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14. *Correlational Study of Near Earthquake Waves.*

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1. Introduction

This paper is the second report on the study of near earthquake waves by a correlational method. In the first report (1958)[1], we gave the result of a preliminary investigation which was made by means of a temporary tripartite net at Kamakura in southern Kanto. This tripartite net was a very small one having an average span of about 100 m, and the seismometers were placed on sandstone which is not very hard. The length of span was appropriate for waves having a velocity less than 3 km per sec, but too small for waves of higher velocity. In the present study, the length of span was elongated to about 500 m, and the temporary tripartite net was set up on granitic rocks exposed near Mt. Tukuba in eastern Kanto.

By the preliminary investigation at Kamakura, it was found that spatially random waves coexist with high velocity waves having the direction of propagation pointing to the origin of the earthquake and also with low velocity waves which have different directions not dependent upon the origin and sometimes are not plane waves within the network. Earthquake waves were divided into consecutive portions each of 2 to 3 seconds time interval for the analysis, and it was found that the fraction of average power (average power here means time average of the square of the amplitude.) carried by the regular plane waves is largest in the initial portion, that it gradually decreases with the time, again becomes large at the arrival of the S wave, and later tends to become almost negligible.

The knowledge of this kind of facts is essential, if one wants to make use of the whole seismogram instead of few particular phases for purposes such as locating small earthquakes and estimating earthquake energy. Also to study them for different frequency ranges and at vari-

1) K. AKI, M. TSUJIURA, M. HORI and K. GOTO, *Bull. Earthq. Res. Inst.*, **36** (1958), 71-98.

ous places will help the understanding of complicated phenomena such as the scattering of waves due to local heterogeneity and local vibrations of the ground induced by the incidence of primary waves coming from the origin of earthquake.

The present writers are especially interested in locating very small earthquakes which have the frequency range of vibration from 1 cps to 50 cps. As well known the frequency of occurrence of earthquakes increases with decreasing magnitude according to Gutenberg-Richter's formula (1949)[2], or an equivalent one called Ishimoto-Iida's formula in Japan. It was revealed by Asada (1958)[3] that very small earthquakes are recorded at a frequency as high as 200 per day at Mt. Tukuba in eastern Kanto during a normal period in accordance with the above formulas. Therefore if we can determine the hypocentres of these earthquakes, we shall have more reliable and continuous information about the seismic activity at a particular place under the ground. This information may be important in connection with the occurrence of larger earthquakes.

For the purpose of locating small earthquakes, the use of a single station having a small tripartite net of seismometers has greater advantages than that of separate stations each having a single seismometer: first because a considerable number of earthquakes recorded at one station are not recorded at other stations even if the sensitivity of the seismometers is very high; secondly, it is not easy in any sense to maintain many separate stations having highly sensitive seismometers with amplifiers and accurate clocks; and thirdly, it will take much more time at the central station to gather and process the data in the separate station system than in the single station system. But the accuracy of determination of hypocentre is certainly better for separate stations than for a single tripartite station using the times of the initial motions of P and S waves. It is intended, therefore, to make use of the whole seismogram instead of initial motions of a few phases.

The details of the method used in the present study were given in previous papers (1957, 1958)[4,5], so only a brief description of it will be given here in comparison with other methods that can be applied for the same purpose.

2) B. GUTENBERG and C. F. RICHTER, "*Seismicity of the Earth*", Princeton, N. J., 1949.

3) T. ASADA, *Jour. Phys. Earth.*, **5** (1957), 83-113.

4) K. AKI, *Bull. Earthq. Res. Inst.*, **35** (1957), 415-456.

5) K. AKI *et al.*, *loc. cit.*, 1).

If the correlation function of two signals, coming from two seismometers placed at a certain distance, is computed, shifting one with respect to the other in time, a curve will be obtained in which the shift time giving the maximum correlation is equal to r/V, where r is the distance between seismometers and V is the apparent velocity of the wave through the two points. This method will give a very accurate determination of local velocity, if the wave in question has a flat frequency spectrum and a single definite velocity which is independent of frequency. In this case, the correlation curve will take the form of a δ-function. But, if the frequency spectrum of the wave is not flat, not only will the sharpness of the peak in the correlation curve be lost, but also secondary peaks will appear at different shift times depending on the shape of the spectrum. Further, if the wave is composed of more than two partial waves of different velocities which may depend on frequency, it will be very difficult to interpret the obtained correlation curve in order to learn the value of velocity and the fraction of average power for each component wave.

Therefore it is at least desirable to eliminate approximately the influence of frequency spectrum upon the final correlation curve. This is possible if one observes the wave in a limited frequency range by filtering it with a band pass filter. Then the correlation coefficient may be computed (the use of the " coefficient " removes the amplitude effect) shifting the seismometer signal in space instead of time, in other words, the autocorrelation of the wave in space, particularly for a limited frequency range, may be computed instead of the crosscorrelation of vibrations at two points. If the band width of the frequency range is taken very small, a correlation coefficient curve of the following form for the wave having a single velocity V will be obtained.

$$\rho(f, r)=\cos\left(\frac{fr}{V}\right) \tag{1}$$

where f is the frequency of centre in the band.

This type of correlation curve is no longer influenced by the spectral distribution of wave, besides it has a greater advantage that we can easily find its theoretical form for various models of complicated waves.

For instance, if the wave in question is a dispersive wave, it can be shown that Eq. 1 holds also in this case without any modification except the substitution of the function $V(f)$ of frequency f for the constant velocity V. If the wave is composed of two partial waves having different velocities V_1 and V_2, and fractions of average power P_1 and P_2, then the correlation curve will take the following form

$$\rho(f, r) = \frac{P_1 \cos\left(\frac{fr}{V_1}\right) + P_2 \cos\left(\frac{fr}{V_2}\right)}{P_1 + P_2} \tag{2}$$

under the condition that the partial waves are independent of each other. This formula is useful to separate waves propagating in different directions through the same place at the same time, and can be generalized to the case of an arbitrary number of partial waves. On the other hand, if the wave in question is purely random in space, we may express the correlation curve in the following form,

$$\left.\begin{aligned} \rho(f, r) = \frac{\delta(fr)}{\delta(0)} &= 1\,, \qquad fr = 0 \\ &= 0\,, \qquad \neq 0 \end{aligned}\right\} \tag{3}$$

And if a regular wave of velocity V is mixed with this type of random noise we may express this in the following formula

$$\rho(f, r) = \frac{P_0 \frac{\delta(fr)}{\delta(0)} + P_1 \cos\left(\frac{fr}{V_1}\right)}{P_0 + P_1} \tag{4}$$

where P_0 and P_1 are the fractions of average power carried by the random noise and the regular wave respectively.

The above method of analysis, however, requires a great number of seismometers, amplifiers, and channels in the recorder which should have a recording time of more than one minute, when the correlation curve is to be plotted against spacial distance, r. Therefore the curve is plotted against the pass frequency, f, of the filter in our investigation. By this conversion the least number of seismometers required for the analysis of earthquake waves is reduced to three.

2. Observation and analysis

We observed local earthquakes at the foot of Mt. Tukuba which is located about at the centre of eastern Kanto. The reason why we chose this place as our field of experiment is as follows: first, this is one of the places where the noise level is the lowest throughout Kanto;

secondly, we can find there granitic rocks exposed on the surface; thirdly, there is a permanent observatory, which belongs to the Earthquake Research Institute, Tokyo University, provided with various kinds of modern equipment and power supplies.

We selected six sites for seismometer locations around the observatory. Fig. 1 shows the map of sites and the distances among them measured by a precise triangulation. At each site, a house for the seismometer made of concrete blocks is cemented to exposed granitic rock which is cut and smoothed to fit the house.

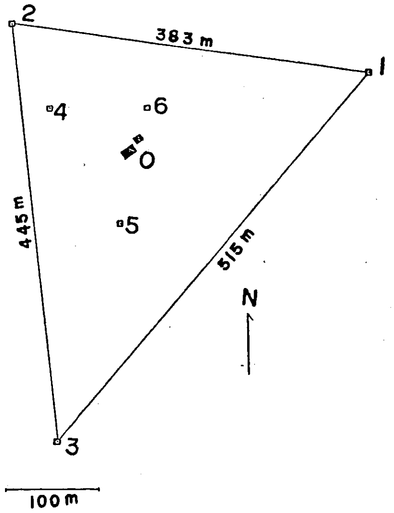

Fig. 1. Map of seismometer location. No. 6 is the Tukuba Observatory seismometer vault and O is its office building.

Seismometers used are moving coil type transducers sensitive to vertical motion, and have a free frequency of 3 cps. We used in the present observation, as in the preliminary one at Kamakura, the U.H.F. radio telerecording seismograph, RTS–II, made by S. Miyamura and M. Tsujiura (1957)[6] of our Institute. This apparatus transmitted the seismometer signals from Mt. Tukuba to the receiving station which was set up in the ninth story of a house of the medical department of Tokyo University at Hongo, Tokyo. The use of this radio telerecorder is not essential for the purpose of the present study, but remarkably increased the efficiency of study by making our field observation a kind of laboratory experiment.

We used two sets of RTS–II in the present observation. In each set, three seismometer signals are amplified, and modulate the frequency of three carriers of audiofrequency, 800, 2670 and 8900 cps respectively. The modulated signals are mixed together and again modulate the frequency of a carrier of 417.7 Mc/s, which is sent from the transmitter as radio waves to the receiver. At the receiving station the received signals are demodulated with respect to the carrier of 417.7 Mc/s and recorded on a magnetic tape, being still modulated by the signals of seismic frequency. Since our magnetic recorder has two channels, we

6) S. MIYAMURA and M. TSUJIURA, *Bull. Earthq. Res. Inst.*, **35** (1957), 381-394.

are able to make a simultaneous recording of six seismometer signals with the two sets of RTS-II.

The recording was made by using a special reel which was designed to enclose an endless tape. The tape enclosed in the reel works as long as 7 minutes for a speed of 7.5 inches per sec.

During the period of observation, the tape recorder is continuously in action waiting for an earthquake, and when one takes place, it is made to stop by means of a delayed trigger circuit after recording of waves.

The record of earthquake motions stored in the endless tape is reproduced, demodulated with respect to the carriers of audiofrequency, and analysed according to the method stated in the introduction by means of a specifically designed computer. This computer consists of two identical filters, a correlation computer, a gate circuit, and a checking circuit. The filter used here is of phase shift type and its resonance frequency covers from 2 cps to 18 cps. The function of this filter is described in the paper of K. Aki (1957)[7] referred to before, with the explanation of the correlation computer in which decatron tubes are used as an integrator and an indicator. The gate circuit controls the computer to make a separate computation for each consecutive portion into which the record of the earthquake is divided. The time interval of the portion is taken as 2.5 seconds which was found to be appropriate for the present purpose in the preliminary experiment.

A checking circuit is provided to check whether the computer is in order. This circuit generates two checking signals, each of which is a series of pulses recurring every 2.5 seconds, one of which is shifted 1/7.5 sec from the other. If we put these two signals, through a pair of identical filters having the resonance frequency f and sufficiently high Q value, into the correlation computer and observe the variation of correlation coefficient by changing the frequency of filters, we shall get a curve of the following form,

$$\rho(f)=\cos\left(2\pi\frac{f}{7.5}\right).$$

Each time before the computation this check was made by the comparison of the result with the above theoretical value; the deviation in terms of the correlation coefficient was usually less than 10%. An example of the correlation curve for the checking signals is shown in

7) K. AKI, *loc. cit.*, 5).

Fig. 2.

In the present investigation, if the phase characteristic of one channel differs from the phase characteristics of the others the result will be seriously distorted. Therefore the phase characteristics of seismometers, amplifiers, modulators and de-modulators were carefully examined, and it was found that the phase differences among the responses of different channels to an input of sinusoidal form can be reduced to about 10 degrees over the frequency range to be studied in the present experiment[8]. This magnitude of difference is small enough for the present purpose.

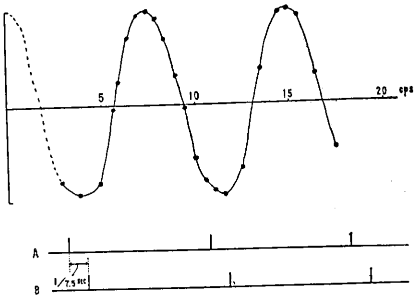

Fig. 2. A correlation curves for the checking signal.

3. Discrimination of plane waves from irregular noises

As far as the noises in earthquake waves are purely random in space, the discrimination of a plane wave from the noises is not difficult. As stated in the introduction, the correlation curve takes the cosine form for a regular plane wave and the form of a δ-function for a purely random noise. One can easily learn the velocity and fraction of average power for the plane wave when it coexists with the purely random noise in the earthquake waves.

But, sometimes we find a correlation curve of intermediate nature between the above two, suggesting the existence of a group of waves which are neither purely random nor perfectly regular. In some of these cases, the form of the correlation curve appears to involve a cosine form and lead us to make a misinterpretation that a plane wave exists in that wave. But, one can decide whether the plane wave really exists in the wave or not, by comparing the values of velocity obtained from every correlation curve for different pairs of seismometers. For instance, if lines are drawn on a map parallel to the lines connecting seismometer pairs and intersect at a point 0, and a point is marked on each line at the distance equal to the value of the apparent velocity along the line from the point 0, then all of the points should lie on a straight line in the case of the plane wave. And this straight line is parallel

8) M. TSUJIURA and S. MIYAMURA, *Bull. Earthq. Res. Inst.*, **37** (1959), 193-206.

to the wave front and the distance between this line and the point 0 is equal to the value of true velocity of the plane wave. This is a simple graphical criterion for the regularity of waves as well as one of the simplest methods of obtaining the direction of propagation and the velocity of the wave.

As a typical example, two sets of correlation curves are shown in Fig. 3a. The curves on the left are obtained from the wave portion for 2.5 sec after the initial motion, and those on the right are from the wave portion including the S wave, both of an earthquake wave recorded on Aug. 22, 1958. The pair of numbers shown on the upper left of each curve indicates the pair of seismometers which are located at the places having the same numbers in Fig. 1 and between which the correlation curve is computed.

In any of these curves a cosine form is discernible though it is more or less distorted by the presence of irregular noise. The dotted curves in the same figure are drawn in order to show how we find the

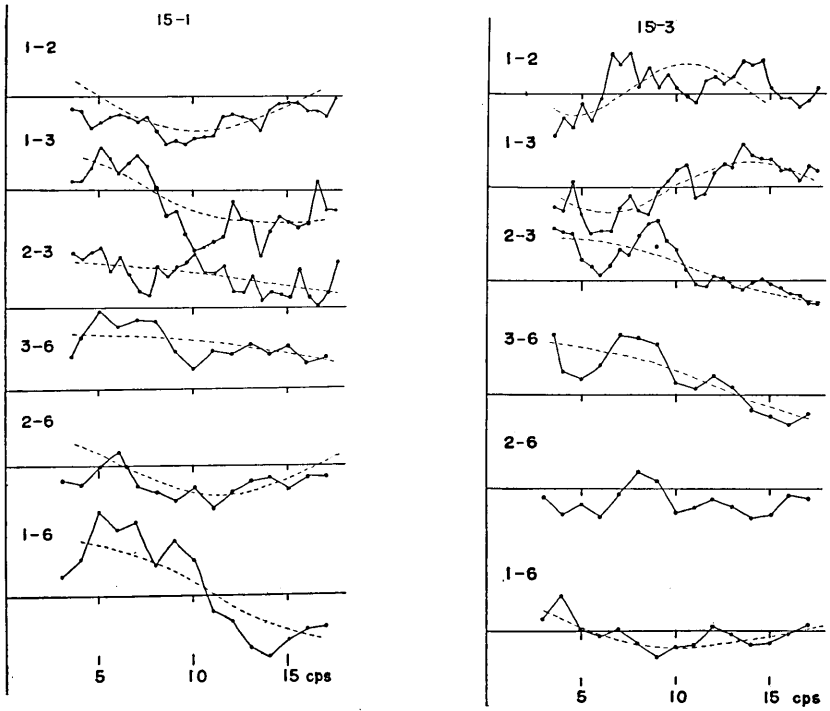

Fig. 3a. Correlation curves for the P and S wave portions of the earthquake No. 15 recorded on Aug. 22, 07 h 09 m.

smoothed curves of nearly cosine form in the original. To check whether these cosine forms are truly from a plane wave, the value of apparent velocity obtained from each of the smoothed curves is plotted in Fig. 3b in the way described above. The points lie fairly well on a straight line, and we can be sure that a plane wave exists in this wave portion. In this case, the velocity of the wave portion after the P phase is 7.9 km/s and that of the portion including the S phase is 4.1 km/s. Their directions of propagation are slightly different.

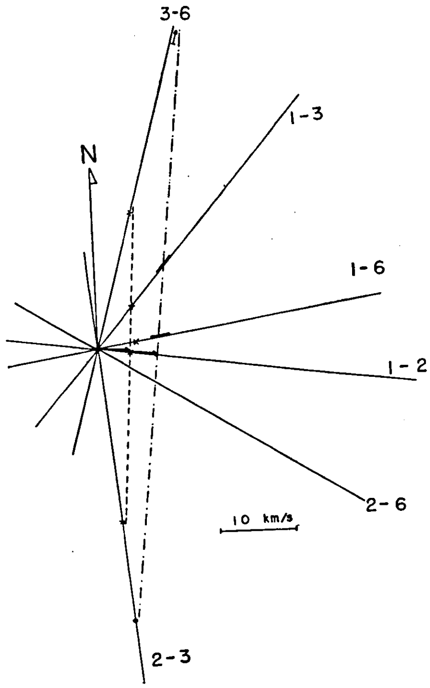

Fig. 3b. Graph to check the regularity of waves.

In this way, plane waves are sought in every portion, having a 2.5 sec interval, of 18 earthquake waves recorded at Mt. Tukuba. These earthquakes are local ones and their P–S times are in the range from about 5 sec to 20 sec. We found well established plane waves in the first wave portion of all the earthquakes but one. In this exceptional one, the direction of initial motion at one place is found to be reverse to those at the other places, therefore an accidental noise at the place can be probably considered as the cause of irregularity of that wave portion. The fraction of average power carried by these plane waves in the initial wave portions varies considerably among earthquakes, it was 80% in the largest case and about 40% on the average. This value of the fraction is not strictly independent of the frequency, and usually large in low frequencies and smaller in higher frequencies. The value given above is obtained by a rough estimation averaging with respect to the frequency.

In the wave portion including the S phase, it is less easy to find plane waves than in the case of the initial portion, and the fraction of average power carried by the plane waves is smaller. We found plane waves in the S portions of 14 earthquakes and failed in 4 earthquakes. The fraction of average power carried by these plane waves was 60%

in the largest case, and roughly 20 to 30% in most cases.

The 18 earthquakes studied contain 76 wave portions between the initial one and the S wave portion. The time interval of each portion is 2.5 sec. We found plane waves in 47 portions out of these 76 portions. Therefore the probability of existence of plane waves in these portions is 0.62 which is less than 0.78 in the case of the S wave portion. The fraction of average power carried by plane waves sometimes is very large, especially in the portions next to the initial ones. In most of the earthquakes, the correlation curves for the initial portion is quite similar to those for the next one. So these successive two portions are usually nearly the same in the fraction of average power, direction of propagation and value of velocity of plane waves which they involve.

We studied 97 wave portions which came after the S wave portions. From these portions we found plane waves in 29 portions. The probability of existence of plane waves is about 0.30 which is much less than

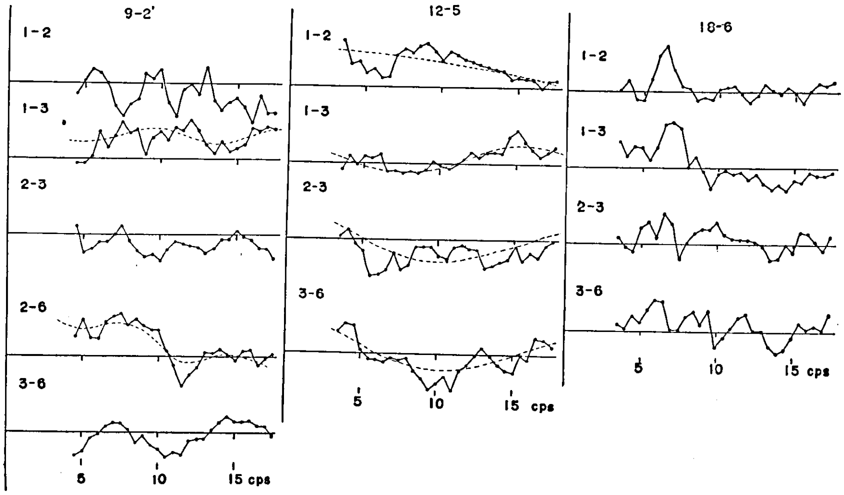

Fig. 4. Typical examples of correlation curves for the coda wave portions.

that for the portions between P and S. The fraction of average power carried by them is very small and never exceeds 30%. Typical examples of the correlation curves for these portions are shown in Fig. 4. The set of curves on the left in the figure refers to low velocity waves, and the curves in the middle refer to high velocity waves. In both cases, the amplitude of cosine forms discernible in these curves is small, that is, the fraction average power carried by the plane wave involved in the portions is small.

The wave portions where we could not find any plane wave are not necessarily purely random in space. As a matter of fact, there were no wave portions which can be concluded as purely random from the point of their correlation curves. If a wave portion is purely random in space, every correlation curve for that portion should take the form of a δ-function. But the actual curves are like that shown on the right of Fig. 4. The value of correlation is sometimes significantly different from zero. It is impossible to interprete these curves unless we have a detailed theoretical model of the wave which is difficult to establish.

4. Comparison of the propagation of the initial wave portion having the duration of 2.5 sec with that of the initial motion

One of the principal purpose of the present study is to learn whether it is possible to increase the accuracy of direction determination of wave propagation by the use of wave portions of longer duration instead of the use of initial motion only. First we shall examine the propagation of the plane waves found in the initial portions. Unfortunately, most of the earthquakes we have recorded are too small to be recorded by the routine network of seismological stations which is operated by the Meteorological Agency in Japan. So, we have no information from this source about the true hypocentres of these earthquakes with one exception, for which an estimation is given. As an alternative procedure to study the problem we can take the following method[9].

As we have earthquake records stored in magnetic tapes, it is very easy to reproduce them on a paper which runs at a speed as high as 10 cm per sec. Therefore, if the wave form at the commencement of the initial motion is sharp enough, we can determine its direction of propagation with a good accuracy. We found that the initial motion is quite sharp for about half of the earthquakes observed. For the rest of them the initial motion is not very sharp but we could determine the apparent velocity and direction of propagation[9]. We compared the direction of propagation of the initial wave portion with that of the initial motion in the case of sharp commencement and in that of vague commencement separately. In each case the mean value of the absolute value of the difference in angle between the two directions is computed.

9) S. MIYAMURA and M. TSUJIURA, *Bull. Earthq. Res. Inst.*, **37** (1959), 359-374.

It was 2.8 degrees in the case of sharp commencement and 12.3 degrees in the case of vague commencement. The difference is apparently significant, and should be attributed to the difference in sharpness of the waveform of initial motion. So, it is a reasonable conclusion that at least for the case of vague commencement the direction determination by the use of the initial wave portion of 2.5 sec duration is much more accurate than that by the use of the initial motion only.

Besides, in the exceptional case in which the epicentre is determined by the network of routine stations, the direction pointing to this epicentre, the direction of propagation of the initial motion and that of the initial wave portion coincide within 5 degrees. And the last one was between two others. This fact perhaps allows us to extend our conclusion.

On the other hand, the value of velocity of the initial wave portion does not coincide very well with that of the initial motion even when the coincidence between their directions is very good. This was expected because the initial wave potion involves partial waves which travelled along different paths from that of the initial motion and have different velocities along the horizon. Therefore, the velocity we obtained from that wave portion is an averaged one over those partial waves and may represent the velocity of the most prevailing partial wave in that portion.

If the velocity and thickness of the layers under the surface are known, the minimum value of the apparent velocity of initial motion along the horizon for an earthquake of a given epicentral distance can be obtained very easily. Fig. 5 (dotted lines) shows this minimum value of velocity for the area in question obtained from the result of seismic explosion experiment (1958)[10]. In the same figure are plotted the velocities of initial motions of earthquake waves investigated in the concurrent study by S. Miyamura and M. Tsujiura[11] against the P–S duration. The good agree-

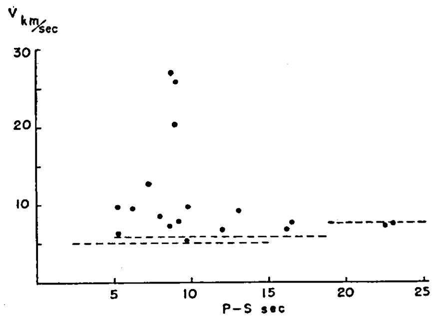

Fig. 5. Apparent velocity of propagation of initial motion plotted against P–S time. (After S. Miyamura and Tsujiura, *loc. cit.*, 9).).

10) RESEARCH GROUP FOR EXPLOSION SEISMOLOGY, *Zisin*, *Ser.* [ii], **11** (1958), 102–113.
11) S. MIYAMURA and M. TSUJIURA, *loc. cit.*, 9).

ment of theory and fact assures that the values of velocity of initial motion obtained for these earthquakes are reliable.

On the other hand, the velocity of the initial wave portion differs from that of the initial motion, being sometimes twice as large as the latter and sometimes one half, though most of them coincide within a relative error of 20%. The ratio of the velocity for the initial wave portion to that for the initial motion is plotted against the *P–S* time in Fig. 6. The ratio is not systematic with respect to the epicentral distance as shown in the figure. Therefore it is reasonable to think that the wave which prevails in the initial wave portion of 2.5 sec duration varies in a way depending mostly on the condition at which the earthquake starts and generates waves. On the average this prevailing wave is similar to the initial motion.

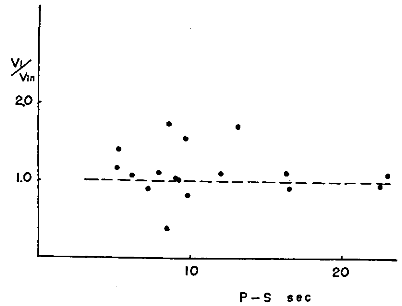

Fig. 6. The ratio of the velocity of the initial wave portion to that of initial motion plotted against *P–S* time.

For each individual earthquake the velocity and direction of propagation are given as a vector in Fig. 8 to Fig. 25 for every wave portion (1, 1′, 2, ...), wherever a plane wave is found, with the velocity vector for the initial motion (In) as well as the seismogram.

5. The nature of the wave portion including the *S* wave

As can be seen in Figs. 8 to 25, the wave portion, in which the *S* wave is considered to be prevailing from the inspection of seismograms, has a velocity sometimes too high and sometimes too low for the *S* wave. This is partly owing to our procedure in which the wave portions are picked up consecutively with a given time interval and they may involve partial waves having the nature of *P* waves and the surface waves. But if we shorten the interval of wave portion to eliminate the influence of the other waves than the *S* wave, the result will be less reliable because of the smaller size of samples. This kind of difficulty is inherent in the correlational study of waves.

The direction of propagation of this wave portion generally differs from that of the initial wave portion of the same earthquake. The discrepancy in angle is as small as 2 degrees in one case and as large as 60 degrees in another, and 25 degrees on the average. This discrepancy is quite natural if the wave portion is of the nature of the surface waves, because we are dealing with the waves having wave lengths of several hundred meters and the horizontal heterogeneity in topography and geology near the surface may easily cause this discrepancy. But there are a few cases in which the wave portion should be considered as an S wave and still has a considerably different direction from that of the initial wave portion. The earthquake recorded on Aug. 8, 1958 is a good example of this. In this case the velocity of the initial wave portion is 26.6 kmps and that of the S wave portion is 14.9 kmps which gives a reasonable value of velocity ratio P to S. These high velocities indicate that the earthquake occurred nearly right underneath the recording station. The P–S time is about 9 sec. The correlation curves for these two portions are given in Fig. 7 in order to show that the

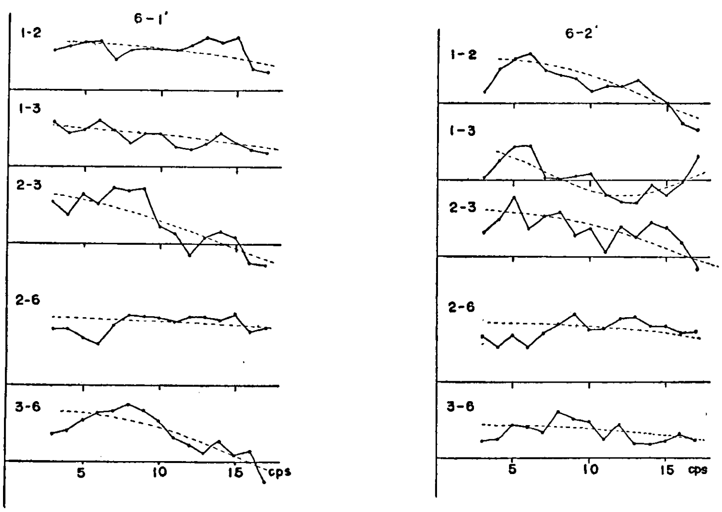

Fig. 7. Correlation curves for the P and S wave portions of the earthquake No. 8 recorded on Aug. 8, 04 h 35 m.

difference between the directions of propagation of these portions is quite significant. Form the ratio of the characteristic frequencies (by characteristic frequency we mean the wave length of the cosine correlation curve) for P and S wave portions for each pair of seismometers. If these two wave portions differ only in velocity and not in direction of propagation, then these ratios should all be the same. But in the present case, this ratio differs among the seismometer pairs, more for the pairs 2–3, 2–6 and 3-6 and less for the rest. This gives rise to a difference in angle of about 60 degrees between the directions of propagation of *P* and *S* wave portions.

This can be explained as follows; if the distribution of velocity of the *P* wave under the ground differs from that of the *S* wave in some way due to, for instance, the heterogeneity in distribution of material having a different Poisson's ratio or the existence of melted material at some places, the wave path will be different between the two waves. Further, if their wave paths are almost perpendicular to the earth's surface, as they are in the present case, the difference in the direction of propagation along the horizontal plane will be very large, even if the net amount of difference in angle between the two wave paths is small.

If we make, therefore, a more systematic study of this direction difference for the earthquake waves coming from just underneath the recording station, we shall be able to learn very precisely the difference between the distributions of velocity of *P* and *S* waves under the ground. This study can be done successfully only by the application of the present method, because it is very difficult to determine accurately the time of the commencement of the *S* wave on the seismogram.

6. Some remarks on the wave portions other than those including *P* and *S* wave

Figs. 8 to 25 show the velocity vectors of all the plane waves found in the records of 18 earthquakes. The following facts are worth noting here.

As can be seen in the figures, for the second 2.5 sec wave portion we find almost the same direction of propagation and value of velocity as for the initial wave portion in most of the cases. But after this, the velocity vector becomes variable and sometimes differs considerably from that of the initial wave portion. Therefore for the purpose of locating the earthquake epicentre the wave portion for the first 5 sec can be safely used.

The value of velocity of the wave portions located between *P* and *S* waves is usually very large. But in three earthquakes out of the eighteen, plane waves having a velocity of 3.5 to 3.8 kmps are found to exist. These may be surface waves generated, not near the origin of the earthquake, but near the recording station.

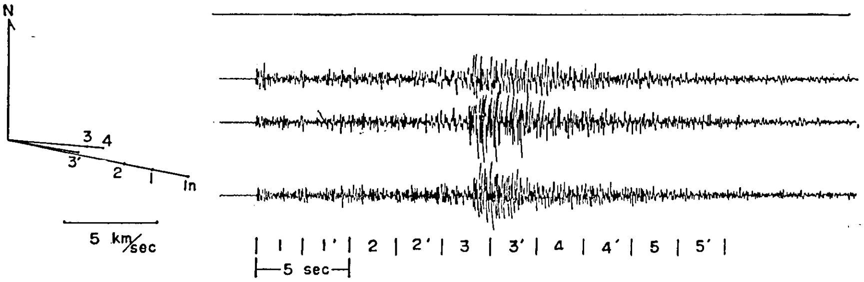

Fig. 8. Velocity vectors for the earthquake of Aug. 22, 07 h 09 m, No. 15.

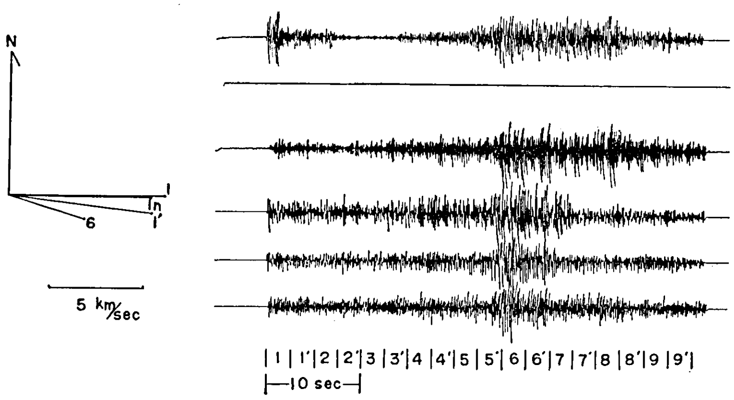

Fig. 9. Velocity vectors for the earthquake of Aug. 7, 14 h 13 m, No. 7.

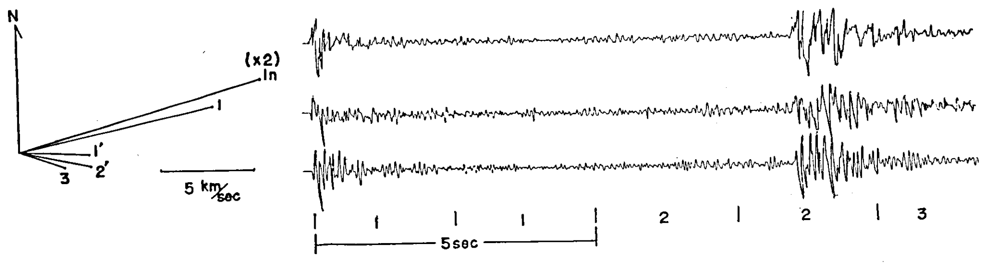

Fig. 10. Velocity vectors for the earthquake of Aug. 29, 04 h 35 m, No. 21.

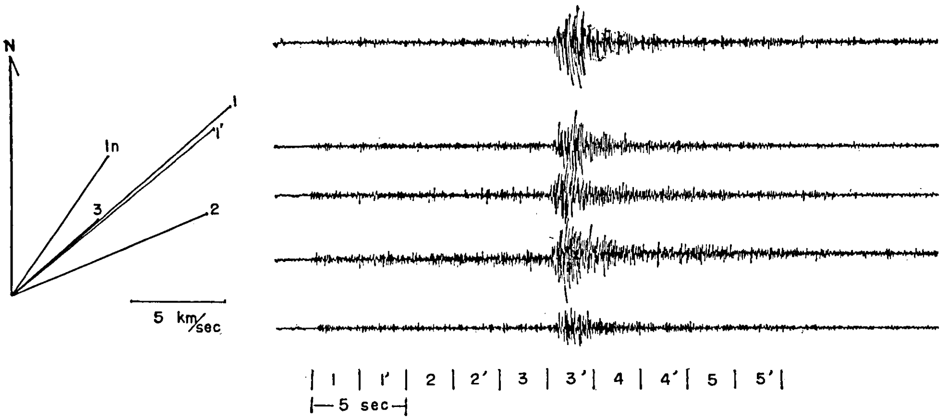

Fig. 11. Velocity vectors for the earthquake of Aug. 25, 19 h 38 m, No. 18.

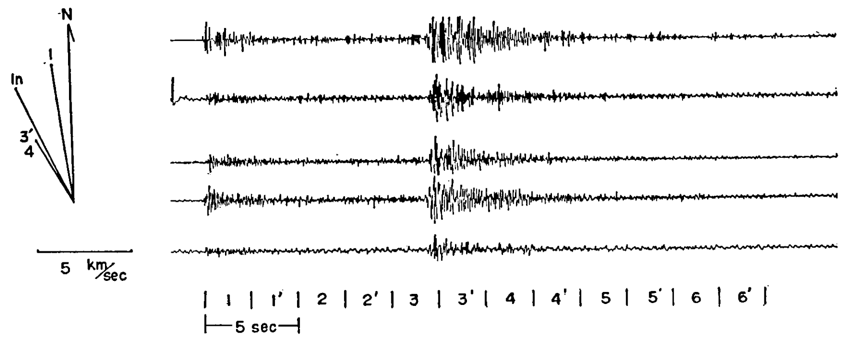

Fig. 12. Velocity vectors for the earthquake of July 29, 02 h 03 m, No. 2.

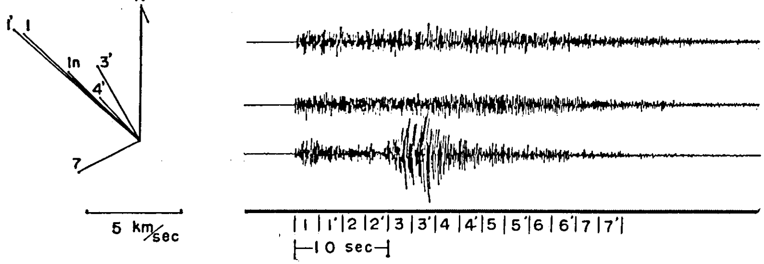

Fig. 13. Velocity vectors for the earthquake of July 29, 13 h 53 m, No. 3.

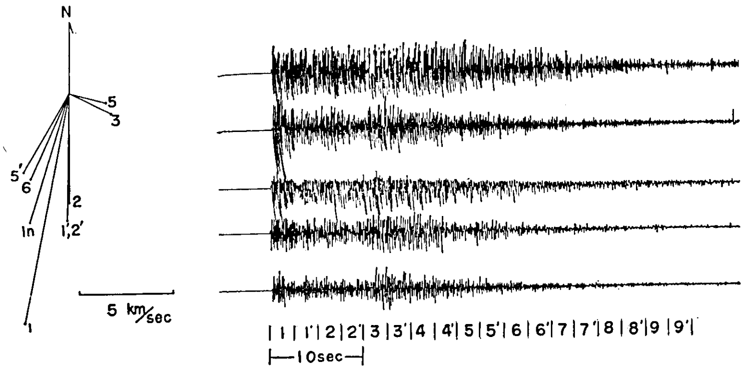

Fig. 14. Velocity vectors for the earthquake of July 30, 00 h 05 m, No. 4.

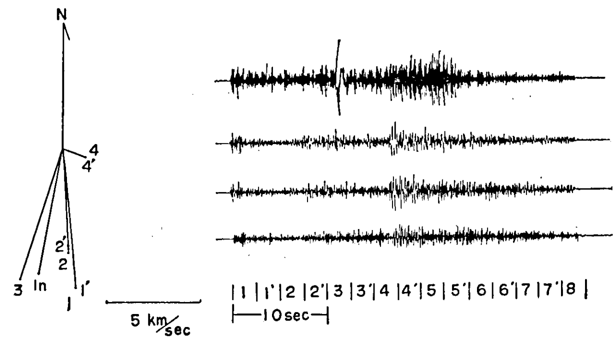

Fig. 15. Velocity vectors for the earthquake of Aug. 9, 13 h 02 m, No. 10.

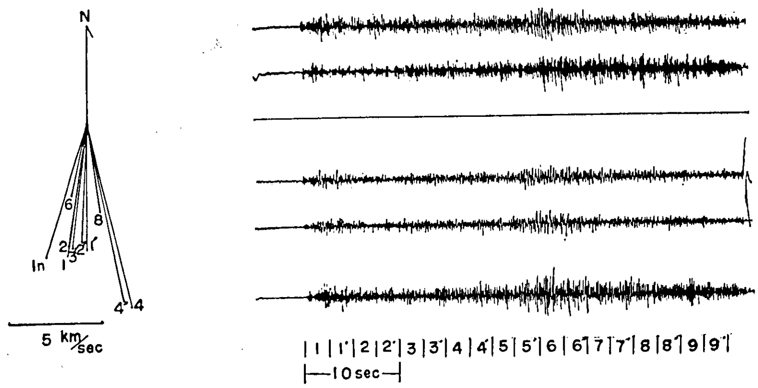

Fig. 16. Velocity vectors for the earthquake of Aug. 15, 18 h 49 m, No. 11.

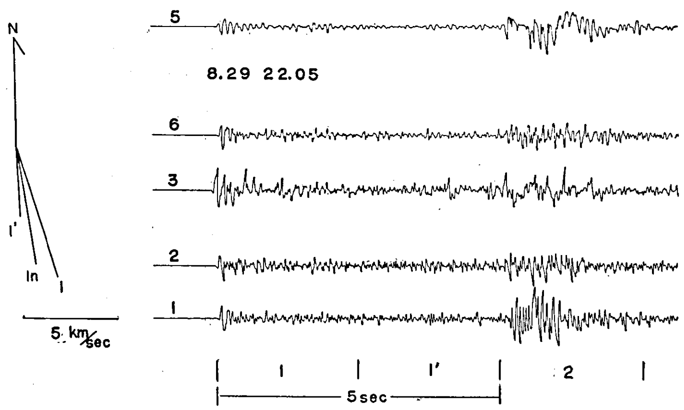

Fig. 17. Velocity vectors for the earthquake of Aug. 29, 22 h 03 m, No. 22.

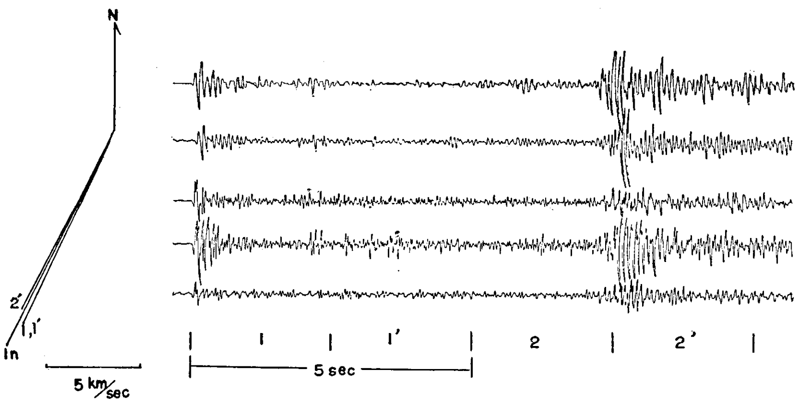

Fig. 18. Velocity vectors for the earthquake of July 30, 19 h 50 m, No. 5.

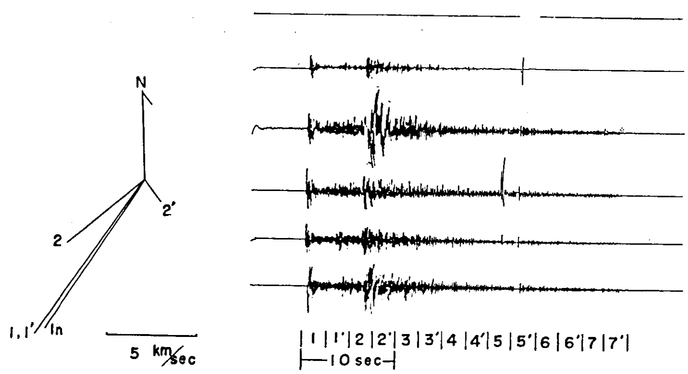

Fig. 19. Velocity vectors for the earthquake of Aug. 9, 10 h 37 m, No. 9.

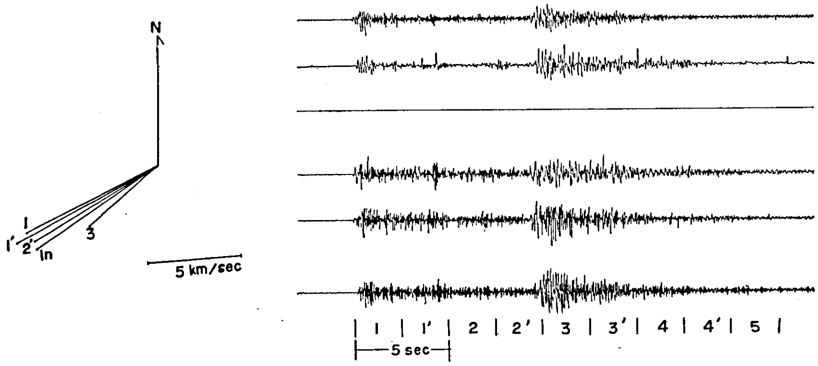

Fig. 20. Velocity vectors for the earthquake of Aug. 24, 04 h 55 m, No. 16.

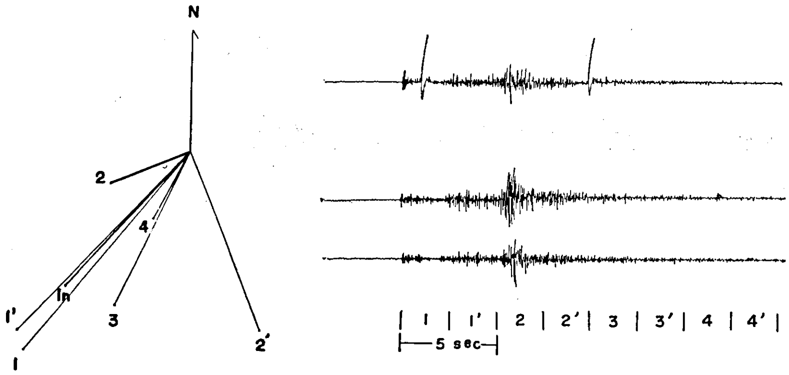

Fig. 21. Velocity vectors for the earthquake of Aug. 28, 13 h 06 m, No. 20.

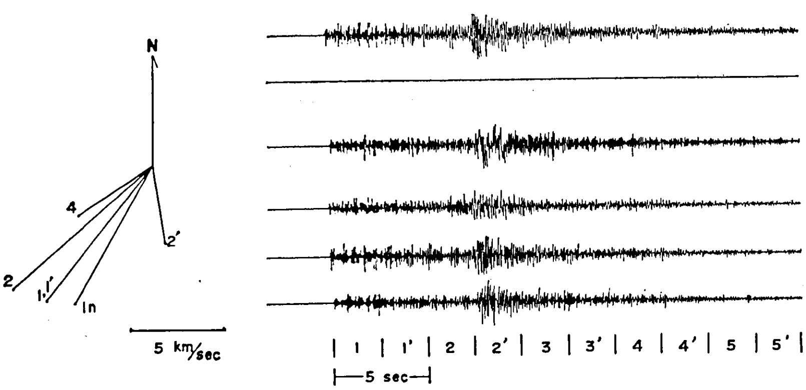

Fig. 22. Velocity vectors for the earthquake of Aug. 8, 04 h 35 m, No. 8.

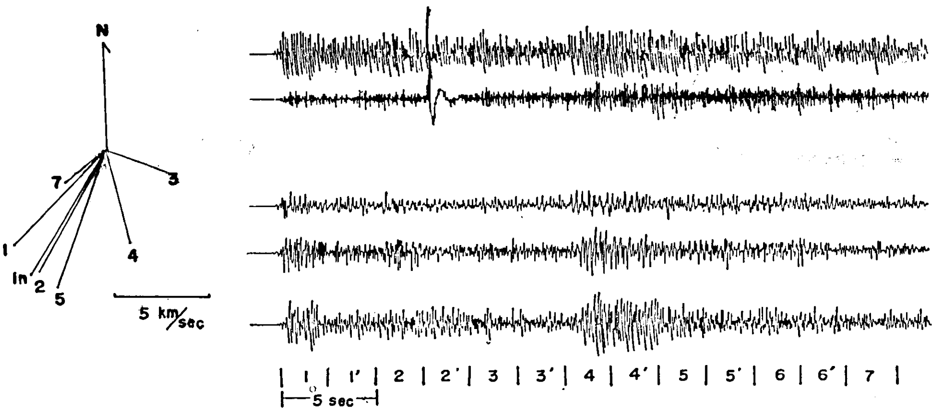

Fig. 23. Velocity vectors for the earthquake of Aug. 15, 15 h 34 m, No. 13.

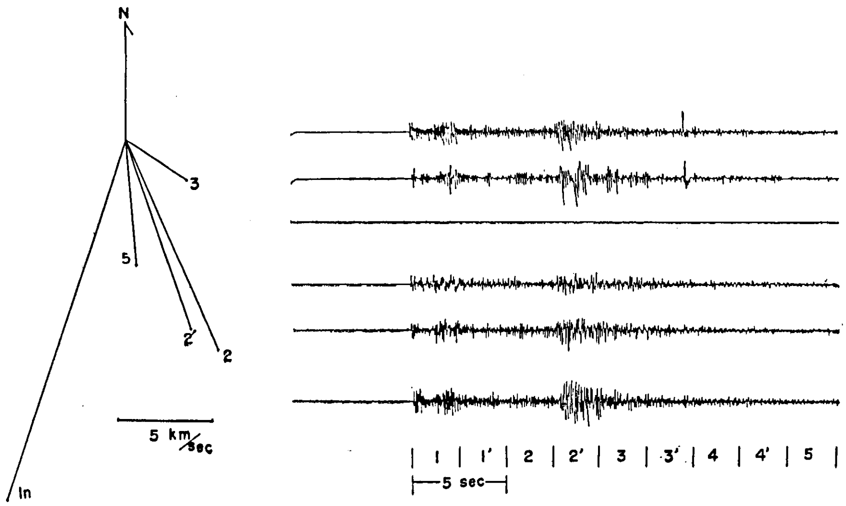

Fig. 24. Velocity vectors for the earthquake of Aug. 24, 14 h 40 m, No. 17.

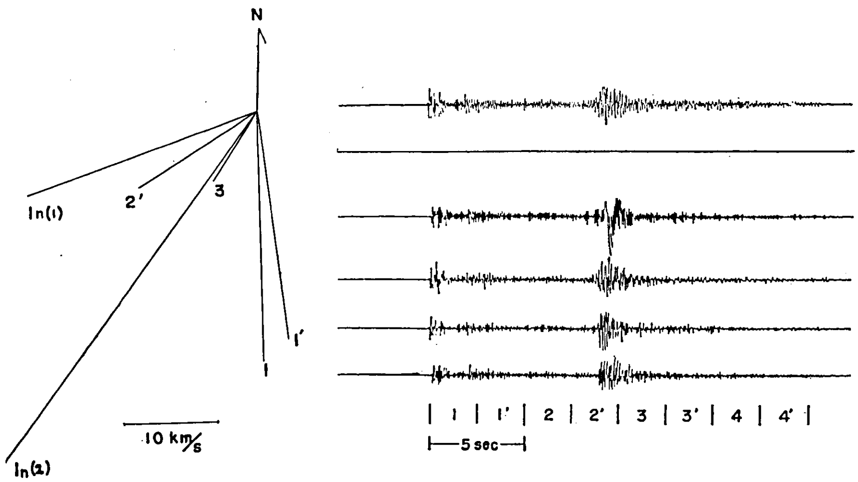

Fig. 25. Velocity vectors for the earthquake of Aug. 6, 23 h 22 m, No. 6.

The wave portions after the S wave portion usually have low velocity, 2 to 4 kmps in most of the cases. But in four earthquakes out of the eighteen, plane wave having a velocity of 7 to 10 kmps are found to exist. These waves may be reflected ones from a certain boundary, and they appear in some cases about 10 sec after the P arrival and about 20 sec in other cases. This problem also requires more systematic study.

7. Comparison of the present result with that obtained by the preliminary investigation at Kamakura

The result of the present investigation made at Mt. Tukuba agreed with that of the preliminary study made at Kamakura in most aspects despite differences in geology and topography of the two places. The followings are common conclusions; a plane wave having the direction which points to the origin of the earthquake fairly well has a comparable average power to irregular noises in the wave portion for the first several seconds after the initial motion. The fraction of average power carried by the plane wave is greatest in the initial wave portion, decreases later, becomes slightly larger at the arrival of the S wave, and very small in the wave portions after the S wave.

On the other hand, plane waves which were found to have a low velocity of several hundred mps at Kamakura are not discernible in the waves recorded at Mt. Tukuba. This may be attributed to the difference between the surface rocks of these two places, because seismometers were placed on sandstone at Kamakura, but on granitic rocks at Mt. Tukuba.

8. Remarks

Local earthquakes occurring in Kanto have been studied by various authors, e.g., Imamura (1957)[12], Nasu, Hagiwara and Omote (1936)[13], Nasu and Yasuda (1941)[14], and Asada (1958)[15]. Fig. 26 shows the distribution of epicentres of earthquakes which occurred in Kanto during the period from 1914 to 1940 quoting from the paper of Asada[16]. On

12) A. IMAMURA, *Bull. Earthq. Res. Inst.*, **3** (1927), 165–185.

13) N. NASU, T. HAGIWARA and S. OMOTE, *Bull. Earthq. Res. Inst.*, **14** (1936), 427–437.

14) N. NASU, C. YASUDA, *Bull. Earthq. Res. Inst.*, **18** (1941), 476–491.

15) T. ASADA, *loc. cit.*, 3).

16) T. ASADA, *loc. cit.*, 3).

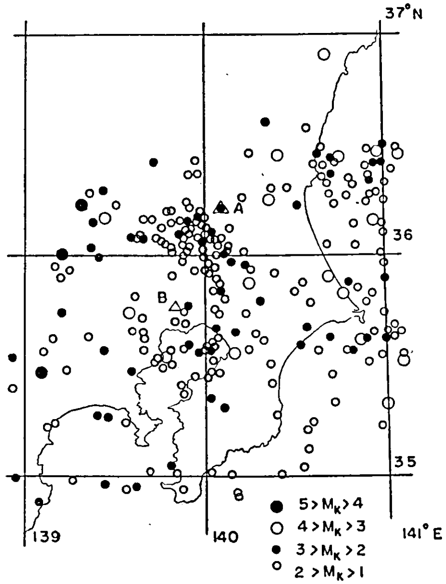

Fig. 26. Epicenters of local earthquakes in Kanto. The mark A indicates the location of Mt. Tukuba. (After T. Asada, *loc. cit.*, 3).)

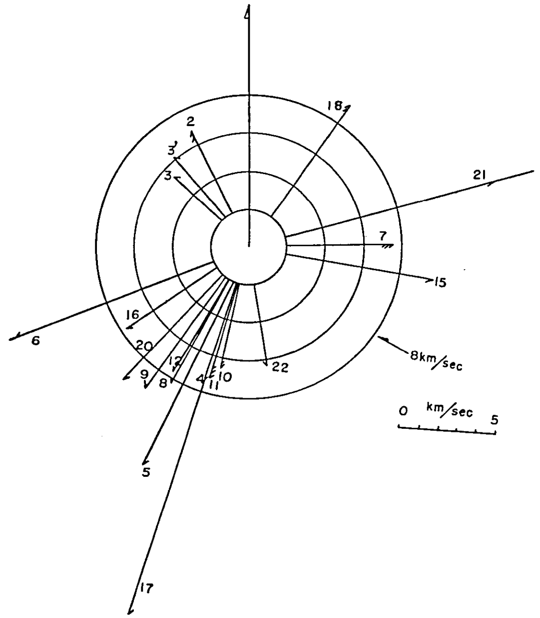

Fig. 27. Distribution of velocity vectors for the initial motion of earthquakes recorded in the present study. (After S. Miyamura and M. Tsujiura).

the other hand, the distribution of the velocity vectors for the propagation of initial motion of earthquakes which we recorded in the present investigation is shown in Fig. 27. The coincidence between these two maps is remarkable; Both maps show that the frequency of earthquake occurrence is very high in the south-west of Mt. Tukuba, low in the north-west and the south-east, intermediate in the east. Thus we conclude with S. Miyamura and M. Tsujiura[17], whose study is concerned partly with the same earthquakes studied here, that it is possible to guess the distribution of larger earthquakes for a long period by learning that of smaller earthquakes for a period as short as one month. There may be some objections to the above statement, because in some cases the distribution of earthquake foci depends on their magnitude and also because the earthquakes have a tendency to occur as a swarm for a limited period and in a limited area which varies from time to time. But it is valuable to study how the small earthquakes occur and how they are related to the larger ones at various places on the earth. One of the present writers is now undertaking a similar study by the method used in the present paper on the local earthquakes occuring in California. This investigation in California will be interesting also with respect to the question of the fraction of average power carried separately by the plane waves and irregular noises which coexist in earthquake waves, because this is expected to depend on the geology and topography of the related area in some ways.

Acknowledgement

The writers express their thanks to Prof. Chuji Tsuboi for his constant guidance and encouragement. The writers' thanks are also due to Assist. Prof. Setumi Miyamura for his kind help and valuable advice given in the course of investigation, especially in the operation of seismograph systems developed hy him, without which the present work could not have been carried out. The writers are indebted to Mr. M. Watanabe of Mt. Tukuba Observatory for his very kind help in the location of seismometer sites and to Mr. M. Hori, Mr. H. Matumoto and Miss H. Kanehira for their help in carrying out the observation and analysis. We owe to Mr. A. Okada the determination of the position of seismometers, and to Mr. A. Jitsukawa his help in making a temporary receiving station.

17) S. MIYAMURA and M. TSUJIURA, *loc. cit.*, 10).

This manuscript was partly prepared, while one of the writers was working in the California Institute of Technology as a visiting research fellow. The writers thank Prof. C. Hewitt Dix for his kindness in reading the manuscript and giving them helpful discussion and criticism.

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14. 近地地震波動の相関法による研究

地震研究所 {安芸敬一 / 辻浦賢}

この研究は，表題はちがうが，著者らによる既報「近地地震波動のスペクトル的研究 (1)」の続報である．表題をかえた理由は，スペクトル的研究ということばが，地震波の時間的スペクトルのみの研究という意味に誤解されやすいからである．

われわれの目的は，地震波の空間スペクトルをしらべることによつて，これまでのように，ただ特別な相の初動にのみ注目するかわりに，地震波動の任意の部分の性質を，統計的にあきらかにすることである．この種の知識はちいさい地震の震源をきめたり，地震のエネルギーを評価するなどの問題で，地震波のはじめからおわりまで全部を利用しようとするばあい是非とも必要なことである．

前報では鎌倉における予備的な研究結果を報告したが，本研究では，つぎの 4 つの点で改良をくわえた．

1. 地震計を鎌倉の砂岩のかわりに，筑波山近隣の花崗岩の露出しているところに設置し，台数を3台から6台にまし，各地震計間の距離を平均約 100 m から，約 500 m にのばした．

2. 観測には，6 台の上下動地震計，2 台の無線地震波記録装置 RTS-II，2 軌道のテープレコーダーがもちいられたが，これらの位相特性を各チャンネル間で，位相差 10 度内外におさえることができた．

3. 解析装置に検算回路を附設した結果，相関係数のあたいにして 10% の誤差で計算できる状態に装置を維持することができた．

4. 受信点を東京大学の病院新館 9 階に設置し，記録解析一切を，そこでおこなつたので，研究の能率がいちぢるしく増進した．

約 1 個月の観測で，約 20 個の地震を磁気テープに記録した．第 1 報でのべたのとおなじ方法，装置で解析した結果えられたおもな結論はつぎのようなものである．

1. 地震波には，どの部分にも，複雑な雑音がまじつている．この研究では，各部分のながさを共通に 2.5 秒にとつたが，初動からはじまる最初の部分では，ほとんど，つねに規則ただしい平面波がみいだされる．この平面波がうけもつ Power のわりあいは，最大 80%，平均 40% である．S 波をふくむ部分に平面波がみいだされる確率は，約 80% で，そのうけもつ Power のわりあいは，大体 20・30% である．P と S のあいだの部分では，平面波が約 60% の確率でみいだされ，そのうけもつ Power のわりあいは，P の部分のすぐあとでは，しばしばおおきく，その他では，おおきいばあいもあり，ちいさいばあいもあつて，まちまちである．S の部分のあとでは，一般に平面波をみとめることは困難で，その確率は 30% 程度であり，そのうけもつ Power のわりあいもきわめてちいさく，30% をこえることはない．平面波をみいだしえなかつた部分の性質はきわめて複雑であつて，空間的白色雑音と断ずることはできない．
2. われわれは記録をテープにとつているので，それを非常にはやいおくりの記録紙上に再生することが容易である．したがつて，P 波のたちあがりを精度よくよみとることができる．ただし P 波の立上りの波形がするどいときには信頼できるが，にぶいときには，信頼できない．そこで，たちあがりのするどい地震ばかり（全体の約半数）をあつめて，それらについて，初動の

到来方向と最初の 2.5 秒間の波動部分にみいだされた平面波の進行方向とをくらべてみると，角度差の絶対値の平均が 2.8 度であつた．するとくないばあいには，おなじ平均が 12.3 度であつたので，このちがいは，あきらかに初動の方に責任がある．したがつて伝播方向の精度に関しては，初動だけよりも，初動から 2.5 秒間の波動全部をつかつた方がよいことがわかる．われわれのしらべた地震のうち，ひとつだけ気象庁の観測網で震源がもとめられたが，この震源への方向，初動の伝播方向，初動部分の伝播方向の三者とも 5 度くらいの範囲におさまり，最後のものが他のふたつの中間をさしていた．初動の到来方向とみかけの速度については，宮村・辻浦[9]により別にくわしく報告されている．

3. S 波をふくむ部分の波動としての性質は一般に複雑であつて，えられた速度も表面波にちかいばあいや，P 波にちかいばあいなど，いろいろである．その伝播方向も P 波部分と一般に一致せず，その差の絶対値の平均は 25 度もある．この差は，もし，この部分に表面波が卓越しているとすれば，表面波は地表のあさいところの不一様性によつて，(われわれは，波長数百米の波動を対象としているので) 屈折しやすいから，当然期待されてよいとおもう．しかし，一方あきらかに，S 波がおもに卓越しているとみられる地震のばあいに，非常にはつきりしたくいちがいがみいだされた．このばあいには P 波部分の速度 26.6 km/s に対して，S 波部分が 14.9 km/s の速度をもつていて，たしかに表面波ではない．速度がおおきいからこの地震は観測点のほとんど直下でおこつたとおもわれる．したがつて，P 波と S 波にわずかな伝播方向の差があつても，地表面にそうみかけの方向には，おおきな差を生じうる．P 波と S 波のわずかな方向の差というものは，Poisson 比のちがう物質の分布とか，とけている物質部分の存在によつて，容易に説明される．いいかえれば，観測点のほとんど直下におこる地震をつかえば，地下における P と S の速度分布のくいちがいを非常に精密にしらべることが可能である．この問題には，もつと系統的な研究が必要である．

4. 筑波山と鎌倉での結果は，ほとんどの点で，大差なかつたが，ただ鎌倉でみいだされた数百米毎秒のおそい速度をもつ波動は，筑波ではみとめられかつた．あさいところの地質がちがうためであろう．

5. この観測でえられた地震の震源の筑波山からの方向分布をつくると，宮村・辻浦[9]ものべているように 1914 年から 1940 年にかけて生じた地震分布図と微細な点まで一致している．1 個月程度の観測で約 30 年間の地震分布図を推測できるということは，興味深い．

このような研究は，いろいろな場所でおこなうことが重要である．著者の 1 人は，現在カリフォルニアの局地地震をしらべるべく，現地においてこのための装置をつくり，観測をはじめている．