The velocity of *P*-waves below the "20° discontinuity"

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Summary. — Velocity values for *P*-waves below the "20° discontinuity" are calculated by applying the Wadati-Masuda method, and by using separately calculated travel-time curves for the *P*-waves and the *PdP*-waves of each of the five earthquakes examined.

The velocities calculated have a mean value of 10.21 ± 0.24 km/sec, and are without exception greater than those obtained in a previous piece of research (9.57 km/sec).

As regards the value of the velocity of the *P*-waves above the "20° discontinuity" (8.96 km/sec), the percentage difference between the velocity above and the velocity below the "20° discontinuity" is 14%.

Application of the Herglotz-Wiechert procedure in the zone of the mantle below the "20° discontinuity" does not give significant variations in the correlation between velocity and depth, in comparison with the results obtained by other authors.

Riassunto. — Si determinano i valori della velocità delle onde *P* al disotto della "discontinuità 20°" mediante l'applicazione del metodo di Wadati e Masuda, usufruendo delle dromocrone delle *P* e delle *PdP* relativamente a cinque terremoti studiati.

Tali velocità, con un valore medio di 10.21 ± 0.24 km/sec, risultano sistematicamente più elevate di quella ottenuta in una precedente ricerca (9.57 km/sec).

Relativamente al valore della velocità al disopra della "discontinuità 20°" (8.96 km/sec) il contrasto di velocità nella zona risulta del 14% circa.

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L'applicazione del metodo di Herglotz-Wiechert nella zona del mantello sottostante alla discontinuità non produce significative variazioni della legge di velocità in funzione della profondità rispetto a quelle dedotte da altri autori.

1. – Introduction

In a previous article (4) (Girlanda and Federico, 1966), the equation for the travel-time curve for the P-waves internal to a spherical surface with a radius $r_* = 5833.389 \pm 10.648$ km, was deduced from a study of the Sicilian earthquake of 23 December 1959. This spherical surface was assumed to be the site of a discontinuity in the velocity of the longitudinal waves (20° discontinuity).

The equation for this travel time curve is the following:

$$t_* = (10.633 \pm 0.263)\,\Delta_* - (10.77 \pm 1.67) \cdot 10^{-2}\Delta_*^2 + \\ + (25.3 \pm 3.5) \cdot 10^{-4}\Delta_*^3 \qquad [1]$$

valid between 0° and 14°.

The values for the P-wave velocities, calculated above and below the "20° discontinuity", are respectively $v_{*+} = 8.958 \pm 0.093$ km/sec and $v_{*-} = 9.575 \pm 0.223$ km/sec, with a percentage difference of 6.89%.

In addition, by applying the Herglotz-Wiechert procedure to the spherical model with radius r_*, several pairs of v_k, r_k were calculated using equation [1] in the $0° \mapsto 14°$ range.

The results, obtained for six different values of Δ_*, are shown in Table 1.

Table 1

Results from the Herglotz-Wiechert procedure.

Δ_* deg	p sec/rad	r_k km	h_k km	v_k m/sec	$\Delta v_k/\Delta h_k$ sec^{-1}
2	586.2755	5822.60	548.40	9.93	0.015
4	566.8076	5802.33	568.67	10.24	0.011
6	550.8196	5778.53	592.47	10.49	0.008
8	538.3111	5753.96	617.04	10.69	0.006
10	529.2824	5731.10	639.90	10.83	0.003
14	521.6638	5702.78	668.78	10.93	

Table 1 shows a gradient $\Delta v_k / \Delta h_k$ notably higher than that deducible from the calculation of Jeffreys (1939) (11), Gutenberg (1948 and 1953) (8), Herrin (1968) (9) and others. Such a notably high gradient, which contrasts with the relatively modest jump in velocity, seems to indicate that the above-mentioned discontinuity is a second order discontinuity.

Later (Federico, 1971)(5) the hypothesis of the existence of reflected waves *PdP* on the "20° discontinuity" was proposed, and the travel-time curves of these reflected waves were calculated for various focal depths. To verify the hypothesis of the existence of waves reflecting of the "20° discontinuity", the calculated travel-times were compared with the travel-times deduced from an examination of the impulses attributed to this type of wave and registered on the 14 seismograms examined.

For various focal depths the coefficients of the travel-time curves of the *PdP* waves of the type $t = A + B\Delta^2 + C\Delta^3 + D\Delta^4$ are given in Table 2.

TABLE 2

Coefficients of the *PdP* travel-time curves for different deep-focus

h km	A	B	C	D	[vv]
0	130.267770	0.705062	—2.0497 · 10⁻²	2.30 · 10⁻⁴	1.98 · 10⁻²
40	123.981917	0.721901	—2.1500	2.47	2.19
80	118.934408	0.745470	—2.2938	2.72	2.25
100	116.407652	0.751493	—2.3676	2.85	2.29
150	110.202418	0.791612	—2.5836	3.24	2.49
200	104.136778	0.830776	—2.8415	3.72	2.73
250	98.172171	0.876169	—3.1551	4.33	2.93
300	92.292136	0.928868	—3.5363	5.11	3.13
350	86.481284	0.991352	—4.0156	6.15	3.25
400	80.728694	1.066804	—4.6343	7.59	3.22
450	75.021918	1.160681	—5.4703	9.69	2.95
500	69.345282	1.282802	—6.6829	13.13	2.21

For the seismograms reproduced in the above-mentioned article and in this article, it is possible to see that the impulses attributed to the *PdP*, and the observed times of which are consistent with the calculated times, are of an amplitude comparable to that of the *PP*. The basic elements of the new information are given in Table 3.

TABLE 3

Earthquake data

No.	Date	Origin Time hr min sec	Locality	Depth km	Δ_{Me}(*) Deg	t_{PdP} sec	O–C sec
1	17 Oct 64	09 50 28.0	Crete	18	8.55	167.78	+2.22
2	13 Jun 65	20 01 50.8	Turkey	33	10.87	186.07	—2.87
3	26 Jul 67	18 53 01.1	Turkey	30	19.36	274.78	+1.12
4	30 Jul 67	01 31 01.0	Turkey	18	11.84	197.26	—5.96
5	4 Jul 68	21 47 53.6	Greece	20	6.08	149.14	+1.26
6	28 Mar 69	01 48 29.5	Turkey	4	10.14	183.85	+1.35
7	31 Mar 69	07 15 54.4	Egypt	33	18.65	264.30	0.00
8	14 May 70	18 12 28.0	E. Caucasus	44	24.35	328.08	+0.72

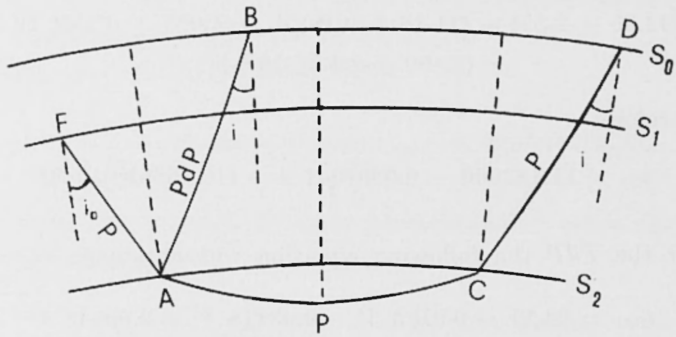
(*) Messina epicentral distance

There are two possible explanations of the sizeably gradient of the velocity P-waves below the "20° discontinuity". It may be because equation [1] was deduced from the adoption of laws of velocity as a function of depth, which, in their details, are rather different from those previously proposed by Jeffreys, Gutenberg and others; or, it may be that the equation [1] is valid within too limited a range for a successful application of the Herglotz-Wiechert procedure. A comparison between the amplitude of the PdP waves and the amplitude of the P-waves suggests, furthermore, a coefficient of reflectivity in the zone of discontinuity higher than that suggested by the jump in velocity found in the preceding piece of research.

1.2. – It is necessary, therefore, to re-examine the jump in velocity across the "20° discontinuity". This can be done by reducing to the surface of the "20° discontinuity" P-wave travel-time curves which have been deduced at sufficiently wide intervals ($20° \leqslant \Delta° \leqslant 90°$), and which have been deduced from the data for a sufficiently large number of seismic events, the hypocentral parameters of which have been determined with sufficient precision and wealth of data.

Such a reduction can be easily obtained using, together with the travel-time curves of the P-waves and the travel-time curves of PdP waves, the procedure adopted separately by Wadati and Masuda (1934)([13]) and by Gutenberg and Richter (1935)([6]) in order to determine the velocity of the P-waves in earth core on the basis of the trasmission times of the ScS and SKS and of the PcP and PKP, respectively.

The procedure adopted by the authors cited above, and which has been extended to the case being studied, can be summarized as follows: let S_1 be the focal depth surface of a given earthquake, and let S_2 be surface of the "20° discontinuity". The seismic ray of a P-wave originating in F under angle i_o intersect the surface of the discontinuity in A. Let p be the parameter of the seismic ray FA. Two longitudinal waves originate in A: a reflected wave AB, and a refracted wave ACD. The refracted wave, after it has traversed the zone below the surface of the discontinuity S_2 for a distance AC, intersect at D under



an angle i, which, given the symmetry connected with the spherical model adopted, is equal to that of the reflected wave AB.

The time required to cross the zone below the "20° discontinuity" can be calculated by taking seismic rays which emerge at the same angle and by subtracting the propagation time t_{PdP} of the reflected wave from the propagation time t_P of the refracted wave. It is necessary, in other words, to calculate from the equations for the travel-time curves $t_{PdP}(\Delta)$ and $t_{Pr}(\Delta)$ the respective $p_{PdP} = \dfrac{dt_{PdP}}{d\Delta}$ and the $p_{Pr} = \dfrac{dt_{Pr}}{d\Delta}$.

The travel-times and the distances relative to the zone underlying the "20° discontinuity" are deduced, for the same value of the seismic ray parameter, from the difference of the travel-times and of the distances of the two travel-time curves $t_{Pr}(\Delta)$ and $t_{PdP}(\Delta)$.

1.3. – The list of the earthquakes chosen for this study is given in Table 4.

Since the procedure for reducing the experimental travel-time curve to the surface of the discontinuity is always the same, only that relative to earthquake N. 1 of January 3, 1960 is given in detail. For the others the global results only are given.

The travel-times relative to distances between 18° and 93° were calculated using the results obtained by Federico (1963)(3) and using, in addition, other observation data given in the *International Seismological Summary* (I.S.S.).

Using data from 43 stations, the following equations for the Pr were obtained:

$$t_{Pr} = (44.35 \pm 2.57) + (11.18 \pm 0.18)\,\Delta - (4.035 \pm 0.355)\,10^{-2}\Delta^2 + \\ + (3.499 \pm 2.130)\,10^{-5}\Delta^3 \qquad [2]$$

and therefore:

$$p_{Pr} = 11.182006 - 0.0807041\,\Delta + (10.496518)10^{-5}\Delta^2 \qquad [3]$$

For the PdP the following equation was obtained:

$$t_{PdP} = 94.11 + 0.9125\,\Delta^2 - 0.03418\,\Delta^3 + 0.00048\,\Delta^4 \qquad [4]$$

the coefficients of which were obtained by using Table 2. From equation [4] the following derivation is obtained:

$$p_{PdP} = 1.8251\,\Delta - 0.10255\,\Delta^2 + 0.0019\,\Delta^3 \qquad [5]$$

In equation [3] the values of p_{Pr} relative to distances of 18° and 93° are, respectively, 9.76330944 sec/Δ° and 4.584369014 sec/Δ°. It is necessary, therefore, to try to obtain from equation [5] the interval of the distances which correspond to values of p_{PdP} included between the above values of the parameter. With this aim in view by attributing to Δ in equation [5] values of each half-degree between 3° and 9.5°, the corresponding values of p_{PdP} were calculated. These, when substituded in equation [3], gave the values of Δ_{Pr}. Then, by using equation [2] and equation [4], the travel-time of t_P and t_{PdP} respectively were calculated relative to these distances. The results of these calculations are contained in Table 5.

The differences $\Delta_{Pr} - \Delta_{PdP}$ and $t_P - t_{PdP}$ make up the pair of values Δ_* and t_* of the distances and of the travel-times for seismic rays, relative to a predetermined value of p, which radiate exclusively in the zone of the mantle below the surface of the "20° discontinuity". (Table 6).

From the 14 pairs of values Δ_*, t_* the following relationship was obtained

$$t_* = (10.034 \pm 0.035)\,\Delta_* - (2.7218 \pm 0.1249)10^{-2}\Delta^2_* + - (2.9 \pm 1.027)10^{-5}\Delta^3_* \qquad [6]$$

valid between 9° and 90° and which makes up the equation of the travel-time curve of the P-waves which radiate exclusively in the zone of the mantle beneath the "20° discontinuity".

From equation [6] it follows that:

$$p_* = \frac{dt_*}{d\Delta_*} = 10.034 - 0.05443\,\Delta_* - 0.0000875706\,\Delta^2_* \qquad [7]$$

and therefore the value of the velocity immediately below the surface of the discontinuity is:

$$v_{*-} = \frac{r_*}{\frac{180}{\pi}(p_*)_{\Delta=0}} = 10.1464 \text{ km/sec}$$

The same procedure, applied to earthquakes 2, 3, 4, 5 of Table 4, gives the following equations:

Earthquake N. 2 of 25.3.62 (Southern Tyrrhenian sea)

$$t_* = (9.9253 \pm 0.0059)\,\Delta_* - (23.3527 \pm 0.2211)\,10^{-3}\,\Delta_*^2 + - (6.00 \pm 0.20)\,10^{-5}\,\Delta_*^3 \qquad [6a]$$

$$v_{*-} = 10.2577 \text{ km/sec}$$

Earthquake N. 3 of 22.7.67 (Turkey)

$$t_* = (9.9575 \pm 0.0021)\,\Delta_* - (24.3451 \pm 0.0785)\,10^{-3}\,\Delta_*^2 + - (5.35 \pm 0.07)\,10^{-5}\,\Delta_*^3 \qquad [6b]$$

$$v_{*-} = 10.2246 \text{ km/sec}$$

Earthquake N. 4 of 26.7.67 (Turkey)

$$t_* = (9.8616 \pm 0.0105)\,\Delta_* - (24.6139 \pm 0.3863)\,10^{-3}\,\Delta_*^2 + \\ - (4.98 \pm 0.33)\,10^{-5}\,\Delta_*^3 \qquad [6c]$$

$v_{*-} = 10.2204$ km/sec

Earthquake N. 5 of 30.7.67 (Turkey)

$$t_* = (9.9679 \pm 0.0123)\,\Delta_* - (24.6905 \pm 0.4532)\,10^{-3}\,\Delta_*^2 + \\ - (5.17 \pm 0.39)\,10^{-5}\,\Delta_*^3 \qquad [6d]$$

$v_{*-} = 10.2139$ km/sec.

The Herglotz-Wiechert procedure, applied at distance intervals from $0^\circ \mapsto 90^\circ$ and, separately, to the five relationships [6], [6a], [6b], [6c], [6d], gives the results summarized in Table 7.

In Fig. 9 the results listed in Table 7 are given in diagram form and are compared with those obtained by other authors (curves 2 and 3). The circles represent the limits of the interval of minimum and maximum values of the propagation velocity found in corrispondence with various depths, in the five different sets of calculations.

2. – Conclusions

2.1. – The hypothesis which formulated the coexistence of the "20° discontinuity" and of the "low velocity channel" is supported not only by the experimental evidence of the reflected waves (Federico, 1970); it also receives further support from the results of the present study which tend to indicate a clearer velocity contrast across the discontinuity. These results are consistent with those suggested by a comparison of the *PP* phases with the *PdP* phases on the seismograms examined. In fact, from the procedure which reduces the experimental travel-time curves of the five earthquakes studied to the surface of the discontinuity, propagation values for the *P*-waves immediately below the surface of the discontinuity were obtained which are systematically higher than those previously obtained (Girlanda-Federico, 1966). In percentage terms, the average increase in relation to the value of the velocity of the *P*-waves above the "20° discontinuity" (8.96 km/sec), is 14%.

2.2. – As it can be seen in Table 7 and even more clearly in the diagram (Fig. 9), the application of the Herglotz-Wiechert method produced, for each depth value, an almost uniform distribution of the velocity of propagation of the P-waves, with moderate standard deviation, of the order of 0.04 km/sec.

In addition, by comparing the diagram of the Fig. 9 with curve 1, it is possible to note the elimination of the high initial gradient $\frac{\Delta v_k}{\Delta h_k}$ previously found.

2.3. – The analysis of the velocity values calculated for the part of the mantle below the depth of 700 km does not show significant variations in the correlation of depth and velocity in respect to that obtained by other authors (Jeffreys, 1952; Gutenberg, 1959; Herrin, 1969).

However, immediately below the surface of the discontinuity and down to a depth of 700 km, the propagation values for the P-waves are systematically higher than those obtained by the above mentioned authors.

This difference is to be explained, in part, by the particular characteristics of the reduction method adopted, and, more importantly, by the structure of the velocity models proposed by the other authors for the upper mantle.

Jeffreys, in fact, does not take into consideration low velocity zones under the crust; Gutenberg, on the contrary, excludes the possibility of discontinuities in the zone; Herrin envisages neither of these hypotheses. In contrast, the present writer adopted a model in which the asthenosphere and the "20° discontinuity" coexist.

However, if the travel-time curves of the five earthquakes here studied are to be consistent with the law which correlates velocity and depth, it is necessary, in all of the models used, that the mean velocity values, in a wide interval around the surface of the discontinuity, are almost equal. These mean values corresponding to depth of 400, 450, 500, 550, 600 km are the following: 9.49 $\pm$ 0.39 km/sec (Gutenberg, 1959), 9.68 $\pm$ 0.48 km/sec (Jeffreys, 1952), 9.68 $\pm$ 0.41 km/sec (Herrin, 1968)([9]) and 9.52 $\pm$ 0.89 km/sec (present work).

As it can be seen, the value obtained by present author is inside the range of the previously calculated values but shows a higher scattering, justified, in fact, by the presence of the discontinuity surface. *The higher scattering explains, therefore, both the lower velocity values above the discontinuity and the higher values below it.*

TABLE 4

Earthquake epicentral data

Nº	Date	Locality	Latitude (deg)	Longitude (deg)	Origin-Time (hr min sec sec)	Depth (km)	Data (*) Source
1	3 Jan 60	S. Tyrr. Sea	39.061 ± 0.041	15.413 ± 0.07	20 19 34.5 ± 0.40	284.5 ± 6.1	F
2	25 Mar 62	S. Tyrr. Sea	39.065 ± 0.033	14.558 ± 0.04	21 38 26.1 ± 0.40	337.6 ± 4.0	B
3	22 Jul 67	Turkey	40.670 ± 0.027	30.690 ± 0.02	16 56 58.0 ± 0.15	33.0 ± 3.3	BCIS
4	26 Jul 67	Turkey	39.540 ± 0.024	40.380 ± 0.02	18 53 01.1 ± 0.14	30.0 ± 2.3	BCIS
5	30 Jul 67	Turkey	40.720 ± 0.018	30.520 ± 0.02	01 31 01.8 ± 0.57	18.0 ± 4.3	BCIS

(*) F = Federico (1963)[3]
B = Bottari and Lo Giudice (1975)[1]
BCIS = Bureau Central International de Séismologie

TABLE 5

Wadati and Masuda's method. Values of the seismic ray parameter and corresponding times and distances for the PdP and Pr.

p sec/Δ^o	Δ_{PdP} deg	Δ_{Pr} deg	t_{PdP} sec	t_{Pr} sec
4.60502037	3	92.66264070	101.4405248	761.8733907
5.21521103	3.5	82.86507080	103.8973235	713.7826383
5.78435016	4	74.00524340	106.6488938	665.0676947
6.31389836	4.5	65.98302570	109.6750751	616.5493310
6.80531625	5	58.71515660	112.9564372	568.8816767
7.26006443	5.5	52.13119560	116.4742798	522.5837080
7.67960352	6	46.17063320	120.2106334	478.0629987
8.06539412	6.5	40.78078190	124.1482586	435.6341376
8.41889685	7	35.91519620	128.2706463	395.5332855
8.74157231	7.5	31.53247400	132.5620175	357.9299717
9.03488112	8	27.59532320	137.0073240	322.9367488
9.30028388	8.5	24.06983090	141.5922476	290.6172725
9.53924121	9	20.92487830	146.3032005	260.9931086
9.75321371	9.5	18.13166840	151.1273247	234.0495506

TABLE 6

Wadati and Masuda's method. Values of the seismic ray parameter for the distances and travel-time under the "20° discontinuity".

p sec/Δ^o	Δ_* deg	t_* sec
9.75321371	8.6316684	82.9222259
9.53924121	11.9248783	114.6899081
9.30028388	15.5698309	149.0250247
9.03488112	19.5953232	185.9294248
8.74157231	24.0324740	225.3679542
8.41889685	28.9151962	267.2626392
8.06539412	34.2807819	311.4858790
7.67960352	40.1706332	357.8523653
7.26006443	46.6311956	406.1094282
6.80531625	53.7151566	455.9252395
6.31389836	61.4830257	506.8742559
5.78435016	70.0052434	558.4188005
5.21521103	79.3650704	609.8853148
4.60502037	89.6626407	660.4328659

TABLE 7

Results from the Herglotz-Wiechert procedure

Δ (deg)	Earthquake 1 h (km)	Earthquake 1 V (km/sec)	Earthquake 2 h (km)	Earthquake 2 V (km/sec)	Earthquake 3 h (km)	Earthquake 3 V (km/sec)	Earthquake 4 h (km)	Earthquake 4 V (km/sec)	Earthquake 5 h (km)	Earthquake 5 V (km/sec)
2	544.018	10.2468	543.591	10.3452	543.705	10.3155	543.736	10.3123	543.745	10.3059
4	555.847	10.3405	554.674	10.4268	554.900	10.4002	555.070	10.3979	555.098	10.3918
6	571.310	10.4306	569.226	10.5054	569.791	10.4818	569.931	10.4802	569.985	10.4744
8	589.801	10.5180	586.690	10.5817	587.538	10.5611	587.739	10.5602	587.832	10.5548
10	610.960	10.6033	606.761	10.6565	607.913	10.6387	608.174	10.6385	608.316	10.6334
12	634.563	10.6869	629.238	10.7301	630.709	10.7150	631.026	10.7153	631.229	10.7107
14	660.443	10.7690	653.981	10.8026	655.782	10.7902	656.147	10.7911	656.423	10.7869
16	688.475	10.8500	680.886	10.8744	683.020	10.8646	683.425	10.8658	683.788	10.8622
18	718.559	10.9299	709.873	10.9455	712.340	10.9382	712.774	10.9398	713.236	10.9368
20	750.616	11.0088	740.878	11.0162	743.675	11.0113	744.124	11.0132	744.701	11.0108
22	784.579	11.0869	773.850	11.0865	776.969	11.0839	777.420	11.0860	778.126	11.0843
24	820.390	11.1642	808.741	11.1565	812.178	11.1562	812.612	11.1584	813.465	11.1575
26	858.002	11.2410	845.533	11.2264	849.263	11.2282	849.666	11.2305	850.680	11.2304
28	897.372	11.3171	884.180	11.2961	888.193	11.3000	888.544	11.3024	889.737	11.3031
30	938.464	11.3928	924.663	11.3660	928.940	11.3718	929.218	11.3741	930.609	11.3758
32	981.245	11.4680	966.960	11.4359	971.481	11.4436	971.665	11.4457	973.273	11.4484
34	1025.686	11.5430	1011.055	11.5059	1015.797	11.5155	1015.863	11.5173	1017.707	11.5211
36	1071.761	11.6176	1056.932	11.5762	1061.870	11.5875	1061.794	11.5890	1063.894	11.5939
38	1119.446	11.6920	1104.579	11.6468	1109.686	11.6598	1109.440	11.6609	1111.819	11.6670
40	1168.720	11.7662	1153.986	11.7179	1159.232	11.7324	1158.790	11.7328	1161.468	11.7403
42	1219.565	11.8403	1205.144	11.7894	1210.497	11.8053	1209.830	11.8051	1212.830	11.8140
44	1271.962	11.9143	1258.045	11.8615	1263.472	11.8787	1262.549	11.8778	1265.896	11.8882

cont. Table 7

	Earthquake 1		Earthquake 2		Earthquake 3		Earthquake 4		Earthquake 5	
Δ (degr)	h (km)	V (km/sec)	h (km)	V (km/sec)	h (km)	V (km/sec)	h (km)	V (km/sec)	h (km)	V (km/sec)
46	1325.896	11.9883	1312.684	11.9342	1318.151	11.9527	1316.940	11.9509	1320.659	11.9629
48	1381.551	12.0624	1369.057	12.0076	1374.526	12.0273	1372.995	12.0245	1377.110	12.0382
50	1438.315	12.1365	1427.162	12.0819	1432.593	12.1025	1430.707	12.0986	1435.245	12.1141
52	1496.775	12.2109	1486.996	12.1571	1492.349	12.1786	1490.072	12.1735	1495.060	12.1909
54	1556.720	12.2854	1548.559	12.2332	1553.790	12.2556	1551.085	12.2491	1556.553	12.2685
56	1618.140	12.3602	1611.853	12.3106	1616.917	12.3335	1613.745	12.3255	1619.722	12.3471
58	1681.026	12.4354	1676.879	12.3891	1681.729	12.4126	1678.048	12.4029	1684.566	12.4267
60	1745.369	12.5110	1743.640	12.4690	1748.227	12.4928	1743.995	12.4813	1751.086	12.5075
62	1811.162	12.5871	1812.142	12.5504	1816.414	12.5744	1811.597	12.5609	1819.283	12.5897
64	1878.398	12.6638	1882.389	12.6334	1886.293	12.6575	1880.825	12.6417	1889.162	12.6733
66	1947.071	12.7410	1954.389	12.7182	1957.870	12.7421	1951.711	12.7239	1960.727	12.7584
68	2017.177	12.8191	2028.150	12.8050	2031.149	12.8285	2024.250	12.8076	2033.982	12.8453
70	2088.710	12.8980	2103.683	12.8938	2106.139	12.9167	2098.448	12.8931	2108.935	12.9340
72	2161.670	12.9778	2181.000	12.9850	2182.849	13.0071	2174.310	12.9803	2185.595	13.0248
74	2236.051	13.0585	2260.113	13.0786	2261.289	13.0997	2251.844	13.0695	2263.971	13.1178
76	2311.855	13.1405	2341.038	13.1750	2341.471	13.1948	2331.061	13.1610	2344.076	13.2133
78	2389.079	13.2236	2423.792	13.2744	2423.411	13.2927	2411.971	13.2548	2425.922	13.3115
80	2467.725	13.3082	2508.396	13.3772	2507.123	13.3935	2494.588	13.3512	2509.526	13.4126
82	2547.794	13.3943	2594.871	13.4835	2592.628	13.4976	2578.926	13.4506	2594.906	13.5169
84	2629.289	13.4821	2683.243	13.5937	2679.944	13.6052	2665.003	13.5530	2682.081	13.6247
86	2712.216	13.5718	2773.540	13.7083	2769.098	13.7168	2752.839	13.6589	2771.075	13.7363
88	2697.578	13.6635	2865.794	13.8277	2860.115	13.8326	2842.453	13.7686	2861.914	13.8523
90	2882.383	13.7575	2960.040	13.9524	2953.026	13.9533	2933.874	13.8825	2954.627	13.9730

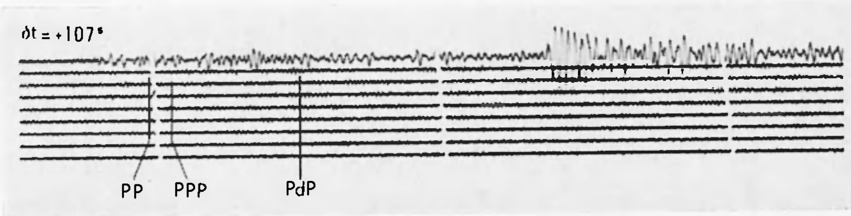


Fig. 1 – Earthquake N. 1 of October 17, 1964 (Crete). Z – Sprengnether seismogram. $\Delta_{ME} = 8.55°$.

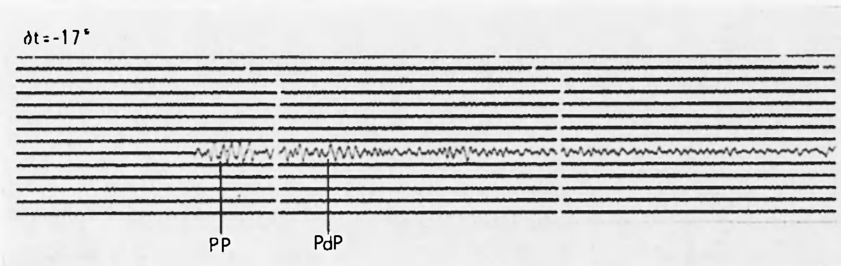


Fig. 2 – Earthquake N. 2 of June 13, 1965 (Turkey). Z – Sprengnether seismogram. $\Delta_{ME} = 10.87°$.

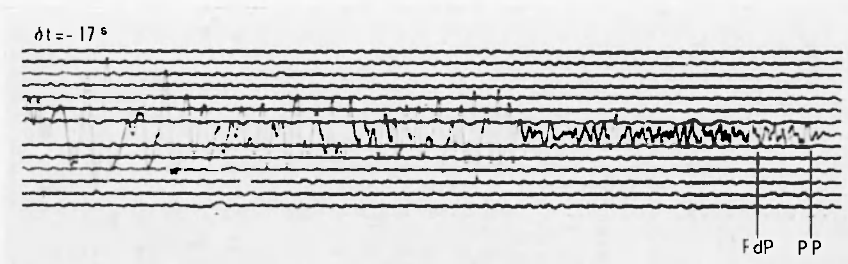


Fig. 2a – Earthquake N. 2: N-S Galitzin seismogram.

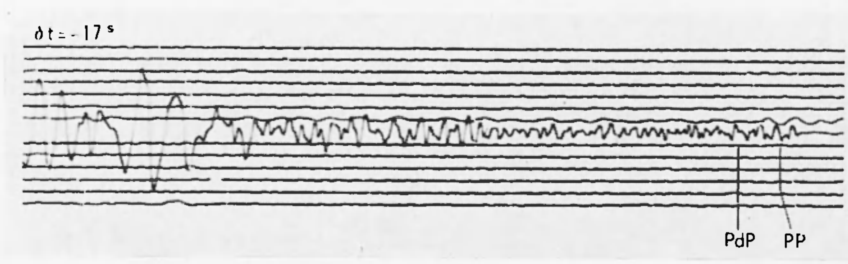


Fig. 2b – Earthquake N. 2: E-W Galitzin seismogram.

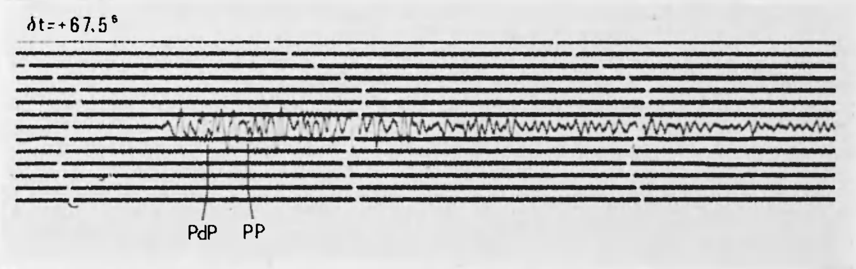


Fig. 3 – Earthquake N. 3 of July 26, 1967 (Turkey). Z – Sprengnether seismogram – Δ_{ME} = 19.36°.

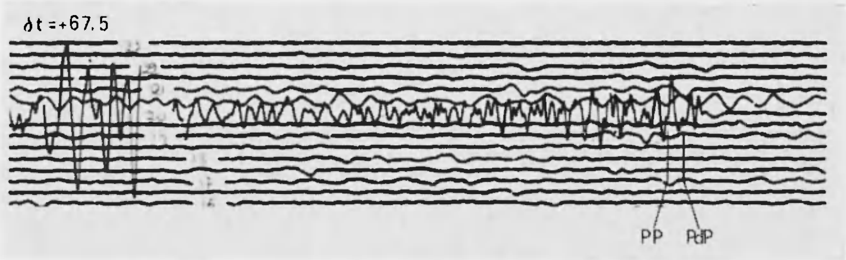


Fig. 3a – Earthquake N. 3: N-S Galitzin seismogram.

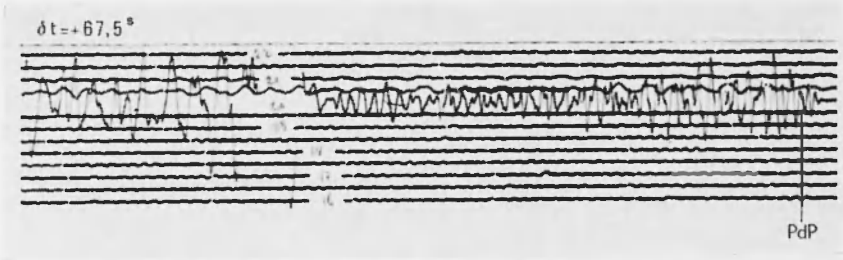


Fig. 3b – Earthquake N. 3: E-W Galitzin seismogram.

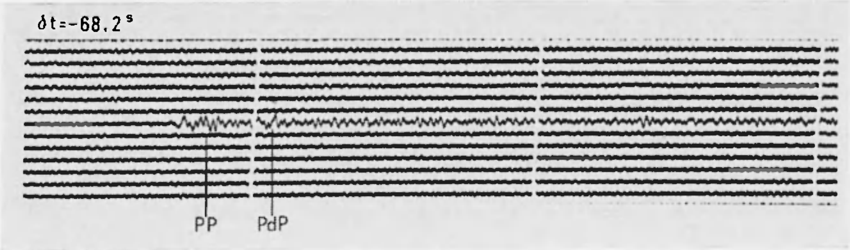


Fig. 4 – Earthquake N. 4 of July 30, 1967 (Turkey). Z – Sprengnether seismogram – Δ_{ME} = 11.84°.

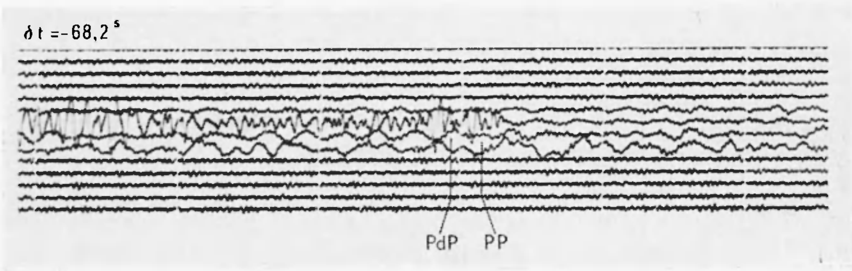


Fig. 4a – Earthquake N. 4: N-S Galitzin seismogram.

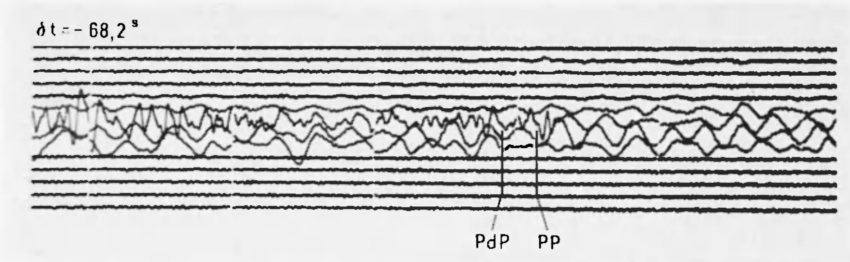


Fig. 4b – Earthquake N. 4: E-W Galitzin seismogram.

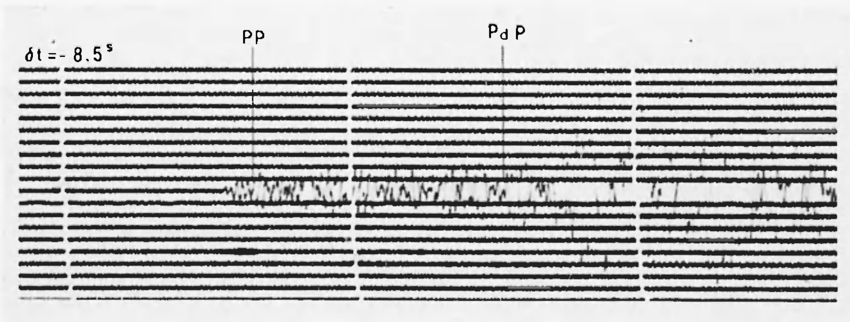


Fig. 5 – Earthquake N. 5 of July 4, 1968 (Greece). Z – Sprengnether seismogram – $\Delta_{ME} = 6.08°$.

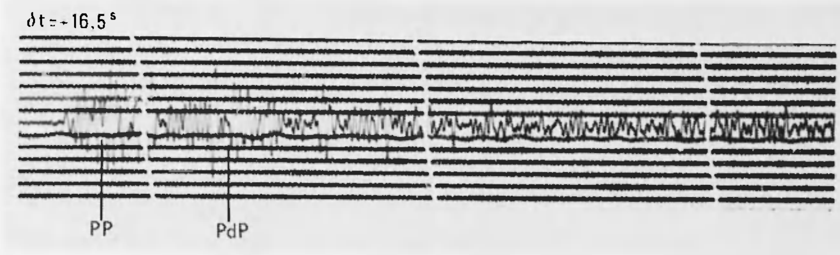


Fig. 6 – Earthquake N. 6 of Mars 28, 1969 (Turkey). Z – Sprengnether seismogram – $\Delta_{ME} = 10.14°$.

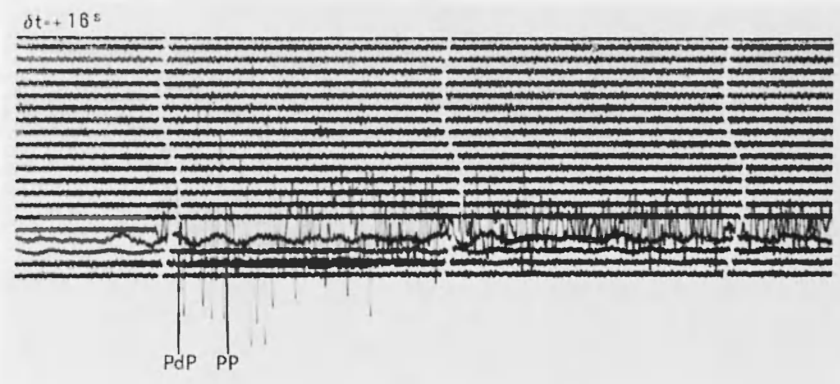


Fig. 7 – Earthquake N. 7 of Mars 31, 1969 (Egypt). Z – Sprengnether seismogram – $\Delta_{ME} = 18.65°$.

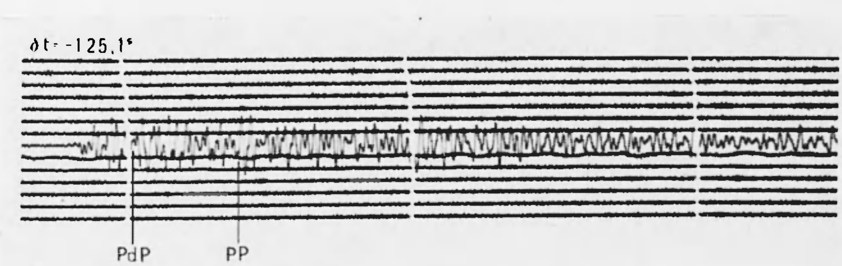


Fig. 8 – Earthquake N. 8 of May 14, 1970 (Eastern Caucasus). Z – Sprengnether seismogram – $\Delta_{ME} = 24.35°$.

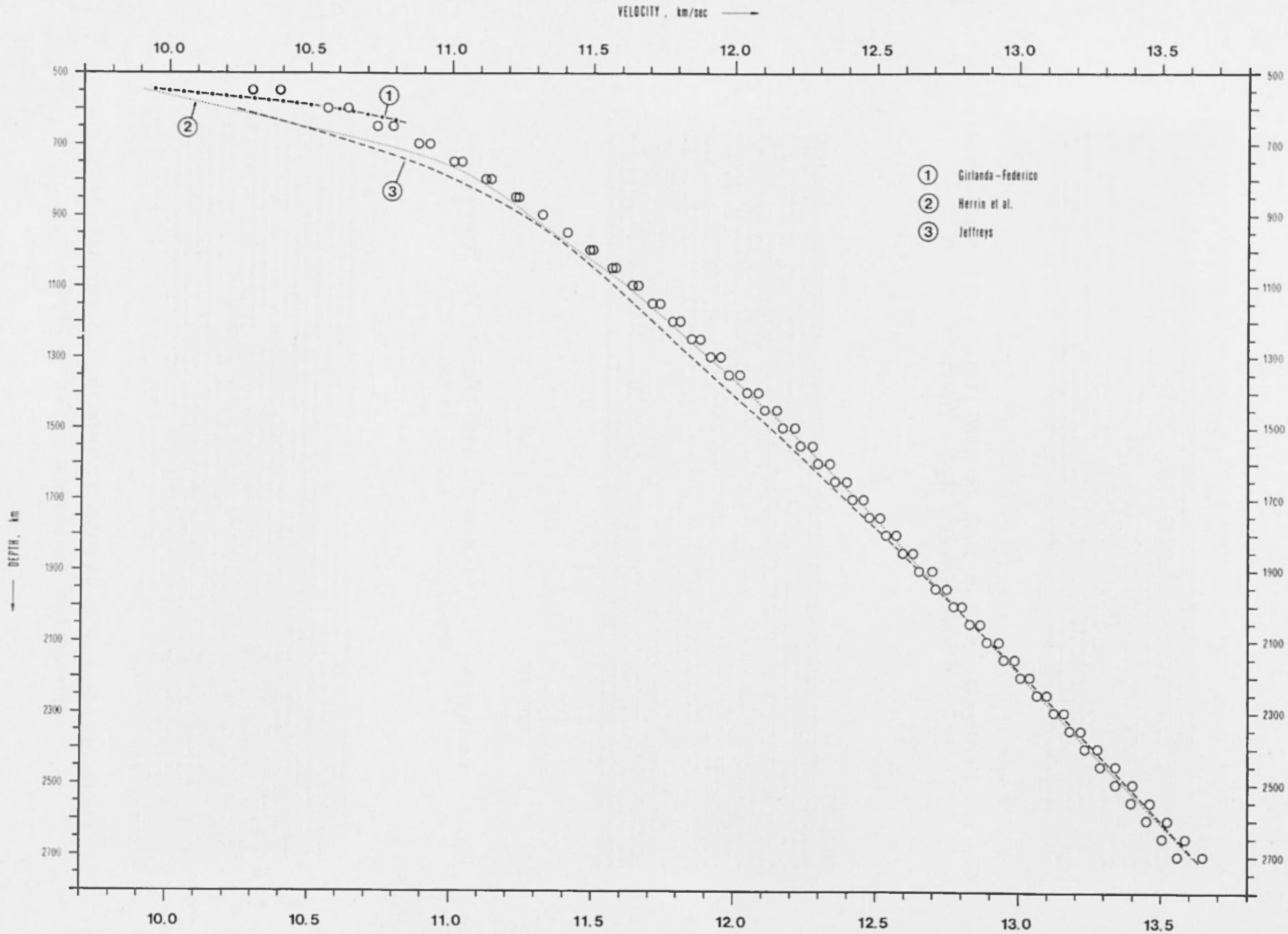


Fig. 9 – Diagram of the *P*-wave velocity. The circles represent the limits of the interval of minimum and maximum values found in the five different sets of calculations.

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