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Lithospheric Structure and Teleseismic *P*-Wave Reflection Delays Under Fennoscandia and Siberia

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Abstract. Differential $(PP-P)$ travel-time residuals from published teleseismic data are used to locate anomalous velocity structure near reflection points in Fennoscandia and Siberia. Reflections up to 2 s early correspond to shield areas, such as the Siberian platform, while later reflections (slower velocity structure), with positive residuals, tend to occur under less stable areas, such as the West Siberian lowlands. Exceptionally late reflections (about 2 s positive) in part of the latter region may be related to major lithossphenic sutures. The reflection delays are generally consistent with trends observed elsewhere for the residuals of *P*-waves.

Key words: *PP* reflections – Lithospheric structure – Seismic velocity – Travel-time residuals

Introduction

The travel-time delays or residuals of telseismic *P*-wave reflections (*PP* waves) may be used to locate regions of anomalous velocity structure in the crust and upper mantle near the reflection point, approximately half-way between the source and receiver. The effects of structure near the source and receiver can be minimised by taking differential $(PP-P)$ residuals, where P is the direct (unreflected) compressional wave (Stewart 1976). Positive $(PP-P)$ residuals correspond to later reflections than would be expected from the assumed Earth model used in theoretical travel-time calculations. Late reflections could be due to slower than normal velocity structure near the reflection point. For example, a delay of one second in the *PP* reflection time would be caused by an increase in crustal thickness of about 10–15 km compared to average values elsewhere. The minimum lateral dimension of anomalies which can be resolved by the *PP* residuals is of the order of 100 km, which is comparable to the size of the first Fresnel diffraction zone for reflections at teleseismic distances.

Published data have been used to correlate areas giving abnormal $(PP-P)$ reflection delays with regional tectonic features for parts of the Arctic ocean and Greenland (Stewart in press 1980). The extension of the data set into northern Eurasia is considered here, and it is shown that trends in reflection residuals tend to be consistent with the known lithospheric structure and with *P*-wave residuals observed elsewhere.

Data Analysis

The *PP* and *P*-wave travel-time residuals are taken from the Bulletins of the International Seismological Centre (I.S.C.) from 1964 through 1976. Some of the problems associated with using this source of data and in analysing the reflection residuals have been discussed by Stewart (1976 in press 1980). Only phases identified by the I.S.C. as *PP* waves are used, to reduce the possibility of errors arising from incorrect identification of waves by the seismograph operators. The $(PP-P)$ residuals used in this study are restricted to the range -8 to $+9$ s, in order to exclude values which are likely to be due to incorrect phase identifications or excessive errors in reading the seismograms. Since genuine reflection delays may approach these values (Stewart and Keen 1978), minimal restrictions are placed on the data range used in the analysis.

The data are given here for two areas, from the Urals westwards, and for Siberia. Most of the data are from earthquakes in the vicinity of the north-western margin of the Pacific Ocean, recorded at stations in Europe. For the area west of the Urals, the values of the $(PP-P)$ residuals in seconds at the reflection points, averaged over regions 1/2° by 2/5° geocentric in area, are given in Fig. 1 Allowance is made for the focal depth in calculating the reflection points. In Fig. 1, letters correspond to negative residuals (early reflections), with A equivalent to -2–-1 s, and so on. The number of data used to give each value in Fig. 1 is shown in Fig. 2. The data in Fig. 1 and 2 comprise 971 reflections from 306 events. The data for Siberia are shown in Fig. 3, averaged as for Fig. 1, with the data density indicated in Fig. 4. The area of this data set partially overlaps that given in Fig. 1, and includes 4055 points from 656 events. Good data coverage is achieved around 90° E longitude, approximately half-way between most of the events along the north-west Pacific margin, and the European seismograph stations.

Although the data can be contoured to show the broad trends in values, this process tends to smooth out details of residual variations which may occur at tectonic boundaries. High-order contour surfaces can show such changes, but tend to be unstable in areas of lower data density, especially where the noise may be several seconds in amplitude. Following the procedure used by Stewart (in press 1980), the lateral variations are shown as averages along great circle profiles. Points within $2^1/_2$ degrees geocentric of a profile are projected perpendicularly on to the line, and a running mean within a window of length 3 geocentric degrees is then taken along the profile. Individual data points are used in this process, rather than the averaged values shown in Fig. 1 and 3.

The locations of the profiles across Fennoscandia and the Russian Platform are indicated in Fig. 5, which also shows the main tectonic and bathymetric features. The residual values along the profiles are given in Fig. 6. Values derived from averages of less

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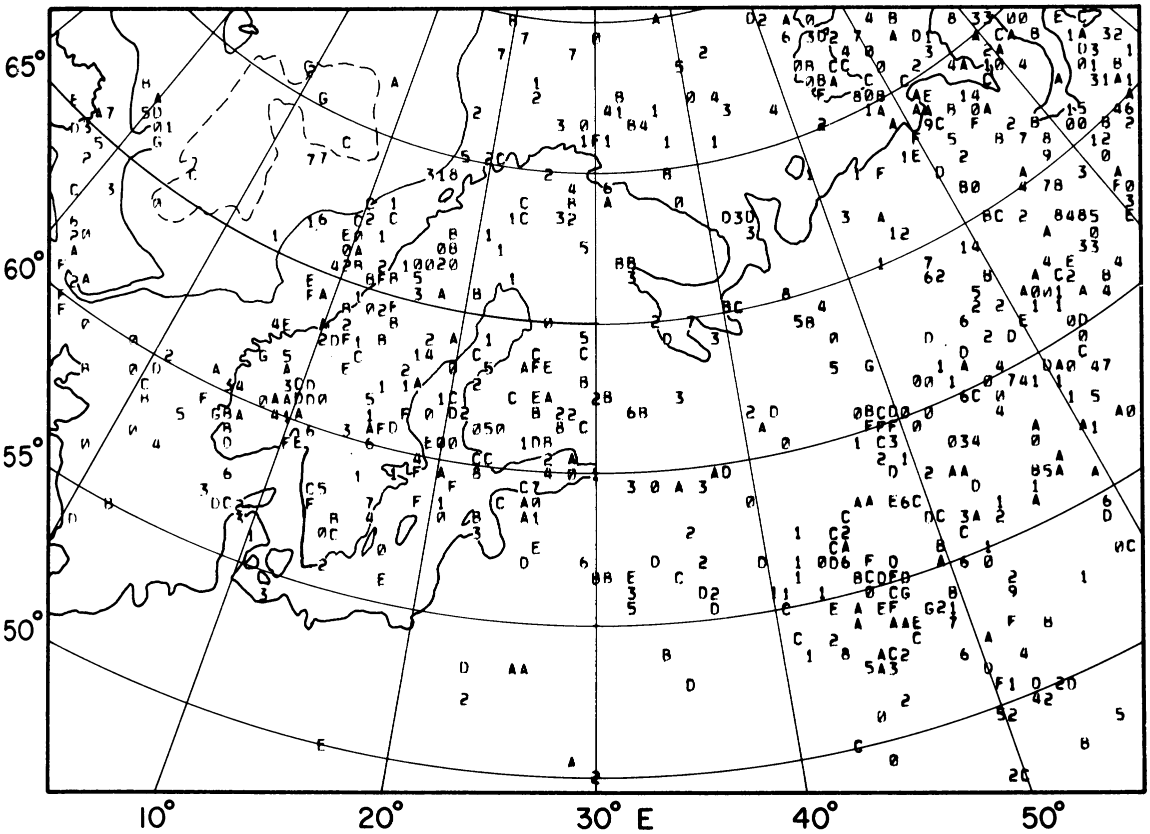

Fig. 1. ($PP-P$) travel-time residuals for reflection points in Fennoscandia and the Russian platform. The values in seconds are averages over areas 2/5° by 1/2° geocentric. Letters correspond to negative values, with *A* equivalent to -2–-1 s, *B* -3–-2 s, and so on

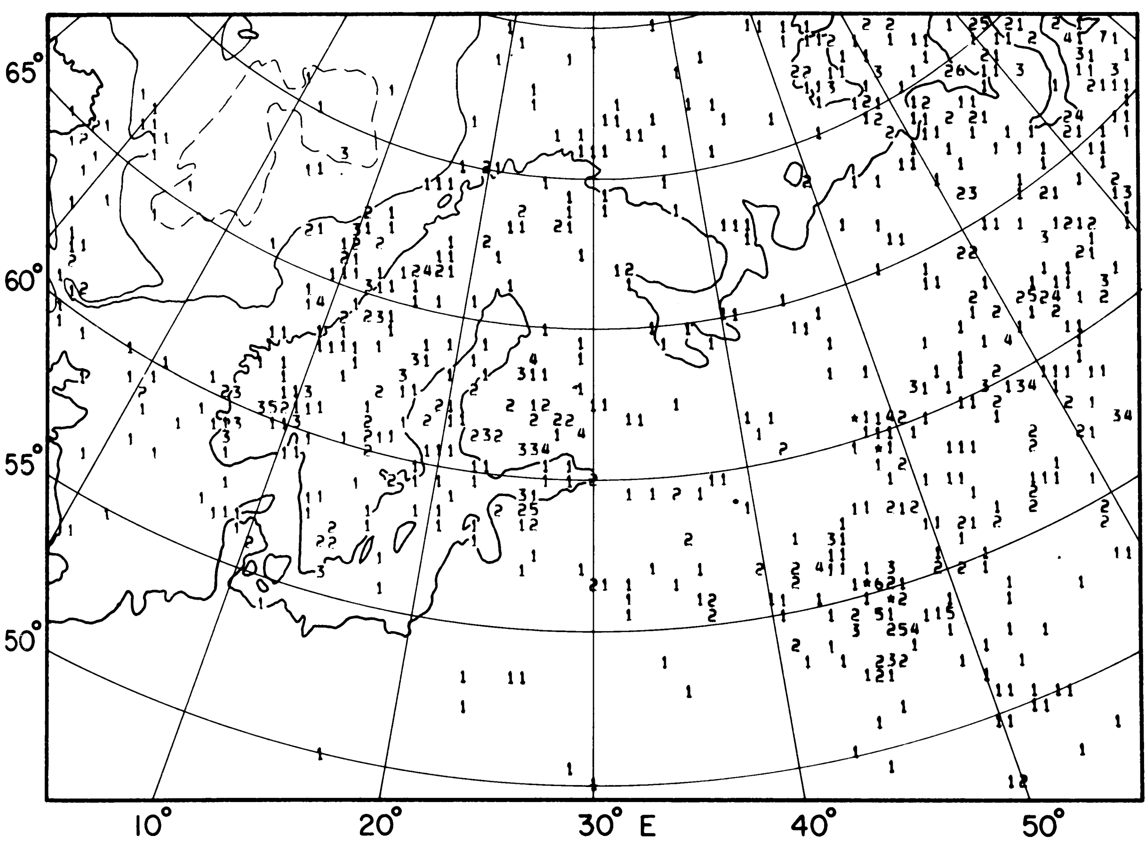

Fig. 2. The number of reflections averaged to give each data value in Fig. 1. Asterisks correspond to more than 9 points per unit area

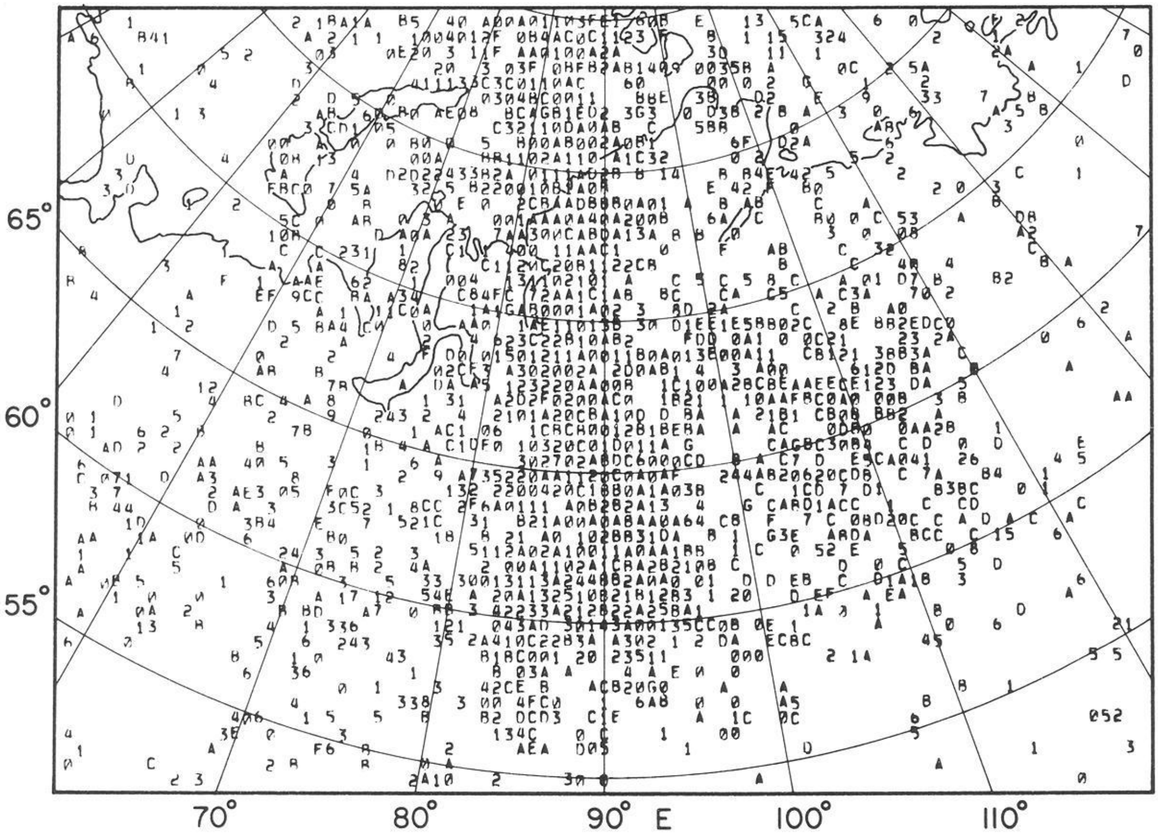

Fig. 3. Travel-time residuals in seconds for reflections under Siberia, plotted similarly to Fig. 1

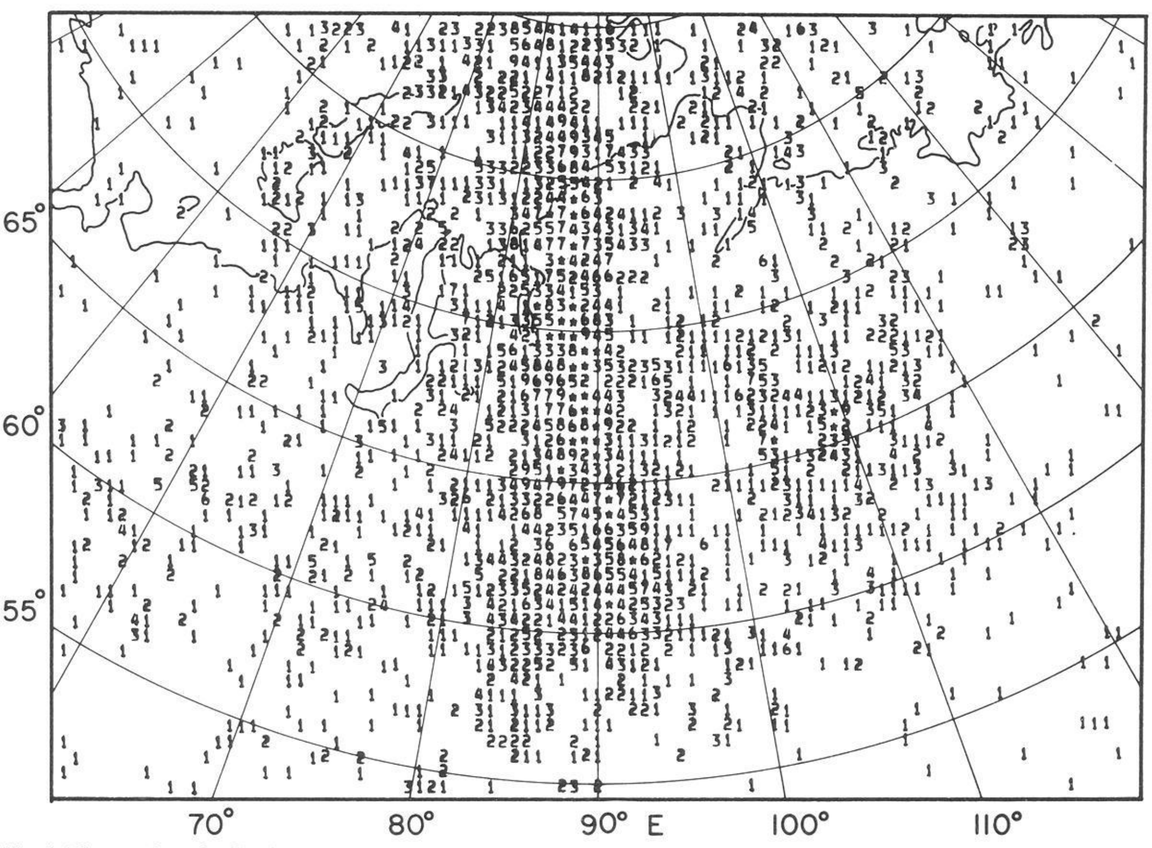

Fig. 4. The number of reflection residuals averaged to give each value in Fig. 3

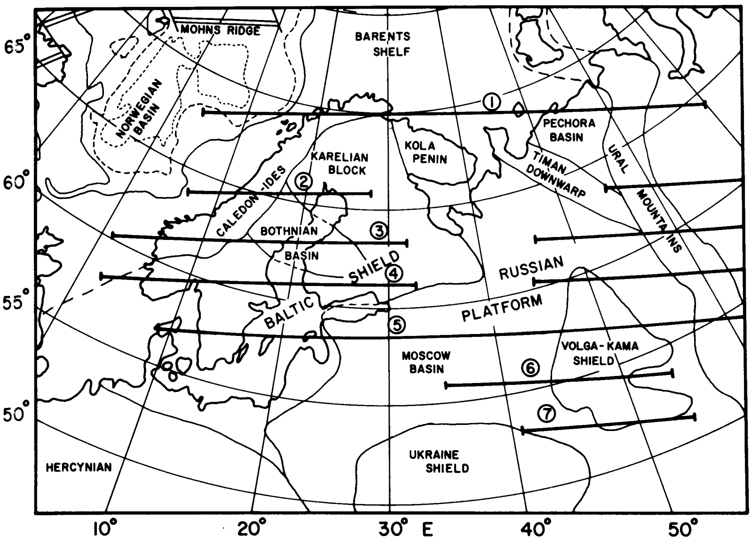

Fig. 5. The main tectonic features of the area covered by Fig. 1. The 1000, 2000, and 3000 m isobaths are also shown. The locations of the profiles along which reflection residuals are averaged in Fig. 6 are also indicated

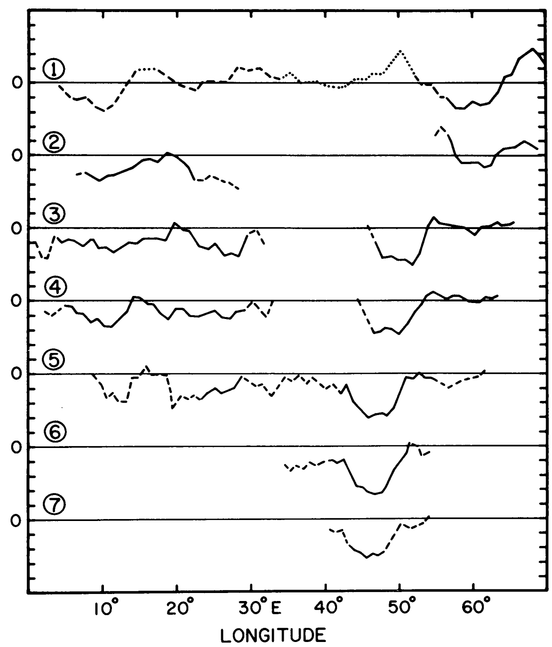

Fig. 6. Averages of $(PP-P)$ residuals along great circle profiles. Data within $2^1/_2°$ geoecentric of a profile are averaged within a running window of length 3° along the line. Averages of less than 10 and 20 data points are shown dotted and dashed respectively. The *profile numbers* correspond to those given in Fig. 5

than 10 data points are shown dotted (which only occurs on Profile 1 in Fig. 5), and averages of less than 20 points are dashed. For the regions of better data coverage in Fig. 5, up to about 50 points are averaged to give each profiled value.

The profile locations for the Siberian data are shown in Fig. 7, and the profiles are given in Fig. 8. A greater density of data occurs for parts of Siberia compared to the Fennoscandian data, and hence some sections of the profiles in Fig. 8 represent averages of more than 400 points.

In the areas which are well covered by reflection points, the effects of random errors should be minimised by averaging along the profiles. A given area of reflections may include data from a range of source-receiver geometries. Hence it is unlikely also that trends in abnormal reflection delays observed in the areas of better coverage are due entirely to any source or receiver effects remaining after taking differential $(PP-P)$ residuals. However, the profiles should be treated with caution in the areas of fewer data.

The standard deviations of the profile values tend to be in the range $2-3$ s. This is comparable to the standard deviations observed for residuals at single stations using data from restricted source regions of the order of 1°–2° across, and may be due largely to random errors in the data. The statistical significance of the variations along the profiles may be tested by using a two-tailed t-test to compare residual values in adjacent areas (Stewart in press 1980). The more prominent anomalies tend to be significant at least at the 95% confidence level. In the areas of better coverage, variations of only 1 s along the profiles are significant at better than the 99% level.

As discussed by Stewart (in press 1980), the features with shorter wavelengths (a few hundred kilometers) on the profiles are probably due to structure within the tectosphere. Variations in residuals with wavelengths of the order of 1,000 km or more are probably due largely to deeper lateral changes in velocity. Romanowicz (1979) suggests that about 70% of the variation in travel-time delays in North America is caused by lateral heterogeneity within 250 km of the surface, although lateral changes in velocity below Eurasia may be significant down to 700 km depth (England et al. 1977).

Profiles in Fennoscandia and the Russian Platform

Much of the Baltic Shield appears to give early reflections, with residuals up to about 2 s negative (Fig. 6). This is consistent with observations of direct P-wave arrivals, which tend to be up to 1 s early in shield areas (Herrin and Taggart 1968). The variations given here (of the order of 2 s) for reflection residuals in Fennoscandia are consistent with lateral changes of about 1 s in P-wave residuals given by Noponen (1977). Similar variations are also observed by Brown (1973) for P-wave residuals at distances less than 55°.

For the present data set, few reflections exist in the vicinity of any of the seismograph stations. Hence the P-wave delays at

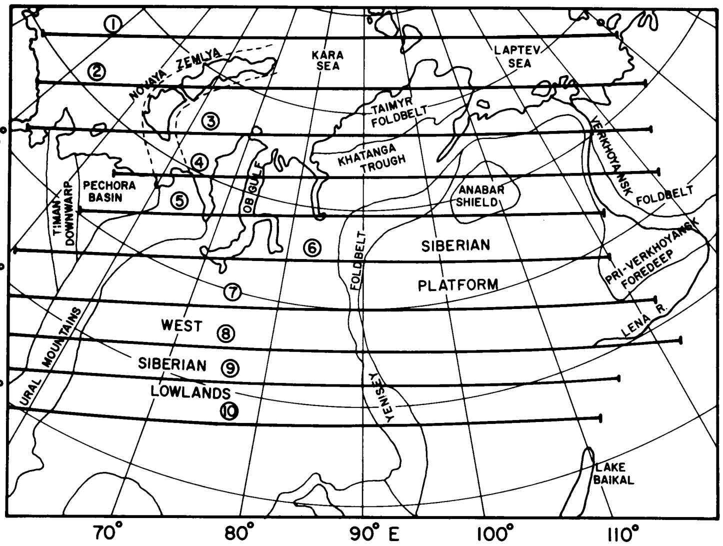

Fig. 7. The main tectonic features of Siberia and the locations of the profiles given in Fig. 8

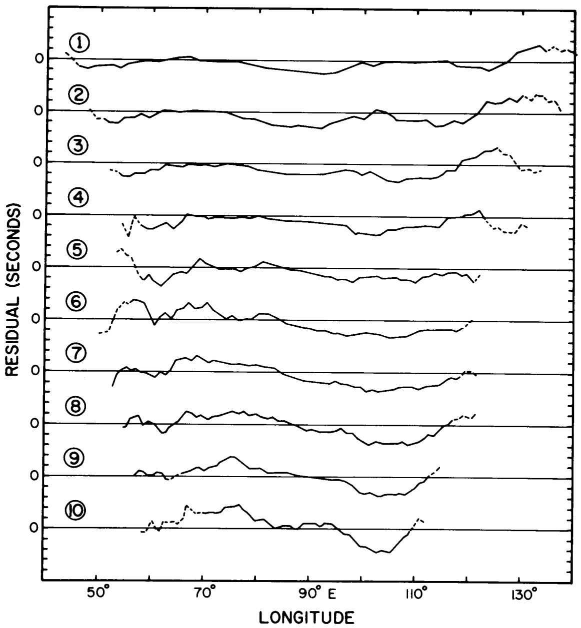

Fig. 8. Averages of $(PP-P)$ residuals along profiles across Siberia, obtained in a similar fashion to those in Fig. 6

individual stations cannot be compared readily with the $(PP-P)$ residuals, which are also subject to noise. It should also be noted that in general there is not good agreement between the various estimates of P-wave residuals at Scandinavian stations (Enayatollah 1972; Brown 1973; Noponen 1977; Poupinet 1979). Brown (1973) suggests that the P-wave residuals have large azimuthal components (up to several seconds in amplitude), which may be due to a large-scale dipping discontinuity in the mantle at about 600 km depth. The effects of a dipping interface should to some extent be cancelled out in the reflection residuals, if the boundary is traversed by both the upgoing and downgoing sections of the PP ray. The PP travel time may therefore yield the non-azimuthally dependent residual for the region.

The Bothnian basin tends to give reflection residuals later than in adjacent areas by about 1 s (Profiles 2 and 3 in Fig. 6). Local crustal thickening of the order of 5 km in the vicinity of the Gulf of Bothnia and central Sweden (Sellevoll 1973) would delay reflections by only about 0.2 s. However, in general the data density in Scandinavia is too low to permit a comparison of the reflection residuals with localised tectonic features. There appears to be no consistent correlation of trends in early reflections (high seismic velocities) with areas of low heat flow in Fennoscandia or the Russian platform (Čermák and Hurtig 1978).

The density of reflection data varies across the Russian platform, but values tend to be negative over much of the region. The seismograph stations in the U.S.S.R. for which P-wave residuals have been published (Herrin and Taggart 1968; Poupinet 1979) are not located in the areas of good $(PP-P)$ data coverage, and hence the residuals cannot be compared. Prominent negative reflection residuals (up to about 3 s early) coincide with part of the Volga-Kama shield. This may result from an abnormally thin low-velocity layer or a cooler than normal upper mantle, perhaps analogous to the possible "cold spot" in West Africa proposed by Chapman and Pollack (1974). Although the regional heat flow may not be exceptional (Kutas et al. 1978), it is possible that the upper mantle is relatively cool with a Curie isotherm depressed by up to 20 km in the region of the Volga-Kama shield (Belyaevsky et al. 1973). It is difficult to correlate the relatively few $(PP-P)$ residuals elsewhere in the Russian platform with major structures given by Nalivkin (1976) or with relief on the Moho discontinuity (Belyaevsky et al. 1973). Crustal thickness in the platform may vary by up to 20 km (Belyaevsky et al. 1973), which could contribute more than 1 s to the changes in residual values.

Siberian Profiles

Over much of its length, the Paleozoic geosyncline forming the Ural mountains is associated with reflection residuals which are of the order of 1 s negative towards the east, compared to values which are near zero or even positive towards the west (Fig. 8). While the Russian platform has an average crustal velocity of 6.5 km s^{-1} (Beliayevsky et al. 1968), velocities tend to be lower in the vinicinity of the Urals by up to 0.5 km s^{-1} for a given depth (Alekseev et al. 1973), and the crustal thickness is also increased by perhaps 5 km (Belyaevsky et al. 1973), Velocities of about 8.0 km s^{-1} below the Moho along the main axis of the Urals are about 0.2 km s^{-1} lower than in adjacent regions, and

Beliayevsky et al. (1968) indicate that velocities as low as 6.8 km s^{-1} occur under the area of greatest sedimentary thickness, which may reach 10 km. Thus the observed variations within the lithosphere could account for changes in reflection delays of the order of 2 s. Čermák and Hurtig (1978) also suggest that the temperature at the Moho under the Urals may be higher by 100°–200° C than in the platform regions to the west.

From seismic profiling, the boundary of the Urals to the east may continue under the West Siberian lowlands for up to 100 to 200 km (Beliayevsky et al. 1968), with velocities exceeding 7 km s^{-1} at depths of only a few kilometers. Relatively high velocity material, exceeding 6.4 km s^{-1}, comprises 80% of the crustal thickness of the eastern Urals (Belyacvskiy et al. 1971). The crustal variations may account for about 1 s in the negative residuals on the eastern margin of the Urals. The geology of the Urals has been interpreted in terms of a former subduction zone dipping to the east (Hamilton 1970). While high velocities can be observed in dipping slabs which have become inactive only in the past few million years (Solomon and Butler 1974), it is unlikely that any velocity contrast extending below the lithosphere would remain from the Urals subduction (Toksöz et al. 1971), especially if the lithosphere is decoupled from the upper mantle.

Much of the West Siberian lowlands appears to give late reflections (positive residuals), with localised anomalies up to about 2 s in magnitude (Profiles 6–10 in Fig. 8). If the profiles are obtained by averaging over smaller areas, then the larger anomalies can even exceed 3 s. The positive residuals generally coincide with relatively low crustal and upper mantle velocities. Average crustal velocities in the West Siberian lowlands are 5.8–6.0 km s^{-1}, which is about 0.5 km s^{-1} lower than in the adjacent platform areas, but velocities tend to increase towards the eastern margin of the lowlands (Belyaevskiy et al. 1971).

There appears to be no consistent correlation of the residuals with crustal thickness (Beliayevsky et al. 1968; Vashchilov 1971). Changes in crustal parameters alone are unlikely to account for the larger anomalies exceeding 1 s in Fig. 8. A velocity decrease of several percent would have to extend vertically over at least 100 km to give the observed late reflections. The later reflection anomalies are comparable in amplitude to those expected from observations of *P*-wave residuals in active rift zones, which are of the order of 1 s (Fairhead and Reeves 1977), where reduced upper mantle velocities and a thinner than normal lithosphere usually prevail. The later reflections around 70° E longitude may lie in the vicinity of a Triassic boundary underlying the lowlands, between the Siberian and Northern European lithospheric blocks (Burrett 1974). The zone of very late reflections is approximately bounded on the east by the major Paleozoic suture extending south from the Ob gulf, postulated by Burke et al. (1977). The region of the late reflections also underwent widespread volcanism related to the Urals subduction during the Paleozoic (Hamilton 1970).

The physical properties of the lithosphere appear to vary appreciably across the lowlands, in both the crustal and sub-crustal layering (Vashchilov 1971). To the south of the Ob gulf, the depth to the top of the asthenosphere may be reduced, which could imply that the isotherms are elevated in the area. In the region of the late reflections, a local maximum occurs in the heat flow, with values of 60–70 mW m^{-2} (Kutas et al. 1978). For the eastern part of the West Siberian lowlands, the heat flow values are typically less than 50 mW m^{-2} (Jessop et al. 1976). Pollack and Chapman (1977a, b) predict a local minimum in the mantle heat flow under the lowlands, with an abnormally great lithospheric thickness of about 300 km. However, in the region of the late reflections around 60° N, the temperature at the Moho may be elevated by perhaps 100°–200° C compared to adjacent areas (Fotiadi et al. 1969). Gravity observations are not available for the area for comparison with the reflection data.

Much of the Siberian platform is well covered by the present data set. The platform is separated from the West Siberian lowlands by the Yenisey foldbelt, which was formed by east-dipping subduction during the late Precambrian (Hamilton 1970). Early reflections tend to occur under this foldbelt, and the residuals become more negative towards the interior of the Siberian platform. This is consistent with observations of *P*-wave residuals in shield areas (Poupinet 1979). The residuals in the platform appear to have no correlation with the thickness of the sedimentary sequences, which may exceed 6 km in some parts, nor with the depth to the Moho, which only varies by a few kilometers over most of the region (Fotiadi et al. 1978).

Negative residuals occur at the western end of the Paleozoic Taimyr foldbelt, and become more positive towards the east, where the crust thickens by several kilometers (Fotiadi et al. 1978). These geosynclines continue offshore into the southern Kara sea (Hamilton 1970). Reflections along the Khatanga trough (North Siberian lowland) vary progressively from about zero in the west to around 2 s negative in the east (Profiles 3 and 4 in Fig. 8). This Mesozoic trough may have been caused by regional tension (Hamilton 1970), with more than 3 km of subsidence (Bazanov et al. 1976) and a crust which may be thinner than adjacent areas by more than 6 km (Fotiadi et al. 1978).

On the eastern continuation of the trough, approximately 20 late reflections occur, which may represent a significant local trend possibly several hundred kilometers in extent near the Lena river delta (Profiles 1, 2 and 3 in Fig. 8). If this is a real anomaly, then the residuals could indicate the location of former hot spot activity (Stewart and Keen 1978). The Arctic mid-ocean spreading center (Nansen or Gakkel Ridge) probably continues across the Laptev sea, through the region of these late reflections, and into a broad zone of seismicity which extends eastwards from the Verkhoyansk foldbelt (Chapman and Solomon 1976). The Khatanga trough may then be an aulacogen, with the Mesozoic Verkhoyansk foldbelt and the spreading center as other tectonic features radiating from a former hot spot. This would be analogous to triple junctions centered on hot spots in other locations (Burke and Dewey 1973).

Poupinet (1979) notes that *P*-wave residuals at Tiksi (71.6° N, 128.9° E), near the Lena delta, tend to be exceptionally early (−1.5 s). Unfortunately no *PP* data are plotted in Fig. 3 within at least 100 km of this seismograph station. The late reflections giving the positive trends on Profiles 1 to 3 in Fig. 8 tend to be to the north and west of Tiksi, while the few reflection values to the south tend to be negative. A hot spot centered possibly several hundred kilometers from Tiksi would not affect most of the *P*-wave residuals observed at that station. Hence the reflection data are not necessarily inconsistent with this single published *P*-wave residual for the region.

Conclusion

The broad variations in the $(PP-P)$ travel-time residuals can be correlated with the major tectonic features in the areas of better data coverage, especially Siberia. As expected from directly-observed *P*-wave residuals, shield areas such as the Siberian platform tend to give reflections early by about 1–2 s compared to less stable regions, such as the West Siberian lowlands. The larger residuals, which may be several seconds positive or negative (late or early), probably indicate changes in structure throughout the lithosphere, and possibly also in the thickness of the asthenosphere. An extensive area of late reflections occurs in the central sections of the West Siberian lowlands, and locally the anomalies become about 2 s positive. The region may have been the location of a major lithospheric suture. The variations observed across the Paleozoic Ural geosyncline, with residuals changing from positive in the west to about 1 s negative towards the east, are probably consistent with the known velocity structure of the lithosphere.

Where suitable source-receiver configurations exist, the reflection residuals enable large areas to be examined more readily than is possible with directly observed *P*-wave residuals, particularly in remote regions. While local anomalies of small extent (less than 100 km across) cannot be observed well by teleseismic *PP* waves, larger-scale lateral variations in whole lithospheric and upper mantle structure can be detected.

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