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SEMIANNUAL TECHNICAL REPORT
NORSAR PHASE 3

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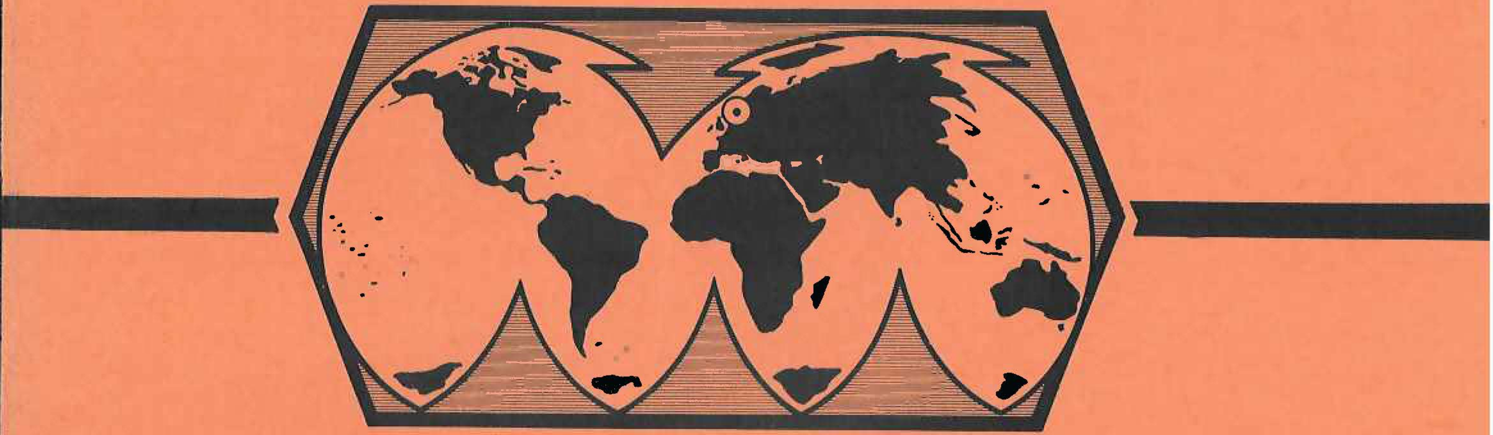

A. THE ANELASTICITY OF THE INNER CORE

Core phases traversing the inner core are well known to be relatively weak, if observable at all. Although the physical parameters in the inner core and near the inner core boundary are still less constrained than in any other region of the earth, there is at present no indication to attribute the "anomalous" attenuation of the inner core phases to the boundary in the velocity-density model. As reported in the previous Semiannual Report, models with a complicated transition zone have no observational basis anymore, following the array analysis of precursors to core phases. A single sharp boundary with a jump in rigidity and density is consistent with observations of normal modes (Dziewonsky and Gilbert 1973) and of PKiKP at short epicentral distances (Engdahl et al 1970). What remains to be investigated is the effect of anelasticity. The construction of an anelasticity model for the inner core in terms of Q for P waves and a study of some implications have recently been completed (Doornbos 1974). The Q model has been derived from spectral ratios of core phases with common source and receiver and with ray paths in and just outside the inner core. A frequency band around 1 Hz has been analyzed. In order to reduce the effect of attenuation sources (including scattering) outside the inner core, it is necessary that each two phases are chosen so that their ray paths outside the inner core are nearly like each other. This condition stresses the problem of finding suitable data. Short period NORSAR data have been used to extract the relevant pairs of phases, including (PKSKP, PKiKP), (PKSKP, PKP(BC)), (PKSSKP, PKiKP), (PKSSKP, PKKP(BC)). The array is needed in particular to separate PKSKP and PKiKP, and to identify PKSSKP. These phases are shown in Figures A.1 and A.2.

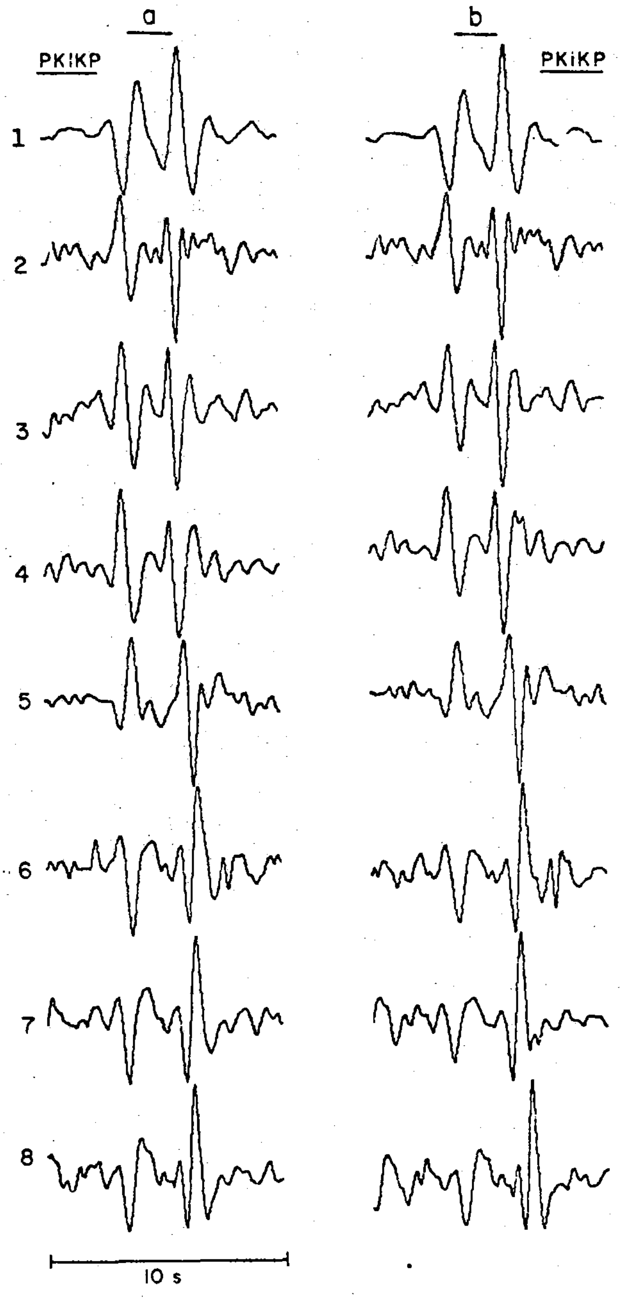

Figure A.1 Array beams steered at PKSKP(a) and PKiKP(b) from deep focus events in the Fiji Islands region. Epicentral distances from NORSAR are between 138°-141°. The traces have been normalized to peak amplitude.

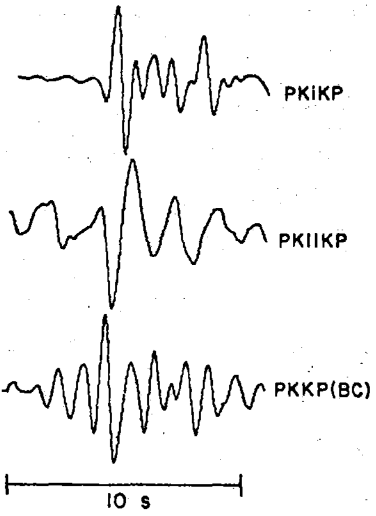

Figure A.2 Core phases from an event with or.time=4 Apr 1972, 22hr, 43 min, 6.7 sec., magn.=6.6, depth=377 km. Epicentral distance from NORSAR is 108.4°. The traces have been normalized to peak amplitude.

Some of the spectral ratios have been derived from spectra forms, but mostly the array beam should be formed, in order to reduce interfering signals and noise. Rather than using the conventional power spectrum of the array beam, I have preferred to derive the spectral ratios from the maxima of instantaneous spectral amplitudes, as introduced by Dziewonsky et al (1969) in the analysis of surface waves. It can be shown that the instantaneous amplitude at center frequency ω_o follows from

$$A(\omega_o,t) = \left|2 \int_{-\infty}^{0} F(\omega,\omega_o)e^{i\omega t} \frac{d\omega}{2\pi}\right| \qquad (1)$$

where $F(\omega,\omega_o)$ represents the filtered beam trace in frequency domain. It is convenient to use a Gaussian filter:

$$F(\omega,\omega_o) = B(\omega)\cdot\exp\{-\alpha(\omega-\omega_o)^2\} \qquad (2)$$

The parameter α should be chosen so as to compromise between the effects of time and frequency windowing. Figure A.3 gives an example of the instantaneous spectral amplitudes of PKSKP and PKiKP with $\alpha=0.35$. The figure shows that in a fixed time window interference, not only of PKSKP and PKiKP, but also of PKP precursors may be important. Use of instantaneous spectral amplitudes minimizes the effect of this interference.

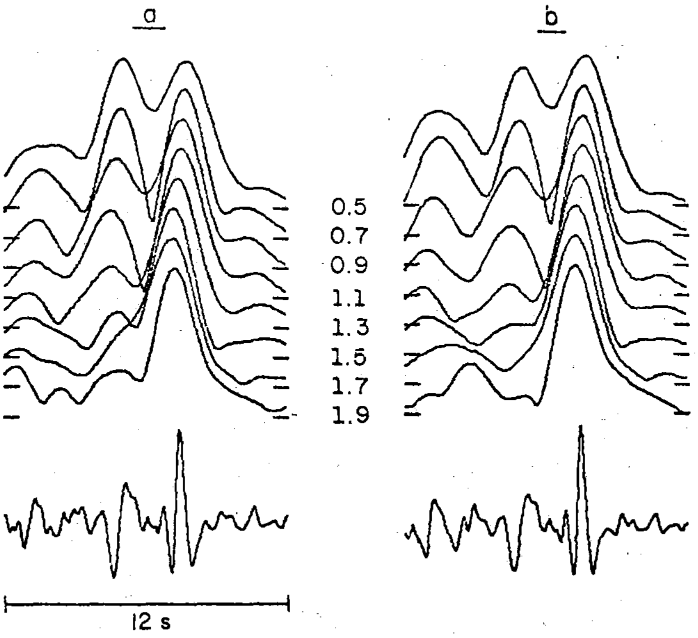

Figure A.3 Instantaneous spectral amplitudes at center frequencies 0.5-1.9 Hz, of array beams steered at PKSKP (lower trace, a) and PKiKP (lower trace, b) from a Fiji Islands event (nr. 8, Figure A.1). The traces have been normalized to peak amplitude.

Under certain assumptions we obtain from the j-th event

$$(t_1^*-t_2^*)_j = a_j + \pi \int_{SC} Q^{-1}(r)\ dt \qquad (3)$$

where the travel time integration is through the inner core, a_j includes terms arising from ray path differences outside the inner core, and the differential attenuation $(t_1^*-t_2^*)_j$ follows from the spectral ratio. The above system of operations can be discretized and inverted. The results corresponding with three different velocity models for the earth are shown in Figure A.4, where it is emphasized that the value Q=1000 in the lower part of the inner core does not rest on firm observational evidence. In this part Q values between limits 600 and 2000 are still possible with the present evidence.

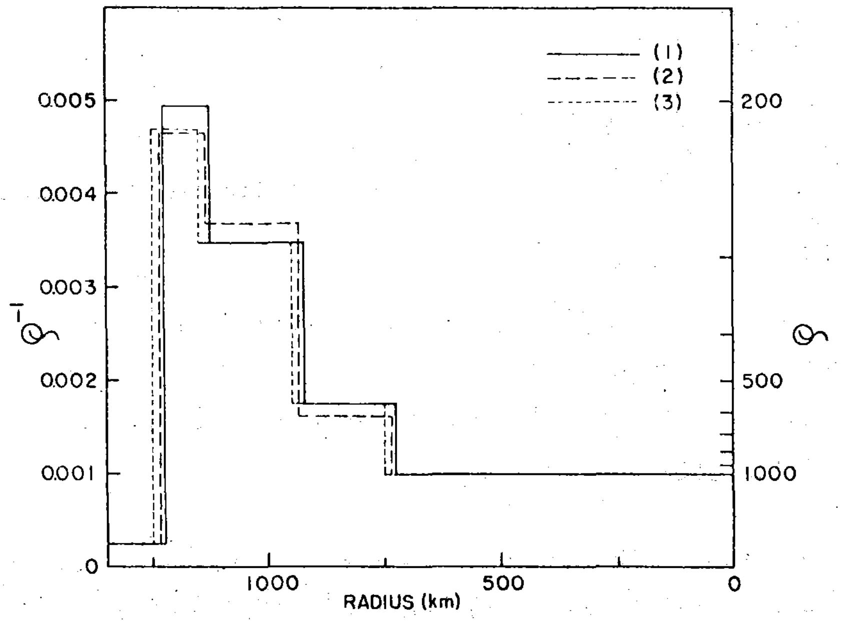

Figure A.4 Q_α at 1 Hz in the inner core, corresponding with different velocity models: (1)=Buchbinder 1971 (Bull. Seism. Soc. Am., 61, 429), (2) Jeffreys-Engdahl 1968 (Ph.D. Thesis Engdahl, Saint Louis University), (3) Gilbert et al 1973 (Proc. Nat. Acad. Sci., 70, 1410). In the lower 750 km of the inner core a Q value 1000 has been suggested.

The low Q values of the models in Figure A.4 have several implications. First, it is suggested that the temperature at the inner core boundary is close to the temperature of inner core material (presumably iron) and that partial melting occurs in at least part of the inner core. Second, the amplitudes of inner core phases, in particular phases like PKSSKP and PKSKP, are much reduced. Both computational and observational results predict that PKSSKP is observable only in exceptional cases and that PKSKP is too small to be observed.

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