On the Periods of Seismic Waves Observed in Local Earthquakes

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ON THE PERIODS OF SEISMIC WAVES OBSERVED IN LOCAL EARTHQUAKES

BY

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ABSTRACT

The periods of body waves in local earthquakes as observed in Wakayama District were studied.

The periods of impulsive initial waves increase with an increase of travel-distance, but are not related to earthquake magnitude and other effects, within the range of our observation. The increasing rate of the period of the *P*-waves is more rapid than that of the *S*-waves. The reason for this was discussed.

The average periods of the *P*- and *S*-wave groups have a certain fixed value in every station, irrespective of focal distance, shock-magnitude, etc.. They may be attributed to free oscillations of superficial layer.

1. Introduction

The periods of the seismic waves observed at any station are presumably related to the conditions of the earthquake's origin, of the path of wave propagation, and of the observation points: that is, firstly, the emission mechanism of elastic waves at the focus, the earthquake magnitude or the dimension of focal region and the crustal structure near the origin; secondly, the nature of the medium, the structure of the crust and the focal distance, travelled by the earthquake-waves; and thirdly, the underground structure in the vicinity of the observation stations and the frequency response of the seismometer used. Because of these various factors the recorded seismograms should be complicated.

According to many investigations hitherto made, it is generally accepted that the seismic wave periods tend to increase with the increasing travel-distance and an increase in the magnitude of the earthquake, and that the predominant periods depend upon the thickness of the superficial layer near the station.

In 1954 and 1956, precise seismometric observations were carried out in the epicentral area of local shocks in Wakayama District for the purpose of studying the nature of micro-earthquakes (**1**). In this paper the relationship between the seismic wave periods, especially of body waves, obtained from the recorded seismograms in these observations, and the various factors mentioned in the following will be discussed: firstly, the focal distance; secondly, the earthquake magnitude, and thirdly, azimuthal distribution, and so forth.

2. Seismograms

The data used in the present analysis were 200 seismograms of 92 local earthquakes recorded at six stations; Idakiso, Wakanoura, Fuyuno, Nokami, Kainan and Yoro. All of these stations are situated in a metamorphic rock zone. Their positions were shown in the previous paper (**1**).

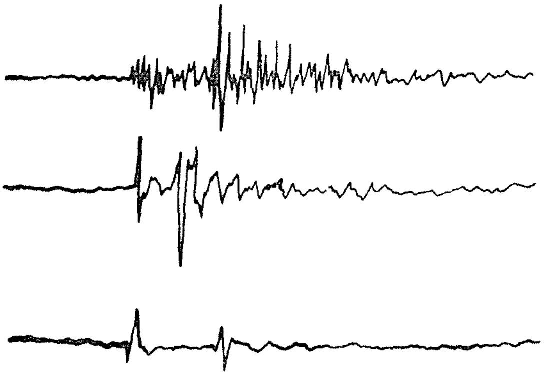


Fig. 1. Examples of seismograms.
(a) A-type, seismogram recorded at Nokami (No. 228)
(b) B-type, seismogram recorded at Wakanoura (No. 228)
(c) B-type, seismogram recorded at Yoro (No. 118)

In these observations, registration was made by direct connection of the electromagnetic seismometers of 0.45 sec in natural period with a galvanometer of 0.40 sec in free period. The paper speed of the oscillogram was about 5 mm/sec, and the time was marked every second on the seismogram by the JJY standard shortwave. The instruments used at all stations were of the same type. It is, therefore, possible to compare the seismograms at the respective stations with each other without taking into consideration the mutual differences of the characteristics of the instruments.

The recorded seismograms can be roughly divided into two classes; A and B, as shown in Fig. 1. In type-A, the initial motions of both *P*- and *S*-waves are followed by fairly regular oscillations similar to sinusoidal waves (i.e., the seismograms recorded at Idakiso and Nokami). In type-B, on the contrary, only the impulsive initial motions of the *P*- and *S*-waves are clearly recorded and are not followed by sinusoidal oscillations. In both cases the recorded times of the first zero of initial motion for both waves were measured, and the average periods for four or five oscillations of the *P*- and *S*-wave groups were read from seismograms of type-B. The periods of *S*-waves were measured only from clear records. Measurements of these periods were made mainly for the records of vertical component.

3. Periods of initial motion of *P*- and *S*-waves

(1) *Correction for recorded period*

It is difficult, in general, to take out the true wave-form of the initial motion of incident seismic waves from the recorded seismograms. Therefore, in order to deduce it, the method of comparing the record calculated from the previously assumed form of incident wave with the real seismogram is usually adopted. Now, both our seismometer and galvanometer are critically damped and their natural periods are nearly equal. Hence, in this paper, the hypothetical record was calculated from the assumed incident wave, according to Kawasumi's method (2) based on Galitzin's results and

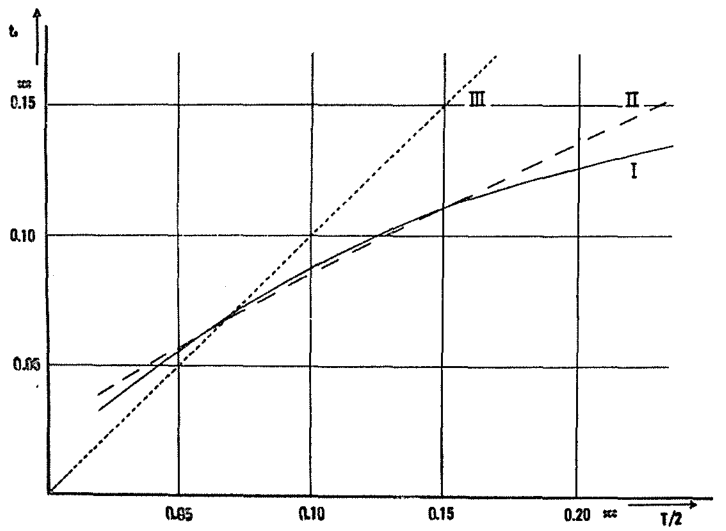


Fig. 2. Theoretical relations between the half-period of initial motion of incident waves and the time of first zero on the ercord, in the case of our seismometer-galvanometer connection.

T: the period of the incident wave
t_0: the time of first zero on record

to Tajime's method (3). Fig. 2 indicates the relation between the half-period of initial motion for the incident wave and the time of the first zero of waves recorded which could be obtained by calculation. The full line (I), broken line (II) and dotted line (III) in the figure correspond to the following three cases respectively assuming the incident wave to be of the form;

i) $f(t)=0$ for $t<0$ and $f(t)=\sin pt$ for $t\geqq 0$;
ii) $f(t)=0$ for $t<0$ and $f(t)=3\sin pt-\sin 3pt$ for $0\leqq pt\leqq\pi$;
iii) the stationary state in the first case.

As can be seen in Fig. 2, the recorded times of the first zero in the cases of i) and ii) make no appreciable difference in the range $0.05<T/2<0.17$ sec or $0.05<t_0<0.12$ sec, whichever form of the incident wave was assumed to be. In the present case, the recorded times of the first zero in 85% of our 163 seismograms, for the shocks of less than 5.0 sec in their $P\sim S$ times, are within the said range. Accordingly, the half-period of initial wave was calculated from the recorded time of the first zero on the seismograms, assuming the form of incident wave to be a disturbance resembling to case ii) owing to the reasons which will be mentioned subsequently. The same procedure was pursued for the S-waves.

Table I shows the frequency distribution of the half-period of initial motion at each station, obtained by the above-described method. It should be further remarked that the measuring error of the period is less than 1/100 sec.

Table I. Frequency distribution of half-period of initial motion.

Station / Period	I		F		W		N		K		Y		Total	
	P	*S*	*P*	*S*	*P*	*S*	*P*	*S*	*P*	*S*	*P*	*S*	*P*	*S*
s s 0.03 — 0.06	7	2	5	2	7	1	3	1	10	0	3	2	35	8
0.06 — 0.08	12	3	6	4	7	1	4	4	2	1	5	1	36	14
0.08 — 0.10	4	4	3	3	1	0	9	5	2	1	3	3	22	16
0.10 — 0.12	1	5	2	1	0	1	6	2	1	0	5	1	15	10
0.12 — 0.15	7	2	1	0	0	0	12	10	2	1	4	4	26	17
0.15 — 0.20	0	1	0	0	1	0	6	7	3	3	1	2	11	13
>0.20	0	0	2	0	2	0	7	2	2	2	5	5	18	9
Total	31	17	19	10	18	3	47	31	22	8	26	18	163	87

(2) *Relation between the half-period of the initial motion and the focal distance*

The relationship between the period of seismic waves and the distance travelled by them, concerning distant earthquakes as well as near quakes, has been investigated by many researchers (**4**, **5**, **6**, **7**, 8, **9**, **10**). There are, however, few investigations about the period of initial motion or of the very near shocks as recorded

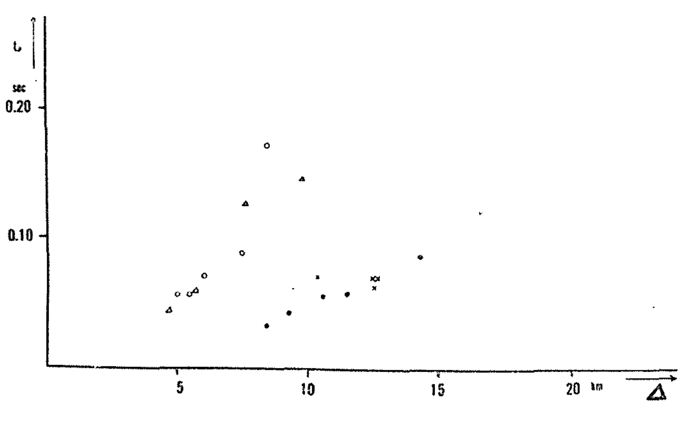


Fig. 3. Relation between the half-period of initial motion of P-wave and the hypocentral distance.

○: No. 218 ●: No. 230 △: No. 250 ×: No. 228

in this observation, and so the problem on the period of initial waves of such local shocks and their travel-distance will be treated here.

Firstly, as to several clearly recorded earthquakes (Nos. 218, 228, 230, 250) (**1**) the half-periods of initial motion of *P*–waves observed at the respective stations were plotted against their hypocentral distances, as shown in Fig. 3. This relation is not uniquely determined for all shocks, but the half-periods tend to increase remarkably with the increasing distance in each shock.

Secondly, Figs. 4 (a)∼(f) show the relationship between the half-periods of initial *P*– and *S*–waves or their period ratios and the travel-distance in all shocks observed at each observatory. For convenience's sake the abscissa indicates the *P*∼*S* times instead of focal distances. The circle (○), dot (●) and cross (×) show the half-period of initial *P*–waves, t_P, that of *S*–waves, t_S, and their period ratio, t_S/t_P, respectively.

Figs. 5, 6 and 7 illustrate the above-described relations for all shocks observed at all stations. That is, in these figures, the preceding relations in the respective observatories shown in Figs. 4 (a)∼(f) are superposed.

Half-periods of both the initial *P*– and *S*–waves are spread over a fairly wide range. On the whole, however, they have a tendency to increase with a certain range, as the *P*∼*S* time or the travel-distance increases, and the increasing rates become less gradually. The rate of this increase for the *P*–wave at the two stations, Kainan and Yoro, seems to be slightly less than that at other stations.

Thirdly, as illustrated in Fig. 7, the ratio of the half-period of the initial *S*–wave to that of the *P*–wave has no fixed value and rather decreases rapidly with increase in travel-distance. This fact may be considered to indicate that the increasing rate relative to the focal distance for the latter is greater than that for the former, as can also be seen by comparing Fig. 5 with Fig. 6. This contrasts sharply with the case in the average periods of the both waves, which will be mentioned subsequently. The similar tendency that some difference is recognizable between the increasing rates for the *P*– and S–waves was observed also by A. E. Jones (**5**) in Hawaii.

(3) *Relation between the periods of the initial motion and the earthquake magnitude*

With a view to investigate the relationship between the half-periods of initial *P*– and *S*–waves and the corresponding earthquake magnitude, the magnitude of shocks was classified into three degrees with the product of recorded maximum amplitude and the *P*∼*S* times at each station, $A_m \times (P{\sim}S)$. This quantity means the maximum amplitude of every earthquake at a certain fixed distance. The triangle, circle and square in Figs. 5 and 6 illustrate the half-periods in the shock of which $A_m \times (P{\sim}S)$ is less than 20, in the range from 20 to 50, and larger than 50, respectively, at the

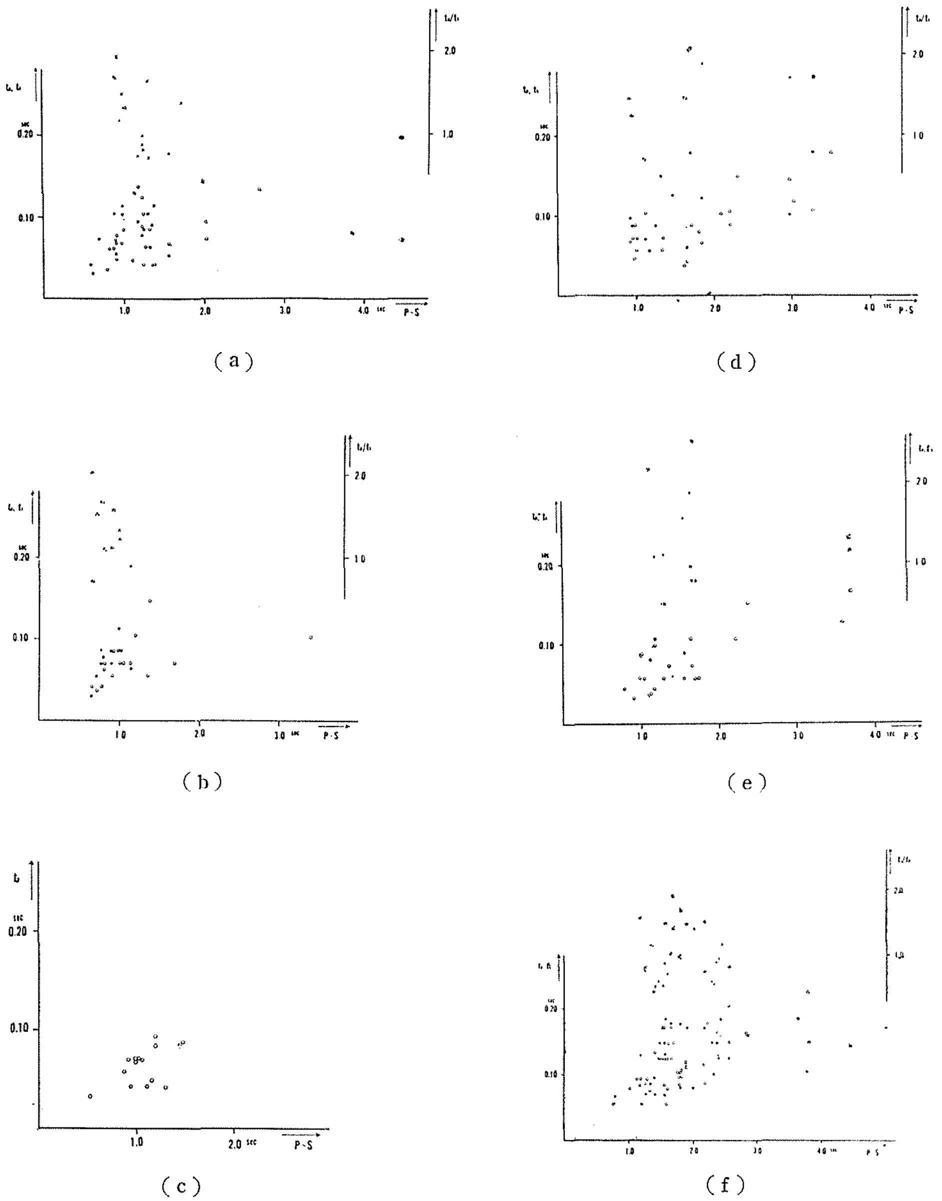


Fig. 4. Relations between the half-periods of initial P- and S-waves or their periods ratio and the focal distance, observed at each station.

○: t_P, half-period of initial P-wave ●: t_S, half-period of initial S-wave
×: t_S/t_P, periods ratio of both waves
(a) Idakiso, (b) Fuyuno, (c) Wakanoura,
(d) Kainan, (e) Yoro, (f) Nokami

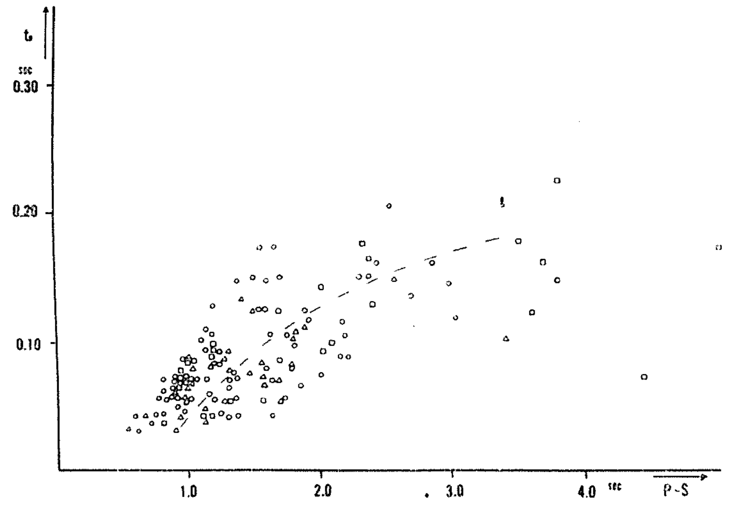


Fig. 5. Relation between the half-period of initial *P*–wave and the focal distance, for all shocks observed at all stations.

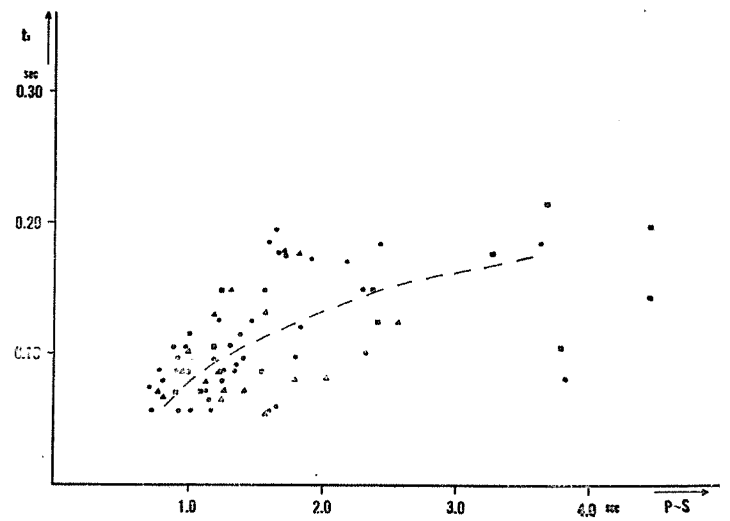


Fig. 6. Relation between the half-period of initial *S*–wave and the focal distance, for all shocks observed at all stations.

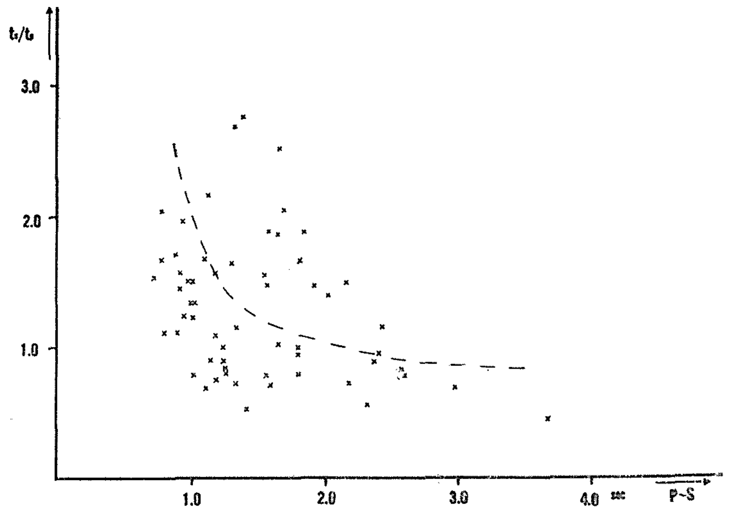


Fig. 7. Relation between the ratio of the half-periods of initial *P*– and *S*–waves and focal distance.

standard station Nokami. The energies released from these earthquakes are roughly estimated (**9**) to be less than 10^{11} ergs, in the order of $10^{11}\sim 10^{12}$ ergs, and more than 10^{12} ergs, respectively. As we can see in these figures, no remarkable difference in the periods concerning the magnitude of the earthquake can be recognized within these ranges observed, though a close correlation between the period and the magnitude in natural earthquakes or the amount of explosives in field experiments has been found out frequently.

(4) *Azimuthal distribution of the period of initial motion*

Several shocks with the already determined hypocentre (**1**) were selected in order to study the azimuthal distribution of the periods of initial motion. In each earthquake, the angle, θ, between one of nodal lines and the line connecting the epicentre and each station was measured counter-clockwise. However, the half-period of the initial *P*– and *S*–waves observed at each station seems to be unrelated to $\cos\theta$, in all the shocks. Furthermore, the relationship between the half-period and the ratio of maximum amplitude of the *S*–wave group, A_S, to that of the *P*–wave group, A_P, at each station was examined for all the observed shocks. A_S/A_P or $A_S/A_P(P\sim S)$ is considered to be proportional to $\cot 2\theta$ (**11**). The value of A_S/A_P was corrected by magnification response of our seismometer for the respective periods of the *P*– and *S*–waves. The half-period of the *P*–waves seems to decrease slightly with an increase in $A_S/A_P(P\sim S)$. In none of the afore-mentioned cases, however, any clear evidence touching azimuthal distribution of those periods depending upon an earthquake generation mechanism (**12**) was detected. Moreover, any difference in period, in connection with the 'push' and 'pull' in the direction of the initial motion, or relating to the travel-time anomalies previously reported, could not be ascertained.

4. Average periods of the *P*– and *S*–wave groups

As mentioned previously, the average periods of the *P*– and *S*–wave groups were measured from the seismograms of type–A recorded at Idakiso and Nokami. When the incident wave is in stationary oscillation as in the form i), stated in the preceding paragraph, the period of the recorded wave is equal to that of the incident one, as shown in Fig. 2. Therefore, the average period can easily be obtained from a seismogram without correction. The frequency distribution of the average periods thus obtained is tabulated in Table II.

Table II. Frequency distribution of average period of *P*– and *S*– wave groups.

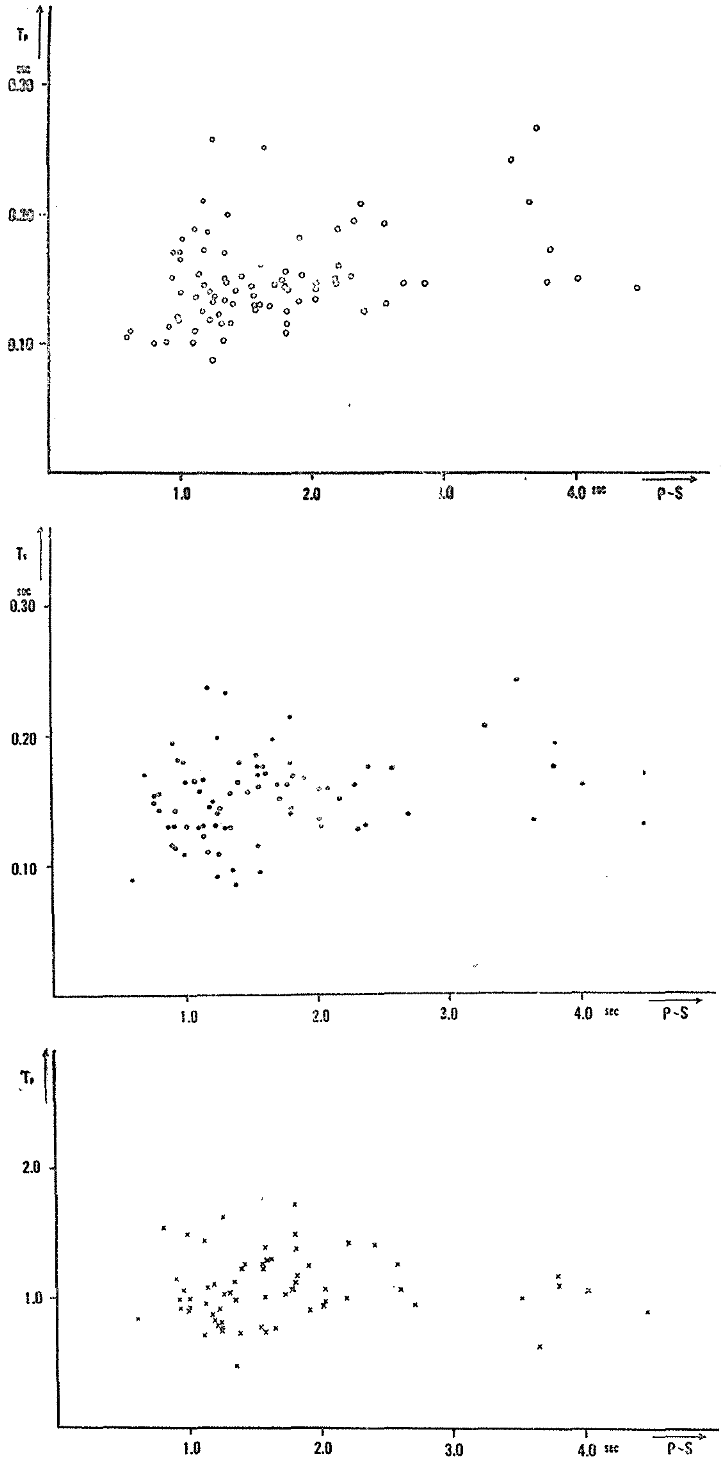
Period \ Station	I		N		Y		Total	
	P	*S*	*P*	*S*	*P*	*S*	*P*	*S*
s <0.10	2	4	0	0	0	0	2	0
0.10 – 0.13	13	8	7	2	2	1	22	11
0.13 – 0.15	11	11	18	8	4	3	33	22
0.15 – 0.17	5	7	7	14	1	3	13	24
0.17 – 0.20	1	2	6	12	4	3	11	17
>0.20	0	0	3	1	5	8	8	9
Total	32	32	41	37	16	18	89	83

The average periods scatter less widely than in the case of half-period of initial motion. When plotted, they have approximately fixed values irrespective of the $P \sim S$ times or travel-distances and earthquake magnitudes, as shown in Fig. 8, and differ from the cases of half-periods. There is a slight difference in the mean values of average periods observed at two of the stations.

5. Discussion

(1) *Periods of the initial P– and S–waves*

The cause of increase in the period of earthquake waves with an increase in travel-distance has been considered to be mainly attributable to the viscosity of the medium for wave propagation, since the other effects, such as scattering due to heterogeneity in the medium or superposition (**13**) of the various waves, seems to be comparatively little. The attenuation factor of amplitude in relation to travel-distance for the waves propagating in visco-elastic medium (for example, the Voigt model) is



(a) Average period of *P*–wave group.

(b) Average period of *S*–wave group.

(c) Ratio of the average periods of *P*– and *S*–wave groups.

Fig. 8. Relations between the average periods of *P*– and *S*–wave groups or their period ratio and the focal distance.

proportional to the square of the wave frequency (**14, 15, 16**). Hence, the waves with longer periods survive as the travel-distance increases, because the wave components of higher frequency decay more rapidly. K. Sezawa studied theoretically the problem of wave propagation in the visco-elastic medium of Voigt solid when several types of disturbance were generated at the origin (**17**). From these results, B. Gutenberg established and expressed the relation between the period of seismic wave and the focal distance in the formula (**18**);

$$T = \sqrt{T_0^2 + \frac{\alpha}{V^3}\Delta}, \qquad (1)$$

where T and T_0 are the observed period and the original period emitted from the focus, respectively, Δ is the hypocentral distance, V the velocity of wave propagation and α a constant.

Next, we try to explain, from the standpoint of the visco-elastic medium, the fact that the half-periods of the P-wave increase more rapidly than that of the S-wave as to the focal distance. As one of the simplest model of visco-elastic medium, the Voigt model was adopted in the present case. Our observed results shown in Figs. 5 and 6 indicate that the respective mean curves are in fairly good accordance with the above Gutenberg's formula. Hence, the relationship between the half-periods and the travel-distance of the P- and S-waves can be represented by the following formulae:

$$\left.\begin{aligned} T_P &= \sqrt{T_{P0}^2 + \frac{\alpha}{V_P^3}\Delta} \quad \text{for } P\text{-wave}, \\ T_S &= \sqrt{T_{S0}^2 + \frac{\beta}{V_S^3}\Delta} \quad \text{for } S\text{-wave}, \end{aligned}\right\} \qquad (2)$$

and

$$\alpha = K\frac{\lambda' + 2\mu'}{2\rho}, \quad \beta = K\frac{\mu'}{2\rho}, \qquad (3)$$

where T_{P0} and T_{S0} are the original periods and V_P and V_S are the propagation velocities, for the P- and S-waves, respectively. λ' and μ' are the viscosity coefficients analogous to Lamé's elastic constants λ and μ. μ' and $\lambda' + \frac{2}{3}\mu'$ are the so-called shear viscosity and dilatational viscosity respectively, ρ is the density and K is a numerical constant determined from the form of disturbance at the origin.

From (2), we get

$$\alpha = V_P^3 \frac{dT_P^2}{d\Delta}, \quad \beta = V_S^3 \frac{dT_S^2}{d\Delta}, \qquad (4)$$

and therefore

$$\frac{\alpha}{\beta} = \left(\frac{V_P}{V_S}\right)^3 \cdot \frac{dT_{P2}^2}{d\Delta} \Big/ \frac{dT_S^2}{d\Delta}. \qquad (5)$$

Also from (3), we have

$$\frac{\alpha}{\beta}=\frac{\lambda'}{\mu'}+2. \qquad (6)$$

The coefficients α and β can be determined from the observed results by the method of least squares, but were roughly estimated in the present treatment by means of a simple procedure. In this calculation the following values of V_P, V_S and Omori's constant k were used, which have already been determined on the crustal structure in Wakayama District (**1**).

$V_P=4.3$ km/sec, $V_S=2.5$ km/sec and $k=6.0$ in the surface layer,
$V_P=5.5$ km/sec, $V_S=3.2$ km/sec and $k=7.6$ in the second layer.

The observed period contain all those in the cases that the focus is both in the upper layer and in the second layer. Considering two extreme cases in which the crust is only of the uniform medium of 4.3 km/sec or 5.5 km/sec in propagation velocity, the real values of α and β should be calculated as the values between these two. Taking these circumstances into consideration, we find

$$7.2\times10^{9}<\alpha<1.2\times10^{10} \quad \text{c.g.s.},$$
$$1.0\times10^{9}<\beta<1.7\times10^{9} \quad \text{c.g.s.}.$$

The calculated value of α agrees in its order of magnitude with the previously obtained ones by Gutenberg (**18**), Asada and Suzuki (**8**) and others. No result as to the value of β for S–waves can be traced.

Taking the following values as the mean density (ρ) and rigidity (μ) in the upper layer of metamorphic rocks and in the second layer of granite, namely:

$\rho=2.64$, $\mu=2.30\times10^{11}$ c.g.s. for the upper layer,
$\rho=2.67$, $\mu=2.87\times10^{11}$ c.g.s. for the second layer,

λ' and μ' can be calculated by Eq. (3) as follows, assuming the value of K to be $8\sim11$ (**18**):

$$2.5\times10^{9}<\lambda'<6.4\times10^{9} \quad \text{c.g.s.},$$
$$4.8\times10^{8}<\mu'<1.2\times10^{9} \quad \text{c.g.s.},$$
$$1.7\times10^{-3}<\mu'/\mu<5.2\times10^{-3}.$$

The viscosity coefficients thus obtained are in approximate accordance with the values determined by Jeffreys (**19**), Sezawa and Kanai (**20**) and others, and the value of μ'/μ agrees in its order both with laboratory data on various rocks and with field experiments (**21**).

Moreover, the following results are obtained from Eqs. (5) and (6),

$$\frac{\alpha}{\beta}=7.36, \quad \frac{\lambda'}{\mu'}=5.36.$$

The ratio of dilatational viscosity to shear viscosity, $\lambda'/\mu'+2/3$, is calculated to be about 6. Few values of λ'/μ' have been determined by laboratory experiments. K. Iida (**22**) published his experimental result of $\lambda'/\mu'=6.43$ for pumice at zero water content. It is also reported by another investigator (**23**) that λ'/μ' is nearly 36 for a kind of metal (nickel). Therefore, our calculated value is not an unexpected one.

As to the ratio of the period of the initial *P*-wave to that of the *S*-wave, K. Kanai discussed theoretically (**24**) in the case of a perfect elastic medium, and showed that the latter is about twice the former, taking into account the correction due to a constant of seismograph (**25**). On the other hand, this ratio becomes nearly 1.8, which can be obtained from Jeffreys' theoretical calculation (**26**). But these results indicate that the ratio of the above two periods do not change as to the travel-distance.

In order to explain the results illustrated in Fig. 7, the formulae (2) in the case of visco-elastic medium can be cited and rewritten in the form:

$$\frac{T_S}{T_P}=\sqrt{\frac{T_{S0}^2+\beta\Delta/V_S^3}{T_{P0}^2+\alpha\Delta/V_P^3}}\,. \qquad (2')$$

As the hypocentral distance Δ increases, T_S/T_P changes from T_{S0}/T_{P0} to $\sqrt{\frac{\beta}{\alpha}\left(\frac{V_P}{V_S}\right)^3}$. The original periods T_{P0} and T_{S0} cannot easily be determined by extrapolation of the results in Figs. 5 and 6. The observed minimum half-period of the *P*-wave is 0.03 sec, and that of the *S*-wave is 0.06 sec. If these values correspond to the original half-periods approximately, T_S/T_P decreases from 2 to 0.8 with an increase in focal distance. The results shown in Fig. 7 can be known from these points of view.

Comparing, then, the increasing rates of the periods of the initial *P*- and *S*-waves with each other, we find theoretically the following relations from the formulae (5) and (6):

$$\frac{dT_P^2}{d\Delta}\gtreqless\frac{dT_S^2}{d\Delta}, \qquad \text{according as} \quad \lambda'/\mu'\gtreqless 3.1\,,$$

assuming Poisson's ratio to be 0.25. That is, it is the case of $\lambda'/\mu'>3.1$ when the increasing rate of the period of *P*-wave with increasing distance is larger than that of the *S*-wave. The presently obtained value of λ'/μ' is within this range, and may be considered to be reasonable one from the above consideration about the experimental data and from the aspect of the attenuation factor of amplitude obtained by some researchers.

Consequently, our observed results are explainable, if we consider the dilatational viscosity in the medium of wave propagation to be about six times as large as the shear viscosity.

However, the above relation cannot hold without any modification for more distant earthquakes. Otherwise, the period of the P-wave come to be longer than that of the S-wave at a long distance. This is not generally acceptable. This problem may probably be solved, if the value of λ'/μ' at depth is smaller than at the shallower crust and $\lambda'/\mu' < 3.1$ on the ground that the initial wave propagates more deeply in the crust to a distant station.

On the other hand, N. Ricker stated in his wavelet theory on the visco-elastic medium that the wavelet breadth is proportional to the square root of the travel time of its centre (**28, 29**). He confirmed his theory by field experiments and calculated the viscosity of shale from the relation between the wavelet breadth, b, and travel time, t; $b^2/t=\text{const.}$ (**30**). This can be rewritten into the formula relating the period, T, with the travel-distance, Δ, as follows:

$$T = 4\sqrt{\frac{\eta}{\rho V^3}\Delta}, \qquad (7)$$

where η is the viscosity of the medium for wave propagation, and can be expressed by $\eta=\frac{3}{4}(\lambda'+2\mu')$ for the P-wave, and by $\eta=\mu'$ for the S-wave, in terms of λ' and μ'. In this case the form of disturbance at the origin was assumed to be of a rectangular type. If Eq. (7) is applied to our data, η is estimated to be of the order of 10^9 c.g.s. for propagation of the P-wave and 10^8 c.g.s. for that of the S-wave, adopting the previously mentioned crustal structure. Here, however, the ratio of the period of the initial P-wave to that of the S-wave is constant, regardless of travel-distance. Therefore, Ricker's formula cannot give a complete explanation to our observed results.

Consequently, we may consider that Gutenberg's formula can most suitably express our obtained data.

(2) *Average period*

The average periods of the P- and S-wave groups may be considered to represent the approximate predominant periods in their respective groups. The fact that there is a different predominant period in every station has been approved from many observations since some ten years ago (**31**). The predominant period was attributed (**32**) to the period of free oscillations of superficial layer, which is secondarily excited by multiple reflections in that layer in the case of incidence of impulsive initial shocks, and the possibility of free oscillations of the surface layer was also studied theoretically (**33, 34**). According to those investigations, free oscillations have a periodicity when the length of disturbance in the impulsive wave is longer as compared with the thickness of surface layer. Hence, the predominant period depends upon the thickness of the surface layer, and the fact that it has a certain definite value scarcely relating to the travel-distance of seismic waves, the earthquake magnitude etc., has been ascertained (**35, 36**). This agrees with our present findings.

Judging from seismograms recorded at Wakanoura, Kainan and especially at Yoro, the initial *P*– and *S*–waves seem to be fairly impulsive, as shown in Fig. 1. Therefore, the sinusoidal oscillations following the initial motion recorded at Idakiso and Nokami may be considered to be free oscillations of the superficial layer, composed of weathered schist, excited by the initial motion. The relationship between the surface thickness and the period of body waves has been estimated to various forms, not only in observations of near earthquakes (**6**, **37**, **38**, **39**) but also in field experiments (**40**, **41**). According to the results of these researches, the thickness of the superficial layer near the above two stations is presumed to be less than 100 m, even if the velocity in this layer is assumed to be nearly 2 km/sec. The delay in travel-time due to the existence of this layer is within 0.03 sec and has no effect upon the travel-time anomalies discussed in the previous paper.

6. Concluding remarks

The following results were obtained from a period analysis for the body waves of local earthquakes observed in Wakayama District.

i) The initial waves are considered to be of an impulsive character. Their periods increase with increasing distance travelled by the waves, but they are related neither to the magnitude of earthquake nor to other effects within the range of our observation. No clear evidence on azimuthal distribution for the periods was recognizable.

ii) The cause for the increase of periods of initial waves with an increase in focal distance is attributable to the viscosity of the medium for wave propagation. The relation between the periods of initial *P*– and *S*–waves and the travel-distance can approximately be expressed by Gutenberg's formula, but the increasing rate of the former is more rapid than that of the latter. This may be by reason that the dilatational viscosity of the medium is considerably larger than the shear viscosity. Our present results can be explained by the hypothesis that the dilatational and shear viscosities have the order of $10^9 \sim 10^{10}$ and $10^8 \sim 10^9$ in c.g.s. units, respectively, and the former is about six times the latter.

iii) The average periods of the *P*– and *S*–wave groups have a certain fixed value irrespective of the focal distance, the earthquake magnitude and other effects, within the scope of this observation. Namely, they may be closely related to the conditions of the observation station.

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