Tomographic imaging of the $P$-wave velocity structure beneath the Kamchatka peninsula

A. Gorbatov,1,* J. Domínguez,1 G. Suárez,1 V. Kostoglodov,1 D. Zhao2 and E. Gordeev3

1 Instituto de Geofísica, Universidad Nacional Autónoma de México, Del. Coyoacan, 04510, México D.F., México
2 Department of Earth Sciences, School of Science, Ehime University, Bunkyo-cho 2–5, Matsuyama 790–8577, Japan
3 Geophysical Service, Academy of Sciences of Russia, Piip Avenue 9, Petropavlovsk-Kamchatsky, 683006, Russia

SUMMARY
A total of 5270 shallow and intermediate-depth earthquakes recorded by the 32 stations of the regional seismic network of the Geophysical Service of Russia are used to assess the $P$-wave velocity structure beneath the Kamchatka peninsula in the Western Pacific. The tomographic inversion is carried out in three steps. First, a 1-D tomographic problem is solved in order to obtain an initial velocity model. Based on the 1-D velocity model, 3-D tomographic inversions with homogeneous and heterogeneous starting models are obtained. The Conrad (15 km depth) and Moho (35 km depth) discontinuities determined from the 1-D tomographic inversion, and the upper boundary of the subducting slab are taken into account in the heterogeneous starting model for the traveltimes and ray-path determinations. Both velocity structure and hypocentral locations are determined simultaneously in the inversion. The spacing of the grid nodes is a half-degree in the horizontal direction and 20–50 km in the vertical direction. A detailed $P$-wave tomographic image is determined down to a depth of 200 km. The resulting tomographic image has a prominent low-velocity anomaly that shows a maximum decrease in $P$-wave velocity of approximately 6 per cent at 30 km depth beneath a chain of active volcanoes. At depth, low-velocity anomalies are also observed in the mantle wedge extending down to a depth of approximately 150 km. These anomalies are apparently associated with the volcanic activity. The sedimentary basin of the Central Kamchatsky graben, to the west of the volcanic front, and the accretionary prism at the trench correlate with shallow low-velocity anomalies. High-velocity anomalies observed at a depth of 10 km may be associated with the location of metamorphic basements in the Ganalsky–Valaginskoe uplift and upper crust of Shipunsky cape. The results also suggest that the subducted Pacific plate has $P$-wave velocities approximately 2–7 per cent higher than those of the surrounding mantle and a thickness of approximately 70 km.

Key words: Kamchatka subduction zone, tomography, volcanic activity.

1 INTRODUCTION
The Pacific plate subducts westwards beneath the Kamchatka peninsula at a rate of $\pm 8$ cm yr$^{-1}$, with a dip angle of about 55° from $\approx 50°$N to $\approx 54°$N. A Wadati–Benioff seismic zone defines the downgoing slab to a maximum depth of $\approx 500$ km in the southern part of the Kamchatka subduction zone (KSZ); this depth shallows gradually to $\approx 300$ km towards the north. North of $\approx 54°$N, the subducted lithosphere is sharply deformed and the subducted slab becomes shallower (Gorbatov et al. 1997). These drastic changes in dip angle are reminiscent of those occurring in younger subduction environments such as Peru (Cahill & Isacks 1992) and Mexico (Pardo & Suárez 1995). In Kamchatka, however, they take place in one of the oldest ($\approx 80$ Ma) (Rea et al. 1993) subducted slabs in the world. This abrupt shallowing of the subducted slab produces a sudden offset of the volcanic front (Kliuchevskoi and Sheveluch volcanoes) to the northwest. This complexity of the subducted...
slab and the sharp offset of the volcanic front has prompted several authors (e.g. Tatsumi et al. 1994) to suggest that Kluachevskoi and Sheveluch are not typical subduction-related volcanoes, but that instead they originated from back-arc volcanic activity and are part of a second volcanic arc.

The seismic velocity structure of the Kamchatka subduction zone is very complex due to the presence of an active volcanic belt, a deep trench and thick sedimentary basins located in the Central Kamchatky graben (Fig. 1). Also, the subducted lithosphere and the abundant volcanic activity could cause large seismic velocity heterogeneities within the crust and upper mantle. Many of these tectonic features of the Kamchatka subduction zone should be visible as seismic velocity anomalies. However, detailed regional studies of the seismic velocity structure such as a 3-D tomographic inversion for the whole Kamchatka subduction zone, including inland and offshore regions, have not been carried out to date.

Several studies have proposed 2-D crustal velocity models beneath Kamchatka using arrival times observed from seismic explosions on land and in the ocean (e.g. Anosov et al. 1978; Selivestrov 1983; Balesta & Gontovaya 1985). These studies suggest that the oceanic lithosphere has higher P-wave velocities (≈ 7 per cent) than the continental crust located near the trench. At these shallow depths, the estimated value of P-wave velocity decreases down to ~ −4 per cent beneath the volcanic front in the central part of the Kamchatka peninsula.

Arrival times from regional earthquakes were also used to study the crustal structure of Kamchatka by Gorshkov (1958) and Slavina & Fedotov (1974). Although studies using refraction methods exist for limited zones such as several individual volcanoes (Gorshkov 1958) or the crust covered by the regional seismic network (Slavina & Fedotov 1974), their results consistently show relatively low seismic velocities beneath the volcanic front at the apparent crust–mantle boundary (≈ 7.5 km s\(^{-1}\)), and relatively high velocities (≈ 7.9 km s\(^{-1}\)) related to the subducted slab.

A 3-D tomographic inversion for the area between the coast and trench was performed by Slavina & Pivovarova (1992). Although the results are restricted by the limited area of their study, the data suggest that the P-wave velocity values are ≈ 8.2 km s\(^{-1}\) in the oceanic plate and ≈ 7.6 km s\(^{-1}\) in the lower continental crust.

The extensive Kamchatka Regional Seismic Network (KRSN) has operated since 1962, reporting about 600 earthquakes each year. The abundant seismic activity and the relatively dense seismic station coverage (Fig. 2) provide the opportunity to determine in detail the tomographic images of the crust and upper mantle beneath the Kamchatka peninsula. The purpose of the present study is to infer the 3-D velocity structure of the crust and upper mantle in the KSZ, applying a 3-D tomographic inversion to the regional seismic data. This information may help to improve our understanding of the
dynamic processes that control the subduction regime and the complex volcanic activity in the region.

2 DATA AND METHODS OF ANALYSIS

2.1 Regional earthquake data

The arrival times recorded by the KRSN from January 1985 to December 1992 were used in this study (Fig. 2). The network consists of 32 three-component seismic stations covering half of the Kamchatka peninsula. The majority of the stations are located in the central part of Kamchatka and along the coast (Fig. 2). A total of 5270 events occurring in the area 50–57° N, 150–165° E and registered by more than four seismic stations with a formal error of hypocentral determination of less than 10 km (Gorbatov et al. 1997) were selected for the analysis.

The distribution of the number of earthquakes versus focal depth shows that most earthquakes occurred at estimated depths shallower than ≈200 km; a relatively small number of events are deeper than ≈200 km (Fig. 3).

To obtain a tomographic image of the Kamchatka subduction zone, three inversions were carried out sequentially. First, a 1-D tomographic inversion (Roecker 1981, 1982) was applied in order to obtain a starting 1-D velocity model for the subsequent 3-D inversions. Then, a 3-D tomographic problem with a laterally homogeneous starting model was resolved using the technique developed by Zhao (1991) and Zhao et al. (1992). Finally, using the same method, a 3-D inversion with a laterally heterogeneous starting model was conducted.

© 1999 RAS, GJI 137, 269–279
2.2 1-D inversion

Although several studies have suggested velocity models for the KSZ (e.g. Gorskikh 1958; Slavina & Fedotov 1974; Anosov et al. 1978; Selivestrov 1983; Balesta et al. 1985; Slavina & Pivovarova 1992), there are differences amongst them. To obtain the initial 1-D velocity model, the method proposed by Roecker (1981, 1982) was used. The initial homogeneous medium was divided into 19 flat layers (each 5 km thick) down to a depth of 95 km. Layers with a thickness of 25 km were then used for the depth range 95–300 km. The solution for hypocentres and velocities was obtained by a two-part iterative procedure. The first part was the separated solution of \( V_p \) and \( V_s \) velocities and station corrections. The singular value decomposition (SVD) method was applied to resolve the damped least-squares problem. The second part was the relocation of the hypocentres using the 1-D velocity model obtained and station corrections.

For the inversions, only 1657 events with estimated errors in hypocentral determination of less than 5 km were used. The thickness of the layers down to a depth of 40 km was changed slightly (in 1 km steps) in order to refine the resulting model. In order to regularize the inversion problem, neighbouring layers with differences between velocity values smaller than 0.15 km s\(^{-1}\) were merged into thicker layers, and the inversion was performed again using the modified model as the initial model. The value of the damping parameter was varied and accepted at its lowest level when the average root mean square (rms) traveltime residual started to decrease continuously during the inversions. After six inversions, the velocity variance in the inversion scheme reached 0.1 per cent and the inversion process was stopped.

Table 1 shows the final 1-D velocity model consisting of eight layers where the average Poisson ratio is 1.71. The standard errors of velocity estimates (Table 1) are determined from the covariance matrix. The resolution coefficient, estimated from the resolution matrix, is approximately 1 for the depth range 15–200 km; however, the resolution coefficient is not good for the first two shallow layers, for which it is 0.28 (0–5 km) and 0.56 (5–15 km). The layers deeper than 200 km do not have sufficient resolution (the resolution coefficient is \( \approx 0.1 \)) due to a limited number of earthquakes located beneath that depth. Three layers are distinguished in the crust (Table 1). The upper (5 km) low-velocity layer is apparently associated with large sedimentary basins and volcanic ash deposits. Although the resolution is relatively poor for that layer (Table 1), the presence of the low-velocity upper layer in the 1-D model gives a smaller rms than in the model lacking that layer. The average rms traveltime residual is 0.62 s for the final 1-D velocity model when the upper low-velocity layer is included. The rms traveltime residual increases to 0.67 s in the case when that layer is not included. Several other studies (e.g. Slavina & Fedotov 1974; Balesta et al. 1985) also suggest the presence of a low-velocity layer. The velocity discontinuities at depths of \( \approx 15 \) and \( \approx 35 \) km correspond apparently to the Conrad and Moho discontinuities, respectively.

2.3 3-D inversion with a laterally homogeneous starting model

The inversion scheme of Zhao (1991) and Zhao et al. (1992) is used to carry out the 3-D tomographic inversion. This method differs from other frequently used 3-D velocity inversion schemes. The main differences are as follows:

(a) a 3-D inhomogeneous velocity model with several velocity discontinuities of complex shape (such as the Conrad and Moho discontinuities, or the upper boundary of the subducted slab, for example) can be included in the inversion as initial a priori information;
(b) the velocities at any point in the model are calculated by the linear interpolation of the velocities at eight grid nodes surrounding that point;
(c) the ray tracing method of Zhao (1991) and Zhao et al. (1992) iteratively uses the pseudo-bending technique of Um & Thurber (1987) and Snell's law to determine the ray path;
(d) converted and refracted phases (such as \( Pn, P^*, PS, SP, Sn, Sp \)) can also be included in the inversion.

First, a 3-D tomographic inversion is carried out using a 1-D velocity starting model. Previously, a regional 1-D velocity model was determined in the depth range 0–200 km. Part of all the available seismic data to be used for the 3-D inversion extends down to a depth of 350 km, hence the earlier regional 1-D model cannot be applied down to this maximum depth. An appropriate starting 1-D model can be constructed as a combination of the regional 1-D model and one of the global mantle models such as the IASP91 earth model (Kennett & Engdahl 1991). Were we to merge formally these two models (regional 1-D model for 0–200 km depth and IASP91 for depths greater than 200 km), the resulting 1-D model would have an artificial sharp discontinuity at 200 km depth due to the inconsistency between the models. To avoid this feature, the starting 1-D model is comprised of only the crustal part (0–35 km) of the regional 1-D velocity model and the mantle part of the IASP91 earth model. This generic starting model better represent the details of the crustal structure in the KSZ and the unperturbed upper mantle.

In order to conduct the 3-D tomographic inversion with the initial laterally homogeneous starting model, the studied area was gridded in half-degree steps in the horizontal direction and at depth steps of 10, 30, 60, 100, 150, 200 and 350 km. The velocity structure, station corrections and hypocentral locations were determined simultaneously in each iteration. The conjugate gradient type solver, the LSQR algorithm of Paige & Saunders (1982), was used to solve the large and sparse system of observation equations. The rms traveltime residual, which was 0.625 s before the inversion, was reduced to 0.517 s after the inversion. Only grid nodes with more than 10 rays passing near each grid node (hit counts) were used in

<table>
<thead>
<tr>
<th>Depth (km)</th>
<th>( V_p ) (km s(^{-1}))</th>
<th>Error (km s(^{-1}))</th>
<th>Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>0–5</td>
<td>3.65</td>
<td>0.13</td>
<td>0.28</td>
</tr>
<tr>
<td>5–15</td>
<td>5.74</td>
<td>0.24</td>
<td>0.59</td>
</tr>
<tr>
<td>15–35</td>
<td>6.74</td>
<td>0.05</td>
<td>0.99</td>
</tr>
<tr>
<td>35–60</td>
<td>7.37</td>
<td>0.02</td>
<td>1.00</td>
</tr>
<tr>
<td>60–75</td>
<td>7.64</td>
<td>0.03</td>
<td>1.00</td>
</tr>
<tr>
<td>75–100</td>
<td>7.84</td>
<td>0.03</td>
<td>1.00</td>
</tr>
<tr>
<td>100–160</td>
<td>8.00</td>
<td>0.06</td>
<td>0.99</td>
</tr>
<tr>
<td>160–200</td>
<td>8.22</td>
<td>0.25</td>
<td>0.90</td>
</tr>
<tr>
<td>200–250</td>
<td>8.30</td>
<td>0.79</td>
<td>0.08</td>
</tr>
</tbody>
</table>
the inversion. A cross-section of the 3-D P-wave percentage perturbations estimated by the inversion is shown in Fig. 4. The perturbations vary between ~ -7 and 7 per cent. A low-velocity perturbation is observed mainly beneath the active volcanoes. An inclined high-velocity zone dipping to the west is clearly observed in the upper mantle. This high-velocity zone has a P-wave velocity 2–7 per cent higher than ‘normal’ mantle velocities. This high-velocity structure reflects the high-velocity subducted slab.

2.4 3-D inversion with a laterally inhomogeneous starting model

The Conrad, Moho and upper boundary of a subducted slab are sharp seismic velocity discontinuities (e.g. Fisher et al. 1983) which should be taken into account in the ray tracing scheme in order to improve the final tomographic image. When these discontinuities are included in the model, the hypocentres are better located and ray paths can be refined because ray trajectories are more accurately traced than under the assumption of a simple medium without the crustal reflectors and the subducted slab (Zhao 1991; Zhao et al. 1992, 1994; Zhao & Hasegawa 1993).

To refine the results obtained with the laterally homogeneous starting model, a tomographic inversion with a laterally inhomogeneous starting model was carried out. The study area was divided into three layers separated by the Conrad (15 km depth) and the Moho (35 km depth) discontinuities; these three layers correspond to the upper crust \( V_p = 5.74 \text{ km s}^{-1} \), lower crust \( V_p = 6.74 \text{ km s}^{-1} \), and upper mantle. The starting 1-D velocity model was the same as that used in the homogeneous tomographic inversion. The results of the homogeneous 3-D inversion suggest that the subducting Pacific plate has P-wave velocities 2–7 per cent higher than the normal IASP91 mantle. Therefore, for the following inversion a high-velocity zone with an average velocity 4 per cent higher than the surrounding mantle was introduced in the starting model as a priori information to represent the subducting Pacific plate. The configuration of the upper boundary of the subducted slab was taken from Gorbatskov et al. (1997) (Fig. 5).

The hypocentral location of the earthquakes, station corrections and the P-wave velocity structure were estimated simultaneously. Several inversions were carried out, changing the initial thickness and the P-wave velocity of the subducted slab in the starting model. Although some variations occurred in the inversion residuals (Figs 6a and b) and in the magnitude of the velocity anomalies, the resulting tomographic images did not change appreciably when the initial slab thickness was changed from 10 to 110 km, and the initial slab velocity perturbation was varied from 1 to 7 per cent. These tests indicate that the stability of the tomographic inversion does not depend strongly on the assumed initial model. However, the reduction in rms residuals shown in Fig. 6 indicates broad minima in both the definition of the slab thickness and the average fractional velocity perturbations. Also, these tests show that the inversion performed with an initial slab thickness of 70 km and slab velocity 4 per cent higher than that of the normal mantle gives the minimum residuals in a rms sense. In this particular case, for example, the average rms traveltime residual was 0.602 s before the inversion and was reduced to 0.419 after three iterations (Fig. 7).

Although the rms residual of the inversion with a priori assumed information was reduced by 19 per cent compared to that of the inversion with a laterally homogeneous starting model, an \( F \)-test (Draper & Smith 1966; Zhao et al. 1994) was performed in order to evaluate the statistical significance of the rms residual reduction. The application of that test is valid because the arrival time residual distribution was close to Gaussian (Fig. 7). The \( F \) ratio is defined as \( F = [(\text{SSR}_1 - \text{SSR}_2)/\text{DF}_2]/[(\text{DF}_1 - \text{DF}_2)/\text{SSR}_2] \), where \( \text{SSR} \) is the sum of the squared residuals and \( \text{DF} \) is the number of degrees of freedom for the inversion with homogeneous

Figure 4. Cross-section b (Fig. 1) of the fractional P-wave velocity perturbations resulting from the homogeneous 3-D velocity inversion. The scale relating the size of the symbols to the velocity perturbations is shown in the inset. Solid triangles are volcanoes.
Figure 5. Isodepth contours in km approximating the upper surface of the Pacific subducted slab (from Gorbatov et al. 1997). Other symbols are the same as in Fig. 1.

Figure 5. Isodepth contours in km approximating the upper surface of the Pacific subducted slab (from Gorbatov et al. 1997). Other symbols are the same as in Fig. 1.

(subscript 1) and inhomogeneous (subscript 2) starting models. Taking into account that the number of P-wave arrivals is 60,935 and that the number of nodes with hit counts greater than 10 is 697 for the inversion with a homogeneous starting model, \( DF_1 = 60,935 - (697 + 4 \times 5270) = 39,158 \). For the inversion with an inhomogeneous starting model, \( DF_2 = 60,952 - (985 + 4 \times 5270) = 38,887 \). The other values are \( SSR_1 = 12,911.14 \) and \( SSR_2 = 12,153.73 \). The resulting \( F \) ratio is 8.94. Considering that \( DF_2 \) is a very large number, a value of infinity is used to select the \( F \) ratio from the tables. The corresponding \( F \) ratio for the decrease of 271 degrees of freedom \( (DF_1 - DF_2) \) is \( F(271, \infty, 0.99) < 1.32 \) (Beyer 1991). The results of the \( f \)-test show that the \( F \) ratio obtained in the test is larger than the value given by the \( F \) probability distribution at the 1 per cent level. Therefore, the inversion with an inhomogeneous starting model significantly improves the final results at a 1 per cent significance level.

3 RESOLUTION TEST

It is commonly recognized that a solution of an inverse problem must be interpreted based on its resolution. The LSQR algorithm applied in this study does not estimate the resolution matrix during the inversion. Therefore, a checkerboard resolution test (CRT) (Humphreys & Clayton 1988; Zhao et al. 1992) was applied to evaluate the resolution of the tomographic results. A simple checkerboard synthetic structure with positive and negative perturbations of 3 and \(-3\) per cent was applied to the starting 1-D model. Two tests were carried out: first with a grid interval of the medium of about 50 km and second with a grid interval of about 100 km. Sets of traveltime delays resulting from ray tracing through the synthetic test structures were determined. Random errors corresponding to a normal distribution with a standard deviation of 0.1 s were added to the arrival times calculated for the synthetic model. Finally, the synthetic traveltime delays were used in the tomographic inversion in order to recover the initial synthetic structure. A comparison of the synthetic inversion image and the source synthetic structure allows us to recognize the areas where good resolution exists for the spatial distribution of available earthquakes and seismic stations in Kamchatka. The resolution is good where inversion results have the same velocity amplitude and checkerboard pattern as those of the original synthetic structure. Conversely, in areas where the initial pattern and amplitude are distorted the resolution is poor.

The checkerboard tests show that adequate resolution is obtained when the node separation in plan view is greater than 50 km. Fig. 8(a) shows the result of a checkerboard resolution test with a grid separation of \( \approx 50 \) km. The resolution is good for most of the study area down to a depth of 60 km (Fig. 8a). In the case of the 100 km deep slice, the resolution is acceptable inland and only near the area where the subducted Pacific plate is located, decreasing to the north and south of the seismic network. For depths beneath 150 km the resolution is poor and a correct test pattern is retrieved only in the central part of the study area. Another checkerboard resolution test with a grid separation of about 100 km (Fig. 8b) shows better results. The checkerboard pattern and anomaly amplitude are
Figure 6. (a) Traveltime residuals of the inhomogeneous 3-D inversion in a rms sense versus initial slab thickness taken as a priori information in the inversion. The P-wave velocity of the slab is assumed to be 4 per cent higher than that of normal mantle. (b) Rms traveltime residuals versus initial fractional P-wave velocity perturbation for a 70 km thick subducted slab.

correctly reconstructed for depths of 150 and 200 km in the central part of the KSZ. In conclusion, the resolution of the 3-D inversion is good for most of the study area and it is best in the central part of the KSZ where the correct test pattern is recovered down to a depth of 200 km with a spatial resolution of 50–100 km.

4 DISCUSSION

The results of the 3-D tomographic inversions (Figs 9 and 10) show a prominent high-velocity zone in the upper mantle dipping landwards, which may be associated with the subducted Pacific plate. The introduction of a subducted Pacific plate as a priori information in the initial model gives a final rms traveltime residual smaller than that of the inversion with a laterally homogeneous starting model. Therefore, the tomographic image obtained with the 3-D tomographic inversion using an inhomogeneous starting model was taken as the final result. The inversion suggests that the subducting Pacific plate has an average thickness of about 70 km and a P-wave velocity 2–7 per cent higher than that of normal mantle. The estimated average thickness of the subducted slab fits the results obtained from studies of Rayleigh-wave dispersion data (e.g. Leeds et al. 1974) of an oceanic lithosphere with an age of ≈ 80 Ma, similar to that of the Pacific plate at the KSZ (Rea et al. 1993). The Wadati–Benioff seismic zone is located near the upper surface of the subducted slab (Fig. 10). The lower plane of a double-planed seismic zone (Gorbatov et al. 1994), which can be seen in profiles (a) and (b), is traced approximately in the central part of the subducted slab. The Wadati–Benioff zone extends down to a maximum depth of ≈ 500 km in southern Kamchatka, and ≈ 300 km in the northern part of the peninsula (Gorbatov et al. 1997). Unfortunately, the regional seismic data used in the study allow us to image only structures shallower than ≈ 200 km. Regional and teleseismic data that trace rays to depths greater than 200 km would be necessary to determine the deeper structures in the KSZ.

A sharply defined feature is the broad low-velocity anomaly located beneath the volcanoes and apparently associated with the present volcanic activity in the peninsula (Fig. 9). The low-velocity zone is clearly defined beneath the volcanic front of Kamchatka, from the upper crust down to a depth of ≈ 100–150 km (Fig. 10). The largest negative velocity perturbations (≈ −6 per cent) occur at a depth of ≈ 30 km and are observed along the whole volcanic front (Fig. 9). These low-velocity perturbations form a continuous belt that is almost linear beneath the volcanic front between 51°N and
increases, reaching a depth of \( \approx 150 \) km. Although the volcanic front coincides approximately throughout the KSZ with the depth of the subducted slab of \( \approx 100 \) km (Fig. 5), the low-velocity zones related to the active volcanoes extend to depths of \( \approx 150 \) km (Fig. 10). Unfortunately, the lack of resolution in the northern part of the KSZ (profiles e and f in Fig. 10) limits our observations to a depth of \( \approx 100 \) km.

The good resolution of the inversion for most of the study area down to a depth of 60 km (Fig. 8) provides the opportunity of correlating the resulting velocity heterogeneities, located at depths of about 10 km, with the upper crustal geology of the KSZ. For example, a low-velocity band is spatially correlated with the Kamchatka trench (Fig. 9). Apparently, this low-velocity band is associated with the accretionary prism. In particular, the low-velocity zone (\( \approx 4 \) per cent) in Kamchatsky Bay correlates with the large sediment deposits located in that area (e.g. Dickinson 1978; Selivestrov 1983).

Other prominent and shallow low-velocity zones are observed in the Central Kamchatsky graben and probably correspond to the large sedimentary basins, which reach a thickness of up to \( \approx 5 \) km (e.g. Moroz 1984; Zinkevich 1993). Several local high-velocity structures exist in the upper continental crust (10 km layer in Fig. 9). Most of these zones are located close to the volcanoes and probably correspond to intrusive bodies. On the other hand, one of these zones can be seen clearly about 25 km east of the Bakening volcano on profile (b) in Fig. 10. Apparently, this high-velocity zone is associated with the Gansalsky–Valaiginskoe uplift, formed by metamorphic rocks and uplifted Cretaceous basement (Rikhter 1993; Konstantinovskaya et al. 1993; Zinkevich 1993). Shipunsky Cape (Fig. 1) is also characterized by the presence of high-velocity zones in the upper crust (Figs 8 and 9). The Cape is comprised of uplifted exotic Cretaceous basement (Zinkevich 1993) which is apparently responsible for the high-velocity structures of that area.

5 CONCLUSIONS

3-D body-wave tomography was used to study the seismic velocity structure beneath the Kamchatka peninsula. A total of 5270 regional earthquakes and 32 seismic stations were used to infer detailed P-wave tomographic images down to a depth of 200 km. The results suggest that the subducting Pacific plate has an average thickness of about 70 km. The P-wave velocities in the subducted plate are \( \approx 2–7 \) per cent higher than that of normal mantle. Prominent low-velocity zones exist in the crust and in the upper mantle beneath the active volcanoes down to a depth of \( \approx 150 \) km. The lowest values of these velocity anomalies lie at a depth of \( \approx 30 \) km beneath the volcanic front. High-velocity zones were found in the upper crust that apparently correlate with the intrusive bodies, uplifted metamorphic rocks and Cretaceous basement that are observed in the local geology. The low-velocity zones in the upper crust are associated with the thick sedimentary basins of the Central Kamchatsky graben and with the accretionary prism of the Kamchatka trench.

ACKNOWLEDGMENTS

We thank H. Inoue for useful discussions and suggestions. We benefited from discussions with Yu. Taran, A. Gusev and N. Shapiro. U. Achauer and H. Bijwaard provided us with helpful comments and suggestions which improved the manuscript.
Figure 9. Tomographic image of Kamchatka shown as fractional P-wave velocity perturbations for various depth slices. Red and blue shading shows slow and fast velocities, respectively, according to the scale shown on the right. The depth of each layer is shown in the lower right corner of each map frame. White triangles are volcanoes. Dashed lines mark the Kamchatka and Aleutian trenches.
Figure 10. Tomographic image of Kamchatka shown as fractional $P$-wave velocity perturbations projected on 40 km wide cross-sections (Fig. 1). Red and blue shading reflects the slow and fast velocities, respectively, according to the scale shown at the bottom. Purple triangles are volcanoes and white dots are the earthquakes used in the inversion. A light-coloured mask covers the area where resolution is considered to be poor. Vertical black bars mark the location of the coastline and arrows indicate the location of the trench.

REFERENCES


