

A Study in the Velocity Structure of December 5, 1997, $M_w = 7.8$ Kronotskii Rupture Zone, Kamchatka

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Abstract—The object of the present study was to obtain and investigate the 3D velocity structure of the rupture zone of a large earthquake, to be specific, the great ($M_w = 7.8$) Kronotskii earthquake that occurred in Kamchatka on December 5, 1997. The event was preceded by a foreshock swarm (December 3–5, 1997) and followed by a long aftershock sequence. We investigated the V_p velocity distribution for different time periods: December 3–7, 1997 (when the chief events occurred, viz., the main shock and the larger aftershocks) and for subsequent periods of decaying aftershock activity until December 1998. The velocity distribution in the rupture zone proved to be inhomogeneous. Three regions have been identified: the northeastern (the main shock and foreshocks), the central, and the southwestern, which differ both in the character of seismicity and in velocity. The V_p distribution was found to be time-dependent. The velocity was below the standard values in the foreshock–aftershock area in December 1997, subsequently the velocity increased. These results may indicate the absence of a continuous rupture zone, with the main shock and the two largest aftershocks that occurred in the southwest probably being independent events rupturing a transverse fault during the stress rearrangement following the main shock.

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INTRODUCTION

The assessment of the state of the earth in the rupture zone of a large earthquake before the main shock and immediately after it, during the aftershock sequence, as well as determination of the state of the earth when the aftershocks have stopped occurring, is certainly an important and poorly investigated problem. This problem has been most commonly studied in lab specimens. However, estimation of the velocity after the main rupture has occurred is rather difficult in the laboratory. Regarding studies in the rupture zones of large earthquakes, mention should be made of S.S. Aref'ev's work [1] in which the velocity field was estimated by seismic tomography. The present study is the first attempt at reconstructing the velocity field within a rupture zone based on P and S travel times of aftershocks recorded by the local seismograph network. The advantages of this method are set forth in [5–10, 13] and will be discussed below.

The data are those recorded by the telemetered network, Kamchatka Seismological Technique Testing Team, Geophysical Service, Russian Academy of Sciences (KSTTT in what follows). The hypocenter parameters were computed by a standard method employed at the KSTTT. No corrections have been applied to the raw data.

The $M_w = 7.8$ (energy class $K_S = 15.5$) earthquake which occurred near the Pacific coast of Kamchatka off

the Kronotskii Peninsula at 11:26 a.m. December 5, 1997, is the largest to have been recorded in the area during the period of detailed seismological observations (1961–1997). The foreshock swarm began on December 3 in the northeast part of the future main-shock rupture zone.

The main shock ($M_w = 7.8$) occurred at 46 hours after the start of the foreshock swarm, somewhat to the south of the foreshock area at a depth of 10 km. The instrumental epicenter is based on 28 P and 3 S phases recorded by the KSTTT regional network of telemetered and analog seismic stations [3].

The aftershocks began occurring immediately after the main shock and involved the depth range 1–50 km. The rupture zone in map view is described by an ellipse (140–220 km) situated along the Benioff zone (azimuth 37°). The ellipse is centered at the point $\phi = 54.15^\circ$ N, $\lambda = 162.1^\circ$ E [3].

From the figures presented in [3] one clearly observes a mosaic, nonuniform distribution of aftershocks in the rupture zone, a comparatively low density of epicenters in the central part of the rupture zone, and a northwest trending epicenter concentration. The rupture process was related, in the opinion of Levina et al. [3], to activation of both northeast–southwest and northwest–southeast tectonic features. The Kronotskii earthquake rupture process involved strong fault activation in the southwestern part of the rupture zone.

The reconstruction of mainshock rupture propagation based on inversion of broadband P waves recorded by the world network allows association of aftershock concentrations with the asperities ruptured by the earthquake. The SW asperity zone, where the slip reached 90 cm, is characterized by increased aftershock activity, while the activity was much lower in the NE asperity zone where the slip was greater (up to 240 cm) [3].

We have tried to reconstruct the P -wave velocity field in the Kronotskii foreshock–aftershock zone using the reciprocal wave technique [5]. The technique can be used to find seismic velocities within hypocenter concentrations based on travel times between a set of hypocenters and several seismic stations in the region.

CHOICE OF COMPUTATION PARAMETERS

The technique for finding compressional velocities in a 3D earth is based on the principle of reciprocity of wave fields. The use of that principle allows one to treat travel times of a seismic wave from small earthquakes to a seismic station as travel times of the reciprocal wave traveling from that station to the hypocenters in the area of study.

The problem of computing seismic velocity in the vicinity of a point $M_0(x_0, y_0, z_0)$, which is a hypocenter, reduces to the problem of finding the front of the reciprocal wave and its velocity in a region G around the point. The velocity is assumed to be smooth enough so that it can be treated as constant in G .

The region G should be large enough to contain at least six other hypocenters with travel times recorded at the station. On the other hand, the region should be small enough so that the seismic velocity can be assumed to be constant in it and so that the front of the reciprocal wave can be fitted by a plane.

The main features of the algorithm are as follows:

- the seismic velocity is found in the region of the hypocenters only;
- the differences of first arrival times are only used for a set of hypocenters at a fixed station, thus reducing the influence of the earth structure under the station and along the wave path;
- differences of only hypocentral coordinates are used in a set of hypocenters to reduce the influence of systematic errors in hypocentral coordinates.

We used a Cartesian set of axes centered at the center of the aftershock ellipse for the Kronotskii earthquake as determined in [3] ($\varphi = 54.15^\circ$ N, $\lambda = 162.1^\circ$ E), the coordinate axes being identical with the axes of the ellipse. The positive OX -axis points northwest and OY southwest.

The most essential problem in the entire algorithm is the choice of parameters for G which specify the size of spatial heterogeneities in the resulting velocity field. This study assumes the averaging region G to be a vertical empty cylinder 9 km high and the radii of the base

Table 1. Number of points, average velocity V_{ave} and velocity uncertainty $\sigma(V_{ave})$ in the 15–35 km layer for the period December 3 to 31, 1997

	Region I	Region II	Region III
Number of points n	121	147	229
Average velocity V_{ave} , km/s	6.05	6.34	6.25
$\sigma(V_{ave})$, km/s	0.06	0.05	0.04

equal to $R_{min} = 5$ km and $R_{max} = 10$ km. The point M_0 is set at the center of G .

IDENTIFICATION OF 3D ZONES AND TIME INTERVALS

A total of 1396 earthquakes were located in the rupture zone by KSTTT workers in December 1997 [3]. Compressional velocities were determined from data for 1095 events only, with most of the points being in the depth range 15–35 km. Preliminary examination of the compressional velocity in that layer revealed considerable velocity heterogeneity along the zone in question.

The average velocity for all points considered in this depth range of 15–35 km was 6.23 km/s. However, the average velocity was appreciably lower in the northeast. We identified three regions of hypocenter concentration where the P velocity was determined, starting from the Kronotskii rupture zone from the north:

Region I, northeast: $-120 \text{ km} \leq Y \leq -60 \text{ km}$;

Region II, central: $-60 \text{ km} \leq Y \leq 10 \text{ km}$;

Region III, southwest: $10 \text{ km} \leq Y \leq 100 \text{ km}$.

Table 1 lists the number of points n , the average velocity V_{ave} and the associated uncertainty $\sigma(V_{ave})$ in the layer 15–35 km for the regions selected based on the data for the entire period December 3 to 31, 1997.

All subsequent calculations of compressional velocities for all depths were carried out separately for each of the three regions. This somewhat reduced the number of points for which the velocity was computed.

As the main object of this study was to reveal the time behavior of the velocity field during the evolution of the Kronotskii earthquake process, all the data were divided into three time periods, viz., December 3 to 7, 1997; December 8 to 20, 1997; and December 21, 1997, to December 31, 1998.

Table 2 gives average P velocities for the layers specified there during the time intervals considered. The seismic process is seen to be largely nonstationary, the aftershocks being nonuniform over the area. Most of the seismicity in the north of the zone was concentrated at depths of 5–35 km, with the velocity being determined at 107 points for the period December 3 to 7, 1997, and at 20 only for December 8 to 20, 1997. In contrast to this, the second region contains only

Table 2. Average velocities V_P , km/s and n —the number of points at which the velocity has been determined—for three spatial regions and three time intervals at various depths

Region	Depth range, km		Dec. 3–7, 1997	Dec. 8–20, 1997	Dec. 21, 1997–Dec. 31, 1998
Region I, northeast	0–5	V_P			
		n	4	0	0
	5–15	V_P	5.51		5.29
		n	53	4	41
	15–35	V_P	5.96	6.17	6.23
		n	50	16	221
Central region (II)	Total number of points		107	20	262
	0–5	V_P		3.81	3.98
		n	0	19	5
	5–15	V_P	5.53		5.44
		n	10	0	34
	15–35	V_P		6.21	6.24
		n	2	58	440
	35–55	V_P	6.61	6.62	6.78
		n	10	23	65
	Total number of points		22	100	544
Region III, south-west	0–5	V_P	3.48	3.56	3.41
		n	87	50	126
	5–15	V_P	5.22	5.50	5.27
		n	22	29	86
	15–35	V_P	6.11	6.23	6.31
		n	35	48	265
	35–55	V_P	6.63	6.73	6.84
		n	10	65	381
	Total number of points		144	192	858

22 points where velocities were determined for the first period. The seismicity of the third region was the highest, enough hypocenters for velocity determination were available at all depths in the range 0–50 km.

This space–time heterogeneity in hypocenter distribution subsequently compelled us to combine the time intervals in order to construct vertical velocity cross-sections in the first and second zones, so that calculations were made for the period December 3 to 20, 1997. When the number of velocity points was below 10, we regarded these values as not accurate enough.

It is seen from the numbers of points as given in Table 2 (i.e., hypocenters in several depth ranges) that the Kronotskii rupture zone has a rather complex structure. It appears that V_P for the three regions was increasing with depth during the time intervals selected. Considerable activity was recorded in the 15–35 km layer in regions I (northeast) and III (southwest). The P velocity went as low as 6.0 km/s at that location. During the subsequent seismic activity (1998) the velocity increased to reach 6.2–6.3 km/s. The velocity variation over depth

and the time intervals will be discussed below as the velocity cross-sections are described.

For the period December 3 to 20, 1997, we compared the positions of foreshocks and large aftershocks with the $K_S \geq 13$ (Table 3) that actually occurred on December 5 and 6, 1997, as well as the numerous aftershocks in whose vicinities the velocity was determined for that period. We emphasize that there should be at least six hypocenters within 10 km of a point in order to determine the velocity at that point by the algorithm we are using. Velocities are thus indeterminate in regions of sparse hypocenters. A preliminary suggestion can thus be ventured that higher seismic activity was recorded where velocities have been determined.

Figure 1 shows the positions of the main shock, foreshocks and aftershocks with $K_S \geq 13$ (Fig. 1a), as well as the hypocenters of numerous smaller earthquakes distributed over depth (Figs. 1b and 1c) where the velocity was determined for the period December 3 to 20, 1997.

Table 3. Hypocenter parameters for the main shock and large ($K_S \geq 13.0$) aftershocks in the area of the December 1997 Krontskii earthquake

Date	Time	φ°	λ°	X, km	Y, km	Z, km	K_S	Region
5 Dec 97	11:26:51.0	54.64	162.55	7.04	-61.4	10	15.5	I
5 Dec 97	11:35:24.1	53.86	161.19	30.2	60.2	32	13.9	III
5 Dec 97	11:48:38.5	54.29	162.46	-10.3	-26.1	10	13.5	II
5 Dec 97	18:48:21.2	53.68	161.85	-16.4	52.1	24	14.2	III
5 Dec 97	19:04:03.9	53.82	161.72	-0.6	44.2	8	13.0	III
6 Dec 97	06:38:09.6	54.07	160.83	62.7	54.4	0	13.1	III
6 Dec 97	10:05:00.2	53.78	162.42	-40.5	21.8	13	13.1	III
6 Dec 97	10:59:07.6	53.85	162.03	-15.3	29.9	7	13.6	III
6 Dec 97	12:37:21.8	54.82	162.24	35.0	-66.3	21	13.2	I
7 Dec 97	23:05:49.2	53.66	162.07	-29.5	45.8	23	13.1	III

Comparing the positions of the larger aftershocks in Fig. 1a and the concentrations of smaller events in Figs. 1b and 1c, one notices that the velocity in the north of the region, where the foreshocks had been recorded, is relevant to depths of 5–35 km. The points in the depth range 15–35 km in the south of region I are related to the December 6, 1997, hypocenter (no. 9 in Table 3) whose coordinates are 54.82° N and 162.24° E at a depth of 21 km. However, the mainshock hypocenter is outside the region of subsequent smaller aftershocks.

In the second region one notes a spatial relationship between the December 5, 1997, earthquake (Table 3) whose coordinates are 54.29° N and 162.46° E at 10 km depth and the many hypocenters in the depth range 5–15 km.

In the second region we see considerable seismicity at depths of 15–35 km as well, but this occurred since

December 8, 1997, rather than during the first day, which may be related to the activity in the Benioff zone.

Seven events of $K_S \geq 13$ occurred in the south of the third region, among them two of the largest aftershocks with $K_S = 13.9$ and $K_S = 14.2$. The large events were accompanied by the occurrence of smaller shocks at all depths considered. The active zone trends northwest–southeast across the Benioff zone. A fault striking in this direction may have showed higher activity or else an asperity failed. It should also be mentioned that the boundary of the aftershock zone of the 1923 and 1959 earthquakes with $M = 8.5$ and 7.6 , 7.0 is located in this zone [11].

A STUDY OF THE COMPRESSIONAL VELOCITY FIELD

The procedure described above was used to find V_p in the selected regions and time intervals. Because the aftershock distribution in the rupture zone is mosaic in

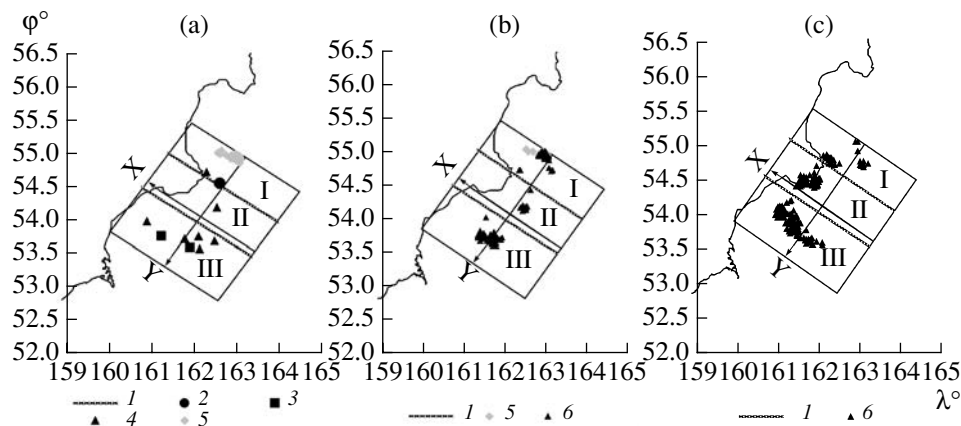


Fig. 1. A map of the study area: (a) main shock, foreshocks, and large aftershocks, distribution of foreshocks and aftershocks for the period December 3–20, 1997, in the depth range (b) 5–15 km and (c) 15–35 km. (1) region boundary, (2) main shock, (3) $K_S \geq 13.4$ aftershocks, (4) $13.0 \leq K_S \leq 13.3$ aftershocks, (5) foreshocks, (6) earthquake hypocenters for which velocities have been determined.

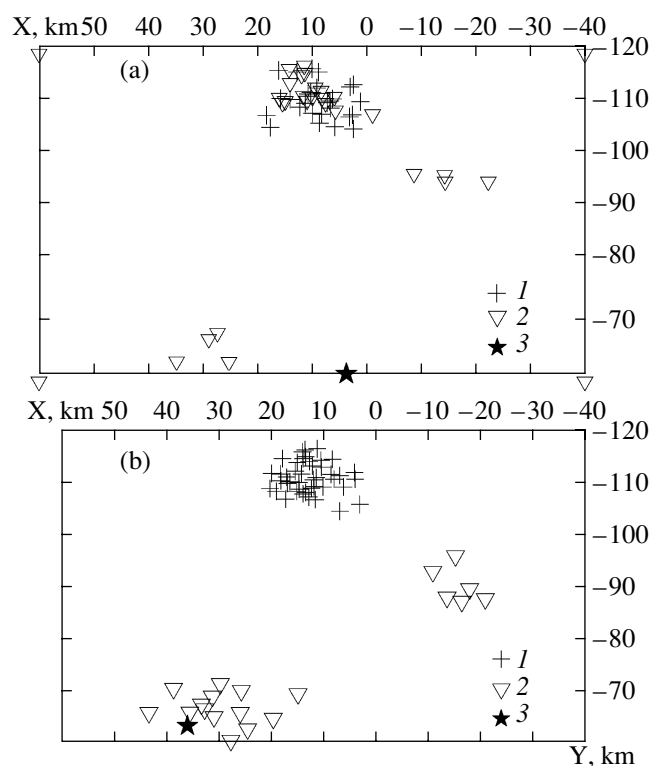


Fig. 2. Hypocenters in region I for which velocities have been determined based on the December 3, 1997–December 20, 1998, data: (a) 5–15 km layer, (b) 15–35 km. (I) foreshocks, (2) aftershocks, (3) hypocenters of large ($K_S \geq 13$) earthquakes.

character, and areas of no $K_S \geq 8.5$ aftershocks are available, it has not been possible to make velocity distribution maps for the entire rupture zone in different depth ranges. Averaging over a horizontal cross section would have involved considerable arbitrariness due to the averaging method chosen. Filling the empty grid points with the average velocity for the selected layers did not yield good, reliable results. For this reason the results are given below as cross sections of velocity against depth for the selected regions and time intervals. To assess the reliability, and to get an idea of the amount of data on which the cross sections are based,

Table 4. Average V_P and the number of hypocenters n where the velocity has been determined in region I for the period December 3 to 20, 1997

Depth	5–15 km		15–35 km	
	V_P , km/s	n	V_P , km/s	n
Foreshocks, December 3 to 5, 1997	5.48	49	6.01	46
Aftershocks, December 5 to 20, 1997	5.72	8	6.06	26

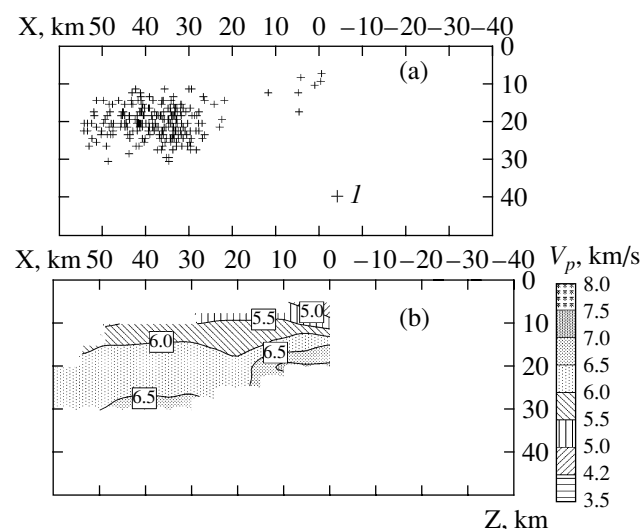


Fig. 3. Results from calculation of V_P in region I based on the December 21, 1997–December 31, 1998, data. (a) Hypocenters for which velocities have been determined, (I) hypocenters; (b) vertical cross-section of $V_P(Z)$.

we show the distribution of the hypocenters in whose vicinities V_P has been determined.

We now consider the hypocenter distribution in the selected zones and the vertical cross sections of V_P based on these.

In region I the positions of foreshocks and aftershocks where the P velocity was determined are concentrations of points in the 5–35 km layer. Figure 2 shows the points in the 5–15 and 15–35 km layers based on the data recorded from December 3rd to December 20th, 1997. The position of the large ($K_S = 13.2$, depth 21 km) aftershock is marked in the 15–35 km layer.

It has not been possible to construct the vertical or horizontal cross sections of the velocity field. Table 4 lists the average velocity V_P in some hypocenter concentrations at different depths and the number of epicenters for which the velocities have been determined.

The velocity values in both layers are below the standard ones (these latter were assumed to be those for the Kuzin model which are used for hypocenter location [2]), especially in the 15–35 km layer. During the subsequent period, aftershocks were only recorded in the south of region I, so that Fig. 3 shows the hypocenter distribution and V_P values in the vertical XZ-plane.

During the period December 3–20, 1997, the average velocity in the 15–35 km layer was 6.0 km/s (Table 4), but increased later to reach 6.23 km/s (Table 2). The velocity field of this region as displayed in Fig. 3b contains isolines of 6.5 and 7.0 km/s.

The vertical velocity cross-sections and the positions of hypocenters in region II at which V_P has been determined are shown in Figs. 4 and 5.

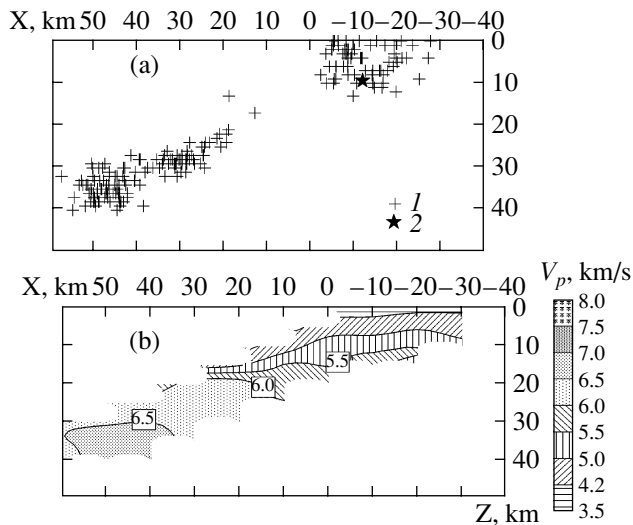


Fig. 4. Results from calculation of V_p in region II based on the December 3–20, 1997, data: (a) hypocenters for which velocities have been determined, (b) vertical cross-section $V_p(Z)$: (1) hypocenters, (2) $K_S \geq 13.0$ aftershocks.

The hypocenters make two regions of hypocenter concentration in both time intervals. One zone is denser with hypocenters, dips landward at an angle of 30 degrees, and extends from 12–15 km depth to 40–45 km; it is likely to be associated with the Benioff zone. The other, more compact zone is in the southeast at depths of 0 to 15 km. The activity in that zone decayed after December 8, 1997 (see Table 2).

The velocity field in the cross sections is an inclined structure extending from southeast to northwest at depths of 0–20 km in the southeast to 40–45 km in the northwest. The velocity varies from 4 km/s in the upper part to 7 km/s in the lower, the 6.5–7.0 km/s layer plunging from 18 km depth in the southeast to 40–45 km in the northwest. The central part of the cross section, $X = 10$ –40 km, (Fig. 5) has an increasingly thicker layer bounded by the isolines of 6.0–6.5 km/s, 5 to 18 km in thickness.

The velocity field in the Benioff zone ($X \geq 10$ km) at depths of 12–15 km contains a gradient zone with velocity increasing from 5.0 to 6.0 km/s (Fig. 4). However, the velocity is lower at shallower depths for $X \leq 15$ km, with the $V_p = 5.0$ km/s isoline descending below 10 km depth. Subsequently, after December 20, the velocity increased in that region, with the $V_p = 6.0$ km/s isoline being located at 15 km depth (Fig. 5).

The most numerous aftershocks were recorded in the southern region III at depths of 0–45 km (Figs. 6–8). Considerable seismic activity was recorded for $X \leq 0$ km during the first days, December 5–6, 1997, i.e., in the southeastern part of the Kronotskii rupture zone at depths of 0–45 km. Subsequently the activity in the region was decreasing, with aftershocks being mostly recorded for $X \geq -10$ km after December 20, 1997.

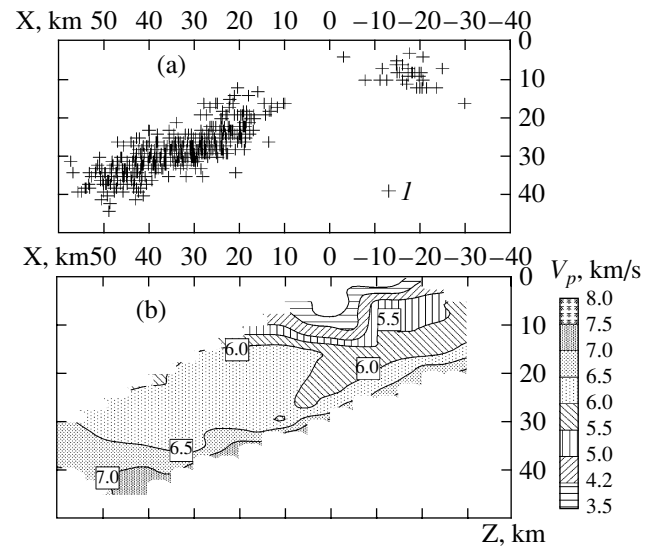


Fig. 5. Results from calculation of V_p in region II based on the December 21, 1997–December 31, 1998, data: (a) hypocenters for which velocities have been determined, (b) vertical cross-section $V_p(Z)$. Symbols are as in Fig. 4.

For the third region we constructed vertical velocity cross sections for the three time intervals, because these had enough hypocenters. As to the first time interval (we recall that it was in this interval that the largest aftershocks of $K_S = 13.9$ and $K_S = 14.2$ and five events of $K_S \geq 13$ occurred), the increase in activity was in the central part of the cross section, $-20 \text{ km} \leq X \leq 20 \text{ km}$, as well as in its southeastern part for $-40 \text{ km} \leq X \leq -20 \text{ km}$ at

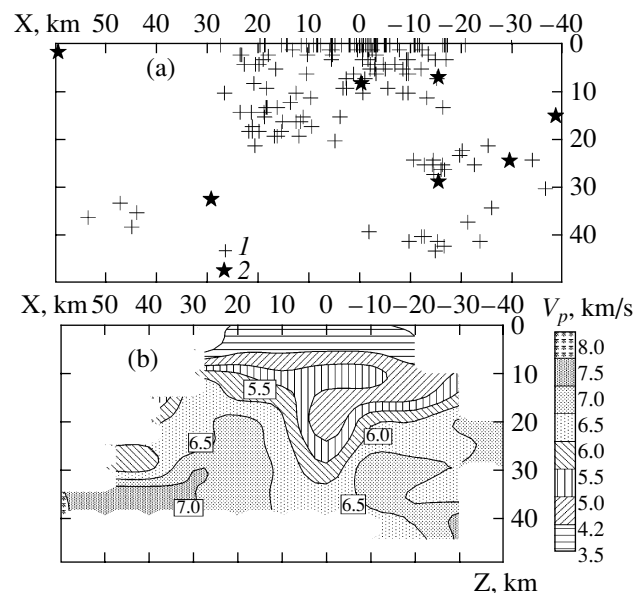


Fig. 6. Results from calculation of V_p in region III based on the December 3–7, 1997, data: (a) hypocenters for which velocities have been determined, (b) vertical cross-section $V_p(Z)$. Symbols are as in Fig. 4.

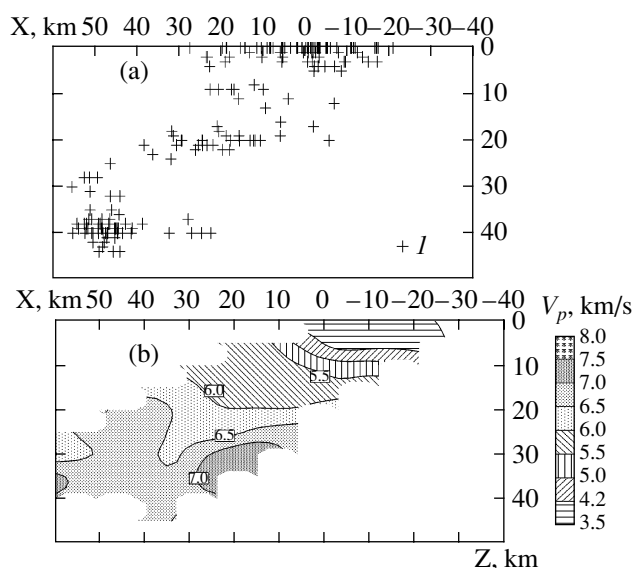


Fig. 7. Results from calculation of V_p in region III based on the December 7–20, 1997, data: (a) hypocenters for which velocities have been determined, (b) vertical cross-section $V_p(Z)$. Symbols are as in Fig. 4.

depths of 0–20 and 20–45 km, respectively. The resulting velocity cross sections show changes in V_p over time as the aftershock process was going on in the region. The velocity field was complex (Fig. 6b) during the first time interval, December 3 to 7, 1997. It is much different from that in late December 1998 (Fig. 7). The velocity values are appreciably below the standard ones for the respective depths. One sees a core of relatively low velocity in the central part of the cross section, $-20 \text{ km} \leq X \leq 20 \text{ km}$ (Fig. 6b), with the isolines of 5.5 and 6.0 km/s penetrating the field of higher velocities (6.5–7.0 km/s) down to 30 km depth. In contrast to this, the isoline of $V_p = 7.0 \text{ km/s}$ rises to reach 20 km depth at the border of the region. It is in this zone that the largest earthquake of this region occurred, with $K_S = 14.2$ and depth $Z = 24 \text{ km}$.

During the second period (December 8 to 20, 1997) a dipping (at an angle of 45 degrees) zone of hypocenters was developing (Fig. 7a) extending from northwest to southeast from the depth 0 to 40 km (the value 40 km is a preliminary one, because the hypocenter band at these depths may indicate hypocenter depth uncertainties).

The velocity field changed later (December 8 to 20, 1997). Low velocities (down to 6.0 km/s) were now observed down to a depth of 20 km only (Fig. 7b), and the isoline $V_p = 6.5 \text{ km/s}$ was seen at depths shallower than 30 km, i.e., the velocity in that region was increasing. High velocities (7.0–7.5 km/s) appeared at depths of 25–40 km in the southeastern part of the cross section.

Figure 8, based on the December 20, 1997, to December 31, 1998 data, shows a quiet velocity field,

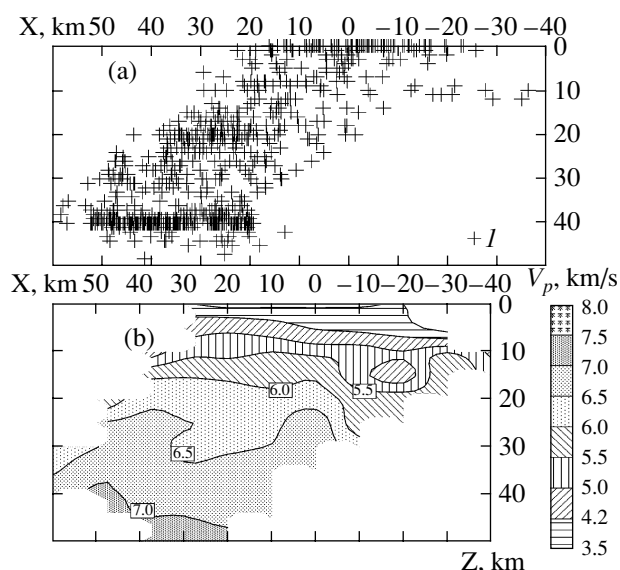


Fig. 8. Results from calculation of V_p in region III based on the December 21, 1997–December 31, 1998, data: (a) hypocenters for which velocities have been determined, (b) vertical cross-section of $V_p(Z)$. Symbols are as in Fig. 4.

the isolines are nearly horizontal, with the velocity increasing with depth from 4.2 km/s in the upper portion to 7.0 km/s at ~40 km depth.

A DISCUSSION OF THE RESULTING VELOCITY FIELD

The assessment of the state of the medium in the rupture zone of a large earthquake and during the aftershock sequence and the recovery of the medium after the aftershocks have died away is a poorly studied problem. Among the studies of rupture zones of large earthquakes we mention that by S.S. Aref'ev of the $M = 7.6$, 1995 Neftegorsk, Sakhalin I. earthquake rupture zone by tomography [1]. He notes that the velocity field seems to be inhomogeneous. There are relatively low velocity values in the rupture zone compared with those obtained by exploration geophysics and low velocities at great depths. However, S.S. Aref'ev did not notice or study the variation of velocity over time during the aftershock sequence.

We have determined and studied the 3D velocity field structure of the Kronotskii earthquake rupture zone. It turned out that compressional velocity is lower immediately after a large earthquake than is usually supposed at a given depth. The velocity was increasing during the decay of aftershock activity in 1998, i.e., the velocity structure was returning to its original state. We note that the velocity in the foreshock area is also appreciably below the standard values according to [2]. This result may be compared with the data obtained by Gorbatoev and Gordeev [12]. Their 1D velocity model features $V_p = 5.74 \text{ km/s}$ in the 5–15 km layer, 6.74 km/s for 15–35 km depth, and 7.37 km/s deeper than 35 km.

According to our calculations, the velocity in the 15–35 km layer was as low as 6.0–6.3 km/s in December 1997.

Comparison of the resulting velocity cross sections in the three regions (north, center, and south of the Kronotskii earthquake zone) reveals appreciable differences, both in the numbers of small earthquakes and in the velocity for those regions. It can thus be asserted that the Kronotskii earthquake rupture zone is not a single whole regarding velocity, but consists of tectonically different features showing increased activity during the release and rearrangement of the stress field. There was a noticeable area of foreshocks in the north where small earthquakes also occurred after the main shock. However, no significant seismic activity was observed there later. The central part of the study area contained a region of aftershocks associated with the Benioff zone. A complex transverse feature has been identified in the southwest, probably a fault that jumped into activity during the stress rearrangement. We also note that same feature (transverse with respect to the Benioff zone) may have acted as a barrier for the propagation of aftershocks following the large earthquakes of 1923 and 1959 [11].

Our calculations of the velocity field show that there was no continuous rupture zone, i.e., no simple rupture. The main shock and its two largest aftershocks which occurred 9 minutes and 7 hours in the southwest after the main event are likely to have been independent events occurring at the southwestern boundary of the underthrusting Pacific plate during the stress rearrangement after the main shock. They do not limit, as is usually supposed, the end of the rupture, but occurred during the stress rearrangement after the main shock and are confined to the transverse feature, which is an independent fault and a boundary to the activated block. The spatial distribution of small earthquakes is quite consistent with the “piano–keyboard model” of the Benioff zone [4].

Our calculations of compressional velocity certainly prove that the velocities vary both in space and time during the aftershock sequence. We have remarked on a similar effect when investigating the velocity field observed during the 1996 eruptions of Karymskii and Akademii Nauk volcanoes [6, 7, 9].

The smallest velocity variation was recorded at depths of 40–50 km. This may indicate a decay and limitation of the activity at this depth. It should also be emphasized that the velocity field was reverting to its original values by the end of 1998, and this process is here related to a recovery of the state of the medium, i.e., fault healing.

CONCLUSIONS

The present method for velocity determination based on inversion of *P* and *S* waves, as well as the use of the difference principle for finding the wave front

and the velocity of its propagation, yield satisfactory results in regions of hypocenter concentration in zones of seismic and volcanic activity.

The Kronotskii earthquake rupture zone has a complex structure, consisting as it does of three regions that differ both in the velocity and the morphology of the velocity field. The field was found to vary over time in the selected regions during the aftershock sequence. In zones of higher activity where hypocenters of larger aftershocks occurred, we found lower velocities and velocity penetration to greater depths.

The velocity field was found to have recovered its original values by the end of 1998, i.e., during the decay of the aftershock process. This we relate to the fault healing process going on in the earth. The smallest changes in velocity were observed at depths of 40 km or greater. This may indicate that these depths were not involved in the rupture.

We believe that the main shock, which occurred at 11 h 26 min on December 5, 1997, ($M_w = 7.8$, $M = 7.0$, $K_S = 15.5$ as reported in the KSTTT catalog) in the northeastern part, and the two larger “aftershocks,” occurring in the southwestern region on December 5, 1997, with $K_S = 13.9$ and $K_S = 14.2$, are interconnected, but independent events. They did not limit, as is usually supposed, the end of the main rupture, but occurred during the stress rearrangement after the main shock, and were confined to a transverse feature which is an independent fault and serves as a boundary to the activated block.

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