k(p) + \epsilon_{ijk}(p)$$

where $\delta\tau^S_i(p)$ and $\delta\tau^R_k(p)$ denote respectively the tau perturbations, at a given slowness, attributed to region $i$ at the source (S) and region $k$ at the receiver (R). The function $\tau^T_i(p)$ is associated with the turning-point (T) path segment, parametrized by the tectonic region $j$ corresponding to its radial projection on the Earth's surface. $\epsilon_{ijk}(p)$ is a random variable with zero mean.

Statistical estimates of various tau quantities may be determined through a simple algebraic formulation. Denote the expectation operator for index $m$ applied to a weighted vector function by

$$E_m[w\phi] = \sum_m w_m \phi_m.$$ 

Next, impose the constraint that the weighted means of the source and receiver perturbations equal zero at any given slowness

$$\sum_{i=1}^{N} w_i \delta\tau^S_i(p) = \sum_{k=1}^{N} w_k \delta\tau^R_k(p) = 0$$

where

$$\sum_{n=1}^{N} w_n = 1.$$ 

Then the operation

$$E_k[w_k \tau_{ijk}(p)] = \tau^T_i(p) + \delta\tau^S_i(p) + \epsilon_{ij}(p) = \tau_{ij}(p)$$

eliminates the receiver perturbation, and

$$E_l[w_l \tau_{ij}(p)] = \tau^T_i(p) + \epsilon_{ij}(p)$$

eliminates the source perturbation, thereby isolating the turning-point tau. The source and receiver perturbation terms are obtained as expectations of differences

$$\delta\tau^S_i(p) = E_l[\tau_{ij}(p) - \tau^T_i(p)]$$

$$\delta\tau^R_k(p) = E_l[\tau_{ijk}(p) - \tau_{ij}(p)].$$

Separating from the tau functions contributions attributed to the velocity structure at the source and receiver regions is similar to the ‘time term’ analysis method in refraction seismology (Willmore & Bancroft 1960). It is important to note, however, that the formulation here treats these perturbations separately for the source and receiver regions and explicitly as functions of slowness.

Since a complete travel-time curve for the mantle involves distances out to $100^\circ$, exceeding the geographical dimensions of the tectonic regions, most seismic rays will propagate through more than one region along the path from source to receiver. Consequently, a
Table 1. Distribution of tau data $\tau_f^T(p)$ according to turning point sampling for all slowness intervals considered. Results are prior to uniform reduction analysis (Jeffreys 1961). Also given is the fractional surface area of regions according to the modified GTR1 (Fig. 2) geographical grid.

<table>
<thead>
<tr>
<th>Tectonic region</th>
<th>Fractional area</th>
<th>Paths</th>
<th>Fraction of total paths</th>
</tr>
</thead>
<tbody>
<tr>
<td>(1) Young oceans</td>
<td>0.13</td>
<td>46 184</td>
<td>0.037</td>
</tr>
<tr>
<td>(2) Intermediate-age oceans</td>
<td>0.34</td>
<td>246 091</td>
<td>0.195</td>
</tr>
<tr>
<td>(3) Old oceans</td>
<td>0.13</td>
<td>111 012</td>
<td>0.088</td>
</tr>
<tr>
<td>(4) Active continents</td>
<td>0.19</td>
<td>468 728</td>
<td>0.372</td>
</tr>
<tr>
<td>(5) Continental platforms</td>
<td>0.10</td>
<td>221 407</td>
<td>0.176</td>
</tr>
<tr>
<td>(6) Continental shields</td>
<td>0.07</td>
<td>55 810</td>
<td>0.044</td>
</tr>
<tr>
<td>(7) Oceanic trenches</td>
<td>0.04</td>
<td>110 523</td>
<td>0.088</td>
</tr>
<tr>
<td>Total</td>
<td>1.00</td>
<td>1 259 755</td>
<td>1.000</td>
</tr>
</tbody>
</table>

'single-region' equivalent tau curve for the mantle is constructed for each region from the contributions attributed to the critical path segments

$$\tau_{ijj}(p) = \tau_f^T(p) + \delta \tau_f^S(p) + \delta \tau_f^R(p).$$

Regionalization of seismic data is intimately related to the problem of uniform sampling. It is expected that the distribution of epicentres and seismographic stations prohibits an even sampling of the Earth's surface. A three-point classification which includes the turning-point projection not only extends the sampling to depth but allows for the correlation of properties attributed to structure at depth with surface heterogeneity. However, in spite of the large amounts of data available, there are certain source, receiver, and turning-point region combinations which are not sufficiently sampled to ensure a dense distribution of ray paths and provide confident statistical estimates. Biases introduced by the over- and undersampling of characteristic velocity structures at the source and receiver regions are thereby decreased by weighting the tau estimates. In this study the weights are chosen as the fraction of the Earth's surface area contained in region $n$ (Table 1). This ensures a uniform areal sampling of the earth by source and receiver raypath segments.

3 Reduction of data

3.1 RAW TRAVEL-TIME DATA

The travel times of seismic P-waves from the Bulletin of the International Seismological Centre (ISC) provide the basic data set. Over 1.25 million ray paths (Table 1) for the periods 1971 February to 1974 December, 1978 January to 1981 January, and 1981 April to 1981 October, in the epicentral range of 10-100° and for source depths less than 70 km were compiled. Travel times are compared to the Jeffreys & Bullen (1958) P-wave table and considered a direct arrival if the residual is within 50 s. Gross errors in the travel-time readings are removed by uniform reduction analysis (Jeffreys 1961) to obtain a Gaussian distribution of observed times at fixed distance intervals. Unlike for core phases (Johnson & Lee 1984), there is insufficient resolution in the data to obtain estimates of tau functions for the separate branches of the upper-mantle triplications between approximately 15 and 30°. Tau estimates in this interval thereby correspond to a smoothed travel-time curve.
Crustal and focal depth corrections are applied to move the source and receiver to a common depth of 33 km using Jeffreys' (1939) velocity model. Ellipticity corrections are also applied. The data are then separated according to the tectonic regions of the epicentre, receiver station, and surface projection of the turning point. Each of these data groups is further divided into $p-\Delta$ intervals and then the method of Bessonova et al. (1976) is used to estimate $\tau_{ijk}(p)$ and its uncertainty at the 99.9 per cent confidence level.

The 29 lower-mantle $p-\Delta$ intervals obtained by Lee (1981) and Lee & Johnson (1984) are also used here for all the regions. The 11 upper-mantle intervals, empirically determined as part of this study, are different for each region and do not indicate the complications with regularity criteria cited in Lee (1981). Evidently, the regionalization separates different modes in the tau distributions resulting from the variations in the lateral velocity structure of the upper mantle.

3.2 REGIONALIZATION OF TRAVEL-TIME DATA

The travel-time data were initially regionalized according to GTR1 (Jordan 1981), which consists of three types of oceanic and three types of continental regions, with a $5 \times 5^\circ$ geographical gridding. The oceanic regions are defined by equal increments in the square root of crustal age ($< 25$, $25-100$, and $> 100$ Myr). The three primary continental regions are orogenic zones and magmatic belts, platforms, and Precambrian shields and platforms. Using the method outlined below, this regionalization was modified by introducing a seventh region for oceanic trenches.

For each $p-\Delta$ interval the effect of outliers is first removed by applying Jeffreys' (1961) method of uniform reduction (see Lee 1981; Lee & Johnson 1984, for details). Both Kolmogorov–Smirnoff (Hays & Winkler 1970) and $\chi^2$ tests are then used to detect deviations from a normal distribution. Significant deviations indicate that the sample is not obtained from a single population and therefore the regionalization is suspect. The regionalization is thereby checked at different depths in the mantle. The Kolmogorov–Smirnoff test is a parti-

![Figure 1](https://academic.oup.com/gji/article-abstract/86/2/475/683295)

**Figure 1.** Normalized distributions of frequency of tau estimates $\tau_{ijk}$ at a slowness 8.10 s deg$^{-1}$ for turning point $j$ below continental platforms (region 5) and one prescribed raypath endpoint region $i$ and/or $k$. 


Variations in mantle P velocity

Fig. 1 shows an intermediate stage of adjusting the regionalization. Histograms of residuals about mean interval tau are used to identify non-normal distributions. The bimodal distribution for sources or receivers in intermediate-age oceans (region 2) and orogenic zones and magmatic belts (region 4) suggested contamination from a region with characteristically faster velocity, giving tau anomalously less than the expected interval mean. The coordinates of the epicentres, receiver stations, and surface projections of the turning points, for ray paths contributing to these anomalous distributions, were located on a Mercator projection of the Earth and 5 x 5° geographical gridding of the regionalization and led to further modifications of the gridding of region 7. This re-characterization of data from regions 2 and 4 into region 7 is also consistent with the inadequacy of GTRI in areas of transitional crust (e.g. continental margins, island arcs, and oceanic plateaux adjoining continental crust) which often border intermediate-age oceans and active continental regions (Jordan 1981, fig. 1).

The data were also subdivided according to the regionalization of Okai (1977), which includes a region for oceanic trenches and provides a better characterization of transitional

![Diagram](https://academic.oup.com/gji/article-abstract/86/2/475/683295)
crust. Although tau estimates about interval means were approximately normally distributed, the $15 \times 15^\circ$ geographical discretization restricted the resolution of anomalous regions.

The technique of allowing the seismic data to check the consistency of the regionalization and avoid travel-time biases is a useful addition to the usual methods of regionalizing the Earth. Fig. 2 shows the modified global tectonic regionalization of Jordan (1981) obtained in this study by the adjustment process outlined above. Although this current regionalization meets the assumptions for the estimation of regionalized tau functions, the non-normal distributions could be used to refine the regionalization further, particularly in distinguishing between active and passive orogenic zones and magmatic belts. Initial tests suggested this differentiation may be significant in the Tibetan plateau. However, present results will not be significantly altered unless the number of observations in the less sampled groups is brought to a few hundred data points. This would require a fourfold increase in the initial ISC data set.

3.3 Tau Estimate Selection Criteria

The definition of selection criteria is the result of several experiments designed to optimize the tau estimation process and improve the tectonic regionalization. It is an important step in the application of the statistical method of estimating tau (Bessonova et al. 1976) to regionalized travel-time data, where the number of estimates required increases as $N^mL$ for $N$ tectonic regions, an $m$-point raypath parametrization, and $L_p-\Delta$ intervals.

The limited number of samples in certain tau estimates $\tau_{ijk}(p)$ leads to minimum requirements which have to be satisfied by the data in order for the estimates to be acceptable. No data subdivision with less than 10 samples is used. Based on the previous work of Lee (1981) and Lee & Johnson (1984), the variances of lower-mantle tau estimates are expected to be on the order of $1s^2$ but are allowed to reach a maximum of $4s^2$ to account for significant levels of scatter when the number of observations is small (<50). Similarly, the variances of upper-mantle tau estimates are allowed to reach a maximum of $5s^2$. If the number of rays is greater than 50, a variance larger than $4s^2$ but less than $6s^2$ is acceptable.

The statistical aspects of the method presented in this paper are straightforward to interpret if the tau data are approximately normally distributed. To ensure this, the method of uniform reduction (Jeffreys 1961) was used to remove outliers, a common non-normal component of travel-time data, and the resulting distributions checked for normality by the Kolmogorov–Smirnoff and $\chi^2$ tests at the 90 and 99.9 per cent confidence levels respectively. Only data which passed these tests were accepted and used to perform the operations in Section 2.3. Thus, given normally distributed $\tau_{ijk}(p)$ data, it follows that the results of the linear combinations in equations (8) to (11) are also normally distributed.

4 Discussion of results

Following the procedure outlined in the previous sections, estimates of tau and its uncertainty were obtained for each of the seven tectonic regions. While a discussion of differences between regions could take place in this tau-slowness space, we have chosen to use the equivalent and more familiar velocity--depth space.

Regionalized radial velocity functions $v_r(r)$ were obtained by inverting estimates of single-region tau functions $\tau_{ij}(p)$ (equation 11). The 99.9 per cent confidence intervals on $\tau_{ij}(p)$ were also inverted using the method of Bessonova et al. (1976) to obtain corresponding extremal bounds on the radial velocity functions for each region. While the use of a 99.9 per
cent confidence level is consistent with previous studies of this type. Using a 95 per cent confidence level would reduce the width of the bounds on both tau and velocity by about 30 per cent. The average depth resolution of the velocity models is 75 km, given the 40 independent estimates of tau for the mantle below each region. The nodal velocity—depth values are interpolated linearly to eliminate features often introduced by higher-order fitting functions.

A reference mean velocity model is defined by

$$v_0(r) = \sum_{j=1}^{N} w_j v_j(r)$$

where the weights $w_j$ are the fractional surface area of the Earth for each region $j$ (Table 1). Extremal bounds on $v_0(r)$ are obtained from the extremal bounds on $v_j(r)$ by a similar weighting.

Fig. 3 compares the mean velocity model $v_0(r)$ with the ‘equivalent’ isotropic PREM (Dziewonski & Anderson 1981) at a reference period of 1 s. At 220, 400 and 670 km, $v_0(r)$ falls in the centre of the velocity discontinuities recognized by PREM. Since travel-time triplications were not resolvable in the tau data (Section 3.1), the inverted velocity functions do not show sharp discontinuities. Above 220 km, the differences from PREM are due to inversion of body wave travel times as opposed to using long-period surface wave data. In the upper-mantle transition zone and in the lower mantle, the PREM velocity model lies within the extremal bounds defined for $v_0(r)$, which are always within ±0.1 km s$^{-1}$ and typically about ±0.05 km s$^{-1}$. The decrease in the PREM velocity gradient approaching the core boundary is also evident in $v_0(r)$.

Fig. 4 shows the residual mean velocity functions and extremal bounds of each tectonic region referenced to $v_0(r)$. For young and old oceanic regions and continental shields, the relatively fewer travel-time data points (due to a lack of regional seismicity and/or seismographic stations) is reflected by the larger widths of the extremal bounds.

Figure 3. Comparison of weighted mean $P$-wave velocity model $v_0(r)$ (solid line) and its 99.9 per cent extremal bounds (dotted lines) with PREM (Dziewonski & Anderson 1981) (dot-dashed line).
Surface mean velocities correspond to a 33 km depth datum (Section 3.1) and are associated with the segment of the tau curves smoothly extrapolated to 0s. The trend in these velocities shows a progressive increase from 7.72 km s\(^{-1}\) in oceanic ridges (region 1) to 7.88 km s\(^{-1}\) in intermediate-age oceans (region 2) and to 7.97 km s\(^{-1}\) in old oceans (region 3). However, these velocity differences tend to decrease with depth. The gradients of the velocity residuals are well pronounced and positive to 200 km and then fairly constant to 275 km below region 1, essentially constant to 280 km below region 2, and negative to 285 km below region 3. The increase in the mean P-wave velocity proceeding from young to old oceanic regions is consistent with plate tectonic models of heat flow and magnetic anomaly variations and oceanic temperature profiles, which indicate that seismic velocities are proportional to the square root of crustal age (Parsons & Sclater 1977). The regionalization of the oceans (Jordan 1981) follows this crustal age dependence.

There is also an increase in the mean surface velocities from magmatic belts and orogenic zones (region 4) to continental platforms (region 5) and shields (region 6) of 7.91 to 7.95 and 7.98 km s\(^{-1}\) respectively. Unlike the pronounced gradients in oceanic regions 1 and 3, the mean velocity residuals in the continents are fairly uniform above 250 km. At a depth of 100 km, active continental regions show a significant mean velocity residual of \(-0.03\) km s\(^{-1}\) (\(-0.3\) per cent of the reference velocity). Continental platforms show significant positive residual velocities of \(0.07\) km s\(^{-1}\) at 165 km (0.9 per cent of the reference velocity) and \(0.05\) km s\(^{-1}\) (0.6 per cent) at about 250 km, and shields show residuals of up to \(0.08\) km s\(^{-1}\) (1.0 per cent) also at approximately 250 km.

In stable continental regions the positive velocity anomalies in the upper 250 km are compensated by negative anomalies between about 350 and 700 km. This effect is particularly pronounced below shields, with mean velocity residuals of up to \(-0.11\) km s\(^{-1}\) (\(-1.3\) per cent of the reference) at 350 km, while below continental platforms the profile is similar but the magnitude of the anomalies is smaller, with residuals of \(-0.06\) km s\(^{-1}\) (\(-0.6\) per cent) at
Variations in mantle P velocity

485

a comparable depth. These differences are suggestive of an age dependence of sub-continental structure and perhaps deep continental roots (Jordan 1975, 1978, 1981; Alvarez 1982). The compensation in the upper-mantle velocity structure below stable continental regions may be related to chemically fractionated magma reservoirs (Jacobsen & Wasserburg 1979; DePaolo 1980) and variations in depletion.

Oceanic trenches (region 7) do not indicate significant variations from the reference model. There is a suggestion of positive velocity residuals between 400 and 700 km, although these features are not significant at the 99.9 per cent confidence level. There are no upper-mantle features which can be clearly identified with the effects of subducted lithospheric slabs. This may be due to limitations in the regionalization of oceanic trenches. For instance, the 5° geographical gridding is rather coarse for the trenches and the known lateral structure at depth from dipping slabs has not been explicitly incorporated. The regionalization of oceanic trenches does, however, remove biases in the tau estimates of the other regions (Section 3.2).

Lateral variations in upper-mantle velocity structure could be attributed to fundamental differences in composition or phase, differences in temperature, and to variations in the depth of thermal or chemical boundary layers and of boundaries associated with phase changes (Bott 1971; Jeanloz & Thompson 1983; Lees, Bukowskini & Jeanloz 1983). The pressure and temperature dependence of thermodynamic phase equilibria and the physical mechanisms which determine whether such boundaries are elevated or depressed, given a local change in temperature, and whether such a perturbation is stabilizing or destabilizing in view of mantle convection, are reviewed by Schubert & Turcotte (1971) and Schubert, Yuen & Turcotte (1975). The age dependence of these variations, as shown in this study, suggests significant differences in the upper 700 km.

Thus it is important to determine if the depths to the 400 and 670 km discontinuities vary below different tectonic regions to help constrain estimates of pressure and temperature conditions relevant to the upper mantle. However, the coarseness of the velocity profiles (Fig. 4) at upper-mantle transition zone depths and the complications encountered in the initial data reduction at epicentral distances between 25 and 30° limit the ability of this study to resolve lateral variations in these discontinuities. Experiments on mantle components for potential phase or compositional changes suggest transition widths typically of 10-50 km (cf. Lees et al., 1983). The average depth resolution obtainable by the tau estimates is approximately 75 km, corresponding to slowness intervals of 0.15 s deg⁻¹.

The residual velocity functions in Fig. 4 also contain interesting anomalies between depths of 700 and 950 km. The basic feature appears to be a negative anomaly flanked by smaller positive anomalies. It is most pronounced below young oceanic regions, where it is centred at a depth of about 780 km with residuals between −0.09 km s⁻¹ (−0.8 per cent of the reference) and −0.07 km s⁻¹ (−0.7 per cent), and below active continental regions, where it is centred at about 880 km with residuals of up to −0.055 km s⁻¹ (−0.5 per cent). Evidence for similar variations below old oceans is perhaps hinted at and, although not significant, lends support to the idea that there is a progressive deepening of this velocity anomaly in going from region 1 to 4. A similar feature may be present below region 7 between 750 and 790 km, but it does not appear to be present below the stable continental regions.

This interesting velocity anomaly occurs at a depth where it could be interpreted in terms of thermal and chemical boundary layers between the upper and lower mantle. Note that in Fig. 3 it corresponds roughly to the depth at which there is a transition from the high velocity gradient of the upper mantle to the lower gradient of the lower mantle. This may also be the same structure that led Dziewonski (1984) to obtain a higher level of perturba-
tions below the 670 discontinuity. Earlier studies which have also suggested velocity structures in this depth range include Repetti (1928), Gutenberg (1958), Chinnery & Toksoz (1967), Chinnery (1968) and Johnson (1969). The fact that these earlier results were not entirely consistent with one another may be explained by the finding of this study that the depth of the structure varies systematically with the type of tectonic region.

Below a depth of 950 km the residual velocities are less pronounced and do not show systematic anomalies between the different tectonic regions. Velocity residuals of ±0.3 per cent of the reference model at 1600 and 1760 km depth are shown below active continental regions, as are kinks at 1490 and 2295 km below old oceans, at 2215 km below oceanic trenches, and at 1700 km below continental platforms. These residuals are not significant but are suggestive of variations in \( dT/d\Delta \) measurements obtained by Chinnery & Toksoz (1967) and Chinnery (1968) for depths of 1150–1300 km and of 2000 km. Similar anomalies have also been presented by Vvedenskaya & Balakina (1959), Bugayevskiy (1964) and Johnson (1969). Although the depths are not in strict agreement, there appears to be a dependence of the ability to observe these velocity variations on the type of tectonic region.

Within about 250 km of the core–mantle boundary (CMB), the residual velocity profiles corresponding to active continental regions, continental platforms, and oceanic trenches indicate significant positive residuals, of up to 0.08 km s\(^{-1}\) (±0.6 per cent of the reference) and 0.05 km s\(^{-1}\) (±0.4 per cent) at approximately 150 km above the CMB below regions 5 and 7 respectively. Similarly, young and old oceanic regions indicate velocities slower than the reference at these depths, with a significant residual mean velocity of up to −0.07 km s\(^{-1}\) at 150 km above the CMB below oceanic regions. The increase in the level of variations agrees with the results of Dziewonski et al. (1977), Dziewonski (1984) and Dziewonski & Anderson (1983). Although lateral variations in the velocity structure of this \( D'' \) region are clearly suggested, an interpretation in view of the surface regionalization cannot be provided with confidence. Similarly, comparison with other studies which have suggested lateral variations at the base of the mantle (e.g. Lay & Helmberger 1983) may be susceptible to differences due to sampling bias.

5 Conclusions

This study has developed a method of estimating tau functions for a tectonically regionalized earth. Perturbations attributable to near-surface heterogeneity associated with the seismic source and receiver regions were determined through a simple algebraic formulation. Separate tau functions were estimated for each tectonic region, allowing the assessment of lateral velocity variations at depth. Attention was given to the inherently non-uniform sampling of the earth and several criteria were defined to decrease travel-time biases and ensure the consistency of the tau estimates with the assumptions of the global regionalization.

\( P \)-wave velocity functions for seven tectonic regions based on Jordan's (1981) global regionalization were obtained by inverting single-region equivalent tau functions. The velocity variations are expressed as deviations from a regionally weighted reference model. Extremal velocity bounds at the 99.9 per cent confidence level constrain the significance of the residuals. Significant variations are observed to depths of about 950 km and within 250 km of the core–mantle boundary. The observed differences in the regionalized tau estimates are of order ±0.1 s with causative \( P \)-wave velocity anomalies typically less than ±1 per cent of the reference mean. These values are somewhat smaller than obtained in previous studies (Dziewonski et al. 1977; Dziewonski & Anderson 1983; Dziewonski 1984). However, a thorough consideration of the near-surface heterogeneity was undertaken and attention given to the problem of uniformly sampling the mantle. No baseline corrections were required.
In both oceanic and continental tectonic regions the velocity variations in the upper mantle seem to correlate with age, with surface velocities showing a systematic increase. Beneath oceans differences at the surface tend to be compensated by the velocity gradients in the upper 250 km, whereas beneath continents the surface differences extend to about 250 km and compensation takes place between 350 and 700 km. This suggests fundamental differences between oceans and continents that extend to depths of at least 700 km.

This study reveals evidence for a velocity anomaly between depths of 700 and 950 km which may be indicative of lateral variations in a boundary layer between the upper and lower mantle. Between 760 and 800 km below oceanic ridges and between 830 and 930 km below active continental regions, there are significant negative velocity residuals which may be related to anomalies suggested by previous authors. The level of velocity variations decreases below 950 km, although significant lateral variations are indicated within 250 km of the core boundary.

A number of ways in which this study could be extended are fairly obvious. A more accurate and more detailed regionalization would be useful, so long as there is sufficient sampling of the new regions which are introduced. Making the regionalization a function of depth is possible, but may significantly increase the problems of undersampling. Given the results of the present study it is also possible to relocate all the sources while taking into account lateral variations in the Earth's velocity structure. Ideally, one would like to do this with a completely new data set, however, efforts to relocate a few events are currently in progress. It is furthermore desirable to develop a technique of tracing rays through an arbitrary number of tectonic regions each with a characteristic tau curve. Finally, a similar study using regionalized S-wave data would help constrain inferences about the potential causes of the velocity variations and could also lead to estimates of density variations and a comparison with geoid anomalies.

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References


D. M. Tralli and L. R. Johnson


Variations in mantle P velocity


Repetti, W. C., 1928. New values for some of the discontinuities in the Earth, dissertation, St Louis University.


