thermal and flexural thickness. We believe that this improved parameterization has significant advantages over the elastic and uniform viscoelastic plates yet remains sufficiently simple in construction and number of parameters to be of use in tectonic models involving lithospheric flexure.

The universality of values for Q and η_b provides the major test of the model. That the preferred values of $Q \sim 170$–$250\ \mathrm{kJ\ mol^{-1}}$ are $\leq 350 \pm 50\ \mathrm{kJ\ mol^{-1}}$ (refs 18–20) observed in the laboratory for creep controlled by 'wet' olivine has also been found[20,21] and may be due to nonlinear creep; reflect the difference between polycrystalline and single crystal creep; be caused by impurities; or be a consequence of significant reheating, during the intrusion of seamounts, which resets the thermal age of the lithosphere. Furthermore, the results cannot be attributed to flow within the crust, for which low values of Q would be more appropriate, because the oceanic crust chills rapidly to temperatures at which no viscous creep occurs. It is also unlikely that significant brittle fracturing, which would reduce effective thickness, occurs outboard of seamount loads in contrast to subduction zones[21].

The Natural Sciences and Engineering Research Council provided financial assistance with this research through a strategic grant to C.B. and a graduate fellowship to R.C.C. We thank Dan McKenzie for useful comments on the manuscript.

Received 5 May; accepted 8 July 1983.

1. Oldenburg, D. W. *Geophys. J.R. astr. Soc.* **43**, 425–451 (1975).
2. Parsons, B. & Sclater, J. G. *J. geophys. Res.* **82**, 803–827 (1977).
3. Parsons, B. & McKenzie, D. P. *J. geophys. Res.* **83**, 4485–4496 (1978).
4. Sclater, J. G., Jaupart, C. & Galson, D. *Rev. Geophys. Space Phys.* **18**, 269–311 (1980).
5. Haxby, W. F. & Turcotte, D. L. *J. geophys. Res.* **83**, 5473–5478 (1978).
6. Forsyth, D. W. *Tectonophysics* **38**, 89–118 (1977).
7. Leeds, A. R. *Phys. Earth Planet Inter.* **11**, 61–64 (1975).
8. Bodine, J. H., Steckler, M. S. & Watts, A. B. *J. geophys. Res.* **86**, 3695–3707 (1981).
9. McNutt, M. K. & Parker, R. L. *Science* **199**, 773–775 (1978).
10. Ashby, M. F. & Verrall, R. A. *Phil. Trans. R. Soc.* **A288**, 59–95 (1978).
11. Goetze, C. & Evans, B. *Geophys. J.R. astr. Soc.* **59**, 463–478 (1979).
12. Watts, A. B. & Cochran, J. R. *Geophys. J. R. astr. Soc.* **38**, 119–141 (1974).
13. Walcott, R. I. *J. geophys. Res.* **75**, 3941–3954 (1970).
14. Beaumont, C. *Geophys. J.R. astr. Soc.* **55**, 471–497 (1978).
15. Courtney, R. C. thesis, Dalhousie Univ. (1982).
16. Peltier, W. R. *Rev. Geophys. Space Phys.* **12**, 649–664 (1974).
17. Wu, P. & Peltier, W. R. *Geophys. J.R. astr. Soc.* **70**, 435–486 (1982).
18. Carter, N. L. *Rev. Geophys. Space Phys.* **14**, 301–360 (1976).
19. Post, R. *Tectonophysics* **42**, 75–110 (1977).
20. Kirby, S. H. *J. geophys. Res.* **85**, 6353–6363 (1980).
21. McNutt, M. K. & Menard, H. W. *Geophys. J.R. astr. Soc.* **71**, 363–394 (1982).
22. Watts, A. B., Bodine, J. H. & Steckler, M. S. *J. geophys. Res.* **85**, 6369–6376 (1980).

Possible heterogeneity of the Earth's core deduced from PKIKP travel times

G. Poupinet*, R. Pillet† & A. Souriau‡

* IRIGM, University of Grenoble, BP 68, 38402 Saint Martin d'Hères, France
† ORSTOM, 24, rue Bayard, 75008 Paris, France
‡ GRGS, 18, Avenue Edouart Belin, 31055 Toulouse, France

The core of the Earth is usually described by spherically-symmetrical velocity models. The core is made of two main spherical layers: the fluid outer core with a radius close to 3,480 km and a P-velocity increasing with depth from 8 km s^{-1} to 10.3 km s^{-1}, and the solid inner core with a radius of 1,220 km and a P-velocity close to 11 km s^{-1} (refs 1, 2). Station residuals of the seismic core phase PKIKP have been computed for 400 seismological observatories worldwide using 5 yr of the International Seismological Centre (ISC) Bulletins. PKIKP travel times can be corrected for upper mantle propagation by subtracting P delays; thus PKIKP − P residuals are a measurement of the average vertical travel times in the lower mantle and in the core of the Earth beneath seismic stations. A spherical harmonic development of PKIKP − P delays up to degree 4 explains 58% of the variance in the data. PKIKP − P exhibit a latitudinal dependence: polar stations tend to be faster than equatorial stations. We show here that this pattern may reflect a departure from spherical symmetry in the P-velocity distribution in the vicinity of the inner core boundary of the Earth.

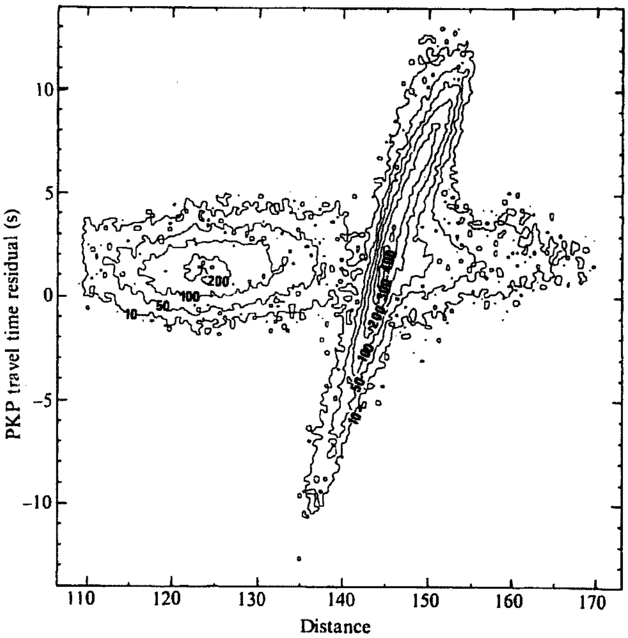

Fig. 1 Travel-time residuals in seconds of the core phase reported as PKIKP by the ISC, as a function of epicentral distance. The PKIKP residuals are computed with respect to Jeffreys–Bullen tables.

The ISC collects readings of P arrival times for all major earthquakes and locates them with Jeffreys–Bullen's tables[3]. Core phases are reported but not used in the computation of the hypocentres; among them, PKIKP is a wave entering the inner core. PKIKP delays are computed with respect to Jeffreys–Bullen's tables for core phases using Shimshoni's[4] depth corrections. PKIKP or PKP (DF) travel times have been extracted from 5 yr of ISC tapes[5] and Fig. 1 shows all phases reported as PKIKP. The coordinates are the difference in arrival time between the observed PKIKP and the theoretical travel times, and the epicentral distance in degrees. The number of PKIKP arrival times in cells of 0.2 s and 0.5° is plotted with level lines at 10, 50, 100, 200, 300 and 400. Four hundred observations are reported at 143° for a PKIKP delay of 1.5 s. The PKIKP diagram should be a horizontal linear ridge with a mean value of 1.5 s. The main feature observed on Fig. 1, beside the expected horizontal DF ridge, is a narrow ridge transversal to the DF branch. Short period precursors to PKIKP in the distance range 120–143° have been interpreted as: (1) one or several refracted branches due to one or several small velocity gradients in the outer core[6–8]; (2) energy diffused by heterogeneities in the lower mantle near the core–mantle boundary[9]; this hypothesis clearly explains the small and weak precursors which are reported at distances <135° and which are spread in time; (3) diffraction from the PKP caustic B[10,11]. Hypotheses (1) and (2) have already been invoked to interpret the transverse ridge from Fig. 1 obtained from limited sets of data[12,13]. The transverse ridge is precisely timed and a ray parameter of 3.4 s per deg is observed. On long period data, the diffraction from PKP at B has a ray parameter of 3.43 s per deg (ref. 14): therefore the most probable interpretation of the transverse branch in the PKIKP reports of the ISC is that it is the short period diffraction from PKP caustic B. Such an interpretation does not exclude the existence of phases related to hypotheses (1) and (2): the ISC only reports first arrivals and, on seismograms, energy is often observed in the time interval between PKIKP and the transverse branch of Fig. 1.

The multibranching in Fig. 1 makes the computation of PKIKP station residuals more complex than the computation of P station residuals. Usually a P or PKIKP residual is an arithmetic mean of travel-time residuals computed for various

0028–0836/83/370204—03$01.00

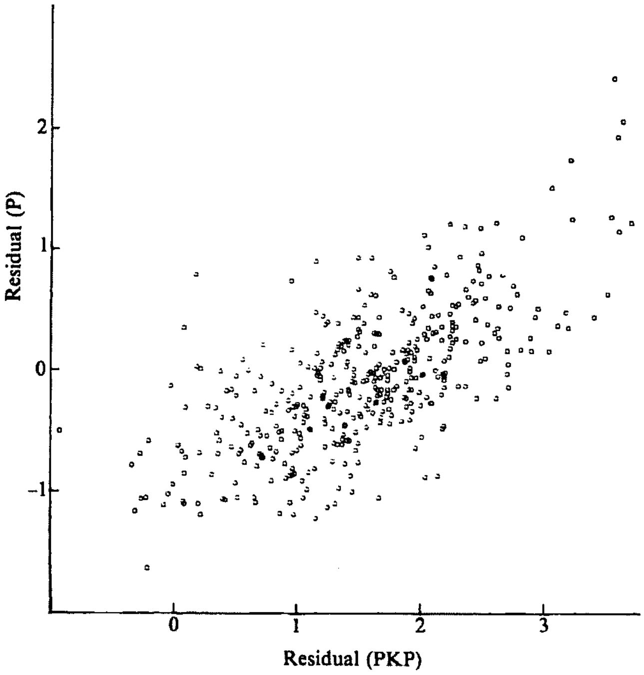

Fig. 2 Correlation between P station residuals and PKIKP station residuals for seismological observatories worldwide. Both axes are graduated in seconds.

earthquakes[15]. Weighting and special tests are applied to eliminate dubious data. To compute PKIKP station residuals from the data presented in Fig. 1, we should eliminate the transverse branch. The first solution would be to eliminate data from 135° to 155°, but the number of data would be too small. To get rid of arrivals that do not belong to the DF branch, we do not retain data from between two parallel lines: $R = 1.6 \times \text{distance} - 223.165$ and $R = 1.6 \times \text{distance} - 240.2$, where R is in seconds and the distance is in degrees. A histogram of residuals is then computed for each station and data outside the +/−3 s interval around the median are rejected. A PKIKP station residual is computed as the arithmetic mean of the remaining data. Figure 2 shows PKIKP station residuals plotted as a function of P residuals[9]: the mean value for PKIKP is 1.5 s and P and PKIKP are correlated within a band of 2 s. A P station delay is essentially a measurement of the vertical P travel time in the upper mantle beneath the station, whereas PKIKP always remains in a narrow cone beneath the station and is sampling the average structure of the upper mantle as well as deeper parts of the Earth.

To correct PKIKP times for propagation in the upper mantle, we compute PKIKP station residual minus P station residual: in theory these delays should give an average propagation time in the lower mantle and in the outer core. Figure 3 is a histogram of PKIKP−P residuals for 400 seismological observations worldwide. The cells are 0.1 s wide. Most PKIKP−P are within an interval of 0–3 s. The seismic stations are referenced by their ISC codes. The errors on PKIKP are of the order of 0.5 s for a station that recorded 100 arrival times and many stations reported more than 100 PKIKP[5].

PKIKP−P are geographically correlated. Late region (larger number) are found in South America, for stations which are not in the same tectonic environment. South American stations report a large number of core phases and their average PKIKP−P is well defined. The trend in the data is to indicate slower propagation times near the Equator compared with faster travel times near the North and South Poles.

The PKIKP field was filtered by performing a spherical harmonics expansion up to a degree 4 (Fig. 4). A simplified pattern very similar to the one observed from the raw data explains 58% of the variance in the data. It exhibits a strong latitudinal dependence: two slow regions are observed near the Equator in South America and near the coast of Brazil for the first region and near the Philippines for the second region. The fastest regions are near the Bering Sea, Alaska and Northern Canada and centred between New Zealand and Antarctica. The travel time difference between slow and fast regions is of the order of 2 s and is resolved by the data. Central Asia and Europe do not show any clear trend.

The first guess, considering such a pattern, is that PKIKP data were not corrected for ellipticity: large travel times would be observed at the Equator and small ones near the poles. But the ISC applies Bullen's[16] ellipticity corrections to all phases reported in the bulletins and therefore there should not be any major latitudinal dependence of the PKIKP residuals. The existence of a systematic bias in PKIKP due, for instance, to a poor distribution of sources and receivers, cannot be disproved. However, results on P residuals, processed in the same manner, show that travel-time delays computed from raw ISC data are

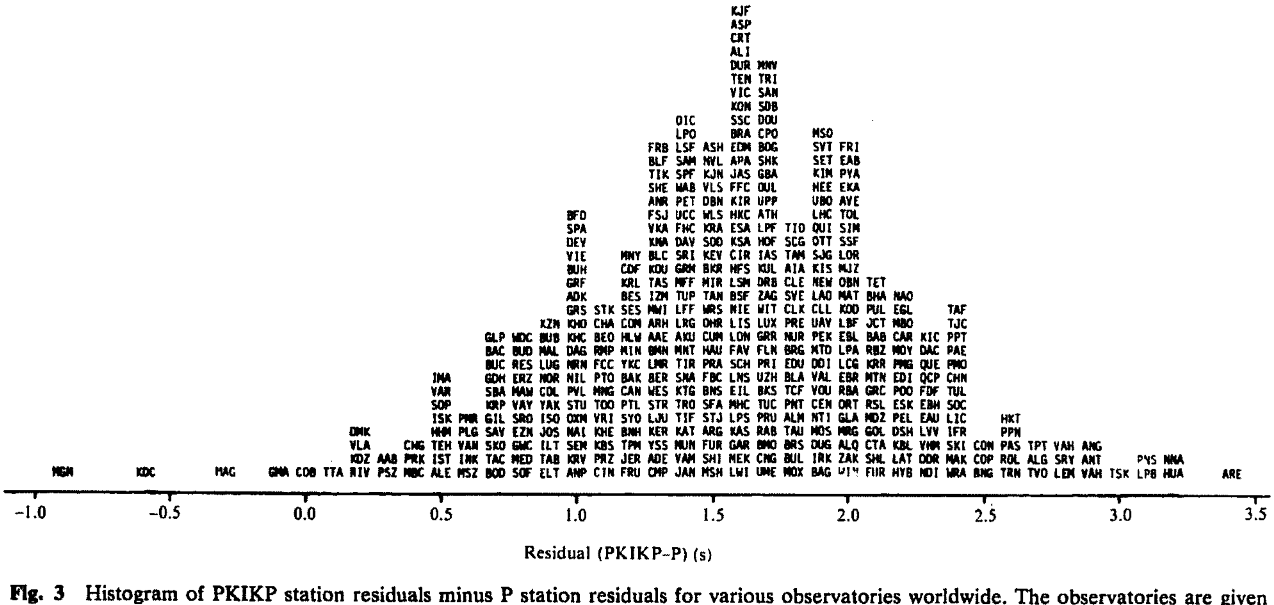

Fig. 3 Histogram of PKIKP station residuals minus P station residuals for various observatories worldwide. The observatories are given by their ISC code. A PKIKP−P residual reflects the average structure deep in the Earth beneath a seismic station.

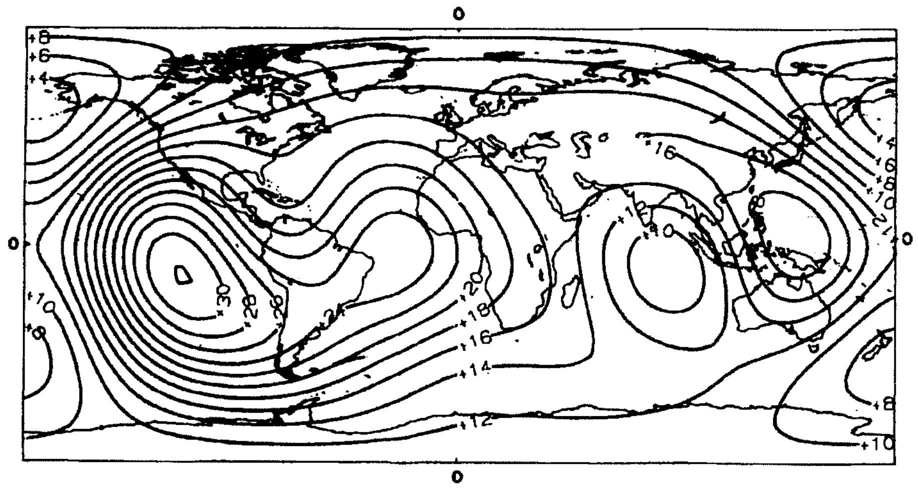

Fig. 4 Spherical harmonics expansion up to a degree 4 of the PKIKP–P travel-time residual field. Values are in tenths of a second. A low number means a fast velocity in the core.

meaningful only when there are sufficient data: similar tectonic regions have the same P delays, whatever their geographical position[17].

From studies of velocity heterogeneities, the upper mantle and a shell of a few hundred kilometres near the core–mantle boundary, in the lower mantle, are known to be the most heterogeneous layers. Our data are corrected for the upper mantle and PKIKP only travel a small part of their path in the lowermost mantle. Dziewonski[18] does not find any evidence for massive vertical flux of material in the lower mantle between the 670-km discontinuity and the core–mantle boundary. The lower mantle would not explain travel-time differences of 2 s in PKIKP: a 1% change in P velocity in the lower mantle in a 100-km layer gives a delay of ~0.1 s. Therefore PKIKP – P are strongly related to the vertical travel times in the core. The observation of precisely timed PmKP, a wave reflected up to seven times in the core, limits the size of lateral heterogeneities in the outer core[19,20]. With the resolution of seismology, the external part of the outer core is very homogeneous. When entering the inner core, PKIKP rays common to a station are spread. The pattern observed cannot be related to the structure deep in the inner core. Therefore it should be explained by heterogeneities near the inner core boundary. The PKIKP – P pattern may be an indication of slow vertical travel times in the vicinity of the inner core boundary near the Equator, compared with fast vertical travel times near the poles. If we assume that such a pattern is related to any kind of convection in the lower part of the fluid core, or to a broad heterogeneous transition region at the inner core boundary[2], this would indicate slow velocity or less dense material in an equatorial belt and more dense material beneath the poles. A prolate inner core would also create the same travel-time pattern: a change of 200 km in the radius of the inner core explains a difference of 2 s in PKIKP: the inner core would be elongated at the poles. The frequent observation of precursors to PKIKP 2–3 s early to the DF branch[21] can be an argument in favour of heterogeneities near the inner core boundary, as for distances near 130°, DF is composed of interferences between PKmIKP[2] and multiple reflections in the inner core would not be precisely timed.

The pattern of the PKIKP – P field filtered up to degree 4 has a resemblance to the intensity of the International Geomagnetic Reference Field (1965)[22]: low isogamma regions would correspond to slow regions. The observations and arguments about the inner core lack substance: the delays observed are of the order of one-thousandth of the total travel time of PKP and disadvantages such as mislocations can obscure our data. New seismological data should be processed to confirm the pattern that we have presented. At this stage, we can only ask if present seismological observations could give us some clues on the convection flow within the core or on the history of the growth of the Earth's core[23].

Received 27 May; accepted 11 July 1983.

1. Dziewonski, A. M. & Anderson, D. L. *Phys. Earth planet. Inter.* **25**, 297 (1981).
2. Choy, G. L. & Cormier, V. F. *Geophys. J. R. astr. Soc.* **72**, 1 (1983).
3. Jeffreys, H. & Bullen, K. E. *British Association for the Advancement of Science* (Gray Milne Trust, London, 1958).
4. Shimshoni, M. *Geophys. J. R. astr. Soc.* **13**, 471 (1967).
5. Pillet, R. thesis, Univ. Grenoble (1979).
6. Adams, R. D. & Randall, M. J. *Bull. seism. Soc. Am.* **54**, 1299 (1964).
7. Bolt, B. A. *Bull. seism. Soc. Am.* **54**, 191 (1964).
8. Buchbinder, G. G. R. *Bull. seism. Soc. Am.* **61**, 429 (1971).
9. Cleary, J. R. & Haddon, R. A. W. *Nature* **240**, 549 (1972).
10. Jeffreys, H. *Mon. Not. R. astr. Soc., Geophys. Suppl.* **4**, 548 (1939).
11. Hai, N. *Ann. Geophys.* **19**, 285 (1963).
12. Agrawal, R. C. *Geophys. J. R. astr. Soc.* **53**, 459 (1978).
13. Andersen, R. S. & Cleary, J. R. *Phys. Earth planet. Inter.* **23**, 207 (1980).
14. Buchbinder, G. G. R. *Bull. seism. Soc. Am.* **64**, 33 (1974).
15. Cleary, J. R. & Hales, A. *J. geophys. Res.* **76**, 7249 (1971).
16. Bullen, K. E. *An Introduction to the Theory of Seismology* (Cambridge University Press, 1963).
17. Poupinet, G. *Earth planet. Sci. Lett.* **14**, 149 (1979).
18. Dziewonski, A. M. *EOS* **63**, 1035 (1982).
19. Engdahl, E. R. *Science* **161**, 263 (1968).
20. Buchbinder, G. G. R. *Earth planet. Sci. Lett.* **14**, 161 (1972).
21. Dorbath, C. *C. r. hebd Séanc. Acad. Sci., Paris* **290B**, 99 (1980).
22. Le Mouel, J. L. in *Traité de Géophysique Interne* (Masson, Paris, 1976).
23. Stevenson, D. J. *Science* **214**, 611 (1981).

BIRPS deep seismic reflection studies of the British Caledonides

Jonathan A. Brewer*, Drummond H. Matthews*, Michael R. Warner*, Jeremy Hall†, David K. Smythe‡ & Robert J. Whittington§

* Bullard Laboratories, Department of Earth Sciences, University of Cambridge, Madingley Road, Cambridge CB3 0EZ, UK
† Department of Geology, The University, Glasgow G12 8QQ, UK
‡ Institute of Geological Sciences, Murchison House, West Mains Road, Edinburgh EH9 3LA, UK
§ Department of Geology, University College of Wales, Aberystwyth, Dyfed SY23 3DB, UK

The Western Isles–North Channel (WINCH) traverse (Fig. 1), an extension of the successful MOIST profile[1], was recorded in 1982 at sea along the west coast of Britain for BIRPS (British Institutions Reflection Profiling Syndicate) by the Geophysical Company of Norway (GECO). The purpose of the WINCH traverse was to study crustal structure of the Caledonian foreland, to cross the Caledonian orogen, and to establish the three-dimensional geometry of mantle reflectors originally seen on MOIST. We describe here the data, first emphasizing British Caledonian structures, and then discussing features of the deep crust of wider significance. The data are of very good quality and contain clear Moho and upper mantle reflections. The lower crust is surprisingly reflective, deeply penetrative thrusts are observed and there is firm evidence that several of the Mesozoic basins round the shores of the United Kingdom were formed by reactivation of older features.

0028–0836/83/370206—05$01.00