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Upper mantle velocity structure in the western Pacific rim estimated from short-period recordings at Matsushiro Seismic Array System

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We investigate lateral variation of the upper mantle *P*-wave velocity structure in the western Pacific rim using travel times of *P* phases from regional shallow events. Onsets of direct and later arrivals are manually picked up from short-period waveforms at Matsushiro Seismic Array System in central Japan by consulting the correlation among the waveforms at seven array stations. For the path along Kurile and Alaska seismic zones, observed first arrivals of shallow events ($d = 5$–50 km) are well explained by a standard travel time table for Japan. Observations provide no strong evidence of a distinct low-velocity zone in the uppermost mantle, and triplicated branches are adequately explained by a slight modification of model 12 by Sugiyama and Nakanishi (1989, 1999). Observed travel times of direct *P* phases from events along the eastern and western Philippine Sea are slightly larger at around 15° and slightly smaller at around 25°, and the upper mantle is slower and the transition zone and uppermost lower mantle faster in the eastern and western Philippine Sea than in the northwestern Pacific. Such differences could reflect a high temperature in the upper mantle, and accumulation of subducted material at the base of transition zone as a result of the past subduction.

1. Introduction

The most direct information of the velocities in the upper mantle can be derived from travel times and apparent velocities of seismic phases. An abrupt change of apparent velocity of direct *P* phase (e.g., Kishimoto, 1956) and triplicated arrivals in regional distances (e.g., Niazi and Anderson, 1965; Walck, 1984) have been used to construct upper mantle velocity models. Several velocity models have been presented for the western Pacific region, some of which are constructed with measurements of travel times and apparent velocities of *P* phase (Kanamori, 1967; Fukao, 1977; Inatani and Kurita, 1980; Sugiyama and Nakanishi, 1989 and 1999), some are by direct waveform analyses (Erdögan and Nowack, 1993; Yamazaki and Hirahara, 1996).

Two major issues concerning the upper mantle velocity structure in oceanic regions are properties of the low-velocity zone in the uppermost mantle, and depths to and magnitudes of velocity jump at the transition zone discontinuities. The low-velocity zone appears to be an indispensable feature in the oceanic region (e.g., Turcotte and Schubert, 1982; Sato *et al.*, 1989), and its possible absence beneath the matured northwestern Pacific basin (e.g., Sugiyama and Nakanishi, 1989 and 1999) would be an important constraint on the potential temperature in the convective oceanic mantle, though recent evidences strongly suggest that the origin of the low-velocity zone may not be solely thermal (Karato, 1995; Hirth and Kohlstedt, 1996; Gaherty *et al.*, 1999; Kato and Jordan, 1999). Rapid increases of velocities approximately at 410 km and 660 km depths are often interpreted as signature of phase transitions of the olivine species (Ringwood, 1975; Bina, 1991). Lateral variation of their depths and magnitudes of velocity jump might be used to constrain the mineralogy and the temperature of the upper mantle.

We investigated the velocity structure of the upper mantle beneath the western Pacific rim by studying travel times of direct *P* phase from regional events. Variations of the upper mantle velocities among regional models are larger than those in global tomographic models (e.g., Fukao *et al.*, 1992; Su *et al.*, 1994), and regional models are needed to further understand the lateral variation of physical properties in the mantle (Nolet *et al.*, 1994). Regional variation has been often discussed by comparing models that are derived using different types of techniques and different types of data (Nolet *et al.*, 1994). One of the scopes of this study is to investigate how much we can infer regional variation from short-period recordings by equalizing characteristics of data and analysis technique.

Our data sets comprise travel times of *P* phases which are measured from waveform data recorded at Matsushiro Seismological Array Systems (MSAS) in Japan. MSAS is located approximately at the intersection of three subduction zone arcs, Kurile-Alaska, Izu-Mariana, and Ryukyu-Philippines, and the data from MSAS would be suitable to investigate variation of upper mantle velocity structure in these three regions. Observed travel times of the first arrivals in three regions are consistent with a standard travel time table in Japan (Hamada, 1984). We present a one-dimensional *P* wave velocity model for the Kurile-Alaska

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Table 1. Stations of MSAS.

Code	Latitude (°N)	Longitude (°E)	Dep	Elev	Vp	T
TKM	36.560	138.242	46.7	598	6.1	1.00
JZO	36.522	138.250	43.3	616	3.6	1.02
SGD	36.539	138.319	75.4	1231	5.3	0.98
DIR	36.482	138.298	53.0	838	4.8	1.02
IRK	36.468	138.254	43.3	761	4.1	0.97
MAT	36.543	138.207	0.0	406	6.0	1.01
WDR	36.487	138.214	65.0	961	4.3	1.01

Dep: Depth of sensor measured from surface (meter). Elev: Elevation of sensor (meter). Vp: Velocity of boring sample (km/sec). T: Characteristic period of vertical component sensor (second). From Osada *et al.* (1984).

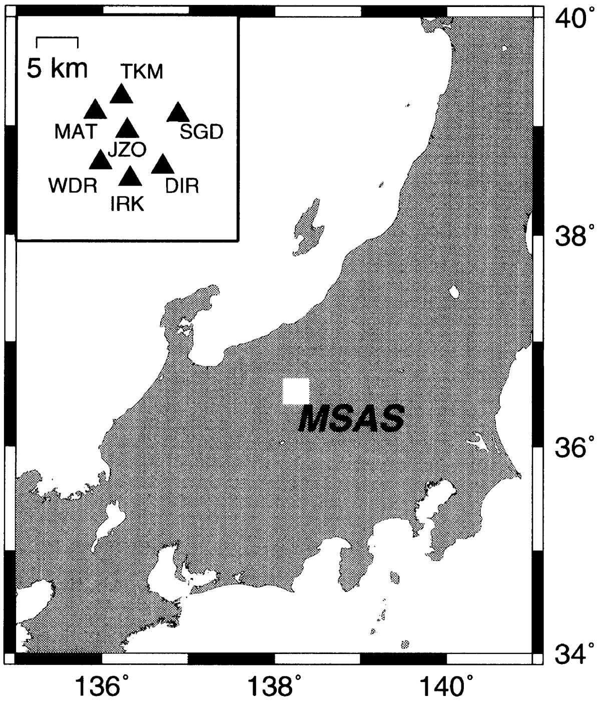

Fig. 1. Location of Matsushiro Seismic Array System. See also Table 1. Inset map shows the location of array stations. Sensors are located in borehole at all stations, except MAT at which sensors are located in a horizontal vault.

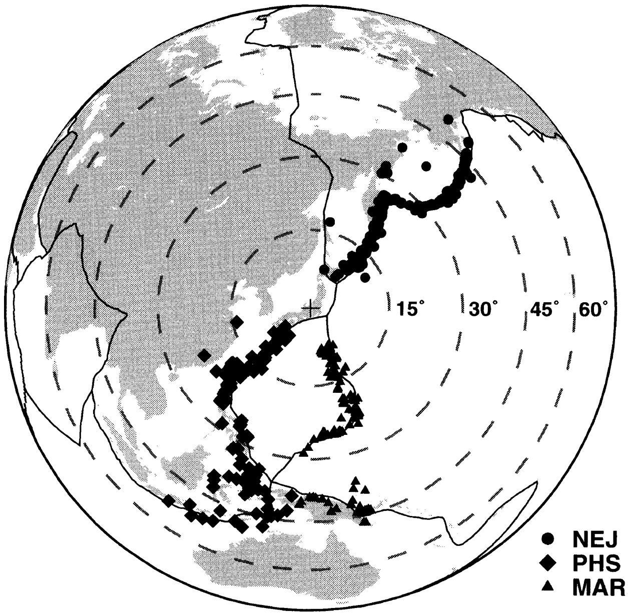

Fig. 2. Location of events. Circles, diamonds, and triangles denote events for NEJ, PHS, and MAR paths, respectively. Dash lines represent equidistant lines for epicentral distances of 15, 30, 45, and 60 degrees. Orthographic projection is used, the center of projection being at MSAS.

path. Observed travel times of initial *P* arrivals for two Philippine Sea paths differ slightly from the prediction of our Kurile-Alaska model, and this difference should indicate that the upper mantle beneath these Philippine Sea paths is characterized by the slow and possibly heterogeneous uppermost mantle and the fast transition zone and uppermost lower mantle.

2. Data

MSAS, located in central Japan (Fig. 1), is operated by Japan Meteorological Agency (Osada *et al.*, 1984). MSAS consists of seven stations and an average distance between stations is approximately 5 kilometers (Fig. 1, Table 1). Each station is equipped with an identical three-component velocity-type short-period sensor, whose characteristic frequency is approximately 1 Hz, responses being flat between 2 and 10 Hz (Table 1). Sampling frequency is 80 Hz (Osada *et al.*, 1984). Sensors at MAT and other six stations are installed in the horizontal vault and the boreholes at 40 to 70 meters depth (Table 1), respectively. We use records of vertical component.

Surface geology beneath the array stations appears to be very heterogeneous; the compressional velocities measured with the drilled samples vary from 3.6 to 6.1 km/s (Osada *et al.*, 1984), and a refraction survey revealed substantial topography of subsurface layer boundaries in the crust (Asano *et al.*, 1969). Simple two-layer crust models have not been fully successful in explaining observed travel time residuals within the array (Nagai and Kashiwabara, 1985; Maki *et al.*, 1987). Complexity of the surface and topography geology results in strong site effects at array stations (Kato *et al.*, 2000), which hinder application of conventional array techniques, such as delay-sum stacking or semblance analysis. We then assume MSAS as a network with small inter-station distances, and arrivals of later phases are visually inspected and manually picked up considering the correlation among waveforms.

Source regions in our study are three arcs in the western Pacific rim. The northeastern Japan (NEJ) path spans the Kurile arc to the Alaska seismic zones, the western Philippine Sea (PHS) path does the Ryukyu-Taiwan-Philippine arc, and the Mariana (MAR) path does Izu-Bonin to Mariana arc and the New Guinean seismic zones (Fig. 2). High-frequency recordings of shallow events (d = 0–100 km) are used in this study. Event sizes are between mb 5.0 and 6.8. Quality and quantity of NEJ data set are best among them and we mainly focus on this data set in the study. Waveforms of several hundred events are initially selected for each path, and quality of waveforms is visually evaluated based on signal-to-noise ratio in the raw records, and then based on the correlation among seven waveforms. Onsets of direct P phase are picked up manually from the filtered records with a band of 0.1–2.5 Hz (Fig. 3). We did not deconvolve the empirical source time function (e.g., Yamazaki and Hirahara, 1996), considering the limited bandwidth in our data and probable contamination of site effects (Kato *et al.*, 2000); correlation among these regional and teleseismic waveforms becomes low after a couple of cycles following the onset of P phase, even though the inter-station distances are small. Examples of array waveforms for the NEJ dataset are given in Fig. 4.

Strong site effects also hinder easy identification of later arrivals. In previous studies (e.g., Fukao, 1977; Inatani and Kurita, 1980; Sugiyama and Nakanishi, 1989 and 1999), apparent velocities are measured in the record sections that span more than a couple of degrees, which are then used to identify later phase branches. Due to a small diameter of MSAS, which is approximately 10 km, it is difficult to conclusively identify any branches with waveforms from a single event solely based on the difference of apparent velocities. Our approach in this study is to visually examine the correlation of the waveforms and identify later arrivals. We also compare measurements for events with similar epicentral distances to identify possible branches. Such approach should be effective in reducing the possibility that an incoherent arrival of energy due to scattering beneath the array would be erroneously identified as a phase from the deep upper mantle.

All travel time measurements are corrected for the source depth using a standard velocity model for the Japanese Islands (Hamada, 1984). Due to a larger scatter in the depth-corrected travel times, and we first excluded data from the events whose depths are deeper than 50 km. Our measurements are mostly well explained by this standard travel time table (Fig. 5), except for data from the shallowest events (0–5 km). We speculate that this disagreement is due to the mislocation of the shallow events, specifically of the source depth. Relocation of these events would not solve the problem unless arrival times of P and depth phases are reexamined with the original waveform data. We therefore removed all waveforms of shallowest events from the dataset. Final dataset includes picks from 276 events for the NEJ path, whose depths are thus between 5 and 50 km.

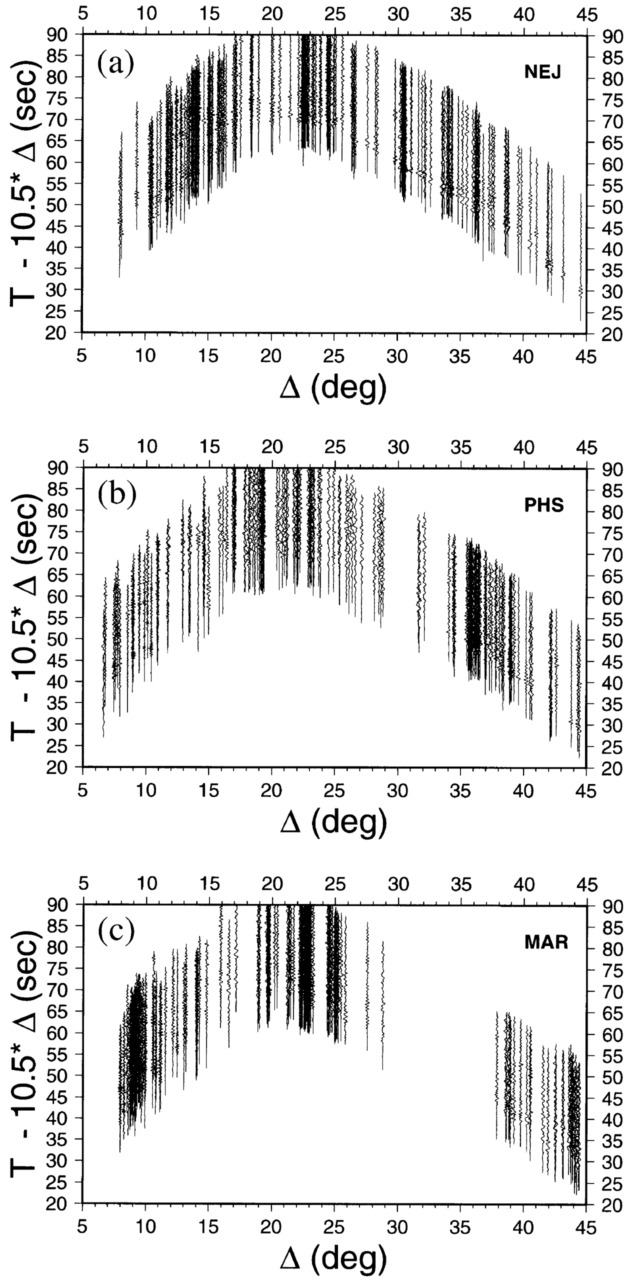

Fig. 3. Record section of observed waveforms at station MAT for NEJ (a), PHS (b), and MAR (c) paths. Each trace is filtered (0.1–2.5 Hz) and is normalized by its respective maximum amplitude. Reduction slowness of 10.5 second/degree is used.

3. Model for the NEJ Path

Travel times of P phase from regional events alone do not constrain velocities in the uppermost mantle tightly, and several assumptions on crustal and lithospheric structure and velocities are needed to be employed. We first examine several published models for the Kurile-Alaska path (Kanamori, 1967; Fukao, 1977; Erdögan and Nowack, 1993; Yamazaki and Hirahara, 1996; Sugiyama and Nakanishi, 1989 and 1999) as well as those for the other part of Pacific (Walck, 1984 and 1985; Gaherty *et al.*, 1996), which are tested whether they could explain several possible branches that are

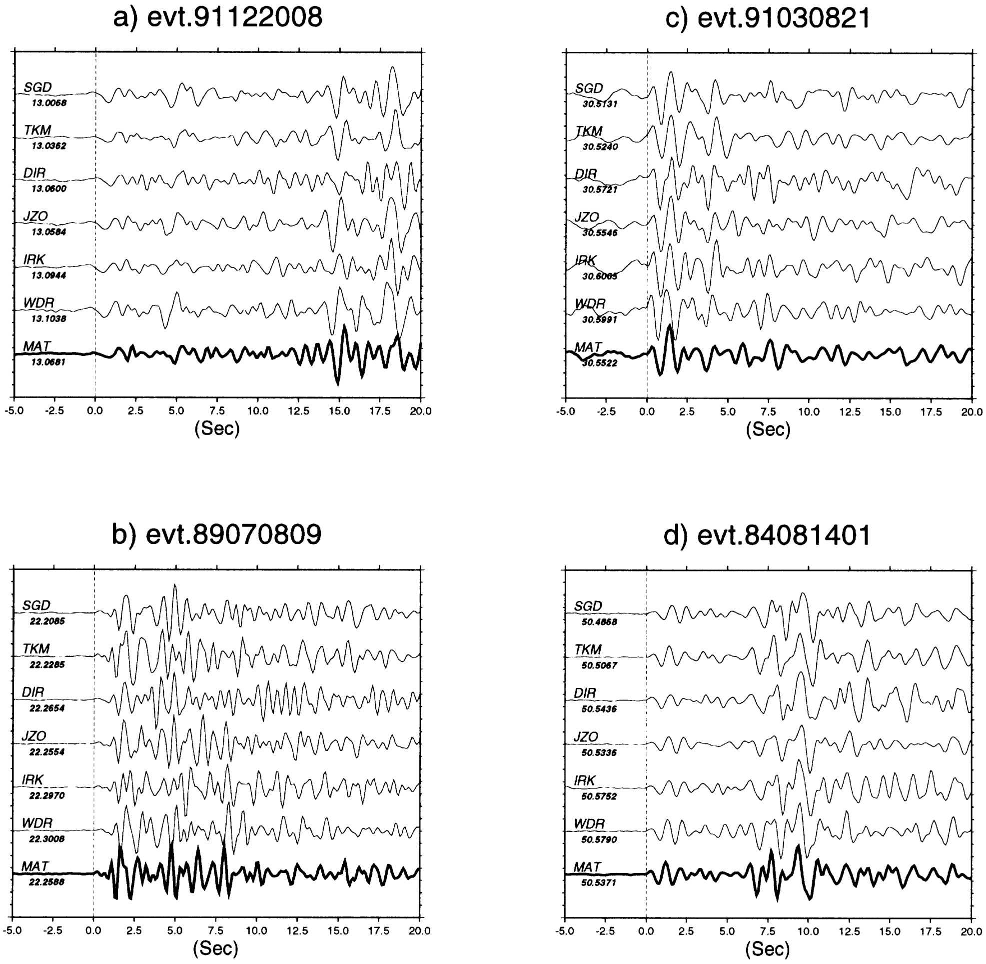

Fig. 4. Example waveforms for the NEJ dataset. Traces are plotted on the same scale for each event. The origin of the plot is set at the *P* arrival time for each record. Epicentral distances are plotted beneath the station code. a) Records from an event at approximately 13 degrees from MSAS (45.13N, 151.25E, 48 km). The phase at 14 sec. is example for the A branch in Fig. 6(a). b) Records from an event at approximately 22 degrees (52.87N, 159.78E, 18 km). c) Records from an event at approximately 30 degrees (60.91N, 167.29E, 10 km) d) Records from an event at approximately 50 degrees (61.81N, 149.11W, 13 km). Even for such teleseismic distances, the correlation of the waveforms within the array is low, which indicates that effects of the sites cannot be negligible at MSAS (Kato *et al*., 2000).

identified on the observed Δ-T plot as well as travel times of direct *P* phase. Our approach differs from those in previous studies in that the observed travel times are not binned according to their epicentral distances or their relative travel times from direct *P* phase; proper binning of data points should require good prior knowledge on the location of triplicated branches on the Δ-T plot. Observed travel time curve appears to be continuous between 10 and 15 degrees. Large scatter in these distances should be related to the heterogeneous structure in the uppermost mantle, and observations could be explained with a model without a distinct low-velocity zone in the uppermost mantle. Model 12 of Sugiyama and Nakanishi (1989, 1999), which is a slight modification of ARC-TR (Fukao, 1977) and is constructed for the path connecting the Kurile-Alaska arc and the stations in Hokkaido, explains travel times of direct *P* phase as well as some later phases fairly well (Fig. 6(a)), though the small difference between the prediction and observation is recognizable in the distance range between 25° and 35°.

Three clusters of later arrivals are not at all well predicted by Model 12 and appear to be out of place. Two of them are located in the typical epicentral range of the triplication that are caused by the discontinuities in the mantle transition zone; slownesses are as large as 10.7–10.8 sec/degree for cluster A between 13 to 17 degrees, and as small as 9.8–9.9 sec/degree for cluster B between 23 and 27 degrees. Associating these branches with direct *P* phases requires velocities in the transition zone to be inconsistent with the constraint from the mantle mineralogy (e.g., Ita and Stixrude, 1992), and is thus quite difficult. These "branches" should be associated with depth phases, such as *pP* and *sP*. Considering the range in the event depths in our data and the strategy of identification of later phases, the possibility of identifying false branch remains, and these linear alignments should be purely coincidental and artificial, and would not be related to the triplication of direct *P* phase. The third one is located at approximated at 30° (C in Fig. 6), which would also be related to depth phases or scattered wavefield excited in the source region.

Model sn12c is our model for the Kurile-Alaska path (Fig. 7, Table 2), which is a slight modification of Model 12 (Sugiyama and Nakanishi, 1989 and 1999). Since our observations are adequately well explained by Model 12, we try to keep the structural features of Model 12, such as depths

Table 2. Model sn12c.

Depth (km)	Vp (km/s)
0	5.57
15	5.57
33	6.50
33	7.84
95	8.10
165	8.10
200	8.36
205	8.41
225	8.40
240	8.40
260	8.41
280	8.44
300	8.47
320	8.52
340	8.57
360	8.64
385	8.74
400	9.27
520	9.72
660	10.23
670	10.29
690	10.39
707	10.83
800	11.03
900	11.19
1000	11.34
1200	11.64

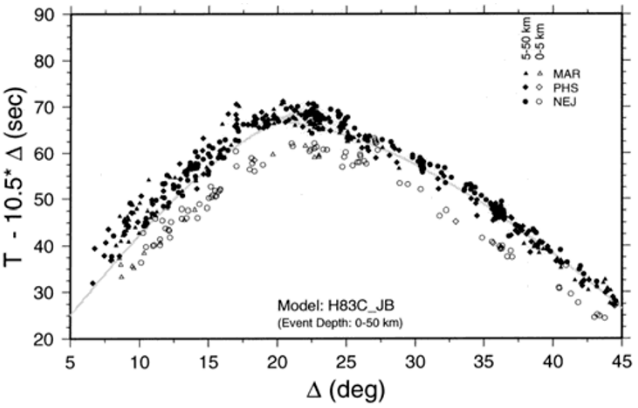

Fig. 5. Picked travel times of the first arrivals. Circles, diamonds, and triangles represent data from NEJ, PHS, and MAR data set, respectively; open and filled symbols are of shallow ($d = 0$–5 km) and deep ($d = 5$–50 km), respectively. Solid line indicates *P* arrival times for the standard time table of Hamada (1984). Note that there is an offset of approximately 5 seconds for some data points, all of which are of shallow (0–5 km) events.

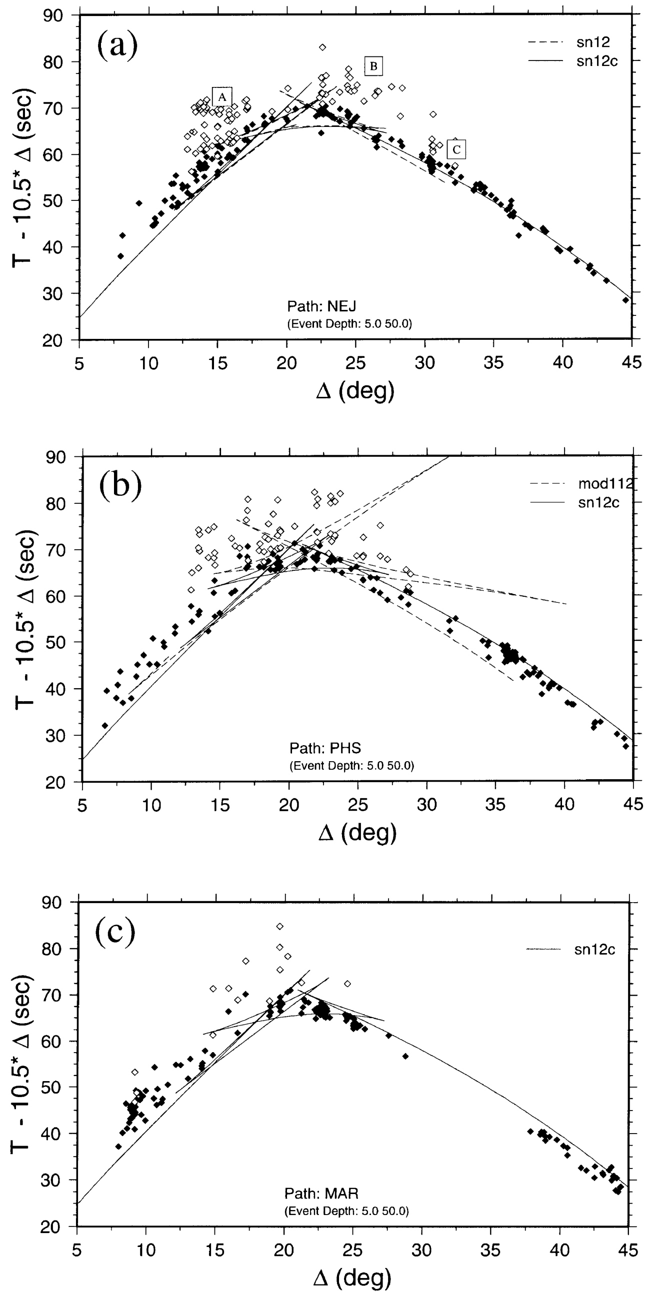

Fig. 6. All picks of travel time for NEJ (a), PHS (b), and MAR (c) data sets. Data from events whose depths are between 5 and 50 km are plotted. First arrivals are shown with filled diamonds, and later arrivals are shown with open diamonds. Also plotted is the predicted travel time curve for (a) model 12 (Sugiyama and Nakanishi, 1989 and 1999), and model sn12c, (b) model 112 (Inatani and Kurita, 1980), and model sn12c, and (c) model sn12c. Inset letters, A to C, in (a) signify clusters of data points which appear not to be well explained by model 12. See text for details.

of discontinuities, as much as possible, and made only three minor modifications. First, the velocity is slightly lowered between 185 and 400 km. This is to explain travel times of the first arrivals between 15 and 20 degrees. Second, the velocity gradient between the 410-km and 660-km discontinuities is simplified. Sugiyama and Nakanishi (1989, 1999) used direct measurements of slowness at 20 degrees to constrain velocity gradients in the transition zone. Our observations are made at the surface and events are shallow, and do not provide such constraint on the deep fine structure in the transition zone (e.g., Tajima and Grand, 1995; Brudzinski *et al.*, 1997), and our preference is a simple model. The third modification is the reduction of velocity jump of the 660-km discontinuity (at approximately 700 km) and an increase of velocity gradient in the uppermost lower mantle. This modification is required to match travel time in the distance range of 25 degrees and beyond. Though the observations show some scatter, model sn12c explains the observations well on average (Fig. 6(a)).

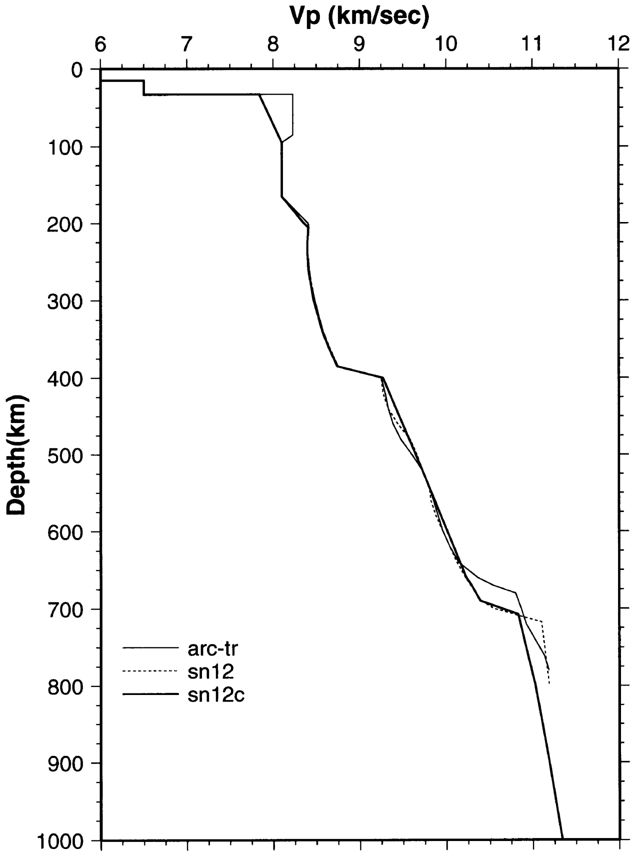

Fig. 7. Our velocity model for the NEJ path, sn12c. Also plotted are model 12 (Sugiyama and Nakanishi, 1989 and 1999: dotted line), and ARC-TR (Fukao, 1977: dash line). Model 12 is slightly modified from ARC-TR, and sn12c from model 12.

4. Results for the Philippine Sea Regions

Quantity and quality for the PHS and MAR datasets are lower than that for the NEJ, and we do not attempt to model velocity structures with any respective dataset. Signal-to-noise ratios of observed waveforms are lower in these two datasets and observed travel times show larger scatters as well (Figs. 6(b), (c)). Moreover, there is a gap in data, and we do not have observations at critical distances for triplicated branches. We then compare observations for these two paths with the predicted travel times from model sn12c to find the regional variations. Inferences from such comparison will be free from the assumptions that we use in modeling the velocity structure, and we could discuss the difference in structure in the upper mantle qualitatively. Data are processed in the similar manner as for the NEJ data set, and final datasets include picks from 175 events for the PHS path, and 172 events for the MAR path.

Compared to the Kurile-Alaska path, the Philippine Sea and the Mariana paths appear to be characterized by slightly lower velocities in the uppermost mantle and higher velocities in the transition zone and in the uppermost lower mantle. Although the scatter in the observation is large, predicted travel time from model sn12c is smaller than observations between 10 and 15 degrees, and larger between 30 and 45 degrees (Figs. 6(b), (c)). Slow upper mantle for the Philippine Sea is consistent with the findings in the low-frequency seismology (Katzman *et al.*, 1998; Kato and Jordan, 1999) as well as global tomographic models with high-frequency phase delays (e.g., Fukao *et al.*, 1992). High velocities in the transition zone beneath the Philippine Sea have been reported in the one-dimensional model (Kato and Jordan, 1999) as well as in three-dimensional tomographic models (Suetsugu and Nakanishi, 1987; van der Hilst *et al.*, 1991; Fukao *et al.*, 1992).

5. Discussion

Existence of a low-velocity zone beneath the matured oceanic basin has been the focus of several decades of researches in the northwestern Pacific region. While some models in this region have well-established low-velocity zones (Fukao, 1977; Erdögan and Nowack, 1993), model sn12c as well as its mother model 12 of Sugiyama and Nakanishi (1989, 1999) does not have a distinct lid; velocity increases smoothly between Moho and the 410-km discontinuity. Strong signature of the low-velocity zone such as shadow zone or discontinuous jump in travel time curve is not seen in our observations. Shallow structure is constrained by observations in the shortest epicentral range, but scatter in observations are large in this epicentral range. Sugiyama and Nakanishi (1989, 1999) showed that incoming directions of P phases do not always propagate along the great circle that connects the network and events in such epicentral ranges, suggesting effects of lateral heterogeneity. Although the scatter in the data inhibits us from strongly concluding so, our observations do not appear to imply a well-established lid-low-velocity-zone contrast. A high-gradient zone above the 410-km discontinuity is seen in model sn12c (Fig. 7). This zone characterizes the oceanic upper mantle, and appears to be a signature of relatively high geotherm (Gaherty *et al.*, 1999).

Model sn12c presented here is not a unique model and should be treated as a representative of a class of models. Sugiyama and Nakanishi (1989, 1999) performed a statistical test in their τ-p inversion (Stark and Parker, 1987); the structure in the shallowest 200 km is fixed and the error bounds for the transition zone velocities are estimated from the scatter in their observations. Their results indicate that while velocity gradients above and in the transition zone are well constrained by the data, shifting the depths of the 410- and 660-km discontinuities by ± 20 and ± 35 km, respectively, can be easily afforded, and they concluded that the difference in depths of the 660-km discontinuity in Model 12 (at approximately at 700 km) and ARC-TR (between 630 and 670 km) might be insignificant. We tested our model with a qualitative approach; a number of models are generated by slightly perturbing Model sn12c, and the predicted travel times are compared with the observations and the sn12c prediction. Our conclusion is similar to that of Sugiyama and Nakanishi (1988, 1999), that is, our regional travel time data bound the discontinuity depths weakly ($\pm \sim 30$ km) due to scatter in data at critical epicentral distances. Similarly, disagreement among previous models in the Northwest Pacific regions might be fictitious and might not reflect the regional variations. In most studies, travel times are picked from the composite record sections, or waveforms of more than one event that are recorded at a number of stations. When the number of events becomes large, the error in source locations would bias the identification of later branches. Use of large aperture arrays appears to be promising in reducing such bias, but, especially when inter-station distances are large, weak correlation of waveforms between adjacent stations, owing to the site effects (Kato *et al.*, 2000), would be disadvantageous in identifying branches of later phases. Station corrections for the short-period waveforms (Yamazaki *et al.*, 1996) should be still a difficulty for waveforms recorded in regional distances owing to the nature of complex wave field.

Lateral heterogeneity in the transition zone velocities has been reported for the Philippine Sea region (Brudzinski *et al.*, 1997; Tajima and Grand, 1998; Nowack *et al.*, 1999) so that it might be crude to discuss the structure in detail with our results from a one-dimensional approach. Nevertheless, our findings of high velocity anomaly in two paths qualitatively agree with high velocity anomalies that are found in other regional studies (e.g., Suetsugu and Nakanishi, 1987; van der Hilst *et al.*, 1991; Fukao *et al.*, 1992). High velocity anomaly in the transition zone beneath the Philippine Sea has been linked to the accumulation of subducted slab which do not penetrate into the lower mantle effectively (Okino *et al.*, 1989; van der Hilst and Seno, 1993; Ohtaki and Kaneshima, 1994; Tajima and Grand, 1998). The origin of the high velocity anomalies could be both thermal and chemical considering the compositional difference between the oceanic plate and the ambient mantle (e.g., Gaherty *et al.*, 1999).

6. Concluding Remarks

We investigated the upper mantle velocity structure in the western Pacific region using array recordings at MSAS. Direct and triplicated *P* phases are picked up from waveform data, and upper mantle velocities in three regions are investigated from observed Δ-T curve. First arrivals for the Kurile-Alaska path are satisfactorily explained by a standard travel time model for this region. Our observations of direct and triplicated *P* branches are well explained by a velocity model sn12c, which is a slight modification of Model 12 (Sugiyama and Nakanishi, 1989 and 1999). No strong evidence for the low-velocity zone in the uppermost mantle is found for this region. For the Ryukyu-Taiwan-Philippine path and Izu-Bonin-Mariana path, construction of velocity models is impeded by the quality and quantity of data. Nevertheless, the observation implies a slow uppermost mantle and a fast transition zone and uppermost lower mantle in these regions. The differences in the velocity structure could be the outcome of the past activity of subduction in the Philippine Sea region.

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