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SEISMOMETRIC INVESTIGATION OF THE EARTH'S INTERIOR

PART II. ON THE STRUCTURE OF THE EARTH'S CRUST

BY

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ABSTRACT

The structure of the earth's crust was determined by an analysis of the refracted seismic waves of shallow earthquakes propagated along the discontinuous surfaces in the earth's crust. The presumed structure was ascertained, on the other hand, mainly by the study of the transformed waves in earthquakes of deep focus. The structure of the earth's crust thus obtained around Kyoto was found to be composed of four distinct layers of different property whose thickness and propagation velocity of P– and S–waves are respectively, listing from the earth's surface to the deeper layers, 1.5 km and 3.8, 2.2 km/sec for the sedimentary layer; 6 km and 5.4, 3.1 km/sec for the granitic layer; 14 km and 5.8, 3.4 km/sec for the granodioritic layer; 10 km and 7.0, 4.0 km/sec for the basaltic layer. The velocities of P– and S–waves of the layer underlying the basaltic layer were estimated to be nearly 8.1 km/sec and 4.7 km/sec, respectively.

Also in this article the regional differences in the crustal structure, especially the local character of the thickness of the basaltic layer, were discussed in some detail, and moreover the correlation between the regional differences of the crustal structure and the focal depths of shallow earthquakes in their respective area is mentioned.

1. Introduction

The seismometric investigation of the structure of the earth's crust was begun early by A. Mohorovičić in his pioneering study of the Kulpa Valley Earthquake of October 8, 1909 (**1**). He confirmed the existence of a sharp discontinuous boundary named the Mohorovičić discontinuity which seismometrically separates the earth's crust from the underlying mantle. Since the discovery of the Mohorovičić discontinuity the crustal structure has been investigated more and more in detail by many researchers such as, for example, B. Gutenberg (**2**), H. Jeffreys (**3**), V. Conrad (**4**) and others who made great contributions to the development of our knowledge in the many-layered structure of the earth's crust.

On the other hand the seismometric investigation of the crustal structure by artificial explosions, which was commenced by O. Hecker on the Oppau Explosion of September 21, 1921 (**5, 6, 7**), has become greatly active parallel with that of natural earthquakes. And after the Second World War the seismometric investigations of the crust with artificial explosions have had a wonderful development in recent years in succession of the great artificial explosion on April 18, 1948 at Heligoland.

In reflection upon the forty years' progress of the seismometric investigation of the crust through natural earthquakes and artificial explosions, the many important facts on the structure and property of the earth's crust have certainly been made clear, but, on the other hand, many more essential and attractive problems on the nature of the crust have been presented in the course of research, and they remain, for the most part, unsolved and are awaiting future development in research. Needless to say, such problems are never completely solved by the seismometric method only, and therefore synthetic investigations are strongly recommended combining the seismometric method with those of gravimetry, structural geology, laboratory experiment of rock under high pressure and temperature, theoretical treatment of physical state and property of rock at deep underground as well as other effective types of research.

In the following, a short review will be made of some interesting problems concerning the crustal structure proposed and discussed in comparatively recent times. In U.S.A. especially in the Eastern District (New England and others) and the Western District (California), the crustal structure has been studied in detail by means of natural earthquakes and artificial cxplosions, and two different values were assigned by some researchers for the velocity of P-wave in the granitic layer such as 5.6 km/sec from the analysis of natural earthquakes and 6.1 km/sec from those of artificial explosions. But P. Byerly (8, **9**) contends that the velocity is determined in California as nearly same value of 5.6 km/sec from both natural and artificial shocks. Also L. Don Leet (**10**, **11**, **12**) proposed a consistent value of about 6 km/sec for the *P*-velocity in the granitic layer derived from both methods and attributed the inconsistency formerly discussed mainly to the personal misinterpretation of seismic waves. On the other hand, B. Gutenberg, in his early investigation, had an opinion (**13**) on this problem of inconsistency between the velocity values derived from the natural earthquakes and artificial explosions that there exists a sedimentary layer of high velocity ($V_p > 6$ km/sec) overlying the granitic layer of velocity ($V_p \approx 5.6$ km/sec), but recently he proposed an entirely different crustal structure from the former conception (**14**, **15**, **16**, **17**, **18**). Namely he assumed the existence of comparatively low velocity layers at depths 15 km and 80 km below the ground surface, and by such crustal structures explained not only the inconsistency of velocity values derived from both natural and artificial shocks but also gave a different interpretation to the seismic waves which were formerly considered as refracted waves (forerunners) propagated along the crustal discontinuity. Although his theory of low velocity layers is in itself a very attractive and suggestive postulation, it yet contains too many question-able points to be generally approved. The difficulties are considered to exist in such points as, for example, an unpractical assumption that earthquakes concerned should occur mainly in the low velocity layer, a complex mode of seismic wave-propagation which is somewhat unreasonable in the point of curvature of trajectory of seismic wave, especially at the small epicentral distance, and some other matters. But this theory of low velocity layer in the earth's crust is certainly an interesting and promising idea which relates to the original formation and secular change of the earth's crust, and may be applicable to, though invalid in the case of a problem affecting the whole earth, any particular region such as, for example, a volcanically and seismically active area, the magmatic reservoir and other allied structures being reasonably considered to exist underneath their ground surfaces. This theory should be earnestly examined, in near future, from various standpoints of research. On the problem of inconsistency of obtained values of seismic velocity from natural and artificial shocks, A. Kubotera (**19**) explained it from the standpoint of the rheological property of the crust, and D. S. Hughes and J. H. Cross (**20**) discussed the difference between the velocity values of rocks determined by laboratory experiments and those obtained by natural earthquakes.

In Japan the systematic and detailed observations of seismic waves aroused by five artificial explosions successively fired in 1950–1953 in Northern Honshû (at Isibuti and Kamaisi) (**21**, **22**, **23**, **24**, **25**) should be highly appreciated and their obtained velocities are listed in the last of this paragraph, the velocity of the upper layer being somewhat more highly estimated compared with that of other districts in our country. The seismometric investigation of the earth crust made by T. Matuzawa (**26**, **27**, **28**, **29**) from the analysis of natural earthquakes has been the most comprehensive and detailed in Japan, the obtained results being also listed in the last. And the crustal structure determined by E. A. Hodgson (**30**, **31**) using the data of the Tango Earthquake, in Japan, March 7, 1927 are considerably different from that of Matuzawa. All these results and others suggest implicitly the complexity of the crustal structure in our country.

The problems of the number of layers in the crust and the thickness of the crust have become a more and more attractive object of research in proportion to the recent trend towards accurateness of observation and analysis. H. E. Tatel, M. A. Tuve and others (**32**, **33**) studied the crustal structure in U. S. A. by the analysis of data of 250 artificial explosions, and obtained nearly 6.1 km/sec as the P–velocity of the upper layer, and they could not find any definite intermediate discontinuity between the ground surface and the Mohorovičić discontinuity. On this problem L. Mintrop (**34**) H. Reich and others (**35**) ascertained the existence of two layers in the crust of the granitic ($V_p \approx 5.5$ km/sec) and the gabbroic ($V_p \approx 6.2 \sim 6.6$ km/sec) layers by the Heligoland Explosion in 1948, but P. L. Willmore (**36**) and J. P. Rothé (**37**) expressed the opinion that the existence of the gabbroic layer above mentioned is highly problematical. In this case the existence of the intermediate discontinuity would certainly be influenced by the locality of the crustal structure, but moreover the uncertainty of its existence is reasonably explained from the fact that the determination of intermediate discontinuity is quite difficult compared with that of the Mohorovičić discontinuity, especially in the analysis of travel time-distance curve, because the degree of discontinuity in the intermediate layer is considered to be less distinct than that in the Mohorovičić discontinuity. Therefore, in the detailed investigation of the crustal structure, the analysis of travel time-distance curves should be appropriately supplemented by other methods such as the analysis of the reflected waves at the discontinuity, the duration of refracted waves (forerunners) along the discontinuity or the transformed waves (*Wechselwellen*) changed at the discontinuity and other available phases of seismic waves.

Regarding the methods of analysing the reflected waves, the transformed waves and the duration of refracted waves at any crustal discontinuity for the study of the crustal structure, their practical applications are considerably few, especially in case of taking into consideration the amplitudes of their respective seismic waves, compared with those of the analyses of only the travel time-distance curves. M. A. Tuve, H. E. Tatel and others (**32, 33**) had observed the very distinct phases of the reflected wave at the Mohorovičić discontinuity in some areas of North America and discussed the regional differences in the crustal structure. They obtained the same strong reflection under some areas such as, the Atlantic coast, the Tennessee mountains and the pre-Cambrian shield in Minnesota and the Appalachians, the Canadian Shield, and the Colorado Plateau, but in some regions these strong reflections were never observed, no matter how carefully they might search. Under the Colorado Plateau where the elevation is over 2 km, a Mohorovičić discontinuity was found at a depth of some 35 km contrary to the expected depth of 70∼80 km from the isostatic compensation. B. Gutenberg (**17, 38**) observed the reflected waves at a surface a little above the Mohorovičić discontinuity and moreover H. Reich (**39**) discussed the same problem. L. Mintrop (**34**) obtained and discussed not only the crustal reflection but also the reflected waves at the discontinuity in the mantle. On the analysis of the transformed waves T. Matuzawa and others (**40, 41**) had early in 1928 discussed in detail the transformed waves changed at the interface of the sedimentary layer named the Kantô loam and the underlying granitic layer, and P. Byerly (**42**) also treated various transformed waves in the observation of the Nevada Earthquake in America.

Reviewing generally the development of our knowledge on the crustal structure

Table 1.

No. of Reference	Author (Date)	Locality	Structure (V_p km/sec, V_s km/sec (Thickness km)) Sedimentary layer	First layer	Second layer	Third layer	Mantle surface
(1)	Mohorovičić (1910)	Jugo-Slavia		5.6~ 3.2~ 5.7, 3.6 (0—50)			7.8, 4.2 (50—)
(2)	Gutenberg (1915)	Germany		5.6, 3.2 (0—30)	6.2, — (30—44)		8.2, 4.4 (44—)
(3)	Jeffreys (1926)	England		5.4, 3.3 (0—12)	6.3, 3.7 (12—30)		7.8, 4.4 (30—)
(4)	Conrad (1928)	Austria		5.6, — (0—12)	(12—37)	6.5, — (37—45)	8.1, — (45—)
(5)	Hecker (1922)	Germany (Oppau)		5.73, 3.0 (—)			
(6)	Wrinch Jeffreys (1923)	"		5.4, 3.15 (0—15)			
(7)	Gutenberg (1926)	Germany	2~4, — (0—2)	5.5~ 5.6, — (2—)			
(8)	Byerly Wilson (1935)	North California		5.6, — (1—13)	6.6, — (13—25)	7.3, — (25—31)	8.0, — (31—)
(9)	Byerly (1939)	North California		5.61, 3.26 (0—9)	6.72, — (9—)	7.24, — (—32)	8.02, — (32—)
(11)	Leet (1941)	New England		6.13, 3.45 (0—16)	6.77, 3.93 (16—29)	7.17, 4.27 (29—36)	8.43, 4.62 (36—)
(15)	Gutenberg (1951)	South California		5.7~ 6.0, — (0—6)	6.5, 3¾ (6—10)	(10—40)	8.1~ 8.2, — (40—)
(18)	Gutenberg (1954)	South California		5.9~ 3.2~ 6¾, 4± (1—10)	6¾~ 4±~ 6±, 3¼ (10—15)	6±~ 3¼ 7±, (15—40)	8.15, 4.55 (40—)
(21)~(25)	R. G. E. S. (1954)	Northeastern Japan	2.51, — (0—2)	5.75~ 5.85, — (2—)	6.10~ 6.20, — (—)		7.5~ 8.0, — (20~25 —)
(26)~(29) (41)	Matuzawa (1929)	Japan	1.94, 1.14 (1—4)	5.0, 3.15 (4—20)	6.2, 3.7 (20—50)		7.5, 4.5 (50—)
(30) (31)	Hodgson (1932)	Central Japan		6.3, — (0—16)			7.75 — (16—)
(32)	Tatel & others (1953)	U. S. A.		6.2, — (0—30~50)			8.0, — (30~50 —)
(34)	Mintrop (1949)	Germany	3.5, 2 (0—4)	5.2, 2.9 (4—13)	6.5, 3.6 (13—28)		8.1, 4.5 (28—)
(35)	Reich & others (1951)	Germany	3.6, — (0—6)	5.4, — (6—10)	6.18~ 6.6, — (10—27)		8.19~ 8.32, — (27—)
(36)	Willmore (1947)	Germany	(0—7)	5.95, — (7—27)			8.18, — (27—)
"	"	"	4.4, — (0—6)	5.5, — (6—14)	6.5, — (14—29)		8.18, — (29—)
(37)	Rothé (1947)	Germany		5.5, — (0—30)			8.1, — (30—)
	Kishimoto (1955)	Central Japan	3.8, 2.2 (0—1.5)	5.4, 3.1 (1.5—7.5)	5.8, 3.4 (7.5—21.5)	7.0, 4.0 (21.5—31.5)	8.1, 4.7 (31.5—)

Table 2. List of the earthquakes used in the present investigation. Latitude and longitude of epicenter and focal depth are those reported by the Central Meteorological Observatory of Japan. Observatories K, A, Aso and B denote the Kamigamo Geophysical Observatory, the Abuyama Seismological Observatory, the Aso Volcanological Laboratory and the Beppu Balneological Laboratory respectively. k denotes the Omori's constant ($k=\Delta(\text{km})/P\sim S(\text{sec})$).

No.	Date					Latitude and longitude of epicenter		Epicentral region	$P\sim S$	Focal depth	Duration of forerunner		Δ	k	Observatory
			d.	h.	m.	N °	E °		s	km	s	s	km		
a-Group															
1	1945	Jan.	12	—	—	—	—	near Okazaki City	16.0	—	1.4 ,	2.3	132	8.2	K
2	1949	June	17	23	14	34.7 ,	137.6	Lake Hamana	22.0	10	1.5 ,	3.7	171	7.8	K
3	1947	Feb.	21	15	—	34.8 ,	138.0	Shizuoka Pref.	23.0	20		3.4	205	8.7	K
4	1945	June	24	—	—	—		〃	23.2	—		2.8	—	—	K
5	1936	Oct.	20	23	15	35.0 ,	138.2	Shizuoka Pref.	25.2	—		3.6	223	8.9	K
9	1949	June	17	23	14	34.7 ,	137.6	Lake Hamana	23.1	10	1.7 ,	3.9	187	8.1	A
7	1936	Oct.	20	23	15	35.0 ,	138.2	Shizuoka Pref.	28.0	—		4.2	243	8.7	A
8	1947	Mar.	11	14	16	34.9 ,	138.4	〃	29.2	20		4.6	242	8.3	K
9	1944	Dec.	9	10	54	34.4 ,	139.0	Izu Peninsula	36.0	20∼40		6.0	306	8.5	K
b-Group															
1	1935	Nov.	19	08	36	33.8 ,	134.1	Tokushima Pref.	23.2	—		3.1	206	8.9	K
2	1950	Mar.	16	02	40	33.6 ,	134.6	〃	23.2	10		3.0	192	8.3	K
3	1953	May	30	23	37	34.2 ,	133.2	Central Setoumi	30.1	20		5.1	253	8.4	K
4	1944	Aug.	3	10	42	—		Shimane Pref.	30.7	—		5.8	—	—	K
5	1935	Nov.	19	08	36	33.8 ,	134.1	Tokushima Pref.	31.3	—		5.1	244	7.8	B
6	1953	July	30	17	25	34.1 ,	132.8	Western Setoumi	37.3	20		6.9	308	8.2	K
7	1953	May	17	06	20	35.1 ,	132.6	Shimane Pref.	30.4	40		3.7	288	9.5	K
8	1950	Aug.	22	11	04	35.2 ,	132.7	〃	31.6	30		3.8	280	8.9	K
9	1953	June	8	22	50	35.0 ,	132.8	Hiroshima Pref.	31.8	10		3.9	270	8.5	K
10	1941	Apr.	6	01	50	34.6 ,	131.7	Yamaguchi Pref.	44.0	—		5.7	375	8.5	K
c-Group															
1	1941	Mar.	4	—	—	37.5 ,	138.0	near Naoetsu	32.3	—		6.3	336	10.4	K
2	〃	〃	7	12	00	—		Nagano Pref.	33.6	<10		6.6	290	8.6	K
3	1937	July	4	00	23	36.9 ,	138.0	Niigata Pref.	40.0	—		7.4	249	6.2	K

d-Group														
1	1947	May	9	03	32	33.7 ,	135.3	Tanabe-Bay	18.4	—	3.8	154	8.4	K
2	1947	Jan.	17	02	44	33.8 ,	134.2	Tokushima Pref.	19.7	10	2.2 , 4.7	200	10.1	K
3	1946	May	9	05	17	33.6 ,	136.0	Katsuura Wakayama Pref.	22.6	0	5.2	162	7.2	K
4	1947	Apr.	29	14	35	33.0 ,	135.0	off Shionomisaki	24.0	—	5.8	238	9.9	K
5	1949	May	18	07	48	34.0 ,	133.8	Tokushima Pref.	25.0	20	6.5	214	8.6	K
6	1953	July	31	04	24	33.2 ,	134.8	Nankaido	27.2	20	4.6 , 7.4	224	8.2	K
7	1947	Feb.	16	18	19	33.1 ,	134.6	〃	28.0	—	7.6	242	8.6	K
e-Group														
1	1943	July	1	13	40	36.2 ,	140.0	Ibaraki Pref.	45.6	50	9.5	410	9.0	K
2	1945	Oct.	24	05	15	36.3 ,	139.8	Gumma Pref.	45.9	40	9.0	390	8.6	K
3	1954	Feb.	25	02	28	35.9 ,	139.9	Chiba Pref.	45.9	40∼50	8.9	390	8.6	K
4	1950	Sept.	10	10	07	35.3 ,	140.5	Kujukuri-Hama	46.5	30∼40	8.5	440	9.5	K
5	1943	Jan.	20	06	24	36.1 ,	140.5	Ibaraki Pref.	50.2	40	9.5	448	8.9	K
6	1952	Nov.	2	10	43	36.2 ,	140.6	Ibaraki Pref.	50.3	45∼50	9.1	458	9.1	K
7	1954	July	18	07	36	35.6 ,	141.0	near Choshi	50.5	40	9.0	485	9.6	K
8	1942	Feb.	19	01	52	35.8 ,	140.8	〃	52.1	—	9.5	470	9.0	K
9	1938	Oct.	29	22	11	35.4 ,	141.0	Kashima-Nada	55.2	—	9.7	482	8.7	K
10	1953	May	11	04	50	36.2 ,	141.1	〃	59.2	40	10.5	510	8.6	K
11	1953	Apr.	4	14	53	35.8 ,	141.9	〃	61.3	40	11.4	570	9.3	K
12	1946	Apr.	3	22	07	35.9 ,	141.3	〃	61.4	40	10.7	517	8.4	K
13	1927	Oct.	11	10	16	—		〃	63.4	—	11.1	—	—	K
f-Group														
1	1929	May	22	01	30	31.8 ,	131.8	Hyuga-Nada	47.1	—	13.2	514	10.9	K
2	1931	Nov.	2	19	09	32.4 ,	132.1	〃	47.4	—	12.6	448	9.5	K
3	1948	May	9	11	09	32.0 ,	131.5	〃	50.0	—	12.6	520	10.4	K
g-Group														
1	1933	May	24	01	38	31.4 ,	131.7	Hyuga-Nada	27.7	—	8.0	203	7.3	Aso
2	〃	〃	〃	〃	53	31.2 ,	131.4	〃	29.3	—	8.5	225	7.7	Aso

deduced from the seismometric investigation in the past more than forty years, the most important and promising problem concerned is the complexity, the regional difference or, in other words, the local character of the earth crust. On the thickness of the crust, the number of discontinuities in the crust, the velocity of the seismic wave, elasticity and density in the crustal layer, all are diverse from region to region. In the present article the durations of refracted waves (forerunners) propagated along the crustal discontinuities are analysed in detail and the regional differences in the crustal structure in Japan are discussed.

2. Forerunner of long duration

In the previous Part I (**43**), the structure of the earth's upper layer (the sedimentary and granitic layers) was determined by an analysis of the refracted seismic waves (forerunners) of short duration (less than 2 sec) recorded mainly at the Kamigamo Geophysical Observatory with regard to the local and near earthquakes of shallow origin around Kyoto. The structure thus determined around Kyoto was as follows:

Layer	Velocity of P-wave km/sec	Velocity of S-wave km/sec	Layer-thickness km
Sedimentary layer	3.8	2.1	1.5
Granitic layer	5.4	3.1	6.0
Granodioritic layer	5.8	3.3	?

In the present investigation the forerunners of longer duration (2～13 sec) were analysed from the seismograms recorded with the large Wiechert Seismographs at the Kamigamo Geophysical Observatory (in the main), the Abuyama Seismological Observatory, the Aso Volcanological Laboratory and the Beppu Balneological Labora-

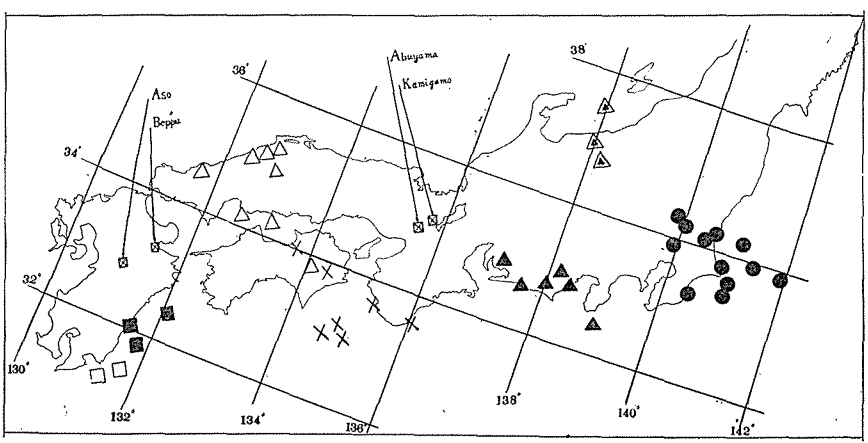

Fig. 1. Positions of the epicenters of earthquakes listed in Table 2 and the Observatories.

tory, all attached to Kyoto University. And from their analyses the structure of the deeper layers of the crust was derived and discussed in some detail.

The observational data of earthquakes used in the present analysis are listed in Table 2, and their epicenters and the referred Observatories are shown in Fig. 1.

Referring to Table 2 and Fig. 1, the duration of each earthquake is plotted, in Fig. 2, with their $(P\sim S)$ times as the abscissa in different symbols (▲, △, ◬, ×, ●, ■ and □) according to their respective groups (a, b, c, d, e, f and g). As seen in Fig. 2 it is considered to be convenient to treat them separating in four classes, namely the first class (Group-a, b and c), the second class (Group-d and g), the third class (Groupe-e) and the fourth class (Group-f) respectively.

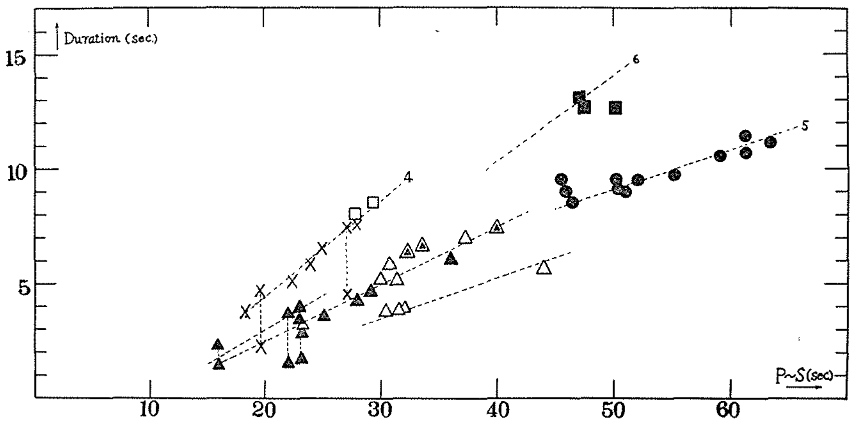

Fig. 2. Relation between the duration of forerunners and the $P\sim S$ time for each Group.

1. On the first class

Group-a, -b and -c contain shallow earthquakes whose epicenters were in the region of Tôkai District (East from Kyoto), Chugoku-Setoumi District (West from Kyoto), and Hokuriku District (North from Kyoto) as shown in Fig. 1. And concerning forerunner duration they are considered to be attributable, roughly speaking, to the same class as seen in Fig. 2. Especially they ought to be clearly distinguished from those of the second class of Group-d. But, treating on the forerunners of the first class in detail, it is convenient to subdivide them into three sub-classes of X, Y and Z, as in Fig. 3, according to their focal depths. Namely, the sub-class X contains three earthquakes of comparatively long forerunner duration and their focal depths are roughly estimated to be within 10 km by the Central Meteorological Observatory (C.M.O.) in Japan. And the sub-class Y contains the major part of earthquakes in the first class, their focal depths being estimated by C.M.O. to be some 20 km. The sub-class Z contains four earthquakes of comparatively short duration and their focal depths are estimated by C.M.O. to be more than 30 km. In this place it is appropriate to mention briefly the problem on the estimation of the focal depth of an earthquake. Generally speaking, it is a very difficult task to determine an accurate value for the focal depth of an earthquake, and especially in Japan it is not too much to say that it is impossible for the following three reasons. First, for the purpose of accurately determining the focal depth, it is necessary to determine accurately the position of the epicenter and the arrival time of the *P*- and *S*-waves at the epicenter. Nevertheless the accurate determination of the epicenter is generally very difficult in Japan because the epicenters of most of the earthquakes occurring in Japan are situated at the open sea or near the sea coast, and therefore there is no way but to estimate roughly the epicenter position from the data obtained at some observatories which are considerably remote from the epicenter and moreover situated in some limited azimuths seen from the epicenter. Secondly, the focal depth calculated from the arrival time of *P*- and *S*-waves at the epicenter depend, needless to say, upon the crustal structure of the epicentral region. That the calculated value of the focal depth depends on the crustal structure and that the crustal structure is determined by the analysis of seismic wave of known focal depth is an endless circular argument, and unfortunately in the case of our country the situation becomes more and more unfavourable because of the local character of the crustal structure. And as the third trouble, the hypocenter of an earthquake is never a point in a geometrical sense but has some extent, and it is reasonably considered that its extention is proportional to the scale of earthquake. And really the order of magnitude of several kilometers is presumably assigned to the extent of hypocenter in case of a large earthquake. From these three and other reasons it may safely be said, except for a favourable case of inland earthquake in the region of simple crustal structure observed by many observatories of dense and uniform distribution near the epicenter, that regarding the value of the focal depth determined from a shallow earthquake, an inevitable error of several kilometers should be borne in mind. And in the present investigation the values of focal depth determined by the Central Meteorological Observatory in Japan were tentatively adopted except a case of special mention, allowing an inaccuracy of several kilometers.

But to return to the main problem, the forerunners of sub-class X, Y and Z should be individually investigated. On the sub-class X it is characteristic that three earthquakes belonging to this sub-class showed two prominent forerunners. Namely, the first forerunner (Pb) has, as seen in Figure A (a), a comparatively long period of oscillation and small amplitude, and the second forerunner (Pg_2) shows a short period of oscillation and small amplitude which is succeeded by the clear arrival of direct wave of large amplitude (Pg_1). As seen in Fig. 3 the duration of the second forerunner is just on the line of duration of the forerunner treated in the previous Part I, therefore the second forerunner is naturally interpreted as the refracted wave transmitted along the discontinuous surface between the granitic layer and the granodioritic layer which will hereafter be designated as 'the *g–g* discontinuity'. And the first forerunner in this case is considered to be the refracted wave transmitted along the discontinuous surface between the granodioritic layer and the basaltic layer which will hereafter be called 'the *g–b* discontinuity'. Therefore the foci of these three earthquakes of very shallow origin are considered to lie in the granitic layer, and the first and second forerunners are interpreted as the refracted waves through the *g–b* discontinuity and *g–g* discontinuity which preceded the direct wave through the granitic layer. Hence, as a natural consequence, the forerunners of sub-class X will serve the purpose of determining the thickness of granodioritic layer and the velocity of basaltic layer, but the numbers of earthquakes in the X–class are too scanty to calculate accurately the required quantities. And this problem will reasonably be transferred from the scanty X–class to the abundant Y–class.

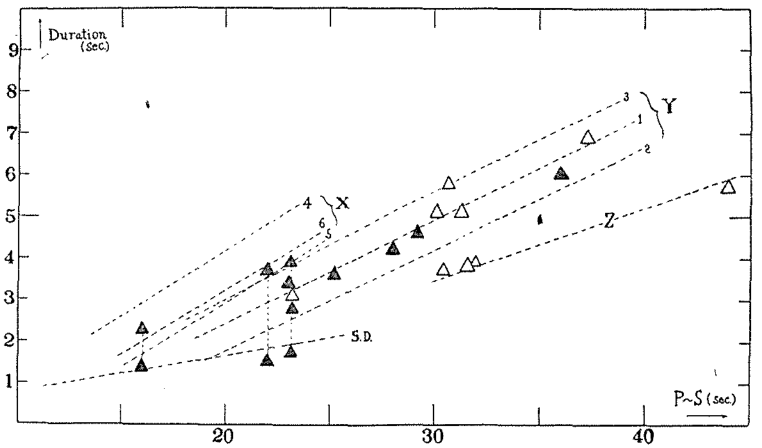

Fig. 3. Relation between the duration of forerunners and the $P{\sim}S$ time for the first class.
S. D. denotes the duration curve obtained in Part I.

As seen in Figure A (b), each earthquake grouped in the sub-class Y is preceded by only one forerunner of long duration (3∼7 sec in the range of epicentral distance from 180 km to 300 km). From these points it is deduced that the foci of earthquakes of Y–class lie in the granodioritic layer and the forerunner observed is interpreted as the refracted wave transmitted along the surface of the *g–b* discontinuity. It ought to be mentioned here that, as will be described in a later paragraph, the refracted wave transmitted along the discontinuous surface between the basaltic layer and the upper part of the mantle which will be called 'the Mohorovičić discontinuity' or simply '*M*-discontinuity', shall clearly precede the above-mentioned forerunner of *Y*-class at the larger epicentral distance beyond some 300 km. For the purpose of determining the seismic velocity in the basaltic layer a similar method to that applied in the previous Part I was used here. Namely, of the earthquakes in *Y*-class the diagram of the forerunner's duration-($P \sim S$) time in Fig. 3 was transformed into an ordinary travel time-distance curve shown as Fig. 4. In this transformation the travel time of the direct wave was calculated under the supposition that the seismic focus lay in the granodioritic layer and at a point 10 km below the *g-g* discontinuity. And it was already discussed in Part I that the difference in the tentatively assumed value of focal depth has little influence upon the calculation of the travel time of a direct wave at a station of considerable distance from the epicenter. Then the velocity of the *P*-wave in the basaltic layer is deduced as 6.8~7.2 km/sec from the inclination of the travel time curve of the refracted wave in Fig. 4. And this sort

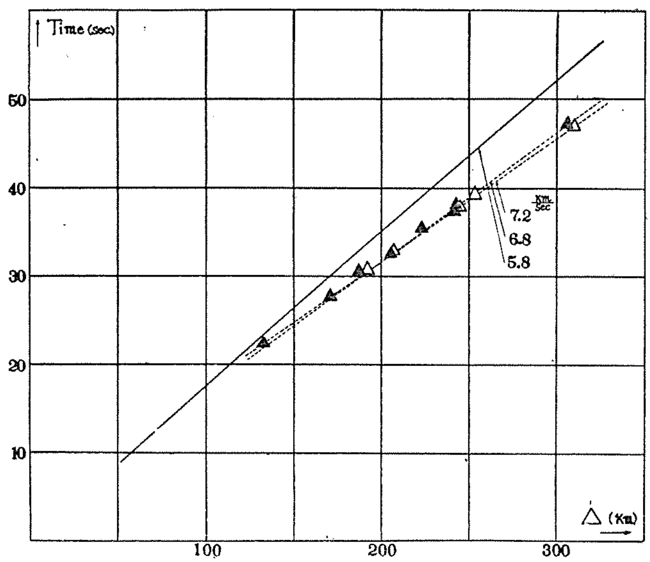

Fig. 4. Travel time-distance curves for Pg_2 (full line) and Pb (broken lines).

of inaccuracy in the value of velocity determined is considered to be inevitable because the foci of earthquakes in the *Y*-class may be at various depths in the granodioritic layer and the number of earthquakes used in calculation is not large enough to smooth the individual effect caused by the difference of focal depth. As an example the travel time-distance curve of an earthquake which occurred on May 30, 1953 at the central part of Setoumi is plotted in Fig. 5. This earthquake belongs to the *Y*-class and its focal depth was estimated to be 20 km, and the velocity of refracted wave is roughly calculated, as seen in Fig. 5, to be 7.0 km/sec. From these factors, the value of the P-wave in the basaltic layer (logically speaking, the layer name of 'basaltic' should be designated after determination of velocity of its layer) was tentatively estimated to be 7.0±0.2 km/sec and the velocity of S-wave may reasonably be assumed to be 4.0 km/sec under the assumption on the value of Poisson's ratio as 0.26.

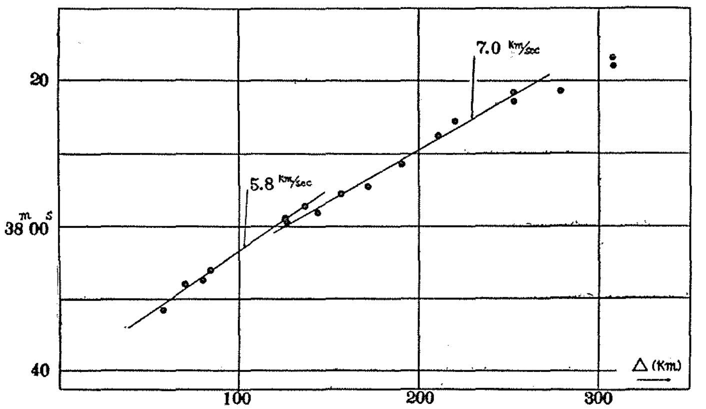

Fig. 5. Time-distance curve (P-wave) of the earthquake on May 30, 1953.

Concerning the thickness of the granodioritic layer the following procedure was applied to the forerunners belonging to the sub-class Y. Under the assumption that the granodioritic layer is approximately a horizontal layer around Kyoto, the most plausible value of thickness determined to fit the durations of all forerunners of the sub-class Y in Fig. 3 was estimated to be nearly 14 km, and the calculated curve for the duration in case of 14 km-thickness of the granodioritic layer and placing the focus at the middle point in the granodioritic layer is shown as curve 1 in Fig. 3. And every point which deviates more or less from the curve 1 is all included in a narrow zone between the curves of 2 and 3 calculated in case of a 14 km-thickness of the granodioritic layer and placing the focus at the uppermost and lowest positions in the granodioritic layer. From these considerations the value of 14 km for the thickness of the granodioritic layer is considered to be the most favourable for the available data of the sub-class Y.

Returning again to the sub-class X the durations of three forerunners in sub-class X are included in a zone of the curves of 4 and 5 in Fig. 3 which are calculated under the assumption that the focus is at the lowest point in the granitic layer and the thicknesses of the granodioritic layer are 10 km and 15 km respectively. Also in this case the value of the granodioritic layer of 14 km is estimated as the most plausible mean value which is expressed as curve 6.

On the last sub-class Z their durations are, as seen in Fig. 3, and Figure A(c), considerably short compared with the other sub-classes of X and Y. And from the fact that their focal depths were estimated to be some 30～40 km, their foci being considered to be in the basaltic layer, and then their forerunners are interpreted as the refracted waves transmitted along the Mohorovičić discontinuity. As will be discussed in detail in a later paragraph, the thickness of the basaltic layer is supposed to vary widely from region to region in Japan. Therefore the individual crustal structure in each region could not be determined from the present method of forerunner analysis, and should be examined by the analysis of the reflected waves of near earthquakes having their epicenters in the respective regions. From the analysis of forerunners only the *mean* crustal structure is determined under the assumption of horizontally parallel structure of layers.

As already discussed or will be further discussed in a later paragraph, the velocities of the basaltic layer and the surface layer of the mantle are approximately estimated as 6.8～7.2 km/sec and 8.0～8.2 km/sec respectively. And the thickness of the basaltic layer derived from the analysis of forerunner duration under the assumption of horizontally parallel structure depends on the value of velocity of basaltic layer and the mantle surface and moreover the position of focus in the basaltic layer in the present case of sub-class Z. Taking the circumstances into consideration, some values for the thickness of the basaltic layer were calculated to fit the observed durations of forerunners in sub-class Z under various plausible assumptions on the layer velocity and the focal depth. The mean value of thickness of the basaltic layer thus obtained is estimated as nearly 14 ± 3 km in the present case of sub-class Z.

In concluding the section *1* on the forerunner of the first class the following structures were determined. The thickness of the granodioritic layer is considered to be uniform, as a first approximation, around Kyoto whose value is estimated about 14 km. The velocities of the P- and S-waves of the basaltic layer underlying the granodioritic layer around Kyoto were calculated as 7.0 ± 0.2 km/sec and 4.0 ± 0.1 km/sec respectively. But on the thickness of the basaltic layer it is appropriate to presume the regional difference which will be discussed in detail in the next section *2*.

2. On the second class

The forerunners of earthquakes belonging to the Group-d are those observed at the Kamigamo Geophysical Observatory, whose epicenters are in the district of the southern part of Kii Peninsula, the south-eastern part of Shikoku and the neighbouring open sea, all being included, roughly speaking, in the district of Nankaido. (The forerunners belonging to the Group-g are those observed at the Aso Volcanological Laboratory and whose epicenters are in the district of the southern part of Hyuga-Nada as shown in Fig. 1.) Those forerunners have considerably longer durations compared with those of the first class and should be separately treated with the forerunner of the first class.

As a first step the forerunners of Group-d will be treated. The seismograms obtained of the earthquakes in the Group-d show, as seen in Figure A(d), two distinct forerunners. And the durations of the second forerunners are plotted just on the mean curve in case of the sub-class Y in the first class as seen in Fig. 2. Moreover the foci of earthquakes in Group-d are estimated as some $10\sim20$ km deep, therefore the second forerunners observed of Group-d are interpreted as the refracted wave transmitted along the *g–b* discontinuity when the focus is in the granodioritic layer. Here another proof on the uniform structure of the granodioritic layer around Kyoto was obtained from the above consideration. Then as what sort of forerunner should the first forerunner of the Group-d be interpreted? Why isn't the forerunner corresponding to the first forerunner of Group-d observed in case of sub-class Y in the first class? The problem could not be solved unless the regional difference of the thickness of the basaltic layer is postulated. In the following the problem will be treated from the standpoint that the thickness of the basaltic layer or, more generally speaking, the depth of the Mohorovičić discontinuity or, in other words, the thickness of the earth's crust may be widely different from place to place.

In the first class the velocity of the layer underlying the M-discontinuity was tentatively assumed to be about $8.0\sim8.2$ km/sec for the P-wave, and in the present paragraph it will be determined for the Nankaido District by an analysis of the earthquake of October 16, 1951. The earthquake in question occurred at open sea off Muroto-Misaki, in the Nankaido District and its focal depth was estimated to be about 40 km. The fact that the seismogram obtained of this earthquake at the Kamigamo Geophysical Observatory was not preceded by any forerunner, as shown in Figure B(a), tells us that its focus was at the point under the M-discontinuity and moreover depth of the M-discontinuity around this district was estimated as less than 40 km from the ground surface of the sea bottom. On such a method of locating the focal depth of the earthquake by the examination of forerunner some detailed discussion will be made in a later paragraph. From the analysis of the above-mentioned earthquake the apparent velocity of the layer under the M-discontinuity in this district was estimated to be about $8.0\sim8.2$ km/sec as seen in the travel time-distance curve of Fig. 6. From this value it was concluded that the velocity of the layer underlying the M-discontinuity or, in other words, the surface velocity of the mantle is nearly $8.0\sim8.2$ km/sec all for the District of Tôkaido, Hokurikudo, Chugoku and Nankaido.

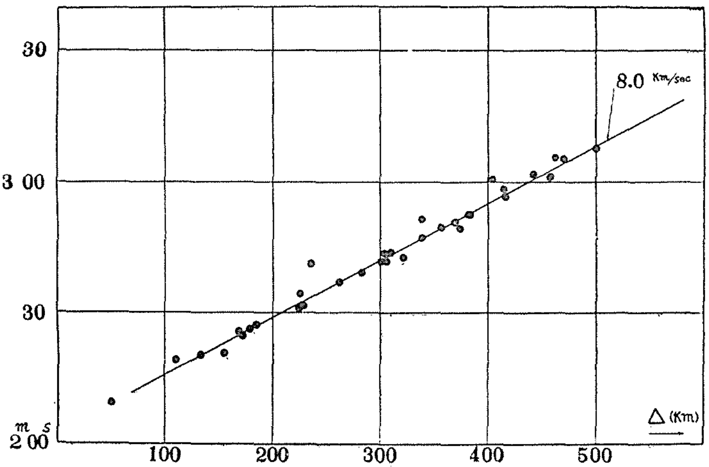

Fig. 6. Time-distance curve (P-wave) of the earthquake on Oct. 16, 1951.

Under the circumstances that the thickness of granodioritic layer is 14 km both for the Nankaido and Kyoto, and moreover the P-velocity in the layer underlying the basaltic layer is of the same value of about 8.0 km/sec for both regions, what value should be attributed to the mean thickness of the basaltic layer under the region between Nankaido and Kyoto? As already stated, the thickness of the basaltic layer under any region is not uniquely determined only from the analysis of an observed forerunner. It depends certainly on the crustal structures of the hypocenter, the observation point and the propagation path of seismic waves. And, moreover, it is generally affected by the assumed value of the propagation velocity of seismic waves in the layers concerned and considerably associated with the orientation of seismic focus. In the present case the most plausible mean value of the thickness of the basaltic layer has been determined to be 6.5 ± 1.2 km under the assumption of horizontally parallel structure under the region between Nankaido and Kyoto, and, moreover, averaging the various values obtained from the some combinations of the velocities for the P-wave in the basaltic layer ($6.8 \sim 7.2$ km/sec), the mantle surface ($8.0 \sim 8.2$ km/sec) and the focal depth in the granodioritic layer. A calculated duration curve for the case of a horizontally parallel structure of the basaltic layer, placing the focus at the middle in the granodioritic layer, and assuming the propagation velocity of 7.0 km/sec for the basaltic layer and 8.2 km/sec for the mantle surface, is expressed as curve 4 in Fig. 2.

Secondly the Group-g will be discussed. This Group contains the earthquakes which occurred in the southern part of Hyuga-Nada, as seen in Fig. 1, and their forerunners were observed at the Aso Volcanological Laboratory. As shown in Fig. 2

the Group-g may be treated approximately in the same manner as the Group-d, in other words, the structural relation of the crust in the southern Hyuga-Nada District to that of Aso is, roughly speaking, similar to the relation between the Nankaido District and Kyoto. The crustal structure around Aso was approximately ascertained by an analysis of somewhat deep earthquakes near Aso as will be discussed in a succeeding article, and the thickness of the earth's crust around Aso is of the same order of magnitude as that around Kyoto. From these it is tentatively stated that the thickness of the basaltic layer in the southern part of Hyuga-Nada is considered to be of the same order of magnitude as in the Nankaido District, namely some several kilometers, which is considerably thin as compared with the other Districts. On this problem the earthquakes having occurred at Hyuga-Nada will be again treated in the Group-f of the fourth class.

In concluding section *2* on the second class it is convenient to mention that the distinct regional difference of the thickness of the basaltic layer was ascertained in this section by analysing the forerunners of earthquakes having occurred in the Nankaido District, and on the other hand, no distinct regional difference was found in the thickness of the granodioritic layer overlying the basaltic layer in the region within some 300 km distance from Kyoto. The thicknesses of the basaltic layer thus determined under the assumption of horizontally parallel structure (which is contrary to the real structure) are roughly estimated to be 6.5 km and 14.0 km for the region of Nankaido-Kyoto and Shimane-Kyoto respectively.

3. On some destructive earthquakes

As discussed in the preceding section *2* the existence of a regional difference in the crustal structure was ascertained from the analysis of the duration of the forerunners of weak or medium earthquakes observed at one station, mainly at the Kamigamo Geophysical Observatory in Kyoto. Too large or destructive near earthquakes are unsuitable for the analysis of forerunners in case of observation with the large Wiechert Seismograph, but for the purpose of accurate determination of the crustal structure the concurrent use of the forerunner with the time-distance curve constructed with the data observed at many stations of a destructive earthquake is the most desirable. In this case the same type of seismograph is recommended for all observatories and their magnifications should be adequately controlled to be able to record ground oscillations of large amplitude at stations near the epicenter and to catch the minute ground oscillations at the distant station. And the accurate time determination for any phase is an important task in this case, and moreover, the identification for any phase is, above all, the most difficult and essential problem. In the present state the above-mentioned conditions are not yet perfectly fulfilled in the seismometric system of our country, but as a great improvement has recently been made in seismometry in Japan, the development of an accurate and detailed seismometric investigation on the regional difference of crustal structure will be expected in the near future. As will be seen in the following treatment of some destructive earthquakes, it is almost impossible, at present, to get a clear picture of the crustal structure from the analysis of the travel time-distance curve only. The difficulty results from the uncertainty of the position of epicenter and hypocenter, the finite extent of seismic focus, the inaccuracy of travel-time determination, the locality of crustal structure around the observatory, the insufficient or excessive magnification of the seismograph, and other artificial or natural sources of errors. Therefore it is recommended that in an accurate seismometric investigation of the structure of the crust or, more generally, the earth's interior, the analysis of travel time-distance curve should be completed by the analysis of forerunner duration, the comparison of observed amplitude of each seismic phase, the analysis of the reflected and the transformed seismic waves, and other available and effective means.

(a) The Fukui Earthquake of June 28, 1948

The Fukui Earthquake of June 28, 1948 was the most severe and destructive in recent years with a damage of twenty thousand casualties and forty thousand ruined buildings. Its epicenter was estimated to be $36°\,08'$ N and $136°\,16.8'$ E, and the seismic magnitude was $7\frac{1}{4}$ in Pasadena Scale. The focal depth was diversely estimated by several investigators, but in our case it is considered to be most plausible to place it in the granodioritic layer as discussed in the succeeding. This earthquake was by far the most completely and precisely investigated compared with other past destructive earthquakes, and its travel time-distance curve is said to be very much trustworthy as it was derived from the accurate location of an inland epicenter, the recent advance of time determination at the seismological observatories and the authority of synthetic investigation by various fields of science (**44**).

In Fig. 7 is shown the travel time-distance curve studied by the writer using the C.M.O.-data, in which the symbols of ● and + denote the arrival time of the first motion of seismic waves observed at the westerly and the easterly observatories from Fukui respectively. And the symbols of ⊙ and ⊕ denote the arrival time of the distinct phases of seismic waves having appeared after the first motion and which were examined by the writer from the available seismograms. From these arrival times three branches of straight lines connecting the points belonging to the same system are tentatively drawn, as seen in the figure, each expressing the apparent velocity of 8.0, 7.0 and 5.9 km/sec respectively. These seismic P–waves of 8.0, 7.0 and 5.9 km/sec–velocity are interpreted respectively as the refracted waves transmitted along the M–discontinuity, the refracted wave along the g–b discontinuity,

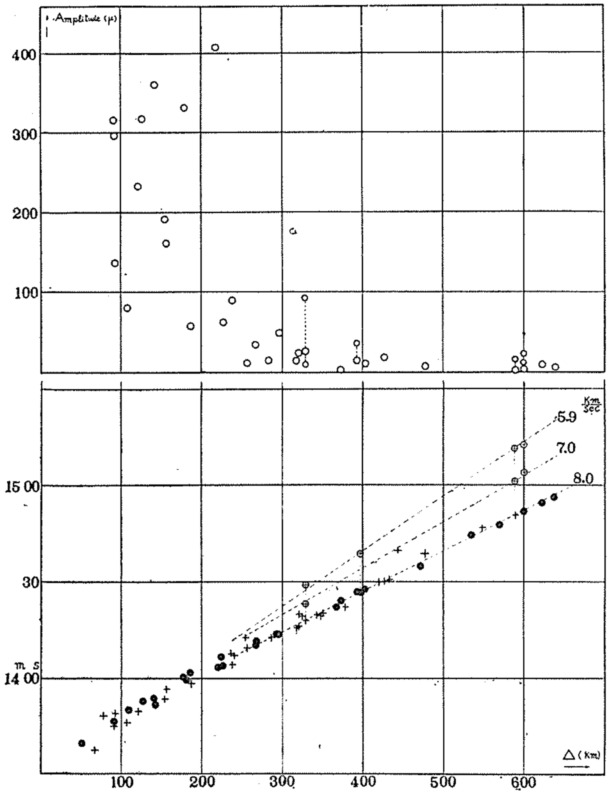

Fig. 7. Time-distance (lower) and amplitude-distance (upper) curves for the P-waves of the Fukui Earthquake on June 28, 1948.

and the direct wave through the granodioritic and granitic layer in case of having the focus in the granodioritic layer. And, moreover, this is ascertained by the observed amplitude relation between three phases mentioned above, namely the observed amplitudes of the seismic waves of first arrival decrease suddenly at the epicentral distance of some 230 km, as expressed in Fig. 7, at the distant places beyond the critical point of 230 km the amplitude of first motion of seismic waves being estimated to be less than one-tenth of those observed at the short distance within the critical point. Beyond the critical point the amplitude relation between the first forerunner, the second forerunner and the direct wave are, as shown also in Fig. 7, in appropriate proportion. It ought to be mentioned here that, in practical cases, the clear separation of three phases can safely be made of the seismograms obtained at places outside the epicentral area of some 300 km. Concerning the problem of placing the focus of the Fukui Earthquake in the granodioritic layer the following examination was also applied. As shown in Fig. 8 the $P\sim S$ times in the ordinate and the epicentral distance in abscissa are expressed of the data observed at the observatories within some 200 km-epicentral distance where the observed seismic waves of first arrival are considered as the direct wave. The two curves of 1 and 2 in the figure are calculated for the case in which the seismic focus is in the granodioritic layer and the basaltic layer respectively. The structure assumed in this case are (5.4 km/sec, 3.1 km/sec, 6 km) for the granitic layer, (5.8 km/sec, 3.4 km/sec, 14 km) for the granodioritic layer and (7.0 km/sec, 4.0 km/sec, —) for the basaltic layer, the foci being in the granodioritic layer 7 km below the *g–g* discontinuity for the curve 1 and in the basaltic layer 6 km below the *g–b* discontinuity for curve 2. Curve 2 is, in this case, not suitable for expressing the plausible curve connecting the data, and we are convinced that the placing of the focus in the granodioritic layer is favourable in the case of the Fukui Earthquake.

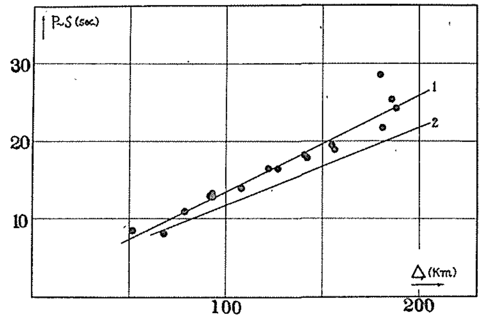

Fig. 8. $P \sim S$ time-distance curve of the Fnkui Farthquake.

Now for the purpose of calculating the thickness of the basaltic layer using the data obtained from the Fukui Earthquake the travel time-distance curve in Fig. 7 will be analysed. But in the present analysis the so-called critical distance (the epicentral distance of the place where the arrival times for both the direct and refracted seismic waves are the same) was not used for the reason that the value of the critical distance could not usually be determined with sufficient accuracy in the analysis of natural earthquakes contrary to that of artificial explosions. In cases of the analysis of natural earthquake including data obtained from such large epicentral distances as several hundreds kilometers, this is due to the fact that the determination of the critical point is greatly disturbed by the local character and regional difference of crustal structure around the respective observatories, and by the inaccurate estimation of the position of seismic focus. Therefore, in the present case, the durations of the first and second forerunners were used to calculate the thickness of the basaltic layer. This method, needless to say, contains also the inaccuracy caused by the same disturbances mentioned above, but it will soon be understood that the loss in this case is very much small compared with that in case of the critical distance. Speaking more in detail, the problem of critical distance concerns the data in a comparatively short distance and that of forerunner mainly the data at far distance, therefore the local and regional difference of the crustal structure, the instrumental and time errors and other disturbances become more and more prominent, as seen in Fig. 7, and greatly spoil the calculated value of layer-thickness in the former case. But, also in the latter case, the obtained values of apparent velocities of the refracted and the direct seismic waves, and the thencefrom derived values of the layer-thickness could not be free, in some measure, from the above-mentioned disturbances, and therefore the estimated values of the velocity and the layer-thickness in the present investigation are considered to be, roughly speaking, in the degree of accuracy with the mean error of ± 0.2 km/sec for the velocity and ± 5 km for the layer-thickness respectively. Considering these circumstances the most plausible structure derived from the forerunners of the Fukui Earthquake is the basaltic layer of 7.0 km/sec–P velocity and some 20 km–thickness, and the velocity of 8.0 km/sec–P wave for the layer underlying the M–discontinuity when assuming the granodioritic layer of 5.8 km/sec–P velocity and 14 km–thickness, and placing the seismic focus at the middle point of the granodioritic layer. As already discussed in the preceding paragraph the thicknesses of the basaltic layer were estimated to be diverse from place to place, and accordingly the above obtained value of some 20 km–thickness of the basaltic layer is considered as a roughly averaged value for a wide area. Therefore, in order to obtain the accurate structure of the crust around any region, the travel time-distance curve and the forerunner should concurrently be analysed with regard to the data observed at the observatories in any one direction and within some suitable distance from the epicenters of a swarm of earthquakes in the same locality. When the forerunner durations obtained at the observatories in the Kanto district are especially analysed, the averaged value of the horizontally parallel structure of the basaltic layer under the region between Fukui and Kanto is estimated, as in the former case, to be 18.7 ± 1.5 km.

Simply speaking, the crust is considered to be shallow, mediate and deep under the region of Nankaido-Kyoto, Shimane-Kyoto and Kanto-Fukui respectively.

(b) Travel-time distance curves for some other destructive earthquakes

In the above paragraph the travel time-distance curve obtained from the data of the Fukui Earthquake of June 28, 1948 which is considered as the most reliable in recent times in Japan was analysed and it was concluded that even the most trustworthy travel time-distance curve as in case of the Fukui Earthquake gives only a roughly averaged value for the crustal structure. In the present paragraph the travel time-distance curves are presented for some destructive earthquakes having occurred in Japan in past twenty years. In Table 3 the materials obtained by the C.M.O. of these destructive earthquakes are listed and their epicenters are shown in Fig. 9. Their tentatively drawn travel time-distance curves are plotted in Figs. 10～17.

Table 3. List of the destructive earthquakes

No.	Earthquake	Date	Magnitude in Pasadena Scale	Epicenter		Focal depth
				° E	° N	km
1	Kita-Izu	Nov. 26, 1930	7.1	139.0 ,	35.1	—
2	Sanriku	Mar. 3, 1933	8½	144.7 ,	39.1	—
3	Hyuga-Nada	Nov. 19, 1941	7.1	132.4 ,	32.3	—
4	Tottori	Sept. 10, 1943	7½	134.2 ,	35.5	0
5	Tonankaido	Dec. 7, 1944	8	136.2 ,	33.7	0~30
6	Mikawa	Jan. 13, 1945	7.1	137.0 ,	34.7	0
7	Nankaido	Dec. 21, 1946	8.2	135.6 ,	33.0	0~30
8	Fukui	June 28, 1948	7¼	136 17′,	36 08′	20
9	Tokachi-Oki	Mar. 4, 1952	8¼	143 52,′	42 09′	45

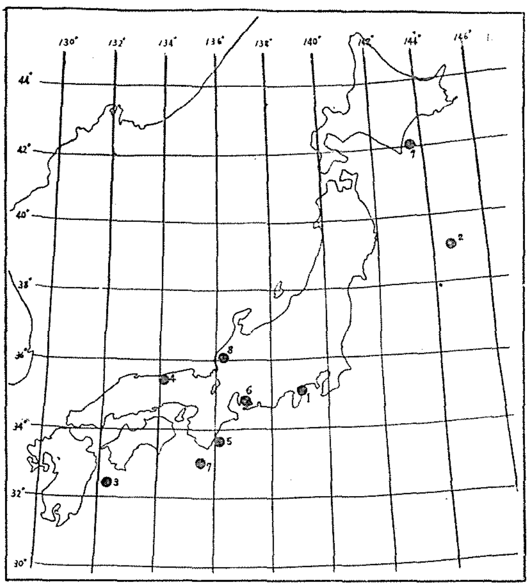

Fig. 9. Positions of nine destructive earthquakes listed in Table 3. The annexed figures are referred to Table 3.

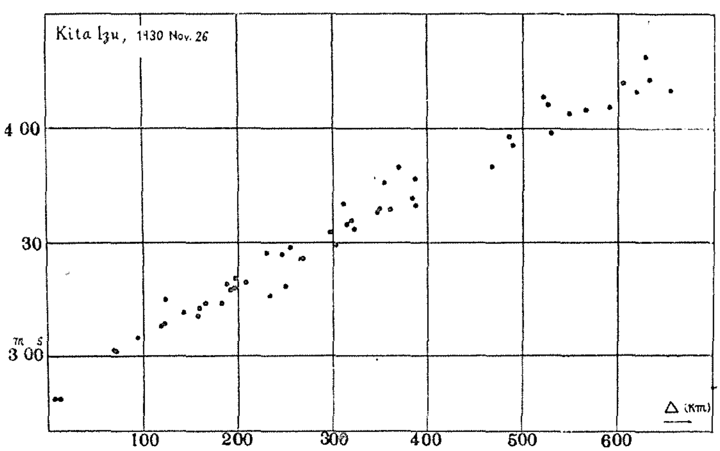

Fig. 10. Time-distance curve (P-wave) of the Kita Izu Farthquake.

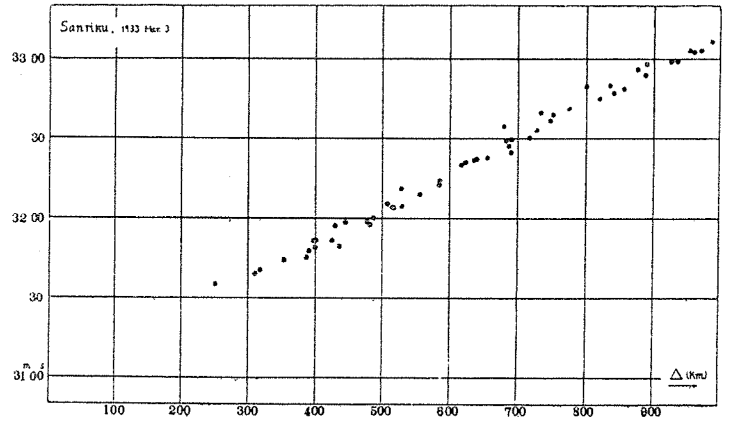

Fig. 11. Time-distance curve (P-wave) of the Sanriku Earthquake.

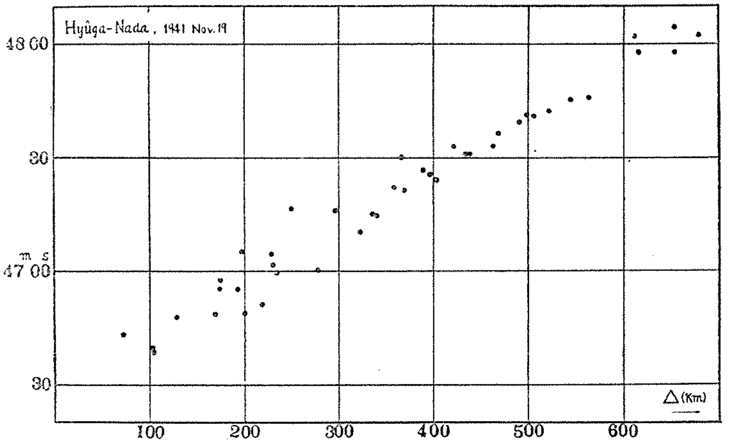

Fig. 12. Time-distance curve (P-wave) of the Hyuga-Nada Earthquake.

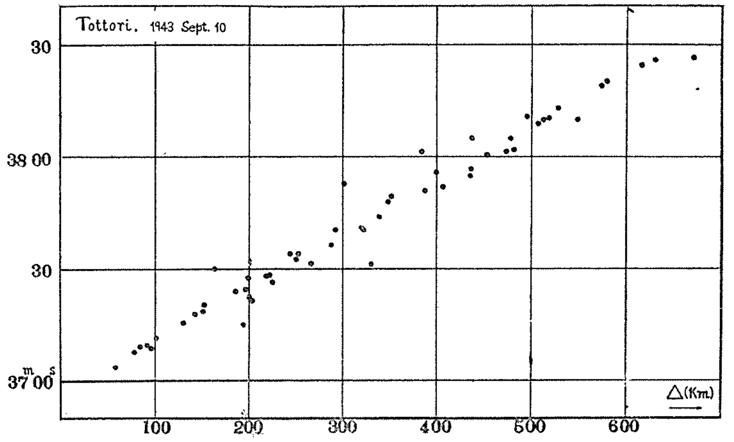

Fig. 13. Time-distance curve (P-wave) of the Tottori Earthquake.

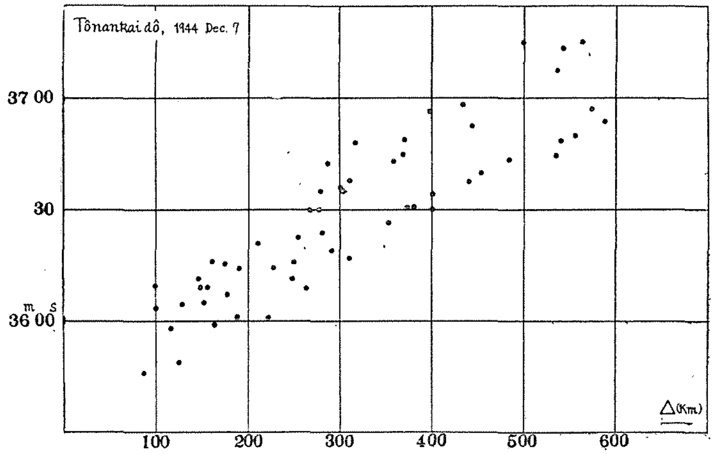

Fig. 14. Time-distance curve (P-wave) of the Tonankaido Earthquake.

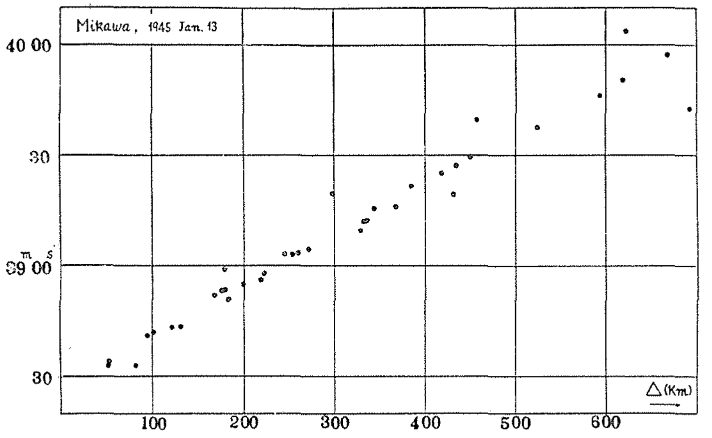

Fig. 15. Time-distance curve (P-wave) of the Mikawa Earthquake.

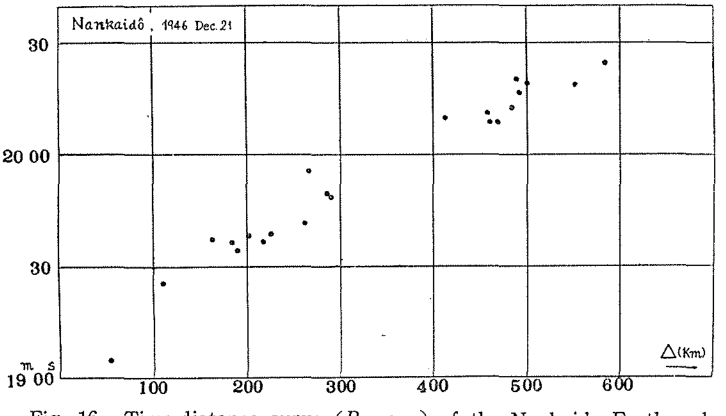

Fig. 16. Time-distance curve (P-wave) of the Nankaido Earthquake.

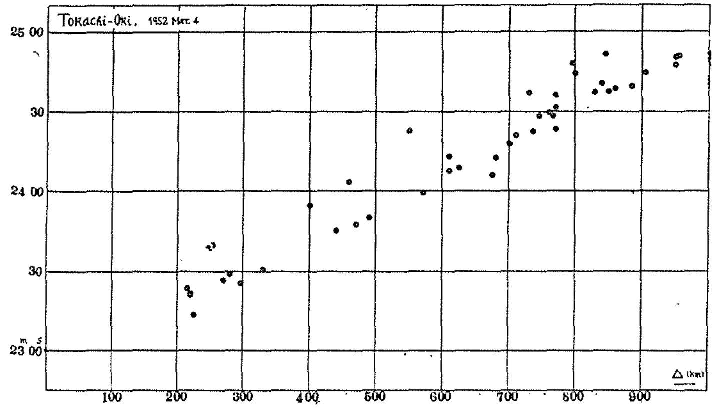

Fig. 17. Time-distance curve (P-wave) of the Tokachi-Oki Earthquake.

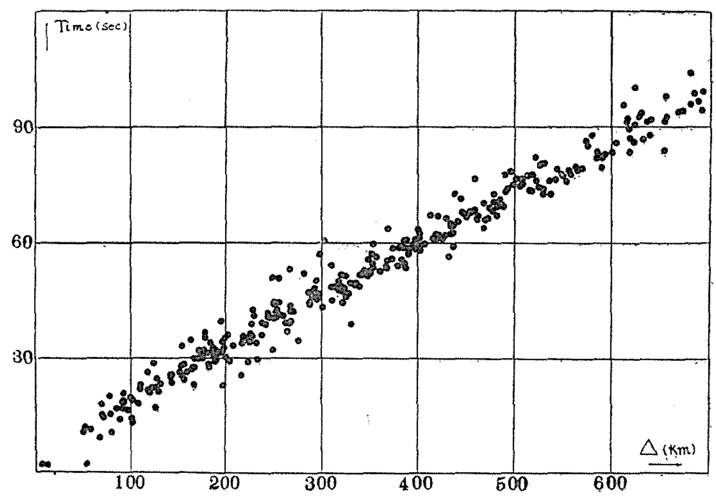

Fig. 18. Superposed time-distance graph (P-wave) of nine destructive earthquakes listed in Table 3.

Comparing these travel time-distance curves derived from nine destructive earthquakes which occurred in the last twenty years in Japan, the difficult circumstances are generally understood for the precise determination of the crustal structure from the analysis of these travel time-distance curves. The difficulties are considered to originate, as already discussed, from the regional differences of the crustal structure, the inaccuracy of time determination, the uncertainty of focal depth and other various disturbances. In a rough treatment all the curves were gathered, as a trial, in a diagram as seen in Fig. 18. In this figure the apparent velocities of the P-wave are roughly estimated as 6.2 km/sec and 8.0 km/sec for the direct and refracted waves respectively, and the critical distance is also tentatively calculated as 150 km. And these velocities of 6.2 km/sec and 8.0 km/sec are considered to be the mean values of the granodioritic and basaltic layers for the former and that of the mantle surface for the latter respectively. In case of rough calculation of the depth of the M-discontinuity under this assumption of one layer structure of the crust, what value of the thickness of the crust will be derived from the above obtained travel time-distance curve superposed with nine destructive earthquakes of various origins? The value of some 40 km is obtained with regard to the depth of the M-discontinuity, in other words, the thickness of the crust, under the assumption that the mean focus is at a point 15 km from the ground surface in the 6.2 km/sec-layer. Under various circumstances it is reasonably presumed that the value of some 40 km denotes approximately the averaged thickness of the crust in Japan.

4. *On the third class*

The third class consists of the earthquakes of the Group-e, and their epicenters are, as seen in Fig. 1, in the eastern part of the Kanto District and its neighbouring open sea. The focal depths of these earthquakes were estimated by C.M.O. as 40∼50 km below the ground surface, and deduced from the observed fact of existence of forerunners, shown in Figure B(b), their foci may correctly be placed in the basaltic layer. In this case the precise structure of the upper crust, namely the granitic and granodioritic layers, was not yet determined by the forerunner analysis method in the same sense of our present investigation. But from the crustal structure obtained by T. Matuzawa (**26, 27, 28, 29**) and the velocities of the basaltic layer and the surface layer of the mantle determined from the superposed travel time-distance curve, as shown in Fig. 19, of thirteen remarkable earthquakes of Kashima-Nada's origin in this Group-e, the most plausible curve 5 for connecting the durations of forerunners of the third class is drawn, as in Fig. 2. The curve 5 is calculated under the conditions that the velocity of the P-wave in the basaltic layer is 6.8 km/sec, that in the surface layer of the mantle 8.2 km/sec, the thickness of the basaltic layer of horizontal structure between the region of Kanto-Kyoto is 18 km. It ought to be mentioned here that the horizontally parallel structure of the basaltic layer between Kanto and Kyoto assumed above is adopted simply for the sake of convenience, and a conclusive answer to this problem deduced from the detailed investigation on regional differences in the crustal structure of each locality is eagerly awaited, because it is possible to imagine that the mountainous region in the middle of Kanto and Kyoto will show a considerably large anomaly on the depth of the M-discontinuity. And moreover a value of some 52 km for the thickness of one layered crust is deduced from the data in Fig. 19, that is 7.3 km/sec, 7.9 km/sec and 300 km for the P-velocity in the crust and the mantle, and the critical distance respectively.

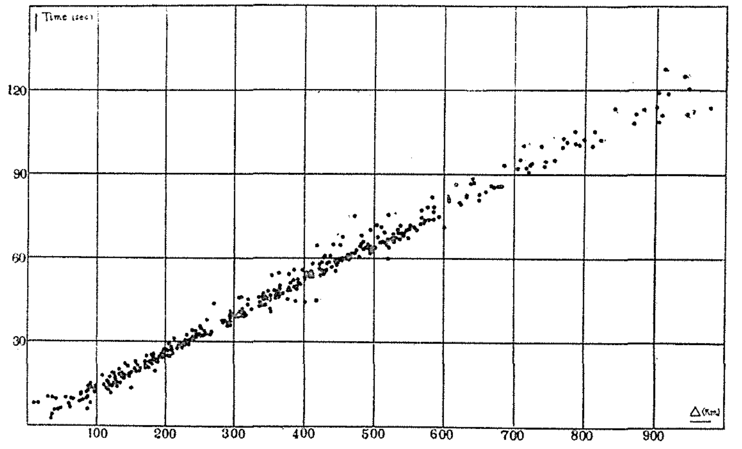

Fig. 19. Superposed time-distance graph (P-wave) of thirteen earthquakes of the third class (e-Group) listed in Table 2.

And it is desirable that a definite conclusion should reasonably be postponed to the day of the accomplishment of a precise and synthetic study on the forerunner and travel time-distance curves of the earthquakes in the Kanto District as deduced from the accurate observed data obtained at some reliable observatories in the same district.

5. On the fourth class

This class contains the earthquakes of Group-f whose epicenters are in the Hyuga-Nada District, as shown in Fig. 1. The epicenters and focal depths of earthquakes belonging to this Group are not accurately determined because their epicenters are generally off in the open sea of Hyuga-Nada and the surrounding observatories are scanty in number and too distant from the origin. But to fit the durations of the forerunners plotted in Fig. 2, it is impossible to place the focus in the basaltic layer no matter what structure the crust might have in a reasonable range. And, as seen in Fig. 20 which is the superposed travel time-distance curve of ten Hyuga-Nada earthquakes, the velocity of the direct wave and the mantle-surface are approximately calculated as 6.0 km/sec and 7.9 km/sec for the P-waves respectively. From these facts the most plausible curve 6 is tentatively drawn, as seen in Fig. 2, under the conditions that the focus is at the middle of the granodioritic layer and the P-velocities of the basaltic layer and the mantle-surface are 7.0 km/sec and 8.1 km/sec respectively. And, moreover, the thickness of the basaltic layer is assumed as horizontally parallel and 15 km for the region between Hyuga-Nada and Kyoto. But this consideration on the crustal structure at Hyuga-Nada, needless to say, should reasonably be postponed till a more precise investigation be made in the future of the data obtained at the observatories near this Hyuga-Hada District. Also in this case, the averaged value of rough estimation on the crustal thickness referred to the materials in Fig. 20 is calculated as 23 km under the condition that the focus is at the middle of the one layered crust, and the P-wave-velocities of the crust and the mantle-surface are 6.0 km/sec and 7.9 km/sec respectively, and the critical distance is 100 km. These considerations referring to the travel time-distance curve are not expected to draw any definite value for the crustal thickness, but it may safely be said that, under the assumption of one layered crust, the value of some 23 km obtained from the superposed travel time-distance curve of the Hyuga-Nada earthquakes is distinctly smaller than that (52 km) of the Kashima-Nada earthquakes as derived from Fig. 19, and thencefrom an essential difference of the crustal thickness can be detected and discussed to a certain degree by the comparison of the materials derived from the superposed travel time-distance curve around any two regions with different crustal structures, considerably distant and diverse from each other.

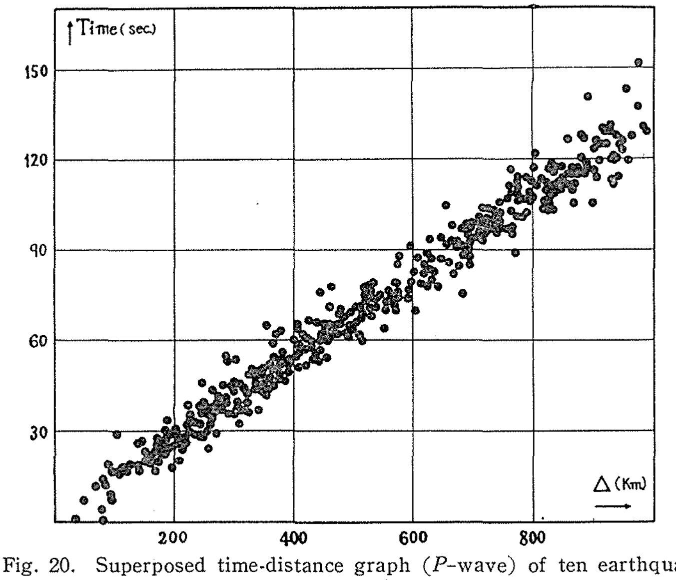

Fig. 20. Superposed time-distance graph (P-wave) of ten earthquakes of the fourth class (f-Group).

In concluding the section 2 of "Forerunner of long duration" the following may be summarized: from the analysis of the forerunners supplemented with the materials of travel time-distance curves the crustal structures under various regions in Japan, especially detailed in the Kinki District around Kyoto, were determined and discussed. And from them the existence of large regional differences in the crustal structure in Japan was generally ascertained. In the succeeding section, the problem of these regional differences in crustal structure will be ascertained by other means.

3. Thickness of the basaltic layer

1. Thickness of the basaltic layer under Kyoto

In the preceding paragraphs the existence of regional differences in crustal structure was ascertained by comparing the durations of refracted waves of earthquakes having occurred in various regions. But, in that case, only the mean thickness could be discussed, as already mentioned, under the unpractical assumption of the horizontally parallel structure of the basaltic layer. In order to determine the absolute value of thickness of the basaltic layer under any point, the reflected or transformed seismic waves of near earthquakes to that point should be investigated. On these two methods the detailed treatment will be postponed to a succeeding paper, and in the present place the transformed seismic waves of the earthquakes near Kyoto of somewhat deep focus are tentatively examined for determination of the thickness of the basaltic layer around Kyoto.

In the present treatment the data concerning the various layers around Kyoto are reasonably assumed as follows:

Layer	V_p km/sec	V_s km/sec	Thickness km	Density c.g.s.	Discontinuity
Granitic layer	5.4	3.1	6	2.6	
					………… g–g
Granodioritic layer	5.8	3.4	14	2.7	
					………… g–b
Basaltic layer	7.0	(4.0)	?	3.0	
					………… M
Mantle surface	8.2	(4.7)	—	3.4	

It is to be mentioned here that the values of V_s in parentheses are calculated from the value of V_p taking the value of Poisson's ratio to be 0.26, while the values of density adopted in this table are the most plausible ones obtained by many authors.

The earthquakes used in the present are, as listed in Table 4, those of mediate focal depth (nearly 70 km) whose epicenters were nearly 100 km distant and in the south direction from Kyoto. As shown in Figure C many distinct phases are observed between the arrivals of direct P– and S–waves in their seismograms. As inferred from their focal depths estimated by C.M.O. and the type of their seismograms, namely unaccompanied by forerunners and the lack of predominating surface waves or others which characterize the earthquake of shallow origin, their foci are safely considered to be under the Mohorovičić discontinuity. Taking these conditions into consideration the distinct phases above described may be interpreted as: (a) transformed waves at the discontinuities in the crust, or, (b) reflected waves at the earth surface and again at the discontinuities in the crust, or, (c) transformed or reflected waves in any discontinuities of first order, if exist, in the mantle. In the present treatment the case of (c) is reasonably excluded, and only two cases of (a) and (b) will be discussed. In our investigation the name of 'transformed wave' is designated for the wave which is transformedly refracted from P-wave to S-wave or from S-wave to P-wave at any discontinuous surface, namely 'Wechselwellen'. For the purpose of identification of observed distinct phases the expected amplitude ratio of (a)- and (b)-waves with regard to the earthquakes concerned should be theoretically calculated under the conditions of crustal structure above listed. In this case the propagating seismic wave is assumed to be a plane wave of no damping. And with regard to the g-g discontinuity, the difference in velocity and density is small compared with the other two g-b and M-discontinuities as seen in the above table, therefore the present calculation concerns mainly the g-b and M-discontinuities. The calculated amplitude ratios are as follows, taking the amplitude of the first incident wave as unit:

	Medium	Transformation	Emergent Angle	g-b Discontinuity	M-Discontinuity	After K & S
(1)	hard → soft	$P \rightarrow S'$	26°	0.151	0.158	0.195
(2)	,,	$P \rightarrow P'$	26	—	0.929	0.885
(3)	soft → hard	$P \rightarrow P_1$	54	0.089	—	0.097
(4)	,,	$P \rightarrow P_1$	49	—	0.106	0.105
(5)	,,	$P \rightarrow S_1$	54	0.093	—	0.105
(6)	,,	$P \rightarrow S_1$	49	—	0.19	0.095

Of the above table some remarks should be made. Here 'hard' medium in the first column means that of large velocity and density compared with 'soft' medium. And the 'prime' ($'$) and 'suffix' ($_1$) denote the refracted and reflected waves at any discontinuity respectively. 'K & S' in the last column is abbreviation of H. Kawasumi and T. Suzuki and the numericals in this column are the amplitude ratio calculated by them (**45**). In 'K & S' case the ratios of hard/soft are 1.1 in density, 1.5 in seismic velocity and 0.25 in Poisson's ratio. As both values obtained by 'K & S' and us are roughly in an agreement, the calculated values after 'K & S' may reasonably be used in our investigation as the amplitude ratios in other various cases. And moreover the calculated values after T. Matuzawa (**46**) are also to be applied as the amplitude ratio of the reflected wave at the free surface of the earth. After these considerations, the transformed and reflected waves worth discussing in the present case are as follows:

$$P \rightarrow S'(M), \quad P \rightarrow S'(g-b), \quad S \rightarrow P'(M), \quad S \rightarrow P'(g\text{-}b), \quad P \rightarrow P_1 \rightarrow P_1(FS,\ M),$$
$$P \rightarrow P_1 \rightarrow P(FS,\ g\text{-}b)$$

The other waves except the above may safely be excluded in the present discussion by reason of their too small amplitudes and relations of arrival time. In the

Table 4. List of the earthquakes used for the investigation of the transformed waves

Date (d. h. m.)	Epicenter and focal depth	Epicentral region	$P \sim S$ (s)	Incident phases: Observational No.	Observational Duration (s)	Observational (Compo-nent,	Observational amplitude ratios)	Calculated Phase identification	Calculated Duration (s)	Calculated (compo-nent,	Calculated amplitude ratios)
1953 Sept. 1 23 15	34, 0°N 135, 7°E 70km	Southern part of Nara Pref.	13.2	1	4.0	$(H,$	0.3)	$P \rightarrow S'$ (M)	4.3	$(H,$	0.31)
				2	6.1 6.3	$(H,$ $(V,$	0.29) 0.2)	$S \rightarrow P'$ (M)	6.2	$(H,$ $(V,$	0.27) 0.17)
				3	7.7	$(H,$	0.24)	$P \rightarrow P_1 \rightarrow P_1$ (FS, M)	7.6	$(H,$	0.07)
				4	9.2 9.1	$(H,$ $(V,$	0.37) 0.26)	$S \rightarrow P'$ $(g-b)$	8.8	$(H,$ $(V,$	0.46) 0.25)
1952 Aug. 8 21 41	34, 5°N 135, 8°E 65km	Central part of Nara Pref.	10.2	1	5.2	$(V,$	0.25)	$S \rightarrow P'$ (M)	5.3	$(V,$	0.21)
				2	6.2	$(H,$ $(V,$	0.38) 0.11)	$P \rightarrow P_1 \rightarrow P_1$ $(FS, g\text{-}b)$	6.2	$(H,$ $(V,$	0.06) 0.07)
				3	6.9	$(H,$	0.37)	$S \rightarrow P'$ $(g-b)$	7.0	$(H,$	0.53)

above described waves the figures *M*, *g–b* and *FS* in the parentheses denote the refraction and reflection at the *M*–, *g–b* discontinuity and the free surface of the earth respectively. It is a very interesting and essential problem to discuss in detail the individual waves mentioned above for investigating the crustal structure around Kyoto or other regions, but its treatment will be reserved for another occasion. And here the $P \to S'(M)$ and $S \to P'(M)$ will mainly be discussed. The thickness of the basaltic layer around Kyoto was roughly estimated to be in the range of 5∼15 km after the above described investigation of forerunners, therefore the travel time-distance curves for individual phases are calculated and shown in Figs. 21, 22 and 23 under the assumed thicknesses of the basaltic layer as 5 km, 10 km and 15 km respectively. Comparing these three figures it is natural to conclude that the value of 10 km is the most plausible for the thickness of the basaltic layer under Kyoto. Detailed comparison of the thickness of the basaltic layer under various regions or more general study for the regional difference of the crustal structure by the analysis of the transformed or reflected waves will be postponed to a succeeding investigation.

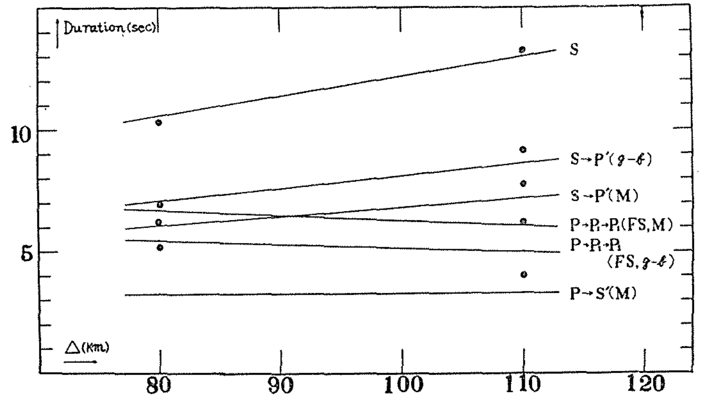

Fig. 21. Durations calculated from the first arrival of *P*–wave of the transformed and reflected waves in case of 5 km thickness of the basaltic layer.

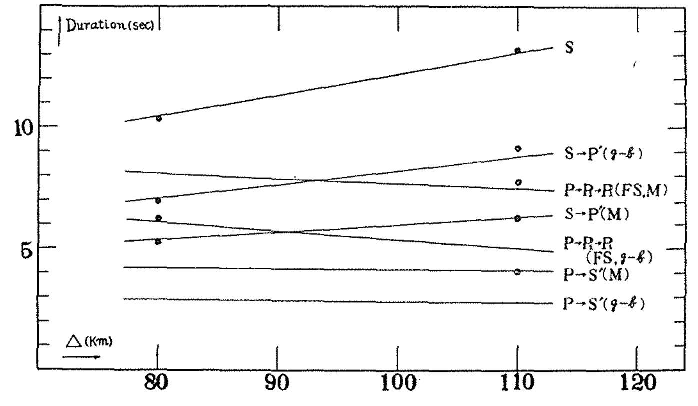

Fig. 22. Durations calculated from the first arrival of *P*–wave of the transformed and reflected waves in case of 10 km thickness of the basaltic layer.

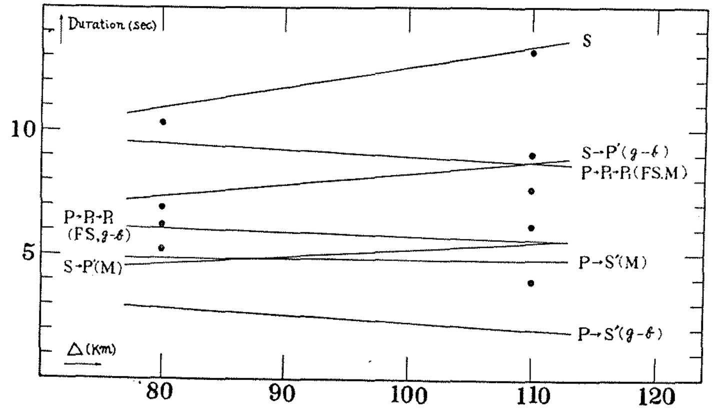

Fig. 23. Durations calculated from the first arrival of P-wave of the transformed and reflected waves in case of 15 km thickness of the basaltic layer.

2. Some examples of locating the depth of M-discontinuity

In case of seismometric observations of an earthquake at some proper distance the non-appearance of forerunners can safely be interpreted as a positive proof for the fact that the focus of the concerned earthquake is situated under the M-discontinuity, unless an absurdly anomalous crustal structure is assigned to the area concerned. Under this consideration the depth of M-discontinuity might roughly be located by comparing two earthquakes, one accompanied and one unaccompanied by forerunners with their epicenters in the same area. As examples, two cases will be reported, namely the earthquakes with their epicenters in Nankaido District and Kanto District. As shown in Figure D on the seismograms obtained with the large Wiechert Seismographs at the Kamigamo Geophysical Observatory, the earthquakes on March 9, 1950(a) and on May 8, 1952(c) were not accompanied by forerunners, and the earthquakes on July 31, 1953(b) and on July 1, 1943(d) inserted for comparison were clearly preceded by forerunners. The epicenters of earthquake (a) and earthquake (b) are close in less than 140 km distance in the open sea of Nankaido ($\varphi=33.2°$ N, $\lambda=136.3°$ E for (a) and $\varphi=33.2°$ N, $\lambda=134.8°$ E for (b)) and the focal depth was estimated by the C.M.O. as 40 km for (a) and 20 km for (b) respectively. From these examples, if the value of focal depth determined by C.M.O. is provisionally confirmed, the depth of the M-discontinuity in this area is roughly located as 30~40 km or less than 30 km from the other examples. And this depth value for the M-discontinuity is well consistent with that derived from the analysis of forerunners of earthquakes in the Nankaido District. On the other hand, the epicenters of earthquakes (c) and (d) are close in less than 80 km distance in Chiba and Ibaraki Prefectures ($\varphi=35.5°$ N, $\lambda=140.2°$ E for (c) and $\varphi=36.2°$ N, $\lambda=140.0°$ E for (d)) and the focal depth is estimated by C.M.O. as 50~60 km for (c) and 50 km for (d). From the same consideration as the former example, the depth of the M-discontinuity in this area is roughly located as 50 km after the numerical value ascribed by C.M.O., and this figure of 50 km also corresponds to the mean depth of the *M*–discontinuity roughly obtained by the analysis of superposed travel time-distance curves of thirteen Kashima-Nada earthquakes. This method of locating the depth of the *M*–discontinuity has certainly a weak point in the uncertainty of focal depth value, but a wide application and fruitful study might be expected in near future by applying this method of location for suitable earthquakes of accurately determined foci.

3. *Focal depths in relation to crustal thickness*

It has early been noticed that a swarm of earthquakes occurring in any fixed area has nearly the same focal depths, and in this section a problem will be examined as to whether or not there is any relation between the focal depths of frequent occurrence in any fixed area and the crustal thickness under that area. Before entering the treatment the problem of value of focal depth determined by seismometric observation should again be discussed. The accurate determination of focal depth is, as already mentioned, a very difficult task, especially for earthquakes in Japan. But putting the problem of difficulty aside for a while, the difference in values calculated from the crustal structure obtained by C.M.O. and the present investigation will be treated in this section under the assumption that the travel-time of the *P*–wave or the *S*–wave were accurately determined at the epicenter or its neighbouring points. The crustal structures adopted by C.M.O. for the calculation of focal depth and others are those by K. Wadati and others (**47, 48**), and the crustal structure adopted in the present investigation for comparison is as shown in Fig. 24. In this figure the travel-times of the *P*–wave and the *P*∼*S*–wave at the epicenter are also plotted for comparison under the crustal structure by C.M.O. and the present investigation. As seen in the figure, the relation between the values of focal depth determined by the same travel time of the *P*–wave at the epicenter of the crustal structure with C.M.O. and the present paper is as follows:

C.M.O.	10 km	20	30	40
The present structure	12	23	34	46

And in Fig. 25 the travel time-distance curves by both structures are plotted under the conditions of various positions of focus. In this figure the full lines of 1∼4 refer to the present structure in cases of the focus at 1) the middle point in the granitic layer (Ca. 5 km–depth), 2) the middle point in the granodioritic layer (Ca. 15 km–depth), 3) the middle point in the basaltic layer (Ca. 27 km–depth) and 4) the bottom of the basaltic layer (Ca. 32 km–depth), and the broken lines of 5∼7 are calculated by C.M.O.–structure under the conditions of focal depth of 5) 0 km, 6)

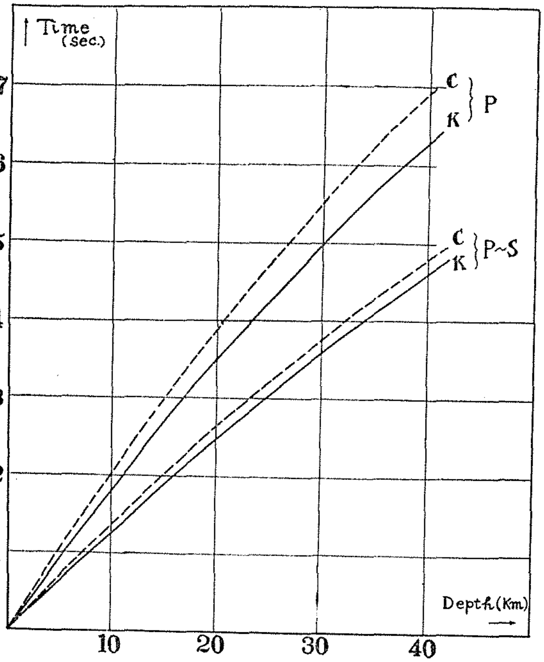

Fig. 24. Travel times of P- and ($P \sim S$)-waves at the epicenter for various focal depths. The full lines (K) and the broken lines (C) represent those calculated by Kishimoto and C.M.O. respectively.

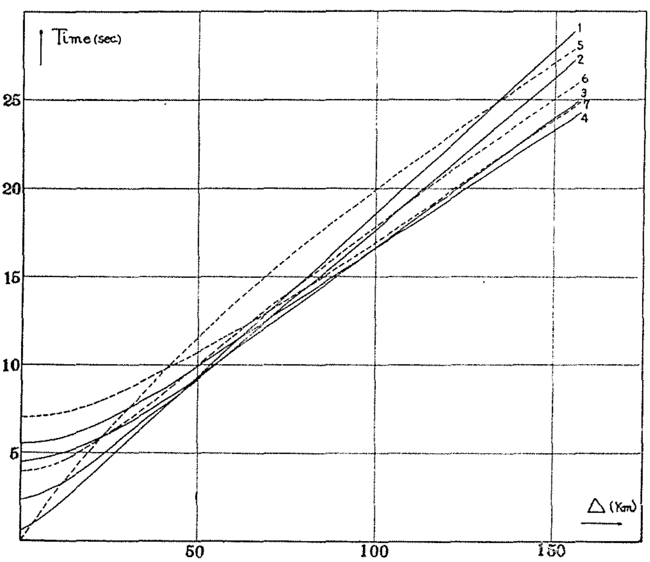

Fig. 25. Comparison of travel time-distance curves between those calculated under the structure of Kishimoto (full lines) and C.M.O. (broken lines) for various focal depths.

20 km and 7) 40 km respectively. From the above described discussion, it is generally understood that the determination of focal depth contains many difficult problems in itself, but in case of permitting errors of some several kilometers the values of focal depth could certainly become an object of discussion, especially in a procedure for

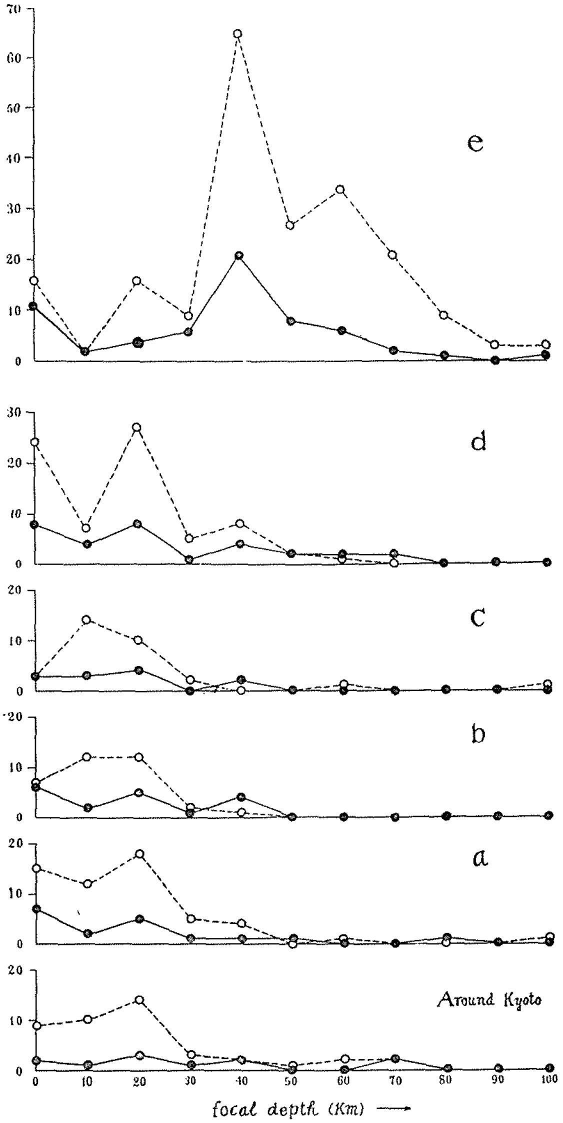

Fig. 26. Relation between frequency of earthquake occurrence and their focal depths classified in the various districts.
The dots and circles denote 'small' and 'moderate and remarkable' earthquakes respectively.

comparing two considerably different values of focal depth.

Remembering the circumstances mentioned above the curves for frequency of occurrence against the focal depth of the earthquakes of 1943 to 1953 at various districts of a ~ e already denoted in Fig. 1 are plotted in Fig. 26, according as three seismic grades of 'remarkable' (sensible by human feeling at $\Delta > 300$ km), 'moderate' (sensible at 300 km $> \Delta >$ 200 km) and 'small' (sensible at 200 km $> \Delta >$ 100 km). In this case the focal depths adopted are determined by C.M.O. In the figure the difference of the peak value of focal depth with two regions of (d) and (e) is very remarkable, and these peak values of 20 km and 40 km for the earthquakes of the Nankaido Group and Kashima-Nada show a well consistent correspondency with the crustal thickness of 52 km and 24 km obtained by the analysis of travel time-distance curves and the duration of forerunner for their respective regions. In allowance for a speculative argument the earthquakes are considered to have tendency to occur frequently at the bottom of the crust.

4. Summary

In the present investigation the analysis of seismic refracted wave (forerunner) was made in detail of the seismograms of natural near and mediate distant earthquakes obtained mainly at the Kamigamo Geophysical Observatory in Kyoto. For the purpose of investigating the crustal structure the method of analysing the travel time-distance curve is usually applied, but it contains many inevitable defects as, for example, inaccuracies in determination of seismic focal depth, arrival time, apparent seismic velocity, critical distance and others, in addition to local effects in the area of the hypocenter and observatory. But the concurrent use of analysis of the forerunner duration, the reflected wave and the transformed wave with the travel time-distance curve may greatly improve the research procedure and raise the reliability of the obtained numerical values. In the present treatment some discontinuous layers in the earth crust were successively investigated from the uppermost layer to the lowest layer by separately analysing the durations of seismic wave forerunners according to their direction and the distance of the epicenter from Kyoto, the study being appropriately supplemented by the analysis of travel time-distance curves and transformed waves. From the present treatment the crustal structure under Kyoto was tentatively determined as follows:

	V_p km/sec	V_s km/sec	Thickness km	Depth of the bottom km
Sedimentary layer	3.8	2.1	1.5	1.5
Granitic layer	5.4	3.1	6	7.5
Granodioritic layer	5.8	3.4	14	21.5
Basaltic layer	7.0	4.0	10	31.5
Mantle surface	8.1	4.7	—	—

A remarkable fact deduced from the present treatment is that of the regional differences of the thickness of the earth's crust which is considered to be intimately related to the distribution of gravity anomaly (**49**) and, moreover, the distribution of seismic focal depths of frequent occurrence. The depth of the Mohorovičić discontinuity or, in other words, the thickness of the crust as deduced from the present investigation is schematically shown in Fig. 27. In case of a monotonous inclination, instead of the horizontally parallel structure above mentioned, of the basaltic layer under the assumption of the value of crustal thickness of 30 km at Kyoto, the crustal thickness at various regions around Kyoto are tentatively calculated as shown in the figure. In the present case, the data observed only at the Kamigamo Geophysical Observatory were used, but the same method is recommended to many observatories at various places for the development of our knowledge on the micro-structure of the earth's crust.

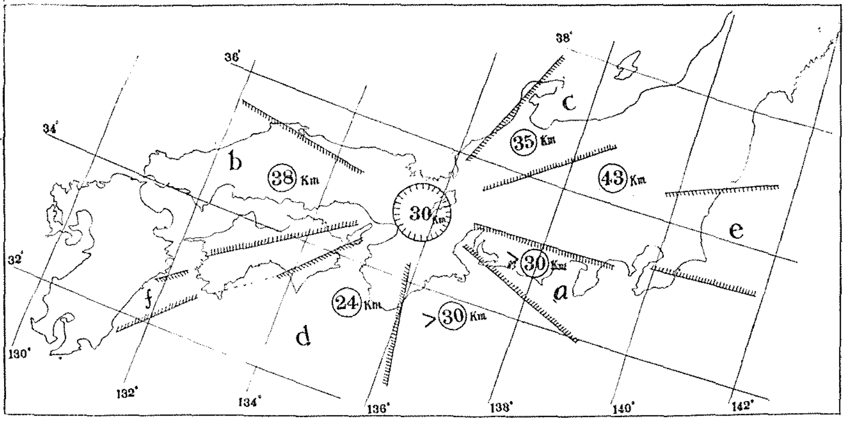

Fig. 27. Schematical representation of the crustal thickness in the various districts.

Acknowledgment

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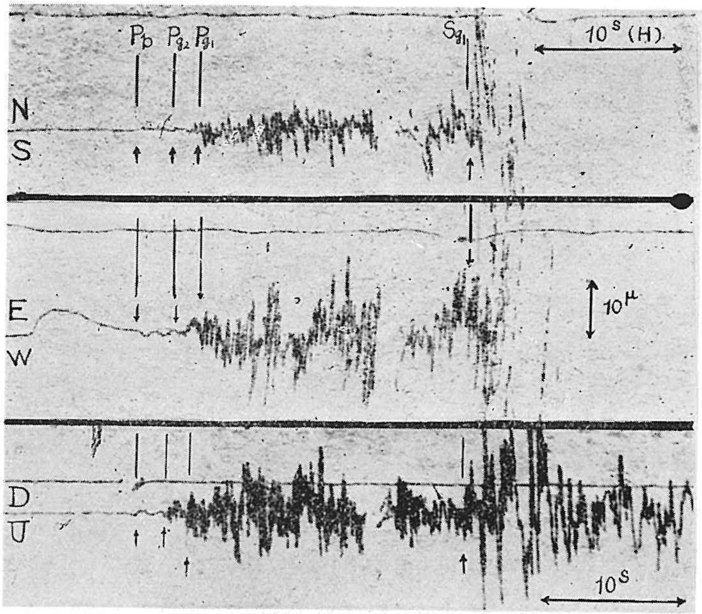

Fig. A (a). Wiechert-seismograms, June 17th 23h 14m, 1949, Lake Hamana, $Pg_1 \sim Sg_1 = 22.0$ sec, $Pb \sim Pg_1 = 3.7$ sec, $Pg_2 \sim Pg_1 = 1.5$ sec, focal depth $= 10$km [sub-class X in the First class; a-No. 2 in Table 2]
Pg_1, Pg_2 and Pb denote the direct wave through the g_1-layer, the refracted wave at the g-g discontinuity and the refracted wave at the g-b discontinuity.

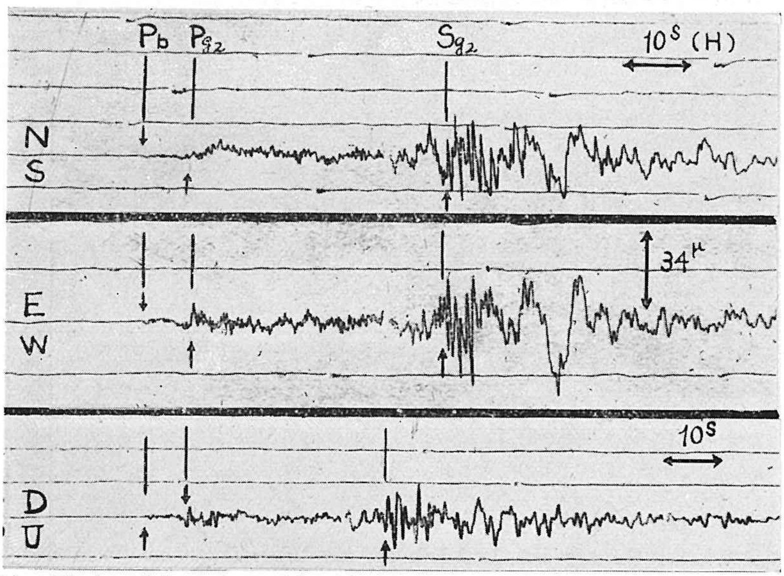

Fig. A (b). Wiechert-Seismograms, May 30th 23h 37m, 1953, Central Setoumi, $Pg_2 \sim Sg_2 = 30.1$ sec, $Pb \sim Pg_2 = 5.1$ sec, focal depth $= 20$km [sub-class Y in the First class; b-No. 3 in Table 2]
Pg_2 and Pb are the direct wave through the g_2- and g_1-layers and refracted wave at the g-b discontinuity.

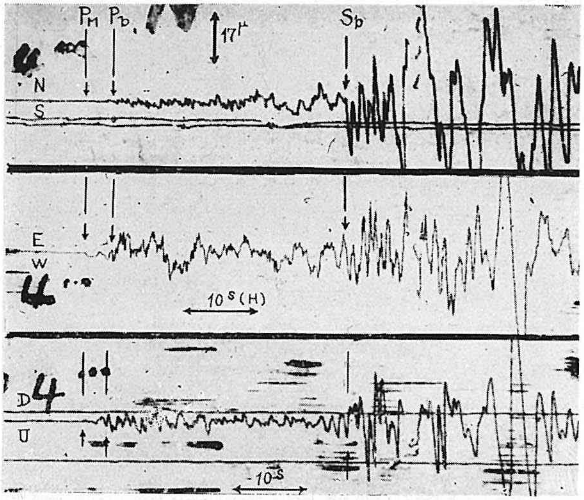

Fig. A (c). Wiechert-Seismograms, Aug. 22nd 11h 04m, 1950, Shimane Pref., $Pb \sim Sb = 31.6$ sec, $P_M \sim Pb = 3.8$ sec, focal depth=30 km [sub-class Z in the First class; b-No. 8 in Table 2]
Pb and P_M denote the direct P-wave through the b-, g_2- and g_1-layers, and the refracted wave at M-discontinuity.

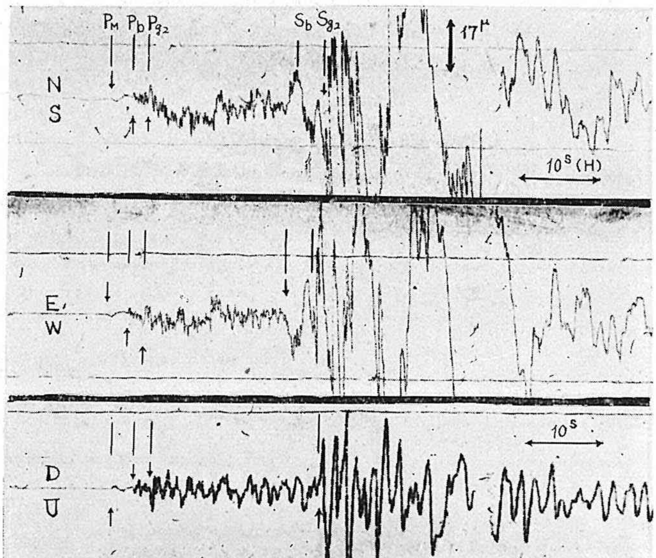

Fig. A(d). Wiechert-Seismograms, Jan. 17th 02h 44m, 1947, Tokushima Pref., $Pg_2 \sim Sg_2 = 19.7$ sec, $Pb \sim Pg_2 = 2.2$ sec, $P_M \sim Pg_2 = 4.7$ sec, focal depth=10km [d-Group in the Second class; d-No. 1 in Table 2]
Pg_2, Pb and P_M denote the direct wave through the g_2- and g_1-layers, the refracted wave at the g-b discontinuity and the refracted wave at the M-discontinuity.

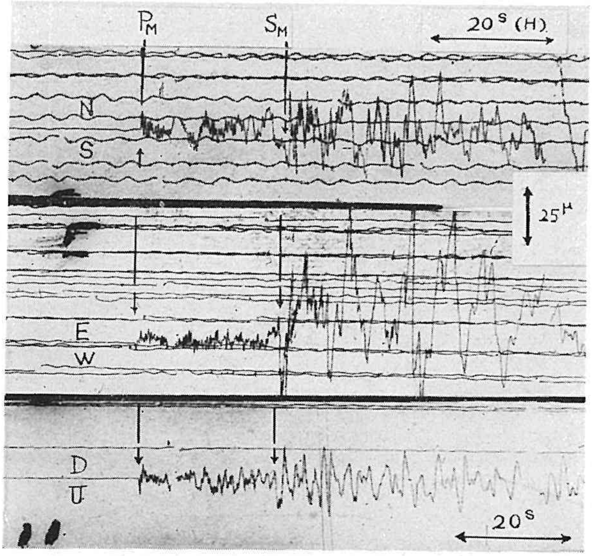

Fig. B (a). Wiechert-Seismograms, Oct. 16th 06h 02m, 1951, off Muroto-Misaki, $P_M \sim S_M = 23.7$ sec, focal depth=40km
P_M denotes the direct wave through the mantle and crust.

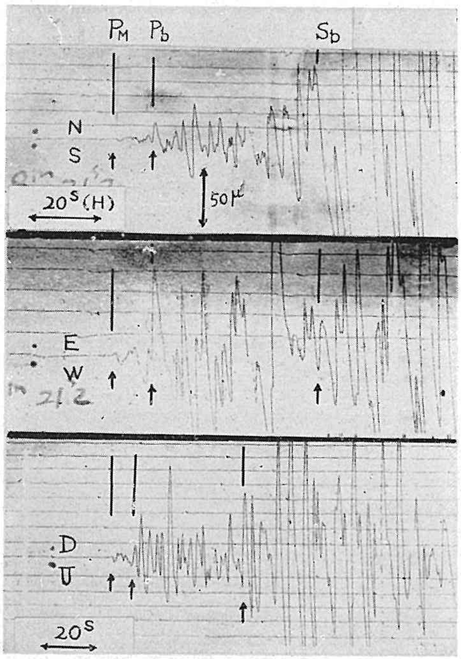

Fig. B (b). Wiechert-Seismograms, July 18th 07h 36m, 1954, near Choshi, $Pb \sim Sb = 50.5$ sec, $P_M \sim Pb = 9.0$ sec, focal depth=40km [e-Group : e-No. 7 in Table 2]
Pb and P_M denote the direct wave through the b-, g_2- and g_1-layers and the refracted wave at the M-discontinuity.

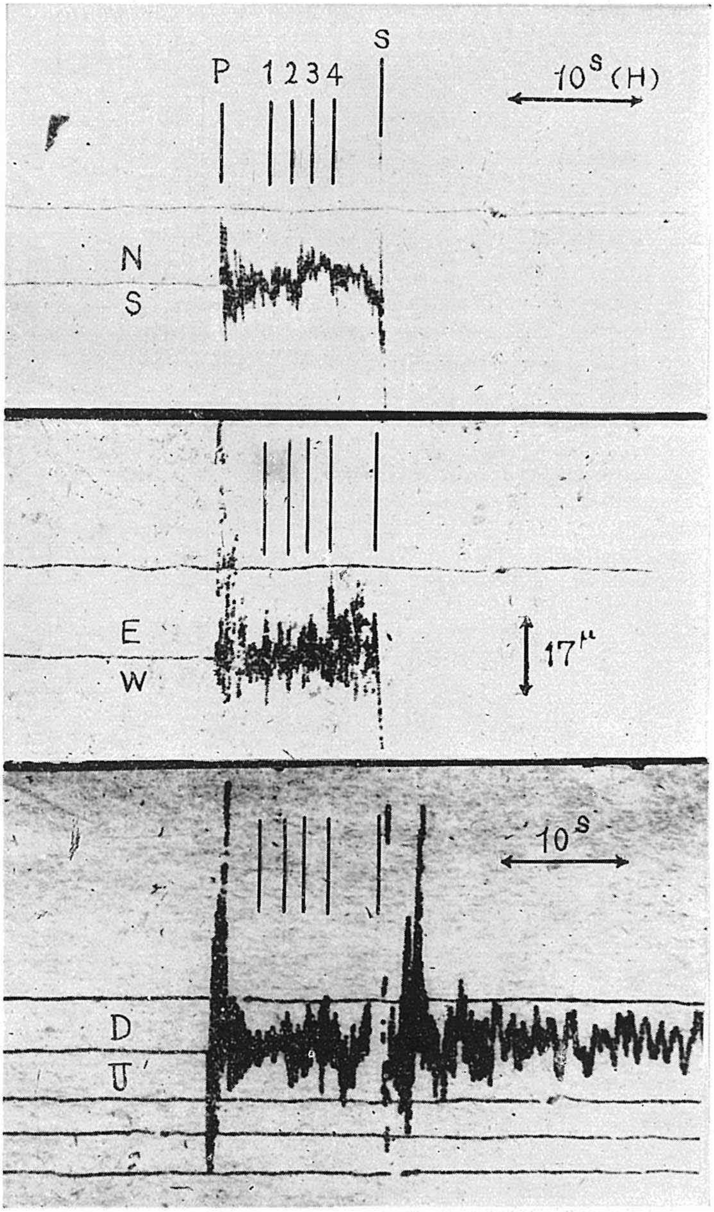

Fig. C. Wiechert-Seismograms of the mediate deep earthquake; Sept. 1st 23h 15m, Nara Pref., $P \sim S = 13.2$ sec, focal depth $= 70$ km. The annexed figures may be referred to Table 4.

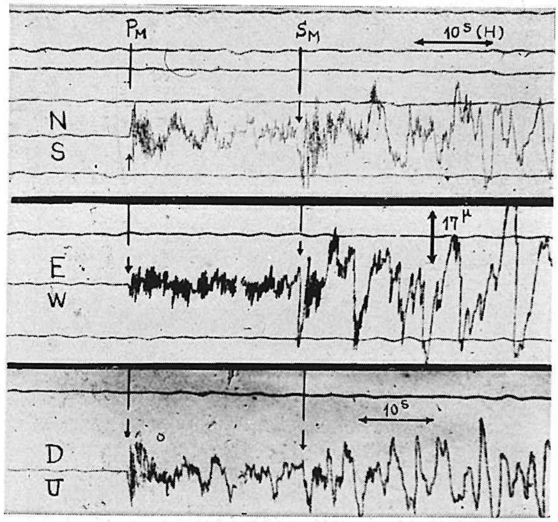

Fig. D (a). Wiechert-Seismograms not accompanied by the forerunner in the d-district; Mar. 9th 17h 59m, 1950, off Shionomisaki, $P_M \sim S_M = 21.9$ sec, focal depth=40km. P_M denotes the direct wave through the mantle and crust.

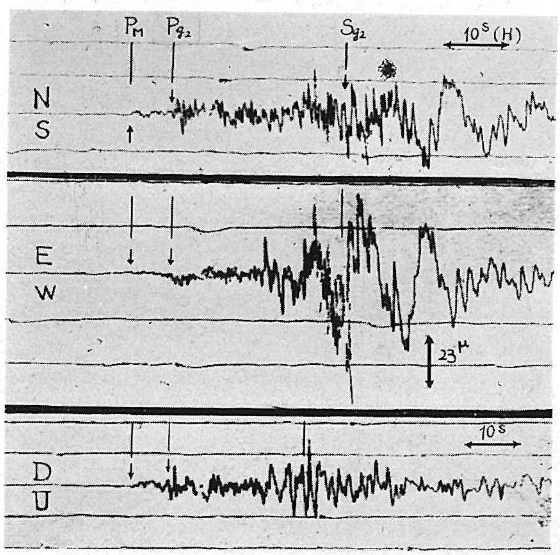

Fig. D (b). Wiechert-Seismograms accompanied by the forerunner in the d-district; July 31st 04h 24m, 1953, Nankaido, $Pg_2 \sim Sg_2 = 27.2$ sec, $P_M \sim Pg_2 = 7.4$ sec, focal depth= 20km [d-Group in the Second class; d-No. 6 in the Table 2]
Pg_2 and P_M are the direct wave through the g_2- and g_1-layers and the refracted wave at the M-discontinuity.

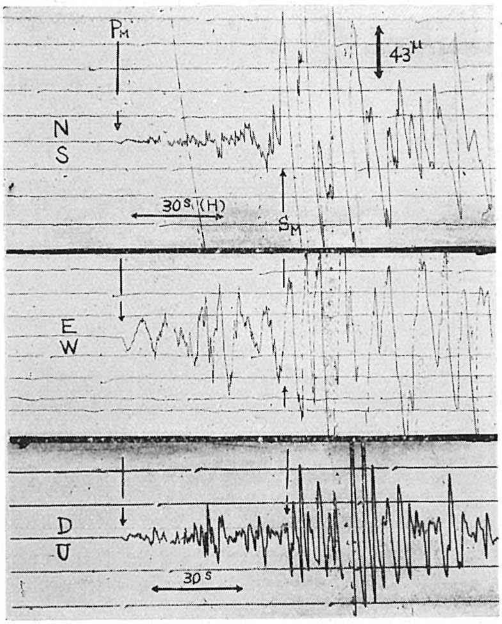

Fig. D (c). Wiechert-Seismograms not accompanied by the forerunner in the e-district; May 8th 09h 58m, 1952, Chiba Pref., $P_M \sim S_M = 52.4$ sec, focal depth $= 50 \sim 60$km
P_M denotes the direct wave through the mantle and crust.

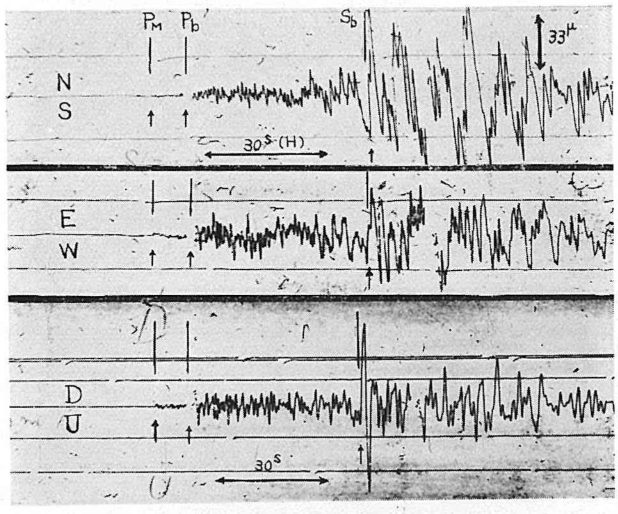

Fig. D (d). Wiechert-Seismograms accompanied by the forerunner in the e-district; July 1st 13h 40m, 1943, Ibaraki Pref., $Pb \sim Sb = 45.6$ sec, $P_M \sim Pb = 9.5$ sec, focal depth $=$ 50km [e-Group; e-No. 1 in Table 2]
Pb and P_M denote the direct wave through the b-, g_2- and g_1-layers and the refracted wave at M-discontinuity.