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The constitution of the core: seismological evidence

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The present best estimates of seismic velocities in the core are compared with the 1939 solution of Jeffreys, with emphasis on the remaining uncertainties and present resolution capability. The relative contributions of measurements of seismic body waves and terrestrial eigenspectra to the inverse problem of the determination of elastic parameters, density and damping in the core are compared. Linear perturbation algorithms and smoothing functions used with the spectral data reduce their capacity for fine structural definition.

Between radii of 1400 and 3300 km (shell E), the outer core appears to be substantially homogeneous and non-stratified, with small or zero rigidity and a dimensionless seismic quality factor, Q, of order 10^4. It is a sufficient but still not a necessary condition that density ρ follows precisely the Adams–Williamson equation in E; for an averaging interval of 400 km, estimates of ρ have a standard error there of about $0.2\ \mathrm{g\ cm^{-3}}$. There is as yet no unequivocal seismological evidence for or against a boundary shell (thickness less than about 200 km) at the top of the liquid outer core. At the bottom of the outer core, the evidence is becoming stronger that any reduction in the rate of increase with depth of P wave velocity α is confined to a minor transition layer little more than 100 km thick.

The inner core has a sharp outer boundary at about 1216 km radius, but below it only average physical properties are estimated with any confidence. The average seismic compressional and shear velocities are about $\alpha = 11.2$ and $\beta = 3.5\ \mathrm{km\ s^{-1}}$ and $12.5 < \rho < 13.6\ \mathrm{g\ cm^{-3}}$, yielding a peculiar mean Poisson ratio of 0.44 or greater. At the inner core boundary, jumps in parameters are: $\Delta\alpha \approx 0.65$, $\Delta\beta = 2.0\text{–}3.0\ \mathrm{km\ s^{-1}}$ and $\Delta\rho \approx 1.0\ \mathrm{g\ cm^{-3}}$. Recent travel-time and waveform synthetics suggest a strong increase of P (and perhaps S) velocity in the upper 300 km of the inner core, which could be interpreted as a mixing or melting effect. Damping properties in the inner core may have an unusual dependence on wave frequency with an order of magnitude increase in Q from 1 Hz to 4 mHz vibrations.

1. Introduction

The discovery of the Earth's core is credited to Oldham (1906), the inner core (shell G) to Lehmann (1936), the fluidity of the outer core (shell E) to Jeffreys (1926), and the solidity of the inner core to Bullen (1946). The extraction from observed seismograms of seismic wave velocities, density, elastic and inelastic parameters in the Earth's core is a mathematical inverse problem subject to statistical variability, incomplete resolution, and lack of uniqueness.

Early quantitative work used the observed travel times of short-period (approximately 1 Hz) PKP and SKS waves from earthquakes that had penetrated the core. By the use of such data in the Herglotz–Wiechert integral equation, Jeffreys (1939) produced, with certain assumptions, a P wave velocity distribution that has long served as a reference standard. In the ensuing years, many seismologists attempted to refine this solution. Additional core wave travel-time data have been augmented with broad frequency-band recordings and terrestrial eigenspectra. As a result, estimates of core parameters derived from the Jeffreys velocity model have required only minor revisions. It has been demonstrated, however, that the 1939 P velocity inversion has an incorrect form just outside the inner core (shell F). Also, the preferred modern radii for the boundaries of the inner and outer core are about $R_{\text{i.c.b.}} = 1216 \pm 2$ km (Bolt 1977) and $R_{\text{m.c.b.}} = 3486$–3482 km (Bullen 1975). The Jeffreys values were $R_{\text{i.c.b.}} = 1250$–1243 km and $R_{\text{m.c.b.}} = 3473 \pm 3$ km.

In the remainder of this paper, outstanding problems on the fine structure of the core will be reviewed, assessments given of the uncertainties remaining in the core velocity and density distributions, and a brief critique offered on current controversies on core constitution. A comparison will be made between two average Earth models, called PREM and CAL8, that satisfy within narrow limits many of the recent core wave and eigen data.

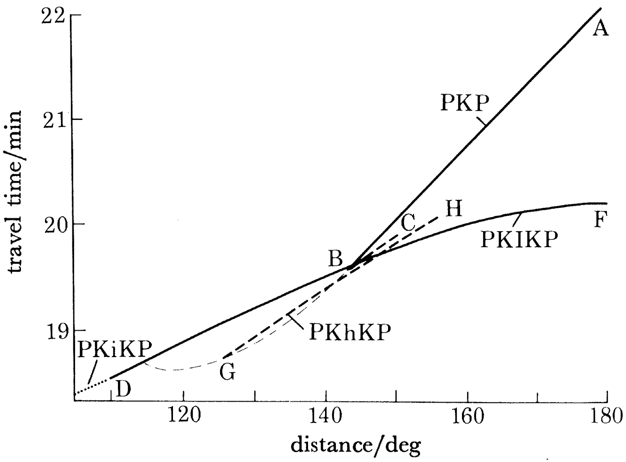

FIGURE 1. The PKP travel-time curve (Jeffreys & Bullen 1940), with branches AB and BC (outer core transit only), and DF (PKIKP rays). Waves reflected from the inner core PKiKP correspond to the receding branch CD (or HD), not drawn here, with a continuation back to $\Delta = 0$ beyond the cusp D. The light broken (parabolic) line before B is the locus of theoretical minimum times of arrival of PKP rays scattered near the mantle–core boundary (Cleary & Haddon 1972). The heavier broken line GH gives the theoretical times of arrival of PKhKP waves reflected from a discontinuity near $R = 1806$ km (Bolt 1968).

2. The P velocity distribution

(a) *Travel times of core waves*

Many recent studies use classical ray-theory procedures with travel times from short-period core phases (Bolt 1968; Cleary & Hales 1971; Randall 1970; Anderssen & Cleary 1980). Inversion from times to velocities is usually by the Herglotz–Wiechert integral or linear perturbation theory. As well as the main phases PKP and SKS, direct measurements for core phases not available to Jeffreys have now been used, such as SKKS, PKhKP (precursors to PKIKP for $120° < \Delta < 145°$), PKiKP (direct reflexions from the inner core boundary, i.c.b.) and PmKP, $m = 2, \ldots, 13$ (multiple reflexions inside m.c.b.).

For reference, the basic form of the PKP travel-time curve is drawn in figure 1. The travel times from the DF branch (PKIKP) are now known within 1 s and for AB within 2 s (Bolt 1968; Qamar 1973; Anderssen & Cleary 1980), but the crucial location of the cusps B, D and C are still not very exact. For the caustic B, Jeffreys (1939) gives $\Delta_{\text{B}} = 143.0°$; Qamar has $\Delta_{\text{B}} = 144°$. For D, Δ_{B} ranges from 110° (Jeffreys) to 120° (Qamar) to 121° (Choy & Cormier

1982). Selection of the travel-time samples in some studies is open to criticism; the use only of data identified by the I.S.C. by an arbitrary association rule may lead to circularity and bias.

The cusp C is, like B, in a portion of the time–distance diagram where branches overlap (see §5*b*). Diffraction (head waves), not accounted for by ray theory, also sets in near these cusps.

The precursors to PKIKP have received much attention. Two explanatory hypotheses, not necessarily exclusive, have been advanced. First, in 1962, Bolt suggested that they were waves called PKhKP reflected from a discontinuity near the base of the outer core. Travel times of the first scattered arrivals could be explained by an additional branch GH (figure 1). Qamar's analysis showed a very much smaller velocity jump (less than 0.02 km s^{-1}) would suffice. His velocity model KOR5 had a small velocity gradient at the bottom of the outer core and a large velocity gradient at the top of the inner core (shell G).

The alternative hypothesis (Cleary & Haddon 1972; Haddon & Cleary 1974) proposed that the PKP precursors were due to scattering from irregularities near the m.c.b. Travel times could not discriminate but wave slowness ($dT/d\Delta$) measured across arrays could. Most such slowness measurements are clearly in agreement with the m.c.b. scattering hypothesis (Doornbos & Husebye 1972; Doornbos 1974; King *et al.* 1974; Husebye *et al.* 1976; Chang & Cleary 1978). The possibility of ambiguities and large biases from wave interference and near array structure has been noted, however, and some experimenters (e.g. Wright 1975; van den Berg *et al.* 1978) find some $dT/d\Delta$ values consistent with reflecting interfaces near the base of the outer core.

From travel-time inversions, the P velocities in the outer core, averaged over 50 km or so, must be considered to be now resolved for $1400 < r < 3300$ km to within 0.03 km s^{-1} (see figure 3). Although further refinements are marginally possible, inherent sources of scatter in travel-time and slowness studies (such as source location error and structural variations) preclude much greater precision in core velocity estimates by using the Herglotz–Wiechert method. Further application is likely to use differential travel times (e.g. PKP(AB) – PKIKP) from individual sources and seismograms (see figure 2) as powerful constraints to models.

(*b*) *Wave amplitudes and synthetic seismograms*

Because controversies on fine core structure entail travel-time differences of less than 2 s, other tools with more resolving power must be used. Measured amplitudes of core waves can test models (see, for example, Qamar 1973; Müller 1973). For example, Bullen & Haddon (1973) point out that the observed small amplitudes of the earlier PKhKP precursors are consistent with the superposition of waves scattered from neighbouring points on the m.c.b.

Theory now permits the computation of synthetic seismograms of ground displacement in the 1 Hz frequency range. Account may be taken of plausible source models and diffraction effects along the path of the propagating body waves. At least for simple sources and seismograms from calibrated digital instruments, it is then possible to make direct overlays between the observed and predicted pulse pattern for models of interest. Although no inversion procedures are at present available, trial and error permits a rejection of competing models and a narrowing of possibilities.

The new method has been applied to only a few core studies (Cormier & Richards 1977; Kind & Müller 1977; Choy 1977; Rial & Cormier 1980; Choy *et al.* 1980). Certain weaknesses of the method should be mentioned. The effect of damping on the pulse shapes can be significant and there is often a trade-off between Q values and gradients in the elastic parameters. The effect may be important when the method is applied to achieve superior resolution in transition zones, where the damping may also be anomalous. Also, the method so far invites only visual comparisons and lacks a statistical basis for discrimination decisions.

A recent important core structure study of this type is by Choy & Cormier (1982) who compute theoretical waveforms appropriate for a deep-focus earthquake in Brazil. Comparison was made with eight PKP waves digitally recorded at standard stations over a distance range of 127.2–165.7°. A representative comparison from the work is given in figure 2 and inferences will be discussed in §5*b*.

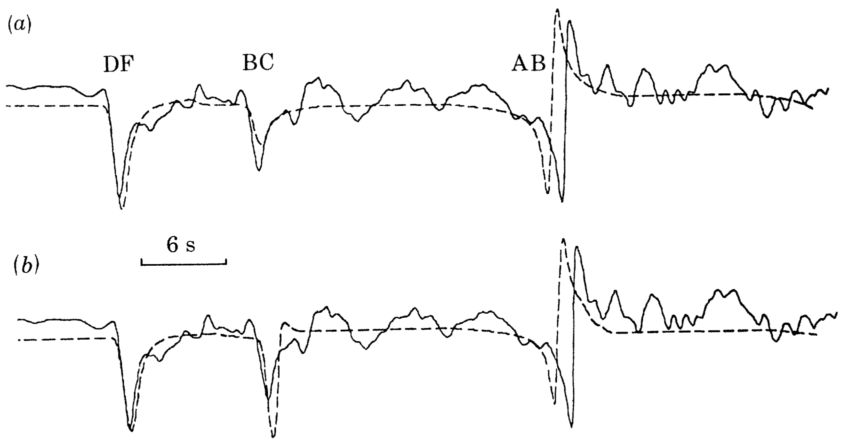

FIGURE 2. A comparison of synthetic wave forms (broken lines) and observed ground displacements (solid lines) of core waves from the deep focus earthquake in Brazil on 11 July 1978 at SHIO (Δ = 156.3°). In (*b*), the comparison velocity model is KOR5 (or CAL8) with a weak P velocity gradient in shell F. In (*a*), the comparison is with PREM with a higher velocity gradient at the bottom of the outer core (see figure 3). The pulse BC is significantly affected by structure near the i.c.b. (Courtesy G. L. Choy & V. F. Cormier 1982.)

3. The density distribution

(*a*) *Information from eigenvibrations*

The free oscillations of the Earth are described by ordinary differential equations with elastic parameters as coefficients. The problem is, given the measured eigenfrequencies of the system, to determine these coefficients. Mathematical formulations show that, without infinite spectra, unique solutions for the coefficients cannot be determined by inverse procedures (Sabatier 1978; Hald 1980). Nevertheless, from more than 1000 eigenfrequencies of normal modes that have already been measured, Earth models have been derived that predict these measurements.

In 1965, Jeffreys warned that eigenspectra for an inelastic Earth were different from those of the perfectly elastic one. The effect is geophysically significant, with eigenfrequency observations differing from the predicted elastic case by 8 s (0.25 %) for the fundamental spheroidal eigenfrequency ${}_0S_2$. Consequently, core models that are solutions to the eigenvibration inverse problem but that do not incorporate anelasticity are suspect and will be neglected here in making quantitative inferences. Two models that do allow for damping are PREM (Dziewonski & Anderson 1981) and CAL8. The latter is an appropriately modified version of CAL6 (Bolt &

Uhrhammer 1981). Because it was derived independently from similar basic seismic measurements but with different starting elastic distributions and mathematical procedures, CAL8 serves an important comparative purpose with PREM (see figure 3).

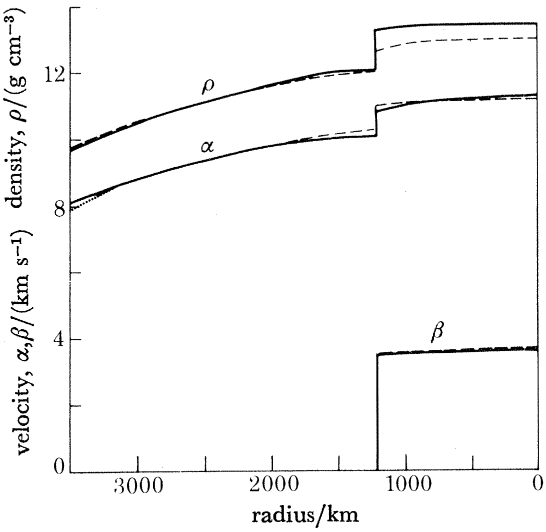

FIGURE 3. Comparisons between P and S velocity distributions (α and β) and density ρ in the core for the globally averaged Earth models PREM (dashed line) and CAL8 (full line). The dotted line is the Hales & Roberts (1971) solution.

(*b*) *Homogeneity*

Two quantities have been used to derive evidence on core constitution from seismic velocities. The first (Bullen 1949; Birch 1952) depends upon the gradient of the seismic parameter $\phi = \alpha^2 - \frac{4}{3}\beta^2 = k_s/\rho$, in the form

$$F(r) = 1 - g^{-1}\,\mathrm{d}\phi/\mathrm{d}r = \mathrm{d}k_s/\mathrm{d}p + \alpha\phi\tau/g, \tag{1}$$

where g is gravity, k_s incompressibility, α the expansion coefficient and τ the super adiabatic gradient. Because appropriate high-pressure laboratory results are few for the core, constitutional inferences from (1) depend upon plausible equations of state. In the lower mantle $F(4500) \approx 3.2$. In the outer core, $1400 < r < 3300$ km, $F(r) \approx 3.5$, averaged over 100 km, for both PREM and CAL8. Both values agree with Birch's theoretical prediction (1952) for homogeneous adiabatic conditions. Just outside and inside the inner core, resolution of $\mathrm{d}\phi/\mathrm{d}r$ is uncertain (see §§5*b* and 5*c*). If $\mathrm{d}\phi/\mathrm{d}r \approx 0$ in the former region, from (1), either $\mathrm{d}k_s/\mathrm{d}p$ must drop sharply or the temperature gradient is abnormal.

A more convenient index of homogeneity related directly to the Adams–Williamson equation (Bolt 1957) is

$$\eta = -(\phi/\rho g)\,\mathrm{d}\rho/\mathrm{d}r, \tag{2}$$

Further, when excess temperature gradients are ignored (Bullen 1963)

$$\eta = \mathrm{d}k_s/\mathrm{d}p + g^{-1}\,\mathrm{d}\phi/\mathrm{d}r. \tag{3}$$

From (2), if Adams–Williamson assumptions hold, $\eta = 1$; if $\eta > 1$, the density gradient exceeds the purely compressional effect or the temperature gradient is superadiabatic. Masters (1979) used η as a stratification index; for $\eta \approx 1$ physical stratification is unlikely.

Representative values of η are given in table 1 for PREM and CAL8. A curious feature of the PREM model is that $\eta = 1.00$ to the first decimal throughout the entire core. A strong constraint is implied because any linear perturbation of basic parameters in the inversion algorithm will produce fluctuations in η. For example, in (2) for PREM, ϕ and ρ are only slowly varying in G (see figure 3). Yet g decreases rapidly towards the centre as r so that η should also decrease as r approaches zero (as it does in the unconstrained CAL8). Thus in G, and probably F, $d\rho/dr$ in PREM is fixed by $\eta \approx 1$. Incidentally, a polynomial density function that satisfies this constraint in the core was derived by Bolt (1957). The important consequence is that PREM shows the *sufficiency* but not the *necessity* of the Adams–Williamson condition in the core.

By contrast, the CAL8 inversions that do not start with an Adams–Williamson core do not converge exactly to unity; η varies between 1 and 1.3 in E and, in G, η decreases like $1/r$ with depth (see table 1).

If α has a large positive gradient with depth (*ca.* 0.003 km s^{-1} per km) at the top of G and dk_s/dp has normal values (3.5–4) then (3) entails the unstable condition $\eta < 0$ unless $\beta > 0$ in G. Qamar (1973) showed by integration of (3) that his KOR5 velocity solution with $\eta \approx 1$ implies a rapid increase in β with depth to a value $\beta_0 + 0.50$ km s^{-1} at $r = 1000$ km, where β_0 is the poorly known shear velocity at the top of the inner core.

4. Anelastic properties

(a) *Short-period seismic core waves*

Amplitudes of core waves will depend upon geometrical spreading, energy partition at reflecting boundaries, and damping. Estimates of damping come from calculated ratios of amplitude spectra of core wave pulses recorded from the same event on the same seismogram.

For a damping factor exp $(-\pi f t/Q)$, where t is the travel time along the ray, average Q values may be determined from the spectral ratios. Let $R(f)$ be the spectral ratio of P7KP to P4KP (see figure 4). Then, to the first order,

$$\ln R(f) = a - \{(3\pi T/Q) + b\} f, \tag{4}$$

where T is the time for one core leg and a, b constant. Regression of (4) against log (spectral ratio) yields Q. Application of (4) has given $Q \approx 10000$ for P waves through the outer core (Sacks 1971; Doornbos 1974; Qamar & Eisenberg 1974; Cormier & Richards 1976). A similar procedure in the inner core gives Q values two orders of magnitude less. Buchbinder (1971) found $Q = 400$, Sacks (1971) inferred $Q = 170$ at the top of G, rising to 600 at the centre. Qamar & Eisenberg obtained $120 < Q < 400$; Bolt (1977) estimated $Q = 450 \pm 100$ from the spectral ratio for PKIIKP/PKiKP as recorded by L.A.S.A. If the doubly reflected phase is identified correctly, the last estimate should have the greatest resolution.

(b) *Damping of eigenvibrations*

The recorded eigenvibrations of the Earth, $S(t)$, can be represented (Bolt & Brillinger 1979) as a linear combination of exponentially decaying cosinusoids, superimposed upon noise with specified statistical properties so that

$$S(t) = \sum \alpha_k \exp(-\beta_k t) \cos(\gamma_k t + \delta_k). \tag{5}$$

The β_k determine the rates of decay of the oscillations and are usually given as $\beta_k = \gamma_k/2Q_k$.

Because each Q value is related to a particular wavelength or radial displacement in the Earth, the estimated values can themselves be inverted to determine corresponding damping properties for specified depths in the Earth. Those eigenvibrations with significant particle motions in the Earth's core thus provide a measure of damping within the core itself. The inferred Q value for the outer core appears very high, of order 10^4 (Dziewonski & Anderson 1981).

For the inner core, Masters & Gilbert (1981) argue, based on their identification of modes with shear energy in the inner core from an earthquake in Tonga in 1977 that $Q_{i.c.} \approx 3500$ at low frequencies (millihertz).

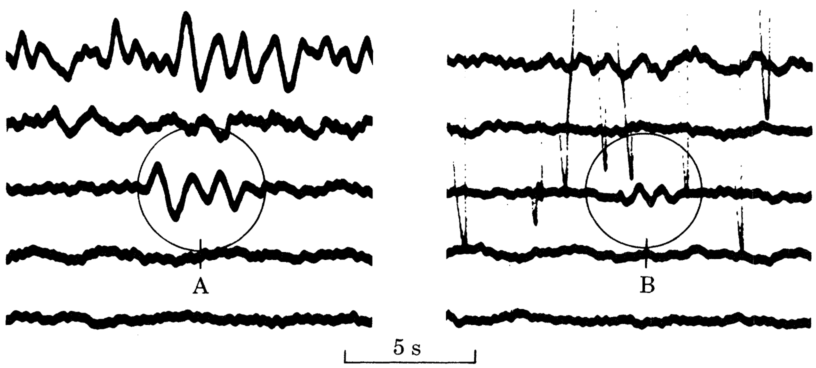

FIGURE 4. Previously unpublished seismograms from the Jamestown station in California, showing P4KP, labelled A, and P7KP, labelled B, phases from an underground nuclear explosion in Novaya Zemlya (Δ = 68.9°). The event occurred on 12 September 1973, and the records are from the vertical component short-period seismograph. Ground motion in P7KP is about 10^{-9} m.

5. DISCONTINUITIES AND STRATIFICATION

(a) E shell

Present seismic evidence does not rule out a transition shell E′ up to about 200 km thick at the top of the liquid core (Bullen 1975). The cubic splines used throughout E in PREM and CAL8 prevent the eigenspectra inversion from discriminating for or against E′. Resolution depends upon high-frequency body wave analysis (§2*a*). The Herglotz–Weichert inversion of Jeffreys (1939) used observed travel times of SKS and PKP. Because only the former bottom in a shell between r = 3480 and r = 3000 km, the velocity in this region is entirely dependent on the SKS waves. However, before 83°, SKS is obscured by the direct S waves on seismograms. For this reason, Jeffreys was forced to interpolate the K travel-time curve from 35° back to 0°, using a cubic spline. He obtained a velocity at the core of 8.10 km s^{-1}. With later SKS data, Randall (1970) obtained 8.26 km s^{-1}. The assumed extrapolation of K was reduced to only 14° by Hales & Roberts (1971) by using differences in arrival time of SKKS and SKS. They found α = 7.91 km s^{-1} at the core. Caution must be used with SKKS because of phase shifts relative to SKS (Choy 1977). It is of interest that the smoothed model PREM has a value α = 8.06 km s^{-1}, close to the Jeffreys value, even though the model is reported to be based on the Hales & Roberts's velocities (Dziewonski & Anderson 1981). There is no definitive work precluding $7.8 < \alpha < 8.2$ km s^{-1}, or 5 % variation at the top of the outer core.

Recently, Murtha (1981) has investigated the implication of a plausible E′ shell on SKS core waves. Reflexion and refraction at the m.c.b. produces a triplication in the S–ScS–SKS travel-time curve. The direct S branch precedes the SKS branch for $\Delta < 83°$ while the cusp A between the receding ScS branch and SKS has been put at 62° (Jeffreys & Bullen 1940). She observes SKS onsets back to at least 69° and these tentatively suggest $7.9 < \alpha < 8.1$ km s^{-1}.

Below the speculative boundary layer E′ there is little evidence for first or second order discontinuities in the monotonic velocity distribution. A jump $\Delta\alpha = 0.2$ km s^{-1} near $r = 2620$ km has been inferred by Kind & Müller (1977) but not yet independently checked, and there are theoretical reasons against the inference (Bolt & Uhrhammer 1981).

Trade-off curves between the width of the averaging kernels (spread) and standard error have been computed for density in the core (Bolt & Uhrhammer 1981). Eigenperiods of 39 core modes were used. At radii of 1000 and 2000 km, the standard errors were about 0.2 g cm^{-3} for a spread of 400 km. This resolution is lower than expected.

(*b*) *F shell*

The peculiarity of Jeffreys's decrease in velocity just outside the inner core was shown by Bolt (1968) to be unnecessary on the PKP data. It was replaced by a constant velocity shell about 450 km wide with $\Delta\alpha \approx 0.4$ km s^{-1} at its upper boundary and, like the Jeffreys solution, a sharp i.c.b. At first, numerous independent studies (see Bullen 1975) tended to confirm the existence of an F shell with smaller $\Delta\alpha$ and a weak velocity increase with depth. Even with the discontinuities removed (see §2*a*), shell F persisted because of the change in velocity gradient from Herglotz–Wiechert inversions. The solution KOR5 has α increasing by only about 0.15 km s^{-1} from $r = 1800$ to $r = 1250$ km compared with 0.34 km s^{-1} for PREM.

For a small velocity gradient, the second term in (3) is negligible so that, assuming normal compression, the value of η is 3 to 4, entailing moderate changes in constitution in F. (The Jeffreys velocity decrease gave the unlikely value $\eta \approx 30$ (Bullen 1975).)

There is now some independent evidence from wave shapes of core wave seismograms that $\mathrm{d}\alpha/\mathrm{d}r$ remains almost constant with depth, at least up to about 100 km from the i.c.b. Choy & Cormier (1982) compared synthetic with broadband seismograms of PKP (see figure 2). Comparison with the use of a range of core velocity models such as PREM and KOR5 leads these authors to conclude that the agreement of the synthetic and observed waveforms requires that 'there is no zone of small or negative P velocity gradient at the bottom of the outer core', but their results do imply strong gradients in the P and S velocities in the upper 200–300 km of the inner core (see §5*c*). Such a definite conclusion is perhaps premature because only one earthquake was used and the effect of anomalous conditions near this source and along these paths must be checked against other cases.

(*c*) *G shell*

The Jeffreys model (1939) had a sharp boundary at the top of the inner core, while Gutenberg (1957) considered a more gradual boundary. Subsequently, observations of the high-frequency reflected wave PKiKP (Bolt & Qamar 1970; Engdahl *et al.* 1970) demonstrated clearly that the physical properties must change abruptly within 1 or 2 km. Among other less direct evidence, observed travel-time differences PKiKP – PcP and PKIIKP – PKiKP at short distances are compatible with $r_{\mathrm{i.c.b.}} = 1216 \pm 2$ km and an average P velocity in G of 11.14 ± 0.02 km s^{-1} (Bolt 1977).

Evidence for significant rigidity at the inner core is fourfold. First, travel times of PKP and amplitudes of PKiKP at nearly normal incidence fix closely the P velocity and density on each side of the i.c.b. On Bullen's argument (1946), for the jump in compressibility, $k_s = \rho\,(\alpha^2 - \frac{4}{3}\beta^2)$, to remain plausible (less than 10 %, say), β must jump from 0 to about 3.1 km s^{-1} (Bolt 1972).

Secondly, one observation of the seismic body wave phase PKJKP with a shear wave leg in the core has been reported by using an array (Julian *et al.* 1972), but subsequently not confirmed; the average velocity was $\beta = 2.95 \pm 0.1$ km s^{-1}. Thirdly, observed eigenvibrations with significant particle motions in the inner core provide information. For example, core modes $_6S_2$ and $_7S_3$ indicate $\bar{\beta} = 3.52 \pm 0.03$ km s^{-1}, for $r_{\text{i.c.b.}} = 1216$ km (Masters & Gilbert 1981). Fourthly, very smooth solutions for PREM and CAL8, with a data set incorporating a number of core-sensitive modes, have $3.50 < \beta < 3.67$ and $3.49 < \beta < 3.60$ km s^{-1}, respectively. Pending a full analysis of resolution and uniqueness, an average rigidity $\mu \approx 170$ GPa is about all that can be adopted. Suggestions that there is a relatively rapid increase in β with depth at the top of G are referred to in §§3*b* and 5*b* (see also Häge 1981).

TABLE 1. AVERAGE CORE PARAMETERS: (i) PREM (ii) CAL8

location	α/(km s^{-1}) (i)	α/(km s^{-1}) (ii)	β/(km s^{-1}) (i)	β/(km s^{-1}) (ii)	ρ/(g cm^{-3}) (i)	ρ/(g cm^{-3}) (ii)	η (i)	η (ii)	Q (1 Hz)
top of E	8.06	8.09	0	0	9.90	9.82	1	1.1	—
middle of E (r = 2500 km)	9.34	9.37	0	0	11.19	11.20	1	1	*ca.* 10000
bottom of F	10.36	10.19	0	0	12.17	12.17	1	0.2	—
top of G	11.03	10.89	3.50†	3.49†	12.76	13.34	1	1.3	—
middle of G (r = 500 km)	11.22	11.29	3.64	3.60	13.03	13.58	1	0.3	450

† $\beta \approx 3.0$ km s^{-1} is perhaps more likely.

A jump $\Delta\alpha \approx 0.65$ km s^{-1} at the i.c.b. has been obtained in several independent studies (Bolt 1972; Müller 1973; Häge 1981), but the higher figure of Cormier & Richards (1977) must be discounted.

Reflections of PKiKP and (possibly) PKIIKP, at steep incident angles on the i.c.b. (see Bolt 1977) require jumps in *both* ρ and k at the i.c.b. because a sudden increase in rigidity alone cannot plausibly produce the observed amplitudes. The lower bound is thus $\rho > 12.0$ g cm^{-3}. The upper bound and distributed density in G are estimated by two independent arguments. First, the assumption of density, P velocity and attenuation above the i.c.b. and the P velocity on each side and the amplitude of PKiKP gives the density ratio across the i.c.b. The method (Bolt 1972) gives, with limited data, $\rho_F/\rho_G = 0.87 \pm 0.04$. This entails $\rho = 14.0$ g cm^{-3} on the PREM and CAL8 models as a probable upper bound at the top of G. The second method involves the inverse density problem, but it is only weakly constrained, with trade-offs with the β distribution in G. Masters (1979) inferred the uncertainty $\Delta\rho = 0.87 \pm 0.32$ g cm^{-3} at the i.c.b. The two comparison solutions PREM and CAL8 give mean inner core densities of 13.0 and 13.5 g cm^{-3}, respectively. Comparisons at key points are given in table 1.

The present situation is that more statistically precise and mathematically rigorous evidence from seismology is needed because details of constitution such as the presence of 'mushy' transition shells have crucial implications for core dynamics (Fearn *et al.* 1981) and equations of state (Stiller *et al.* 1980). The condition of part of the inner core may be close to melting. Its average Poisson ratio is about 0.44 compared with 0.5 for a fluid and 0.35 in the solid mantle. There are also the recent indications that damping may vary by an order of magnitude in the inner core with the frequency of excitation (see §5*b* and Cormier (1981)).

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