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Detection of *PKJKP* at intermediate periods by progressive multi-channel correlation

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Abstract

We have identified the seismic phase *PKJKP*, traveling through the inner core of the Earth as a shear wave, in intermediate-period (2–10 s) records of the deep 1996 Flores Sea earthquake at eight stations of the French seismic network. This constitutes direct evidence of the solidity of the inner core which, while generally recognized, was until now only inferred from indirect evidence. The arrival times on stacked seismograms require a shear-wave velocity at the top of the inner core $\beta_{ICB} = 3.65$ km/s in agreement with values suggested from normal mode observations. Julian *et al.*'s 1972 observation of a lower velocity (2.95 km/s) is easily reconciled with our result if interpreted as the surface reflection *pPKJKP*. The high Poisson ratio ($\nu = 0.44$) can be reconciled with a normal crystalline structure without invoking partial melting on account of the overwhelming pressure at the center of the Earth.

Keywords: PKP-waves; S-waves; inner core; earthquakes

1. Introduction

The solidity of the Earth's inner core has generally been inferred from relatively circumstantial evidence, such as the observed jump in compressional wave velocity at the inner-outer core boundary [ICB] [1, 2], the reflected phases *PKiKP* and *PKIIKP* [3], or the pattern of coupling between certain poloidal normal modes of the Earth and the inferred shear modes of the inner core [4]. Direct evidence, *i.e.*, the recording at the Earth's surface of a body wave such as *PKJKP*, traveling through the inner core in the shear mode *J*, had until recently eluded seismologists ever since Birch [5] first proposed in 1940 that the discontinuity earlier discovered by Inge Lehmann [6] indeed corresponded to the freezing of the innermost part of the planet under the overwhelming pressure reached at its center. The only exception to this lack of observations, the 1972 report by Julian *et al.* [7] proposing an inner core shear velocity $\beta_{ic} = 2.95$ km/s, was generally dismissed because it could not be reconciled with the observations of normal mode coupling [4].

The occurrence in 1994–96 of three large deep earthquakes (with moments greater than 3×10^{27} dyn-cm), and the implementation of large-aperture digital seismic arrays provide new opportunities to

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apply recently developed sophisticated data processing techniques in the hope of directly detecting inner core shear waves. In this report, we use the Progressive Multi-Channel Correlation [PMCC] method [8] to extract the phase *PKJKP* at intermediate periods (2–10 s) generated by the deep Flores Sea, Indonesia earthquake of 17 June 1996, from records of the French short-period seismograph network. An independent observation by Deuss *et al.* [9] at lower frequencies (typically 30 mHz) is being reported concurrently.

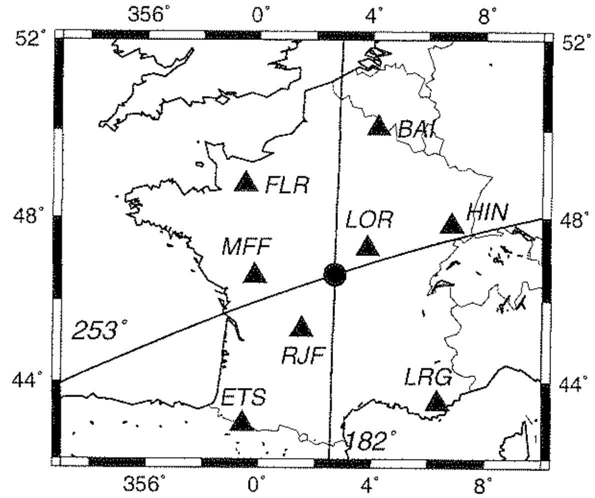

Fig. 1. Map of the receiving array used in this study. The individual stations, with 3-letter codes, are shown as triangles; the dot shows the centroid of the array, and the transverse arcs are the great circles to the epicenters of the 1996 Flores Sea event (azimuth 253°) and the 1994 Fiji event (azimuth 182°).

2. Methodology

The PMCC method is an improved time-frequency analysis algorithm which can be summarized as follows. For each window in the time series, and within each frequency band, cross-correlation functions are computed for all pairs of available seismograms and the consistency of their maxima is analyzed: in the presence of a coherent wavetrain throughout the array, closure relations of the form

$$\tau_{ijk} \equiv t_{ij} + t_{jk} + t_{ki} = 0 \qquad (1)$$

are expected to hold between the delays t_{ij} maximizing the various cross-correlations; on the other hand, the failure of closing relations ($\tau_{ijk} \neq 0$) is characteristic of uncorrelated noise. When coherence is achieved, the system keeps track of the optimal slowness vector $\mathbf{p}$ across the array (with azimuth ζ and modulus $p = C^{-1}$ where C is the apparent phase velocity), as well as of the maximum number of coherent records, with all this information computed and stored for each "pixel" combination of time and frequency. The advantage of the method is that it analyzes each pixel irrespective of its neighbors, for which energy may be propagated at a different $\mathbf{p}$. It has been successfully applied to the detection of weak signals masked at single stations by overwhelming noise [8, 10].

We applied the PMCC method to 8 stations of the French short-period seismic network, shown on Fig. 1, and comprising a regularly shaped array, 600 km across. Distances from the Flores Sea epicenter ranged from 112.9° to 119.3°. We considered windows of 20 s duration, extending from 1602 to 1912 s after origin time which, for a source depth of 584 km, should be adequate for the detection of *PKJKP* and *pPKJKP* for values of β_{ICB}, the shear velocity at the top of the inner core, ranging from 2.5 to 4.0 km/s. The frequency band investigated was 0.1 to 0.5 Hz; the response of the short-period instruments is inadequate at lower frequencies while higher frequencies are not expected to survive anelastic attenuation in the inner core [11], and are at any rate dominated by the various branches of the fully compressional phase *PKKP*.

For characterization of the phase *PKJKP*, we note that because of the steep refraction at the ICB, it curls around the center of the Earth and emerges as a retrograde phase, expected to sweep the French network from a geometrical back-azimuth $\zeta = 253°$. In addition, any phase penetrating the inner core must have a slowness $p = r \sin i / v$ less than its value $r_{ICB}/\alpha_{oc,ICB}$ = 2.06 s/° for a K ray grazing the ICB at horizontal incidence. The corresponding apparent phase velocity constraint is $C \geq 54$ km/s.

After verifying that the phase *PKKP* is correctly detected in the higher-frequency band 0.5–2 Hz (including its branch *df* which penetrates the inner core), we proceeded to build the PMCC diagram shown on Fig. 2. Fig 2d. expresses the signal consistency between stations, as the root-mean-squares (in s) of the closure misfits τ_{ijk} in (1) for all

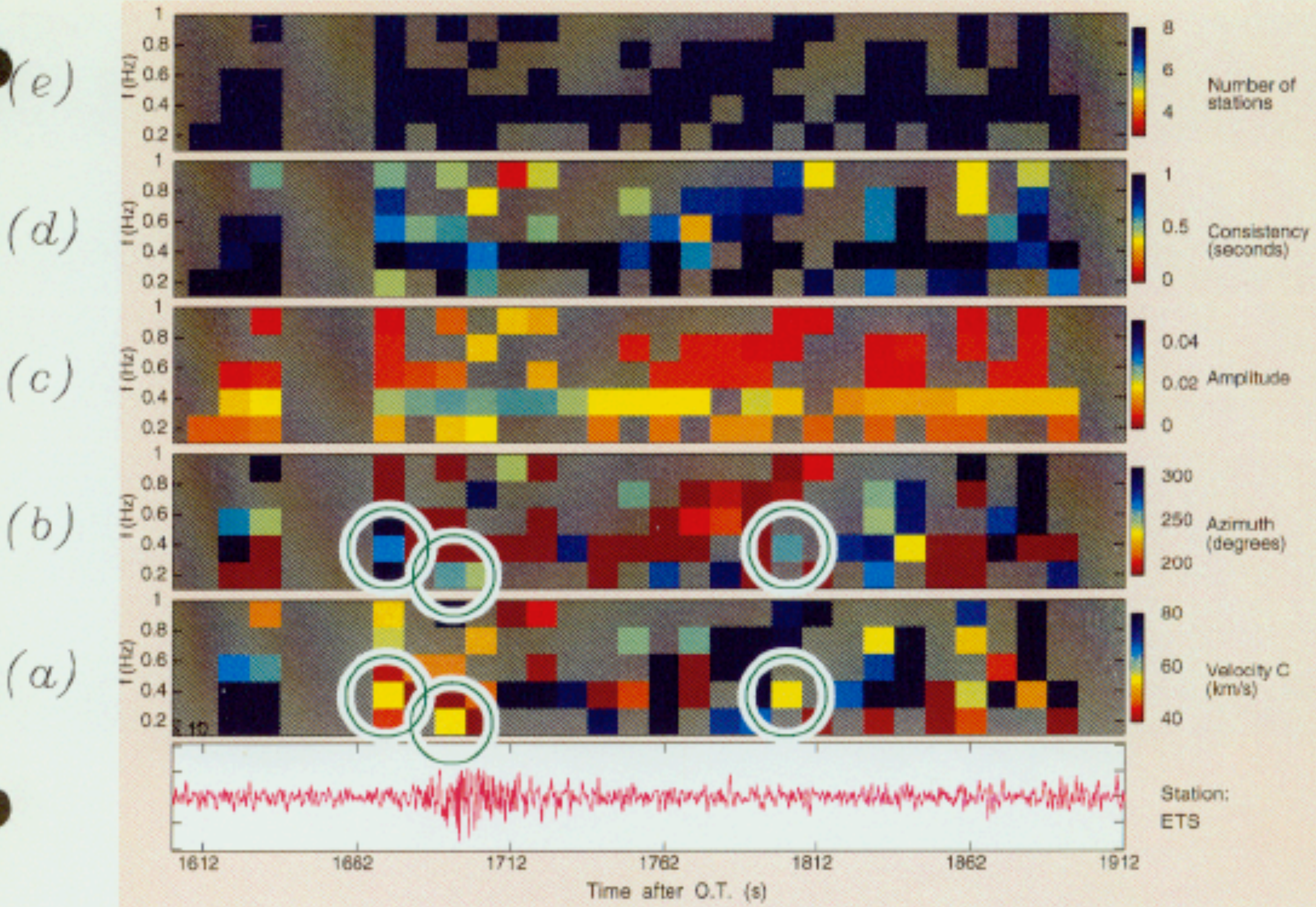

Fig. 2. Output of the PMCC algorithm applied to the Flores Sea dataset. Each of the top five frames is decomposed in a series of time-frequency pixels. In each case when coherence is achieved, the pixel is colored according to the value of the parameter best-fitting the correlation between seismograms. From the bottom: (a) apparent phase-velocity across the network in km/s; (b) back-azimuth of slowness vector; (c) amplitude (arbitrary but common scale) of signal; (d) consistency, defined as the root-mean-squares (in s) of the closure misfits among times maximizing cross-correlations; (e) Number of stations participating in coherent arrivals (in the present case, all stations are coherent for all coherent pixels). The bottom trace is an example of seismogram at ETS. The white and green rings indicate the only pixels where the combination of velocity *(a)* and back-azimuth *(b)* is adequate for *PKJKP*.

possible combinations of i, j, k. Figs. 2a and b present the polar coordinates of the best-fitting slowness vector **p** for those pixels where a consistency is achieved at less than 1 s (note that the modulus of **p** is expressed through its inverse, C, in km/s). Fig. 2c keeps track of the amplitude of the resulting slant stack in each pixel, and Fig. 2e expresses the number of coherent stations (in this particular example, all eight stations are coherent in all consistent pixels). Pixels shaded in background gray failed the 1-s consistency test.

Estimates of the travel times and slownesses of *PKJKP* were computed for a variety of models of the inner core assuming the same velocity gradients $(\partial \ln \beta_{ic})/(\partial r)$ as in model PREM [12] but with shear velocities at the ICB ranging from 2.5 to 4.5 km/s. At the target distance for the center of the network, the apparent phase velocity C is expected to range from 55 to 70 km/s. We thus explored the diagram on Fig. 2 for arrivals at such values of C and at an azimuth ζ close to 253°. This amounts to seeking on Fig. 2 a pixel combination of time and frequency which would show up as yellow-to-greenish in Fig. 2a, blue green-to-light blue in Fig. 2b, and other than red (to feature sufficient amplitude) in Fig. 2c.

3. Results

We identified three such pixels, circled in white-and-green on Figure 2, at respectively C = 55.6 km/s, ζ=272°; C = 56.8 km/s, ζ=262°; and C = 56.9 km/s, ζ=264°. These are the only windows in time and frequency where sufficient energy is sweeping the array at the proper combination of phase velocity and azimuth. We regard the slight discrepancy in azimuth in the vector slowness relative to the geometrical retrograde direction (253°) as typical of the effects of lateral heterogeneity under the receiving array [13]; note that these will be exacerbated when the scalar slowness becomes very small, as is the case for all waves sampling the inner core.

In order to further interpret these arrivals, we built spectrograms for stacked time series obtained by beaming the array using the relevant slowness vectors. Fig. 3 shows the case of the second stack (C = 56.8 km/s; ζ = 262°). The time scale on that figure is the unaltered time at the first station reached by the wave packet (in the case of this retrograde phase, the most distant one, ETS, at 119.3°). The center frame on Fig. 3 shows a spectrogram covering the frequency range 0.1–1.0 Hz. It is dominated by the phase $PKKP_{df}$ at frequencies close to 0.6 Hz (the slowness of $PKKP_{df}$ across the array (1.88 s/° or C =59.1 km/s) is too close to the stacking slowness (see Fig. 4a) to fully annihilate this high-amplitude phase from the stack). The bottom frame, which is a close-up spectrogram at lower frequencies, shows a practically continuous arrival of energy between 1634 s and 1677 s in the frequency range 0.15 to 0.45 Hz.

For the purpose of identifying this wave packet, we note that any candidate phase must match both travel time and vector slowness constraints, as well as feature a low-frequency spectrum, peaked below 0.5 Hz. The latter is of course, best achieved by requiring that at least of portion of the path be traveled as a shear wave. In particular, the observed slowness (1.96 s/°) is less than 2.06 s/°, meaning that the relevant ray must have its turning point (if it has one or more) in the inner core. This eliminates mantle phases such as PS or PPS whose travel times (but not slownesses) would approach the observed ones. Furthermore, the phase cannot be $PKKP_{df}$, whose slowness is comparable, but which arrives 1706 s after origin time (*i.e.,* 30 s after the end of this low-frequency wave packet) and is well detected by PMCC at 1–2 Hz; nor can it be $PKIIKP$ which at a similar distance ss higher-frequency and would arrive much earlier (1280 s) and with a much lower slowness (1.40 s/°).

The observed wave could also, in principle, correspond to a ray reflected upwards before it penetrates the inner core, either a multiple PcP, a multiple ScS, or a multiple $PKiKP$; however no feasible combination thereof would fit both the relevant travel time and retrograde slowness. In particular, phases diffracted at the core-mantle boundary, such $pPcPPKP_{diff}$ would have a much greater slowness. Finally, Earle and Shearer [14] have interpreted arrivals with similar travel times and slownesses as precursors to $PKKP$ resulting from scattering by lateral heterogeneities along the core-mantle boundary; the low-frequency character of our arrivals would however argue against such an interpretation.

We therefore suggest that the wave packet observed between 1634 and 1677 s on the stacked seismogram is indeed $PKJKP$. The duration of the phase is consistent with the source duration of 20 s inverted from body waves by Goes *et al.* [15] for the Flores Sea event. It may be longer than that of $PKKP_{df}$ due to the significant anisotropy in the inner core. According to various models [16, 17] derived from the study of $PKIKP$, the latter is expected to be as large as 3% with uniaxial symmetry about an axis close to the Earth's rotation axis. In our geometry, the inner core is traversed for 660 s along a ray inclined 61° to the rotation axis, which would lead to a substantial splitting of the two polarizations along the J segment, with a time lag of 20 s between them. When combined with source duration, the total duration of $PKJKP$ could be expected to be as long as 40 s.

The arrival time of the first pulse of $PKJKP$ at ETS, 1634 s after the reported origin time at the hypocenter, can be used to assess the shear velocity in the inner core, by comparing with the models shown on Fig 4b. In this respect, we choose to interpret the travel-time rather than the observed slowness, since the latter will be affected more by lateral heterogeneity under the receivers. The result is β_{ICB} =3.65 km/s. This value is in good agreement with that inferred from normal modes [4] and also with Deuss *et al.*'s observation at much lower fre-

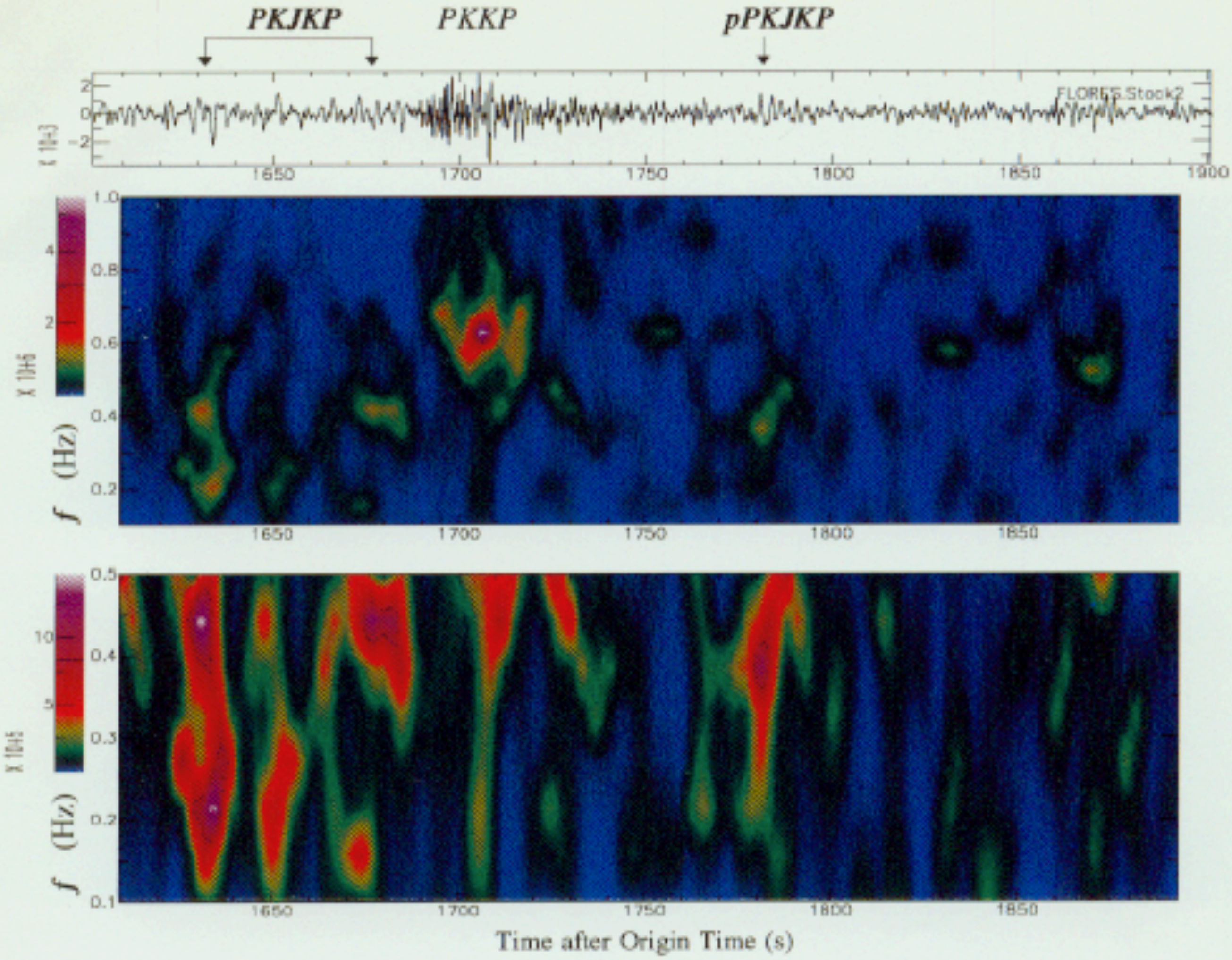

Fig. 3. Top: Stacked record obtained by beaming the array with $C = 56.8$ km/s; $\zeta = 262°$. The time scale assumes no lag at station ETS ($\Delta = 119.3°$). *PKJKP* is comprised of the energy spread out between 1632 and 1677 seconds. Note that the phase *pPKJKP* is clearly observable in the time domain. The higher frequency phase $PKKP_{df}$ is also present. Center and bottom: Spectrograms of the record using 15-s windows lagged by 1.5 s. The center frame covers the range 0.1–1.0 Hz; note that it is dominated by the remnants of the phase $PKKP_{df}$, not fully annihilated by the stacking procedure. The bottom frame is a close-up in the range 0.1–0.5 Hz, where the dominant phases are *PKJKP* and *pPKJKP*.

quencies [9], but significantly higher than suggested by Julian *et al.* [7]; the reason of this discrepancy is discussed below.

The later arrival observed 1782 s after origin time, with $C = 56.9$ km/s; $\zeta = 264°$ in the frequency band 0.3–0.5 Hz, lags 130 s behind the bulk of the *PKJKP* packet and can be interpreted as the source-side surface reflection *pPKJKP*; it is prominent in the stacked time series (Fig. 3), despite a shorter duration, weaker amplitude, and simpler spectrum than *PKJKP*. The origin of these discrepancies is presently not clear. As for the phase *SKJKP*, predicted around 1840 s, it could not be detected, probably due to an unfavorable focal mechanism (the radiation coefficient at the source [18] is $R^{SV} = -0.05$ as opposed to $R^{P} = -0.74$).

As for other large deep earthquakes, we failed to detect *PKJKP* or *SKJKP* from the Bolivian event of 09 June 1994, the largest deep shock ever recorded. At the relevant distance (88° at the center

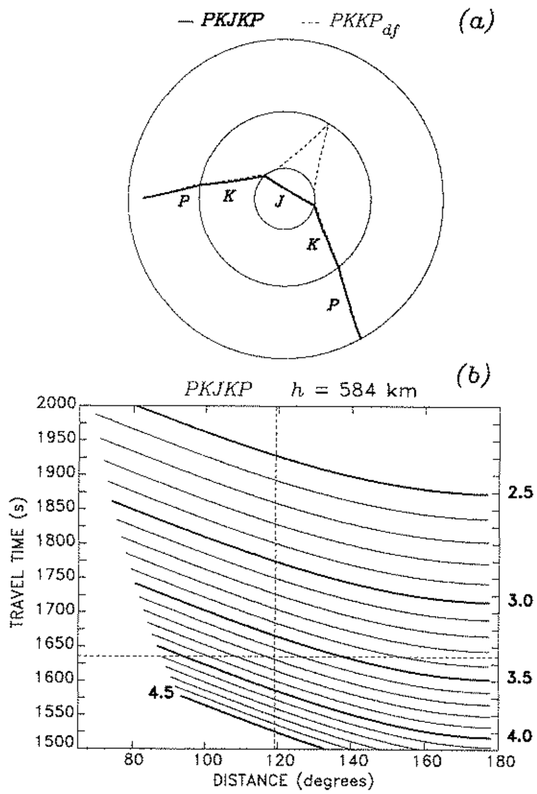

Fig. 4 (a): Paths of *PKJKP* and $PKKP_{df}$ through the Earth in the geometry of the Flores-to-France experiment (h = 584 km; $\Delta = 119°$). Note the proximity of the slowness values, as evidenced by the near coincidence of the paths through the mantle and outer core. (b): *PKJKP* travel-time curves computed for various values of the shear velocity in the inner core. The logarithmic velocity gradient is taken after PREM [12], and the velocity at the ICB, β_{ICB} is sampled at 0.1 km/s intervals, with multiples of 0.5 km/s drawn fat and labeled at right (in km/s). The dashed lines correspond to the observations on the stacked record.

of the French array), this can be explained easily from the unfavorable focal mechanism in the case of *PKJKP* ($R^P = -0.03$) and the significant geometrical spreading in the case of *SKJKP* ($d\Delta/dp = 440\,(°)^2$/s as opposed to only $61\,(°)^2$/s for the Flores Sea event *PKJKP*, and $81\,(°)^2$/s for its undetected *SKJKP*).

In the case of the third large deep earthquake of the past few years, the Fiji event of 09 March 1994, the PMCC method gives a weak signal 1732 s after origin time (with C = 102 km/s (p = 1.09 s/°) and $\zeta = 167°$ in the frequency band 0.3–0.5 Hz) which can be interpreted as *pPKJKP* with $\beta_{ICB} = 3.62$ km/s, for the significantly larger distances involved (150.4° at the center of the array where the geometrical retrograde back-azimuth would be 182°). Note in particular that this arrival at a retrograde slowness cannot be interpreted as a compressional phase such as *PKIKP* from one of the many aftershocks (some of them most probably unreported) which followed the Fiji event [19].

4. Discussion

It has generally been argued [20] that *PKJKP* should not be observable, on account of both poor conversion coefficients at the ICB, and strong anelastic attenuation in the inner core. We wish to discuss and quantify these assertions.

First, we show on Figure 5 the full transmission coefficient T for the phase *PKJKP*, defined in the notation of Aki and Richards [18] as the product of the transmission coefficients $\grave{P}\grave{S}$ at the $K \rightarrow J$ conversion and $\acute{S}\acute{P}$ at the $J \rightarrow K$ conversion, as a

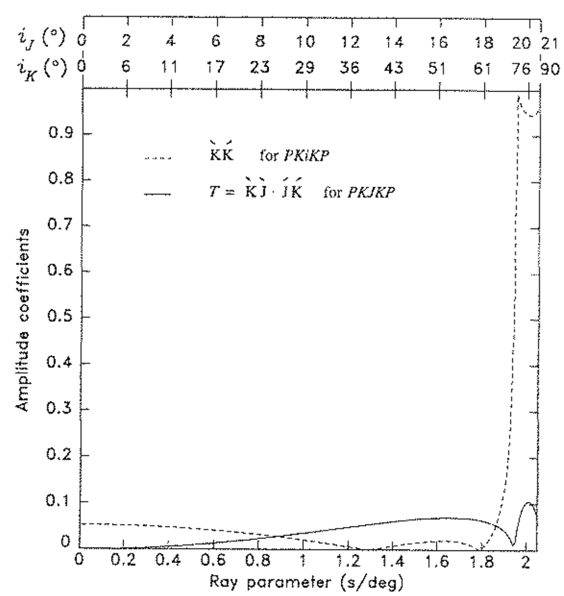

Fig. 5. Reflection and transmission coefficients computed at the ICB. The solid line is the expected full transmission coefficient for *PKJKP*, resulting from both the $K \rightarrow J$ and $J \rightarrow K$ conversions. For comparison, the dashed line is the reflection coefficient for *PKiKP*. The horizontal scale is linear in slowness p; the top bar gives the corresponding angles of incidence at the ICB for the K segment, and of refraction into the inner core for the J ray. Note that the latter is always very steep.

function of the ray parameter p, and computed using density and P–wave values from the PREM model [12], together with the β_{ICB} value measured in this study (3.65 km/s). Note that T grows with the angle of incidence on the ICB, and in particular reaches a maximum of 0.105 when the phase K becomes post-critical for I ($1.93 \leq p \leq 2.06$ s/° or $54 \leq C \leq 58$ km/s), precisely in the range of slownesses observed in the Flores-to-France geometry. While this figure of 10% remains admittedly small, it is comparable to, if not larger than, the reflection coefficient for *PKiKP* at short distances, a phase that has been successfully observed and studied at high frequencies on single-station records following nuclear blasts [21].

As for the effect of attenuation, we note that an average value $Q_\mu = 110 \pm 25\%$ has been proposed for the whole inner core based on normal mode measurements [11, 22], but that many recent models suggest a "mushy" zone at the top of the inner core (possibly ~200 km thick and related to its ongoing solidification at the expense of the outer core), where attenuation could be twice as strong [23, 24, 25]. Also, some authors have proposed that low values of the quality factor of *PKIKP* may be due to finite bulk attenuation Q_K^{-1} [26]. The phase *PKJKP* would be insensitive to the latter, and because of its steep refraction towards the center of the planet, would be exposed only minimally to the high attenuation in any mushy outer shell of the inner core. Thus, in such models, and because a normal mode integrates attenuation over the volume of a sphere, while Q^{-1} for a ray is computed as a line integral through the structure, Q_J would be expected to be greater than Q_μ obtained from normal mode studies. Using the tentative value $Q_J = 140$ (which corresponds to the high end of the error bar in [11]), and combining it with a travel-time of the J phase through the inner core of typically 660 s, this leads to $(t^*)_J = 4.7$ s, which is not significantly different from that for mantle S waves. The frequency content of *PKJKP* would then be expected to be comparable to that of a classic teleseismic S wave; while *PKJKP* would not be expected to be observable at frequencies $f \geq 1$ Hz typical of other deep core phases (such as *PKKP* or *PKiKP*), there is in principle no reason why the phase should not be present at the lower frequencies where we detect it.

The combined effect of refraction and attenuation may well bring the amplitude of *PKJKP* below single-station noise levels; this is precisely the reason why powerful stacking procedures such as the PMCC method are required to extract the phase.

Finally, we believe that Julian *et al.* [7] actually detected the source-side reflection *pPKJKP* but misidentified it as *PKJKP*. In the case of the 1967 Fiji earthquake analyzed by vespagram in their Figure 1, once the source depth is corrected to 658 km [27], the observed stacked time at LASA (17:52:59 GMT or 1865 s after origin time) would fit *pPKJKP* at the relevant distance ($\Delta = 94.3°$) with $\beta_{ICB} = 3.50$ km/s. This time would also fit *SKJKP* but, based on our failure to detect the phase at a similar distance during the Bolivian earthquake, we prefer the former interpretation. As to the difference between the values of β_{ICB} required to fit our results (3.65 km/s) and theirs (3.50 km/s), it may be ascribed to a combination of imprecision in the method, anisotropy, and possibly lateral heterogeneity in the top of the inner core (the Flores-to-France path samples the inner core along an arc located under the South Pacific and South America, while the Fiji-to-LASA one is under the Indian Ocean, Africa and the North Atlantic).

The bottom line remains anyway that Julian *et al.*'s observation is easily reconciled with shear velocities in the range 3.5–3.65 km/s, and does not require the exceedingly low values which they suggested and which led to the general dismissal of their observation; there can be no doubt that these authors had indeed picked up a seismic phase having traveled the inner core in the shear mode J.

In conclusion, this study provides direct evidence for shear waves in the inner core, and hence, for its rigidity. In this respect, it can be seen as a final, and to some extent reassuring, confirmation of the inferences drawn for several decades from more indirect observations. The shear-wave velocity, β_{ICB} at the top of the inner core entices a Poisson ratio $\nu = 0.44$. While this number is undoubtedly high, it does not require that the material be necessarily in an anomalous state involving, for example, partial melting. Indeed, metals such Au and Pb, known for their malleability, but hardly describable as mushy or partially molten, have Poisson ratios of 0.41–0.42 at ambient conditions [28]. Furthermore, Falzone and Stacey [29] have demonstrated that a significant increase in Poisson ratio should be a general property of crystalline materials under pressures becoming comparable to their elastic constants (p/K reaches 1/4 in the inner core). While there remains considerable debate on the question of the exact crystal structure of Fe (and its alloys) in the inner core, it is most probable that the latter can be regarded, at least in its bulk interior, as fully solidified.

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