
Australian Journal of Earth Sciences
An International Geoscience Journal of the Geological Society of Australia
Publication details, including instructions for authors and subscription information:
http://www.informaworld.com/smpp/title~content=t716100753

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H. TkalČić [a]; B. L. N. Kennett [a]
[a] Research School of Earth Sciences, Australian National University, ACT, Australia

Online Publication Date: 01 January 2008

To cite this Article: TkalČić, H. and Kennett, B. L. N. (2008) 'Core structure and heterogeneity: a seismological perspective ', Australian Journal of Earth Sciences, 55:4, 419 — 431

To link to this article: DOI: 10.1080/08120090801888578
URL: http://dx.doi.org/10.1080/08120090801888578

*Australian Journal of Earth Sciences* (2008) **55**, (419–431)

# Core structure and heterogeneity: a seismological perspective*

## H. TKALČIĆ† AND B. L. N. KENNETT

*Research School of Earth Sciences, Australian National University, ACT 0200, Australia.*

The aim of this paper is to give an overview of important historical results and review current knowledge of the Earth's core, as well as to discuss prospects for seismological studies of the core. Although the properties of the core of the Earth can only be determined indirectly, there has been considerable progress in elucidating its structure. The iron-rich core is dense but has lower P-wave speed than the mantle above; the solid inner core has fewer light constituents than the fluid outer core. The density contrast at the inner-core boundary is too large for just a phase transition. The fluid outer core is well stirred by the convective flows associated with the generation of the geodynamo and is expected to have a nearly adiabatic profile. Only inside the tangent cylinder defined by the presence of the inner core might there be some seismic heterogeneity in the bulk of the outer core. Some variability along the underside of the core–mantle boundary due to selective separation of lighter material is suggested by some observations. By comparison, the inner core is rather complex with heterogeneous and anisotropic structures that appear to have hemispherical differences. Significant attenuation occurs just below the inner-core boundary, probably due to a mushy zone associated with the growth of the inner core. A variety of seismic observations help to define inner-core structures, but it is important to take account of the influence of the complex structure at the base of the mantle. A slightly different zone has been suggested around the centre of the Earth, although it is difficult to get good control on this region.

**KEY WORDS: anisotropy, attenuation, Earth's core, geodynamic processes, heterogeneity.**

## INTRODUCTION

At the turn of the nineteenth century, Wiechert's view of the Earth's interior was that it could be subdivided into two shells—a metallic core, surrounded by a silicate shell (Wiechert 1897). Oldham (1906) discovered the existence of the Earth's core and made preliminary estimates of its size from the P waves that he interpreted to propagate more slowly in the core. He also noted a seismic 'shadow zone,' on the side of the Earth opposite the earthquake, where no P waves were recorded and attributed it to P wave refraction along the core–mantle boundary. Gutenberg (1914) estimated the depth of the core–mantle boundary with a value that is not very far from today's figure of about 2889 km (Kennett *et al.* 1995). Jeffreys (1926) observed an S wave shadow zone that begins at an epicentral distance of about 103° from an earthquake. This result indicated that the core was molten, since shear waves do not propagate through liquids. Lehmann (1936) observed P waves in the P shadow zone that she interpreted as refracted from the inner core, a region inside the outer core with different properties. Bullen (1946) suggested that the inner core is solid. In parallel with seismological studies, Birch (1940) suggested that the inner core was solidifying from the outer core and that the inner-core boundary was a phase transition. It was also recognised that the solidification results in latent heat release (Verhoogen 1961) and compositional buoyancy (Braginsky 1963), which drive convection in the outer core. A seismological study by Dziewonski & Gilbert (1971) confirmed the hypothesis that the inner core was solid, based on the observations of Earth's normal modes.

## IMPORTANCE OF CORE STUDIES AND THE ROLE OF SEISMOLOGY

It is clear from the introduction that seismology has played a crucial role in advancing our understanding of this most remote region of our planet. Other important disciplines involved in studies of the core include paleomagnetism, geodesy, isotope geochemistry, cosmochemistry, mineral physics, and experimental petrology. The understanding of the present inner and outer core composition and their differences is very important in the context of understanding core dynamics, because compositional buoyancy plays a crucial role in powering the geodynamo. Some of the important geochemical issues include chemical composition of the

*This is one of a series of invited review papers addressing aspects of the themes for the International Year of Planet Earth.

†Corresponding author: hrvoje.tkalcic@anu.edu.au

ISSN 0812-0099 print/ISSN 1440-0952 online © 2008 Geological Society of Australia
DOI: 10.1080/08120090801888578

core and timing of core formation. The chemical composition of the Earth is important for understanding the early evolution of the solar system (Drake & Righter 2002), because some 'missing' chemical elements are likely to have been incorporated into the core during the differentiation process.

The amount of seismological data is growing rapidly, and the focus in present-day studies of the Earth's core shifts back and forth from topics related to structure and attenuation, to anisotropy in seismic-wave speed and its spatial distribution. There is a war of arguments in the seismological papers dealing with the subject of the core. There are, in our opinion, four groups of topics on which seismological studies are of major interest, presently debated with equal vigour, and deserve more attention than the rest. These topics are: (i) the chemical composition of the inner and outer cores; (ii) the solidification, texture of the inner core and the density contrast at the inner-core boundary; (iii) the anisotropy of seismic-wave speed in the inner core; and (iv) the attenuation and attenuation anisotropy of the inner core. We also include a brief discussion on the differential rotation of the inner core with respect to the rest of the planet.

## SEISMOLOGY AND THE CHEMICAL COMPOSITION OF THE CORE

Seismology provides major constraints on the composition, physical and chemical properties of the core, as well as on its dynamics. The density as a function of Earth's radius can be estimated using seismic speeds and the Adams–Williamson equation (Williamson & Adams 1923) based on the assumption of adiabaticity due to rapid mixing. The uncertainties from the estimates of density translate directly to the uncertainties in the composition of the Earth. This is because seismic data cannot generally distinguish between two chemical elements of the same density. In order to deduce chemical composition, seismological information has to be combined with theoretical studies, extrapolation from meteorites and laboratory results at high pressures and temperatures. Information from meteorites presents an important constraint on the core composition, but because meteorites are not formed from planets whose cores were exposed to quite the same pressure range as the Earth's core, these results must be used with caution.

Besides the composition, the phase diagram of the core is not known. Thus, we do not know if the inner core is a solid solution or forms a eutectic. It is likely that the inner core is not simply a binary system (iron–iron oxide or iron–iron sulfide), but FeO and $FeS_2$ are some of the likely candidates to be present in the core with iron (Bergman 2003). Without knowing the phase diagram, it is difficult to understand the conditions under which anisotropy might form, and it is also impossible to know the precise temperature and thermal history of the inner core (for example, at which point in the Earth's history it was formed).

Nonetheless, it is well accepted in the geophysical community that the core is predominantly made of iron. It is estimated that the outer core is about 5–10% less dense than pure iron at core pressures (McDonough 2004). The inner core represents only about 5% of the core mass. The estimates for the inner-core density deficit (in comparison with pure iron at inner-core pressures) are somewhat lower (3–5%) than for the outer core. This not only argues for the existence of lighter elements in the core, but also supports the hypothesis that less dense material from iron-depleted alloy partitions in the outer core during inner-core solidification.

Although seismological data are insensitive to absolute chemical composition, they are more sensitive to relative densities, and physical properties in general (e.g. the size and orientation of grains, impedance contrast, anisotropy). As an illustration, there are constraints from seismological studies on the density ratio at the inner-core boundary, and although they vary, it is generally accepted that the outer core is several hundreds of kilograms per cubic metre less dense than the inner core. Possibly 200 kg/m$^3$ comes from the fact that the density of iron is lower in liquid phase, and the difference must come from a less dense alloy partitioning in the outer core. In other words, seismological results suggest that the density jump is bigger than it would be just from a phase transition. We give an overview of this topic in the next section.

Nickel is thought to be present (8%) in the core, but because its density is similar to that of iron, it is seismically undetectable and does not seem to play a major role in the core thermodynamics.

As mentioned above, oxides and sulfides are likely candidates for the lighter elements present in the core. Although iron oxide is not present in meteorites and is not soluble in iron at atmospheric conditions, this property of oxygen changes with increasing temperature and pressure. O'Neill *et al.* (1998) showed that oxygen solubility in iron increases with temperature and decreases with pressure, and that about 2% or less oxygen could have been dissolved into a core (although this is not enough alone to explain the density deficit). Helffrich & Kaneshima (2004) used multiple reflections of P waves (P4KP)* from the inner side of the core–mantle boundary, and from only four stacked records concluded that there was no evidence for a layered outer core (and therefore an oxygen-rich outer core), which would be expected to be present if certain combinations of light elements in the core were present. If there were stratification at the top of the outer core, there would be arrivals on seismograms preceding P4KP arrivals. However, Eaton & Kendall (2006) sustained the possibility that such a layer existed, based on the analysis of S4KS/S3KS amplitude ratios. Furthermore, Tanaka (2007) found evidence for a low-velocity layer in the outer core, with a P-wave speed of 7.95 km/s and thickness equal to 90 km. Obviously, more observations of multiple reflections from the inner side of the core–mantle boundary are needed to resolve this controversy. Nevertheless,

*For the nomenclature of seismic body waves, see Chapter 8 (Exploring inside the Earth) in Bolt 1988.

because oxygen is present in the mantle and because it is abundant in the universe, it is most likely present in the core as well. Badro *et al*. (2007) recently derived a compositional model of the core using information from X-ray scattering and an application of Birch's Law at high pressures. Their preferred model contains significant amounts of oxygen and silicon.

The estimated sulfur, carbon and phosphorus contents of the core are too small to account for the core density deficit (McDonough 2004). However, silicon is thought to be a likely light element in the core (Ringwood 1959). There are alternative ranges of both silicon models and oxygen models, but the models that would combine oxygen and silicon are more problematic because these two elements are mutually exclusive in metallic alloy liquids over a range of temperature and pressure conditions. Interestingly, Buffett *et al*. (2000) argued for the presence of sediments at the top of the liquid core, which would form as a result of extraction of FeSi and FeO from the liquid core to form a silicate perovskite. This extraction would create an additional source of buoyancy in the outer core and would enhance convection, as a dense iron-rich liquid would sink into the interior of the core. Seismologically, such a scenario is supported by the observation of heterogeneous structure at the core–mantle boundary. However, if ultra-low-velocity zones (Garnero & Helmberger 1996) are invoked as possible manifestations of this mechanism, they should be more readily observed on a global scale, especially if the sediments would accumulate in the valleys of the core–mantle boundary first.

It is speculative what other elements might be present in the core. Buffett (2002) calculated that some radioactivity might be required to satisfy the energy conditions for the geodynamo. Thus, the presence and the quantity of radioactive elements in the core have become the subject of recent studies. For example, Lee *et al*. (2004) found that although potassium could be an important addition to iron at high temperatures and pressures, it is unlikely that Fe–K alloy is present in the core, because the core differentiation started before the conditions were favourable for potassium to alloy with iron.

Seismology can provide indirect constraints on the chemical composition of the core from density, speed, and elastic parameters. There is a strong potential for imaging possible stratification below the core–mantle boundary, provided that such stratification would be resistant to vigorous convection in the outer core. Apart from the underside reflection P4KP mentioned above, there are other seismic phases such as SKKS, SKKKS, which are more regularly observed, and PKKP, P7KP (although they do not sample quite so shallow parts of the outer core as their S counterparts) or other more exotic multiple arrivals which are seldom observed (the higher the number of multiples, the closer to the core–mantle boundary the sampling gets). In one such recent study, Rost & Garnero (2004) found that the differential travel times of P2KP waves show significant scatter, and speculated that strong heterogeneity in the inner core and along mantle paths may contribute to such observations. There is also a strong potential for better understanding of physical properties of the major boundaries, which might in return say something about the chemical composition. One important test of the Buffett *et al*. (2000) model would be a better mapping of the core–mantle boundary topography, and correlation between deeper regions and the observed locations of ultra-low-velocity zones. It is not quite clear how seismology could advance our knowledge of which radioactive and other trace elements are present in the core, unless their abundance would somehow reflect on physical properties of material present in the core, and the dynamics of the core on relatively short time-scales.

## SEISMOLOGICAL STUDIES OF THE TEXTURE OF THE INNER CORE

Seismological results suggest that the inner core is more complex than might be expected from purely thermodynamical considerations. A schematic overview of some structural features of the inner core that will be discussed in the following sections is given in Figure 1. It is well accepted that the inner core solidifies from the outer core, but the process of solidification is still not entirely understood. It is crucial to understand this process, as it determines the texture of the inner core that will determine the nature of seismic-wave interactions. Most plausible results point to a dendritic or accumulating nature for the growth (Bergman 2003). Apart from the process of solidification, the inner core is most likely experiencing a post-solidification deformation due to an internal stress field (Jeanloz & Wenk 1988; Yoshida *et al*. 1996; Karato 1999; Buffett & Wenk 2001). It is not yet clear which of these two processes is likely to have more impact on possible crystal alignment. In terms of the grainsize, there are some arguments for a very small grainsize but also for a very large size growth (Bergman 2003). It has even been suggested that the entire inner core could be a single crystal (Stixrude & Cohen 1995) as a means of matching seismologically observed anisotropy. However, subsequent studies, including those by the same authors, found that anisotropy of hcp iron at core pressures must be much stronger (see below). Recently, it was proposed that the inner core has a centremost shell, called the innermost inner core (Ishii & Dziewonski 2002). The hypothesis that the top of the inner core is a mushy zone seems to be well established among seismologists, and the estimated grainsize on the order of 1–2 km is well accepted (Vidale & Earle 2000) though not strongly controlled. The rigidity of the inner core has been extremely difficult to prove, although there have been several observations interpreted as shear waves in the inner core (Julian *et al*. 1972; Okal & Cansi 1998; Deuss *et al*. 2000; Cao *et al*. 2005). In general, due to poor signal-to-noise ratios and inability to be observed more readily, these observations are still subjected to scepticism. In a recent paper, Andrews *et al*. (2006) argued that the inner core is strongly attenuating in the normal-mode frequency band in shear, and that there is no discrepancy between attenuation models between normal modes and body waves.

In the following, we will focus on recent inferences of small-scale heterogeneities and the density-contrast

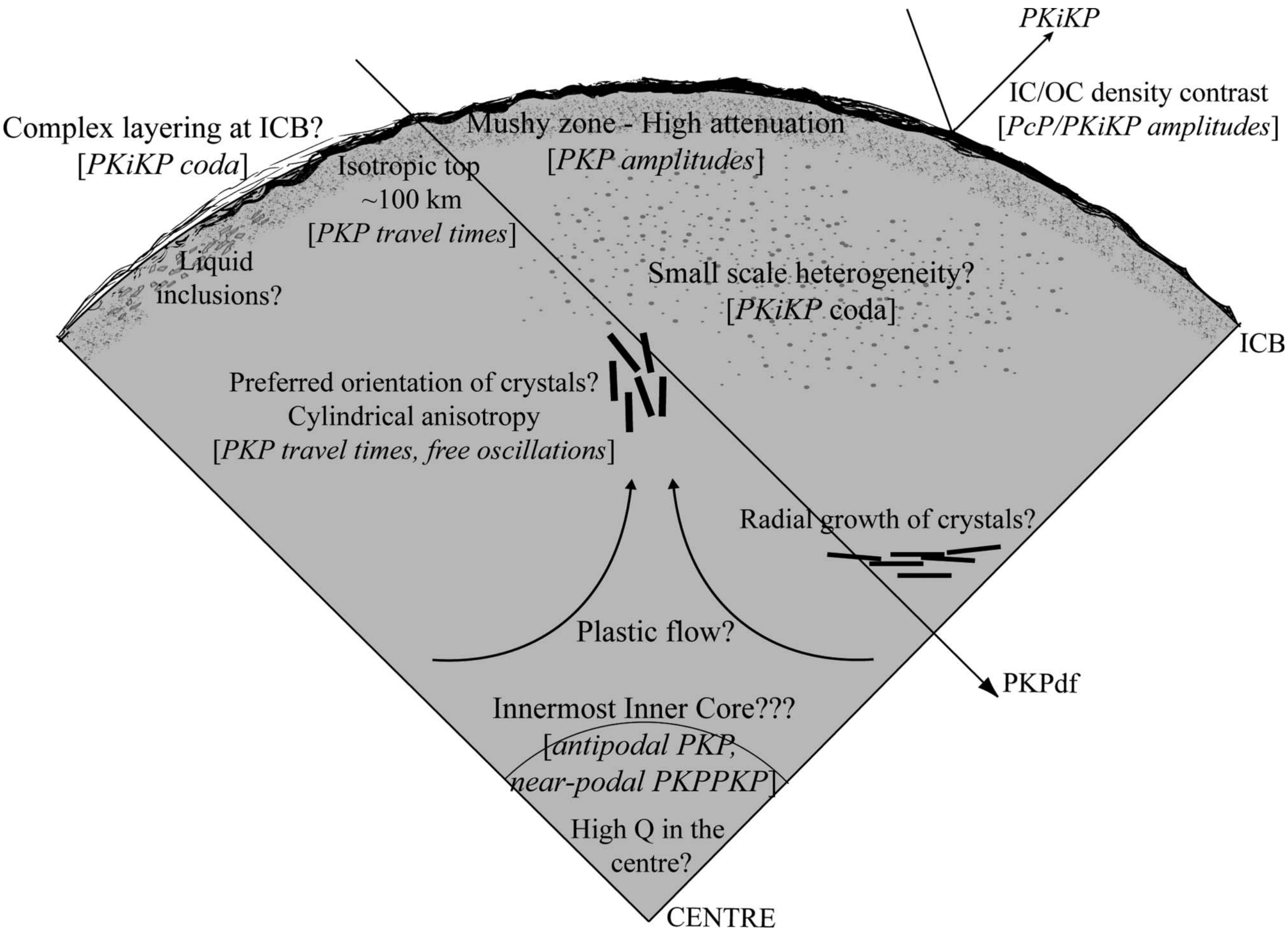


**Figure 1** Schematic representation of processes and structures in the Earth's inner core (ICB, inner-core boundary). The features are drawn at approximately true scale. The seismic phases associated with particular features are indicated in square brackets.

estimate at the inner-core boundary. Vidale & Earle (2000) interpreted their observations of long coda following the arrivals of PKiKP waves as inner-core scattering. They argued that the cause of scattering was a small-scale heterogeneity present at the top of the Earth's inner core. They modelled their observations by 1.2% variations in stiffness with a scale length of 2 km in the top 300 km of the inner core. Their study argues against the inner core having a single crystal structure, because of the proposed existence of heterogeneities in the bulk of the inner core. These heterogeneities can explain the observed attenuation of seismic waves by scattering. This view of the inner core also agrees better with a model of dendritic solidification and the consequent grainsize (Bergman 1998). One important aspect of the Vidale & Earle (2000) study is that if scattering is indeed present, it could explain the discrepancy between the normal mode estimates of Q (higher, e.g. PREM) and the results from body-waves analysis (Cormier *et al.* 1998). Normal modes predict much higher Q than body waves; one reason for this could be a scattering from small-scale heterogeneity to which normal modes are insensitive.

However, Poupinet & Kennett (2004) find very compelling evidence against the inner-core scattering suggested by Vidale & Earle (2000), based on their observations of PKiKP from Australian stations. They find no evidence for inner-core scattering, but do for inner-core-boundary scattering. The PKiKP wave coda at the WRA array is characterised by an envelope that is more constant and smaller in amplitude than the main PKiKP arrivals, rather than building up over time and reaching a maximum after the arrivals of PKiKP. This suggests scattering process confined near the inner-core boundary, and implies a more complex inner-core boundary than previously thought. Such an idea is compatible with a view that during the sedimentary compaction, a complex, crust-like region can develop at the surface of the inner core due to a low porosity at the top (Sumita *et al.* 1996). In addition, high-frequency PKiKP arrivals suggest that the transmission through core–mantle boundary is very efficient and that the inner-core boundary is a very effective reflector of PKiKP. The process of scattering must invoke some sort of a channelling of energy. Moreover, recent observations of PKPPKP waves at short distances (Tkalčić *et al.* 2006) further support the idea that the core–mantle boundary and inner-core boundary are efficient transmission zones for body waves.

Undoubtedly PKP waves are attenuated in the inner core, and this happens after they enter the inner-core boundary, even though they have a similar angle of incidence to PKiKP, which remains much less attenuated (see below). The mechanism of attenuation yet has to be understood.

The density contrast at the inner-core boundary is larger than it would be for just a phase transition. This inference follows from seismological results, which are manifested in 1D reference Earth models (Jeffreys 1926; Dziewonski & Anderson 1981; Kennett & Engdahl 1991; Kennett *et al.* 1995). Kennett (1998) concluded that the constraints imposed on the polynomials for the different depth intervals in available 1D density models are based on mathematical convenience rather than an attempt to allow for different physical processes. The polynomial nature of 1D density profiles imposes a very strong restriction on the nature of possible gradients within the core and the Earth in general. Physically, it is feasible that strong reflections of PKiKP could be observed from the inner-core boundary because the estimated thickness of the so-called 'mushy zone' at the top of the inner core ranges from only several hundred metres (Loper 1983), which is less than the wavelength of P waves in the inner core. The solid fraction rapidly grows with depth and increases one order of magnitude in only several hundred metres. However, there are also interpretations of the mushy zone extending tens of kilometres below the core–mantle boundary, if it is interpreted in terms of melt-fraction content (Cao & Romanowicz 2004b).

The history of the estimates of the density ratio at the inner-core boundary is somewhat symptomatic of the seismology of the core, and we can point to several hotly debated puzzles, whose decipherment follows the same trend. In this light, it is interesting to consider the history of PKiKP/PcP amplitude ratio observations and consequent publications. The first estimates from body waves by Bolt & Qamar (1970) yielded rather high values for the density contrast (1800 $kg/m^3$) in comparison with $\sim$600 $kg/m^3$ later estimated from normal modes (Dziewonski & Anderson 1981; Shearer & Masters 1990). The body-wave technique relies on plane-wave propagation theory and the boundary conditions associated with the solid–liquid interface. Bolt & Qamar (1970) calculated reflection and refraction coefficients from the boundary conditions and assuming some previous knowledge of elastic parameters as a function of depth from 1D models, deduced the density ratio. However, they relied on only one observation. As mentioned, more recent constraints on the density ratio at the inner-core boundary come from the observations of PKiKP and PcP by Shearer & Masters (1990), although they found only two clear observations. Masters & Gubbins (2003) then recalculated the density jump from normal mode data showing that the previous estimate based on normal modes was too low and that it could be raised to about 820 $km/m^3$. Most recently, Cao & Romanowicz (2004a) investigated body waves and, using five observations, found a density contrast of 850 $kg/m^3$. Thus, the results from two independent datasets converged to the same value resulting in a reconciliation of an old discrepancy.

Seismological papers so far have suggested that the inferences from the observations of PKiKP at 10–70° most likely represents extreme conditions (probably enhanced through focusing on mantle heterogeneities), so that the density contrast estimates would actually present an upper bound (Shearer and Masters 1990). One problematic assumption that is commonly made is that the PcP and PKiKP ray paths are very similar, so that attenuation in the outer core with Q of 10 000 or more would simply affect the ray path difference in the outer core. It is not at all clear how well the attenuation is known in the mantle, where these two paths differ, or even if attempts are made to estimate the effect. This means that the shorter the epicentral distance, the better the assumptions become because the ray paths in the regions of common sampling (crust and upper mantle) are almost the same.

Another important factor is the radiation pattern from the seismic source. It is questionable how well the details of radiation patterns are known from the inversion of focal mechanism parameters. The global CMT (centroid moment tensor) moment tensor parameters are taken into account to correct for amplitude differences, but they are based on low-frequency observations. A relatively small error in the orientation of the radiation pattern will have strong impact on assumed radiated energy content. This might well be a major reason why such estimates differ among different groups of researchers. Therefore, it is crucial to observe PKiKP and PcP waves at short distances, because the amplitude difference between the radiated energies into PcP and PKiKP will approach zero as the epicentral distance approaches 0° (both PcP and PKiKP waves would leave the focal sphere at the same place). Such observations at short distances would obviously put more constraints on the estimate of the inner-core boundary density ratio, although they have been extremely difficult to find. Tkalčić & Kennett (2007) reported clear observations of PKiKP at very short epicentral distances ($<10°$). An example is shown in Figure 2. These observations are encouraging, because they open new prospects in putting additional constraints on the boundaries within the Earth.

Undoubtedly, the constraints from seismology are the key constraints on the structure of the Earth's core. They come from both body waves and normal modes. Unfortunately, normal mode sensitivity is an integral over a depth range, and therefore such data can only constrain average structure. Seismic body waves at higher frequency present a more versatile tool to probe the finer details of core structure. More observations of PKiKP waves are definitely needed to provide critical insights into the various controversies. We believe that gradual resolution of accumulated inconsistencies in our interpretation of the inner core processes will be possible, especially in the light of the deployments of several small arrays similar to WRA (the Warramunga Seismic and Infrasound Research Station in the Northern Territory) as part of the International Monitoring System of the Comprehensive Nuclear-Test Ban Treaty. Not all such data are yet readily available, and it is to be hoped that present access restrictions will be relaxed.

With the current expansion of seismic deployments on a global scale and improvements in data quality, we can look forward to further progress in our understanding of inner- and outer-core structures.

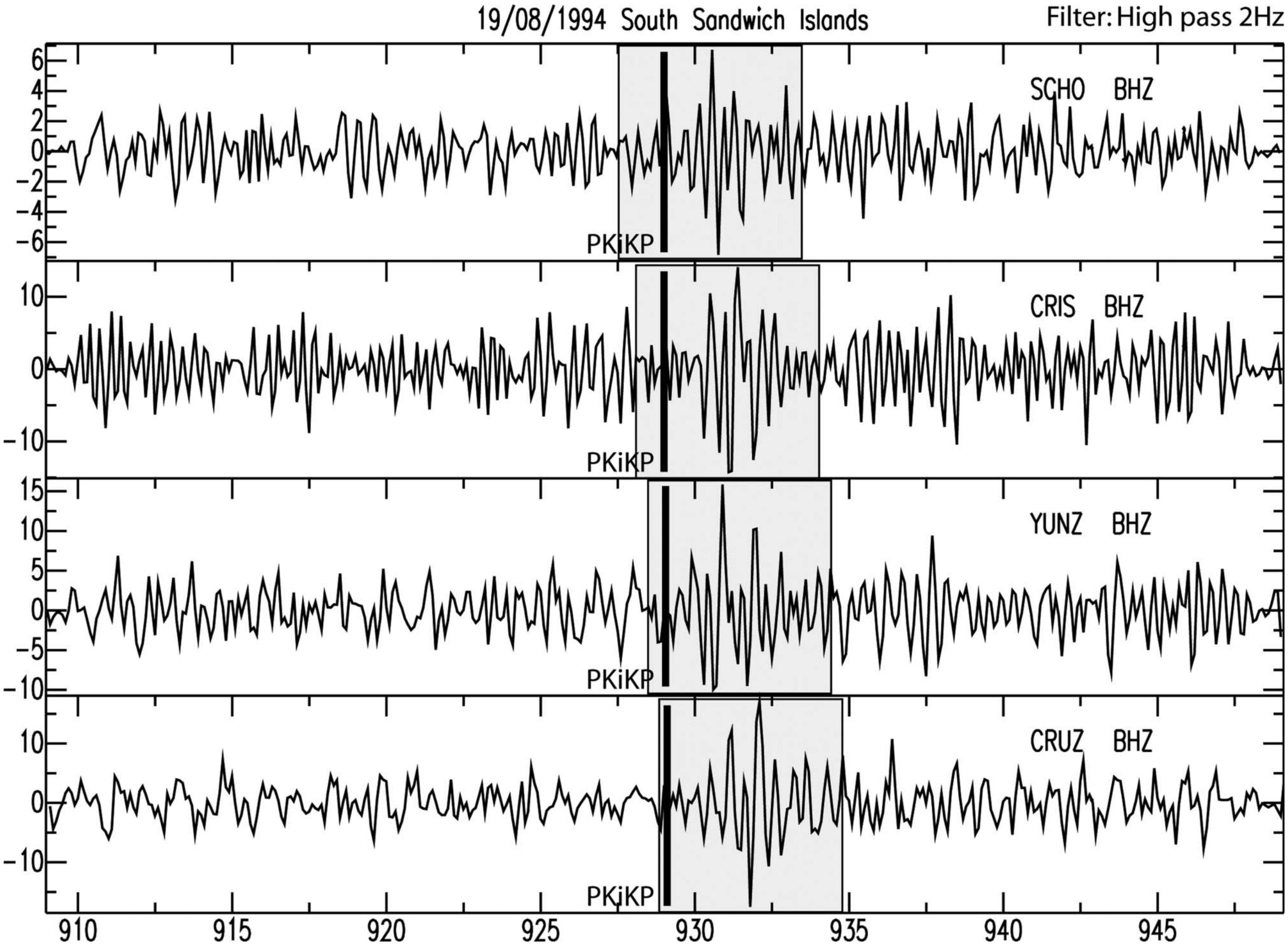


**Figure 2** PKiKP waves observed at very short epicentral distances ($\sim 7^\circ$) for an earthquake from the South Sandwich Islands region. The time windows focusing on PKiKP arrivals are highlighted in light grey. Vertical bars are predicted PKiKP travel times from the radially symmetric model ak135.

## SEISMICALLY INFERRED ANISOTROPY IN P-WAVE SPEEDS IN THE INNER CORE

Seismic anisotropy is a physical property of a medium in which there are different values for the elastic properties when measured in different directions. It is thought that seismic anisotropy is present in the inner core, in a similar way to the Earth's crust and mantle. We will first give a short overview of anisotropy at inner-core conditions from non-seismological studies, and then discuss some seismological observations.

The hexagonal close packed phase of iron (hcp, also known as $\varepsilon$ phase) is stable at core conditions. There are two more phases of iron identified from diamond anvil cell experiments: the cubic close packed (fcc, also known as $\alpha$ phase) at high temperatures and the body centre cubic phase (bcc, also known as $\gamma$ phase) at lower temperatures and pressures. The bcc phase reappears in a narrow stability field just below melting. Although the hcp phase is favoured as the one present in the inner core, other observations such as shock-wave experiments suggest that other stable phases of iron might exist at inner core conditions (Steinle-Neumann *et al.* 2003).

The physical reason for the alignment of crystals in the inner core (which would by accumulation produce the anisotropic effects) is not well known. The hypotheses can be divided into those that invoke solidification, and those that involve post-solidification deformation and/or recrystallisation. The hypotheses that argue for solidification as a cause of anisotropy are: (i) anisotropic paramagnetic susceptibility (Karato 1993); (ii) the single crystal concept (Stixrude & Cohen 1995); and (iii) texturing due to directional solidification (Bergman 1997). The hypotheses that argue for post-solidification deformation as a cause of anisotropy are: (i) inner core thermal convection (Jeanloz & Wenk 1988); (ii) misalignment between the gravitational equipotential and the thermodynamical equilibrium figure of the inner-core field (Yoshida *et al.* 1996); (iii) radial flow due to Lorenz stresses (Karato 1999); and (iv) longitudinal flow due to Lorenz stresses (Buffett & Wenk 2001). It is beyond the scope of this paper to discuss the relative importance of each proposed mechanism. Each one of these hypotheses deserves equal attention and should be taken into account when interpreting seismological results.

Each one of the above hypotheses can be challenged. For example, an attractive mechanism for explanation of crystal alignment is a post-solidification process. If the inner core grows faster in the equatorial direction (because heat flow is the fastest in that direction) it means that it corrects its shape in a similar way to glacial isostatic rebound. This is a relatively quick process, and this leads to an imposed strain. However, the strength of this process is highly unlikely to be sufficient to produce the 3% anisotropy needed to fit some very anomalous travel-time data, and the time-scale of such a process would be very long, longer in fact than the age of Earth.

On the other hand, if inner-core anisotropy is exclusively a consequence of solidification, deeper parts of the inner core have had more time to solidify, and so anisotropy would be expected to be stronger. However, such an effect has not been observed seismologically. If deformation exists, and if anisotropy is entirely due to solidification, then post-solidification processes need to be weak so that they do not destroy the pre-existing fabric.

Laboratory experiments (Bergman 2003) support the hypothesis that the inner core is composed of columnar crystals with cylindrical, not spherical symmetry. The crystals grow as columns perpendicular to the rotation axis of Earth. If the c crystallographic axes correspond to fast axes, anisotropy would depend on how these axes are oriented in the process of solidification. For the equatorial paths of PKP waves, this would mean that there should be depth dependency of anisotropy (although it depends on how the c axes are oriented).

The results of experiments to determine anisotropy associated with the orientation of c axes are quite contradictory. Crystallographic c axes were shown to be faster than the basal plane (Stixrude & Cohen 1995; Bergman 1998), Yet, Mao *et al.* (1998) found that the fastest direction is 45° from the c axes, and Steinle-Neumann *et al.* (2001) found that the c axes are slower than the basal plane. Unfortunately, this ambiguity of mineral physics results makes seismological interpretations very challenging. However, seismological observations using travel times are of the utmost importance in studying the anisotropic properties of the inner core.

PKP (P′) waves are routinely used to study the lowermost mantle and the core, because the geometry of their ray-paths allows the probing of the deepest parts of the Earth. Poupinet *et al.* (1983) observed that the spherical symmetry of PKP travel times is perturbed when the travel times are analysed as a function of the angle between PKP ray-path in the inner core and Earth's rotation axis. A large number of travel times corresponding to the paths that sample the inner core in planes nearly parallel to the rotation axis arrive earlier than predicted by spherically symmetric Earth's reference models. This property was used as a basis for the hypothesis about the existence of a uniform anisotropy in the inner core (Figure 3a), proposed by Morelli *et al.* (1986). The first models of inner-core anisotropy were relatively simple and uniform throughout the radius of the inner core (Figure 3a). Vinnik *et al.* (1994) argued that the same 3% cylindrical anisotropy extends all the way down to the centre of Earth.

However, the finding of Tanaka & Hamaguchi (1997), that only one hemisphere of the inner core is anisotropic (Figure 3b), demonstrated efficiently that the inner core anisotropy is much more complex than previously assumed. Song & Helmberger (1998) analysed PKP waves sampling shallower portions of the inner core and found that they do not show a coherent change with respect to the angle of sampling of the inner core and therefore concluded that the top portion of the inner core is isotropic. In order to reconcile these two findings, Creager (2000) proposed an asymmetric model of inner core anisotropy, which was able to fit the PKP travel times (Figure 3c). Ishii & Dziewonski (2002) proposed the existence of a centremost shell in the inner core (Figure 3d), distinguished by its anisotropic properties from the rest of the inner core; they made this finding analysing a large, but less reliable dataset of PKP travel times from the ISC [International Seismological Centre (<http://www.isc.ac.uk>)] catalogue.

At the same time, a number of papers demonstrated that a significant percentage of the differential travel times of PKP phases could be explained by heterogeneous structure in the lowermost mantle (Bréger *et al.* 1999, 2000; Tkalčić *et al.* 2002) (Figure 4a). The assertion that PKP travel times are affected by mantle structure was based on the observation of coherent patterns of travel-time residuals in relation to geographical locations of earthquakes (Tkalčić 2001), as well as on

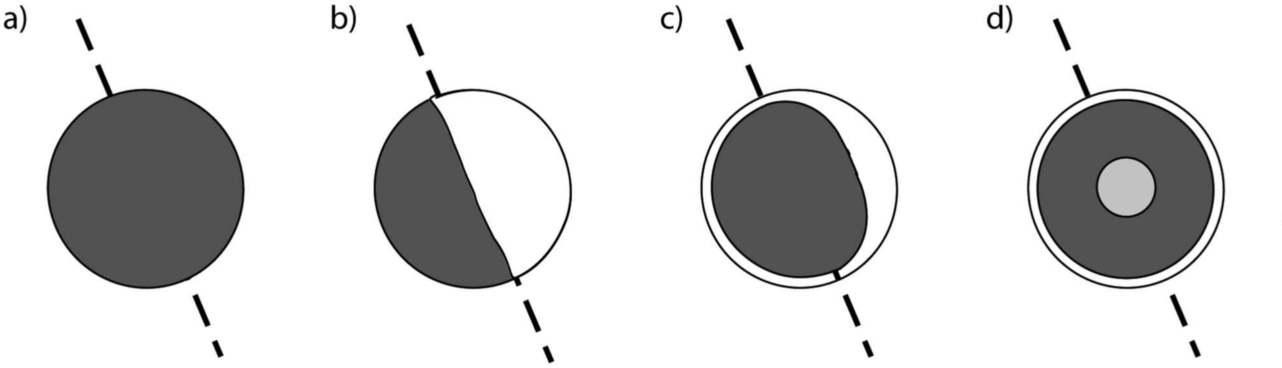


**Figure 3** Schematic representation of the models of anisotropy in the inner core. (a) Uniform inner-core anisotropy proposed by Morelli *et al.* (1986). (b) Hemispherical dependence of inner core anisotropy proposed by Tanaka & Hamaguchi (1997). (c) Combination of hemispherical dependence and isotropic upper part of the inner core proposed by Creager (2000). (d) Innermost inner core proposed by Ishii & Dziewonski (2002).

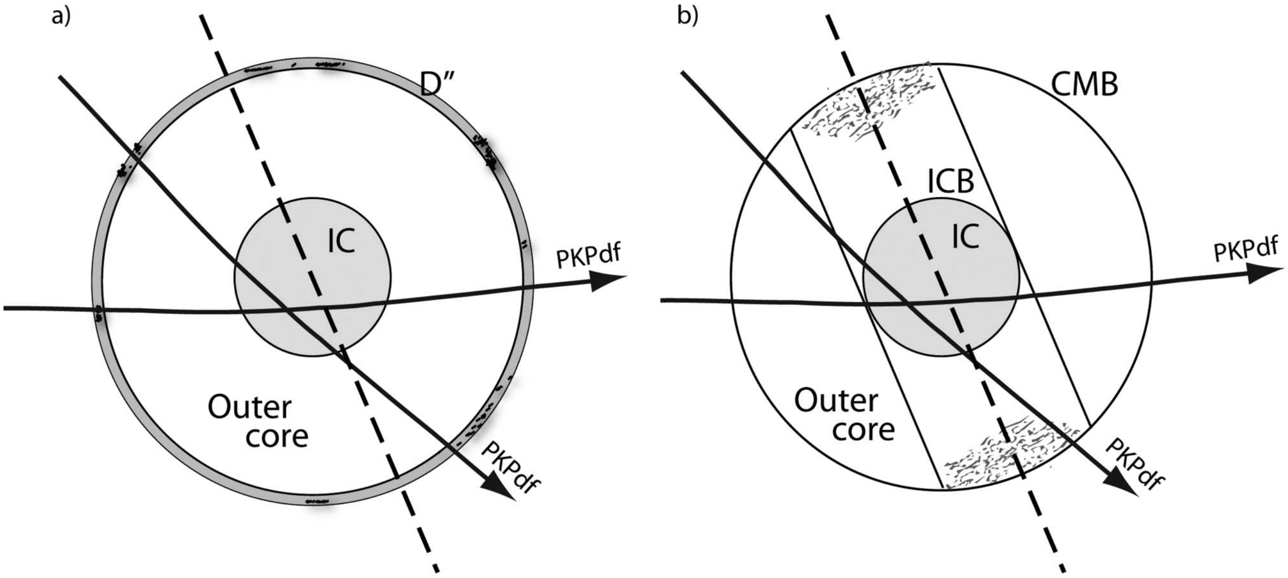


**Figure 4** Schematic representation of alternative models to inner-core anisotropy to explain travel-time variations of PKP waves: (a) various-scale heterogeneity in the lowermost mantle; and (b) small-scale heterogeneity in the outer core confined in the volume of the so called 'tangent cylinder'. CMB, core–mantle boundary; IC, inner core; ICB, inner-core boundary.

numerous observations that demonstrate the existence of different scale heterogeneities in the lowermost mantle. In particular, Tkalčić *et al.* (2002) assessed how much of the core-sensitive PKP travel-time data can be explained by mantle structure alone. Using the highest-quality PKP(AB-DF) and PcP-P differential travel-time data, they developed a 3D compressional velocity model of the lowermost mantle and outlined its relevance for understanding anisotropic structure of the Earth's core. A large amount of PKP(AB-DF) time data was adequately explained with mantle structure alone. However, in order to explain the most anomalous PKP(BC-DF) differential travel-time data observed (mostly originating from the South Sandwich Islands region earthquakes), alternative explanations are needed: (i) much smaller scale heterogeneity in the lowermost mantle than can be resolved with the current spatial sampling of the lowermost mantle [required by the fact that the ray-paths of PKP(BC) and PKP(DF) are much closer to each other in the mantle than the ray-paths of PKP(AB) and PKP(DF) phases]; or (ii) a combination of heterogeneity and anisotropy in the mantle and core.

The latter combination is an attractive and likely explanation for the observed anomalous travel times of PKP waves. Unfortunately, it might be very difficult to get a definitive answer to which of these two explanations is more likely because the coverage of the lowermost mantle by P waves is still relatively poor. Many authors wrongly assume that the mantle contribution to travel times is annulled by the use of differential travel-time analysis. This is not entirely true, especially at the frequencies at which PKP body waves are usually analysed (about 1 Hz). The proximity of the ray-paths for PKP(BC-DF) certainly gives a more reliable type of data than the corresponding pair for PKP(AB-DF), but the ray paths are still well separated in the lowermost mantle and core. Even when travel times are corrected for the mantle structure of current 3D models, because mantle structure is poorly known in the lowermost mantle, it is wise to consider the mantle as a significant source of noise and bias to core-sensitive seismic phases. In support of the idea that localised heterogeneity is a more likely cause of complex pattern of travel-time observations than complex anisotropy, Ishii *et al.* (2002) concluded that a simple model of anisotropy in the inner core combined with complexity in the mantle is a satisfactory model to explain absolute and differential PKP travel times. However, Calvet *et al.* (2006) clearly demonstrated the non-uniqueness of the seismological anisotropy models in the inner core. They showed that at least three different models of inner core anisotropy explain the same data, and each model has very different consequences for the origin of anisotropy in the inner core of the Earth.

One of the biggest problems with the interpretation of inner-core anisotropy is that the most strongly split normal modes are those with sampling in the shallowest parts of the inner core, where, contradictorily, no anisotropy is observed from body waves. As an alternative explanation to the inner-core anisotropy hypothesis, Romanowicz & Bréger (2000) showed that anomalous splitting of normal modes (except for the mode ${}_3S_2$) could be explained by structure in the outer core. Romanowicz *et al.* (2003) investigated this hypothesis, with a focus on explaining the anomalous travel times of PKP waves. They found that a 0.5–1.0% increase in P-wave speed inside the 'tangent cylinder' (a volume of the outer core with distinctly different convection characteristics from the rest of the outer core) could account for the geographical trends of absolute and differential PKP travel-time data (Figure 4b), in particular for an L-shaped pattern of travel-time residuals when plotted as a function of the angle with respect to the rotation axis of the Earth. While arguably unrealistic, these models deserve further investigation and should be compared with the models of inner-core anisotropy.

It is challenging to identify a plausible physical mechanism which would account for hemispherical (longitudinal) dependence of inner-core anisotropy. There are some suggestions that convection in the outer core could be controlled by the mantle (Bloxham & Gubbins 1987; Sumita & Olson 1999), which would then leave a signature on the texture of the inner core. However, this would mean that the inner core is locked to the mantle (with no differential rotation). Such differential rotation aside, the real problem with this hypothesis is that it has been suggested from seismological observations that the inner-core anisotropy hemispherical pattern exists to depths of 500 km, which would require structure in the mantle to persist for 500 million years (since there is convection in the mantle, it is hard to understand why there would be a stagnant zone at the base of the mantle for such a long time).

The contradictions in the interpretations of core-sensitive seismic data (such as anomalously advanced travel times and anomalously split normal modes) mean that it is important to sustain global observations of core-sensitive seismic body waves. One problem with the interpretation of inner-core anisotropy is the difficulty of achieving complete sampling of the inner core in all directions. Such complete sampling is geometrically impossible for the PKP at longer epicentral distance, due to the absence of large earthquakes at extreme latitudes. Almost perfect coverage is achieved for the ray-paths corresponding to PKP in their triplication range (145–155°). For this epicentral distance range, the smallest angles to the polar axis are between 10 and 15°, and there is a large number of data sampling the inner core with angles up to 90°. However, the radius of the bottoming points for such data is still relatively large, and therefore they do not carry information about the deeper inner core, from the region close to the centre of the Earth.

In order to probe Earth near its centre, PKP observations at epicentral distances of about 170° or larger are required. Unfortunately, it is difficult to achieve the same level of sampling with the current configuration of seismographic stations and uneven distribution of large earthquakes worldwide. High-latitude earthquakes and stations and corresponding source–receiver configurations are extremely rare. One group comes from a path between the South Sandwich Islands and stations in Alaska and the Northeast Asia, although this particular geometry does not produce antipodal paths. Travel-time residuals corresponding to these paths are very anomalous, and present the largest percentage of the travel-time data upon which inner-core anisotropy hypothesis is based.

Therefore, it is clear that a broader distribution of PKP data would be extremely useful to improve our current understanding of inner-core anisotropy and in particular its radial dependence inside the inner core. Data from recent temporary networks at extreme latitudes could be very useful (Reading 2006). However, even with new locations for seismic instrumentation, it is reasonable to assume that the source location of the largest earthquakes will not change. This situation presents a significant challenge for the seismology of the inner core.

In an attempt to address this problem, Tkalčić *et al.* (2006) have reported the first results of a systematic search for PKPPKP waves. The idea is to increase spatial sampling of the inner core by the use of ray-paths with different orientations, and introduce new geometries of sampling near Earth's centre. Such phases, due to their long paths through the Earth, are significantly attenuated and, because they arrive long after normal seismic phases, are rarely observed. Although only a few PKPPKP waves at short distances were observed, each new datum presents a valuable constraint on the inner-core structure and anisotropy. Peculiarly, PKPPKP data do not show anomalously delayed or advanced travel times with respect to standard reference Earth models.

Garcia *et al.* (2006) recently assembled a new dataset of PKP waves and travel times using a non-linear inversion method (Chevrot 2002), which enabled them to use shallow earthquakes. Shallow earthquakes were previously omitted from analysis because the complexity of waveforms increases with the interference of depth phases such as pPKP and sPKP. The innovation of this approach is in using synthetic waveforms that match observations instead of reading travel times directly from the seismograms. The method increased the spatial sampling of D″ and the inner core compared with the previous datasets. However, it did not introduce many polar paths with inner core sampling at long epicentral distances, merely because of the lack of respectable size earthquakes at high northern and southern latitudes. Undoubtedly, innovative techniques that have the capability of improving the signal-to-noise ratio will be a very important tool in further studies of the core and lowermost mantle structures, especially to enhance sampling.

## SEISMIC ATTENUATION IN THE INNER CORE

It is generally accepted that the inner core solidifies from the outer core. Some metallurgy experiments suggest that such growth is in a dendritic fashion. This would be compatible with the existence of a mushy zone with possible fluid inclusions at the top of the inner core, and might explain the observed seismic attenuation, at least in the very top part of the inner core.

The existence of a mushy zone at the top of the inner core is associated with seismic attenuation. Liquid pockets could be entrapped in the dendritic texture in the mushy zone (interdendritic fluid). Earlier seismological results suggest that Q changes from about 200 to about 1000 going from the top of the inner core to the centre (Doornbos 1974, 1983; Cormier 1981; Choy & Cormier 1983; Shearer & Masters 1990). However, such depth dependence is not well resolved (Bhattacharyya *et al.* 1993), and there is little depth resolution from the normal modes. There are differences in mineral physics results, some invoking partial melt and some not (Jackson *et al.* 2001). In addition, Jackson *et al.* (2001) found pronounced viscoelastic relaxation and marked anisotropy in the fcc phase of iron, thus making it an attractive alternative to the hcp phase, if the fcc phase were to be stabilised at inner-core conditions.

Cormier & Li (2002) confirmed the results of Vidale & Earle (2000), concluding that scattering attenuation is the dominant mechanism of attenuation in the inner core, based on waveform studies of a global PKP dataset. Interestingly, their observations for equatorial paths agree with the model in which c axes are the fast axes (Bergman 2003) in that there is a depth dependency of attenuation. This result is compatible with deeper penetrating rays crossing more crystal boundaries. However, the depth dependency that has been inferred for polar paths does not agree with Bergman's model, for which a depth dependency of attenuation should not be observed (because there is no scattering from the grain boundaries and regardless of the depth of sampling, there will be approximately the same number of crossed crystal boundaries). According to Li & Cormier (2002), there are two types of attenuation active in the inner core: attenuation combines both viscoelastic and scattering effects. Body-wave studies find much higher attenuation present in the inner core than predicted; this discrepancy could be due to either viscoelastic or scattering attenuation. Viscoelastic attenuation would require a large amount of liquid inclusions present in the inner core, but most likely a large amount of liquid is expelled buoyantly to the outer core. Therefore, scattering attenuation seems to be an attractive mechanism to explain the discrepancy between observed attenuation of body waves and that predicted from normal modes.

However, as discussed in the section on inner-core texture, Poupinet & Kennett (2004) found compelling evidence against inner-core scattering, suggesting instead scattering process confined near the inner-core boundary. This discrepancy of the results could be viewed in the light of recent findings of Leyton & Koper (2007). They studied the geographical distribution of PKiKP coda recorded at short-period, small-aperture seismic arrays and found a quasi-hemispherical variation of PKiKP coda. Although their dataset is very limited, they attributed this variation (the complete absence of PKiKP coda at some regions, as opposed to strong signals at other regions) to the variable volumetric scattering from variable solidification texturing of iron crystals in the inner core. They argued for a scattering attenuation in the inner core comparable in size to the intrinsic attenuation ($Q \sim 500$). Also recently, Cormier (2007) proposed a qualitative model for the texture of the uppermost inner core, supporting the idea that lateral variations in flow near the inner-core boundary could be recorded in the texture. According to this model, the quasi-eastern hemisphere undergoes a more active solidification and is more efficient in attenuating PKiKP forward-scattered waves due to the orientation of fabrics.

## ANISOTROPY IN SEISMIC ATTENUATION

Creager (1992) and Souriau & Romanowicz (1996, 1997) observed attenuation anisotropy with 10–30 km wavelength waves that propagate in the direction nearly parallel to the rotation axis of the Earth, exhibiting more complex waveforms and smaller amplitudes. A positive correlation was clearly observed between travel times and amplitudes (fast travel times corresponding to small amplitudes). This observation was opposite to that predicted by a model with small ellipsoidal liquid inclusions in the direction parallel to the rotation axis of the Earth. On the other hand, Bergman's model of a cylindrically radial direction of growth and consequent lattice preferred orientation can account for this observation with scattering of the walls of the grains (grain boundaries). If crystallographic a axes are perpendicular to the rotation axis, and c axes are randomly oriented in the plane transverse to the grain direction, there will be scattering due to impedance contrasts between grains for the waves traversing the inner core in the planes parallel to the rotation axis of the Earth. Cao & Romanowicz (2004b) find a hemispherical variation in the quality factor in the top part of the inner core and suggest that a higher-quality factor might be present in the quasi-western hemisphere due to a faster freezing rate and a higher porosity (better connected liquid inclusions). As discussed in the previous section, Cormier (2007) gives a different interpretation, in that he concludes that the quasi-eastern hemisphere undergoes a more active solidification, and therefore different orientations of inner core fabrics associated with the outer core flow give rise to different scattering and attenuation properties. Thus, there has been increasing recognition that some, if not most, of the attenuation in the inner core might be due to scattering rather than viscoelasticity.

If the observed attenuation is predominantly from scattering, then it should tell us about the grainsize in the inner core. The grainsize on the other hand, can tell us something about the deformation present in the inner core. The larger the stress field present in the inner core, the smaller the grainsize will be. Laboratory experiments (extrapolation) and meteorite samples suggest a grainsize of about 1 km. The grain width obtained from a 1/3 to 1/5 amplitude reduction in the seismic observations is on the order of few hundred metres to a kilometre. Vidale & Earle (2000) argued for a couple of kilometres grainsize from their scattering hypothesis. Stixrude & Cohen (1995) suggested that the whole inner core is a crystal. However, Steinle-Neumann *et al.* (2003) found that this would increase anisotropy to 10%, which is not observed.

Recent observations of high-frequency compressional waves from the inner core are encouraging because they may carry considerable information about the grainsize and attenuation in the core. The wavelengths of PKiKP waves at 2–3 Hz approach a few kilometres. It seems as though there is agreement that the small quality factor (high attenuation) observed at the top of the inner core is an indicator of a mushy zone. As the seismic waves penetrate deeper in the core, they reveal an increase in the quality factor, although it is not well known what that value might be in the Earth's centre. Obviously, more observations of high-quality core-sensitive phases, in particular those sampling near the Earth's centre, will bring improved resolution.

## DISCUSSION

Seismological studies provide critical insights for our understanding of inner-core structure and heterogeneity. With a dramatic increase in the number of modern broadband digital data, there has been a significant progress in recent decades on elucidating the internal workings of the Earth's core. However, we are still far from reaching definitive answers about the dynamic role of the core in the planet's evolution. Progress is partly inhibited by the lack of knowledge of the phase diagram of iron at core conditions, a difficulty driven by the uncertainties when dealing with high pressures and temperatures. There is also a lack of consensus in the mineral physics community on which mineralogical phase of iron prevails in the inner core and what would be the true orientation of the fast crystallographic axis. These uncertainties make seismological interpretations extremely challenging. Moreover, the seismological studies are also restricted by incomplete spatial sampling of the core by the seismic body waves, and more innovative signal processing and phase identification techniques are needed to mitigate this serious problem.

Apart from the inner-core anisotropy hypothesis that was discussed throughout the paper, another phenomenon—inner core differential rotation with respect to the rest of the planet—is an extremely hotly debated topic. Such rotation was predicted from the geodynamo models (Glatzmaier & Roberts 1996). Seismologists reported observations that agree well with the theoretical predictions (Richards *et al*. 1997). The biggest assumption made is that the fast anisotropy axis in the inner core is tilted with respect to the rotation axis of the Earth. With such a tilt and the differential rotation of the inner core with respect to the rest of the planet, the fast axis of anisotropy moves in time relative to any fixed path between a source–receiver pair. Over time there will be a changing angle between the path of the PKPdf waves within the inner core and the fast axis of anisotropy, and this is the proposed cause of the systematic variation in differential travel times of PKP waves. However, the effect has been demonstrated to be very small or even undetectable by the current resolution of seismic probes (Souriau 1998). Body waves and normal modes are much less sensitive to possible differential rotation than to the properties of the core discussed in previous sections, and so results are far more speculative than inferences about structure.

Current models for the core are more complex than the simple concept of radial stratification that prevailed until only a few years ago. The changes have been driven by improved seismological data, and we can anticipate further surprises as further information becomes available.

## ACKNOWLEDGEMENTS

We would like to acknowledge the contributions of many workers on Earth structure who have influenced this paper (even if they might feel that their views are misrepresented). We are grateful to two anonymous reviewers for their constructive comments.

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*Received 20 August 2007; accepted 21 December 2007*