Spatial distribution of lateral heterogeneity in the upper mantle around the western Pacific region as inferred from analysis of transverse components of teleseismic P-coda

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Spatial distribution of lateral heterogeneity in the upper mantle around the western Pacific region as inferred from analysis of transverse components of teleseismic P-coda

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[1] We investigated the spatial distribution of lateral heterogeneity in the upper mantle in the western Pacific region analyzing transverse components of teleseismic P-coda recorded by broad-band seismic networks. Large amplitude transverse components are observed at stations close to plate boundaries such as island arcs for the frequency range of 0.04–2.56 Hz, while small amplitudes are observed at stations on stable continents. Our envelope inversion applied to the observed P-coda reveals large scattering coefficients at depths from 100 to more than 300 km beneath the island arcs, which are interpreted as strong heterogeneities originating from subducting slabs and mantle diapirs. Analyses of a dense seismic network further show a significant difference of heterogeneity between East and West Japan: the former is more heterogeneous than the latter. In contrasts, scattering coefficients less than 20% of the maximum beneath the island arcs are obtained beneath the stable continents, which represents homogeneous and transparent upper mantle.


1. Introduction

[2] Seismic coda analyses have recently succeeded in detecting spatial variation of random heterogeneity of the earth, which is not captured by conventional tomography methods using travel times alone. The obtained spatial changes are well correlated with the tectonic settings such as island arcs [Obara and Sato, 1995], active seismic faults [Nishigami, 2000] and volcanic processes [Nishimura et al., 1997]. However, most of these results are based on analyses of regional seismic data, and few systematic studies on world wide distributions of random heterogeneity have been reported although spatial changes are clearly recognized between island arcs and stable continents [e.g., Korn, 1993; Nishimura, 1996]. In this study, therefore, by analyzing teleseismic P-coda recorded by broad-band seismic networks that have been recently deployed around the western Pacific region, we estimate the spatial distribution of the strength of lateral heterogeneity in the upper mantle and compare the results among the various tectonic settings.

[3] We analyze three-component broad-band seismic data recorded by the Japan-Indonesia Seismic Networks (JISNET, Ohtaki et al., 2000) in Indonesia for the period from 1998 to March 2001, F-net in Japan from 1995 to 2000, and IRIS from 1987 to 2000, which are digitized with a sampling frequency ranging from 20 to 50 Hz and an A/D resolution of 20–24 bits. Only teleseismic P-coda waves from pairs of stations and deep earthquakes (>300km, M5-7) whose epicentral distances are less than 60 degrees of arc are analyzed to avoid contamination by depth phases and converted phases such as PcP [Korn, 1993]. Although shallow earthquake data may be useful for our analyses by applying deconvolution techniques that can eliminate complex source time functions and depth phases, this study is a first attempt so that we only use the deep earthquake data having a simple source time function. We further confine our data set to the seismograms whose amplitude ratio of direct P-wave to noise level is larger than 20. Since previous theoretical and analytical studies indicate frequency dependent scattering waves propagating in the earth structures [e.g., Sato and Fehler, 1999], we analyze band-pass filtered (0.04–0.08, 0.08–0.16, 0.16–0.32, 0.32–0.64, 0.64–1.28 and 1.28–2.56 Hz) seismograms in the followings.

2. Spatial Variation of the Amplitude of Transverse Components

[4] Seismic signals appearing in the transverse component of teleseismic P-coda of deep earthquakes are primarily...
generated from lateral heterogeneity of the crust and upper mantle beneath the recording station. We therefore evaluate the amplitude of the transverse component of teleseismic P-wave at each station as an indicator of the regionality of lateral heterogeneity. First, for each pair of station and earthquake, we calculate the seismic energy, $E_T$, in the transverse component and the total seismic energy of the

![Figure 1](opposite) Spatial distributions of the transverse amplitude of teleseismic P-waves for the frequency bands of 0.04–0.08 Hz and 1.28–2.56 Hz. Red and blue colors represent transverse amplitudes larger and smaller than the mean (0.15 for 0.04–0.08 Hz and 0.43 for 1.28–2.56 Hz) for all data, respectively. Each symbol represents the region where the station is located. A seismicity map is also shown to indicate tectonically active regions (the data is based on the catalogues of the International Seismological Centre (ISC) from 1990 to Feb. 2000, 5 < M < 7). Note that several stations do not show the transverse amplitudes for both low and high frequency bands because the data at these stations is not available together with a good signal to noise ratio for the two frequency bands.

![Figure 2](Spatial distribution of transverse amplitudes of teleseismic P-waves in Japan for the frequency bands of 0.08–0.16 Hz and 0.64–1.28 Hz with a seismicity map (ISC data from 1990 to Feb. 2000, 5 < M < 7). Brown triangles in the bottom figure represent active volcanoes in Japan.)
three components, $E_A$, for 60 s from the P-wave onset. The transverse direction is calculated from the locations of hypocenter and station. The ratios, $ET/E_A$, are then averaged for all events at each station, and we obtain the averaged transverse amplitudes (hereafter referred to simply transverse amplitude) for each station by taking the square root of the ratio.

Figure 1 shows the spatial distribution of the transverse amplitude at low and high frequency ranges (0.04–0.08 Hz and 0.08–0.16 Hz). The envelopes are stacked together by adjusting the theoretical P-wave arrival time (30 s in the figure) of each event. Broken lines represent the best-fit synthetic envelopes obtained by our inversion. Note that the vertical axes represent the normalized amplitude of the envelopes (squared velocity) and that the amplitude of the upper four traces are a half of those of the West Asia and West Japan and one-quarter of those of the East Japan and Indonesia. The number on the bottom left of each trace represents the number of stacked seismograms.

Figure 3. Observed envelope seismograms (fine lines) of the transverse components at the stations located in Australia, East, North and South Eurasia, West Asia, West Japan, East Japan and Indonesia for the frequency bands of 0.04–0.08 Hz (left) and 0.08–0.16 Hz (right). The envelopes are stacked together by adjusting the theoretical P-wave arrival time (30 s in the figure) of each event. Broken lines represent the best-fit synthetic envelopes obtained by our inversion. Note that the vertical axes represent the normalized amplitude of the envelopes (squared velocity) and that the amplitude of the upper four traces are a half of those of the West Asia and West Japan and one-quarter of those of the East Japan and Indonesia. The number on the bottom left of each trace represents the number of stacked seismograms.

Figure 4. Depth distributions of the scattering coefficients for the frequency bands of 0.04–0.08 Hz (solid lines) and 0.08–0.16 Hz (dotted lines). Scattering coefficients are plotted at the middle depth of each layer as open circles with error bars (one standard deviation).
large amplitudes are observed at some stations located in Kyushu, where several active volcanoes are located and the Philippine Sea plate bends downward around the depth of 80 km.

3. Depth Variation of Scattering Coefficients in the Upper Mantle

[7] Temporal variations of the amplitude of envelope seismograms (squared velocity) can further give us information on where strong lateral heterogeneity exists. Figure 3 shows the observed envelope seismograms of the transverse component of teleseismic P-coda. The envelope seismograms are stacked together for the stations located in the same areas (Indonesia, East Japan, West Japan, West Asia, North, South and East Eurasia and Australia) after being normalized by the maximum amplitude of the vertical component for each envelope. Only the events whose seismic-ray passes below 500 km and whose incident angles are in the range of 20 to 40 degrees are used. The envelope seismograms at Indonesia, East and West Japan, and West Asia, where intermediate-depth and deep earthquakes are observed, continue with large amplitude for 60 s after the P-onset. On the other hand, the amplitudes of envelope seismograms observed on stable continents (North and South Eurasia, East Eurasia, and Australia) are small.

[8] To quantitatively evaluate the heterogeneity, we estimate the depth variation of scattering coefficients [e.g., Sato and Fehler, 1999] applying the envelope inversion method of Nishimura et al. [1997] to the observed envelopes. We assume that the teleseismic P-wave is obliquely incident from depth as a plane wave with an apparent incident angle of 30 degrees that fits well with the observations. The incident P-wave is singly and isotropically scattered by lateral heterogeneity in the crust and upper mantle, and the P-wave reflected at the free-surface is also singly-scattered in the crust and upper mantle. PP, PS, SP, and SS scattering processes are considered in the calculation. We assume a velocity structure having three layers that are obtained by smoothing the standard seismic velocity model (iasp91), and calculate travel times and geometrical spreading factors. Poisson’s ratio within each layer is assumed to be 0.25. We divide the crust and upper mantle into six layers (0–80 km, 80–160 km, 160–240 km, 240–320 km, 320–400 km, 400–480 km) each having a thickness of 80 km, and then, we numerically calculate the Green’s function for scattering within each layer. We assume the mantle below 480 km to be homogeneous because our preliminary analyses for some regions with sufficient seismic data indicate a small scattering coefficient below 480 km. Fitting the synthetic envelopes with the observed ones by a non-negative least squares method, we obtain the depth variation of the scattering coefficients. Since multiple scattering is generally dominant for high-frequencies, we apply this method to the observed envelope only at low-frequency bands (0.04–0.08 and 0.08–0.16 Hz). In Figure 3, we compare the theoretical envelopes calculated from the best-fit solutions with the observed envelopes. Overall characteristics of the observed envelopes are well reproduced by the theoretical envelopes. The poor fit found around the onset of the P-wave may be due to our simple assumption about the scattering radiation pattern (isotropic scattering) and variations of the source time function of each earthquake. Figure 4 shows the depth variation of the scattering coefficients for each region. Note that our inversion can estimate only relative (not absolute) values of the scattering coefficients [see Nishimura et al., 1997] and that the scattering coefficients are normalized by the maximum value obtained at East Japan. Large scattering coefficients are obtained at depths from about 100 to more than 300 km beneath island arcs (East Japan and Indonesia). These strong lateral heterogeneity is probably related with subducting slabs and magmatic processes, which are detected as high (+6%) and low (−6%) velocity zones from conventional tomography methods [e.g., Nakajima et al., 2001] as previously mentioned. The largest scattering coefficient in West Japan and West Asia is observed at 100–200 km, which is close to the depths of deepest hypocenters associated with the subducting plates. The scattering coefficients obtained at Australia, North, South and East Eurasia are less than 20% of the maximum of East Japan. That is, the upper mantle is transparent beneath the stable continents although weak lateral heterogeneity is detected at depths of more than 300 km under Australia and East Eurasia.

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