Viscosity Determination in the Earth's Outer Core from ScS and SKS Phases

By

Yasunori SUZUKI and Ryosuke SATO

Geophysical Institute, Faculty of Science, University of Tokyo

Abstract

Attenuation of seismic waves in the Earth's outer core is investigated by using pairs of transmitted waves through the outer core and reflected waves at the core-mantle boundary, which have common ray paths in the mantle.

Taking into account the conditions: that the angle of incidence at the core-mantle boundary is small so that ray theory is applicable; that the phase to be analyzed should be clear and not contaminated by other phases; and that the ray does not penetrate into the inner core; pairs of SKS and ScS are used to determine the viscosity coefficient in the outer part of the Earth's outer core.

Long-period seismograms of three deep-focus earthquakes supplied by the U.S.C. G.S. are used in this analysis. Ratios of the Fourier spectral amplitudes of SKS to those of ScS in the frequency range of 0.04 Hz to 0.15 Hz, after eliminating the effects of crustal layering and the seismograph response, yield a mean viscosity coefficient $\bar{\nu}$ in the outer core. Using 16 pairs of SKS and ScS waves $\bar{\nu}$ in the above frequency range is estimated to be 3×10^{10} to 7×10^{10} dyne·sec/cm², which is consistent with the $\bar{\nu}$ determination from multiply-reflected ScS's at the core-mantle boundary.

§ 1. Introduction

Recently, studies on attenuation of seismic waves in the mantle have been made by various investigators, for example, ANDERSON and ARCHAMBEAU (1964), ANDERSON and KOVACH (1964), KOVACH and ANDERSON (1964), ANDERSON *et al.* (1965), TENG (1966, 1968), KURITA (1966), HIRASAWA and TAKANO (1966), SATO and ESPINOSA (1967), MIKUMO and KURITA (1968), using long-period seismic wave recordings; and TAKANO (1966), KANAMORI (1967 a, b, c), SUMNER (1967), and KURITA (1968), using short-period seismic wave recordings. However, the detailed *Q*-structure in the mantle has not been established as yet, and further work is required to ascertain the variations of *Q* with depth.

There have been several proposed speculations in the constitution of the physical properties of the core, but there are quite a number of unknown parameters. So far, one of the few accepted properties is that the rigidity of the outer core is negligible, if it

exists, and in this sense the outer core is regarded to be liquid.

The velocity distribution in the core has been studied by several investigators, such as JEFFREYS (1939), GUTENBERG (1959), BOLT (1964), ADAMS and RANDALL (1964), and their overall results are similar structures, at least in the outer core. For the inner core, however, it may be stated that the physical properties, even velocity distribution, are still unknown. Taking into account the actual knowledge of the inner and outer core, it was decided in the present study to investigate the attenuation of seismic waves traveling through the outer core only, eliminating the effects of attenuation in the mantle.

There have been several attempts to obtain the rigidity of the core assuming that it is a soft-solid, or the viscosity of the core assuming that it is a viscous-liquid medium. One of these estimations has been done by determining a change of period of the free oscillation of the earth, and some other studies from the reflectivity of seismic body waves

at the core-mantle boundary. These studies are summarized in the paper by SATO and ESPINOSA (1967). There have been no investigations to directly determine the attenuation of seismic waves by analyzing records of waves transmitted through the Earth's outer core.

In this study the outer core is considered to be a viscous-liquid medium, and the mean viscosity is determined from the spectral amplitude ratios of waves transmitted through the outer core and waves reflected at the core-mantle boundary, which have common ray paths in the mantle. An important question concerned with the physical properties of the core arises whether the attenuation in the outer core behaves like that in a solid; where the attenuation factor is proportional to frequency (the dimensionless quality factor Q is independent of frequency), or it behaves like that in a viscous-liquid medium; where the attenuation factor is proportional to the square of the frequency (the viscosity coefficient ν is independent of frequency).

The applicability of a constant Q -model for the outer core in the frequency rang under consideration is also discussed.

§2. Method of Analysis

In Fig. 1 is shown schematically a transmitted wave through the outer core and a reflected wave at the core-mantle boundary which have common ray paths in the mantle. The take-off angle at the focus, the angle of

incidence at the core-mantle boundary, θ_c , and at the Mohorovicic discontinuity, θ_m , as used in this study are shown in this figure.

The observed amplitude spectra of the outer core phase, $A_c(f)$, and of the core-reflected phase, $A_R(f)$, are expressed by

$$A_c(f) = A_0(f) \cdot P_c \cdot T \cdot D_c \cdot C_c(f) \cdot S_c(f) \cdot X_c(f) \quad (1)$$

and

$$A_R(f) = A_0(f) \cdot P_R \cdot R \cdot D_R \cdot C_R(f) \cdot S_R(f) \cdot X_R(f) \quad (2)$$

respectively, where

- f : frequency,
- $A_0(f)$: source spectrum,
- P_c and P_R : radiation pattern effects,
- T and R : transmission and reflection coefficients, at the core-mantle boundary,
- D_c and D_R : divergence factors due to geometrical spreading,
- $C(f)$: transfer function of the crustal layering,
- $S(f)$: amplitude response of seismograph system,
- $X(f)$: effect of attenuation which is given below.

If the attenuation is like that in a solid and Q is frequency independent, then

$$X(f) = \exp\left(-\pi f \int_{\text{ray path}} \frac{ds}{QV}\right)$$

and if the attenuation is like that in a viscous-liquid medium and the viscosity coefficient ν is frequency independent, then

Fig. 1. A pair of core phase and core reflected phase (A schematic picture).

$$X(f) = \exp \left(-2\pi^2 f^2 \int_{\text{ray path}} \frac{\nu ds}{\rho V^3} \right)$$

where V is the seismic wave velocity, ρ is the density, ds is the ray path element and the integral is performed along the entire ray path.

Now let us define the modified ratio A of the amplitude spectra by

$$A \equiv \frac{A_C(f)/C_C(f) \cdot S_C(f)}{A_R(f)/C_R(f) \cdot S_R(f)} \quad (3)$$

From equation (1) and (2), we obtain

$$A = \left(\frac{P_C \cdot T \cdot D_C}{P_R \cdot R \cdot D_R} \right) \frac{X_C(f)}{X_R(f)} \quad (4)$$

We may reasonably regard the factor in parenthesis in the right-hand side of equation (4) as frequency independent, so far as we use the phases for which ray theory is applicable. The factor $X_C(f)/X_R(f)$ is reduced to the attenuation effect in the outer core only, because the phases forming the pair have almost the same paths in the mantle and so the attenuation effects in the mantle are naturally the same. We will discuss the effects of lateral inhomogeneity in the upper mantle in section 5.

If the outer core is a viscous-liquid medium, equation (4) becomes

$$A = \text{const} \cdot \exp \left[-2\pi^2 f^2 \bar{\nu} \int_{\substack{\text{ray path in} \\ \text{(the outer core)}}} \frac{ds}{\rho V^3} \right]$$

where $\bar{\nu}$ is mean viscosity coefficient in the outer core.

Then,

$$\ln A = \text{const} - 2\pi^2 f^2 \bar{\nu} \int \frac{ds}{\rho V^3} \quad (5)$$

$$= \text{const} - f^2 \bar{\nu} J \quad (6)$$

where

$$J = 2\pi^2 \int_{\substack{\text{ray path in} \\ \text{(the outer core)}}} \frac{ds}{\rho V^3} \quad (7)$$

J as defined by equation (7) can be calculated from a given density and velocity structure in the outer core. Taking the Jeffreys-Bullen structure, J is shown in Fig. 5 as a function of angle of incidence at the core-mantle boundary. The use of different

structures in the outer core yields only slight differences in J .

According to the method described above, both pairs of *PKP-PcP* and *SKS-ScS* seem to be adequate for the study of the attenuation in the outer core. But the pair of *PKP-PcP* is inadequate by the following reasons:

1. In order that we can take the transmission and reflection coefficients at the core-mantle boundary as frequency independent, the angle of incidence θ_C should be small. SATO (1969) showed that ray theory is not applicable even for waves of period less than 20~30 sec, if θ_C is greater than 70°, and
2. the *PKP* phase should not be contaminated by *PKIKP* propagated through the inner core.

The epicentral distance (Δ) for *PKP* phases which satisfy the above two conditions is limited to a narrow epicentral distance range at about 150°.

On the other hand for the pair of *SKS-ScS* waves, the appropriate range of Δ for *SKS* phases is from 70° to 120°, and that for *ScS* phases is from 15° to 55°. Except at about 80° where *SKS*, *S* and *SKKS* arrive at the station with very small time intervals, and at about 40° where *ScS* and *SS* arrive almost

Fig. 2. An example of nomogram to find a relation Δ_{SKS} and θ_C for a focal depth of 610 km

Fig. 3. Δ_{ScS} versus θ_C .

Fig. 4. An example of nomogram to obtain an appropriate pair of SKS and ScS which have the same take-off angle at the source.

simultaneously. As is shown in Figs. 2 and 3, θ_C for SKS and ScS in the above epicentral range is from 20° to 45° and so ray theory is applicable for these phases. Taking into account that even if Δ for SKS changes by several degrees, the corresponding change of the angle at the hypocenter is only 1 to 2 degrees, hence we can make use of several observations of SKS for one observation of ScS. In Fig. 4 are shown Δ_{SKS} and Δ_{ScS} as a function of the take-off angle for a focal depth of 610 km. When the SKS and ScS have the same take-off angle they make a proper pair.

According to the reasons and constraints mentioned above, pairs of SKS and ScS are analyzed in this study.

§3. Data

In Table 1 are listed three earthquakes used in the present study. The origin times, epicenters, focal depths, body-wave magnitudes m_b are taken from the Seismological Bulletin of the U.S.C.G.S. Several nomograms are prepared for these earthquakes to obtain the appropriate pairs of SKS and ScS and to find the angles of incidence at Moho and at the core boundary for the phases, such as those shown in Figs. 2 through 5.

Since it is required that both SKS and ScS phases should be clearly recorded, the earthquakes selected are deep focus (590~640 km) and are fairly large magnitudes ($m_b=6.2$). The analyzed records are from long-period

Table 1. Information of the earthquakes analysed (USCGS)

Shock No.	Origin Time	Location	Depth (Km)	m_b
1	Dec. 28, 1964 16 h 16 m 11.0 s	22°.1S, 179°.6 W South of Fiji Islands	611	6.2
2	Mar. 17, 1966 15 h 50 m 33.1 s	21°.1S, 179°.2 W Fiji Islands Region	639	6.2
3	Nov. 3, 1965 01 h 39 m 0.31 s	9°.1S, 71°.4 W Peru-Brazil Border Region	593	6.2

Fig. 5. J , defined by equation (7), versus θ_c . The lowermost scale indicates Δ_{SKS} for a focal depth of 610 km.

horizontal component seismograms obtained at WWSSN stations. Since the SV components are more adequate than the SH components for our study, we tried to analyze the longitudinal components rather than the transverse components. When it was impossible to utilize the SV components, we were forced to use the SH components. There were only two such cases out of twenty-four.

The epicentral distances and azimuths seen from the epicenter are shown for earthquake No. 1 in Fig. 6. The stations are indicated by the three letter codes of the U.S.C.G.S. The records at the stations in brackets are analyzed, but were not used for the determination of the mean viscosity $\bar{\nu}$.

Since the free periods of the pendulum and galvanometer are reported as the same for all the stations, 30 sec-100 sec at the time of No. 1 earthquake and 15 sec-100 sec at the time of No. 2 and No. 3 earthquakes, the seismograph responses were assumed to be the same. So the factor $S_G(f)/S_R(f)$ in equation (3) was set equal to 1.

Crustal structures are determined as follows. The P wave velocity and layer thick-

Fig. 6. Azimuth and epicentral distance of stations seen from the epicenter (No. 1 earthquake).

ness are taken from a compiled result by STEINHART and MEYER (1961), and the density is obtained by using NAFE and DRAKE's velocity-density curve (TALWANI *et al.*, 1959). The appropriate Poisson's ratio was used to calculate S wave velocity.

Travel times of SKS and ScS phases for each earthquake are all in good agreement with those of the Seismological Tables of JEFFREYS and BULLEN (1967).

The original films are enlarged by a factor of 40, and digitized at every 1/4 or 1/2 seconds. The following data window is applied with $T_m = 25 \sim 30$ seconds;

$$w(t) = \frac{1}{2} \left(1 + \cos \frac{\pi t}{T_m} \right); |t| < T_m$$

$$= 0 \quad ; |t| > T_m$$

Except for one case, the onsets of the phases are taken as the origin time of analysis, which implies the signals are assumed to be zero before the onsets. The onset is not sharp in the case of the ScS at RIV of earthquake No. 2, so the origin is set about

the maximum amplitude of the ScS phase, and the whole window is used. Examples of ScS and SKS wave-forms are shown in Figs. 7, 8, and 9 for all three earthquakes. The arrow with the J-B mark indicates the arrival time taken from the Seismological Tables. The time intervals used for digitization are shown on these records by two small arrows.

§4. Results

The mean viscosity coefficient $\bar{\nu}$ is determined by equation (6) in section 2, using the modified spectral ratio A in the frequency range from 0.04 Hz to 0.15 Hz. The spectral ratio A for the SKS-ScS pair and calculated curves for various values of $\bar{\nu}$ are shown in in Figs. 10 to 12. The values estimated from each pair are summarized in Table 2 and most of them are in the range from 3×10^{10} to 7×10^{10} dyne-sec/cm².

Next we consider the case in which the quality factor Q is frequency independent as in the case for a lossy solid. Hence, the

Fig. 7. Waveforms of ScS and SKS from the Dec. 28, 1964 earthquake (No. 1).

NO. 2

Fig. 8. Waveforms of *ScS* and *SKS* from the Mar. 17, 1966 earthquake (No. 2).

NO. 3

Fig. 9. Waveforms of *ScS* and *SKS* phases from the Peru-Brazil earthquake of Nov. 3, 1965 (No. 3).

Fig. 10. a) Determination of $\bar{\nu}$ assuming that it is frequency independent in the frequency range under consideration. Solid curves indicate theoretical curves for a given $\bar{\nu}$ value, and a broken curve is the mean of the modified spectral ratios of three pairs (earthquake No. 1).

Fig. 10. b) Modified spectral ratio of SKS and $S\omega S$ for earthquake No. 1 and theoretical curves for constant $\bar{\nu}$.

Fig. 10. c) $\bar{\nu}$ determination from earthquake No. 1. A somewhat large difference between two $\bar{\nu}$ values obtained from two pairs may be due to large difference in azimuthal direction of COL from other stations (*cf.* Fig. 6).

Fig. 11. a) $\bar{\nu}$ determination from earthquake No. 2. Fig. 11. b) $\bar{\nu}$ determination from earthquake No. 2. Fig. 12. a) $\bar{\nu}$ determination from earthquake No. 3.

Fig. 12. b) $\bar{D}$ determination from earthquake No. 3. Fig. 12. c) $\bar{D}$ determination from earthquake No. 3. Fig. 13. a) $\bar{Q}$ determination from earthquake No. 1.
Observed data are the same as in Fig. 10. a).

Fig. 13. b) $\bar{Q}$ determination from earthquake No. 2. Observed data are the same as in Fig. 11. a).

Fig. 13. c) $\bar{Q}$ determination from earthquake No. 2. Observed data are the same as in Fig. 11. b).

following equation must be used instead of equation (6).

$$\begin{aligned} \ln A &= \text{const} - \frac{\pi f}{\bar{Q}} \int_{\text{(ray path in the outer core)}} \frac{ds}{V} \\ &= \text{const} - \frac{\pi t_k}{\bar{Q}} f \end{aligned} \quad (8)$$

where $\bar{Q}$ is the average quality factor in the outer core, and t_k is the SKS travel-time through the outer core.

Several cases are shown for the $\bar{Q}$ determination in Fig. 13, using the same data as in the ν determination. The $\bar{Q}$ values estimated from each pair are summarized in the

Table 2. Results of $\bar{\nu}$ - and $\bar{Q}$ -determination

Earthq. No.	SKS	ScS	$\bar{\nu}$ ($\times 10^{10}$ c.g.s.)	$\bar{Q}$
1	LPB (EW)	RIV (EW)	3~5	100~200
	QUI (EW)	RIV (EW)	5	100~200
	ARE (EW)	RIV (EW)	5~7	100~200
	Average		5	150
	NNA (EW)	TAU (EW)	2~3	200~300
	GIE (EW)	ADE (EW)	2~3	200~300
2	COL (NS)	ADE (EW)	10	60
	LPB (EW)	RIV (EW)	5~7	100~200
	QUI (EW)	RIV (EW)	3~5	200~300
	ARE (EW)	RIV (EW)	5~7	100~200
	ANT (EW)	RIV (EW)	3~5	200~300
	Average		5	150~200
	BOZ (EW)	*ADE (NS)	1~2	200~300
	ALQ (EW)	*ADE (NS)	2~3	300~500
3	NAI (EW)	TRN (NS)	2~4	200~300
	*LON (NS)	ALQ (NS)	6~8	100
	COR (EW)	ALQ (NS)	10	50
	ESK (EW)	JCT (NS)	1~2	

* indicates transverse component to the ray path.

right column of Table 2, and most of them are in the range from 100 to 300.

We have investigated the attenuation in the outer core in two ways. Though we cannot conclude definitely in what way the seismic waves are attenuated through the outer core, it is most likely that the attenuation is like that in a viscous-liquid medium, where the viscosity coefficient ν is independent of frequency.

§ 5. Discussion

In the previous section we obtained the results that the outer core may be regarded as a viscous-liquid medium and the estimated values of the mean viscosity coefficient are in the range from 3×10^{10} to 7×10^{10} dyne·sec/cm². These value are consistent with the upper limit of 8.6×10^{10} dyne·sec/cm², which was found from multiply reflected ScS phases by SATO and ESPINOSA (1968).

In this paper the seismograph responses are assumed to be the same for all the observing stations since the free periods of seismometer and galvanometer are approximately the same. To check this assumption, we calculated the seismograph response for several stations according to the method developed by ESPINOSA *et al.* (1965), and found the overall response of the system to be similar, to a first order approximation.

Concerning the correction on crustal structure necessary for the calculations of the transfer functions, there remains some ambiguity. The results of near profiles for some stations are directly available (for example, ADE, RIV, ALQ, and ARE), but for some other stations (for example, GIE and NAI) there are no profile data available. For these cases the results of profiles having a similar geological structure were used. Since the spectral ratio not corrected for the crustal

transfer functions is too irregular to determine the mean viscosity coefficient $\bar{\nu}$, the correction of the crustal transfer function will be necessary in a future investigation.

In the preceding sections we neglected the effect of lateral inhomogeneity in the upper mantle. It is shown here that the effect is negligible in the present $\bar{\nu}$ determination.

Instead of equation (6) in section 2, the following equation for the case of laterally inhomogeneous mantle is used;

$$\ln A = \text{const} - f^2 \bar{\nu} J - \pi f \Delta q \quad (9)$$

where $q = T/Q$, T being the travel time through the upper mantle, and Δq is the difference of q between SKS and ScS recording stations.

Here we consider the lateral inhomogeneity effect in the velocity- and Q -structure for the the uppermost 200 km of the mantle. The travel-time from a point with 200 km in depth to the Moho discontinuity is 39 sec for the SKS phase at $\Delta s_{ks} = 70^\circ$, and 38 sec for the SKS phase at $\Delta s_{ks} = 130^\circ$. This means that the angle of incidence due to the difference in the velocity structure must be very small.

If the difference of Q in the upper mantle, beneath the SKS and the ScS recording stations, is ϵ then Δq becomes $T/Q \epsilon/Q$. Using values of 40 sec, 150 and 50 for T , Q and ϵ , respectively, we get $\Delta q \approx 0.09$ sec. The third term on the right-hand side of equation (9) becomes $0.3f$. Using values of 11×10^{-10} and 5×10^{10} for J and $\bar{\nu}$, the second term becomes $55f^2$. In the frequency range from 0.04 Hz to 0.15 Hz presented, the second term is greater than the third term by a factor of 1. Therefore, the effect of lateral inhomogeneity in the upper mantle does not affect the $\bar{\nu}$ determination with the accuracy considered in the present study.

One of the severe constraints of the present investigation is the scantiness of the data which is available for this type of research. However, further investigation with more refined data will yield a detailed information on the attenuation model of the outer core.

§ 6. Summary

Attenuation of seismic waves in the outer

core is investigated by utilising pairs of SKS and ScS phases which have common ray paths in the mantle, eliminating in this way the effects of crustal structure beneath the observing stations. Since the mechanism of attenuation in the outer core is not known at all, two mechanisms are considered. One is that the attenuation coefficient is proportional to the square of the frequency which corresponds to the attenuation in a viscous-liquid medium. The other is that the attenuation coefficient is proportional to frequency, which corresponds to the attenuation in a lossy solid.

Comparison of the theoretical curves under the above mentioned assumptions for a loss-mechanism with the observed data suggests, but not definitely, that the attenuation in the outer core is interpreted well as a viscous-liquid medium. In the frequency range from 0.04 Hz to 0.15 Hz, the mean viscosity coefficient $\bar{\nu}$ is estimated to be 3×10^{10} to 7×10^{10} dyne·sec/cm².

§ 7. Acknowledgements

The authors are indebted to Dr. A. F. Espinosa who read the manuscript critically and offered valuable suggestions. They are grateful to Professor T. Asada for his encouragement during the course of the present study. All the members of Seismological Division of Geophysical Institute, Faculty of Science, University of Tokyo, contributed valuable help and advice.

Computations in this study were carried out by IBM-7090 through UNICON in the International Business Machine Cooperation of Japan, to whom the author's sincere thanks are also due.

References

- Adams, R. D., and M. J. Randall, The fine structure of the Earth's core, *Bull. Seism. Soc. Amer.*, **54**, 1299-1313, 1964.
- Anderson, D. L., and C. B. Archambeau, The anelasticity of the Earth, *J. Geophys. Res.*, **69**, 2071-2084, 1964.
- Anderson, D. 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.
- Anderson, D. L., A. Ben-Menahem, and C. B. Archambeau, Attenuation of seismic energy in the upper mantle, *J. Geophys. Res.*, **70**, 1441-1448, 1965.
- Bolt, B. A., The velocity of seismic waves near the Earth's center, *Bull. Seism. Soc. Amer.*, **54**, 191-208, 1964.
- Espinosa, A. F., G. H. Sutton, and H. J. Miller, A transient technique for seismograph calibration-manual and standard set of theoretical responses, Geophys. Lab., Inst. of Science and Tech., Univ. of Michigan, 1965.
- Gutenberg, B., *Physics of the Earth's Interior*, Academic Press, New York, 1959.
- Hirasawa, T., and K. Takano, Differential attenuation of *P* waves as derived from a Hindu-Kush earthquake, *J. Phys. Earth*, **14**, 49-57, 1966.
- Jeffreys, H., and K. E. Bullen, *Seismological Tables*, British Association for Advance for Science, Gray-Milne Trust, 1967.
- Jeffreys, H., The times of the core waves (second paper), *M.N.R.A.S. Geophys. Suppl.*, **4**, 594-615, 1939.
- Kanamori, H., Spectrum of *P* and *PcP* in relation to the mantle-core boundary and attenuation in the mantle, *J. Geophys. Res.*, **72**, 559-571, 1967a.
- Kanamori, H., Spectrum of short-period core phases in relation to the attenuation in the mantle, *J. Geophys. Res.*, **72**, 2181-2186, 1967b.
- Kanamori, H., Attenuation of *P* waves in the upper and lower mantle, *Bull. Earthq. Res. Inst.*, **45**, 299-312, 1967c.
- Kovach, R. L., and D. L. Anderson, Attenuation of shear waves in the upper and lower mantle, *Bull. Seism. Soc. Amer.*, **54**, 1855-1864, 1964.
- Kurita, T., Attenuation of long-period *P*-waves and *Q* in the mantle, *J. Phys. Earth*, **14**, 1-14, 1966.
- Kurita, T., Attenuation of short-period *P*-waves and *Q* in the mantle, *J. Phys. Earth*, **16**, 61-78, 1963.
- Mikumo, T., and T. Kurita, *Q* distribution for long-period *P* waves in the mantle, *J. Phys. Earth*, **16**, 11-29, 1968.
- Sato, R., Amplitudes of *PcP* and *PcS* obtained from ray and wave theory solutions and amplitudes near shadow boundary, *J. Phys. Earth*, **17**, 1-12, 1969.
- Sato, R., and A. F. Espinosa, Dissipation in the Earth's mantle and rigidity and viscosity in the Earth's core determined from waves multiply reflected from the mantle-core boundary, *Bull. Seism. Soc. Amer.*, **57**, 829-856, 1967.
- Steinhart, J. S., and R. P. Meyer, Explosion studies of continental structure, *Carnegie Inst. Publ.*, 1961.
- Summer, R. D., Attenuation of earthquake generated *P* waves along the Western Flank of the Andes, *Bull. Seism. Soc. Amer.*, **57**, 173-190, 1967.
- Takano, K., Attenuation of short Period *S* waves in the mantle (in Japanese), *Zisin* **2**, **19**, 246-254, 1966.
- Talwani, M., G. H. Sutton, and J. L. Worzel, A crustal section across the Puerto Rico Trench, *J. Geophys. Res.*, **64**, 1545-1555, 1959.
- Teng, T. L., Body wave and earthquake source studies, Chápt. 7, Attenuation of body waves and *Q* structure of the mantle, *Thesis*, Cal. Inst. Tech., 1966.
- Teng, T. L., Attenuation of body waves and the *Q* structure of the mantle, *J. Geophys. Res.*, **73**, 2195-2208, 1968.

(Received March 12, 1970)