Slant-stack Velocity Analysis for One-dimensional Upper Mantle Structure Using Short-period Data From MAJO

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Abstract—In this paper, regional *P*-wave upper mantle structure is investigated using slant-stack velocity analysis of short-period earthquake data recorded at station MAJO (Matsushiro, Japan). Shallow earthquakes from 1980–1986 within 35° of MAJO are used to construct a common receiver gather. Processing of the wavefield data includes focal depth and static time corrections, as well as deterministic deconvolution, in order to equalize pulse shapes and align wavelets on the first arrivals. The processed wavefield data are slant stacked and iteratively downward continued to obtain a regional upper mantle velocity model. The model includes a low velocity zone between 107 and 220 km. Beneath the LVZ, the velocity increases smoothly down to the discontinuity at 401 km. In the transition zone, the velocity model again increases linearly, although there is some suggestion of further complexity in the downward continued wavefield data. At the base of the transition zone, a second velocity discontinuity occurs at 660 km, with a linear velocity gradient below. In addition to slant-stack analysis, travel times and synthetic seismograms are computed and compared with the processed and unprocessed wavefield data.

Key words: Slant-stack velocity analysis, upper mantle structure.

Introduction

In this study, *P*-wave upper mantle structure beneath the active subduction zones of the northwest Pacific is investigated using slant-stack analysis of short-period earthquake data recorded at station MAJO. Shallow earthquakes from 1980–1986 within 35° of station MAJO are used to construct a common receiver gather. Slant-stack analysis and downward continuation are then used to image one-dimensional upper mantle structure. The determination of regional upper mantle structure is important when comparing with other regional and global models, as well as providing a baseline for more detailed tomographic analyses (e.g., FUKAO *et al.*, 1992; VAN DER HILST, 1991).

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An early study of the upper mantle P -velocity structure for the western Pacific was conducted by FUKAO (1977). Fukao obtained a velocity model for the upper 800 km of the mantle beneath the ocean side of the Japan-Kuril Arc by modeling travel time and amplitude data from Kuril-Kamchatka earthquakes recorded at the Wakayama Micro-Earthquake Observatory, Japan. His model ARC-TR shows two zones of high velocity gradient between 390–400 km and 630–670 km. The Fukao model also includes a thick lithospheric lid down to 85 km and a low velocity zone between 85–205 km.

Recently, SHEARER (1991) stacked long-period Global Digital Seismograph Network (GDSN) data in order to construct global record sections. The phases from velocity discontinuities near 410 and 660 km are clear in the stacked waveform images. Possible evidence for a weaker discontinuity at 520 km was also inferred from the waveform images, and from cross correlation analysis of individual seismograms. However, no evidence for a 220 km discontinuity was found. SHEARER (1991) converted travel times to apparent discontinuity depths relative to the earth model PREM. The discontinuity depths obtained from P to SV converted phases at 35 individual stations, including station MAJO, show variations of less than 20 km. The best global averages found from the study by Shearer have discontinuity depths at 415, 519 and 659 km. The apparent discontinuity depths from P/SV conversions at station MAJO were found to be 403 and 658 km.

REVENAUGH and JORDAN (1991a,b,c) analyzed upper mantle layering by using multiple ScS phases and reverberations in the long-period SH polarized seismograms. They found average upper mantle discontinuity depths at 414 and 660 km using PREM as a reference velocity model. For station MAJO, the apparent depths of upper mantle discontinuities are given by Revenaugh and Jordan to be at 402 km and 656 km. In addition, a '520' km discontinuity was found at 513 km for station MAJO.

In this study, a common receiver gather is constructed using short-period earthquake data within 35° of station MAJO. The data analysis includes the data selection process, the construction of a common receiver gather, bandpass filtering, focal depth and static corrections, and deconvolution. Slant-stack velocity analysis and downward continuation are then performed similar to the analysis of WALCK and CLAYTON (1984) and WALCK (1985), except that for this study, only a single station is used for the receiver gather. In addition to slant-stack analysis, travel times and synthetic seismograms are computed and compared to the wavefield data.

Data Analysis

Earthquakes from January 1980 through December 1986 with m_b or M_s greater than 5.5 and focal depth less than 55 km were selected from the NEIC Earthquake

Data Base on CD-ROM. Short-period data within 35° of station MAJO (Matsushiro, Japan) were used to construct a common receiver gather. The selected earthquake epicenters centered on station MAJO are shown on an azimuthal equidistant projection map in Figure 1. An initial set of 166 shallow events with epicentral distances less than 35° from station MAJO were identified for construction of the receiver gather. The seismograms were bandpassed from 0.1 to 1.0 Hz using a Butterworth bandpass filter, and the amplitudes were unit normalized for each filtered record.

AZIMUTHAL DISTRIBUTION OF EARTHQUAKES
BETWEEN 1980-1986 FOR STATION MAJO
($M \geq 5.5$ AND DEPTH < 55 KM)

Figure 1

Distribution of earthquake epicenters centered on MAJO (open triangle) (m_b or $M_s \geq 5.5$, focal depth ≤ 55 km, 1980-1986). The solid circle includes the seismicity within 35 degrees of station MAJO. The map projection used is an azimuthal equidistant projection.

First Arrival Travel Times and Static Corrections

Figure 2A shows the first arrival travel times which were picked interactively for the MAJO record section for a subset of 109 events. The scatter in the travel time data results from picking errors, errors in the origin times and hypocentral depths, as well as from lateral velocity variations. Approximately 40 percent of the shallow events selected had focal depths constrained to 33 km. Nonetheless, time corrections for focal depth are still required because of the trade-off between focal depth and origin time.

In order to correct the travel times for variable focal depth, the Herrin tables (HERRIN, 1968) were used to equalize focal depths. This is consistent with the original location procedures. Also, the shallow source structures were not known well enough for more detailed corrections. An effective zero focal depth equalization was chosen, since this is required for the downward continuation analysis. Figure 2B shows the resulting depth corrected travel times. The depth correction has the effect of shifting the times later, as well as reducing some of the scatter in the picks, particularly in the critical distance range between 2000 and 2700 km. However, additional scatter in the travel time remains.

In Figures 2A and 2B, the open squares are for events along the Japan and Kuril-Kamchatka subduction zones north of MAJO, and the open circles are for events to the south of MAJO. Since the time differences between events to the north

Figure 2(a)

Figure 2(b)

Figure 2

A) First arrival travel-time picks for the MAJO common receiver gather. B) Travel time corrected for focal depths. The dotted line is a smooth static correction line. Squares denote earthquakes north of MAJO and circles denote earthquakes to the south of MAJO.

and to the south of MAJO are on the order of the scatter in the depth corrected travel times, the events have been combined in order to estimate an average one-dimensional velocity structure. For the estimation of laterally varying structure, multiple stations as well as events are required.

Several travel times with large delays are shown in Figures 2A and 2B. For example, the open circle near a range of 1400 km and a reduced time of 54 seconds has been identified to be from a reverse branch of the 400 km discontinuity. However, a clear first arrival for this seismogram was difficult to identify. As a result, this event, as well as several others near a range of 1200 km, were removed prior to the stacking procedure. However, the points on the travel-time plots have been retained to help identify triplication branches.

In order to construct a one-dimensional velocity model, additional static corrections were applied to the travel times. A static correction line was manually fit through the depth corrected travel-time data and is shown by the dotted line in Figure 2B. In order to test whether the static correction line is reasonable, the travel times associated with this line were inverted for a one-dimensional model using the

tau-sum method of DIEBOLD and STOFFA (1981). The resulting velocity model is compared in Figure 3 with the model PREM of DZIEWONSKI and ANDERSON (1981), the model 1066b of GILBERT and DZIEWONSKI (1975) and the model ARC-TR of FUKAO (1977). Since the static correction line is a smoothly varying curve fit to the first arrivals, the resulting inverted model has only limited resolution of upper mantle velocity increases. Still, the trend of the velocity model derived from the static correction line is comparable with the other reference models. Therefore, the static correction line is considered to be reasonable for providing a smoothed time correction for the first arrival travel times.

Since the smallest offset from MAJO is 294 km, there is limited information on crustal structure. From the slope of the static line shown in Figure 2B, a shallow upper mantle velocity of approximately 8.2 km/sec is found. This is similar to

Figure 3

Velocity inversion of the smooth static correction line (solid line) given in Figure 2B compared to PREM (dashed line), 1066b (dotted line) and ARC-TR (dot-dashed line) of FUKAO (1977).

8.23 km/sec found by FUKAO (1977) using dT/dX data for Kuril-Kamchatka events recorded at the Wakayama array in Japan. Local seismic estimates of shallow upper mantle velocity beneath Japan are between 7.5 to 7.7 km/sec (YOSHII and ASANO, 1972; ZHAO *et al.*, 1992). Therefore, the travel time slope found here, as well as by FUKAO (1977), must be an average of the ocean side structure near the event locations and the structure beneath the station MAJO in Japan.

The velocity model of FUKAO (1977) had a 33 km crust with an upper 15 km at 5.57 km/sec above a 6.5 km/sec lower crust. This was constrained by a single event

Figure 4

Deterministic deconvolution of a seismic trace at an epicentral distance of 3241 km. A) A 5 sec extracted wavelet is used to deconvolve the original trace. B) A 7.5 sec extracted wavelet is used to deconvolve the original trace. For each case, 1 is the original trace, 2 is the extracted wavelet and 3 is the deconvolved result.

with an ISC depth of 38 km recorded at several regional stations. However, the resulting travel times for a surface focus event were found to be too fast by about 3.5 s, with respect to the JB tables (FUKAO, 1977; Figs. 14, 15). This discrepancy was ascribed to uncertainties in the event origin time.

YOSHII and ASANO (1972) determined a structural model beneath Honshu, Japan with a 2.5 km/sec near surface layer, 5.9 km/sec down to 15 km, 6.6 km/sec down to the Moho at 33 km, and 7.5 km/sec beneath the Moho. The recent study by ZHAO *et al.* (1992) used simultaneous inversion of local earthquake data to determine a variable depth to the Conrad and Moho beneath Japan. In the vicinity of station MAJO, Conrad and Moho depths of approximately 19 km and 38 km were found using crustal velocities of 5.9 km/sec and 6.6 km/sec, and a sub-Moho velocity of 7.7 km/sec (ZHAO *et al.*, 1992; Figs. 6, 9).

For our study, an average crustal correction for both the source and receiver is needed which matches the near offset, depth corrected travel times. A layered crust with a shallow layer of 2.5 km/sec in the upper 1.4 km, 5.5 km/sec down to 19 km and 6.5 km/sec down to 38 km provides the necessary time correction needed to

BP FILTERED SEISMOGRAMS FROM 0.1 TO 1.0 HZ
IN THE DISTANCE RANGE 0 TO 35 DEGREES FROM MAJO

Figure 5(a)

BP FILTERED SEISMOGRAMS FROM 0.1 TO 1.0 HZ
IN THE DISTANCE RANGE 0 TO 35 DEGREES FROM MAJO

Figure 5(b)

Figure 5

A) Unprocessed MAJO wavefield gather corrected for focal depth. The X's are the depth corrected arrival times and the dashed line is the static correction line of Figure 2B. B) Processed wavefield gather with static corrections and deterministic deconvolution applied.

match the near offset travel times. Nonetheless, this model is only an average of the source and receiver crusts. If the crust and upper mantle velocities beneath MAJO from ZHAO *et al.* (1992) down to 50 km are combined with an oceanic crustal model with an 8.2 km/sec upper mantle, this gives an average vertical time very similar to that given by the above crustal model used to match the depth corrected travel times at near offset. Although the layered crustal model given above is slower than that given by either that of YOSHII and ASANO (1972) or ZHAO *et al.* (1992) for Japan, it gives the appropriate time correction which averages between the source and receiver near surface structure. For a more detailed estimate of laterally varying crustal and upper mantle structure, the use of multiple stations and sources is required.

Deconvolution

The seismograms from the MAJO common receiver gather were further processed using deterministic deconvolution in order to improve the temporal resolution of the different seismic phases in the data. Since each seismogram of the receiver gather is from a different earthquake, deconvolution is necessary to equalize source pulses for better coherence in the slant-stacked image. In addition, deconvolution centers the wavelets on the arrival times.

For deterministic deconvolution, the source wavelet is assumed known and the inverse filter is designed to contract the source wavelet. The deconvolved output

Figure 6

Slant-stacked τ - V wavefield of the processed T - X wavefield data in Figure 5B.

$x(t)$ can be written as,

$$x(t) = IFFT \left\{ \frac{FFT(h(t)) * FFT(y(t))}{FFT(h(t)) * FFT(h(t)) + \theta^2} \right\},$$

where $y(t)$ is the original trace, $h(t)$ is extracted wavelet for deconvolution, FFT denotes the fast Fourier transform, $IFFT$ denotes the inverse fast Fourier transform and $*$ denotes complex conjugate. θ^2 is a damping factor, typically chosen as some factor of the maximum squared spectral amplitude of $FFT(h(t))$. For our study,

$$\theta^2 = 0.05 \times \max\{FFT(h(t)) * FFT(h(t))\}.$$

Figure 7

Downward continued $Z-V$ wavefield using the velocity model obtained from the inversion of the static correction line given in Figure 3.

Finally, the deconvolved results are bandpass filtered back to the original bandwidth of the data.

As an example of deterministic deconvolution, Figure 4 shows a seismogram from an August 26, 1986 earthquake ($M_s = 5.9$) with a shallow focal depth of 8 km and a range of 3241 km. In Figure 4A, the original seismogram, noted by 1, is deconvolved by a 5 sec extracted wavelet, noted by 2. The deconvolved trace, noted by 3, is then filtered back to the bandwidth of the original data. In Figure 4B, a 7.5 sec extracted wavelet is used for the deconvolution. Although both extracted wavelets result in a symmetric first arrival, the shorter wavelet preserves more detail in the deconvolved seismogram.

Figure 8

Downward continued Z-V wavefield using the velocity model 1066b.

Since the earthquake source time functions were generally unknown, the shortest extracted wavelets which provided the simplest, most symmetric deconvolved results were used. The lengths of the extracted wavelets in the deconvolution were obtained interactively for each trace. For the seismograms in the receiver gather, an average extracted wavelet length of 4.5 sec was used, with a minimum of 2.5 sec and a maximum of 7 seconds. In the distance range between 1000 and 3000 km, where the upper mantle triplications occur, shorter extracted wavelets were used. Although some detail is lost using deconvolution, it was found to be a necessary step for the slant-stack analysis when using seismograms from many different events.

The unprocessed wavefield gather for MAJO is shown in Figure 5A. The X's are the focal depth corrected travel times and the dashed line is the static correction line. The processed wavefield gather is shown in Figure 5B, where the processing included static time corrections and deterministic deconvolution in order to improve coherence and align the wavelets on the first arrivals. The processed wavefield data in Figure 5B were muted 5 sec before and 24 sec after the centered first pulse in order to window out large amplitude later arrivals. Also, the first and last traces with offset were cosine tapered prior to slant stacking. Good coherence of the processed wavefield can be seen in Figure 5B, particularly for ranges less than 1400 km and greater than 3000 km. A decrease in relative first arrival amplitudes occurs in the range between 1600 and 2250 km, which is interpreted to result from the upper mantle low velocity zone for this region.

Slant-stack Velocity Analysis

Review of the Method

Slant-stack velocity analysis of waveform data consists of two transformations of seismic T - X data, slant stacking and downward continuation (CLAYTON and McMECHAN, 1981). Slant stacking transforms the wavefield data in the T - X domain into the τ - p domain through the relation,

$$S(\tau, p) = \int_{-\infty}^{\infty} y(t = \tau + px, x) dx,$$

where $y(t, x)$ is the original waveform data, τ is the intercept time, t is the travel time, x is the distance and p is the ray parameter.

The second transformation is the downward continuation from the $(\tau$ - p) domain to the $(p$ - Z) domain with a specified velocity model $v(z)$. Lateral homogeneity along the observational profile is assumed for this step. For a flat earth

model, $S(z, p)$ is given by

$$S(z, p) = \int S(\omega, p) e^{-i\omega\psi(p, z)} d\omega$$

where

$$\psi(p, z) = 2 \int_0^{z(p)} \left[\frac{1}{v(z)^2} - p^2 \right]^{1/2} dz,$$

ω is the frequency and $z(p)$ is the bottoming depth of the ray given by ray parameter p .

Figure 9

Downward continued Z-V wavefield using the velocity model ARC-TR of FUKAO (1977).

Finally, an average phase shift of $5\pi/4$ is applied to the wavefield data as described by CLAYTON and McMECHAN (1981). Since the downward continuation requires an initial velocity model, inversion by downward continuation is inherently an iterative procedure. Convergence is achieved when the maxima in the downward continued wavefield coincide with the specified velocity model. Also, since the transformations are performed assuming a flat-layered earth model, conversion from a flat to a spherical earth is necessary (AKI and RICHARDS, 1980). The resulting procedure used is similar to that of WALCK and CLAYTON (1984), except that for this study just a single station is used for the slant-stack analysis.

Figure 10

Downward continued Z - V wavefield using the earth flattened final velocity model in Table 1.

Slant-stack Analysis of the Waveform Data from the Common Receiver Gather at Station MAJO

The seismic wavefield in Figure 5B was transformed to the τ - p domain by a slant-stack transformation. The resulting slant-stacked wavefield is shown in Figure 6 using equal increments of velocity instead of ray parameter, p , where $v = 1/p$. The velocity axis ranges from 7 to 14 km/sec with a spacing of 0.1 km/sec. The corresponding τ values are from 0 to 140 sec in 0.05 sec increments. The seismic gather used for the slant stack had 104 traces which were unit normalized prior to stacking. Although the slant-stacked wavefield has a somewhat noisy character, loci of energy can be identified. Coherence of the slant-stacked wavefield is affected both by noise, as well as gaps in the T - X data. However, tests using synthetic data with as few as 50 traces still provided useful slant-stacked wavefields for the downward continuation.

Several velocity models were used to initially downward continue the slant-stack data to the Z - V domain. Figure 7 shows the downward continued Z - V wavefield using the velocity model obtained from the static correction line shown in Figure 3. This is a smooth model with no upper mantle discontinuities and results in a broad zone of large amplitudes in the Z - V wavefield. The next velocity model used was the model 1066b with upper mantle discontinuities but no P -wave low velocity zone. The downward continued image is shown in Figure 8. For this model, there

Table 1
P-wave velocities for MAJO

Depth (km)	Velocity (km/sec)	Depth (km/sec)	Velocity (km/sec)
0.0	2.5	577.	9.95
1.4	5.5	608.	10.06
19.	6.5	634.	10.17
38.	8.19	659.	10.23
73.	8.20	661.	10.99
107.	8.21	706.	11.02
117.	8.07	740.	11.06
190.	8.07	775.	11.10
220.	8.45	809.	11.15
260.	8.53	844.	11.19
280.	8.58	879.	11.23
320.	8.66	913.	11.27
360.	8.75	948.	11.32
385.	8.80	982.	11.37
400.	8.83	1017.	11.42
402.	9.37	1052.	11.48
451.	9.53	1086.	11.53
483.	9.64	1121.	11.59
514.	9.75	1156.	11.64
546.	9.85	1190.	11.70

is some convergence of the Z - V wavefield for depths greater than 670 km but refinements are needed for shallower depths. For example, the large wavefield amplitudes near a depth of 100 km are not fit by this model. Figure 9 displays the downward continued Z - V wavefield using the velocity model of FUKAO (1977). This model includes an upper mantle low velocity zone, as well as broad velocity increases near 400 km and 660 km. The use of this model for the downward continuation has the effect of diffusing the wavefield energy in the Z - V domain.

A number of additional iterations of the downward continuation step were performed. The fitting was done interactively in order to compress and align the energy of the downward continued wavefield to coincide with the iteratively updated velocity model. This is similar to the criterion used by CLAYTON and MCMECHAN (1981) for convergence of the iterative process. The positions and sizes of the discontinuities were also updated in the iterative procedure. As a constraint, each intermediate velocity model was ray traced and compared with the observed travel times. The final downward continued wavefield and earth flattened velocity model are shown in Figure 10. The converted radial earth model is tabulated in Table 1

**BP FILTERED SEISMOGRAMS FROM 0.1 TO 1.0 HZ
IN THE DISTANCE RANGE 0 TO 35 DEGREES FROM MAJO**

Figure 11(a)

and shown by the solid line in Figure 14. The final model has velocities similar to the model of FUKAO (1977) in the upper 220 km but with the low velocity zone at a somewhat greater depth. The LVZ is characterized on the $Z-V$ wavefield by a zone of lower amplitudes. Below the LVZ, the velocity model increases linearly down to the velocity discontinuity at 401 km. In the transition zone, the velocity model again increases linearly, although there is some suggestion of additional small-scale structure in the transition zone on the $Z-V$ wavefield. At the base of the transition zone, a velocity discontinuity occurs at 660 km with a linear increase in velocity below.

The computed travel times for the resulting model are shown by the solid line in Figure 11A, displayed on the unprocessed $T-X$ wavefield data. The X's are the depth corrected travel times and the dashed line is the static correction line. Figure

Figure 11(b)

Figure 11

A) Calculated travel times (solid line) for the final model displayed on the unprocessed MAJO seismic wavefield gather corrected for focal depth. X's denote the arrival times corrected for focal depth and the dashed line is the static correction line. B) Calculated travel times (solid line) for the final model displayed on the processed wavefield gather. The dashed line is the static correction line.

Figure 12

Ray theoretical synthetic seismograms (unit normalized) for the velocity model in Table 1.

11B shows the computed travel times displayed on the processed $T-X$ wavefield data. For offsets greater than 1600 km, a decrease in relative first arrival amplitudes is associated with the upper mantle low velocity zone. The geometric arrivals are delayed by the LVZ and can be seen as the larger amplitudes near 1600 km in the processed wavefield data. Wavefield phases associated with both 400 and 660 km discontinuities can also be seen in Figure 11B. In particular, on seismograms near 2200 km, several of the triplication phases can be identified.

Figure 12 shows ray theoretical synthetic seismograms (ČERVENÝ and PŠENČÍK, 1984) computed for the final velocity model. The traces are unit normalized similar to the observed wavefield data. One of the consequences of using ray theory is that diffractions from caustics are not included. Nonetheless, the major geometric phases are obtained. The synthetic wavefield data is slant stacked and downward continued to obtain the $Z-V$ wavefield shown in Figure 13. Although the downward continuation process does not image the lower amplitudes within the low velocity zone, the $Z-V$ wavefield above and below the LVZ correctly images the velocity structure. The downward continued synthetic wavefield in Figure 13 can be compared with the observed $Z-V$ wavefield data in Figure 10.

Figure 13

Downward continuation of the slant stack of the synthetic seismograms in Figure 12 using the velocity model in Table 1.

Discussion

The average radial velocity model obtained from the slant stack and downward continuation of the MAJO common receiver gather of short-period data is shown by the solid line in Figure 14. The velocity models PREM of DZIEWONSKI and ANDERSON (1981) (dashed line), 1066b of GILBERT and DZIEWONSKI (1975) (dotted line) and ARC-TR of FUKAO (1977) (dot-dashed line) are also shown for comparison. The final velocity model for MAJO has a low velocity zone at depths between 107–220 km with the lowest velocity within the LVZ of 8.07 km/sec. On

Figure 14

Comparison of the radial *P*-wave velocity model obtained from the slant stack and downward continuation of the waveform data from the MAJO common receiver gather (solid line) to the velocity model ARC-TR of FUKAO (1977) (dot-dashed line), the PREM model of DZIEWONSKI and ANDERSON (1981) (dashed line), and 1066b of GILBERT and DZIEWONSKI (1975) (dotted line).

the processed seismic gather in Figure 11B, the low velocity zone results in reduced amplitudes of the first arrivals at offsets between 1600 and 2250 km. This was also identified by FUKAO (1977). The lower amplitude first arrivals in this range result from the delay of the geometric arrivals due to the LVZ. For travel-time inversion studies the delayed geometric arrivals are the important phases for identification (SLICHTER, 1932; GERVER and MARKUSHEVITCH, 1966; AKI and RICHARDS, 1980). The incorrect identification of the geometric arrivals would pose interpretation problems for both 1-D inversions and tomographic studies using travel times from this distance range.

Beneath the low velocity zone, the velocity increases smoothly down to the discontinuity at 401 km, where the velocity increases by 5.9%. Since our data has limited resolution in the transition zone, the final transition zone velocities increase smoothly with a gradient of .0034/sec. SHEARER (1990) inferred a global velocity discontinuity at 519 km, based on evidence from the stacked images of long-period seismograms. REVENAUGH and JORDAN (1991a,b) also reported a 520 km discontinuity with a reflection coefficient of 0.014. However, recent short-period studies by CUMMINS *et al.* (1992) and JONES *et al.* (1992) conclude that a sharp 520 km velocity discontinuity is not required, although a density change is not ruled out. At the base of the transition zone, a velocity increase of 7.3% occurs at 660 km. Although there may be further complications below 660 km, the final model from this analysis has the velocity increase smoothly with a gradient of .0013/sec.

In the wavefield downward continuation analysis, wavelet resolution can be approximately used as a measure of model uncertainty. For the downward continuation using the velocity model in Figure 10, the errors are approximately ± 0.2 km/sec in the upper 400 km and ± 0.25 km/sec below. A comparison with other mantle models in Figure 14 reveals that these uncertainties are also the approximate ranges between different models. Nonetheless, the depth-averaged velocities are constrained to fit the absolute travel times. Therefore, systematic deviations over the entire depth section are not allowed. In terms of depths to upper mantle discontinuities, the error estimates are on the order of a fraction of a $Z-V$ wavefield pulse width of approximately ± 12 km, trading off with the velocity estimates.

The upper mantle velocity discontinuities at 401 and 660 km depths in the final velocity model are similar to those obtained for MAJO by SHEARER (1991) and REVENAUGH and JORDAN (1991a,b). SHEARER (1991) found velocity increases for MAJO at 403 ± 18 and 658 ± 9 km depths, using P/SV conversions. REVENAUGH and JORDAN (1991a,b) obtained velocity discontinuities for MAJO at depths of 402 and 656 km, using long-period ScS reflections. The final model discontinuity depth of 401 km, as well as depths estimated for MAJO by SHEARER (1991) and REVENAUGH and JORDAN (1991a,b), are shallower than the global average of 415 km found by SHEARER (1991) and 414 km found by REVENAUGH and JORDAN (1991a,b). However, the depth estimate near 660 km in our final model for MAJO is comparable to the global depth averages of 659 km by SHEARER (1991) and 660 km found by REVENAUGH and JORDAN (1991a,b).

Using long-period SS precursor data, SHEARER (1991) and SHEARER and MASTERS (1992) found a correlation between subduction zones and depth fluctuations of the 410 km and 660 km discontinuities. Along the Kuril-Kamchatka trench, SHEARER and MASTERS (1992) found a broad depression in the 660 km discontinuity where the subducting slab intersects the discontinuity. In our study, the ray

bottoming points are predominantly on the eastern side of the subduction zones of the northwest Pacific. Our inferred discontinuity depth of 660 km is consistent with the depths determined from SS precursors for a 10 degree zone on the eastern side of the Kuril-Kamchatka subduction zone (see Figure 6 from SHEARER and MASTERS, 1992).

The long-period SS precursor data of SHEARER (1991) for the 410 km discontinuity show less depth variability than for the 660 km discontinuity along the Kuril-Kamchatka trench. Our results, based on rays bottoming mostly on the eastern side of the subduction zones of the northwest Pacific, suggest that the 410 km discontinuity is relatively elevated, compared with the global averages.

An advantage of the present work is that it provides an estimate of both velocity structure and discontinuity depths. This is in contrast to techniques which infer only relative discontinuity depths and must use a reference velocity model for interpretation. Also, the results of this study are based on regionally recorded, short-period seismic data which have the potential of providing better resolution than estimates based on long-period data.

Conclusion

The average one-dimensional structure of the upper mantle beneath the active subduction zones of the northwest Pacific has been investigated using short-period earthquake data recorded at station MAJO (Matsushiro, Japan). Shallow earthquakes from 1980–1986 at distances less than 35° were used to construct a seismic receiver gather. The processing of the wavefield data included focal depth and static corrections and deterministic deconvolution to increase coherence and align the wavelets on the first arrivals. The processed data were slant stacked to the τ - p domain and then iteratively downward continued to estimate a regional upper mantle velocity structure. The final model has an upper mantle low velocity zone between 107–220 km. Below the LVZ, the velocity increases smoothly down to the velocity discontinuity at 401 km. In the transition zone, the velocity again increases linearly, although there is some suggestion of further complexity in the downward continued wavefield data. A second velocity discontinuity occurs at 660 km with smoothly varying velocities below. Travel times and synthetic seismograms have been computed and compared with the wavefield data.

Acknowledgments

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