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Inhomogeneities of the Eurasian mantle structure from the travel-times of the nuclear explosions recorded by the Finnish seismic network during 1961–1985

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SUMMARY

During the period 1961–1985, hundreds of nuclear explosions from the area of the Soviet Union and China were well recorded at Finnish seismological stations. The seismic waves recorded on the Baltic (Fennoscandian) shield penetrated through the mantle of the Siberian platform, the Ural Mountains and the East European platform. From the known crustal models, the functions describing the increase of the average velocities of sedimentary cover and crystalline complex of the crust with thickness were found. The corrections for sedimentary cover and crustal thickness were included. The large amount of data permitted the calculation of traveltimes of *P* waves for three sectors up to distances of about 5000 km. In all mantle models, the boundaries '400' and '700' km were found. Comparison of the results shows a difference in the traveltimes of the order of 5 s at a distance of about 4000 km, which reflects the mantle structure differentiation for depths greater than 700 km (lowest velocities for latitudinal direction and highest velocities for longitudinal direction). The average *S*-wave velocity model of the mantle was obtained using the traveltimes of *S*-wave first arrivals. High values of the V_P/V_S ratio were found in the depth interval 200–400 km, while in other depth intervals they were close to 1.73. Our 1-D models are compared with and discussed in connection with other models of the East European and Siberian platforms as well as with global tomographic solutions.

Key words: Earth's mantle models, Eurasia, nuclear explosions, *P*- and *S*-wave traveltimes.

1 INTRODUCTION

During the period 1961–1985 Finnish seismological stations recorded very clearly the nuclear explosions from the area of the Soviet Union and China. Such sources are especially suitable for a study of the mantle structure because of the simplicity of the source function and the absence of depth phases. The purpose of the present paper is to present an analysis of the traveltimes of body waves for the Earth's mantle of the Eurasian continent. The events from Novaya Zemlya were not included in this analysis because the explosions took place very close to the Finnish seismological stations and the area of the Barents Sea differs distinctly from typical Eurasian continental structure. The locations of the nuclear explosions analysed and the major tectonic units of Eurasia are shown in Fig. 1.

The records of over 400 nuclear explosions recorded by 17 Finnish stations were used to study the inhomogeneity and azimuthal differentiation of mantle structure. The locations of the stations are listed in Table 1 (Teikari & Suvilinna 1980; Ahjos *et al.* 1989). The geographical coordinates, origin times and magnitudes of events were taken from the bulletins of the ISC, NORSAR and FOA.

2 TRAVELTIME DATA AND 1-D MODELS

The initial data were the arrival times of successive body-wave phases from the Eurasian nuclear explosions recorded by Finnish seismological stations. The data were collected during 25 years of station operation as routine readings of times and magnitudes. Only observations from stations NUR, KEV and SOD were systematically published in international bulletins. Altogether about 4000 readings of arrival time comprise the preliminary data set. From this number about 2450 are first arrivals of *P* waves and about 200 were interpreted as first arrivals of *S* waves. Other

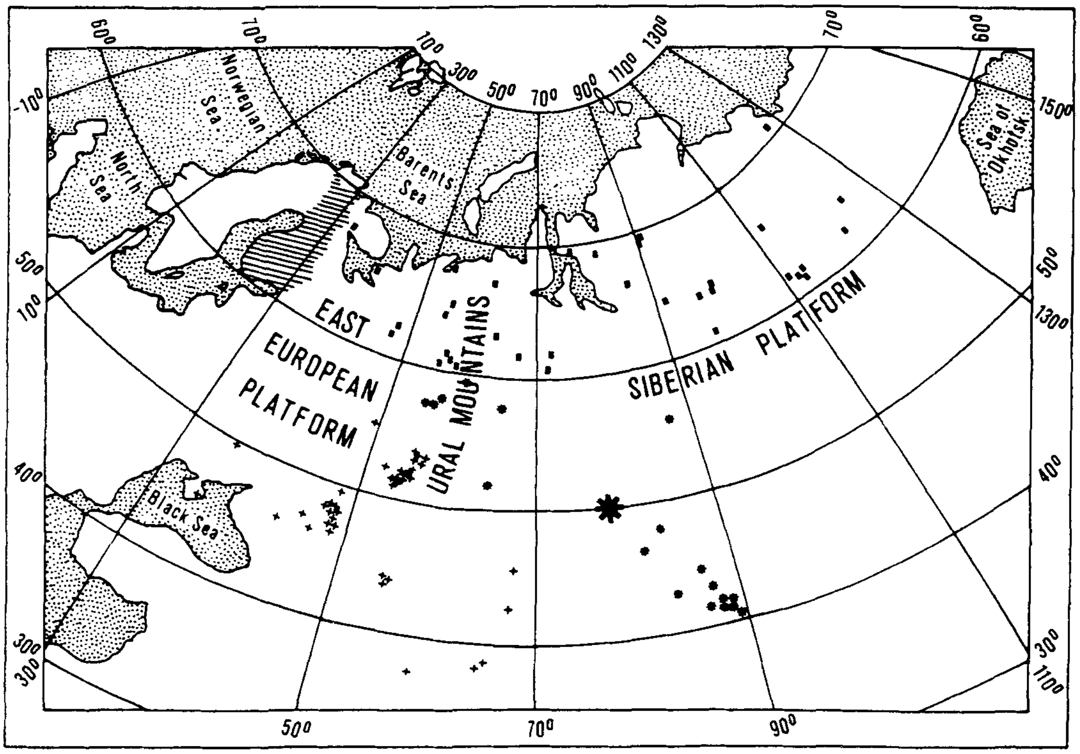

Figure 1. The main tectonic features of the Eurasian area. The locations of nuclear explosions from the East European and Siberian platforms in years 1961–1985 recorded by Finnish seismological stations and discussed in this paper: black squares–sector 1, between 60° and 70° N; stars–sector 2, an average azimuth of about 130°; crosses–sector 3, an average azimuth about of 160°. The large star represents about 200 events in the Semipalatinsk test site. Dashed area–territory of Finland.

phases were provisionally interpreted as reflected or converted body waves and were not used in the present analysis. The observed P-wave traveltimes are shown in Fig. 2, together with theoretical traveltimes for the average MUMEP model for the East European platform (Grad 1987, 1988). The accuracy of the original picks for iP and eP first arrivals is about 0.1–0.2 and 0.5 s, respectively. Taking into account any data selection, and without correcting for the crustal structure, the scattering of the points is not very significant. Practically all the points are scattered within a 10 s band and consequently model MUMEP provides a representative model of the mantle structure for Eurasia.

2.1 Crustal velocities and time corrections

The seismological stations in Finland are located on the crystalline basement of the Baltic shield. In this area the crustal thickness varies from 40 to about 60 km (Luosto 1990). It is obvious that the mean crustal velocity also depends on the crustal thickness. Values for the mean crustal velocities were collected from 1-D and 2-D models for all the deep seismic sounding profiles in Finland. The linear relationship between mean crustal velocity and crustal thickness was calculated for all the data using the

Table 1. Finnish seismological stations.

Station	Lat (°N)	Long (°E)	H (m)	Open	Closed
PKK	60.005	24.517	10	1964	
HYV	60.650	24.770	110	1962	1968
OBB	60.030	24.370	15	1964	1966
PRV	60.357	25.558	30	1964	1979
PRF	60.386	25.681	10	1979	
NUR	60.509	24.651	102	1958	
KAF	62.113	24.306	205	1976	
KEF	62.166	24.871	215	1976	
SUF	62.719	26.151	185	1976	
JOF	62.918	31.312	180	1981	
JOE	62.652	29.695	90	1960	1978
KJN	64.085	27.713	270	1959	
KJF	64.199	27.715	159	1970	1989
OUL	65.085	25.896	60	1963	
SOD	67.371	26.629	181	1956	1992
SO	67.420	26.394	276	1973	
KEV	69.755	27.007	81	1961	

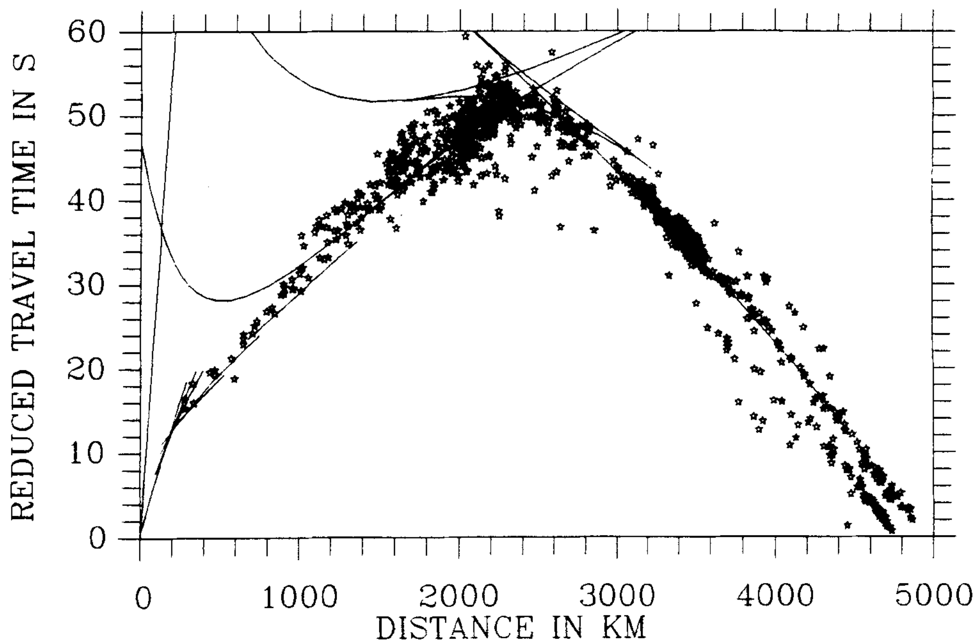

Figure 2. Reduced traveltimes for the nuclear explosions recorded in Finland (stars) compared with the *P*-wave traveltimes for the MUMEP model (solid lines, Grad 1988). This set includes about 2450 arrival times of *P* waves. Reduction velocity $V_r = 10.0\,\mathrm{km\,s^{-1}}$.

least-squares method:

$$\bar{v}(h_M) = 6.10\,(\pm 0.12) + 0.0114\,(\pm 0.0024)\,h_M, \tag{1}$$

where $\bar{v}$ is the mean velocity in $\mathrm{km\,s^{-1}}$, and h_M the crustal thickness in km (Grad & Yliniemi 1991). The standard deviations for coefficients are given in brackets. The mean crustal velocity versus crustal thickness increases from about $6.55\,\mathrm{km\,s^{-1}}$ for a 40 km thick crust to about $6.80\,\mathrm{km\,s^{-1}}$ for a 60 km thick crust. As was shown by Grad & Yliniemi (1991), even for shield regions without any low-velocity sedimentary cover, the observed effect of the crustal structure differentiation was significant, of the order of 0.5 s. The time correction Δt related to differentiation of the crustal structure beneath the seismological station can be expressed in the form

$$\Delta t = \frac{h_M - H_M}{\sin e}\left(\frac{\bar{v}_c - v_m}{\bar{v}_c v_m}\right), \tag{2}$$

where h_M is the Moho depth under the station, H_M the reference depth, e the emergence angle, $\bar{v}_c$ the mean velocity in a crust of thickness h_M, and $\bar{v}_m$ the velocity in the upper mantle, below the Moho. The time corrections for the nuclear explosions were calculated for the reference depth of the Moho $H_M = 44$ km; depths of the Moho h_M under the stations were taken from the map of crustal thickness for the Fennoscandian shield (Luosto 1990); mean crustal velocities $\bar{v}_c$ were calculated from eq. (1); and the average velocity in the uppermost mantle v_m was taken to be $8.2\,\mathrm{km\,s^{-1}}$ (e.g. Luosto *et al.* 1983, 1984, 1985, 1989; Grad & Luosto 1987; Yliniemi 1989; Bannister, Rund & Husebye 1991). The value of the emergence angle e was calculated for the average model MUMEP (Grad 1987, 1988).

The crustal structure of the area of Eurasia investigated is complex and heterogeneous. The crystalline basement usually lies at a depth of 2–3 km, emerging to the surface in the area of the Baltic (Fennoscandian) shield, the Ukrainian shield and the Anabar shield. Characteristic of the marginal zones of the East European platform and Siberian platforms is the existence of large sedimentary basins, where the thickness of sediments exceeds 10–15 km (Polish–German–Danish Depression, Black Sea and Donbass Basins, Ciscaspian Depression, Pechora Basin, Vilyui, Tunguss and Ob-Tasovsk Depressions). In the area investigated the crustal thickness varies widely from 35 to about 60 km (Burmakov *et al.* 1987; Buryanov, Gordienko & Pavlenkova 1980; Egorkin *et al.* 1987; Giese & Pavlenkova 1988; Grad 1992; Guterch *et al.* 1986; Khalevin 1975; Kosminskaya & Pavlenkova 1979; Kosminskaya, Belyaevski & Volvovsky 1969; Kuznetsov 1980; Pavlenkova 1984; Pomeranzeva, Barskova & Mozjenko 1975; Sollogub *et al.* 1978a,b,c; Tarkov & Basula 1983; Yegorkin *et al.* 1980a,b; Zunnunov 1985). Because of such substantial variations in structure, the data on the sedimentary complex and crustal thicknesses, as well as mean velocities in sediments and crystalline crust, are of great importance in our investigations. The analysis of average *P*-wave velocities in sedimentary cover and crystalline crust was made on the basis of data published in more than 30 papers. Altogether, more than 100 values of average velocity in sediments were used in the determination of the power function relating velocity to thickness. More than 160 values of the average velocity in crystalline crust were collected for the shield, platform, orogeny and depression areas of the study area. The relations between mean velocity and thickness were fitted by linear functions. The final results are shown in Fig. 3. For the sediments the relation between mean velocity and thickness has the form

$$v_{sed}(h) = 2.42\,(\pm 0.04)\,h^{0.236\,(\pm 0.009)}. \tag{3}$$

The average relations between mean velocity and thickness

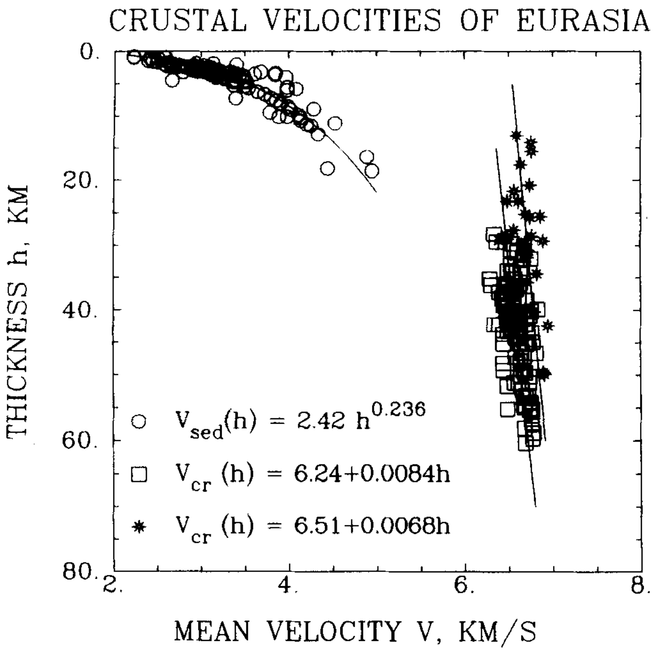

Figure 3. Mean velocities of the sedimentary cover (open circles) and crystalline complex of the Earth's crust for Eurasia (open squares–shield, platform and orogeny areas, stars–deep sedimentary basins and rift zones). Compiled from: Buryanov *et al.* 1980; Egorkin *et al.* 1987; Giese & Pavlenkova 1988; Grad 1976, 1986, 1987, 1988, 1992; Grad & Luosto 1987; Grad & Yliniemi 1991; Grad, Doan & Klimkowski 1991a; Grad, Guterch & Lund; Guterch *et al.* 1986; Khalevin 1975; Kosminskaya & Pavlenkova 1979; Kosminskaya *et al.* 1969; Kuznetsov 1980; Luosto 1986, 1990; Luosto *et al.* 1983, 1984, 1985, 1989; Pavlenkova 1984; Pomeranzeva *et al.* 1975; Sollogub *et al.* 1978a,b; Tarkov & Basula 1983; Vinnik & Ryaboy 1981; Yegorkin *et al.* 1980a,b; Yliniemi 1989; Yliniemi & Luosto 1983; Yliniemi & Grad 1991; Zunnunov 1985.

for shields, platforms, orogeny areas and depressions were fitted by linear functions

$$v_{cr}(h) = 5.97\,(\pm 0.09) + 0.0141(\pm 0.0018)\,h, \quad \text{for shields} \tag{4}$$

$$v_{cr}(h) = 6.47\,(\pm 0.13) + 0.0030(\pm 0.0033)\,h, \quad \text{for platforms} \tag{5}$$

$$v_{cr}(h) = 6.53\,(\pm 0.06) + 0.0001(\pm 0.0013)\,h, \quad \text{for orogeny} \tag{6}$$

$$v_{cr}(h) = 6.51\,(\pm 0.06) + 0.0068(\pm 0.0021)\,h, \quad \text{for depressions.} \tag{7}$$

The average relations for the shield, platform and orogeny areas were very similar, and differences were insignificant. In this case the relation between mean velocity and thickness has the form

$$v_{cr}(h) = 6.24\,(\pm 0.05) + 0.0084\,(\pm 0.0012)\,h. \tag{8}$$

For a crystalline crust of thickness between 30 and 60 km the functions (4)–(6) are within ± 0.1 km s^{-1} of the average line (eq. 8). It gives the difference for time correction according to eq. (2) as smaller than 0.07 s. The velocities of the crystalline crust in the areas of depressions are on average about 0.3 km s^{-1} larger than values determined using eq. (8). In the last six (eqs 3–8) the thickness is given in km, and velocity in km s^{-1}. The results for sediments (eq. 3) and for crystalline crust (eqs 7–8) are presented in Fig.3.

The time correction with regard to the differentiation of the sedimentary complex and crystalline crust structure beneath the source can be expressed in the same form as in eq. (2). The corresponding values of reference thicknesses of the sediments and whole crust were taken as 4 and 44 km, respectively. The mean velocities were calculated from the corresponding eqs (3), (7) and (8). The value of the emergence angle *e* was calculated for the average model MUMEP (Grad 1987, 1988). Finally, the arrival times were corrected for the thickness of the sediments and crust beneath the source, and for the crustal thickness beneath the seismic station.

2.2 *P*-wave traveltimes in sectors and 1-D models

The locations of nuclear explosions from the East European and Siberian platforms discussed in this paper are shown against a background of the main tectonic features of Eurasia in Fig. 1. Corrected for the crustal inhomogeneities, arrival times of *P* waves were divided into three data sets, corresponding to three sectors. Sector 1 contains the events from between 60°N and 70°N ('latitudinal' sector, average azimuth about 90°), sector 2 has an average azimuth about 130°, and sector 3 has an average azimuth about 160° (nearly 'longitudinal' sector). The large star in sector 2 represents about 200 events in the Semipalatinsk test site. Each event was recorded by several stations, and the fitting of these data by a linear function gave 'partial' traveltimes of about 200–700 km in length. These are shown in Fig. 4 for sectors 1, 2 and 3. The scattering of 'partial' traveltimes is not very substantial, especially for sectors 1 and 2. Also, their gradients are well ordered. This fact confirms our opinion of the high quality of seismograms recorded by the Finnish network. The differences in time can be accounted for primarily by inaccuracies in the origin time and location of the events. In further analysis we assumed that the quality of origin time and location would be better for bigger explosions because the event would be recorded by a larger number of seismic stations worldwide. These data were taken from international bulletins. We used polynomials up to second order to fit the middle points of the 'partial' traveltimes in the following distance ranges: below 2200 km, between 2200 and 3000 km, and over 3000 km. The data were weighted by the number of worldwide stations recording the particular event. In each case we have fitted polynomials using a simplex method for minimalization of χ^2. The fit was accepted as satisfactory when the error probability was 0.05. To avoid edge effects we tested the fitting for all the data sets, and for data excluding the last points. The results are presented in Fig. 5. The apparent velocities of the first arrivals of *P* waves are about 8.5, 10.0 and 12.5 km s^{-1} for successive intervals of distance. A comparison of traveltimes for latitudinal and nearly longitudinal directions shows a difference of the order of 5 s at a distance of about 4000 km, which reflects the mantle structure differentiation. The polynomial coefficient errors when propagated to the smoothed traveltime data give uncertainity bounds of ± 0.7 s, so we still have significant differences between sectors.

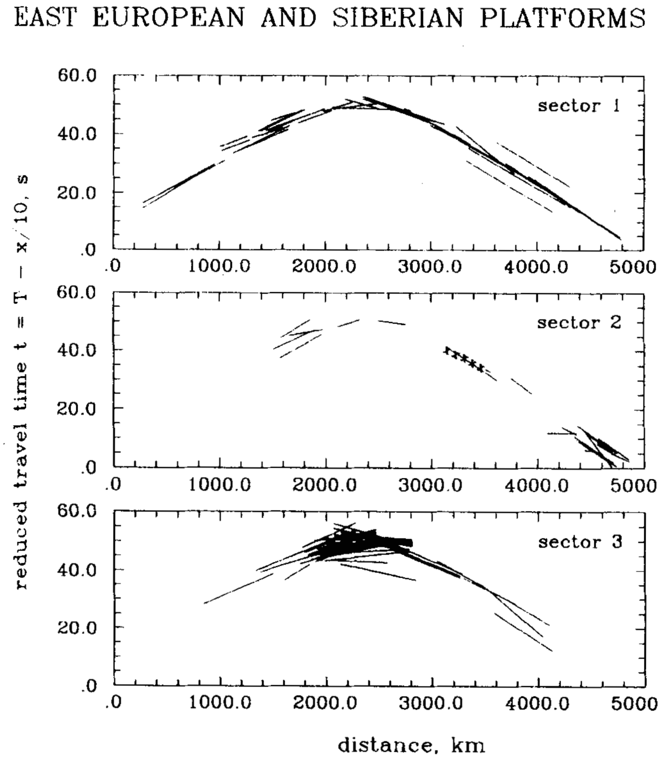

Figure 4. Traveltimes for the nuclear explosions recorded by Finnish seismological stations for three sectors. Each line represents the average traveltime for one event recorded by several stations. Crosses in sector 2—average traveltime for about 200 events in the Semipalatinsk test side.

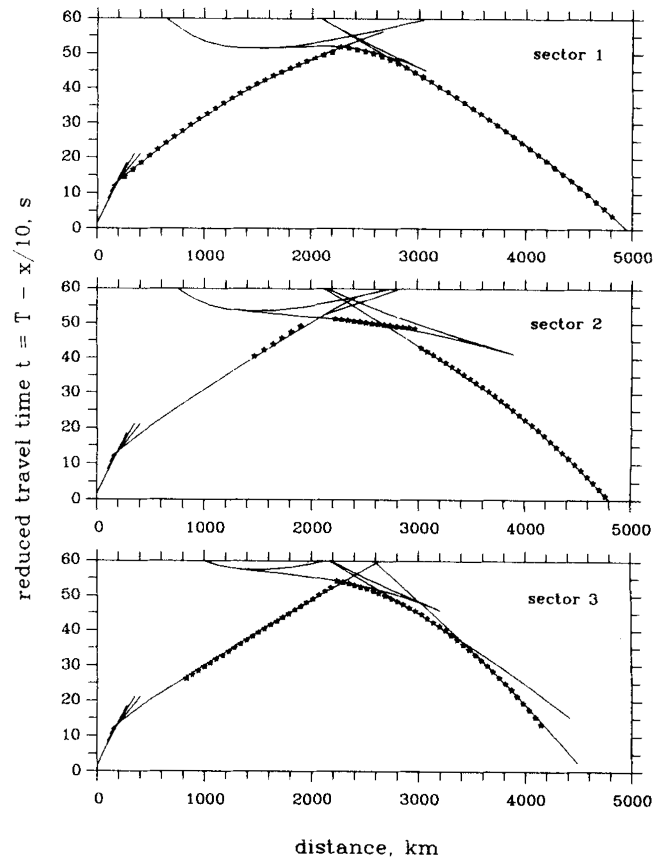

Figure 6. Comparison of average observed (stars) and theoretical (lines) traveltimes for 1-D models for three sectors. See text for details.

From the average traveltimes in the three sectors, the models of the *P*-wave velocity distribution with depth were found by 1-D modelling using a trial and error method. As a reference we took a crust model with 2 km of sediments (an average of 4 km was used for calculation of corrections beneath the source, and there were no sediments beneath the stations on the shield) and 44 km depth of the Moho. It should be noted that uncertainty bounds obtained from traveltime fitting (± 0.7 s) give an uncertainty of velocity lower than $\pm 0.03\ \mathrm{km\,s^{-1}}$. The results are shown in Figs 6 and 7.

2.3 *S*-wave traveltimes and 1-D models

Altogether, about 200 phases were interpreted as the first arrivals of *S* waves in the distance interval up to about

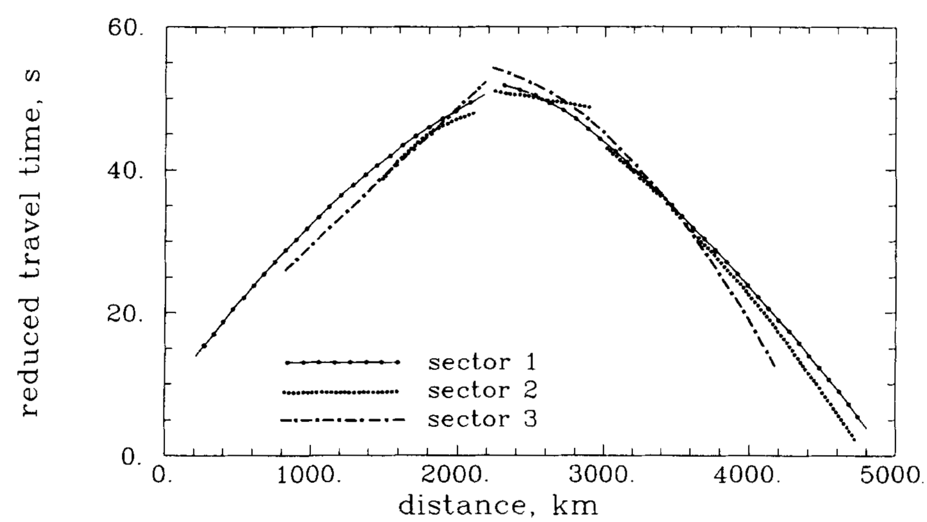

Figure 5. Average traveltimes for three sectors. See text for details.

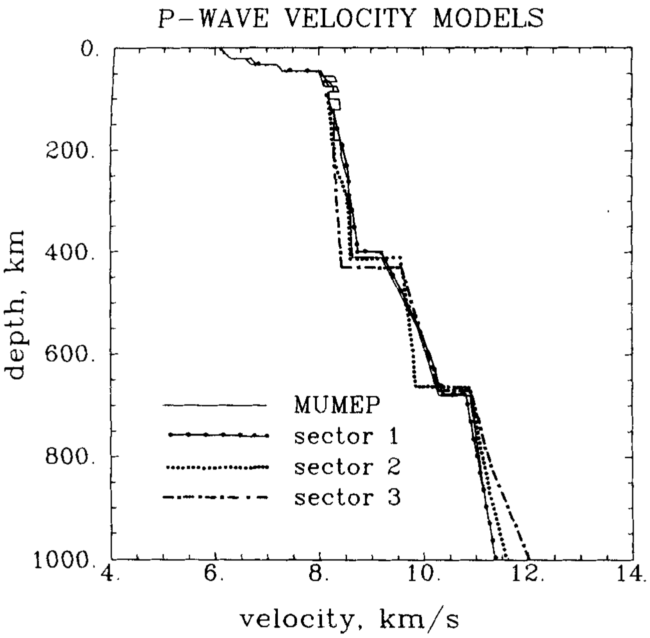

Figure 7. *P*-wave velocity distribution for three sectors of Eurasia. MUMEP model included for comparison (Grad 1988).

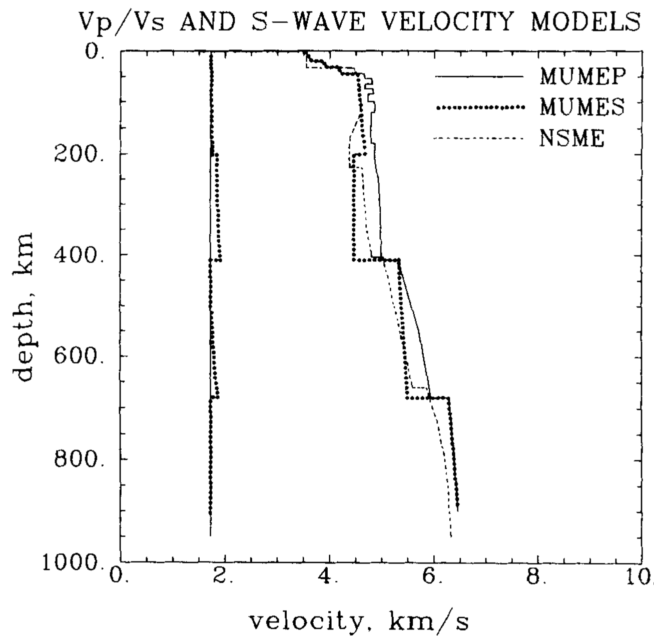

Figure 9. Distribution of the *S*-wave velocities and V_P/V_S ratios for Eurasia. The results are compared for the MUMES (dotted lines—this study), MUMEP (solid lines, assumed $V_p/V_s = 1.73$—Grad 1987, 1988) and NSME models (dashed line—Zielhuis *et al.* 1989).

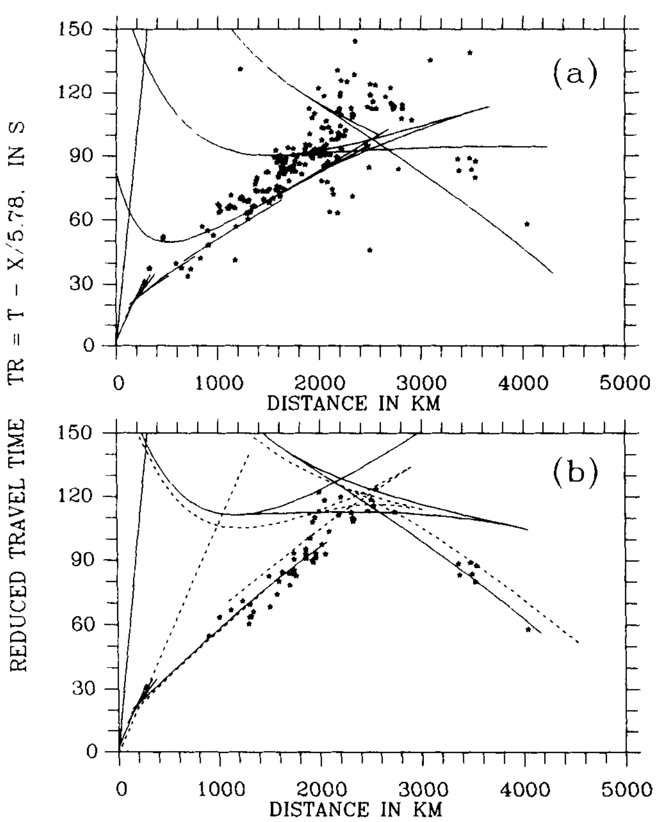

Figure 8. Comparison of observed (stars) and theoretical (lines) traveltimes of *S* waves. (a) All the data; theoretical traveltimes were calculated for the MUMEP model with $V_P/V_S = 1.73$. (b) Only *iS* phases (with exception of distances >3300 km, for which only phases *eS* were recorded); theoretical traveltimes were for the MUMES model (solid lines) and the NSME model (dashed lines). See text for details.

4000 km. The observed times for *S* waves are shown in Fig. 8(a), together with theoretical traveltimes for the average model MUMEP (Grad 1987, 1988), where the velocity ratio $V_P/V_S = 1.73$ was assumed. The scattering of the points is substantial, but a comparison of traveltimes indicates that in general V_S velocities in the model are too high. In further analysis, we used strong phases marked as *iS* (the exceptions are distances >3300 km, where only *eS* phases were recorded); however the number of data was insufficient for detailed analysis in sectors. For about 60 points the 1-D general model MUMES (model of the upper mantle for Eurasia, *S* waves) was found, and a comparison of observed and theoretical traveltimes is shown in Fig. 8(b). Traveltimes for model NSME are also shown here for comparison with the West Europe *S*-velocity structure. The velocity ratio V_P/V_S changes with depth, being 1.76–1.79 down to 200 km depth, about 1.87–1.92 in the depth interval 200 to 400 km, 1.72–1.87 in the depth interval 400 to 700 km, and about 1.73 for greater depths.

P-wave velocities in the three sectors and *S*-wave velocities in the MUMES model are compiled in Figs 7 and 9, and in Table 2.

3 DISCUSSION OF RESULTS AND CONCLUSIONS

Determination of the velocity distributions of seismic waves in the Eurasian mantle has been the subject of a large number of investigations. Both explosions and natural earthquakes recorded by networks of seismic stations and seismic arrays have been used (e.g. Burmakov *et al.* 1987; Calcagnile 1982; Egorkin *et al.* 1987; Enayatollah 1972a,b; England & Worthington 1977; England, Worthington & King 1977, 1978; Garnero, Helmberger & Burdick 1992; Given &

Table 2. *P*- and *S*-wave velocities of Eurasia.

sector 1		sector 2		sector 3		MUMES	
depth (km)	Vp (km/s)	depth (km)	Vp (km/s)	depth (km)	Vp (km/s)	depth (km)	V_S (km/s)
0	2.60	0	2.60	0	2.60	0	1.50
2	2.60	2	2.60	2	2.60	2	1.50
2	6.10	2	6.10	2	6.10	2	3.53
20	6.30	20	6.30	20	6.30	20	3.64
20	6.60	20	6.60	20	6.60	20	3.87
32	6.70	32	6.70	32	6.70	32	3.93
32	7.20	32	7.20	32	7.20	32	4.16
44	7.30	44	7.30	44	7.30	44	4.22
44	8.00	44	8.00	44	8.00	44	4.57
75	8.10	75	8.20	75	8.28	200	4.70
210	8.50	210	8.30	210	8.30	200	4.50
300	8.60	300	8.60	430	8.45	410	4.50
400	8.75	410	8.65	430	9.60	410	5.35
400	9.20	410	9.60	670	10.30	680	5.50
550	9.90	660	9.90	670	10.90	680	6.29
680	10.40	660	10.90	790	11.20	900	6.47
680	10.80	900	11.36	900	11.60		
900	11.18						

Helmberger 1980; Goldstein, Walter & Zandt 1992; Grad 1987, 1988; Hurtig, Grässl & Oesberg 1979; King & Calcagnile 1976; Masse & Alexander 1974; Patton 1980; Pavlenkova & Yegorkin 1983; Ryaboy 1977, 1990; Stewart 1981; Vinnik & Ryaboy 1981; Yegorkin & Chernyshev 1983; Yegorkin & Pavlenkova 1981; Zielhuis, Spakman & Nolet 1989). In this study we attempted to determine *P*- and *S*-wave velocity models from the recordings of nuclear explosions up to a distance of about 5000 km. The results are compiled in Figs 7 and 9 and in Table 2.

In all mantle models the boundaries '400' and '700' km were observed. The depth of the former increases from 400 km in sector 1 and 410 km in sector 2 to 420 km in sector 3. The depth of the second boundary decreases from 680 km in sector 1 to 670 km in sector 3. In the MUMEP model, corresponding depth values are 410 and 680 km. The results are also very close to model K8 for the upper mantle of north-western Eurasia (Given & Helmberger 1980). The average model for *S* waves (MUMES—model of the upper mantle for *S* waves) has these boundaries at depths of 410 and 680 km. In the NSME model (New S-Model for Europe, Zielhuis *et al.* 1989) the depths are 405 and 660 km. The *S*-wave velocity distributions are very similar, and in particular they show that V_S velocities are distinctly lower than velocities in the model MUMEP with $V_P/V_S = 1.73$ (especially in the depth interval 200–400 km). A low velocity of *S* waves in the upper mantle of the Eurasian continent was also obtained by Patton (1980) from the phase velocities of surface waves.

In the last few years the problems of the upper mantle structure on a global scale as well as upper mantle heterogeneity have been investigated (e.g. Woodhouse & Dziewonski 1984; Kennett & Bowman 1990; Kennett & Nolet 1990; Tanimoto 1990; Woodward & Masters 1991; Cummins *et al.* 1992; Pulliam & Johnson 1992). However, it is difficult to compare results that differ in resolution and usually represent averaged velocity over a range of several thousand kilometres. Nevertheless, some results do have similarities. For example, our *S*-wave velocities are low compared to the *P*-wave model of Eurasia, but they exceed by about 0.2 km s^{-1} the *S*-wave velocities in the PREM model (Dziewonski & Anderson 1981). A similar anomaly has been observed for the Eurasian continent in the M48C model (Woodhouse & Dziewonski 1984).

Finally, some conclusions about the accuracy of our results should be drawn. As mentioned earlier, our calculations of traveltimes were based on the original times and geographical coordinates of the explosions as determined and published by international bulletins. In 1989, Soviet seismologists published a description of 96 nuclear explosions conducted from 1961 to 1972 at the Semipalatinsk test site (Bocharov, Zelentsov & Mikhailov 1989; Vergino 1989a,b). These data have made it possible to calculate the precise distances and arrival times for these events. Recalculated traveltimes for 96 explosions in Semipalatinsk, together with Jeffreys–Bullen, IASPEI 1991 and MUMES traveltimes, are compiled in Fig. 10. Apart from about 2–3 s difference between Jeffreys–Bullen and IASPEI 1991 traveltimes, this comparison shows also that our traveltimes are on average about 2 s later than those determined for true data. We did not take this fact into account in our analysis in order to preserve a 'homogeneity' of error. We can expect that the relative errors might also be of the order of ±2–3 s for other events from Eurasia. Thus the knowledge of directly determined origin times and coordinates makes more precise interpretation possible, including the interpretation of other phases, as well as of the dynamic properties of recorded waves.

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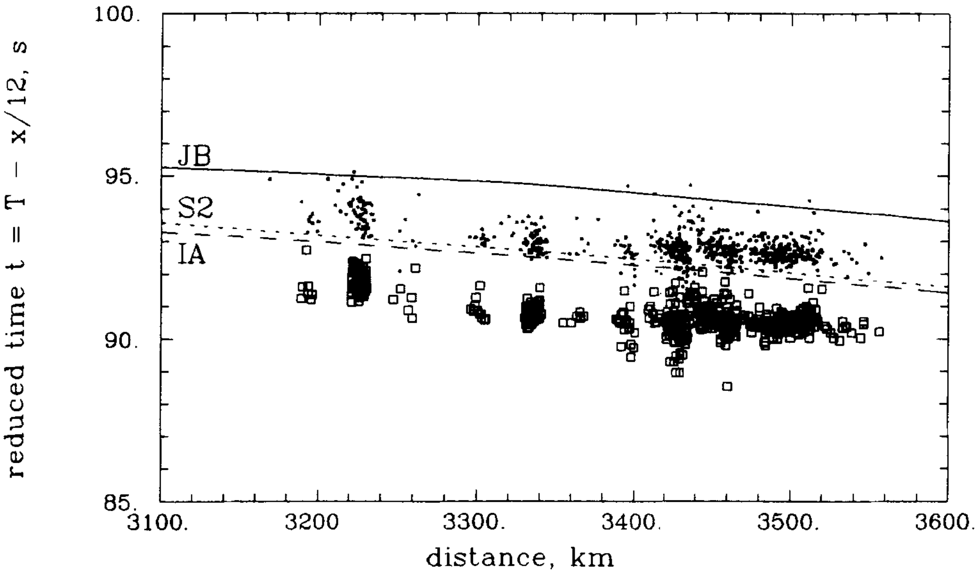

Figure 10. Traveltimes for explosions conducted at the Semipalatinsk test site and recorded by Finnish seismological stations: dots–traveltimes calculated according to BISC location and original time; squares–traveltimes according to original data published by Bocharov *et al.* (1989) and Vergino (1989a,b). For comparison, traveltimes of Jeffreys–Bullen (solid line), IASPEI 1991 (dashed line) and sector 2 (dotted line) are shown (Jeffreys & Bullen 1940; Kennett 1991).

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