Global Images of the Upper 200 km in the Earth by Short-Period Surface Waves

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A global shallow Earth structure is retrieved from short-period surface wave data (40–200 s) using a data set assembled from global digital networks. Phase velocity maps for both Love and Rayleigh waves are retrieved by spherical harmonic inversion. A comparison with results by block inversion is also presented, and the similarity of the results by the two different methods supports the belief that the data are capable of resolving most details in final velocity maps. Strong effects of crustal structure are evident in the patterns of short-period Love wave phase velocity variations, although depth resolution kernels show that the complete resolution of crustal structure is not possible from this period range of data. Two models of S-wave velocity maps are presented, one with an assumed crustal structure and topography and the other without such an assumption. Major differences between the two models occur just below the moho; the one with an assumed crustal structure shows the beginning of tectosphere-like structure immediately below the moho under cratons, while the model without it shows the beginning depths of tectosphere-like structure between 50 and 80 km. It is obvious that the short-period cut-off in this study (40 s) is not sufficiently short to resolve shallow structures in detail. A few interesting tectonic features in the models include: 1) S-wave velocity under ridge axes is slow in the upper 200 km and is very slow in the upper 100–120 km, due possibly to partial melting at shallow depths; and 2) depth slices under the Himalaya-Tibet region show fast velocity anomaly extending from India to the underside of Tibet. While it is difficult to distinguish between a subducted oceanic lithosphere and a thickened continental lid, the shape of the contour lines seems to favor the interpretation that this fast velocity anomaly is the underthrusting Indian ocean lithosphere.

1. Introduction

Retrieving shallow seismic structures, especially within the upper 200 km of the Earth's surface, is quite important to understand the tectonics of the Earth. There have been many studies for this depth range using a number of different seismic phases as data sets. One approach is to use surface-bouncing body waves such as PP, SS, PPP, SSS, ... (e.g., Sheehan and Solomon, 1991; Su and Dziewonski, 1991). The present configuration of global seismic stations, however, is such that those body waves are,

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relatively speaking, deep turning rays, which inherently bring in ambiguity in resolving shallow structures. For example, teleseismic SS phase at a distance of 60 deg has its turning point in the transition zone or in the top part of the lower mantle. For some regions in the Earth, observation of SS phase for smaller distances is not possible. This suggests that the resolution of shallow structures is not possible only with SS-type phases. It is true that this type of bouncing body waves carries information of shallow structures, but the data are also affected (or contaminated) by deeper structures. Another approach is to use surface waves (e.g., Brune and Dorman, 1963; Forsyth, 1975; Mitchell and Yu, 1980; Nakanishi and Anderson, 1984; Nishimura and Forsyth, 1988; Nataf *et al.*, 1986; Zhang and Tanimoto, 1992, 1993) which have their energy concentrated near the surface. It is more advantageous for constraining shallow structures than the body wave approach. A drawback in the surface wave approach is in resolving deep structures. Although long-period fundamental mode phase and higher mode phase can provide information from the deep interior, resolution becomes progressively worse with depth. In this study, for example, the resolution of structures below 300 km depth was very difficult. The surface wave approach also has problems for short periods below about 15 s because of scatterings and multipathings caused by topography and near-surface heterogeneity. These heterogeneities convert a large fraction of surface waves into body waves, and the concept of phase velocity map at a given frequency loses significance at such short periods. For the period range of my study (>40 s), such concerns did not seem to apply, however.

The aim of this paper is to present the results of short-period surface wave inversion for the global Earth structure. The period range of the data set is approximately between 40 and 200 s. One of the improved features of my models over previous results is that the short-period Love wave phase velocity maps (for periods between 40 and 50 s) show strong effects of crustal structure. This may not be surprising, but it can provide a significant improvement for shallow structure studies in the future. A similar study has recently been published by Trampert and Woodhouse (1995), although they only published phase velocity maps.

In the following, we describe the data analysis procedure, phase velocity maps with resolution and error analysis, and S-wave velocity maps. In the main text, we present the results obtained by spherical harmonic inversion. The results by block inversion are given in the APPENDIX. Since the main purpose of this paper is to present a new model of shallow global structure, a detailed discussion of the tectonic implication of the model will be kept to a minimum, although we will point out some interesting features.

2. Analysis

2.1 Basic phase data

I exclusively use fundamental mode surface waves in this paper. One of the most useful properties of fundamental mode surface waves is their generally isolated group velocity from other types of waves. Because their group velocity is different, fundamental mode waves are separated in time from other waves and are thus easy to analyze. In a broadband seismogram, long-period fundamental phase may interfere with higher mode phases, but in general, contamination from other waves can be avoided without

much effort. This issue has been discussed extensively in the past (e.g., Nakanishi and Anderson, 1984; Zhang and Tanimoto, 1993), thus we will not discuss it in detail here. This is in contrast to higher mode waves, which are basically multi-mode phases and thus require a somewhat more complicated analysis technique (e.g., Lerner-Lam and Jordan, 1983; Woodhouse and Dziewonski, 1984; Tanimoto, 1988; Nolet, 1990).

Our basic data are phases, measured at various frequencies, denoted by $\phi_{\text{obs}}(\omega)$, where ω is the angular frequency. We measured this quantity by comparing data with synthetic seismograms for a known Earth model. Phase difference between the observation and synthetic seismogram is always less than 2π at long periods (>150 s) for minor arc data. By connecting the phase from long periods to short periods, ambiguity of the phase measurement can be removed completely for the entire period range. This technique has been used by, for example, Suetsugu and Nakanishi (1985) and Zhang and Tanimoto (1991).

We used the Preliminary Reference Earth Model (PREM, Dziewonski and Anderson, 1981) and the source parameters from the centroid moment tensor (CMT) solutions published by the Harvard University group (e.g., Dziewonski and Woodhouse, 1983) in order to generate the synthetic seismograms. We applied Fourier transformation to both data and synthetic seismograms, and measured phase shift as a function of frequency. This phase shift, given at a constant frequency, should be converted to an eigenfrequency perturbation for the inversion of seismic velocity structures. Picking a particular angular degree (L) and using ϕ_{obs} at the corresponding frequency, the eigenfrequency perturbation is given by

$$\left[\frac{\delta\omega}{\omega} \right]_l = \frac{U}{\omega L} [\phi_{\text{obs}} - \phi_{\text{PREM}}]_{\omega}, \quad (1)$$

where U is the group velocity of this mode and L is the distance between an earthquake and a receiver. The eigenfrequency perturbation is basically the phase velocity perturbation measured at a fixed wavelength; hereafter, we refer to this quantity as phase velocity perturbation or eigenfrequency perturbation.

It is sometimes pointed out that CMT solutions contain systematic origin time bias in some tectonic regions. This bias is typically of the order of a few seconds. Since the signal from fundamental mode surface waves was of the order of 20–30 s in the long-period range of my data set (period 150–200 s) and became larger at shorter periods, the bias in CMT solutions likely did no harm. Even so, in order to reduce this bias, we selected data from earthquakes with good azimuthal coverage.

2.2 Ray tracing

We incorporated ray tracing in the analysis. The introduction of ray tracing makes the problem non-linear and thus iteration is built into the formulation. The formula to solve for local eigenfrequency perturbation from phase data is

$$\phi_{\text{obs}} - \phi_{\text{ref}} = -\frac{\omega a}{U} \int_{\text{raypath}} \left(\frac{\delta\omega}{\omega} \right)_l d\Delta, \quad (2)$$

where a is the radius of the Earth, ϕ_{ref} is the ray-traced phase for a given Earth model

and integration is along a ray path.

In the spherical harmonic approach, the integrand in Eq. (2) is expanded as

$$\left(\frac{\delta\omega}{\omega}\right)_l = \sum_{i,m} s_{lm} Y_l^m(\theta, \phi), \quad (3)$$

where $Y_l^m(\theta, \phi)$ is the spherical harmonic with the angular degree l and the azimuthal order m , and is a function of colatitude θ and longitude ϕ . If defining

$$p_{lm} = -\frac{\omega a}{U} \int_{\text{ray path}} Y_l^m d\Delta, \quad (4)$$

we can write Eq. (2) as

$$\phi_{\text{obs}} - \phi_{\text{ref}} = \sum_{l,m} p_{lm} s_{lm}. \quad (5)$$

We started the inversion process from the Earth model PREM. At each step of iteration, the following procedure was carried out; 1) computation of ray paths for all pairs of source and receiver, 2) computation of ϕ_{ref} and p_{lm} for new ray paths, 3) application of least squares to above Eq. (5), and then solve for s_{lm} , which provided a new phase velocity map.

We developed our own ray tracing codes, but relied heavily on suggestions given in the book by Press *et al.* (1992). The specific technique we adopted was the extrapolation method of Burlisch and Stoer with the mid-point integration rule. Use of the mid-point rule with the Burlisch and Stoer extrapolation scheme saved 20–30% in time over the Runge-Kutta method. The basic equations for surface wave ray tracing in isotropic and anisotropic media were given in many books and papers, so they are not repeated here (e.g., Aki and Richards, 1980; Gjevik, 1974; Tanimoto, 1987). The analysis was done only for the isotropic case in this paper. For completeness, anisotropy should be incorporated, but as was shown in Montagner and Tanimoto (1990), the maximum amplitude of azimuthal variation of seismic velocity is less than one percent, which is small compared with heterogeneity variations near the surface. Additionally, patterns of velocity variations are not strongly affected by the incorporation of azimuthal terms (Tanimoto and Anderson, 1985; Montagner and Tanimoto, 1990), thus to the first order, negligence of azimuthal terms at surface wave periods may be justified for our data set. Of course, anisotropy is an important attribute of Earth materials and should be studied carefully in the future with a much larger data set.

3. Application to Our Data Set

The basic data set consisted of about 4,000 measurements for R1 and G1 separately. The data were mostly collected from the Global Digital Seismograph Network (GDSN) and the French network GEOSCOPE. These data were supplemented by data from the Incorporated Research Institutions for Seismology (IRIS) for earthquakes between 1990 and 1991, but this additional data represents less than 10% of the whole data set. The data were selected to cover the Earth as evenly as possible. It is not, of course,

Fig. 1. Traced rays (solid lines) and great circle paths (dashed lines). Because we collected data from minor arc data, many paths are similar to those in the top figure. When the velocity gradient perpendicular to a ray path is severe, such as a region in the subduction zone (the bottom figure), deviation of the ray path from the great circle becomes large.

possible to make it completely even, but by counting the number of rays within small blocks, we made biases in the data coverage as small as possible (Tanimoto and Anderson, 1985; Montagner and Tanimoto, 1990). Additionally, the effects of bias in coverage can be examined a posteriori in resolution kernels as we will show in a later section.

Figure 1 shows two examples of source-receiver pairs and their ray paths. Dashed lines are the great circle paths which were used in the first iteration. Solid lines represent ray paths for the second iteration. In many pairs, ray path deviations from the great circle paths remain small and are similar to the example in the top figure, which show the path across the Eurasian continent. In other cases, especially when phase velocity has a steep velocity gradient perpendicular to a ray path, significant deviation arises as in the bottom figure. This path is along the strike of trench axes, and thus velocity gradient perpendicular to the path is significant. An example like this may lead to a conclusion that ray tracing must be included in the analysis. However, ray tracing turns

Fig. 2. Reduction of total variances for Love wave data (${}_0T_{200}$). Iteration with ray tracing turned out to be unimportant for our data set, which should be related to the fact that only minor arc data were used.

out to be not so important for our data set. In Fig. 2, the variance reduction at each iteration step is shown for Love wave data, which is equivalent to the toroidal mode ${}_0T_{200}$ (period 45 s). This figure shows that iterations with ray tracing do not improve the overall fit of the present data set. We suspect that this lack of improvement with iterations is related to the fact that our data set consists of relatively short-distance, minor arc data. Use of major arc data, for example, may change this situation.

3.1 Phase velocity maps

Each phase measurement provides Eq. (5) for the unknown coefficients (s_{lm}) and collection of such data can be written as

$$Ax = b, \quad (6)$$

where elements of A are p_{lm} in Eq. (5), elements of x are s_{lm} and elements of b are measured phase deviations ($\phi_{\text{obs}} - \phi_{\text{ref}}$). We refer to the maximum angular degree as L_{max} . We solve this set of equations using damped least-squares; the solution given by

$$\hat{x} = (A^T A + \lambda^2 I)^{-1} A^T b, \quad (7)$$

where λ^2 is the damping parameter.

The least-squares approach limits the problem to the model space. We used the equivalent term “ V -space” in Fig. 3, following the notation of singular value decomposition ($A = U \wedge V^T$) (e.g., Aki and Richards, 1980). In the model space, nearly 100% variance reduction can be achieved for λ^2 smaller than a threshold value. But too small a damping parameter leads to an oscillatory solution. Figure 3 shows that variance reductions for four spheroidal modes (${}_0S_{50}$, ${}_0S_{100}$, ${}_0S_{150}$, ${}_0S_{200}$) and those for four toroidal modes (${}_0T_{50}$, ${}_0T_{100}$, ${}_0T_{150}$, ${}_0T_{200}$) start to decline at about $\lambda^2 = 100$. We selected $\lambda^2 = 200$ for our analysis, which achieved a variance reduction of about 95–99% in the model space.

The phase velocity maps for Love waves (${}_0T_{200}$) and Rayleigh waves (${}_0S_{230}$) are shown in Fig. 4(a) and (b), respectively. The maximum angular degree in these maps

Fig. 3. Variance reduction in the model space. Variance reduction as a function of the damping parameter (λ^2) is shown. Four cases of spheroidal modes are shown in the top figure and four cases of toroidal modes are shown in the bottom figure. In all cases, variance reductions begin to fall from 100% at about $\lambda^2=100$. We selected the damping parameter $\lambda^2=200$, shown by the arrows, for analysis.

is 14 ($L_{\max}=14$). Dark regions correspond to faster velocity than the global average and white regions to slower velocity. Contours are given at every 1%. In order to compare the phase velocity variations with shallow crustal effects, we show theoretical phase velocity variations at the bottom. In this calculation, both the topographic variation (ETOPO5) and the crustal thickness variation (Soller *et al.*, 1981) are taken into account.

Phase velocity variation for ${}_0T_{200}$ shows that long wavelength components of phase velocity variations are completely dominated by crustal effects. The short wavelength components show some deviation from crustal effects, and indicate that there may be additional thermal and compositional variations in the crust and upper mantle. The phase velocity maps for ${}_0S_{230}$ show much larger differences even for long wavelength components. This can be understood by the fact that the energy of spheroidal modes

Fig. 4. Phase velocity maps of ${}_0T_{200}$ (a) and ${}_0S_{230}$ (b). The theoretical phase velocity maps with topography and crustal thickness variations are shown at the bottom. Dark regions are faster velocity regions than the global average and white regions are slower velocity regions. Contours denote 1% intervals. The maximum angular degree is 14 and the maximum peak value is given at the top of each figure (in %). Long wavelength features in the Love wave phase velocity map (${}_0T_{200}$) are clearly dominated by crustal structures.

penetrates deeper than that of toroidal modes in the same period. Therefore, crustal effects for spheroidal modes are not dominant at the period of about 40 s, while they are dominant for toroidal modes.

3.2 Resolution kernels

Resolution for spherical harmonic components can be computed by

$$R = (A^T A + \lambda^2 I)^{-1} A^T A, \quad (8)$$

where the estimated solution $\hat{x}$ is related to the true x by

$$\hat{x} = Rx. \quad (9)$$

It is simple to transform this relation to the spatial domain; if the resolution kernel at a location (θ_0, ϕ_0) , $K(\theta, \phi; \theta_0, \phi_0)$, is defined by

$$v(\theta_0, \phi_0) = \int d\Omega K(\theta, \phi; \theta_0, \phi_0) v(\theta, \phi), \quad (10)$$

where $v(\theta, \phi)$ is the phase velocity at (θ, ϕ) and the integral is over the globe. Then

$$K(\theta, \phi; \theta_0, \phi_0) = \alpha(\theta_0, \phi_0)^T R \alpha(\theta, \phi), \quad (11)$$

where $\alpha(\theta, \phi)$ is a vector whose elements consist of spherical harmonics $Y_l^m(\theta, \phi)$ as

$$\alpha(\theta, \phi) = (Y_0^0(\theta, \phi), Y_1^0(\theta, \phi), Y_1^1(\theta, \phi), \dots)^T. \quad (12)$$

Examples of this kernel at two locations are shown in Fig. 5(a) and (b). The target location in Fig. 5(a) is near Himalaya and that in Fig. 5(b) is near the East Pacific Rise. In both cases, bell-shaped kernels are constructed, which supports the belief that lateral resolution in the phase velocity maps is reasonably good. Close examinations show, however, that the kernel in Fig. 5(b) is elongated to the west and the secondary peaks exist to the northwest and the southeast of the main peak. This is the bias that originates from uneven path coverage, and we will attempt to evaluate its effects in the following uncertainty estimate.

The effects of cut-off in spherical harmonic expansion appear in the oscillations of resolution kernels. When the maximum angular degree is taken as $L_{\max} = 20$, the resolution kernel near the East Pacific Rise (Fig. 5(c)) becomes much more oscillatory than the one with $L_{\max} = 14$ (Fig. 5(b)). While a narrower peak at the target location is desirable, it is obtained here at the expense of widespread short wavelength oscillations. In this paper, we chose the model with $L_{\max} = 14$ because the resulting model contains much less oscillations. Strictly speaking, however, this choice (of $L_{\max}$) is subjective and we have not found a satisfactory objective criterion. Therefore, it seems to be important not to make interpretations that can be affected by the choice of this cut-off.

3.3 Uncertainty in the phase velocity map

The covariance matrix (C_{vv}) for the phase velocity maps can be computed as

$$C_{vv} = (A^T A + \lambda^2 I)^{-1} A^T C_{\phi\phi} A (A^T A + \lambda^2 I)^{-1}, \quad (13)$$

where the covariance matrix for the observed phase ($C_{\phi\phi}$) is assumed proportional to the identity matrix ($C_{\phi\phi} = \sigma^2 I$). We estimated σ^2 from the data misfit by

$$\sigma^2 = \frac{1}{N-k} \sum |Ax - b|^2, \quad (14)$$

where N is the number of data and k is the number of parameters $(L_{\max} + 1)^2$. Using C_{vv} , spatial variation of error ($\varepsilon(\theta, \phi)$) can be computed as

$$\varepsilon(\theta, \phi) = \alpha(\theta, \phi)^T C_{vv} \alpha(\theta, \phi). \quad (15)$$

Error maps for ${}_0S_{200}$ and ${}_0T_{200}$ are shown in Fig. 6. In both cases, spatial variations are approximately between 0.3 and 0.4%, with the average of 0.35%. Contours are given at every 0.02%. The dark region corresponds to larger error than average and

Fig. 5. (a) Lateral resolution kernel near the Himalaya-Tibet region. The same kernel is shown in three-dimensional perspective (top) and in contour plot (bottom). (b) Same with (a) except that this is at a location near the East Pacific Rise. Distortion of the kernel is clear and is caused by uneven path coverage. This distortion (bias or contamination) is included in the error analysis. (c) Resolution kernel with a maximum angular degree of 20 ($L_{\max} = 20$). The target location is the same as (b). Widespread oscillations are obvious, and this led me to choose a smaller $L_{\max}$.

white region to smaller error.

There are some problems in using these error estimates. This is mainly because resolution kernels contain some degradations and are not uniform from one place to another, as Fig. 5(a) and (b) show. We attempted to take this degradation into con-

Fig. 6. Error maps for ${}_0S_{200}$ and ${}_0T_{200}$. Range of error is 0.3–0.4% for ${}_0S_{200}$ and 0.29–0.43% for ${}_0T_{200}$. They do not contain the effects of bias. We evaluated the effects of bias and added them to these errors in the final error estimates (see text).

sideration in the following way; from Eq. (10), the estimate of phase velocity at (θ_0, ϕ_0) , $\bar{v}(\theta_0, \phi_0)$ is written as

$$\begin{aligned}\bar{v}(\theta_0, \phi_0) &= \int d\Omega K(\theta, \phi; \theta_0, \phi_0) v(\theta, \phi) \\ &= \int_{\text{peak}} d\Omega K(\theta, \phi) v(\theta, \phi) + \int_{\text{outside}} d\Omega K(\theta, \phi) v(\theta, \phi), \quad (16)\end{aligned}$$

where the first term is the integration over the peak at the target location and the second term is the integral outside of that region.

The second term is the source of bias and can be bounded by

$$\left| \int_{\text{outside}} d\Omega K(\theta, \phi) v(\theta, \phi) \right| \leq |v_{\max}| \int_{\text{outside}} d\Omega |K|, \quad (17)$$

where $v_{\max}$ is the expected maximum velocity. Since the most annoying secondary peaks

Fig. 7. Total error (including bias) for all the modes in the data set. Global average values are plotted here, because geographical variations at each depth are within 10–20% of these values.

are next to the peak at the target location, approximately between 15 and 30 deg in distance from (θ_0, ϕ_0) , we estimated the bias (c) by

$$c = |v_{\max}| \int d\phi \int_{15^\circ}^{30^\circ} d\theta \sin \theta |K|. \quad (18)$$

We used $v_{\max} = 10\%$ in order to evaluate this formula.

We defined the combined uncertainty or total error (Δv) as

$$\Delta v = (\varepsilon^2 + c^2)^{1/2}, \quad (19)$$

where ε is the error estimate from least squares.

Plots of Δv for various modes are shown in Fig. 7. Typically, total error amounts to about 1% for most modes except for some low angular degree modes. In deriving these values, we neglected the spatial variations which fall within $\pm 0.2\%$, and are thus second order.

4. Depth Inversion

Phase velocity (eigenfrequency perturbation) maps were constructed for toroidal modes between ${}_0T_{50}$ and ${}_0T_{206}$ and for spheroidal modes between ${}_0S_{55}$ and ${}_0S_{234}$. We solved these phase velocity maps for the depth variation of S-wave velocity using the Backus-Gilbert method (Backus and Gilbert, 1967, 1968, 1970). The formulation for phase velocity inversion followed that of Tanimoto (1986). The results are presented in S-wave velocity variation maps; effects from P-wave velocity (for spheroidal modes)

Fig. 8. Depth resolution kernels for two cases of damping parameters, 0.1 (a) and 0.01 (b). In both cases, target depths are 50, 100, 150, and 200 km, and are denoted by arrows. We selected 0.1 (a) for the SSPS and SSPSC models. A smaller damping parameter such as 0.01 (b) introduces (undesirable) oscillations in the depth resolution kernels, as depicted by negative peaks outside the target peaks.

were included by assuming that P-wave perturbation ($\delta\alpha/\alpha$) was 75% S-wave perturbation ($\delta\beta/\beta$).

The choice of trade-off parameters affects depth resolution and error. In Fig. 8(a) and (b), we show two different cases of trade-off parameters. In the figure, the trade-off parameter is written as $tnal$ and is $tnal = 0.1$ (Fig. 8(a)) and 0.01 (Fig. 8(b)). A comparison of the two figures in Fig. 8(a) and (b) suggests that, while the smaller trade-off parameter (0.01) yields narrower peak at target depths, it leads to an oscillatory resolution kernel as depicted by the negative secondary peaks. A smaller trade-off parameter than 0.01 yields even larger oscillations outside the target peak, in particular, at shallow depths. This is related to the limitation of the frequency band in our data set. If it was possible to incorporate wider frequency range data, oscillations could be reduced while making the target peaks narrower. In this study, we chose the case $tnal = 0.1$ for the final models (Fig. 8(a)) because the lack of oscillations (relatively speaking) makes it easier to interpret the final results. Even so, the depth resolution kernels in this case (0.1) contain secondary peaks near the surface, which require some caution in the interpretation of S-wave velocity maps.

Depth variations of error are given in Fig. 9. Thick and thin lines correspond to the trade-off parameters of 0.1 (Fig. 8(a)) and 0.01 (Fig. 8(b)). Errors are laterally

Fig. 9. Error as a function of depth for the two cases of damping parameters (0.1 and 0.01). Error in this figure includes the effects of bias (defined by Eq. (19)) and is the global average value at a given depth.

varying, but variation is within $\pm 20\%$ and is of the second order. Thus, we plotted only the global average (error) values in this figure. This figure shows a well-known trend that a larger trade-off parameter (0.1) leads to smaller error estimates, and simultaneously to wider peaks at target depths.

5. Crustal Structure Correction

We retrieved two S-wave velocity maps for the trade-off parameter of 0.1. One model (SSPSC) was obtained with the topography and crustal thickness corrections (Soller *et al.*, 1981). The other model (SSPS) does not contain such corrections.

The results at four depths (50, 100, 150, and 200 km) are shown in Fig. 10(a)–(d). In all cases, the model without correction (SSPS) is shown at the top and the model with corrections is shown below (SSPSC). Contours are given at every 1% interval, and dark regions correspond to faster velocity regions than the global average as in the phase velocity maps (Fig. 4(a) and (b)). The maximum velocity (in %) in each map is given at the top of each figure. The peak to peak variations in these maps amount to 12–13% at shallow depths, although these numbers must be interpreted carefully, taking into account the widths of the depth resolution kernels.

The aim of this study was to improve the detailed global scale structure with addition of short-period surface waves. One of the points we sought to examine was whether short-period (Love) waves can replace the crustal structure correction that has been commonly used in global structure inversion. This is because, strictly speaking,

Fig. 10. S-wave velocity models, SSPS (top) and SSPSC (bottom), are shown at four depths: 50 km (a), 100 km (b), 150 km (c), and 200 km (d). The SSPSC model includes crustal thickness (Soller *et al.*, 1981) and topography correction (ETOPO5). Dark regions are faster than the global average at each depth and contours are given at 1% intervals. The maximum peak value in each map is given at the top of each figure (in %). Small filled circles are hotspot locations.

Fig. 10

Fig. 11. (a) Same as Fig. 4(a) except that $L_{\max}=4$. Long wavelength components are clearly explained by crustal and topographic effects. An exception is the region near the East Pacific Rise and Pacific Antarctica Ridge, where the upper mantle structure is causing slow velocity because of high temperature and partial melting. (b) Same with (a) except that $L_{\max}=8$. Discrepancies between the observed map and theoretical map increase from (a) ($L_{\max}=4$).

crustal structure is not known globally and there is a possibility that an incorrect crustal structure was mapped into deeper mantle structure in previous studies. As we discuss below, however, the shortest period in our data set (40 s) is not sufficient to constrain crustal structure.

The effects of crustal structure correction can be seen in Fig. 10(a)–(d) as differences between the two models SSPS and SSPSC. Major differences between the two models are found in Fig. 10(a), a velocity map just below the moho. The SSPSC model shows fast velocity under shields in this map while such patterns start at a somewhat deeper depth, about 70 km, in the SSPS model. This is clearly caused by smearing out of crustal structures in the SSPS model due to finite depth resolution.

Other places that show differences are the regions near the Mid-Atlantic Ridge. Slow velocities along the ridge axes are evident in the SSPSC model (Fig. 10(a)), while they are not necessarily slow in the SSPS model. Since S-wave velocity under the Mid-Atlantic Ridge is known to be slow, this may be one indication that the SSPSC model should be favored.

The phase velocity map for $0 T_{200}$ is dominated by crustal structure at least for its

long wavelength components (Fig. 4(a)). However, even for its long wavelength components, some effects of mantle structure are seen; Fig. 11(a) is the same as Fig. 4(a) except that the maximum angular degree in this map is 4 ($L_{\max}=4$). Overall, fast (dark) and slow (white) velocity regions match quite well between the top figure (observed phase velocity) and bottom figure (theoretical phase velocity map due to topography and crustal structure). This indicates that crustal structure has a dominant effect on ${}_0T_{200}$. There are a few regions, however, in which discrepancy becomes apparent. Such a region is near the East Pacific Rise and Pacific Antarctica Ridge. In this region, the observed phase velocity is much smaller than the theoretical phase velocity. This can be interpreted that partial melting and higher temperature under the fast spreading ridges are causing the difference (i.e., mantle structure is the source of the difference). Another region is in the western Pacific, where the observed velocity is much faster than the theoretical velocity; this reflects the fact that the old oceans have a fast velocity lithosphere (lid). The discrepancies between the observed and theoretical maps become more apparent with higher degree maps. Figure 11(b) shows maps with the highest angular degree 8 ($L_{\max}=8$). Differences are found, for example, in the Eurasian continent where the theoretical map shows a single peak dominated by thick crust in the Himalaya-Tibet region, whereas the observed map shows multiple (three) slow velocity peaks in an east-west alignment. It is not clear in this case whether the differences are caused by mantle structure or by unknown variations in the crust.

The above evidence indicates that Love waves for periods below 50 s are dominated by effects from crust, but my data, with a cutoff period of 40 s, also expressed some effects from the upper mantle. The two effects are not separable due to finite depth resolution. Further investigations, particularly with the use of even shorter period data, are required to isolate crustal effects from mantle structure (and vice versa).

6. Some Features

Since the main purpose of this paper is to present a shallow, global Earth structure rather than to discuss the tectonic implications of velocity variations, we simply point out two features that might be of interest in geodynamics.

6.1 Ridge

The existence of slow velocities along ridge axes is clear in Fig. 10(a)–(d), in particular, in the SSPSC model. Depth slice along the slice perpendicular to the East Pacific Rise (EPR, Fig. 12(a)) shows that very slow velocity anomaly, smaller than -1.5% , is confined to the upper 100–120 km under the ridge axes. This feature (confined, very slow velocity anomaly at shallow depths) is similar to the results by Zhang and Tanimoto (1993), although the peak value in this study (-3%) is more pronounced than that in Zhang and Tanimoto (1993). This difference was caused by data sets, particularly the frequency band, as well as by the choice of damping parameter in the inversion. Since shorter period data are included in this study, the results are probably more reliable as far as shallow structures are concerned. This shallow, very slow velocity anomaly should be closely related to partial melting under the ridge axes (Tanimoto and Stevenson, 1994).

Fig. 12

Slices along the EPR and Southeast Indian Ocean Ridge are shown in Fig. 12(b) and (c). Both show the confinement of very slow velocities to the upper 100–120 km. Fast velocity anomaly near the center of the Southeast Indian Ocean Ridge is an anomaly which seems to disrupt the ridge-like feature. This was also found in other studies including Zhang and Tanimoto (1993). It may be related to the Australia-Antarctica Discordant region, although the location is somewhat shifted to the west.

Fig. 12. (a) Depth slice perpendicular to the East Pacific Rise. S-wave velocity structure is generally slow under the ridge, but is particularly slow in the upper 100–120 km as depicted by the -1.5% contour. Contour interval in (a)–(e) is 0.5% . Note that only the depth range 0–200 km is reliable in our velocity maps. (b) Depth slice along the East Pacific Rise. (c) Depth slice along the Southeast Indian Ocean. (d) Location of five slices in (e). (e) Depth slice in the Himalaya-Tibet region. Note the dipping, fast velocity anomaly from right (near India) to left (under Tibet) in A. Underthrusting of the Indian Ocean lithosphere, as opposed to thickened continental lithosphere, is suggested in this figure. Under China, velocities are generally slow.

6.2 Himalaya-Tibet region

Another interesting feature is found in the Himalaya-Tibet region. Four slices along the direction of subduction (slices A–D) and one perpendicular slice (E) are given in Fig. 12(d) and (e). Slice A clearly shows a dipping, fast velocity anomaly under Tibet while velocities are generally slow under China. The fast velocity contours dipping from right to left in A suggest the existence of dipping in the Indian Ocean lithosphere, and are seen to reach 200 km with a shallow angle (about 10–15 deg). Below this depth, the model may have resolution problems, and thus interpretations should not be made. Although fast velocity under the Himalaya-Tibet region has been pointed out by other seismological studies (e.g., Holt and Wallace, 1990; Beghoul *et al.*, 1993), this figure shows the overall shape of the lithosphere from a much broader perspective.

There have been some disagreements regarding the interpretation of this fast velocity anomaly under Tibet. In one view, the fast velocity is regarded as the underthrusting Indian Ocean plate (e.g., Argand, 1924; Ni and Barazangi, 1984), while the other view interprets it as the thickened Eurasian plate (e.g., Molnar *et al.*, 1993). Although my results are relatively low in resolution and may not provide a conclusive answer, the smooth continuation of fast velocity contours from India toward deeper parts of Tibet seems to favor that this fast velocity is the sinking Indian Ocean slab, suggesting that the former view is more compatible with our results.

7. Conclusions

We analyzed surface wave data with periods from about 200 to 40 s and obtained new S-wave velocity maps for the upper 200 km of the Earth. We demonstrated that Love wave phase velocity maps for periods below 50 s show strong effects from crustal structures. The two maps with (SSPS) and without (SSPSC) crustal structures and topography correction show that structures differ mostly just below the moho and agree fairly well for depths below 100 km. The SSPSC model shows the beginning depths of tectosphere-like fast velocity anomaly just below the moho, while the SSPS model shows the beginning depths at a somewhat deeper depth (about 70 km). This is caused by smearing out of the crustal structures in the SSPS model due to finite depth resolution, and shows that my data set (the shortest period of 40 s) is not sufficient to resolve crustal structures. Structures under ridge axes show the existence of slow S-wave velocity, particularly very low velocity anomaly (less than -2%) confined to the upper 100–120 km. Slices along the Himalaya-Tibet region show the existence of a dipping fast velocity anomaly; the shape of contours suggests that this is the dipping Indian Ocean plate rather than a thickened Eurasian plate.

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APPENDIX

Spherical Harmonic Approach vs. Block Approach

The phase velocity maps reported in the main text were obtained by the spherical harmonic approach. The data (measured phase shifts) were solved for the spherical harmonic coefficients of phase velocity variations. Naturally, there arises a question on how reliable these maps are, because the series must be truncated in spherical harmonic expansion. Therefore, in order to examine the reliability of the results, we solved the same problem using an alternative, block scheme. By using a different approach, we examined whether the retrieved maps were reproducible or not.

We adopted the block scheme shown in Fig. A1, which is $5^\circ \times 5^\circ$ in size near the equator. Blocks at higher latitudes were modified to maintain approximately the same area. In each block, phase velocities were assumed to be constant for which the data were solved for. After the inversion, the results were expanded into spherical harmonics and compared with the results obtained by the spherical harmonic approach.

Fig. A1. Block scheme adopted for the block inversion. Each block is divided into approximately the same area. Near the equator, a block is about $5^\circ \times 5^\circ$.

Fig. A2

Fig. A3

Fig. A2. Comparison of the spherical harmonic inversion (top) and the block inversion (bottom) for the toroidal mode ${}_0T_{150}$.

Fig. A3. Same as Fig. A2 except for mode ${}_0S_{150}$.

Figures A2 and A3 show a comparison of the two inversion schemes for ${}_0T_{150}$ (Fig. A2) and ${}_0S_{150}$ (Fig. A3). The figures demonstrate that the two approaches produce extremely close results. Major features in the patterns, especially dark and white regions, are virtually the same, and differences arise only in the detailed patterns of the contour. It seems safe to say that most of the features in the results are independent of the methods that were used to retrieve them.

Lateral resolutions of the data set for the phase velocity maps were also retrieved by the block inversion approach. In Figs. A4–A9, resolution kernels at six locations are plotted in global maps. In each case, a sharp peak is found at the target block, showing fairly even resolving power of the data set. Differences in six cases are mainly found in the trade-offs among adjacent blocks, which seem to be reasonably small in all cases. Overall, these kernels support our contention that resolution over the globe is achieved in a fairly uniform manner.

Fig. A4

Fig. A5

Fig. A6

Fig. A7

Fig. A8

Fig. A9

Figs. A4-A9. Resolution kernels by block inversion. Lateral resolution is fairly uniform over the globe.