Seismological Evidence for Heterogeneity of the Mantle

Hiroo KANAMORI

Earthquake Research Institute, University of Tokyo

(Received January 12, 1970)

Abstract

The latest results concerning the depth variation of Q, P- and S-wave velocities obtained by various seismological methods are reviewed with special emphasis on the regional variation. The body-wave travel time and the slope, $dT/d\Delta$, have been inverted to obtain P- and S-wave velocity distributions within the mantle. The P velocity is now believed to be known with an accuracy of $\pm$ 1% in the mantle below 1000 km. In the upper mantle, a considerable difference has been found among different regions. The long-period surface waves have been used to reveal the regionality of the upper mantle. Based upon the surface wave and body-wave results, it is now agreed that the mantle beneath shields has much higher seismic velocities than those beneath ocean and tectonically active regions. At such tectonically active regions as Japan, basin and range province of the western U.S., and the Pacific ocean side of South America, the upper mantle velocities have been found to be significantly lower than the average. Teleseismic P waves are delayed by 1.0 to 1.5 sec at tectonically active regions as compared with shields. The existence of the low-velocity zone for S waves in the sub-oceanic mantle is almost certain. The shape of the low-velocity zone varies markedly beneath continents. The distribution of Q with depth is not known in detail. However, it is almost certain that the value of Q is higher by more than one order of magnitude in the mantle below 1000 km than in the mantle above. The average values of Q_α (Q for P-waves) and Q_β (Q for S-waves) for the upper mantle range from 100 to 300. Horizontal heterogeneities as large as a few per cent in velocity and one order of magnitude in Q have been found beneath mid-oceanic ridges and island arcs. However, the uniqueness of the structure derived is still questionable.

Introduction

The determination of the earth's internal structure is important not only for its own sake but also for the interpretation of various geophysical data such as heat flow, conductivity anomaly, seismicity etc. During the last decade, as the interest of earth scientists was directed toward unusual regions on the earth, the regionality of the crust and the upper mantle structure became an increasingly important subject of study. This paper describes the current state of our knowledge concerning the earth's structure in special reference to the regional variation of the mantle. It is intended here to summarize the latest results for those who do not primarily specialize in seismology rather than to make an extensive review; for more general discussions on the related subjects reference can be made, for example, to *Anderson* (1965, 1967), *Press* (1964, 1966*a*), *Kovach* (1965), *Nuttli* (1963), *Bolt* (1963), *Savarenskii* (1966), and *Toksöz et al.* (1967).

1. Body-wave studies

The travel-time tables by *Jeffreys and Bullen* (1958), and the associated structure (*e.g. Bullen*, 1963) have been frequently used in many studies. This structure was derived on the basis of many readings of natural earthquakes on the assumption of spherical symmetry. An extensive use of statistical technique was incorporated. Obviously, it was intended to construct a most objective model which is consistent, on the average, with the existing data, rather than to determine fine structures. This fact should be borne in mind whenever the Jeffreys model is considered.

The earth model derived by Gutenberg (see *e.g. Press*, 1966*b*) is also often referred to in literature. This model, too, was derived mostly from travel times of natural earthquakes, but it differs significantly from the Jeffreys model in the upper 500 km of the mantle; the Gutenberg model has a velocity reversal centered at 150 km depth. This difference is perhaps due to the difference in approach used by these two investigators (*e.g. Jeffreys*, 1966), but it may also be partly due to the difference of the region studied. It should be remarked that there are several versions of the Gutenberg model and some confusion has been caused.

Recently *Herrin et al.* (1968) published a new travel-time table for P waves derived from more accurate data for earthquakes as well as for large explosions. The structure inverted from this travel-time curve (*Herrin et al.*, 1968) is very close to the Jeffreys model below 1000km.

These three studies belong more or less to the same category in that they all treat the data "statistically" thereby averaging out possible complicated fine structures. In Gutenberg model, however, the individual nature of the seismograms was also investigated for both amplitude and frequency; this led to the inclusion of the low-velocity layer in the upper mantle. Without low-velocity layers the average travel-time curves can be inverted in a straightforward manner to obtain a structure using the well-known Herglotz-Wiechert technique. However, since, in the averaging process, possible multiplications of travel times are smoothed out, the derived structure should be regarded as a smoothed model approximating the actual structure.

On the other hand, the recent $dT/d\Delta$ (slope of travel-time curve) measurement using seismograph arrays can be regarded as a fine-structure oriented method. By measuring travel times across a seismic array with an aperture of 100 to 200 km, $dT/d\Delta$, the derivative of travel time with respect to distance can be directly determined with an accuracy sufficient for investigating detailed fine structures. From the coherence of the wave train across the array, the $dT/d\Delta$ can be determined not only for the first arrivals but also for later arrivals; this resolves the multiplicity of the travel times thereby revealing fine structures at depths. The application of this technique to the entire mantle has been made by *Johnson* (1969). The earth model derived by Johnson naturally shows many features in the upper mantle but his velocity structure for the lower mantle is very close to that of the Jeffreys model. Figure 1 shows a comparison of the four models derived from different sets of data, upon different assumptions, and by different techniques. From these

results we can conclude that the P velocity below 1000 km of the mantle is now known to $\pm$ 1% and the regional variation of the structure exists primarily in the upper 1000 km. The regionality of the lower mantle has been suggested by *Chinnery and Toksöz* (1967), and *Hales et al.* (1968), but it is of much smaller magnitude than that for the upper mantle.

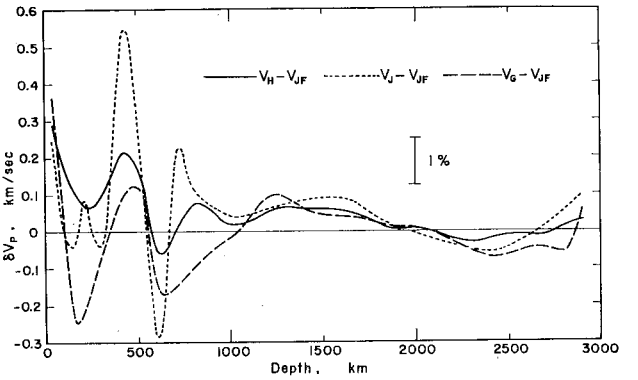

Fig. 1. P wave velocity structures of *Herrin et al.* (1968), *Johnson* (1969), and Gutenberg (see *Press*, 1966*b*) models with respect to Jeffreys model (see *Bullen*, 1963). The curves designated as $V_H - V_{JF}$, $V_J - V_{JF}$, and $V_G - V_{JF}$ show the difference between Herrin *et al.* and Jeffreys, Johnson and Jeffreys, and Gutenberg and Jeffreys models respectively.

It is to be noted here that these structures have been derived primarily from the data obtained at continental stations for earthquakes and explosions on continents and continental margins. Therefore the upper mantle sampled in these studies represents the mantle beneath continents. Thus, our knowledge of the P velocity distribution in the upper mantle beneath ocean is much less precise than that beneath continents. The P velocity distribution beneath ocean can, in principle, be derived from surface wave dispersions along oceanic paths, but it is well-known that surface wave dispersion is not very sensitive to P velocity distribution.

The regionality of the upper mantle structure appears to be demonstrated to a certain extent in Figure 1, but it can be more clearly demonstrated in terms of travel times and $dT/d\Delta$ measured for various regions over a distance range of 2° to 30°.

The first indication can be provided by the variation of P_n velocities. The definition of P_n velocity found in literature is not very clear. In short-range refraction studies, the P_n velocity usually indicates the sub-Moho velocity involving the depth range of several tens of kilometers. The definition which is more appropriate for the discussion of the world-wide regionality would be the average slope of the travel-time curve to a distance of about 10°. The latter definition rests on the fact that the observed travel-time curve is often nearly straight over the range $2° < \Delta < 10°$. This part of the travel-time curve is certainly affected by many unknown features such as dipping interfaces, velocity reversals, and lateral irregularities. Nevertheless, the smoothed slope over this range will represent the overall velocity of the mantle to the depth of about 200 km which may directly reflect the tectonic character of the upper mantle.

The distribution of P_n velocity is best known in the United States (*Herrin et al.*, 1968) (Figure 2). The P_n velocities at continental margins may be somewhat disturbed by oceanic structures. It is seen that the P_n velocity is larger in areas of Precambrian and Paleozoic folding than in those of younger orogenic regions such as the basin and range province (Figure 2).

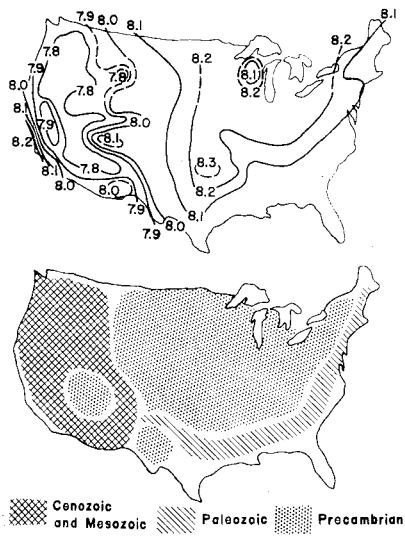

Fig. 2. Distribution of P_n velocity (*Herrin et al.*, 1968) (top) and chronological classification of crust (after *Umbgrove*, 1947) (bottom).

The P_n velocities for other regions are summarized by *Nuttli* (1963). It is found that P_n velocity is as low as 7.8 km/sec for such tectonically active regions as Japan, and is as high as 8.2 to 8.3 km/sec for Precambrian shields such as Canadian shield, Transvaar of S. Africa and a part of Australia. In Europe and central Asia, the P_n velocity is intermediate, from 8.1 to 8.2 km/sec. The range of the variation for various regions amounts to 0.4 to 0.5 km/sec which is comparable to that found within the United States.

We will next consider the difference of the structure beneath several distinct provinces. Figure 3 shows several vertical profiles of P velocity down to 450 km obtained by some of the recent studies. The major feature which characterizes the structure of different tectonic provinces is the velocity distribution in the upper 200 km of the mantle. In tectonically active regions such as Japan and the western United States(basin and range province) the velocity over this depth range is significantly lower than 8.0 km/sec. Whether the P velocity distribution really forms a velocity reversal or not is still a matter of considerable dispute. However, what is more important here is the evidence for gross low-velocity, rather than the velocity reversal. The mantle beneath relatively quiescent regions such as the central Asia and the central US is characterized by relatively high velocity, significantly higher than 8.0 km/sec. The intermediate velocity structure is found for the region of intermediate tectonic activity such as the western-central United States.

The structures shown in Figure 3 probably represent the structures beneath most of typical tectonic provinces beneath continents. Unfortunately, no detailed P wave structure beneath ocean is known.

The difference of the structures as shown in Figure 3 is reflected in observed travel-time residuals. The travel-time residuals can be regarded as indicating the anomaly of travel time averaged over a nearly vertical path beneath the individual station. Figure 4 shows the travel-time residuals from a certain "average" travel time observed at stations in the United States (*Herrin et al.*, 1968). It ranges from 0.75 sec (late) in the basin and range provinces to -0.5 sec (early) in the central United States. A similar comparison can be made at stations over the world. Figures 5 and 6 show the travel-time residuals from the J-B table observed for the large explosions on the Aleutian Island (LONGSHOT) and in Nevada respectively. The stations are classified according to the age of the crust on which they are located. After removing the general trend we find that the arrivals at stations on shields are, on the average, earlier by 1.5 sec than those at stations on orogenic

belts. This difference, 1.5 sec, is comparable to that observed within the United States, and may be compared to the difference of the structure among various tectonic provinces (Figure 3).

Similar studies of S waves are much more limited than those of P waves. This is due to the well-known difficulty in S wave identification, the inappropriate response of widely used seismometers to S waves, and the poor generation of S waves by large explosions.

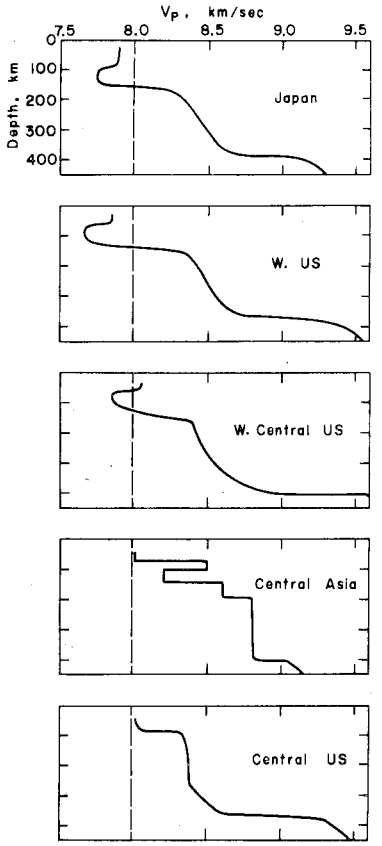

Fig. 3. Upper mantle P velocity structures for various tectonic provinces. The vertical dashed lines at 8.0 km/sec are drawn for reference. The tectonic character generally becomes from "active" to "quiescent" from the top to the bottom. The references are: Japan (*Kanamori*, 1967*a*), Western United States (*Johnson*, 1967), Western-Central United States (*Archambeau et al.*, 1969), Central Asia (*Lukk and Nersesov*, 1965), Central United States (*Green and Hales*, 1968).

The travel-time anomaly at worldwide stations has not been studied but the fragmentary data for S_n velocity compiled by *Nuttli* (1963) indicate a significant regionality; the S_n velocity is 4.35 to 4.55 km/sec for California, 4.63 to 4.70 km/sec for N.E. United States, around 4.7 km/sec for the central U.S., 4.72 km/sec for Canadian Shield, 4.576 km/sec for Europe, 4.608 km/sec for central Asia, and 4.7 to 4.75 km/sec for Australia. It is quite likely that the regional variation of S velocity structure is about the same order as, or even more pronounced than, that of P velocity structure.

The regional variation of S velocity distribution in the mantle is probably determined better by using long-period surface waves rather than body waves.

2. Surface wave studies

The surface wave in a bounded heterogeneous medium is dispersive; the phase velocity is a function of period and depends on the structure of the medium. Therefore, by measuring the velocity of surface waves propagated along the earth's surface

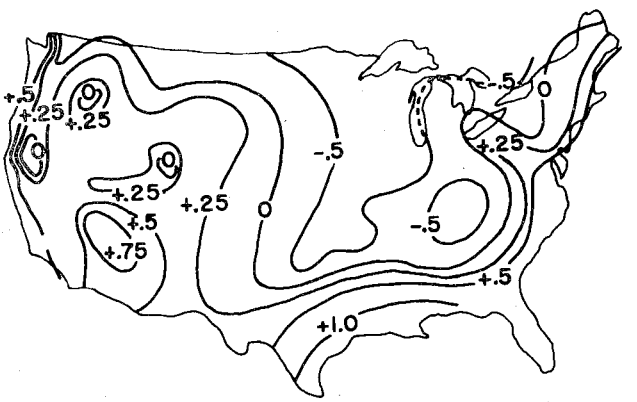

Fig. 4. Travel-time residuals in the United States (*Herrin et al.*, 1968).

as a function of period, we can determine the internal structure of the earth, although not uniquely. In order to determine mantle structures to the depth of several hundreds of kilometers, surface waves of comparable, or longer, wavelength should be used. The use of such long wavelength, *i.e.* long-period, surface waves was initiated by *Ewing and Press* (1954), and a large amount of data has accumulated since.

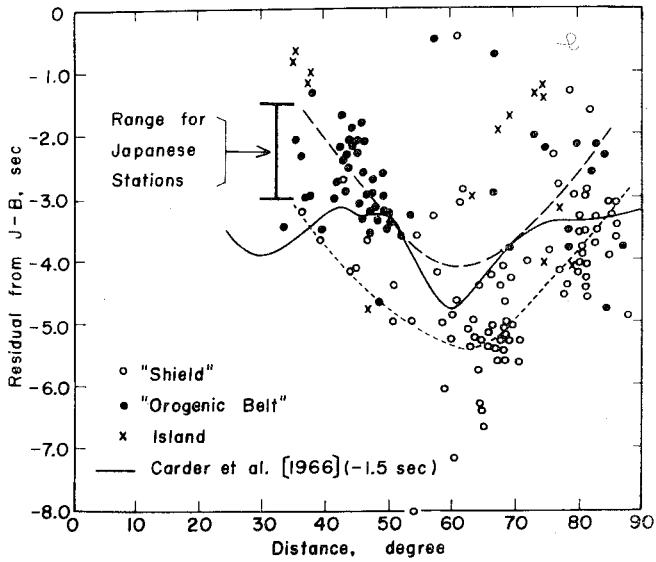

Fig. 5. Travel-time residuals for the LONGSHOT from the Jeffreys-Bullen table. Stations are classified according to the chronological difference of the crust under the station. "Shield" includes Precambrian shields and Paleozoic folded belts. "Orogenic Belts" include folded belts of Mesozoic and later times. The classification is according to *Umbgrove* (1947). The crosses are for mid-oceanic islands. The dashed and dotted curves are approximate averages for "Orogenic Belt" and "Shield". The solid curve is the average travel-time residual given by *Carder et al.* (1966) displaced by −1.5 sec. Range for Japanese stations is included.

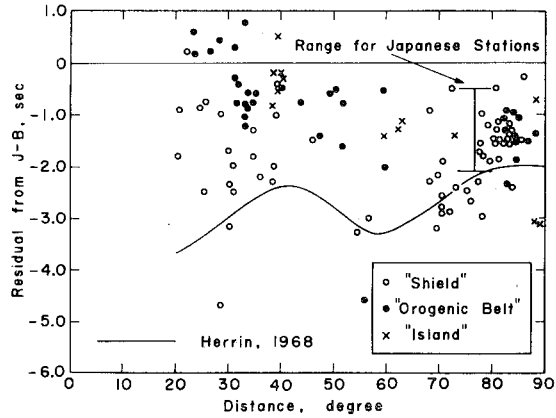

Fig. 6. Travel-time residuals for Nevada explosion of April 26, 1968. The solid curve is the travel-time curve by *Herrin et al.* (1968).

In surface wave studies, the basic data are phase velocities and group velocities as functions of period. The phase velocities can be measured, in principle, by determining the arrival times of a certain phase of a Fourier component of a wave train at two stations located on a great circle path. When the two stations are not exactly on the same great circle path, another station is necessary to determine the direction of wave incidence. On the other hand, in case of major earthquakes ($M \geq 7.5$), a single station often suffices to determine the average phase velocity along a great circle path. In this case, surface waves circuiting several times around the earth can be observed at a single station; hence by comparing the arrival time of a certain phase at a passage with that at another

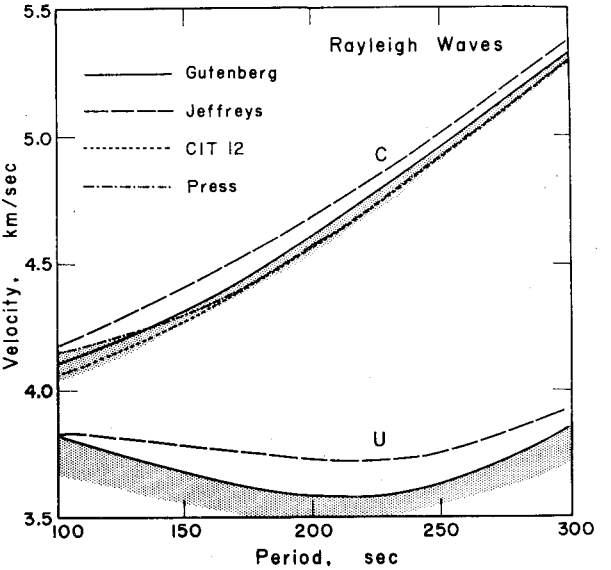

Fig. 7. Experimental Rayleigh wave phase (C) and group (U) velocity dispersion data (dotted zone) compared with theoretical curves for Gutenberg, Jeffreys, CIT 12, and Press models. For detailed models and dispersion data, see *Kanamori and Abe* (1968) for Gutenberg and Jeffreys models, *Toksöz and Anderson* (1966) for CIT 12 model, and *Press* (1970) for Press model.

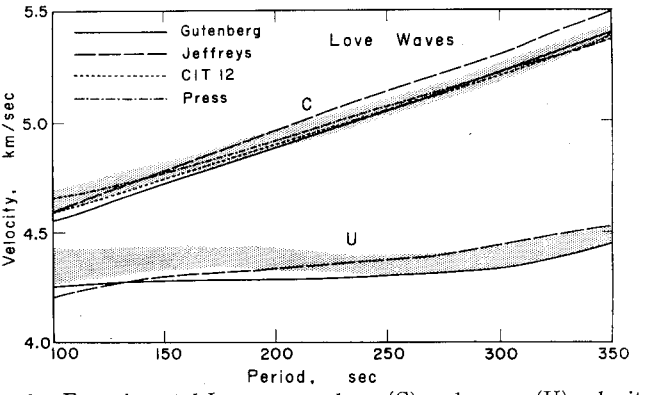

Fig. 8. Experimental Love wave phase (C) and group (U) velocity dispersion data (dotted zone) compared with theoretical curves for Gutenberg, Jeffreys, CIT 12, and Press models.

passage after one or more circuits, the phase velocity can be determined. The observation of the free oscillations of the earth also provides long-period phase velocity data.

The regional variation of the phase velocity can be investigated either by using the two-station method for various tectonic provinces or by using the single-station method for various great circle paths which sample distinct tectonic provinces.

The group velocity is the velocity of propagation of a wave packet having a limited frequency band around a certain frequency. Since wave packets of different frequencies can be considered to take off at the earthquake source almost simultaneously, the group velocity can be determined by measuring, at a single station, the group arrival times of the wave packets. Since only one station is sufficient and moderate-size earthquakes can be used, it is relatively easy to find an appropriate combination of source and station by which a distinct region can be studied. However, it is known that the interpretation of the group velocity data is not so straightforward as that of the phase velocity data.

The phase and group velocities of Rayleigh and Love waves have been determined along various great circle paths. Individual data are too numerous to be listed here; therefore in Figures 7 and 8 the range of the existing data is indicated. The scatter of the data is due not only to the experimental error but also to the regional difference. The accuracy of the recent measurements is probably better than 1%. It should be noted that since 70% of the earth's surface is covered by ocean, these paths are usually heavily weighted by oceanic mantle; the oceanic percentage along these paths ranges from 50 to 90%, but is highly concentrated between 70 and 80%.

These sets of data can now be compared with theoretical dispersion curves calculated for several models. In Figures 7 and 8, phase and group velocities for Jeffreys and Gutenberg models are shown. The dispersion of Love waves is affected by S velocity and density distributions, and that of Rayleigh waves by P, S velocity and density distributions. The S velocity distribution which is most sensitive to the surface wave dispersion is shown for the Gutenberg and Jeffreys models in Figure 9. For P velocity and density distributions used in the calculations, see references given in the figure caption. These two models are first chosen for comparison because they represent mantles of entirely different characters. In the Jeffreys model, shear velocity in the mantle increases monotonically with depth starting with a relatively low sub-Moho velocity, while in the Gutenberg model, the shear velocity distribution is characterized by a pronounced low-velocity zone capped by a lid with a relatively high velocity. The comparison shows that, for Rayleigh waves, the Gutenberg model obviously provides a much better fit to the data than the

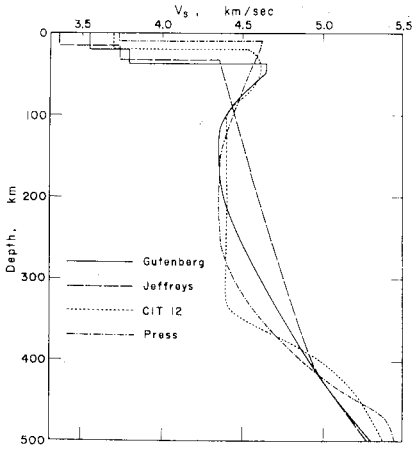

Fig. 9. Shear velocity distributions in the upper mantle for Gutenberg, Jeffreys, CIT 12, and Press models.

Jeffreys model, although the theoretical velocities seem slightly too high as compared with the observed velocities. For Love waves. neither model seems superior to the other, and both models give too low group velocities. It should be remembered here that both models have a completely continental crust while the data are obtained for mixed, but primarily, oceanic, paths. The effect of the crust on the dispersions has been studied by several investigators and is found to be larger for Love waves than for Rayleigh waves, particularly in case of group velocities (*e.g.* fig. 19 of *Kanamori and Abe*, 1968; *Derr*, 1967). Therefore by replacing the continental crust in the Gutenberg model by either an oceanic or an appropriately averaged composite crust, the Gutenberg model can be fitted to both the Rayleigh and Love wave data reasonably well. Thus, it seems almost certain that a velocity reversal or at least a much smaller velocity gradient than that of the Jeffreys model is required for S velocity distribution in the oceanic upper mantle.

Further improvement of the fit between the Gutenberg curve and the experimental curve for Rayleigh waves can be achieved by slightly reducing the velocity in the upper mantle thereby shifting the theoretical phase and group velocity curves downward. A search for a better model along this line has been made by *Toksöz and Anderson* (1966), and *Starovoit et al.* (1968). The CIT 12 model (Figure 9) thus derived by Toksöz and Anderson has an average crust consisting of 3 km thick water layer and 18 km crust, and a lower velocity around the depth of 300 km. The phase velocity curves are shown in Figures 7 and 8 to show a better fit. By using a Monte Carlo procedure *Press* (1970) tested a number of models against various geophysical data. One of the models inverted by Press is shown in Figure 9 and the corresponding phase velocities are shown in Figures 7 and 8. This model which has been selected on a more objective basis again shows a velocity reversal. Thus, aside from details, Gutenberg, CIT 12, and Press models all have common features, a velocity reversal and a relatively low velocity around 300 km depth as compared with the sub-Moho velocity. It can be concluded that these are the characteristic features of the average structure of predominantly oceanic mantle.

The surface wave studies of the mantle structure beneath continents have been made by the two-station or multi-station method. On all continents shield occupies a major part, and tectonically active regions are limited in narrow belts. The application of surface wave technique to spatially limited regions such as orogenic belts is naturally difficult. Consequently most reliable results have been obtained for shields, particularly Canadian shield (*Brune and Dorman*, 1963), and Baltic shield (*Noponen et al.*, 1967). The observed phase velocities for these shields are shown for Love and Rayleigh waves in Figures 10 and 11. Although the period range is limited, it is evident that the observed phase velocities are significantly higher than those for the Gutenberg model. Inverted models are shown in Figure 12 which shows that the shear velocity in the top 150 km of the mantle is 0.3 to 0.4 km/sec higher for Shield models than for the Gutenberg model. Because of the limited period range, the velocity structure below 150 km is not reliable.

A similar conclusion was reached by *Toksöz and Anderson* (1966) who used phase velocities of Love waves along various great circle paths sampling oceanic, shield, and

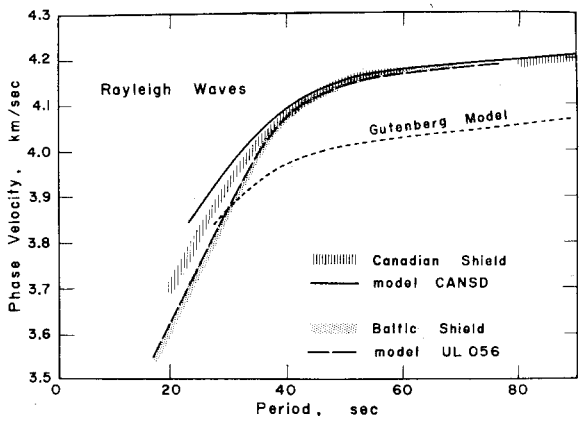

Fig. 10. Phase velocity of Rayleigh waves for Canadian shield (*Brune and Dorman*, 1963) and Baltic shield (*Noponen et al.*,1967). Theoretical curves are for CANSD and UL056 models shown in Figure 12. Gutenberg model is shown for comparison.

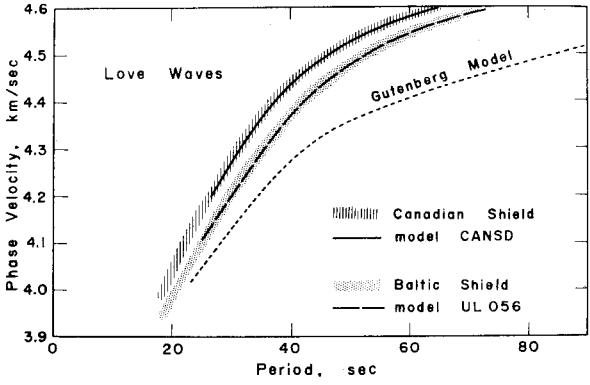

Fig. 11. Phase velocity of Love waves for Canadian shield (*Brune and Dorman*, 1963) and Baltic shield (*Noponen et al.*, 1967).

tectonic mantles. They observed that the Love wave phase velocity is larger for the path having a longer shield path. By extrapolation they constructed phase velocity curves for purely shield, oceanic and tectonic paths, and inverted them to obtain corresponding structures. Although this procedure involves a risky extrapolation, their results seem quite reasonable; their shield structure is quite similar to the CANSD model. The difference between the oceanic and tectonic mantle is relatively small as compared with the marked difference between shield and ocean.

The difference between the mantles beneath shield and tectonically active regions thus suggested from surface wave studies is consistent with the difference of the P_n velocity (Figure 2), travel-time residual (Figures 4, 5, and 6), and structure (Figure 3) between these distinct regions.

3. Q

Another important physical property of the earth is the anelasticity which provides additional information concerning the state, temperature, and the composition of the earth's interior. The anelasticity of the earth can be measured in terms of the attenuation of seismic waves.

Attenuation of seismic waves can be most conveniently described in terms of quality factor Q. The definition of Q is exactly the same as that adopted in other oscillation problems in electric and mechanical engineering. The present state of our knowledge of Q within the earth has been summarized by *Knopoff* (1964), *Sato* (1967), and *Jackson and Anderson* (1969).

The value of Q can be determined, in principle, by measuring the amplitude decay of seismic waves. However, amplitude is also affected by many other factors such as radiation at the source, reflection, refraction, scattering, and seismometer response. In addition, the value of Q may vary not only with depth but also with frequency of the

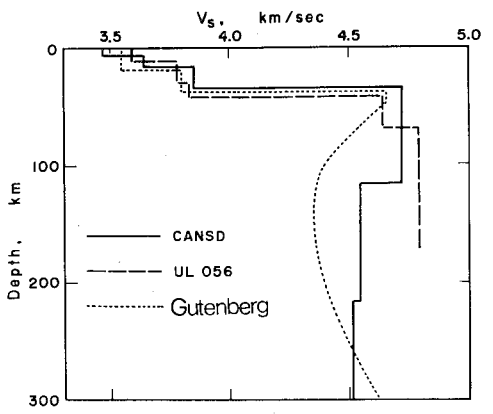

Fig. 12. Proposed shear velocity distributions beneath shields as compared with Gutenberg model. For detailed models see *Brune and Dorman* (1963) and *Noponen et al.* (1967).

waves. Because of these difficulties, neither the detailed distribution of Q with depth nor its frequency dependence is known; the determination of the lateral variability is even more difficult. In the following, only the broad characteristics of the Q distribution within the earth will be discussed.

Probably the first reliable determination of the average (travel-time-weighted harmonic average) Q for S waves, Q_β, for the entire mantle was made by *Press* (1956) who measured the decay of the multiple core reflections, *ScSi*, of nearly vertical incidence, and determined the average Q. In this method the effect of source radiation and the instrument response are removed, and the effect of reflection and refraction are minimized. He obtained $\bar{Q}_\beta \sim 500$ at a period of 11 sec. Later this technique was refined and extended to a wider period range by *Anderson and Kovach* (1964), and *Kovach and Anderson* (1964). They confirmed that $\bar{Q}_\beta \sim 500$ to 600 over a period range 15 to 70 sec. They also determined the average Q_β for the upper 600 km of the mantle to be 150 to 200, and the average Q_β for the lower mantle below 600 km to be about one order of magnitude larger than that for the upper mantle. Since the earthquakes used in these studies are all in South America, these values apply at least to the mantle beneath South America.

A comparable study for P waves is difficult because the reflection coefficient at the mantle-core boundary is very small for nearly vertical incidence; hence no multiple *PcP* phase can be observed for nearly vertical incidence. In order to overcome this difficulty, different combinations such as P and pP (*Teng*, 1968) and P and *PcP* (*Kanamori* 1967*b*) have been used. The results from P and *PcP* combination around the period of 1 to 2 sec show that the average Q_α (Q for P waves) is 410 to 630 for the entire mantle, 180 to 240 for the upper mantle, and 1600 to 6000 for the lower mantle. Since this measurement was made in the western United States, a tectonically active region, these values may be biased toward low Q side. In fact, *Asada and Takano* (1963) suggested, though qualitatively, an existence of high Q mantles. *Archambeau et al.* (1969) also reported a relatively high Q model for 2 to 5 Hz P waves; their model gives an average Q_α for the upper mantle of about 600.

In view of these results and also referring to many other works cited in *Sato* (1967) and *Jackson and Anderson* (1969), it may be said that (1) the value of Q in the upper several hundreds of kilometers of the mantle ranges from 60 to 600 for P waves over the period range from 0.1 to 10 sec, and 60 to 300 for S waves over the period range from 2 to 70 sec. This variation certainly reflects the regional variations, but the experimental error is suspected to be considerable; (2) the value of Q in the lower mantle is extremely large for

both P and S waves. If the value of Q exceeds several thousands, the material can be considered to be perfectly elastic, as far as most seismic phenomena are concerned.

For much longer periods, the attenuation of long-period surface waves and the decay of the earth's free oscillations provide means of estimating the value of Q. The attenuation of long-period surface waves was first measured by *Ewing and Press* (1954) for Rayleigh waves, and by *Satô* (1958) for G waves. The value of Q for free oscillations of various modes was first measured at the time of the 1960 Great Chilean Earthquake.

Following these earlier measurements, many more determinations have been made using more refined techniques. The present status for surface wave Q may be best seen in the plot by *Anderson et al.* (1965) (Figure 13). Here Q is defined in such a way that progressive surface wave amplitude decays with distance Δ as $\exp(-\pi\Delta/QTU)$ where T is period and U is group velocity. Figure 13 clearly shows that the observed Q is frequency-dependent. This does not necessarily mean that the intrinsic Q of the earth's material is frequency-dependent; if Q varies with depth surface waves having different periods would show apparently different attenuation according to the different depth of penetration, even if the intrinsic Q is frequency-independent. It is as yet unknown whether the frequency dependence of Q, the depth variation, or the combination of these two is responsible for the observed frequency dependence of surface wave Q.

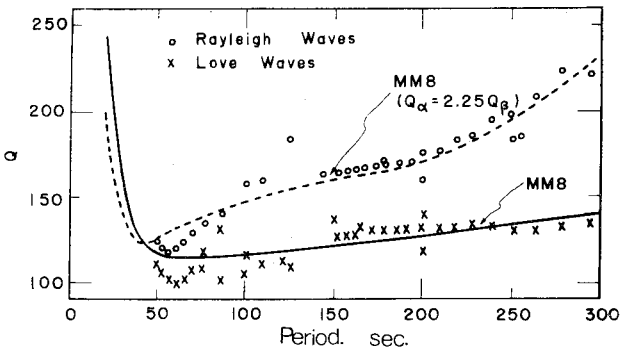

Fig. 13. Experimental and theoretical Q values for Love and Rayleigh waves (after *Anderson et al.* (1965)). The solid curve shows predicted Love wave attenuation for MM8 Q_β model, and the dotted curve shows predicted Rayleigh wave attenuation for MM8 Q_β model coupled with the assumption $Q_\alpha = 2.25 Q_\beta$.

However, it is instructive to interpret the data in terms of the depth variation alone, assuming the frequency independence of intrinsic Q. This has been done by *Anderson and Archambeau* (1964), and *Anderson et al.* (1965). Considering the relatively low value of Q at periods less than 150 sec for Love waves which involve only shear motions, it is evident that the value of Q_β in the shallower part of the mantle cannot be very large, probably around 100. The increase of Q with period reflects the increase of Q with depth. The model thus inverted by *Anderson et al.* (1965) by trial and error method is called MM8 Q_β and shown in Figure 14. This model gives average Q_β of 144 for the upper 600 km of the mantle which may be compared with Q_β determined by body wave studies for the corresponding depth, 60 to 300.

The value of Q_α can be determined from the attenuation of Rayleigh waves which involve both shear and compressional motions. *Anderson et al.* (1965) concluded that $2.5 Q_\beta > Q_\alpha > 1.8 Q_\beta$ which results in the upper mantle average of about 300 for Q_α. The the-

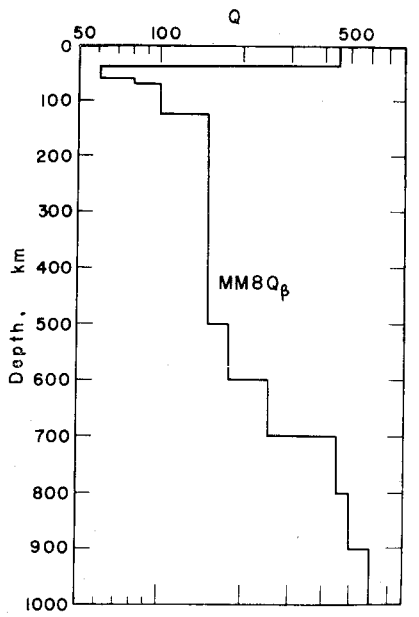

Fig. 14. MM8 Q_β model (after *Anderson et al.* (1965)).

oretical Love wave Q for MM8 Q_β distribution, and the theoretical Rayleigh wave Q for MM8 Q_β with $Q_\alpha = 2.25 Q_\beta$ are shown in Figure 13. They show a good fit with the experimental data.

These values are regarded as representing the average for primarily oceanic paths. Combining these values with those obtained by body wave studies, it may be concluded that, if we assume a frequency independence of Q, the "normal" value of the upper mantle Q is 200 to 300 for Q_α, and 100 to 200 for Q_β with a possible regional variation of a factor of ten. A frequency dependence of factor of 2 to 3 over the seismic frequency range is still a possibility.

4. Special regions

The preceding discussions have been limited primarily to a large scale regionality involving ocean, shield, and tectonic regions. However, in recent years, the existence of large variability in structure within a much smaller region became apparent, especially at mid-oceanic ridges, island arcs and some continental margins (*Otsuka*, 1966; *Bolt and Nuttli*, 1966). At these special regions the structures change so markedly in a horizontal extent of only several hundreds of kilometers that the resolving power of ordinary seismological techniques deteriorates. Consequently, for these regions, the uniqueness and detailed features of the derived structure inevitably become somewhat questionable, although the existence of marked heterogeneity is definite.

Since detailed discussions on individual regions will be made in another session of this symposium, only some examples will be discussed here in order to illustrate the nature of heterogeneity in these regions.

Mid-oceanic region

Ewing and Ewing (1959) suggested, based on the results of their refraction studies, that beneath the Mid-Atlantic ridge in the north Atlantic Ocean an anomalous structure having a velocity of 7.4 km/sec extends to a depth of several tens of kilometers over a width of about 700 km. For the same region, *Tryggvason* (1961) suggested, based on the travel times of earthquake P waves, that a layer of relatively low P velocity, 7.4 km/sec, extends down to the depth of 140 km over a probable width of 1000 km.

Santô and Satô (1966) analyzed the group velocity dispersions of surface waves for various paths which traverse various tectonic regions. They used the "crossing path technique" to assign a characteristic dispersion curve to an individual region. They noted that the group velocity dispersion curves assigned to the East Pacific Rise and the Mid-

Atlantic Ridge show anomalously low velocity towards long period. Although it is not possible to derive a detailed structure because of the limited period range of their data, Santô suggested an existence of a layer of extremely low velocity at a depth of several tens of kilometers. In passing, Santô and Satô did not find the anomalous dispersion for oceanic ridges in the Indian Ocean.

Summarizing these results it can be concluded that the heterogeneity beneath mid-oceanic ridges can be characterized by a low-velocity zone extending to a depth of at least several tens of kilometers; the velocity decrease in this zone is about 0.5 km/sec for P waves. The depth to the bottom of this zone is as yet unknown.

Island Arcs

The structures of island arcs such as Japan, Tonga, New Zealand (*Mooney and Hatherton*, 1969), and the Pacific coast of South America (arc-like structure) (*Saa and Sacks*, 1969) have been extensively studied. A characteristic feature of these regions is an extremely sharp contrast in both velocity and Q within a relatively small horizontal distance. The distributions of high and low velocity, and high and low Q regions are associated with the distribution of intermediate and deep earthquakes. A velocity difference as large as 5% and a ten-fold variation of Q have been suggested. Although the detailed features of the proposed mantle models differ from one arc to another, all the models have a common feature in that the inward (continental) side of the arc is characterized by either a low-velocity or a low Q as compared with the outward (oceanic) side.

These results have been obtained by using (1) travel-time residuals of earthquakes and explosions at stations distributed around island arcs, (2) wave form, amplitude and spectrum of seismic waves for various paths, and (3) the dispersion of surface waves. For example, *Utsu* (1966) interpreted the distribution of seismic intensity (amplitude) observed in Japan in terms of a lateral variation of Q; he suggested an existence of a low-Q region above the inclined seismic zone beneath Japan. *Utsu* (1967) also studied P wave travel-time residuals observed in the northern part of Japan and suggested an existence of a wedge-like high velocity region in which the P velocity is about 6% higher than in the surrounding mantles. *Oliver and Isacks* (1967) compared the wave forms of a number of short-period seismograms observed at the Tonga-Kermadec arc, and found that the wave forms are markedly different for different propagation paths. From these observations they suggested an existence of a plate-like high Q region just beneath the inclined seismic zone. In the high Q region the value of Q is almost one order of magnitude larger than that in the surrounding mantles. *Mitronovas and Isacks* (1969) studied the travel time of P waves for approximately the same region, and concluded that the P velocity is higher by about 5% in the high Q region than in the surrounding mantles at corresponding depths.

Accurate measurements of travel times of teleseismic explosions (*Kanamori*, 1968) showed that the range of the travel-time residuals at Japanese stations amounts to 1.5 sec among the stations only several hundreds of kilometers apart. As shown in Figures 5 and 6, the earliest arrival is comparable to that at stations on shields, and the latest, to that at stations on orogenic belts. Thus we may conclude that the world's fastest and slowest

mantles are juxtaposed in the Japan region. An evidence of a structure more complicated than a simple plate-like or wedge-like structure has been found (*Kanamori*, 1970); the low-velocity region above the inclined seismic zone seems to be locally absent in a region above a gap of earthquake activity.

A study of long-period Love and Rayleigh waves (*Kanamori and Abe*, 1968) suggested that the existence of the anomalously low velocity mantle is not limited only to the mantle beneath the Japanese islands but extends over a much wider region between the Mariana and Ryukyu arcs. This observation seems to be consistent, at least qualitatively, with the results of *Molnar and Oliver* (1969) who found that the transmission of S_n is inefficient in the same region.

5. Concluding remarks

A difference in P velocity of several per cent has been found in the upper 200 km of the mantle between shields and other regions such as orogenic belts and oceans. Regionality of the same magnitude but of a much smaller horizontal scale also exists at especially active regions such as Japan, and Tonga-Kermadec arcs. A difference of the same, or even larger magnitude exists for S waves.

These differences can be interpreted in terms of many factors such as compositions, temperature, and partial melting. The observed variation may be caused by either one, or a combination of two or more of these factors. If we assume that the difference is entirely due to the temperature effect, the P velocity difference of -0.4 km/sec would imply a temperature difference of about 800°C with a commonly used temperature coefficient of P velocity, -5×10^{-4} km/sec. deg. The effect of partial melting depends not only on the concentration of the melt but also on the manner in which it occurs. However, if we simply assume that the rigidity vanishes for a molten portion but that the Lamé parameter λ is unchanged on melting, we can write the velocity V for a material having a liquid inclusion with a small concentration c by $V=V_0\ (1-0.58\ c)$ (*Hashin*, 1962) where V_0 is the velocity for $c=0$. Then the 5% reduction in velocity would imply a 9% partial melting. These values delimit the maximum possible range of temperature difference and partial melting which can be incoprorated in interpreting other geophysical data, although allowance should be made for a possible compositional change.

The effect of temperature and partial melting on Q of earth materials is not well-known. However, some experimental results (*Mizutani and Kanamori*, 1964; *Spetzler and Anderson*, 1968) suggest a sharp drop of Q at the onset of melting. In view of this, the observed ten-fold variation of Q seems to provide a favorable evidence for an existence of partial melting in the mantle of the unusual regions.

Since this paper is not intended to be an extensive review of the subject, the selection of the materials is rather arbitrary, and many of the results are often used without specifically making reference to individual sources. Further, in view of the recent vast outflow of literature in which considerable controversy is sometimes found, it is not possible for the author to make a complete and entirely objective review. What is intended here is to

draw, without going into too much details, several conclusions which seem to the author most reasonable in the light of currently available data.

I thank Allan Cox for kindly reading the manuscript and offering suggestions for its improvement.

Note added in proof

After this paper was submitted for publication, a great many data of great-circle phase and group velocities were collected for both Love and Rayleigh waves. These data with good estimates of errors permit one to make a more comprehensive discussion of mantle structure than that made in section 2 of this paper. For details, reference is made to Kanamori, H., Velocity and Q of mantle waves, *Phys. Earth Planet. Interiors*, **2**, 1970.

References

Anderson, Don L., Recent evidence concerning the structure and composition of the earth's mantle, *Phys. Chem. Earth*, **6**, 1–131, 1965.

Anderson, Don L., Latest information from seismic observasions, Chapter 12, in *The Earth's Mantle*, edited by T.F. Gaskell, pp. 355–420, Academic Press, London, 1967.

Anderson, Don L., and C.B. Archambeau, The anelasticity of the earth, *J. Geophs. Res.*, **69**, 2071–2084, 1964.

Anderson, Don L., A. Ben-Menahem, and C.B. Archambeau, Attenuation of seisimic energy in the upper mantle, *J. Geophys. Res.*, **70**, 1441–1448, 1965.

Anderson, Don L., and R.L. Kovach, Attenuation in the mantle and rigidity of the core from multiply reflected core phases, *Proc. Natl. Acad. Sci. U.S.*, **51**, 168–172, 1964.

Archambeau, C.B., E.A. Flinn, and D.G. Lambert, Fine structure of the upper mantle, preprint, Contribution No. 1589, Division of Geological Sciences, California Institute of Technology, 1969.

Asada, T., and K. Takano, Attenuation of short period *P* waves in the mantle, *J. Phys. Earth*, **11**, 25–34, 1963.

Bolt, B.A., Recent information on the earth's interior from studies of mantle waves and eigenvibrations, *Phys. Chem. Earth*, **5**, 55–119, 1963.

Bolt, B.A., and O.W. Nuttli, *P* wave residuals as a function of azimuth, 1, *J. Geophys. Res.*, **71**, 5977–5985, 1966.

Brune, J., and J. Dorman, Seismic waves and earth structure in the Canadian shield, *Bull. Seismol. Soc. Am.*, **53**, 167–209, 1963.

Bullen, K.E., *An Introduction to the Theory of Seismology*, Third Edition, Univ. Press, Cambridge, 1963.

Carder, D.S., D.W. Gordon, and J.N. Jordan, Analysis of surface-foci travel times, *Bull. Seismol. Soc. Am.*, **56**, 815–840, 1966.

Chinnery, M.A., and M.N. Toksöz, *P*-wave velocities in the mantle below 700 km, *Bull. Seismol. Soc. Am.*, **57**, 199–226, 1967.

Derr, J.S., A comparison of free oscillations of oceanic and continental earth models, *Bull. Seismol. Soc. Am.*, **57**, 1047–1061, 1967.

Ewing, J., and M. Ewing. Seismic-refraction measurements in the Atlantic ocean basins, in the Mediterranean sea, on the Mid-Atlantic ridge, and in the Norwegian sea, *Bull. Geol. Soc. Am.*, **70**, 291–317, 1959.

Ewing, M., and F. Press, An investigation of mantle Rayleigh waves, *Bull. Seismol. Soc. Am.*, **44**, 127–147, 1954.

Green, R.W. E., and A.L. Hales, The travel times of P waves to 30° in the central United States and upper mantle structure, *Bull. Seismol. Soc. Am.*, 58, 267–289, 1968.

Hales, A.L., J.R. Cleary, and J.L. Roberts, Velocity distribution in the lower mantle, *Bull. Seismol. Soc. Am.*, 58, 1975–1989, 1968.

Hashin, Z., The elastic moduli of heterogeneous materials, *J. Appl. Mechanics*, **29**, 143–150, 1962.

Herrin, E., E.P. Arnold, B.A. Bolt, G.E. Clawson, E.R. Engdahl, H.W. Freedman, D.W. Gordon, A.L. Hales, J.L. Lobdell, O. Nuttli, C. Romney, J. Taggart, and W. Tucker, 1968 Seismological Tables for P phases, *Bull, Seismol. Soc. Am.*, 58, 1193–1351, 1968.

Jackson, D.D., and Don L. Anderson, Physical mechanisms of seismic wave attenuation, *Rev. Geophys.*, in press, 1969.

Jeffreys, H., Revision of travel times, *Geophys. J. R. Astr. Soc.*, **11**, 5–12, 1966.

Jeffreys, H., and K.E. Bullen, *Seismological Tables*, British Association for the Advancement of Science, Gray Milne Trust, London, 1958.

Johnson, L.R., Array measurements of P velocities in the upper mantle, *J. Geophys. Res.*, **72**, 6309–6325, 1967.

Johnson, Lane R., Array measurements of P velocities in the lower mantle, *Bull. Seismol. Soc. Am.*, **59**, 973–1008, 1969.

Kanamori, H., Upper mantle structure from apparent velocities of P waves recorded at Wakayama Micro-Earthquake Observatory, *Bull. Earthquake Res. Inst.Tokyo Univ.*, **45**, 657–678, 1967*a*.

Kanamori, H., Attenuation of P waves in the upper and lower mantle, *Bull. Earthquake Res. Inst. Tokyo Univ.*, **45**, 299–312, 1967*b*.

Kanamori, H., Travel times to Japanese stations from Longshot and their geophysical implications, *Bull. Earthquake Res. Inst. Tokyo Univ.*, **46**, 841–859, 1968.

Kanamori, H., Mantle beneath the Japanese arc, *Physics of the Earth and Planetary Interiors*, **3**, 475–483, 1970.

Kanamori, H., and K. Abe, Deep structure of island arcs as revealed by surface waves, *Bull. Earthq. Res. Inst. Tokyo Univ.*, **46**, 1001–1025, 1968.

Knopoff, L., Q, *Rev. Geophys.*, **2**, 625–660, 1964.

Kovach, R.L., Seismic surface waves: Some observations and recent development, *Phys. Chem. Earth*, **6**, 251–314, 1965.

Kovach, R.L., and Don L. Anderson, Attenuation of shear waves in the upper and lower mantle, *Bull. Seismol. Soc.Am.*, 54, 1855–1864, 1964,.

Lukk, A.A., and I.L. Nersesov, Structure of the upper mantle as shown by observations of earthquakes of intermediate focal depth, *Dokl. Akad. Nauk SSSR*, **162**, 14–16, 1965.

Mitronovas, W., and B. Isacks, Seismic velocity anomaly in the upper mantle beneath the Tonga-Kermadec island arc (abstract), *Trans. Am. Geophys. Union*, **50**. 246, 1969.

Mizutani, H., and H. Kanamori, Variation of elastic wave velocity and attenuative property near the melting temperature, *Journal of Physics of the Earth*, **12**, 43–49, 1964.

Molnar, P., and J. Oliver, Lateral variations of attenuation in the upper mantle and discontinuities in the lithosphere, *J. Geophys. Res.*, 74, 2948-2682, 1969.

Mooney, H.M., and T. Hatherton, Upper mantle inhomogeneity beneath New Zealand, abstract, *Trans, Am. Geophys. Union*, **50**, 246, 1969.

Noponen, I., M.T. Porkka, S. Pirhonen, and U. Luosto, The crust and mantle tn Finland, *Journal of Physics of the Earth*, **15**, 19–24, 1967.

Nuttli, O., Seismological evidence pertaining to the structure of the earth's upper mantle, *Rev. Geophys.*, **1**, 351–400, 1963.

Oliver, J., and B. Isacks, Deep earthquake zones, anomalous structures in the upper mantle, and the lithosphere, *J. Geophys. Res.*, **72**, 4259–4275, 1967.

Otsuka, M., Azimuth and slowness anomalies of seismic waves measured on the central California seismographic array, 2, *Bull. Seismol. Soc. Am.*, **56**, 655–675, 1966.

Press, F., Rigidity of the earth's core, *Science*, **124**, 1204, 1956.

Press, F., Long-period waves and free oscillations of the earth, *Research in Geophysics, Vol. 2, Chap. 1*, edited by H. Odishaw, pp. 1–26, MIT Press, Cambridge, Massachusetts, 1964.

Press, F., Seismological information and advances in *Advances in Earth Sciences*, edited by P.M. Hurley, pp. 247–286, MIT Press, Cambridge, Massachusetts, 1966*a*.

Press, F., Seismic velocities, in *Handbook of Physical Constants*, edited by S.P. Clark Jr., pp. 195–218, Geol. Soc. Am., New York, 1966*b*.

Press, F., Earth models consistent with geophysical data, *Physics of the Earth and Planetary Interiors*, **3**, 3–22, 1970.

Saa, G., and S.I. Sacks, The effect of elevation and region on the *P* and *S* waves of some South American Seismograph Stations, *Trans. Am. Geophys. Union*, **50**, 237, 1969.

Santo, T., and Y. Satô, World-wide survey of the regional characteristics of group velocity dispersion of Rayleigh waves, *Bull. Earthquake Res. Inst. Tokyo Univ.*, **44**, 939–964, 1966.

Sato, R., Attenuation of seismic waves, *Journal of Physics of the Earth*, **15**, 32–61, 1967.

Satô, Y, Attenuation, dispersion and the wave guide of the *G* wave, *Bull. Seismol. Soc. Am.*, **48**, 231–251, 1958.

Savarenskii, E.F., Investigation of the upper mantle of the earth by seismic body waves, *Bull. Acad. Sci. USSR, Geophys. Ser., English Transl.*, 133–140, 1966.

Spetzler, H., and Don L. Anderson, The effect of temperature and partial melting on velocity and attenuation in a simple binary system, *J. Geophys. Res.*, **73**, 6051–6060, 1968.

Starovoit, O.E., E.F. Savarensky, and S.A. Fedorov, The structure of the earth's mantle and observations of long-period surface waves, *Gerlands Beitr. Geophys.*, **71**, 205–214, 1968.

Teng, T. L., Attenuation of body waves and the *Q* structure of the mantle, *J. Geophys. Res.*, **73**, 2195–2208, 1968.

Toksöz, M.N., and Don L. Anderson, Phase velocities of long-period surface waves and structure of the upper mantle, 1, *J. Geophys. Res.*, **71**, 1649–1658, 1966.

Toksöz, M.N., M.A. Chinnery, and Don L. Anderson, Inhomogeneities in the earth's mantle, *Geophys. J.*, **13**, 31–59, 1967.

Tryggvason, E., Wave velocity in the upper mantle below the Arctic-Atlantic ocean and northwest Europe, *Ann. Geofis. Rome*, **14**, 379–392, 1961.

Umbgrove, J.H.F., *The Pulse of the Earth*, Second edition, Martinus Nijhoff, Hague, 1947.

Utsu, T., Regional differences in absorption of seismic waves in the upper mantle as inferred from abnormal distributions of seismic intensities, *J. Fac. Sci. Hokkaido Univ. Japan, Ser. VII*, **2**, 359–374, 1966.

Utsu, T., Anomalies in seismic wave velocity and attenuation associated with a deep earthquake zone (1), *J. Fac. Sci. Hokkaido Univ. Japan, Ser. VII*, **3**, 1–25, 1967.