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Seismic Waves near 110°: is Structure in Core or Upper Mantle Responsible?

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Summary

An analysis is given of longitudinal waves which are observed at distances between 105° and 115° some minutes after the arrival of the *P* waves diffracted by the mantle–core boundary. A number of seismograms which clearly show these wave trains is examined; recordings at *WWSS* stations from the Chilean shock of 1965 March 28 are specially studied.

The observations show (i) that discernible *PKiKP* waves with wavelengths of order 10 km are reflected from the boundary of the Earth's inner core back to distances of at least 105°, and (ii) that many longitudinal wave onsets (the *PdP* phase) having travel-times up to 60 s before *PKiKP* and 90 s before *PP* arrive near 110° by means of reflection from the lower side of physical discontinuities in the upper mantle of the Earth.

The first result is consistent with a relatively sharp increase in *P* velocity between the transition zone and the inner core at a radius of about 1220 km. The second suggests the existence of a number of discrete shells of different elastic properties in the Earth's upper mantle above 400 km; in particular, one prominent group of *PdP* waves of order 2 s period is consistent with reflection from a discontinuity near 385 km. This result provides confirmation of the overall high velocity gradient near this depth inferred by L. Johnson; there is an indication, however, of first-order discontinuities in the velocity function assumed to be smooth by Johnson.

1. The problems

In this paper we endeavour to answer two seismological questions:

(a) Does appreciable wave-energy arrive at the Earth's surface from reflections at the outside of the inner core?

(b) What is the history of the waves which are recorded up to about one minute before the expected times of the waves reflected at the inner core boundary?

The first question bears on the degree of sharpness of the outer boundary of the inner core. Although accepting the hypothesis of Lehmann (1936) that *PKiKP* waves for $\Delta < 140°$ arose from a discontinuity within the core, Gutenberg & Richter (1938) adopted in practice a rapid but continuous increase in velocity in a narrow shell within the core. They commented that 'no reflected waves have been found which would correspond to a discontinuity'. By contrast, Jeffreys (1939) adopted

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reflection from a sharp discontinuity 'for convenience, (although) I think that it may well be true, partly from consideration of . . . amplitudes, partly from the length of the receding branch itself'.

Although it might be expected that more recent seismological work, based on the greater global distribution of high-gain short-period seismographs, would have resolved this question, a number of papers in the last decade show the matter has not been settled. Only a few specific observational surveys have been published (e.g. Melik-Gajkazan 1955; Gutenberg 1960; Caloi 1961, 1963; Ergin 1967; Engdahl 1968). On the theoretical side, the reflection hypothesis has been incorporated into core models by Bolt (1962), Hai (1963), Adams & Randall (1964), and Ergin (1967). The adopted elastic contrast at the inner core boundary has ranged widely; the Jeffreys model has a jump in the *P* velocity of 1·76 km/s while Ergin's solution has only 0·6 km/s. In order to allow an experimental study of the question, Bolt & O'Neill (1965) published theoretical work on the wave-energy partition at a sharp inner core boundary. The conclusion was that for a 10 per cent jump in *P* velocity at the boundary, the optimum range for observation of *PKiKP* is $105° < \Delta < 110°$.

Since 1965, we have made a special study of many seismograms from short-period vertical component seismographs. We have found observational confirmation of the hypothesis of reflections at a sharp boundary.

Problem (b) has had little intensive study. The observational reality of the longitudinal bodily waves in question is not in doubt; many published papers have plots of unexplained onsets arriving at $105° < \Delta < 110°$ (e.g. Gutenberg 1960; Hai 1967; Engdahl 1968). Gutenberg discussed this feature several times (e.g. Gutenberg 1960, Fig. 3) and pointed out that early empirical travel-time curves for *PKiKP* started near 105° with a slope almost parallel to that of *PP*; he did not, however, give a specific explanation of the observations. In his remarkable paper, Nguyen-Hai (1963, p. 48) also gave examples of the early phases and wrote of 'a multiplicity of phases recorded between *P* and *PP*' in the distance range $95° < \Delta < 111°$. Of the six earthquakes he studied, three Fijian shocks had focal depths of order 600 km so that the reflected phases of *pP* type mask the time–distance domain in question. Only one of the shallower earthquakes (1958 August 15, Celebes Sea) provided suitable observations. The record analysis is somewhat complicated, however, by the likelihood, established by Nguyen-Hai, that two shocks occurred within a few seconds with depths of 210 and 115 km.

Hai plotted a series of arrival times for $95° < \Delta < 111°$ which occur up to 100 s before those *PP* onsets which correspond to the first shock of the double earthquake. He divided the onsets into two classes: class one fell on hypothetical travel-time curves parallel to *P* and class two on curves parallel to *PP*. The first class he explained in terms of an upward reflection of the down-going *P* wave on the upper surface of a (Repetti) discontinuity at about 920 km depth followed by downward reflection from the Earth's surface. The second class was attributed to reflections of *PP* type on the lower side of discontinuities in the upper mantle. From the material published, it appears that this division into two groups is not required by the data; although Hai's division may be true we examine the consequences of assuming that only the second mode of generation applies. For reference, we denote the longitudinal waves reflected from the lower side of an interface at depth *d* by the symbol *PdP*.

2. Evidence on *PKiKP* waves

The first part of the work consisted of a survey of specially selected seismograms from stations of the University of California at Berkeley network in central and northern California. In a recording interval of ten years (1955–65), the twenty

Table 1

Earthquake data, Berkeley stations

No.	Date	Origin Time (h min s)	Epicentre (deg)	Depth (km)	Mag.	Region	Station Name	Station Δ (deg)
1	1955 Mar. 31	18 17 19	8·1 N, 123·2 E	~100	7½	Philippines	Reno	105·0
2	1956 Apr. 6	07 11 38	36·4 N, 70·7 E	~220	6¾	Hindu Kush	Mt Hamilton	105·8
3	1956 June 9	23 13 51	35·1 N, 67·5 E	0	7½	Afghanistan	Reno	105·4
4	1956 July 18	06 19 34	5·07 S, 130·26 E	~130	7	Banda Sea	Mineral	107·1
							Mt Hamilton	107·4
							Reno	108·6
5	1957 Mar. 23	05 12 43	5·54 S, 130·97 E	147	7	Banda Sea	Mineral	106·8
							Reno	108·3
6	1958 Mar. 28	12 06 24	36·51 N, 70·98 E	188	6½	Hindu Kush	Reno	103·7
7	1960 Oct. 7	15 18 30·8	7·4 S, 130·7 E	45	6¾–7	Banda Sea	Mt Hamilton	108·5
8	1960 Dec. 14	23 51 31·5	3·0 N, 126·3 E	78	6¾	Molucca St	Mineral	104·7
9	1961 Mar. 28	09 35 55·4	0·2 N, 123·6 E	83	6¾–7	Celebes	Mineral	108·6
10	1962 Mar. 19	05 54 36·9	0·2 N, 123·6 E	150	6½	Celebes	Mineral	108·6
							Mt Hamilton	109·4
11	1962 July 6	23 05 32·2	36·6 N, 70·4 E	203	6¾	Hindu Kush	Berkeley	105·0
12	1962 Sept. 12	20 57 00·4	36·5 N, 69·2 E	50	6½–6¾	Hindu Kush	Priest	107·1
13	1964 Jan. 28	14 09 17·1	36·5 N, 70·9 E	207	6·1	Hindu Kush	Mt Hamilton	105·6
							Priest	106·9
14	1964 Apr. 7	13 18 18·9	0·1 N, 123·2 E	150	6·3	Celebes	Oroville	109·2
							Mt Hamilton	109·7
							Priest	110·8
15	1964 July 8	11 55 39·1	5·47 S, 192·77 E	165	6·5	Banda Sea	Oroville	107·8
							Mt Hamilton	108·1
							Priest	109·0
16	1964 Sept. 12	22 07 03·2	49·06 S, 164·21 E	33	6·9	Auckland Is.	Mt Hamilton	108·2
17	1965 Mar. 14	15 53 06·6	36·35 N, 70·73 E	219	6·6	Hindu Kush	Oroville	103·8
							Mt Hamilton	105·8
							Priest	107·1
18	1965 May 16	11 35 46·0	5·29 N, 125·67 E	36	6·2	Philippines	Priest	105·7
19	1965 Aug. 20	05 54 50·0	5·67 S, 128·62 E	326	6·2	Banda Sea	Oroville	108·8
							Jamestown	110·0
							Priest	110·0
20	1965 Nov. 21	10 31 49·7	6·13 S, 130·41 E	93	6·3	Banda Sea	Mt Hamilton	108·0
							Jamestown	108·8
							Priest	108·9

earthquakes listed in Table 1 were found to be of suitable size and location (103° < Δ < 110°). The locations of the shocks were taken from the International Seismological Summary (1955–59), the Bureau Central International de Séismologie (1960–62) and the U.S. Coast and Geodetic Survey (1963–65).

The 35 records were from Benioff vertical-component short-period seismographs at all stations except Reno where only a Sprengnether instrument with $T_0 = 2$ s, $T_g = 2$ s was available. The approximate instrumental magnifications at 1 s period were: Reno (~3600), Mt Hamilton (30 000), Mineral (30 000), Berkeley (30 000), Priest (20 000), Oroville (100 000), Jamestown (270 000). Records from the very-high-gain stations at Oroville and Jamestown were available only from 1963 and 1964, respectively.

Readings of significant onsets were made systematically without attempting to fit a particular travel-time curve. (The onsets identified as the reflections *pP*, etc. which arise from focal depths are excluded in this discussion.) Table 2 is a summary of the longitudinal phases observed up to the approximate arrival-time of the *PP* phase. Residuals (O–C) give the time difference predicted by either the Jeffreys-Bullen tables (*P*, *PP*, *PKIKP*), by a straight-line extrapolation of the Jeffreys–Bullen tables (*P* at Δ > 105°), or by the tables published by Bolt & O'Neill (*PKiKP*).

The quality of selection of each reading is subjectively rated as follows: a = excellent, b = good, c = fair, d = poor. The raw travel-times of the observed data are plotted in Fig. 1 (for *P*) and Fig. 2. The thicker lines represent the theoretical reference curves referred to above. Each datum is adjusted to correspond to a surface focus. We now comment on some properties of the observations.

Figs 1 and 2 are presented mainly to give a measure of the degree of scatter in the readings of a single phase from the diverse set of earthquakes. Because we are not concerned in this study with high precision of travel-times the scatter of each of

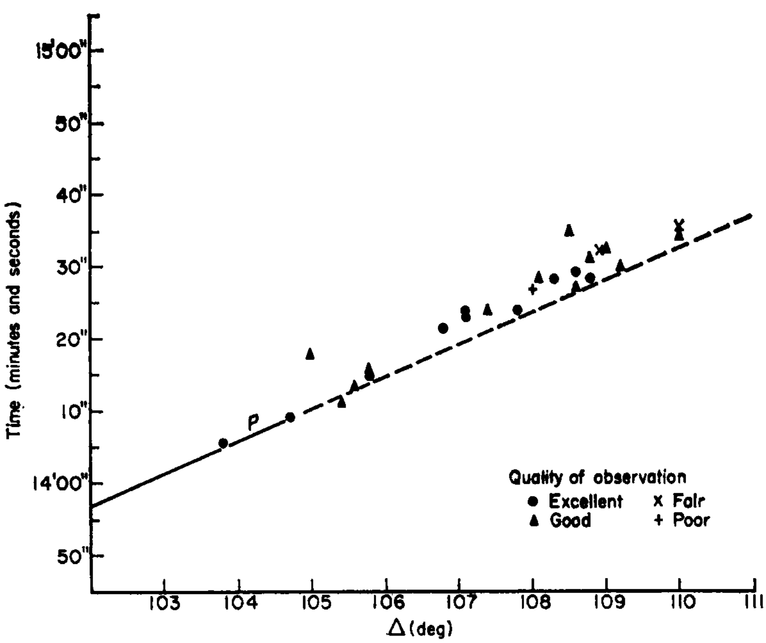

Fig. 1. Observations of times of travel of *P* waves to stations of the Berkeley network (see Tables 1 and 2). The dashed line is a linear extrapolation of the Jeffreys–Bullen table for *P* for a surface-focus.

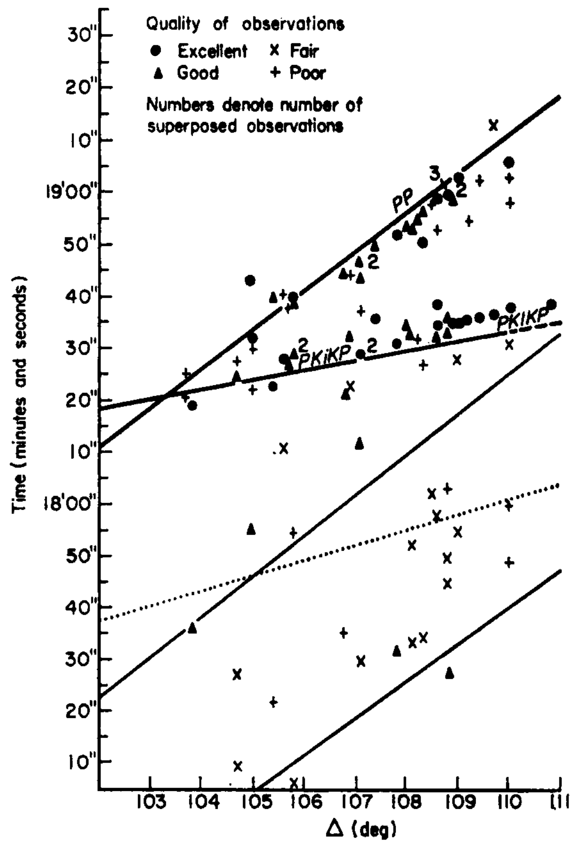

FIG. 2. Observed times of travel of *PP* and precursor-onsets to the Berkeley stations. The thick lines are *PP* and *PKiKP* times computed from the Jeffreys–Bullen tables. The dotted line is a linear extrapolation of the *GH* branch of *PKP* given by Bolt (1962). The thin continuous lines represent theoretical times of the phase *P*160*P* and *P*400*P* using upper mantle velocities derived by Johnson (1967) and the *PP* tables of Herrin *et al.* (1968).

the pooled *P* and *PP* times is not excessive and we may use these data to verify the existence of certain phases. The beginning of the *P* phase (104° < Δ < 110°) was distinct on 25 of the 35 seismograms; arrival-times tended to be several seconds later than times predicted by the linear extrapolation of the Jeffreys–Bullen *P* tables. The first few pulses of the *P* wave sometimes were significantly larger than those following; in Table 2, periods given in parentheses with a B, are values for these beginning pulses. Beyond 105°, the recorded longitudinal wave is generally considered to be diffracted by the mantle–core boundary. Whatever the propagation mechanism, the present study confirms that short-period *P* waves are clear out to 110°. The arrival-times of the *PP* phase (104° < Δ < 110°) were on the average several seconds earlier than times predicted by the Jeffreys–Bullen tables; the *PP* onsets are usually moderately sharp on the records from the short-period instruments although the path is not a minimum for all variations.

It is clear from Fig. 2 that there is a concentration of reliable onsets within a few seconds of the expected times for *PKiKP*. (The tendency for the times to be a few seconds later than the Jeffreys–Bullen *PKIKP* curves is predicted by the recent revisions of the latter curve (Bolt 1962).) The amplitudes were usually smaller and the predominant periods somewhat shorter than those observed for *PP*. Seven

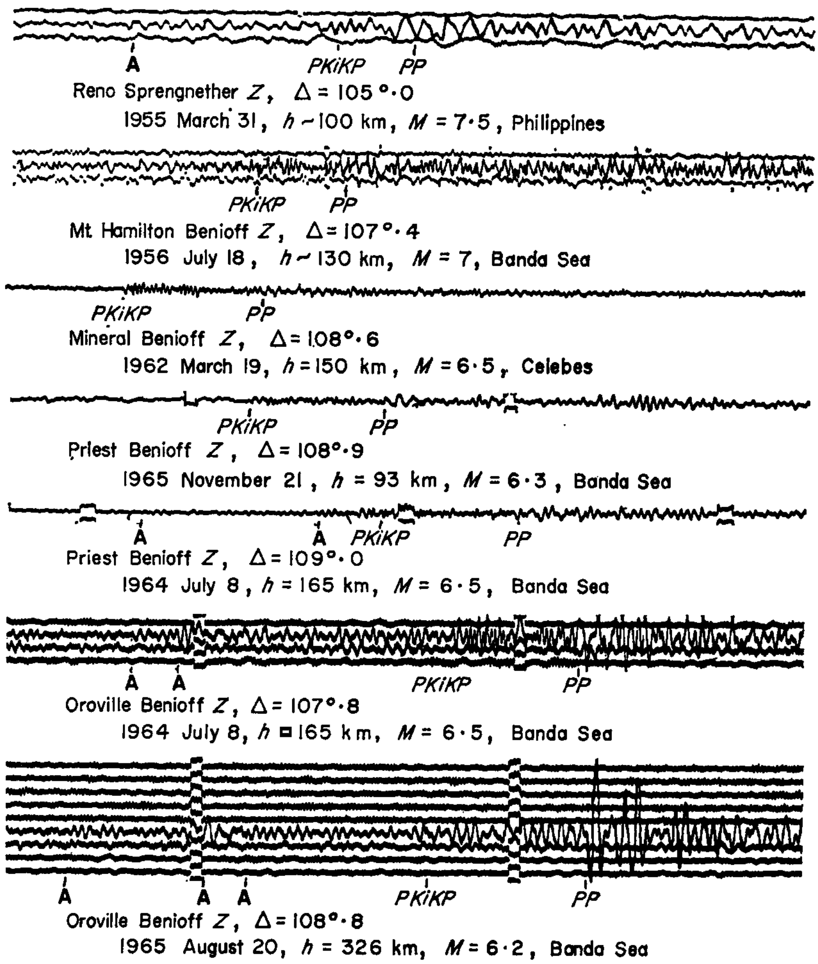

FIG. 3. Reproductions of seismograms from short-period vertical-component instruments at stations of the Berkeley network. The onsets marked **A** are examples of the pulses which occur in the early wave train.

examples of the seismograms are shown on Fig. 3. The occurrence of wave trains near the expected time for *PKiKP* is quite apparent in each case, particularly on the Reno (105°), Oroville (107°·8) and Mineral (108°·6) records.

We conclude from the above observational evidence that longitudinal waves with periods about 1 s can be traced along an extension of the *PKiKP* curve back from $\Delta = 110°$ to at least 105°. Nothing can be said at the adjacent shorter distances because of the *PP* arrivals. The amplitudes tend to decrease with decreasing distance back from 110° in agreement with a steady fall in amplitude computed by Bolt & O'Neill (1965). (No sudden increase in amplitudes near 110° is observed.)

Because the distance where the refracted core wave *PKIKP* first appears can hardly be much less than 110° and may be considerably greater (cf. Gutenberg 1960; Ergin 1967) it is difficult to escape the conclusion that the observed onsets arise from reflections from an inner core boundary. Further, in order to account for the relatively high frequencies of the observed *PKiKP* waves this boundary marks a rapid transition in elastic properties within 10 km or so.

We now digress somewhat to consider a reported observation of *PKiKP* for $\Delta < 105°$. Caloi (1963) drew attention to a small but definite phase recorded by a special vertical-component seismometer at Halifax, Canada some 80° from the large Iranian earthquake of 1962 September 1. The seismometer, designed by Willmore, strongly filtered out periods above 1 s; the maximum magnification of only 4000 occurred between 0·5 and 0·8 s. The observed phase which arrived 4 min and 20 s after *P* was identified as *PKiKP* by Caloi and he deduced from the travel time that the effective radius of the inner core was at least 1500 km. The predicted *PKiKP* amplitude computed by Bolt & O'Neill (1965) indicates that the identification as a reflection is an unlikely one.

We submit as an alternative explanation that the phase cited by Caloi is the *P* phase of an aftershock of the Iranian earthquake. As a test of this theory, seismograms from eleven short-period vertical seismographs at stations which had $30° < \Delta < 90°$ and which provided adequate signal-to-noise ratios were specially examined. Of these, five showed definite onsets about 04 m 20 s after *P*. These stations (with the instrument type and magnification at 1 s) were as follows: Toledo (Benioff, 45 000); Málaga (Benioff, 50 000); Alert (Willmore, 70 000); Mould Bay (Willmore, 100 000); Copenhagen (Benioff, 12 500). The alternative theory thus finds some observational support. If the cited phase was indeed *P* from an aftershock, the magnitude of this earthquake would be between $4\frac{1}{2}$ and 5 as calculated from the magnification curves for Mould Bay and Halifax.

3. Evidence on the early arrivals: $105° < \Delta < 110°$

Other features of the seismograms from the Berkeley stations were discovered which were as striking as the *PKiKP* phases. On many of the seismograms, quite perceptible wave energy was noticed arriving some three minutes after the *P* phase; this interval corresponds to about one min before *PKiKP* and about $1\frac{1}{2}$ min before *PP*. Often individual onsets could be selected and the arrival times of these are listed under ' Additional readings: before *PKiKP* ' in Table 2. The corresponding travel times are plotted in Fig. 2. A number of persuasive cases are reproduced in Fig. 3 where some of the sharper onsets are marked A. The Reno and Oroville records display the properties of the precursors to *PP* well. Pulse-like onsets were observed to be somewhat exceptional; more commonly, a train of waves with a somewhat indefinite onset can be traced for up to 90 s before *PP* proper arrives. On each Oroville record in Fig. 3, a very distinctive onset marked A occurs just on a minute time-break. The time intervals between these onsets and the observed *PP* waves are 75 s and 73 s, respectively; wave energy is clearly arriving, however, before these particular onsets.

The study of the early arrivals was subsequently broadened to include records from the World Wide Standardized Stations network. A useful shock for this purpose is the Chilean earthquake of 1965 March 28 located by the USCGS at 32°·43 S, 71°·20 W, with focal depth 61 km and origin-time $16^h\ 33^m\ 14{\cdot}6^s$. Seismograms from both short- and long-period instruments were examined for the 22 stations available in the range $100° < \Delta < 120°$. This examination confirmed the reality of the early arrivals and the general wave character described above. Whatever their genesis, some of the precursors possess a broad frequency spectrum because wave trains are often observable on the long-period records (see the NAI, AAE seismograms in Fig. 4) as well as on the Benioff records. A selection of records illustrative of these properties is reproduced in Fig. 4. The thick dashed line is the

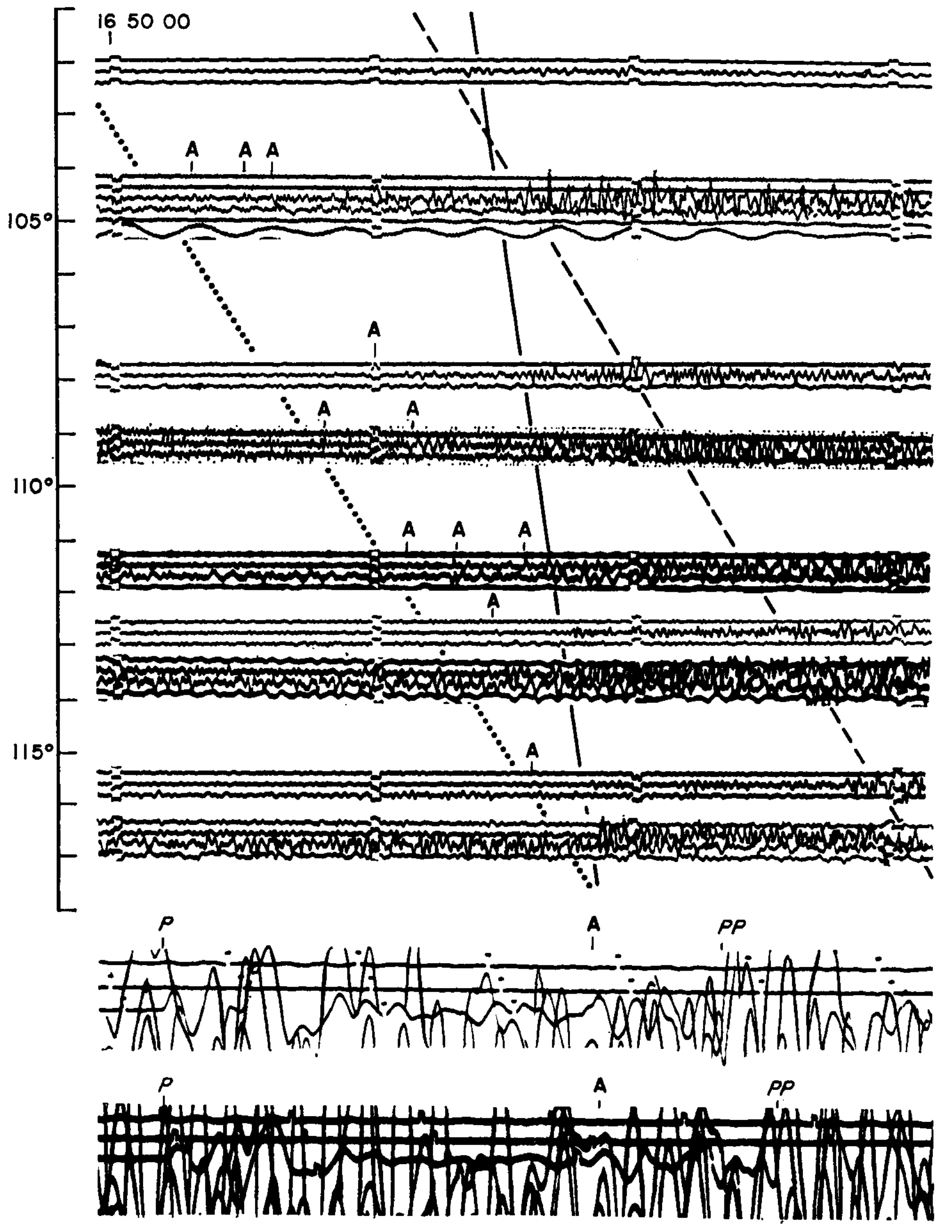

FIG. 4. Reproductions of *WWSS* seismograms from the 1965 Chilean shock. The stations, with distances, in order, are: GDH (102°·2), NAI (104°·5), STU (107°·9), TRI (109°·2), AAE (111°·6), COP (112°·8), COL (113°·5), MUN (115°·6), CTA (116°·6), NAI (104°·5), and AAE (111°·6). The records are from vertical component instruments; the first nine from the Benioff type—the last two from the Press–Ewing long-period type.

appropriate Jeffreys–Bullen curve for *PP*; the thin continuous line is the Jeffreys–Bullen curve for *PKiKP*; the dotted line is the computed travel-time curve for *P400P* as described for Fig. 2. The onsets marked **A** were selected prior to the computation of travel-time curves.

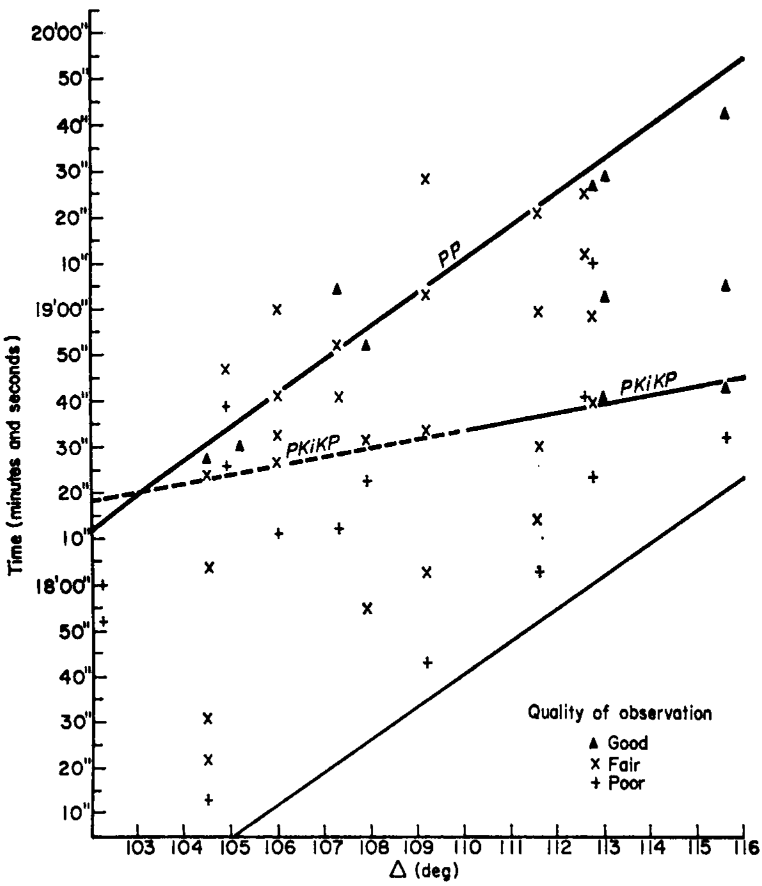

FIG. 5. Observed times of travel of *PP* and precursor onsets from the Chilean shock to *WWSS* stations. The thick lines are drawn from the Jeffreys–Bullen tables; the thinner line corresponds to the computed times for *P400P* derived as for Fig. 2.

The clearer onsets from the Benioff instruments were selected in the same way as in the Berkeley network study and the corresponding travel times are plotted in Fig. 5. The theoretical curves are as in Fig. 2.

We searched particularly seismograms from the 12 stations with $110° < \Delta < 120°$ to discover at what distance only *PKIKP* occurred prior to any other precursors to *PP*. This search indicated that in the case of the 1965 Chilean earthquake, no waves earlier than *PKIKP* can be seen for $117° < \Delta < 120°$.

The reality of the precursors to *PP* reported by seismologists in earlier studies is confirmed. The problem is to explain their existence in the light of the observed wave periods and duration of the train. Two hypotheses need to be considered:

(i) the waves are core phases which correspond to reflections from the outer boundary of the main transition zone between the outer and inner cores;

(ii) the waves never penetrate the core but are reflected from discontinuities in the upper mantle of the Earth (Nguyen-Hai's hypothesis).

The first hypothesis is the weaker. In the solution T2 of Bolt (1962), observed precursors to *PKIKP* for $125° < \Delta < 140°$ fall along a branch of the *PKP* travel-time curve called *GH*. The *GH* branch corresponds to refraction in a transition layer between the outer and inner cores with a boundary at radius 1667 km; it is followed within a second or so with the branch corresponding to the associated reflected waves. In Fig. 2, the dotted line represents the linear extrapolation of the *GH* branch back from the cusp at *G* at 125°. This curve does pass through the scattered early onsets but there is no special concentration of points about it; many onsets were selected up to ½ min earlier which would entail further sharp interfaces at larger radii in the core. Another objection is that the outer boundary to the transition shell separates shells with a small mismatch in elastic properties. The *P* velocity probably jumps by less than 5 per cent above 10 km/s; there are likely to be negligible changes in rigidity or density across the boundary.

We have calculated the amplitude ratio of the incident to the reflected *P* wave on this boundary, assuming a 0·5 km/s increase in *P* velocity, zero rigidity, and continuity of density. A distance of 110° would correspond to an angle of emergence at the outer boundary of the transition zone of 22°. Near this angle the amplitude ratio is about two-fifths to one-quarter. Because the observed energy carried by the refracted *PKP* waves is normally quite small, the calculated amplitude ratio suggests therefore that the energy reflected from the outside of the transition shell would not

Table 3

Travel-time interval for PP minus PdP (shallow focus) at $\Delta = 108°$

d (km)	160	200	250	300	350	400	450
PP–PdP (s)	38	46	56	65	75	84	91

be sufficient to explain the observed amplitudes of the early waves. It is normally quite difficult to observe the *GH* branch and the associated later reflections of *PKP* even for $\Delta < 130°$ so that the appearance of the reflections again at $\Delta < 115°$ would be extraordinary.

The second hypothesis seems physically reasonable on a number of grounds. First, the early wave trains on the records contain relatively longer-period energy than reflected core waves such as *PKiKP*. The periods appear visually to be comparable to those dominant in recorded *PP* waves. Secondly, the more prominent pulses which we have selected, although scattered, suggest that the first arrivals of the early waves fall along a curve which has, more or less, the gradient of *PP*. Thirdly, reflecting surfaces are more plausible for general geophysical reasons in the solid upper mantle than in the (convecting) liquid core.

Finally, the observed ratio of the amplitude of the clearer early waves to that of *PP* on the same record ranges from 0·1 to 0·5. Most work on the structure of the upper mantle based upon recent seismological evidence indicates a considerable change in physical properties between 360 and 420 km in depth. For example, Lane Johnson (1967, Fig. 8) has derived a *P* velocity distribution which shows a rapid increase from about 8·8 to 9·6 km/s in this range. In Johnson's solution there are also rapid increases in velocity near 160 km (bottom of a low velocity layer) and 650 km. Consider a sharp interface at 400 km depth with *P* and *S* velocities and densities above and below of (9·0, 9·5), (5·0, 5·3), (3·6, 3·9) in the usual units. The modulus of the computed amplitude ratio of the reflected to the *P* waves incident at the lower side is only about 1/25 for the angle of emergence which corresponds to the present distance range. (The *PP* ray parameter at these distances is $7·0 < p < 7·5$ sec/deg.) The amplitude comparison on the seismograms is with *PP* which corresponds to the wave refracted at the 400 km interface (amplitude ratio with the reflected waves equals 30). This refracted wave undergoes a reflection at the free surface of the Earth and a further refraction on passing again through the

400 km level. (We ignore other reflection surfaces and the Mohorovicic discontuinity.) The appropriate partition ratios are 0·8 and 0·94, respectively. It follows that the theoretical amplitude ratio between the early waves and *PP* should be greater than $\frac{1}{30}\times 1{\cdot}3\times 1{\cdot}1$, which equals 1/20. However, the *PP* wave proper passes two extra times through any low-velocity zone above 400 km; because attenuation of the higher frequencies in such a shell is likely to be relatively high (perhaps as much as 80 per cent) the observed *PP* amplitude is not representative of the harmonic components contained in waves reflected below the low-velocity zone. It appears, therefore, that the observed amplitudes of the *PP* and early waves do not rule out reflection in the upper mantle.

Suppose the second hypothesis is correct. Then the early waves are *PdP* and the difference in their travel times from that of *PP* can be computed once the depth d is known and some *P* velocity model for the upper mantle assumed. We have assumed the *PP* travel-time tables for a surface focus published by Herrin *et al.* (1968). These times are 6 s less at $\Delta = 110°$ than those of Jeffreys and Bullen and hence are somewhat more in agreement with the observed values (see Table 2). The velocity distribution of Johnson (1967), published as a graph, was adopted* and the usual integral formula used to determine the time correction to *PP* appropriate to reflection as a series of depths d. The resulting time intervals (*PP–PdP*) are listed in Table 3. The values are probably correct to one second.

For $105° < \Delta < 110°$, it is an adequate approximation to take the gradient p of *PdP* to be equal to that of *PP* at all distances.

As indicated by the data plotted on Fig. 2 (see also the first Oroville record in Fig. 3), the first perceptible *PdP* waves arrive, on the average, within a few seconds of 81 s before the observed *PP* waves†. It should be noted in interpreting Figs 2 and 5 that the *P400P* curve is computed from the Herrin *et al.* *PP* curve which if drawn on the figures would fall *earlier* than most of the observed *PP* phases. From Table 3, based on Johnson's velocity model and Herrin's *PP* tables, an interval of 81 s entails a reflection horizon for *PdP* at $d = 385$ km. Similarly, the two large pulses at Oroville shown as A in Fig. 3 with a *PP* minus *PdP* interval of about 74 s can be associated with a depth of 345 km to the reflecting surface.

The uncertainties in these depth-estimates may amount to a few tens of kilometers. The assumed mean *PP* time may be out by, say, 1–2 s; the onset time of *PdP* by 3–5 s. A change of 5 s in the estimated *PP* minus *PdP* time corresponds to a change of about 25 km in d. There is, thus, in broad terms, internal consistency between the appearance, for $105° < \Delta < 110°$, of waves some 81 s before *PP* arrives and the sharp change in the *P* velocity curve between 360 and 420 km depth derived by Johnson. The short-period prominent pulses of *PdP* which arrive at Oroville 12 s after the likely beginning of the train perhaps suggest that first order seismic discontinuities occur in this transition region rather than the smooth variation used by Johnson.

4. Conclusions

For epicentral distances between 105° and 115°, approximately, trains of longitudinal waves occur on seismograms in the interval between the onsets of *P* and *PP*, independently of the geographical location of the source. One set of clear onsets with periods as short as $\frac{3}{4}$–1 s are found to lie within a few seconds of the linear extension of the *PKIKP* curve predicted by the core model T2 of Bolt (1962). These onsets are probably the phase *PKiKP* reflected from the inner core; they add weight

* Herrin's velocities in the upper mantle are not directly derived from observations.

† Energy sometimes arrives *earlier*, however; see the second Oroville record in Fig. 3. The magnitude of the 1965 Chilean shock does not seem to have been quite large enough to produce *PdP* waves as early as 80–85 s before *PP*; they can, however, be seen between 70 and 75 s before *PP*.

to the view that the inner core has a sharp boundary within a few kilometres of 1220 km radius.

The remaining precursors to *PP* between 105° and 110° (ignoring depth phases such as *pP*) probably, as suggested by Nguyen-Hai, are not predominately core phases but most of the energy arrives by means of reflections and back-scattering at a number of physical discontinuities in elastic properties in the upper mantle of the Earth. We have named these waves the *PdP* phases. In particular, the measured times of the earliest identifiable onsets from the larger earthquakes are about 81 s less than the *PP* times to corresponding distances. These onsets provide important independent evidence for either two or more discrete jumps in the *P* velocity curve or a strong increase in its gradient at depths a little less than 400 km in the mantle in close general agreement with the result of Johnson (1967). On the evidence of this study, the first alternative is to be preferred.

Additional study of these early waves may well prove to be of importance in settling questions of fine structure in the upper mantle. In contrast with most possible reflections from the upper sides of discontinuities in the upper mantle, these waves occur in a portion of the time–distance domain which (for all but the deepest earthquakes) is probably free of other perceptible phases. Reliable measurement of the apparent velocities of the earlier arrivals *PdP* is now being sought using the LASA facility in Montana.

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